SIMULATING THE COLD SEASON SNOWPACK: THE IMPACT OF SNOW ALBEDO AND MULTI-LAYER SNOW PHYSICS

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Prepared By:

Waliser, D.,[1] J. Kim,[2] Y. Xue,[2] Y. Chao,[1] A. Eldering,[1] R. Fovell,[2] A. Hall,[2] Q. Li,[2] K. N. Liou,[2] J. McWilliams,[2] S. Kapnick,[2] R. Vasic,[2,3] F. De Sale,[2] and Y. Yu[2]

[1] Jet Propulsion Laboratory/California Institute of Technology
[2] University of California Los Angeles
[3] National Centers for Environmental Prediction

DRAFT PAPER

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Arnold Schwarzenegger, *Governor*

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Preface

The California Energy Commission's Public Interest Energy Research (PIER) Program supports public interest energy research and development that will help improve the quality of life in California by bringing environmentally safe, affordable, and reliable energy services and products to the marketplace.

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The California Climate Change Center Report Series details ongoing center-sponsored research. As interim project results, the information contained in these reports may change; authors should be contacted for the most recent project results. By providing ready access to this timely research, the center seeks to inform the public and expand dissemination of climate change information, thereby leveraging collaborative efforts and increasing the benefits of this research to California's citizens, environment, and economy.

For more information on the PIER Program, please visit the Energy Commission's website www.energy.ca.gov/pier/ or contract the Energy Commission at (916) 654-5164.

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Abstract

This study used numerical experiments to investigate two important concerns in simulating the cold season snowpack: the impact the alterations of snow albedo due to anthropogenic aerosol deposition on snowpack and the treatment of detailed snow physics using a multi-layer snow model. The snow albedo experiments considered qualitatively future changes in anthropogenic emissions and the subsequent increase or decrease of black carbon deposition on the Sierra Nevada snowpack by simply altering the prescribed snow albedo values. The alterations in the snow albedo primarily affect the snowpack via surface energy budget with little impact on precipitation. It was found that a decrease in snow albedo due to an increase in local emissions enhances snowmelt and runoff in the early part of a cold season, resulting in reduced snowmelt-driven runoff in the later part of the cold season. For example, for a 5 percent change in the specified snow albedo, the snowmelt and runoff in late winter changes by 30 percent and 20 percent, respectively, in regions above 2750 meters. An increase in snow albedo associated with reduced anthropogenic emissions results in the opposite effects. The timing and relative magnitude of the sensitivity of snow water equivalent, snowmelt, and runoff vary systematically according to terrain elevation; as terrain elevation increases, the peak response of these fields occurs later in the cold season. For instance, the maximum response of snowmelt to the prescribed snow albedo changes occurs in February in the elevation range 2250–2750 meters, but it occurs in March in the elevation range above the 2750-meter level. The response of area-mean surface albedo, the absorption of surface insolation, sensible heat flux, and the surface and low-level temperatures is also analyzed. The response of snow water equivalent and surface energy budget to the alterations in snow albedo found in this study also shows that the effects of snow albedo on snowpack are further enhanced via local snow-albedo feedback. This experiment's results suggest that a reduction in local emissions, which would provide an increase in snow albedo, could alleviate the early snowmelt and reduced runoff in late winter and early spring caused by the global climate change, at least partially. The most serious uncertainties associated with this part of the study are a quantification of anthropogenic black carbon deposits on the Sierra Nevada snowpack and the relationship between the amount of black carbon deposition and snow albedo—a subject of future study. The comparison of the spring snowpack simulated with a single- and multi-layer snow model during the spring of 1998 shows that a more realistic treatment of snow physics in a multi-layer snow model can improve the simulation of snowpack, especially during spring when snow ablation is significant.

Keywords: Snow albedo, black carbon, snow process, snowpack, the Sierra Nevada Mountains

1.0 Introduction

The snowpack in the Sierra Nevada region is important to California's water resources. The high elevation snowpack serves as a natural reservoir that stores fresh water during the wet, cold season and releases it gradually during the dry, warm season. About 60% of the water supply for Southern California comes from melting Sierra Nevada snowpack. Snowmelt also affects hydropower generation in California (Vicuña et al. 2008). The impact of global warming on the Sierra Nevada snowpack has become one of the leading topics in the regional climate change studies for the California region (Leung and Ghan 1999; Kim 2001; Kim et al. 2002). The snow budget in the Sierra Nevada is affected by a number of factors, such as insolation, air temperature, and orography. Previous studies on the impact of climate change on the Sierra Nevada snowpack have focused solely on the impact of lower tropospheric warming (e.g., Leung and Ghan 1999; Kim 2001; Kim et al. 2002; Cayan et al. 2008) since this affects two important factors determining the snow budget: rainfall-snowfall partitioning and snow ablation. For a more comprehensive understanding and projection of the Sierra Nevada snowpack in future climate, it is necessary to investigate the role of other factors that might also affect the snow budget.

Snow albedo is among the most important local parameters in shaping the spatiotemporal variations in snowpack. Surface insolation, and more specifically the portion of insolation absorbed by the snowpack, is the leading energy source in the evolution of snowpack, especially during the melting period. Thus variations in snow albedo can exert significant impact on snowpack during the course of accumulation and ablation. The surface albedo of sufficiently deep snowpack, and in turn the amount of the insolation absorbed by the snowpack, depends largely on the ice grain size and impurities within or at the surface of ice grains (e.g., Wiscombe and Warren 1980; Warren and Wiscombe 1980; Yang et al. 1997; Mölders et al. 2008). Previously, the impact of snow grain size has been incorporated into snow albedo formulation in terms of snow age or surface temperature or a combination of both, and has been examined in a number of evaluation studies (e.g., Yang et al. 1997; Sun et al. 1999; Molotch and Bales 2006; Mölders et al. 2008). There exist, however, only a limited number of studies (e.g., Hadley et al. 2007) on the alteration in snow albedo and its impact on surface hydrology due to dust and black carbon (BC) particles deposited on snowpack. This is an important concern because the amount of BC deposition on snowpack is closely related with anthropogenic emissions. Thus, anthropogenic emissions that have bearing on the causes and characteristics of global climate change include an influence on local snowpack by altering snow albedo.

