Effects of soot-induced snow albedo change on snowpack and hydrological cycle in western United States based on Weather Research and Forecasting chemistry and regional climate simulations

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[1] Radiative forcing induced by soot on snow is an important anthropogenic forcing affecting the global climate. In this study we simulated the deposition of soot aerosol on snow and the resulting impact on snowpack and the hydrological cycle in the western United States. A year-long simulation was performed using the chemistry version of the Weather Research and Forecasting model (WRF-Chem) to determine the soot deposition, followed by three simulations using WRF in meteorology-only mode, with and without the soot-induced snow albedo perturbations. The chemistry simulation shows large spatial variability in soot deposition that reflects the localized emissions and the influence of the complex terrain. The soot-induced snow albedo perturbations increase the surface net solar radiation flux during late winter to early spring, increase the surface air temperature, and reduce the snow accumulation and spring snowmelt. These effects are stronger over the central Rockies and southern Alberta, where soot deposition and snowpack overlap the most. The indirect forcing of soot accelerates snowmelt and alters stream flows, including a trend toward earlier melt dates in the western United States. The soot-induced albedo reduction initiates a positive feedback process whereby dirty snow absorbs more solar radiation, heating the surface and warming the air. This warming causes reduced snow depth and fraction, which further reduces the regional surface albedo for the snow-covered regions. For a doubled snow albedo perturbation, the change to surface energy and temperature is around 50–80%; however, snowpack reduction is nonlinearly accelerated.


1. Introduction

[2] The presence of soot particles in snow can reduce the snow albedo and affect snowmelt [Warren and Wiscombe, 1980, 1985; Hansen and Nazarenko, 2004; Jacobson, 2004; Flanner et al., 2007]. The 2007 Intergovernmental Panel on Climate Change (IPCC) report listed the radiative forcing induced by “soot on snow” as one of the important anthropogenic forcings affecting climate change between 1750 and 2005 [IPCC, 2007]. Therefore, quantifying and reducing the albedo errors due to this effect is a priority for improving simulations of climate change and the hydrological cycle using climate models.

1.1. Soot in Snow

[3] Soot is produced by incomplete combustion of carbonaceous material, mainly fossil fuels and biomass. Black carbon (BC) is the main component of atmospheric soot particles, which typically have a diameter of around 0.1 μm. “Soot” has been used in different contexts in different studies; in this paper we mean BC when we use the term soot. Because of their dark color, soot particles absorb solar radiation, convert it into internal energy, and emit thermal-infrared radiation, therefore warming the air [Jacobson, 2001]. Soot particles generally are hydrophobic but can mix internally or externally with other hydrophilic aerosols, such as sulfate. Soot particles are removed from the atmosphere within days to weeks by rainout, washout, and dry deposition.

[4] Optical and electron microscopes show that a typical snow crystal contains thousands of particles, including soot and dust [Chylek et al., 1987]. It is hypothesized that wet deposition (via snow and rain) is the primary mechanism for transferring soot from the atmosphere into the snowpack. Dry deposition also can be significant, accounting for several tens of percent of the total deposition [Davidson et al., 1985]. The deposition source of soot into the snowpack is countered by the sink of snowmelt. Typical mixing ratio values for
soot in snow range from 5 to more than 60 ng_{soot} g_{snow}^{-1}, as shown in Table 1.

1.2. Snow Albedo

[5] Surface albedo is an important surface characteristic that determines energy transfer at the surface. Major perturbations to the albedo can be caused by snow over very short timeframes. Snow cover can change the albedo of grassland by a factor of 3–4 and forested regions by a factor of 2–3 [Betts and Ball, 1997; Thomas and Rowntree, 1992]. This has been emphasized by the results of many albedo sensitivity intercomparisons [Cess et al., 1991; Randall et al., 1994]. Additional studies, [e.g., Barnett et al., 1988; Walland and Simmonds, 1996] have shown that the long-term average of snow accumulation or melt patterns may significantly alter regional climate and have a strong impact upon the general circulation. Current land surface models used in climate models have a wide range of methods for parameterizing snow albedo [Barry, 1996; Schlosser et al., 2000]. Different snow albedo parameterizations can produce significantly different snow mass and snowmelt timing as demonstrated in both off-line studies [e.g., Yang and Niu, 2000] and regional climate simulations [e.g., Lynch et al., 1998].

[6] Many factors affect snow albedo, with the effect of soot significantly stronger than the effect of snow grain size [Conway et al., 1996]. For example, the visible snowpack albedo can be reduced from 0.95 for pure snow to 0.1 for dirty snow with 100 ppm by weight (ppmw) of soot. For optically thick snowpack, the broadband visible albedo depends most strongly on the soot content, especially when soot is at the top of snowpack.

1.3. Effects of Soot in Snow on Radiative Forcing

[7] Hansen and Nazarenko [2004] used observations to highlight the importance of soot-snow forcing. They used global-scale estimates of soot concentrations within snow and ice to deduce the surface albedo perturbation, which results in a positive radiative forcing (RF) of 0.15 W m$^{-2}$ and global warming of 0.24°C, yielding a high forcing efficacy [Hansen et al., 2005]. Jacobson [2004] developed a global model that allows the soot to enter snow via precipitation and dry deposition, thereby modifying the snow albedo and emissivity. He found that the simulated concentrations of soot within snow were in reasonable agreement with those from observations. His modeling study showed that soot on snow and sea ice can cause a decrease in the surface albedo of 0.4% globally and 1% in the Northern Hemisphere. Flanner et al. [2007] coupled a snow, ice, and aerosol radiative model to a GCM with prognostic carbon aerosol transport to estimate present-day climate forcing and response from BC in snow. They estimated a global annual mean BC-snow surface RF of 0.049 to 0.054 W m$^{-2}$ and changes to global-mean 2-m air temperature of 0.10°C–0.15°C. They suggested that the anthropogenic contribution to the total soot-snow forcing is at least 80%.

[8] However, all these studies are global in scale and employ coarse spatial resolutions, which cannot resolve the significant orography-related features of snowpack and aerosol deposition [Ghan and Shipert, 2006]. It is well known that snowpack has a strong regional dependence on surface elevation [Marshall et al., 2003]. Furthermore, these studies only estimated the soot-snow induced RF and resulting temperature response and did not investigate the hydrological impacts, such as snowpack and runoff perturbations over regions with complex topography and inhomogeneous land cover such as the western United States (WUS). Understanding the dynamic physical processes controlling snowpack in these regions is of utmost importance for managing water resources [Leung et al., 2003]. In the WUS, mountain snowmelt accounts for more than 70% of the annual stream flows that support irrigation in the semiarid Central Valley and Columbia Basin, and hydro-power generation and navigation in the major river basins. Analysis of measurements has revealed that snowpack levels have dropped considerably throughout the WUS since the 1950s [Mote et al., 2005]. Further reductions in snowpack amount are predicted under plausible climate change scenarios [e.g., Leung et al., 2004].

[9] Some studies have attributed the snow retreat and reduction in the snowfall/precipitation ratio to warming trends in the WUS over the past 50 years [Mote et al., 2005; Feng and Hu, 2007]. The warming trends are consistent with global climate simulations that included greenhouse gas forcing in the past [Christensen et al., 2007]. In this study a potential secondary contributing cause of snowpack change is investigated: the heating of snow through decreased albedo caused by the deposition of soot. This accelerates snowmelt and reduces snowpack. In the WUS, much of the soot available for deposition into snow comes from populated regions west of the mountains. During winter, the prevailing westerly to southwesterly flow transports moisture from the Pacific Ocean as well as anthropogenic aerosols from coastal metropolitan areas into the higher elevations where they are deposited. Vehicular and ship emissions contribute much of the soot production in the WUS [Cooke et al., 1999; Reddy and Boucher, 2007].