The potential importance of BC deposition on snow albedo in the Sierra Nevada region can be inferred from previous studies. In a series of theoretical studies, Wiscombe and Warren (1980) and Warren and Wiscombe (1980) showed that impurities in snowpack such as dust and BC reduce snow albedo primarily in the spectral range shorter than 1 micrometer (μm) where most of the solar energy resides. For an ice grain radius of 1000 μm, for example, their calculations show that the average snow albedo for the wavelengths between 0.4 μm and 1 μm varies from near unity (i.e., almost total reflection of insolation) for pure snow to about 0.4 with a presence of a moderate amount of soot within the snow layer. They also showed that the impact of soot on snow albedo tends to decrease as the snow grain size decreases; however, for the snow grain size of 100 μm, soot concentration of 1 part per million, by weight (ppmw) can still reduce snow albedo from near unity to below 0.9. A reduction of surface albedo by 0.1 can increase the

absorbed insolation by over 60 watts per square meter (Wm^{-2}) during midday under clear conditions and 20 Wm^{-2} for daily averages. Thus their theoretical study shows that the snow albedo and the absorbed insolation at the surface of the snowpack can be altered by a significant amount by the deposition of black carbon. Recently, observational studies (Husar et al. 2001; VanCuren et al. 2005; Painter et al. 2007) reveal that depositions of dusts and BC of local and Asian origins can alter snow albedo and snow ablation in the western U.S. region. Similar impacts of dust and BC depositions on snow have also been observed in the polar region (McConnell et al. 2007). Significant anthropogenic emissions in California, in conjunction with prevailing westerly winds that transport fine particulates into the Sierra Nevada region, can alter the snow albedo in the region. Thus, the sensitivity of the Sierra Nevada snowpack to the deposition of BC needs investigation.

Another challenge in simulating snowpack is the complexity in the physical processes in the interior of the snowpack. Snow models that have been used in climate simulations range from a relatively simple single layer model that considers only limited physical processes within the snowpack to state-of-the-art multi-layer models that can resolve a number of important physical processes (e.g., Yang et al. 1997; Sun et al. 1999; Slater et al. 2001; Ek et al. 2003; Xue et al. 2003). Many regional climate models including the newly developed community model, the Weather Research and Forecast (WRF) community model (Skamarock et al. 2005), use a single-layer snow model. In general, a single-layer snow model overly simplifies important physical processes such as the heat transfer, snow compaction, phase changes in energy balance, and refreezing of snowmelt water. One of the shortcomings of most single-layer snow models in simulating long-term snow variations, especially during the ablation period, is that for snowmelt to occur, the temperature of the entire snow layer must rise above the freezing point[1]. In reality, the near-surface snow layer can readily warm up relative to deeper levels to begin the melting process. Incorporating this realism into a model would alter snowpack loss rate significantly, not only for the spring snow ablation period but also for the winter snow accumulation period. Recognizing the shortcomings of single-layer snow models, several multi-layer snow models have been developed for various land-surface schemes (e.g., Sun et al. 1999; Dai et al. 2003). Evaluation of long-term snowpack simulations using multiple snow schemes against field data showed that multi-layer snow models can outperform single-layer models in simulating long-term snowpack variations in cold regions (Slater et al. 2001; Bowling et al. 2002; Nijssen et al. 2002; Luo et al. 2003; Xue et al. 2003; Rutter et al. 2008). In particular, the multi-layer treatment of snowpack results in faster snowmelt during spring, representing an improvement associated with a systematic error in many single-layer model simulations (Slater et al. 2001). Considering the importance of the long-term snow budget in the water resources for California, the difference in snowpack simulation by a use of more physically based multi-layer snow models needs close examination in order to improve the impact projections of anthropogenic climate change on the cold season snowpack and in turn on California's water resources. This is an especially important issue because the snow scheme in the WRF model that is selected by the authors and many other institutions for regional climate research has not been closely examined yet.

[1] This is not common in all single layer models. Among the exceptions is the Noah land-surface model that calculates snowmelt in terms of the skin temperature as a part of solving the surface energy balance equation, not from the temperature of the snow layer.

This study examines qualitatively the impact of the alteration of snow albedo by BC deposition and the multi-layer treatment of snow physics on simulating the cold season snowpack in two regional climate model (RCM) experiments. The results from this study will be used to design more comprehensive future experiments for quantifying these effects in regional climate study with an emphasis on California and the western U.S. region using the Regional Earth System Model (RESM). Experimental designs for examining the impact of snow albedo alterations and the multi-layer snow scheme are presented in Section 2. Sections 3 and 4 present the results obtained in the snow albedo and the multi-layer snow scheme experiments, respectively. Conclusions and discussions based on these experiments are presented in Section 5.

2.0 Experimental Design

The results presented in the following sections are generated in two separate numerical experiments: (1) a fine-resolution simulation for the California region in which the impact of snow albedo changes on the Sierra Nevada snowpack is investigated, and (2) a coarse resolution simulation over a North American region for investigating the snowpack simulation based on single- and multi-layer snow models. Both experiments used the WRF model version 2.2.1 (Skamarock et al. 2005) but with different land-surface schemes. Both experiments employ the WRF model configured with 28 atmospheric layers. The WRF model parameterizations for atmospheric physical processes used in these experiments include the rapid radiative transfer model (RRTM) longwave radiation scheme (Mlawer et al. 1997), Dudhia (1989) shortwave radiation, and the WRF single-moment with simple ice cloud microphysics scheme (WSM-3). Details of the WRF model and the physics parameterizations can be found on the WRF model website (http://wrf-model.org) and will not be included here. Specifics for the two experiments are presented below.

2.1. The Impact of the Snow Albedo Changes on the Sierra Nevada Snowpack

In the snow albedo sensitivity study, the model domain covers California with a 12-kilometer (km) horizontal resolution (Figure 1a). Note that parameterized cumulus convection is inactive in this fine-resolution simulation. The NOAH[2] land-surface scheme (Chang et al. 1999) with four soil layers is used in this experiment. The NOAH model utilizes a single-layer snow scheme to simulate snowpack. Details on the NOAH and the snow models are presented in Mahrt and Pan (1984), Pan and Mahrt (1987), Kim and Ek (1995), and Chang et al. (1999).

The simulation period covers the seven month period October 2050–April 2051 using the initial and lateral boundary forcing obtained from the results in a 36-km resolution WRF model simulation that in turn is driven by a climate scenario from the National Center for Atmospheric Research Community Climate System Model, version 3.0 (NCAR-CCSM3) corresponding to the *Special Report on Emissions Scenarios* (SRES) A1B emission scenario. Details of the 36-km regional climate simulation are presented in Kim et al. (2008). This allows us to zoom in the fields

[2] NOAH consists of the National Centers for Environmental Prediction (NCEP), Oregon State University (Dept of Atmospheric Sciences), the Air Force (both AFWA and AFRL), and the Office of Hydrologic Development at the National Weather Service.

simulated in a coarse resolution simulation over a region of special interests without incurring excessive computational resources. This also allows us to avoid an excessive spectral gap between the large-scale forcing data and the regional simulation (Kim et al. 2008).

2.2. Snowpack Differences Between Single- and Multi-layer Snow Model Simulations

In the multi-layer snow model study, the model domain covers the entire conterminous U.S. region at an 80 km horizontal resolution (Figure 1b). The WRF model physics selected for the experiment are the same as in the snow albedo sensitivity study except that the Kain-Fritsch cumulus parameterization scheme (Kain and Fritsch 1993, 1998) is activated due to this experiment's coarse horizontal resolution, and that the NOAH land-surface scheme is replaced with the Simplified Simple Biosphere (SSiB) model with three soil layers. In addition, the SSiB/Monin-Obukhov scheme (Xue et al. 1991, 2003) is used for computing surface turbulent fluxes. The simulation is performed for the three-month period April-June 1998 in which snow ablation is important and the differences among snow schemes are largest in general (Slater et al. 2001).

The differences in snowpack simulations in single and multi-layer snow schemes are examined using the WRF-SSiB model. For the single- and multi-layer snow model simulations, the SSiB-1 and SSiB-3 models that use a single- and three-layer snow models respectively, are separately coupled with the WRF model. The physics in the SSiB-1 and SSiB-3 models are identical except the snow scheme. For more details of SSiB and the snow models used in lieu of SSiB-1 and SSiB-3, readers are referred to Sun et al. (1999) and Xue et al. (2003).

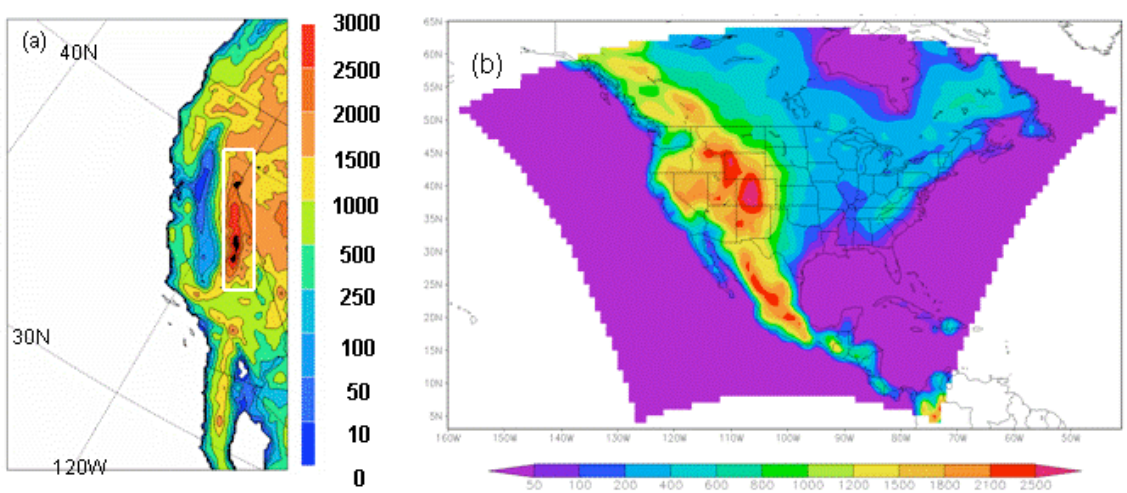

Figure 1. The model domains and terrain representations in (a) the snow albedo sensitivity study (12 km) and (b) multi-layer snow model sensitivity study (80 km). The units are in meters.

3.0 Results

3.1. The Impact of the Snow Albedo Changes on the Sierra Nevada Snowpack

The impact of the alterations in snow albedo on the cold season Sierra Nevada snowpack is investigated in a sensitivity study in which five cold season simulations are performed using five different snow albedo specifications. The control simulation utilizes the default snow albedo values provided with the NOAH model as a part of the WRF-2.2.1 model. The default snow albedo values in this model are currently prescribed on the basis of satellite observations (Robinson and Kukla 1985) without considering the physical processes that can alter snow albedo (e.g., aging, aerosol deposition, compaction). The prescribed snow albedo values are used to calculate the grid-mean surface albedo value by area-weighted averaging the snow albedo and the landuse-dependent snow-free albedo according to the fractional snowcover area calculated in terms of the snow-water equivalent (SWE) within each model grid. More details of the calculation of the fractional snow cover and the corresponding surface albedo value are presented in Chang et al. (1998). The default area-mean snow albedo values used in the control run are 0.6 in the three Sierra Nevada regions described below. In the four sensitivity runs, the default snow albedo values are modified to be 90%, 95%, 105%, and 110% of the control run. The corresponding snow albedo values used in the sensitivity simulations range from 0.54 for the 90% run to 0.66 for the 110% run. Considering Warren and Wiscombe (1980) showed that snow albedo could be reduced from near unity to below 0.94 with 0.1 ppmw BC concentration for 100 μm-size snow grains and even by a larger amount for larger snow grains and/or higher BC concentration, the range of snow albedo variations in these sensitivity studies are realistic possibilities. The two smaller snow albedo runs (0.9 and 0.95 times the default values) represent the cases in which BC emissions in California, thus the BC deposition on the Sierra Nevada snowpack, will continue to increase in the future. The other two runs with larger snow albedo values (1.05 and 1.1 times the default values) represent the cases in which anthropogenic emissions will be reduced by successful implementation of recent mandates by California's governor (Steiner et al. 2006). The sensitivity study based on the snow albedo values prescribed in this way can be used for qualitative examinations; the qualitative approach is inevitable because the amount of aerosol deposits on the Sierra Nevada snowpack and the quantitative relationship between aerosol deposition and snow albedo remain poorly understood. The model data for elevations roughly above the 1750-meter (m) level within the Sierra Nevada region (marked by the white box in Figure 1a) are analyzed according to elevation ranges defined at 500-m intervals.