[10] This study consists of a series of model simulations. It begins with a year-long simulation using the chemistry version of the Weather Research and Forecasting model (WRF-Chem) to simulate an annual cycle of soot aerosol
deposition on snow. This is then used to estimate snow albedo perturbations induced by the soot within the WUS. Then, three regional climate simulations are performed using WRF with the chemistry turned off, but with or without the perturbed snow albedo based on the WRF-Chem simulation.

2. WRF-Chem Simulation for Soot Deposition

2.1. WRF-Chem

[11] In order to address scientific questions related to meteorological-aerosol-radiation-cloud feedbacks at the urban to regional scale, scientists at the Pacific Northwest National Laboratory (PNNL) have developed substantial portions of WRF-Chem [Grell et al., 2005] during the past 3 years. For more detailed descriptions of the changes, see Fast et al. [2006], Gustafson et al. [2007], and Chapman et al. [2008] and the references therein. In summary, the contributions include: the CBM-Z gas-phase chemistry mechanism; the MOSAIC sectional aerosol module; the Fast-J photolysis module; feedbacks between aerosols, clouds, and radiation; activation/resuspension and wet scavenging; aqueous chemistry; and extending the nesting capability of WRF to include the chemistry scalars. Using CBM-Z and MOSAIC, WRF-Chem simulates 55 prognostic gas species and eight aerosol species plus the aerosol number, water content, and an estimate of hypsometric water content. Aerosol species are assumed to be internally mixed and are explicitly carried in two forms, one representing dry/interstitial conditions and one representing particles inside cloud droplets. For this study, WRF-Chem has been further modified to track the deposition budget of BC. Variables have been added to save the deposition mass of BC by size bin for three categories: BC wet deposition concurrent with rain, BC wet deposition concurrent with snow, and BC dry deposition. Each of these quantities was archived hourly during the simulation. The partitioning between rain and snow is determined from the surface temperature, as determined by the Noah land surface model [Chen and Dudhia, 2001], which is used for all the WRF simulations in this study. Precipitation is assumed to be rain if the surface temperature is above freezing and snow if it is colder.

[12] For the WRF-Chem simulation presented here, the MOSAIC aerosol module was configured with four sectional size bins: three size bins for particles up to 2.5 μm diameter and one bin for particles with diameters between 2.5 and 10 μm. This is half the number of bins used in previous WRF-Chem research applications using the MOSAIC sectional aerosol module [Fast et al., 2006; Gustafson et al., 2007; Chapman et al., 2008], but is necessary to reduce computational cost for the climate study presented in this paper. Even with this compromise, the simulation cost is substantial and requires limiting the chemistry run to a single year. This is the first WRF-Chem study using the fully coupled aerosol-cloud parameterizations for extended, climate-length simulations. In order to get proper climate statistics, another three simulations lasting for 5 years are done using the meteorological-only settings of WRF (called WRF-RCM to denote regional climate simulations with WRF alone), as described in section 3, to generate climatologically relevant statistics for the impact of the soot-induced albedo perturbation on climate and the hydrological cycle. [13] The WRF-Chem domain extends over the WUS from the Pacific Ocean to the eastern side of the Rockies (see Figure 1). Meteorological boundary conditions have been provided as in the extended climate simulations described in section 3.2. Coupling between the aerosol and cloud fields is accomplished via feedbacks in the Lin et al. [1983] microphysics scheme and aerosols with radiation in the Goddard Space Flight Center shortwave scheme [Fast et al., 2006]. [14] Trace gas and particulate emissions were compiled from the National Emission Inventory 1999 (NEI99) [U.S. Environmental Protection Agency, 2003] by Stuart McKee of NOAA and provided to the WRF-Chem user community. The input emissions data set and details regarding it can be found at http://ruc.fsl.noaa.gov/wrf/WG11/anthropogenic.htm. As provided to WRF-Chem users, the NEI99 emissions data set employs a 4-km grid spacing and a repeating diurnal cycle representing a typical summer day. For this study, the emissions were regridded to the WRF-Chem domain and the NEI99 chemical speciation mapped to CBM-Z and MOSAIC species and bins (see Figure 1). Although these emissions reflect summertime conditions, the seasonal cycle for anthropogenic emissions in the WUS is small, with 25% of annual emissions apportioned each season in other, global inventories that also utilize U.S. EPA data [e.g., Benkovitz et al., 1996 as modified by E. C. Voldner et al. (http://www.geiacenter.org and http://www.ortech.ca/cgeic/poster.html). Thus, use of the WRF-Chem-supplied inventories for alternate seasons should not greatly impact the results of this study since they are intended to be more qualitative than quantitative given the uncertainties described elsewhere in the paper. Additionally, it should be noted that alternative global emission inventories exist, for example, that of Bond et al. [2004] at 1° × 1°, but they are coarser and unable to capture the spatial complexity required for the high-resolution simulations in this study. Emissions from forest fires have not been included since fire locations vary substantially from year to year. This results in an underestimate of BC emissions, but is preferable to biasing the soot deposition to particular regions with a year-specific fire pattern. Also, most of the fires occur during summer and fall when snow has already melted and is not accumulating. Ship emissions also were omitted. Although on a global basis emissions from oceangoing ships should not be neglected, such emissions are known to be at a minimum during winter months [Corbett et al., 1999]; furthermore their impact is expected to be minimal compared to other anthropogenic emissions for the modeling domain used in this study.

2.2. Soot Concentration in the Atmosphere

[15] The continuous WRF-Chem simulation, from 1 September 2003 to 31 October 2004, provides a complete annual cycle of atmospheric soot concentration and deposition after neglecting the first month as spin-up. We compared the simulated near-surface concentration of BC with observations from the IMPROVE network (http://vista.cira.colostate.edu/IMPROVE/). The spatial plots for each season (not shown) exhibit the strong BC gradients in the WUS. Much of this is due to high-emission regions near areas of complex topography. The complex terrain preferentially forces the pollutants into areas such as the California Central Valley where they accumulate. In general, WRF-Chem reproduces the spatial variation seen in the observations,
but it also reveals how gaps in the IMPROVE network make a thorough comparison difficult because of the gradients. For example, many of the stations in California are at higher elevations in the Sierra Nevada along the edge of the strong concentration gradient. Small location errors in the simulated gradient lead to large differences compared to the observations.

[16] The spatial plots also reveal that the modeled BC concentration is lower in DJF than for the other seasons over much of the domain with the exception of the Central Valley, which has higher concentrations. By comparison, the IMPROVE data also have lower concentrations during DJF, but the difference between the model and data is greater during this season. In general, the model overpredicts BC every month, as seen in Figure 2, which shows monthly averaged BC concentrations for the IMPROVE stations within the domain along with the coincident WRF-Chem grid point averages. The minimum, maximum, 25th percentile, median, and 75th percentile values are all higher in the model than in the observations with the exception of the 75th percentile in October and the maximum in June. The high maximum value observed in June is due to a forest fire, which is not included in the model.

2.3. Soot Deposition

[17] To calculate the soot content within snow, black carbon (BC) deposition is tracked in WRF-Chem as the model integrates forward in time using the deposition budget variables described above. The separation of wet deposition into two categories, with rain or with snow, provides information for parameterizing how carbon accumulates within the snowpack, either depositing on the surface, possibly below the surface, or in the snow crystal.

[18] Figure 3 shows the BC deposited for DJF and MAM. The total deposition of BC is spatially correlated with BC emissions; the largest deposition occurs around large cities, for example, in coastal Southern California, the San Francisco Bay area, and Calgary. Nevada has the least deposition of the states included in the model domain; this is due to both lower emissions and very little rain. In general, the dry deposition (not shown) is larger over the coastal metropolitan areas, for example, Seattle, Portland, Los

Figure 1. (a) Model domain and subregions, (b) elevation (unit: 10 m), and (c) daily black carbon emissions for particle diameter <10 \( \mu \text{m} \) (unit: mg m\(^{-2}\) d\(^{-1}\)).

Figure 2. Monthly averaged BC concentrations for the IMPROVE stations within the domain along with the coincident WRF-Chem grid point averages. The box and whisker plot shows the minimum and maximum value with the bar, the 25th and 75th percentiles with the box, and the 50th percentile by the bar within the box.
Angeles, and San Francisco, as well as a few large cities in southern Canada. The wet deposition (not shown) is more evenly distributed spatially than the dry deposition. However, the wet deposition has a strong seasonal cycle mimicking the precipitation cycle, whereas the dry deposition is more similar from month to month because we neglect the seasonality of emissions. Also, strong orographic influences are seen in the wet deposition owing to orographically forced precipitation, for example, over the coastal mountains of the Pacific Northwest and portions of the Rockies. The partitioning of the precipitation between rain and snow also makes a difference, particularly when determining how much of the BC remains in the snowpack.

2.4. Soot Content in Snow and Induced Albedo Perturbation

The soot content in snow is calculated on the basis of the simulated time series of snow depth and soot deposition using a methodology derived from the algorithms of Jacobson [2004]. Archived hourly output from WRF-Chem is used in an offline program to determine the soot mixing ratio in snow. No attempt has been made to treat different particle diameters differently; deposition for all size bins has been combined into net deposition values.

The algorithm for determining the mixing ratio of soot in snow, $r_{BC}$ with units of ng_BC g$_{snow}$, is summarized as follows. First, the mass of snow deposited during the previous period is determined. For the hourly data here, this time increment equates to $\Delta t = 3600$ s. For dry deposition, the incremental mass addition of soot per unit area, $\Delta m_{BC,d}$, is simply the sum of the dry deposition from the model output over the time $\Delta t$. The incremental mass addition for wet deposition per unit area, $\Delta m_{BC,w}$, is dependent on whether the snow depth increased or decreased during $\Delta t$. If snow is added or stayed the same during $\Delta t$, $\Delta m_{BC,w}$ is the sum of wet deposition concurrent with snow during $\Delta t$. If the snowpack thinned, then any BC that deposited with snow during $\Delta t$ is instead added to $\Delta m_{BC,d}$. Any BC that deposited with rain is assumed to instantly wash out of the snowpack.

Next, the change in mixing ratio of soot in snow due to $\Delta m_{BC,d}$ versus $\Delta m_{BC,w}$ is determined. The mixing ratio of soot in snow due to dry and wet deposition, $r_{BC,d}$ and $r_{BC,w}$, respectively, is tracked separately over time and the sum of the two mixing ratios results in $r_{BC}$. The value $r_{BC}$ represents the soot mixing ratio for the radiatively most important portion of the snowpack, which we refer to as the effective snow depth, $D'$, and is defined as the minimum of the actual snow depth, $D$, or $D'_{max} = 10^{-3}$ m following Jacobson [2004]. Also, $r_{BC}$ is only determined for cells with $D > 10^{-3}$ m to maintain stability in the algorithm. The formulas and methodologies calculating $r_{BC,d}$ and $r_{BC,w}$ are similar to equations (1) and (2) of Jacobson [2004], and the final soot mixing ratio in snow is given by $r_{BC} = r_{BC,d} + r_{BC,w}$.

The resulting values for the soot mixing ratio in snow are shown in Figure 4. The values are around 10–80 ng g$^{-1}$ in winter over the WUS and southern Alberta with the maximum simulated values larger than 100 ng g$^{-1}$ over some small regions such as eastern Idaho and Calgary. These values are in accord with observations and GCM simulations as shown in Table 1.

Once $r_{BC}$ is known, it can be used to estimate the change in snow albedo due to BC. The DJF average of $r_{BC}$ is calculated for each model point. For the few points that have very unrealistic $r_{BC}$ values, the smoothed seasonal average is capped with a value of 500 ng g$^{-1}$, which is the upper limit of the values used to determine the albedo formula below. The result, $r_{BC,DJF}$, is then used to estimate the snow albedo perturbation using the formula

$$A' = 0.5 \times \left( 2.2883 \cdot 10^{-9}r_{BC,DJF} - 2.2314 \cdot 10^{-6}r_{BC,DJF}^2 + 8.5369 \cdot 10^{-4}r_{BC,DJF} \right).$$

This formula is a third-order polynomial fit to the points from Jacobson’s [2004] Figure 1c at a wavelength of 500 nm, which is near the most sensitive spectral region of snow albedo change from BC. The factor of 0.5 is applied to
account for the fact that BC-induced change in snow albedo for broadband solar is approximately half as great as the change at 500 nm, on the basis of Figure 1c of Jacobson [2004], Figure 2 of Warren and Wiscombe [1985], and Figures 2 and 3a of Flanner et al. [2007].

The perturbation represents the difference between the snow albedo for a given soot-snow mixing ratio and when no soot is present. Surface albedo refers to an average for solar broadband in the Noah Land Surface Model (LSM) in WRF. No attempt is made to represent albedo perturbations from snow aging since the Noah soil model does not track snow age. Since both WRF-RCM simulations, with and without the snow albedo perturbation, neglect this physical process, comparing the difference between the simulations still provides meaningful results for the impact of BC, but may have neglected the possible accelerated aging process induced by soot-on-snow effect. It should also be noted that only the DJF season was used to determine the soot content in snow owing to the time-independent handling of maximum snow albedo in the Noah LSM (described below). The net result, based on the BC mixing ratios (Figure 4a) and equation (1), is a 0.5–3% perturbation to the snow albedo over most of the northwestern states during winter (Figure 4b). The maximum simulated reduction in snow albedo reaches 3.5–5.5% over areas coincident with the maximum soot content in snow.

3. WRF-RCM Control Simulation and Evaluation

3.1. Evaluation Data

To simulate the impacts of soot-induced snow albedo change on the hydrological cycle, it is important to realistically capture precipitation, snowpack, and runoff in the control simulation. The following data sets are used for evaluating model simulations.

3.1.1. Surface Air Temperature and Precipitation

The UW/PRISM data set consists of daily maximum and minimum surface temperature and precipitation for 1949–2000. It was developed by the Surface Water Modeling Group at the University of Washington following the methodology outlined by Maurer et al. [2001] based on daily observations made at the National Oceanic and Atmospheric Administration (NOAA) Cooperative Observer (Co-op) stations, with topographic adjustment based on the PRISM precipitation climatology [Daly et al., 1994].

3.1.2. Snow Water Equivalent (SWE)

Three different snow observation data are used in this study to illustrate the uncertainty in mountain snowpack observations and the effects of spatial resolution. The first one, SNOTEL, is point measurements of SWE at more than 550 snow telemetry stations in the WUS between 1981 and 2000 [Leung and Qian, 2003]. The second SWE data set was developed by the National Operational Hydrologic Remote Sensing Center (NOHRSC). NOHRSC uses a statistical methodology [Hartman et al., 1995] to combine station measurements of snow water, satellite estimates of the snowline, and a digital surface elevation model to produce a gridded distribution of snow water at a resolution of 1.5' for WUS. The third SWE data set, developed at the Canadian Meteorological Centre (CMC), is daily snow depth in water equivalent at 0.25' resolution over North America for the period 1979–1996 [Niu and Yang, 2007]. The gridded snow depth combines in situ daily observations from ~8000 U.S. cooperative stations and Canadian climate stations and first-guess fields with an optimum interpolation scheme developed by Brown et al. [2003].