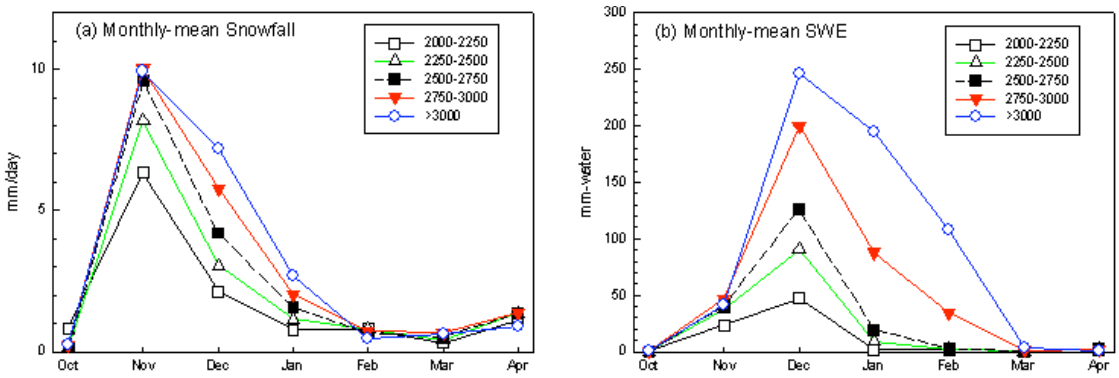

Figure 2. The monthly-mean snowfall and SWE in the Sierra Nevada region above the 2000 m level simulated in the 12-km resolution control run

The snowfall and SWE in the control run (Figure 2) vary significantly according to terrain elevation as well as month. The timing of the maximum snowfall is similar in all elevations, but the amount increases with increasing elevation (Figure 2a). This is because higher elevation regions remain above the freezing level for longer periods during the cold season and because precipitation generally increases with increasing elevation in the region (Soong and Kim 1996; Kim 1997, 2001). The variation according to elevation range is significantly amplified in the SWE field relative to the snowfall amount (Figure 2b). In December, the SWE in the highest elevation range (above 3000 m) is about five times the value in the lowest elevation range (2000–2250 m), even though the snowfall is only twice as large. The large variations in the SWE according to terrain elevation are due to larger snowmelt in the lower elevation ranges. Below 2500 m, snowpack is almost completely depleted in February; above that level, snow depletion occurs one month later. The simulated occurrence of the peak snowfall and SWE in November and December, respectively, in the simulation does not correspond well with the historical data that show the peak snowfall in California climatologically occurs in the period January–February (e.g., Kim and Lee 2003). This may be due to shortcomings of the WRF model, interannual variations in precipitation, and/or due to the warmer climate of the mid-twenty-first century. Note that the large-scale forcing in this study is obtained from a GCM projection that contains its own biases, not from more accurate present-day reanalysis data.

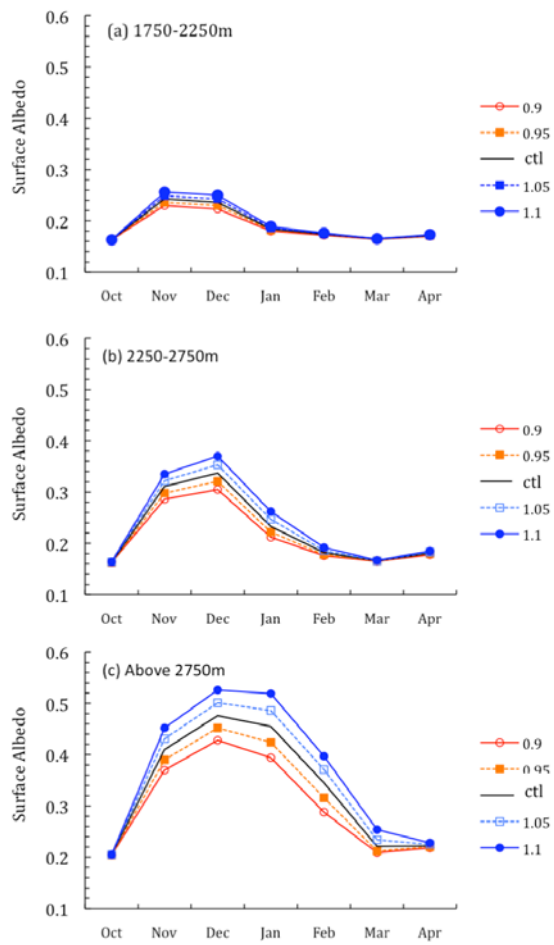

Figure 3. The monthly-mean surface albedo values in the five simulations in the three elevation ranges: (a) 1750–2250 m, (b) 2250–2750 m, and (c) above the 2500 m level. The numbers in the legends indicate the ratio between the snow albedo values used in the corresponding sensitivity study and in the control run (ctl). For example, 0.95 represents the simulation in which the snow albedo values are 95% of the control run.

The area-mean surface albedo varies systematically according to terrain elevations as well as the snow albedo (Figure 3). The maximum surface albedo occurs in December when the SWE is also largest, especially in the regions above 2250 m where snow cover is significant. The differences in the surface albedo among the sensitivity runs increase with increasing elevation. In the lowest elevation range (Figure 3a), the maximum difference in the monthly-mean surface albedo between the ±10% snow albedo change runs is 0.04 in November and December; it becomes 0.11, almost as large as the snow albedo differences between the ±10% runs, in the highest elevation range in January (Figure 3c). This suggests that the impact of snow albedo changes on the surface hydrology is larger in higher elevation regions. It is interesting to note that the period of maximum differences in the surface albedo among the sensitivity runs tend to occur later in the season as terrain elevation increases. In the lowest elevation range, the maximum differences appear in the November–December period (Figure 3a), but the differences are largest during January in the highest elevation range (Figure 3c).

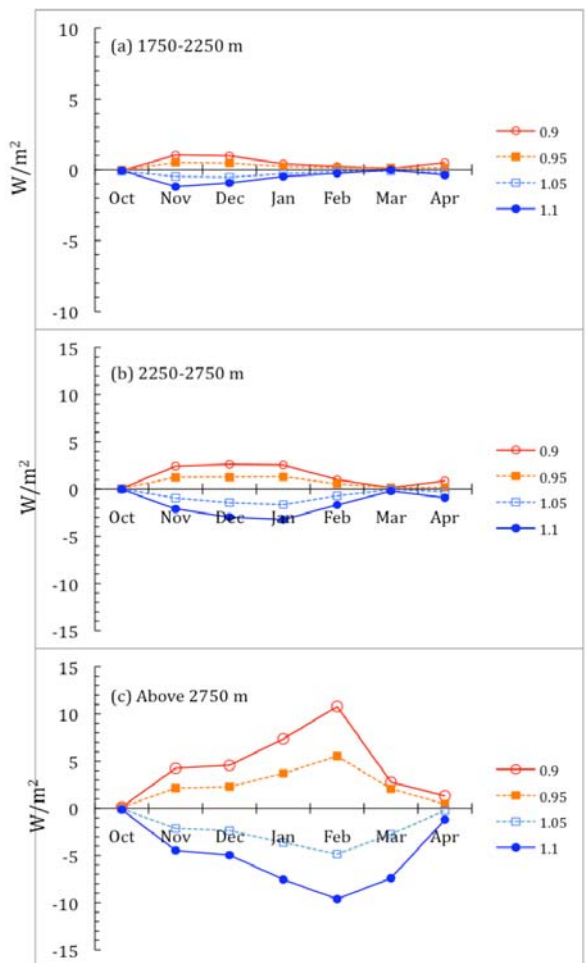

Figure 4. The changes in the absorbed surface insolation due to the snow albedo changes in the three elevation ranges: (a) 1750–2250 m, (b) 2250-2750 m, and (c) above the 2750 m level