3.1.3. Runoff

The runoff data set used in this study is the University of New Hampshire (UNH)-Global Runoff Data Centre (GRDC) global runoff climatology. The UNH-GRDC data set provides monthly climatological runoff fields, which are runoff outputs from a water balance model driven by observed meteorological data and then corrected with the runoff fields that are disaggregated from the observed river discharges. Although a no-time-delay assumption is applied when the gauge-observed discharge is distributed uniformly over a catchment, the resulting runoff fields over a large river basin approximate the real runoff, especially in river basins that contain a sufficiently dense network of rain gauges [Fekete et al., 2000].

3.2. WRF-RCM Configuration

The regional climate model used in this study is based on the Weather Research and Forecasting (WRF) model [Skamarock et al., 2005], with modifications described by Leung et al. [2005], which includes the use of a wider
buffer zone for the lateral boundaries, with the relaxation coefficients of the nudging boundary conditions following a linear-exponential function, updating of the background surface albedo to include seasonal changes, updating of time-dependent sea surface temperature and sea ice from the lower boundary conditions, and estimation of cloud fraction based on work by Xu and Randall [1996]. A series of sensitivity experiments are done to determine the parameterization schemes most fit for regional climate simulation in WUS. The resulting parameterization choices include the CAM3 shortwave and longwave radiation [Kiehl et al., 1996], the modified Kain-Fritsch convection parameterization [Kain and Fritsch, 1993; Kain, 2004], the WRF Single-Moment 6-class cloud microphysics scheme [Hong et al., 2006], the Yonsei University (YSU) boundary layer scheme [Hong et al., 2006], and the Noah land surface model [Chen and Dudhia, 2001].

[31] The model domain encompasses all states over the WUS and extends into the Pacific Ocean (see Figure 1a). It is centered at 40.0°N and 114.0°W, with a horizontal grid spacing of 15 km, and it includes 35 vertical levels extending up to 100 hPa. The boxes in Figure 1a encompass the regions analyzed in this paper, and roughly contain the boundaries of the Columbia River Basin (CRB), the Sacramento—San Joaquin (SSJ) River Basin, the Central Rockies (CR), and the Sierra Nevada (SN) mountains. Figure 1b shows the model topography. At 15-km grid spacing the Rocky Mountain and coastal ranges, such as the Sierra Nevada and the Cascades, are well resolved. Several sensitivity experiments were performed to determine the impact of snow emissivity (SE) on the simulation of climate over the WUS. The simulated snowpack and temperature were found to be sensitive to the value of SE (e.g., set from 0.90 to 1.0). SE is assumed to be 0.98 in this study, since the typical values of snow emissivity range from 0.96 to 1.00 [e.g., Pelke, 1984].

[32] Initial and lateral boundary conditions of the simulation were derived from the NCEP-DOE Global Reanalysis II [Kanamitsu et al., 2002]. Sea surface temperature (SST) and sea ice from AMIP-II [Taylor et al., 2000] were used to construct the lower boundary conditions. The simulation was initialized on 1 September 1993 with the lateral and lower boundary conditions updated every 6 h through 31 December 1998.

3.3. Surface Albedo in Noah LSM

[33] The Noah LSM, based on the OSU LSM described by Chen and Dudhia [2001], was used in this study. As described in section 1, surface albedo is a critical parameter that affects the net solar radiation that drives the terrestrial surface energy and water budget [Qu and Hall, 2006]. The following formula is used in the Noah LSM to calculate the surface albedo, \( \alpha \):

\[
\alpha = \alpha_0 + [1.0 - (f_{\text{veg}} - f_{\text{minveg}})] \times f_{\text{snow}} \times (\alpha_{\text{maxsnow}} - \alpha_0),
\]

where \( \alpha_0 \) refers to the background snow-free surface albedo (in fraction), which is updated by temporally interpolating between the prescribed vegetation-dependent background surface albedo for the summer and winter; \( \alpha_{\text{maxsnow}} \) is the maximum albedo over deep snow; \( f_{\text{veg}} \) is the fractional areal coverage of green vegetation; \( f_{\text{minveg}} \) is the minimum fractional areal coverage of green vegetation; and \( f_{\text{snow}} \) is the fractional snow cover calculated as

\[
f_{\text{snow}} = 1.0 - (e^{-S_a} - R_c),
\]

where \( S_a \) is a tuning parameter (set to 2.6); \( R_c = S_{\text{veg}}/S_a \); \( S_{\text{veg}} \) is the snow depth in SWE; and vegetation-dependent \( S_a \) is a threshold snow depth in SWE that implies 100 percent snow cover. Hence, surface albedo (\( \alpha \)) is calculated on the basis of a simple weighting of the background albedo (\( \alpha_0 \)) and the maximum albedo over deep snow (\( \alpha_{\text{maxsnow}} \)) based on the fractional snow free and snow cover area within a model grid cell.

[34] To limit the values of surface albedo when snow is present, the Noah LSM inputs the measured maximum albedo over deep snow [Robinson and Kukla, 1985], which has a spatial resolution of 1° by 1° and is produced from 109 measured scenes from the winter of 1978–1979.

[35] Figure 5a shows the spatial distribution of the maximum snow albedo (\( \alpha_{\text{maxsnow}} \)) used in the default configuration of WRF.

3.4. Results of Control Simulation

3.4.1. Precipitation

[36] Figures 6a and 6b show the simulated and observed seasonal mean precipitation averaged between December 1993 and November 1998. During the cold season, the observed precipitation at 1/8° resolution shows distinct spatial distributions that resemble the complex orography. The two precipitation bands along the west coast correspond to orographic precipitation associated with the coastal range and the Cascades and the Sierra Nevada that are further inland. East of these mountains, precipitation decreases sharply in the basins and the intermountain zone, before it increases slightly again over the Rockies.

[37] The simulated cold season precipitation clearly has a spatial pattern similar to the observations. The two precipitation bands near the coast are very distinguishable, with precipitation decreasing in the valleys between the two elevated regions. In addition, precipitation on the east side of the Cascades and the Sierra Nevada is significantly less.

[38] Although there is a one-to-one correspondence between most areas of maximum precipitation in the observations and the control simulation, the simulated orographic precipitation along the windward slopes of the Cascades and Sierra Nevada is too strong. This is consistent with results based on real time weather forecasts for the Pacific Northwest [e.g., Colle et al., 2000; Westrick and Mass, 2001] and MM5 regional climate simulation [Leung and Qian, 2003] that show an increasing amount of area averaged precipitation as spatial resolution increases.

[39] During spring, the precipitation pattern is similar but with a smaller magnitude than in winter. Again, WRF overpredicts the orographic precipitation along the Cascades and Sierra Nevada as well as the Rockies. Figure 7 shows the monthly time series of precipitation from December 1993 to November 1998 averaged over CRB and SSJ. WRF captures the seasonal cycle and interannual variability of precipitation very well over the two basins. The annual mean precipitation is 2.21 and 2.60 mm d\(^{-1}\) over CRB and SSJ, respectively, and model overpredicts precipitation by
35% and 37% over the two basins. This behavior is similar to the MM5 model that tends to have higher large-scale moisture convergence during the cold season when driven by global reanalysis at high spatial resolution [Leung et al., 2003].