The surface energy budget component that is most directly affected by the alterations in snow albedo is the amount of absorbed insolation at the surface. The differences in the absorbed surface insolation among the five simulations range from less than 2 Wm^{-2} in the lowest elevation range (Figure 4a) to over 10 Wm^{-2} in the highest elevation range (Figure 4c). These variations in the absorbed solar energy at the surface according to terrain elevation are a direct consequence of the elevation dependence in the surface albedo sensitivity presented in Figure 3 because the incident solar radiation at the surface is similar in all elevation ranges (not shown). Considering that the changes in the surface radiative forcing between the late twentieth century and the late nineteenth century by the combined effects of the increases in greenhouse gases and sulfate aerosols is about 2 Wm^{-2} (Meehl et al. 2003), the results show that anthropogenic alterations in snow albedo via BC deposition can exert significant climate forcing in the regions of significant snow cover.

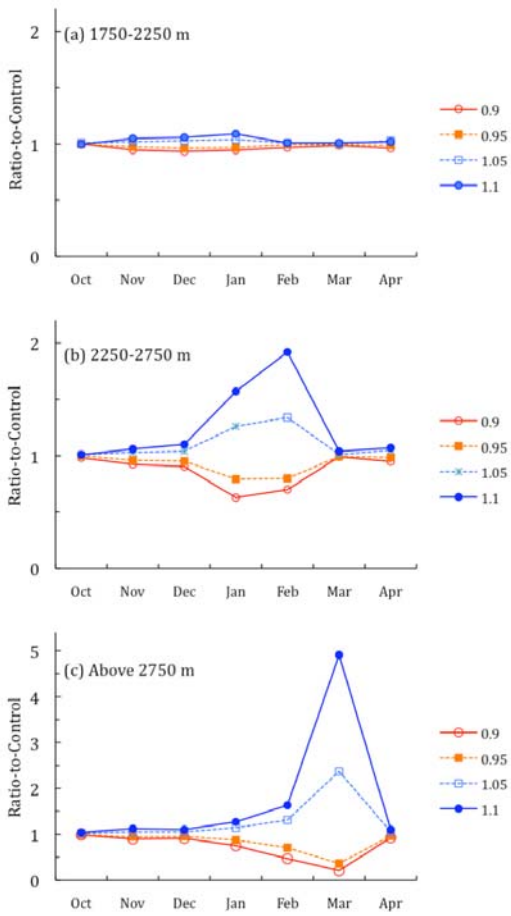

Figure 5. The ratio of the SWE in the four sensitivity simulations to that in the control simulation in the three elevation ranges: (a) 1750–2250 m, (b) 2250–2750 m, and (c) above the 2750 m level

The most notable feature in the sensitivity of the Sierra Nevada snowpack to snow albedo is that the snowpack, represented here as SWE, sensitivity increases with increasing terrain elevation (Figure 5). In the lowest elevation range (1750–2250 m), the SWE varies according to the prescribed snow albedo values, but the magnitude of the SWE sensitivity is small (Figure 5a). The SWE sensitivity to snow albedo becomes larger in higher elevations. With a 10% increase in snow albedo from the control run (or an increase in snow albedo by 0.06 in physical units), the February SWE is doubled in the mid-elevation range, 2250–2750m, (Figure 5b) and the March SWE becomes nearly five times as large as in the control run above the 2750m (Figure 5c). The impact of the decrease in snow albedo on SWE is also larger in higher elevations. Another noticeable result in the experiment is that the timing of the peak sensitivity varies according to both the sign of the snow albedo changes and terrain elevations. The peak percentage reduction in SWE due to a decrease in snow albedo occurs earlier than the peak percentage increase in SWE due to increased snow albedo. This is most evident in the mid-elevation range (2250–2750 m), where the largest reduction in the SWE due to decreased snow albedo (red lines) occurs in January, and the largest impact of the increased snow albedo (blue lines) occurs in February. The actual time lag varies according to the interval of the terrain elevation range in which the average is taken; however, the timing of the peak responses presented above remains qualitatively consistent. The discrepancy between the timing of the peak response of SWE (Figure 5) and the surface albedo (Figure 3) to snow albedo reveals that the alterations in snow albedo are further amplified through local snow-albedo feedback.

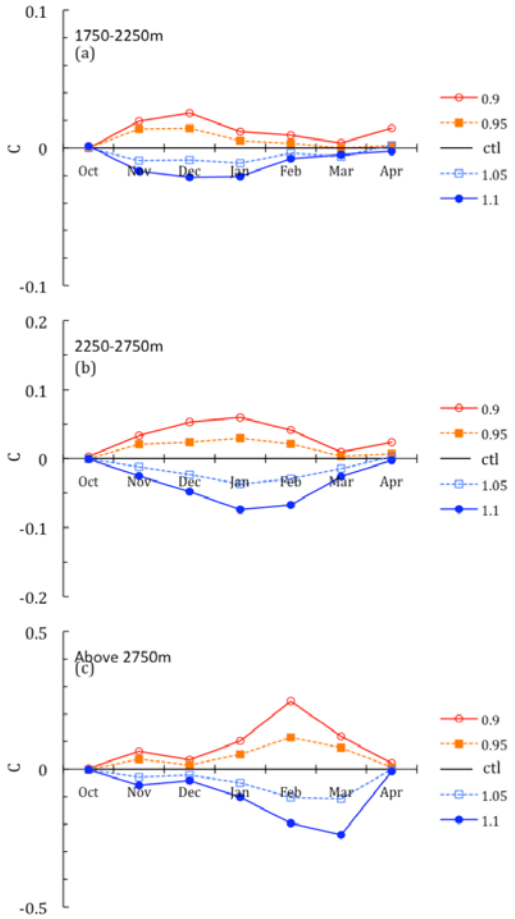

Figure 6. The sensitivity of the 2-m air temperature to the prescribed snow albedo changes

The response of the simulated 2-m air temperature to the prescribed snow albedo changes also varies according to terrain elevation and the magnitude of the snow albedo change (Figure 6). The prescribed snow albedo changes have minimal impacts on the 2-m temperature in the lowest elevation range; with snow albedo changes by ±10% resulting in temperature changes of less than 0.04°C (0.07°F). The 2-m temperature sensitivity to the albedo changes increases in higher altitudes; by nearly 0.1°C (0.18°F) in the mid-elevation range (Figure 6b) and by over 0.2°C (0.36°F) in the highest elevation range (Figure 6c). Not only the magnitude but also the timing of the peak sensitivity varies with both elevation and the sign of the snow albedo change in a similar way as other variables. In general, the peak warming due to the decrease in snow albedo occurs earlier than the peak cooling by the increase in snow albedo. The skin temperature responds to the snow albedo changes similarly, but with almost twice the magnitudes (not shown).

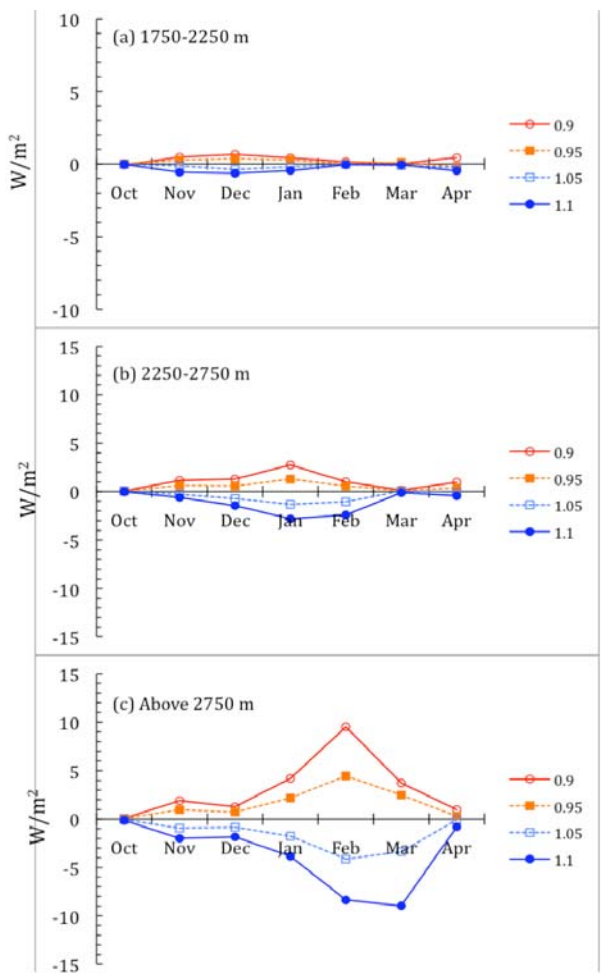

Figure 7. Same as Figure 6, but the sensible heat flux

The sensible heat flux varies according to the changes in snow albedo (Figure 7) similarly as the absorbed insolation. The decrease in snow albedo results in a larger decrease in the skin temperature than in the 2 m air temperature as discussed above, and in turn, results in the increase in the sensible heat flux. The increase in snow albedo causes the opposite effect on the sensible heat flux. The sensible heat flux sensitivity also displays clear dependence on terrain elevation and month. The sensible heat flux sensitivity increases with increasing terrain height; about 1 Wm^{-2} in the lowest elevation range (Figure 7a) and around 10 Wm^{-2} in the highest elevation range in response to ±10% changes in the snow albedo. Also, the peak response to the decreased snow albedo occurs earlier than that to the increase in snow albedo as seen above for the sensitivity of other variables.

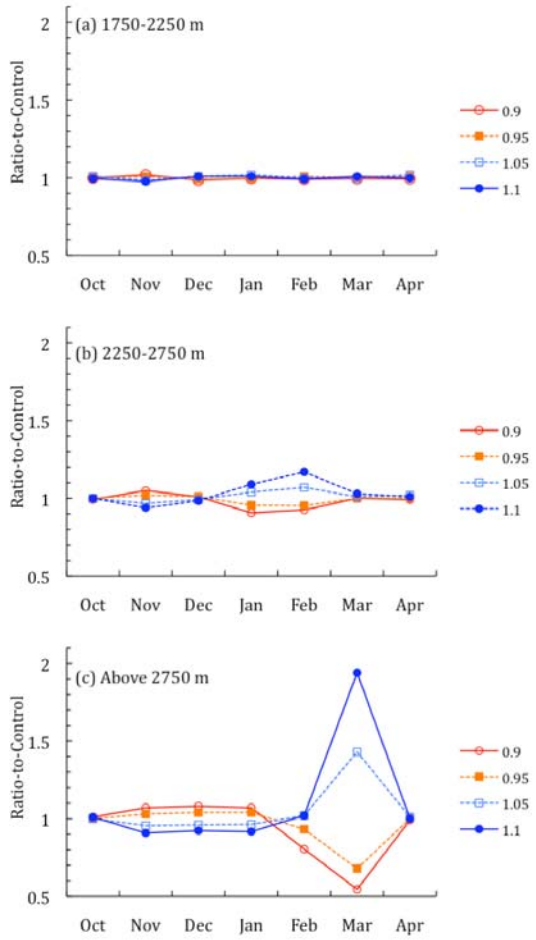

Figure 8. The sensitivity of the snowmelt to the snow albedo changes. The sensitivity is presented in terms of the ratio of snowmelt in each sensitivity study to that in the control run.

With the decrease in snow albedo, snowmelt increases in earlier months of the cold season. The timing of the increased snowmelt also appears in later months as terrain elevation increases (Figure 8). The snowmelt changes in the lowest elevation range (Figure 8a) are negligible because snowfall in the low elevation region melts quickly due to higher atmospheric temperatures. Thus the impact of snow albedo changes on the snowmelt in the low-elevation range is small. In the mid-elevation region (Figure 8b), the snowmelt increases by 5% in November, followed by a 10%

decrease in January in response to the decrease in snow albedo by 10% (red solid line with open circles). Peak snowmelt changes (i.e., early-season increase and late-season decrease) in the highest elevation region (Figures 8c) due to smaller snow albedo values occur one-month later than in the mid-elevation region and with much larger magnitudes. Thus, the most notable impact of the decrease in snow albedo is to enhance early-season snowmelt and to reduce late-season snowmelt, resulting in an adverse impact on warm season water resources in California. The two experiments with larger snow albedo values (the blue solid and dashed lines in Figure 8) show that an increase in snow albedo will suppress snowmelt in the early part of the cold season and will enhance it in the later part of the season. Such a change might help to partially alleviate the adverse impact of global warming on California water resources. The timing of peak impact of the altered snow albedo on the snowmelt also varies with elevation similarly as for SWE—that is, the peak response appears later in higher regions than in lower regions, especially in the cases of increased snow albedo.