### 3.4.2. Temperature

[40] Figures 6c and 6d show the seasonal mean surface air temperature averaged for December, January, and February (DJF) based on the WRF control simulation and observation. The spatial distribution of large-scale features is very consistent between the two. The mesoscale details of temperature are also captured over the mountains in the WRF simulation at relatively high spatial resolution. The magnitude of simulated temperature during the DJF and MAM seasons (not shown) is very close to the observed, with the exception of a small cold bias over the Rockies. As shown in Figure 7 the basin-averaged temperature is slightly underestimated during winter and spring, but the cold bias is generally less than 2°C over CRB and SSJ basin. Surface air temperature is simulated very well during summer and fall. The annual mean temperature is 5.09° and 12.40°C over CRB and SSJ, respectively, and the cold bias of model is only 0.70° and 0.78°C over the two basins.

### 3.4.3. Snow Water

[41] Figures 8a, 8b, 8c and 8d show the simulated and observed distribution of DJF mean snow depth in terms of snow water equivalent. We present three observational data sets of SWE to provide an estimate of uncertainty in the observed quantity. Because of the improved simulation of surface temperature gained through more accurate elevation in the regional model with higher spatial resolution, the phase of the precipitation is more accurate than in a GCM with coarse resolution. This results in improved snow accumulation and melting processes at the surface and the SWE is generally well simulated. The simulation reproduces many of the features of the observed snow distribution, with heavy snow cover over mountains (e.g., Cascades and the Sierra Ranges, and Rockies) and light snow in the valleys (e.g., Central Valley) or basins (e.g., CRB). On the basis of the SNOTEL station data, the CMC data show smaller SWE over both high- and low-elevation areas. Since the NOHRSC data have higher spatial resolution than CMC and include station measurements of snow water in their product, the magnitude of SWE from NOHRSC is very close to the SNOTEL data where measurements are available.

[42] Figures 8e and 8f show the simulated and observed spatial distribution of mean SWE in MAM. Compared to DJF, the snow cover area has significantly decreased in spring owing to snowmelt at the lower elevations. However, the SWE is much higher in MAM than in DJF over the mountains with higher elevation (e.g., the Cascades, Sierra Nevada, and Rockies). During MAM, the model captured the maximum SWE over the Sierra Nevada range and northern Rockies. Since the precipitation is overpredicted in the Cascades, Sierra Nevada, and Rockies, the overpredicted liquid rain on snow may have increased the melt process and runoff (see Figure 9).

### 3.4.4. Runoff

[43] The spatial distribution of modeled runoff generally agrees with the GRDC runoff in winter (not shown). Both the modeled and observed data show higher runoff in wet regions such as the high elevations, which is at least partially related to the overestimated orographic precipitation. Further inland, runoff rapidly decreases coincident with decreasing precipitation in the rain shadows. The main contribution to total runoff during winter is surface runoff generated by liquid rain. The discrepancy between the modeled and observed runoff may also result from the inconsistency of time periods used for averaging. GRDC runoff is estimated through longer past discharge records, while the modeled runoff is an average over the years 1993–1998.

[44] During spring both precipitation and snowmelt contribute to runoff. Generally, the spatial distribution of large-scale features is very consistent between the observa-
4. Simulation of Soot-Induced Snow Albedo Change

To determine the impact of soot-induced changes to snow albedo, a second WRF-RCM simulation, named EXP1, was done using the same settings as the control simulation except that the snow albedo was modified on the basis of soot deposition determined from the WRF-Chem simulation. This was done by modifying the map of maximum snow albedo (Figure 5) used by the Noah LSM, which was described in section 3.3. The albedo perturbation, $A'$, is subtracted from $\alpha_{\text{max,snow}}$. We compared the results of these two 5-year-long WRF regional climate simulations with and without soot-induced snow albedo perturbations, to investigate the effects of soot on the surface energy and water budget in the WUS. The following analyses focus on the changes of variables such as surface solar radiation, temperature, precipitation, evaporation, soil moisture, snowpack, and runoff. These variables are the most relevant to the hydrological cycle and water resources in the WUS.

4.1. Surface Solar Radiation

Figure 10a shows the change in net surface shortwave radiation flux (NSW) for March between the control and soot-sensitivity simulations. The change of NSW is spatially correlated with the change of surface albedo (Figure 10d). The largest changes are over the central Rockies and southern
Alberta, where NSW increases 2–7 W m$^{-2}$ in late winter and 5–12 W m$^{-2}$ in early spring. As seen in Figure 11, the NSW increases in winter and spring, and the maximum change occurs in March. The NSW in March increases 3.6 W m$^{-2}$ over the Central Rockies (CR) and 1.1 W m$^{-2}$ over the Sierra Nevada (SN), respectively (see Table 2a). The change of NSW is caused mainly by the perturbed surface albedo rather than changes to the downward solar radiation flux. As shown in Table 2a the surface NSW increases 2.7% in March over CR and 0.7% over SN, respectively. However, downward solar radiation at the surface only changes 0.3% over CR and 0.1% over SN, respectively. It can be seen from Figure 11 that the albedo change is similar between December and March over both river basins; however, NSW increases significantly during this period and reaches a peak in March because of increased downward solar radiation resulting from the increased solar height angle.

4.2. Surface Air and Skin Temperature

[47] Skin (2-m air) temperature increases by 0.2$^\circ$–1.4$^\circ$C (0.1$^\circ$–1.0$^\circ$C) over the majority of the snow-covered areas in the WUS during late winter to early spring. The temporal and spatial distributions of temperature changes are mainly driven by the change of surface absorbed solar radiation, which is correlated with the NSW change. Similar to the NSW changes, the warming period is from November to May and the maximum change occurs in March. Figures 10b and 10c show the spatial distribution of changes for 2-m air and skin temperatures, respectively, averaged for March. Regional averaged increases of skin (2-m air) temperature are 0.43$^\circ$C (0.34$^\circ$C) over CR and 0.09$^\circ$C (0.06$^\circ$C) over SN, respectively. The increase in surface skin temperature is 20–50% higher than that of surface air temperature owing to partitioning of the energy to multiple sinks, such as sensible and latent heat, melting the snow, and warming the subsurface soil layers.

4.3. Snowpack

[48] As a result of soot-induced reduction in NSW and surface warming, there are significant reductions in snowpack between December and May. Figure 12 shows the spatial distribution of SWE change for March between the control and soot-perturbed simulations. The SWE decreases 2–50 mm over mountain areas in the WUS during late winter to early spring. Maximum reductions of SWE are over CR, SN, and the mountainous areas of western Canada. The
Figure 8. Spatial distribution of (a) simulated and three observed, (b) CMC, (c) NOHRSC, and (d) SNOTEL, snow water equivalent data sets (SWE, unit: mm) for DJF and (e) simulated and (f) CMC SWE for MAM.
Figure 9. Spatial distribution of (a) simulated and (b) observed runoff (GRDC) for MAM (unit: mm d\(^{-1}\)).

Figure 10. Spatial distribution of change in March for (a) net shortwave radiation flux at the surface (NSW, W m\(^{-2}\)), (b) surface (2-m) air temperature (\(^{\circ}\)C), (c) skin temperature (\(^{\circ}\)C), and (d) surface albedo.
precipitation difference between the two simulations is small (not shown). However, the warmer temperature as a result of soot perturbations leads to more precipitation coming in the form of rain rather than snow, while total precipitation amount has no change, resulting in less snow accumulation during the peak snow season. Meanwhile, warmer surface temperatures speed up snowmelt during spring. Driven by reduced snow accumulation in the winter and increased snowmelt in spring, SWE reduction reaches a maximum in March over most of the snow-covered areas except for SN, where the SWE reduction reaches a maximum in April. The mean SWE in March is reduced by 5.6 mm (−3.9%) over the CRB basin, 5.9 mm (−6.7%) over CR, 1.4 mm (−1.6%) over the SSJ basin, and 4.1 mm (−1.6%) over SN, respectively. Comparing the monthly change of SWE over the two river basins (Figure 11), the SWE reduction lasts longer in SSJ than in CRB because SWE was overpredicted in SSJ and the incoming solar radiation is higher at the lower latitudes, both of which allow changes in NSW, ALB, T2/TS, and SWE to extend into the summer.

4.4. Runoff

[49] Figure 13 shows the spatial distribution of runoff from February to May. It can be seen that the runoff increases during late winter because in the soot-perturbed simulation more precipitation comes in the form of rain rather than snow. Figure 13a shows that runoff increases by 0.1–0.7 mm per day in February over the majority of the snow-covered areas. Runoff increases in February by 6.6% averaged over CR and by 1.0% averaged over SN, respectively. However, runoff decreases by 0.1–0.7 mm per day in May over major snow-covered areas as shown in Figure 13d. Runoff decreases in May by 2.1% averaged over CR and by 0.6% averaged over SN, respectively. Since snow accumulation is less during winter the snow line recedes more quickly in the perturbed simulation and runoff from snowmelt is reduced in late spring.

[50] The spatial distribution of runoff changes shows an interesting mixed pattern in March and April. Driven by the higher rain to snow ratio in the soot-perturbed simulation, the runoff increases in March over about half of the areas. Over the other half, less snowmelt from the reduced snowpack leads to runoff decreases for March (Figure 13b). The

Figure 11. Changes in monthly mean snow water equivalent (SWE, unit: 10 mm), albedo (×0.1), surface net shortwave radiation flux (NSW, ×5 W m⁻²), 2-m air temperature (T2, °C), and skin temperature (Ts, °C) averaged over the CRB and SSJ basin.

| Table 2a. EXP1 Mean Changes in March in Maximum Snow Albedo, Surface Albedo, Surface Downward Solar Radiation Flux and Net Solar Radiation Flux, Surface Air Temperature and Skin Temperature, Precipitation, Evaporation, Snow Water Equivalent, Soil Water Content, and Runoff Averaged for CRB and SSJ Basins as Well as CR and SN Regions * |
| --- | --- | --- | --- | --- |
| CRB | SSJ | CR | SN |
| maxsnowalb | −0.013 (−2.76%) | −0.004 (−0.73%) | −0.016 (−2.72%) | −0.008 (−1.41%) |
| albedo | −0.011 (−3.80%) | −0.0021 (−0.92%) | −0.019 (−5.15%) | −0.005 (−1.61%) |
| downsw | −0.137 (−0.07%) | −0.001 (0.00%) | −0.663 (−0.32%) | −0.203 (−0.09%) |
| nsw | 2.003 (1.58%) | 0.454 (0.28%) | 3.594 (2.69%) | 1.111 (0.73%) |
| t2 | 0.212 | 0.049 | 0.340 | 0.064 |
| ts | 0.260 | 0.0355 | 0.433 | 0.090 |
| Prevp | 0.002 (0.074%) | 0.000 (−0.01%) | 0.017 (0.82%) | −0.003 (−0.04%) |
| Evap | 0.015 (1.29%) | 0.006 (0.31%) | 0.023 (2.27%) | 0.010 (0.65%) |
| swe | −1.723 (−1.94%) | −2.481 (−1.62%) | −5.901 (−6.66%) | −4.125 (−1.55%) |
| soilm | 0.092 (0.16%) | 0.038 (0.00%) | 2.292 (0.42%) | 0.316 (0.054%) |
| runoff | 0.004 (0.14%) | 0.009 (0.21%) | 0.014 (0.82%) | 0.035 (0.57%) |

* Notation and units of measure: Maximum snow albedo, maxsnowalb; surface albedo, albedo; surface downward solar radiation flux, downsw (W/m²); net solar radiation flux, nsw (W/m²); surface air temperature, t2 (°C); skin temperature, ts (K); precipitation (mm/d); evaporation (mm/d); snow water equivalent, swe (mm); soil water content (mm); runoff (mm/d). EXP1 and EXP2 are two sensitivity experiments described in section 4. See Figure 6 for regions. Change in percentage is in brackets.
resulting net mean runoff change is very small (0.1–0.2\%) in March averaged over the CRB or SSJ basins (see Table 2a). Generally, the areas with increased runoff correspond to regions with higher elevation (see Figure 1b) and larger SWE (see Figure 8), whereas the areas with decreased runoff correspond to regions with lower elevation and smaller SWE where snow begins to melt earlier and faster. Runoff changes still show a mixed spatial pattern in April but the ratio of decreases to increases grows to around 70\%; the runoff increases only over a small portion of mountain areas such as SN range and eastern CR ranges.

4.5. Surface Water Budget

[51] To illustrate changes in the seasonal cycle of water partitioning, Figure 14 shows the basin mean monthly changes of precipitation, total runoff, evapotranspiration, soil moisture, and snowpack accumulation for the surface water budget over CRB and SSJ. In both basins, precipitation changes are minor throughout the seasonal cycle. This suggests that changes in other components of the water budget are mostly driven by changes in snowmelt or accumulation rather than precipitation amount. As a result of warming, there are significant reductions in snowpack between November and May (e.g., Figure 11), which are reflected in reduced snow accumulation between November and May.

Figure 12. Spatial distribution of change in mean snow water equivalent (SWE, millimeters) for March.

Figure 13. Spatial distribution of change in runoff (mm d\textsuperscript{-1}) from February to May.
Changes in monthly mean surface water budget averaged over the (top) CRB and (bottom) SSJ basins. Shown are changes in precipitation (P), snowpack accumulation rate (SP), runoff (R), soil moisture accumulation rate (SM), and evapotranspiration (ET) in mm d$^{-1}$.

Figure 14. Changes in monthly mean surface water budget and delayed timing in changes of snowmelt and runoff. Warmer temperatures in the soot-perturbed simulation also cause snowpack reductions in SSJ. Additionally, snowmelt decreases occur later in the year than for CRB and last a month longer. As a result of snowmelt change, the runoff decreases between May and August rather than between April and June in CRB basin. The change of soil moisture accumulation is small during most months in SSJ.

4.6. Feedbacks Between Albedo, Solar Radiation, Temperature, and Snowpack

[53] The soot-induced decrease of maximum snow albedo causes the land surface to absorb more solar radiation, increasing the skin and air temperatures. As a result of the perturbed radiation and warming, snow depth and fraction decrease during winter and spring, further bringing down the albedo over areas with at least some snow cover. This is a positive feedback process initialized by soot deposition on snow, which can be illustrated as: soot deposited on snow $\rightarrow$ reduced albedo over deep snow $\rightarrow$ reduced surface albedo $\rightarrow$ increased surface net solar radiation $\rightarrow$ skin and air warming $\rightarrow$ snowpack and snow fraction reduced $\rightarrow$ further reduced surface albedo.