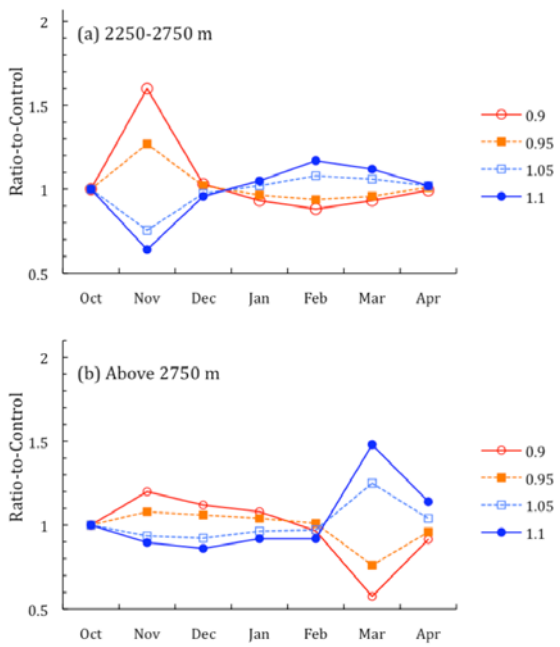

Figure 9. The runoff sensitivity to the snow albedo changes in the mid- and high-elevation ranges

The changes in snowmelt due to the alterations in snow albedo result in notable changes in runoff during the early and late part of the cold season (Figure 9). Decreases in snow albedo result in runoff increases in the early part of the cold season and decreased runoff in the late part of the cold season. Increases in snow albedo result in opposite effects; a decrease in runoff during the early cold season and an increase in runoff in the late cold season. This runoff response to increased snow albedo is qualitatively consistent with the corresponding responses of SWE and snowmelt. Similar to the SWE and snowmelt, the timing of peak response of the simulated runoff to snow albedo occurs later in the cold season as terrain elevation increases.

3.2. Snowpack Differences Between Single- and Multi-layer Snow Model Simulations

An additional uncertainty in simulating snowpack, especially snowmelt processes, derives from the formulation of physical processes within snowpack such as snow compaction, heat conduction, snow grain growth, and the retention of liquid water within snow, among others (Yang et al. 1997; Sun et al. 1999). Accommodating these processes in simulating snowpack requires an efficient snowcover layering system and has led to the development of several multi-layer snow models (e.g., Sun et al. 1997; Dai et al. 2003). Single-layer snow models that are widely used in many broadly-applied climate models often lack important physical processes such as the development of the vertical temperature gradient within snowpack that plays a crucial role in snowmelt. This is an important concern in assessing the impact of anthropogenic climate change on California's water resources. In this study, we compare two snowpack simulations simulated using a single- and three-layer snow models within the context of the WRF-SSiB model in order to examine the snowpack treatment in regional climate simulations.

The SSiB-3 uses the three-layer version of the Simple Atmosphere-Snow Transfer (SAST) model of Sun et al. (1999) that was developed on the basis of up-to-date comprehensive and complex snow schemes (Anderson 1976; Jordan 1991) with a number of simplifications and improvements. The SAST model includes three prognostic variables: specific enthalpy, SWE, and snow depth. Using enthalpy instead of temperature, in the energy conservation equation greatly simplifies the computational procedure for calculating phase change within snowpack. The model also retains important physical processes such as snow compaction, heat conduction, snow grain growth, and snow melting. Details of the SAST model and the coupling of the SAST and SSiB-3 are presented in Sun et al. (1999) and Xue et al. (2003). The SSiB-1 model is an early version of SSiB-3 in which a single-layer snow scheme is utilized. Details of SSiB-1 can be found in Xue et al. (1991).

Two seasonal simulations in which the WRF model coupled with SSiB-1 and SSiB-3 are used for the period April–June 1998 over the North American domain (Figure 1b). Both simulations are initialized from the National Centers for Environmental Prediction (NCEP)/NCAR reanalysis (Kalnay et al. 1996). The large-scale forcing along the lateral boundaries is obtained from the reanalysis data as well. The SWE fields during May 1998 in the two simulations are compared with the observed data over the western U.S. region for evaluation. For more details of the observed snow data, the readers are referred to Mote (2003) and Mote et al. (2005a,b). The differences in the simulated SWE between the single- and three-layer snow models are presented for two sub-regions in the western United States (Figure 10): Pacific coastal region (P) that includes the Coastal Range, the Sierra Nevada and the Cascades, and the Rocky Mountains region (R). The orography in the P region varies with characteristic zonal length scales between 50 and 100 km and is not well represented by the 80 km resolution. The zonal length scale of the terrain in R (Rocky Mountains) is of hundreds of kilometers and is represented better than in P.

The SWE fields simulated with SSiB-1 (Figure 10b) and SSiB-3 (Figure 10c) are compared against the observational data (Figure 10a). Both snow models significantly underestimate SWE in P, perhaps due to insufficient representation of the orographic variations in P due to the relatively coarse resolution. The snow-covered area in Oregon and California compares well with the observed data, but the significant SWE in northern California and in northern

Washington is absent in both runs. Both snow models represent the SWE field in R better than in P. The major SWE centers along the Rocky Mountains in R appear both simulations. However, the spatial details of the observed SWE distribution such as the SWE maximum in the northeastern Arizona region, are missing in both simulations, showing the problem with coarse spatial resolution.