[54] Figure 10d shows the spatial distribution of mean albedo change in March, which is the month with maximum albedo change. It can be seen that the albedo decreased 0.01–0.07 over snow-covered areas of the Central Rockies. Numerically in the Noah LSM, the simulated albedo reduction in EXP1 is caused by two processes: (1) the reduction in maximum albedo $\alpha_{\text{max}}$ due to the soot perturbations, and (2) the reduction of snowpack from the warmer skin temperature. Comparing the $\alpha_{\text{max}}$ and mean albedo changes shown in Table 2a, we can find that the $\alpha_{\text{max}}$ decreased 2.72% but albedo decreased 5.15% over CR, which indicates that only around half of the albedo reduction is directly caused by the $\alpha_{\text{max}}$ change. Snowpack reduction accounts for the remaining half. In the SN region the decrease of mean albedo is only slightly larger than the decrease of $\alpha_{\text{max}}$, which implies that the major contributor to albedo change over SN was $\alpha_{\text{max}}$ instead of the snowpack reduction. In the SN region, snowpack is deep and confined to the high terrain because of the sharp topographic gradients, and the model overpredicted precipitation as well as snowpack. In deep snow, the change in snow fractional coverage due to warming is small. Therefore, the change in albedo is dominated by the change in $\alpha_{\text{max}}$ and the percentage changes of both are smaller in SN and the SSJ basin because the high terrain only accounts for a small fraction of the basin area defined in this study.

4.7. Sensitivity of Impacts to Doubled Snow Albedo Perturbation

[55] A third WRF-RCM simulation, named EXP2, was done using the same settings as the EXP1 except that the snow albedo perturbation is doubled. Table 2b compared the climatic and hydrologic responses to various snow albedo perturbations. The major conclusions are not changed qualitatively in EXP2, but there is a larger magnitude of change than in EXP1. With the doubled snow albedo perturbation $\alpha_{\text{max}}$, the mean albedo reduction increases around 70%, the increase of surface absorbed solar radiation goes up 60–80%, and surface air and skin warming amounts increase.
Table 2b. Same as Table 2a but for EXP2

<table>
<thead>
<tr>
<th></th>
<th>CRB</th>
<th>SSJ</th>
<th>CR</th>
<th>SN</th>
</tr>
</thead>
<tbody>
<tr>
<td>maxsnowalbedo</td>
<td>−0.026 (−5.52%)</td>
<td>−0.008 (−1.45%)</td>
<td>−0.031 (−5.43%)</td>
<td>−0.015 (−2.81%)</td>
</tr>
<tr>
<td>albedo</td>
<td>−0.019 (−6.51%)</td>
<td>−0.0035 (−1.57%)</td>
<td>−0.033 (−9.07%)</td>
<td>−0.010 (−2.99%)</td>
</tr>
<tr>
<td>downsnow</td>
<td>−0.406 (−0.23%)</td>
<td>0.054 (0.025%)</td>
<td>−1.112 (−0.53%)</td>
<td>−0.388 (−0.17%)</td>
</tr>
<tr>
<td>snow</td>
<td>3.277 (2.58%)</td>
<td>0.833 (0.51%)</td>
<td>6.360 (4.77%)</td>
<td>2.066 (1.35%)</td>
</tr>
<tr>
<td>t2</td>
<td>0.319</td>
<td>0.071</td>
<td>0.557</td>
<td>0.106</td>
</tr>
<tr>
<td>ts</td>
<td>0.398</td>
<td>0.086</td>
<td>0.721</td>
<td>0.156</td>
</tr>
<tr>
<td>Prep</td>
<td>0.006 (0.19%)</td>
<td>0.009 (0.18%)</td>
<td>0.021 (1.02%)</td>
<td>0.019 (0.29%)</td>
</tr>
<tr>
<td>Evap</td>
<td>0.025 (2.23%)</td>
<td>0.010 (0.54%)</td>
<td>0.041 (4.04%)</td>
<td>0.020 (1.30%)</td>
</tr>
<tr>
<td>swe</td>
<td>−8.800 (−6.22%)</td>
<td>−2.689 (−3.15%)</td>
<td>−10.363 (−11.70%)</td>
<td>−8.135 (−3.06%)</td>
</tr>
<tr>
<td>soilm</td>
<td>1.607 (0.25%)</td>
<td>−0.154 (−0.025%)</td>
<td>3.788 (0.69%)</td>
<td>0.546 (0.094%)</td>
</tr>
<tr>
<td>runoff</td>
<td>0.015 (0.50%)</td>
<td>0.025 (0.58%)</td>
<td>0.027 (1.57%)</td>
<td>0.083 (1.35%)</td>
</tr>
</tbody>
</table>

around 50% over both CRB and SSJ (see Table 2c). The SWE reduction increases around 58% over CRB, but is almost doubled over SSJ, CR, and SN. This indicates that the change of surface energy and temperature is smaller than a linear response to snow albedo perturbation, however, snowpack reduction is nonlinearly accelerated (i.e., greater than linear with respect to the weighted albedo as well as surface energy and temperature) by the surface energy and temperature changes.

The results from EXP1 and EXP2 do not quantitatively span the uncertainty range of climatic and hydrologic impacts induced by the soot-on-snow effect, but they do provide more information on how sensitive the regional climate response is to the snow albedo perturbation and its biases, which could result from such simplifications as neglect of the snow aging process, assumption of a universal albedo reduction function, and uncertainties in the WRF chemistry simulation such as with the emissions.

Table 3 compares the simulated precipitation, surface temperature, and SWE against observations and the model biases averaged over CR for MAM. As discussed in section 3.4, the control simulation (CONT) significantly overpredicts the precipitation and the wet bias over CR is larger than over the coastal mountainous areas and the two river basins, especially during DJF. CONT shows a cold bias of −1.5°C to −2.0°C for the cold season over CR. As expected, the difference among the precipitation simulations is small, but EXP1 and EXP2 slightly reduced the model cold bias by about 0.2°C−0.3°C (10−15%) after the soot-induced snow albedo perturbation is included in the simulation. Of the three simulations, EXP2 agrees the most with observation of surface air temperature. Consistent with our discussion above, EXP2 has smaller SWE in both DJF and MAM compared to CONT or EXP1, thus highlighting the effects of soot on snow. The SWE from all simulations is within the range of the two observation data sets (i.e., NOHRSC and CMC). Considering the large uncertainty in the snow observations, model biases in simulating snow, and that not all factors that affect snow are included in the simulations, it should be noted that agreement between the observed and simulated snow provides no evidence for or against the soot effects on snow.

5. Conclusion and Discussion

The 2007 IPCC report listed the radiative forcing induced by “soot on snow” as an important anthropogenic forcing affecting climate change between 1750 and 2005. However, it is still uncertain how the soot-induced snow albedo perturbation affects regional snowpack and the hydrological cycle in mountain areas. In this paper we simulate the deposition of soot aerosol on snow using a coupled regional aerosol/chemistry model, WRF-Chem, to estimate the soot-induced snow albedo perturbation. This is then used to modify the maximum snow albedo for sensitivity tests using WRF as a regional climate model. Investigation of the WRF regional climate simulations focuses primarily on the impact of the soot-induced snow albedo perturbation on snowpack and the hydrological cycle in the WUS.

WRF-Chem simulated large spatial variability in BC concentrations that reflect the localized emissions, as well as the influence of the complex terrain and the resulting meteorological processes that affect chemical reactions and transport of pollutants. The simulated BC concentrations in the lower atmosphere are generally higher than the observed values derived from the IMPROVE networks. The climate simulations have been compared against observations for surface air temperature, precipitation, runoff, and snow water. The WRF-RCM model captured the seasonal cycle and interannual variability of precipitation very well. The spatial pattern of precipitation is also reasonably

Table 2c. Comparison of EXP1 and EXP2

<table>
<thead>
<tr>
<th></th>
<th>CRB</th>
<th>SSJ</th>
<th>Percent Change</th>
<th>CRB</th>
<th>SSJ</th>
<th>Percent Change</th>
</tr>
</thead>
<tbody>
<tr>
<td>EXP1-CONT</td>
<td>EXP2-CONT</td>
<td>EXP1-CONT</td>
<td>EXP2-CONT</td>
<td>EXP1-CONT</td>
<td>EXP2-CONT</td>
<td>EXP1-CONT</td>
</tr>
<tr>
<td>maxsnowalbedo</td>
<td>−0.013</td>
<td>−0.026</td>
<td>+100%</td>
<td>−0.004</td>
<td>−0.008</td>
<td>+100%</td>
</tr>
<tr>
<td>albedo</td>
<td>−0.011</td>
<td>−0.019</td>
<td>+73%</td>
<td>−0.021</td>
<td>−0.035</td>
<td>+67%</td>
</tr>
<tr>
<td>snow</td>
<td>2.003</td>
<td>3.277</td>
<td>+62%</td>
<td>0.454</td>
<td>0.833</td>
<td>+83%</td>
</tr>
<tr>
<td>t2</td>
<td>0.212</td>
<td>0.319</td>
<td>+50%</td>
<td>0.049</td>
<td>0.071</td>
<td>+45%</td>
</tr>
<tr>
<td>ts</td>
<td>0.260</td>
<td>0.398</td>
<td>+53%</td>
<td>0.055</td>
<td>0.086</td>
<td>+56%</td>
</tr>
<tr>
<td>swe</td>
<td>−5.582</td>
<td>−8.800</td>
<td>+58%</td>
<td>−1.381</td>
<td>−2.689</td>
<td>+95%</td>
</tr>
</tbody>
</table>

*a*Notation is the same as in Table 2a. CONT denotes control simulation as described in section 3.
simulated during the cold season except for an overestimate of the orographic precipitation along the Cascades and Sierra Nevada mountains. The surface air temperature is spatially consistent between the observations and simulation and the bias is less than 1°C over the Columbia River Basin (CRB) and Sacramento-San Joaquin (SSJ) basins. Accurately simulated surface temperatures in the model result in an accurate reproduction of the phase of precipitation and the snow accumulation and melt processes at the surface. Therefore SWE is simulated well in WRF-RCM. The simulation reproduces many of the features of the observed snowpack distribution. The runoff is also evaluated against observations and the spatial distribution of the modeled runoff generally agrees with that of the GRDC runoff in winter and spring.

The WRF-RCM simulations show that soot-induced snow albedo perturbations significantly change the regional climate and water budget in WUS. NSW, which is closely spatially correlated with the change of surface albedo, increases 2–12 W m\(^{-2}\) from late winter to early spring over the central Rockies and southern Alberta, which drives a surface skin (2-m air) temperature increase of 0.2°–1.4°C (0.1°–1.0°C) over the majority of the snow-covered areas in the WUS. The SWE decreases 2–50 mm over the mountains during late winter to early spring, which is reflected in reduced snow accumulation in winter and less snowmelt in spring. The maximum change occurs in March for albedo, NSW, temperature, and snowpack.

Our simulations show that the precipitation difference is negligible between the perturbed and control simulations. In a warmer perturbed climate, more precipitation comes in the form of rain rather than snow, which results in less snow accumulation but more runoff during the snowy winter period. The runoff decreased in late spring, driven by the reduced snowmelt from the reduced snowpack. This is a positive feedback process initialized by the soot deposition on snow. Our simulations indicate that over the central Rockies, only about half of the albedo reduction is directly caused by the change in maximum snow albedo induced by the soot in snow. Snowpack reduction accounts for the remaining half of the surface albedo change.

The same feedback process described above can also be initialized by increases in greenhouse gases, which lead to skin/air warming and the subsequent changes in the feedback loop. Hence the suite of changes described in section 5 is very similar to what have been discussed by studies that investigated changes in mountain hydrology due to greenhouse warming [e.g., Leung et al., 2004]. But while greenhouse forcings are more uniformly distributed spatially (though the response in snowpack is still highly regional), soot forcing on snow is more regional and depends on the coexistence of snowpack and soot. Therefore, larger uncertainties may be expected in estimating both the forcing and response related to soot-induced snow albedo effects simply because of the inherent spatial scale of processes involved. Additionally, other uncertainties are introduced by the methodology used in this study, which we plan to address in future studies. The most significant of these issues include the following.

### 5.1. WRF-Chem Simulation and Emission Data

Because of the extreme computational cost, the WRF-Chem simulation has been run only for 1 year. The interannual variability of concentration and deposition of soot therefore is not accounted for in this study. We used the EPA NEI99 emission data set in the WRF-Chem simulations and no adjustments were made to bring the 1999 values closer to the 2003 values and no seasonal cycle was included. Also, the NEI99 estimates do not include forest fire, biogenic, or ship emissions. However, most of the fires occur during summer and fall, after the snow has melted, so the impact on snowpack may be not significant.

Since WRF-Chem is a regional model, assumptions must be made regarding trace gases and particulates entering the model boundaries. Fixed profiles were used as inflow boundary conditions, so the model does not reproduce the impact of trans-Pacific transport. The net effect of these assumptions is that the deposition in these simulations should be considered a minimum estimate. However, our simulated BC concentrations are noticeably higher than the observed values in the lower atmosphere, which does not necessarily imply an overestimation of BC deposition on surface. Unfortunately, we lack surface measurements to evaluate the model performance for simulated soot deposition. This suggests a need for more analysis and development to understand and
reduce model biases. The WRF-Chem simulation, however, provided a representative magnitude and spatial distribution of BC for investigating soot effects on snow in a mountainous region.

5.2. Estimation for Snow Albedo Perturbation

[66] In the current Noah LSM, the snow albedo is calculated on the basis of snow fraction, which is parameterized according to simulated snow depth, and the maximum snow albedo $\alpha_{\text{max, snow}}$ which is based on the measured snow albedo over deep snow. The $\alpha_{\text{max, snow}}$ input includes spatial variability, albeit at a coarser resolution than the model resolution, but does not include temporal variation, which is related to other possible effects of snow properties (e.g., grain size) on snow albedo and emissivity. Flanner and Zender [2005, 2006] suggested several positive feedbacks, including vertically resolved snowpack heating, amplifying the first-order warming effect of BC in snow. In this study we estimated soot-induced snow albedo reduction on the basis of very limited measurements and an empirical relationship from Jacobson [2004]. We should incorporate an interactive aerosol-snow surface radiative model [e.g., Flanner and Zender, 2006] in the land surface model in WRF if resources permit in the future, though sensitivity experiments using the present model framework by doubling the current snow albedo perturbation have provided a range for the uncertainties of soot-induced snow albedo effects.

5.3. WRF-RCM Simulations

[67] The control and sensitivity WRF-RCM simulations are only run for 5 years owing to limited resources. It would be better to extend the simulations to better represent the interannual variations of mountain hydroclimate and the significance tests for impacts of soot. More importantly, although the spatial pattern and seasonal cycle of precipitation are simulated reasonably well by WRF-RCM during the cold season, an overestimation of the orographic precipitation is also apparent along the Cascades and Sierra Nevada as well as CRB and SSJ basins. These errors result in biases in simulating snowpack and runoff. Last, WRF-Chem and WRF-RCM are run separately (offline) so interactions between the concentration and deposition of soot and climate that would modify the soot-induced snow albedo perturbation are not addressed.

5.4. Other Types of Aerosols in Snow

[68] In this study we only included and simulated the deposition of soot aerosol on snow and the resulting impact in the sensitivity experiments. In the real world, other natural and anthropogenic atmospheric particles, such as mineral dust and volcanic ash, also deposit on the snow and change the albedo of surface. In fact, the dust deposited on snow likely has a much greater impact than soot in some regions (e.g., the Southern Rockies) during a specific season [Painter et al., 2007]. Therefore, it is desirable in future studies to include the radiative and hydrological effects of other aerosols as well.

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