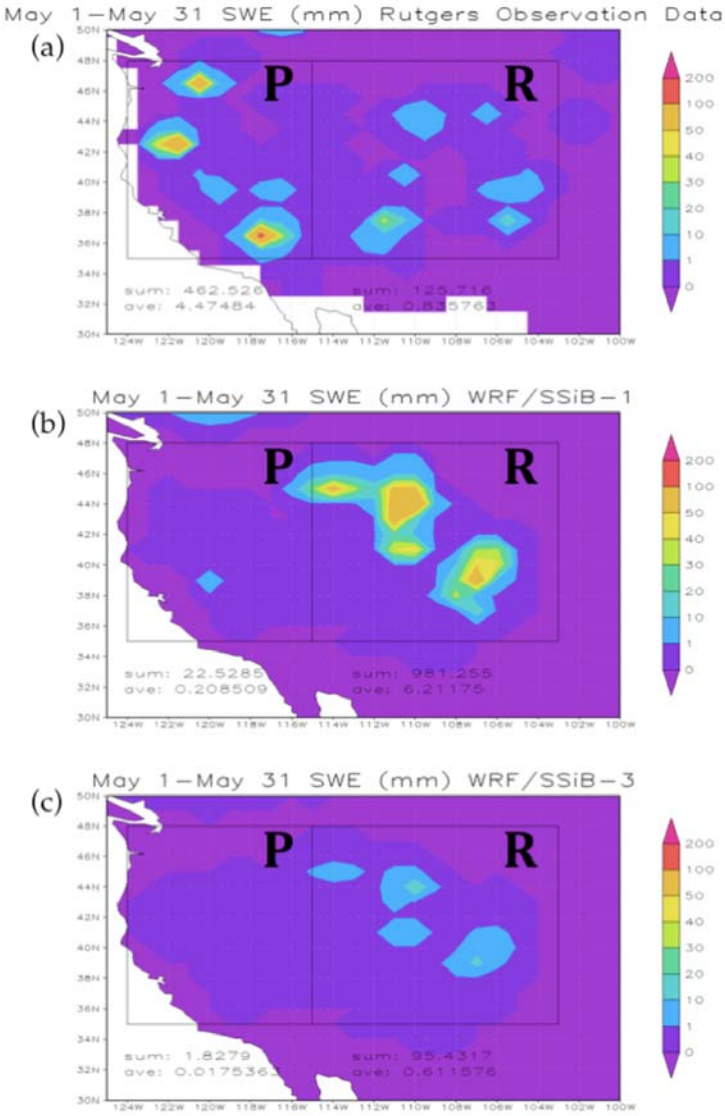

Figure 10. The monthly mean SWE in millimeters (mm): (a) Observation, and simulated with (b) a single layer snow model (SSiB-1) and (c) a 3-layer snow model (SSiB-3)

Figure 11 presents the model errors in terms of the difference between the simulated and observed monthly-mean SWE and the corresponding root-mean-square error (RMSE). In P, both the single- and three-layer model underestimates SWE, especially in the northern California and the central Sierra Nevada regions. The three-layer model results in slightly larger bias (-4.46 mm)

than the single layer model (-4.27 mm) in the area-mean SWE values, but the RMSE in the three-layer simulation is slightly smaller than the single layer model (11.16 mm versus 11.33 mm). Overall, the difference in the area-mean bias and the RMSE between the two snow model simulations is small for the P region. The SWE bias in R shows that the three-layer snow model could improve the snowpack simulation over the single-layer snow model. The significant SWE bias of +5.38 mm in the single-layer snow simulation is reduced to -0.22 mm in the three-layer model results. The three-layer snow model simulation also improved the RMSE from 17.4 mm to 3 mm in the R region. Evaluations of the results for Canada and northeastern United States (not shown) also reveal that a significant improvement in simulating snowpack could be achieved for those regions by the use of a multi-layer snow model.

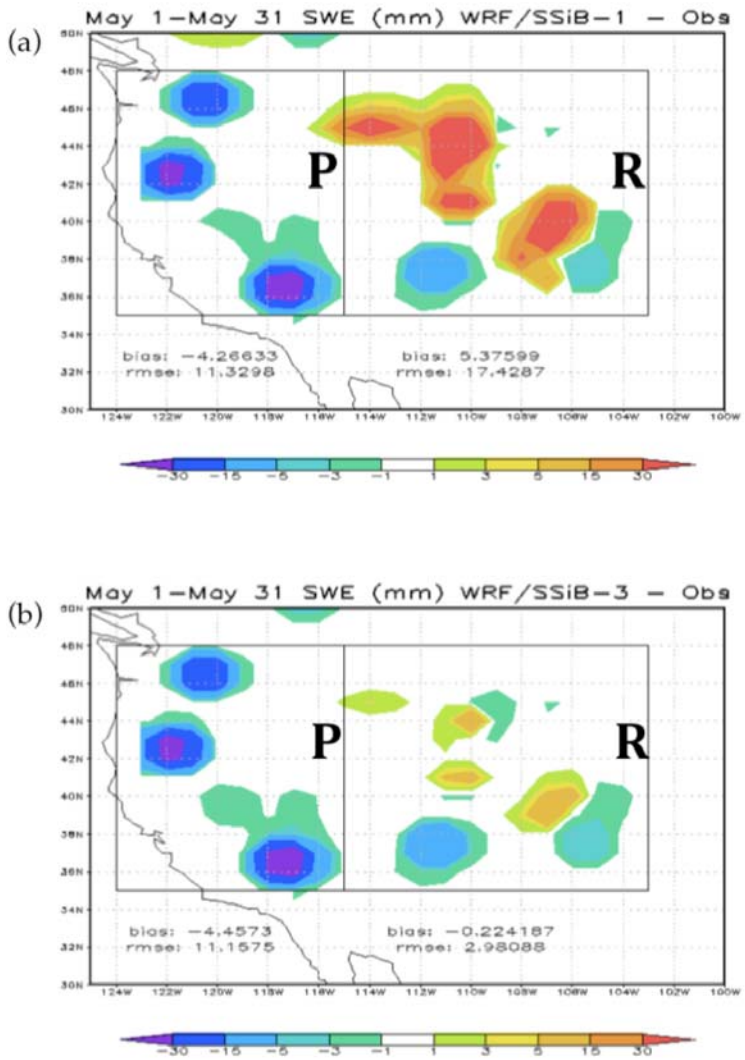

Figure 11. The monthly mean SWE (mm) simulation errors against observation: (a) a single layer snow model (SSiB-1) and (b) a 3-layer snow model (SSiB-3)

These results show that the lack of representation of physical processes within the snowpack, in this case most likely the vertical transfer of solar and thermal energy and the associated development of temperature gradient within the snowpack, can be a source of significant errors in simulating snowpack during the snowmelt period. The results in this study are highly qualitative, mainly due to the coarse horizontal resolution; however, the improvement in simulating snowpack during major ablation periods by the use of a multi-layer snow model is well demonstrated. Moreover, for sometime, climate models (especially those used for global climate projection studies) will be applied with even lower resolution so these results are quite relevant to understanding the uncertainties associated with these projections. A high-resolution experiment in conjunction with more detailed observed data (e.g., the fine-resolution snow analysis by the National Operational Hydrologic Remote Sensing Center (www.nohrsc.nws.gov/) is necessary for a more quantitative analysis of the differences between the single- and multi-layer snow model results. This is a subject of our follow-up experiment.

4.0 Conclusions and Discussions

Cold season snowpack plays a crucial role in determining the warm-season water supply in California and the western U.S. region. Apart from water supply, snowpack in the Sierra Nevada Mountains plays an important role in tourism and in establishing and maintaining the region's ecosystems. Projections of snowpack are uncertain due to a number of reasons, with two paramount ones being the alterations in snow albedo due to possible changes in the deposition of anthropogenic BC and the representation of snow physics in a numerical model. To better understand and quantify the uncertainties in projecting the impact of anthropogenic climate change on California hydrology, specifically snowpack, we have carried out and analyzed two regional climate model (RCM) experiments focused on these two issues.

The possible effects of the deposition of anthropogenic BC on snow albedo and subsequently on the evolution of the Sierra Nevada snowpack have been investigated using the WRF model by varying the default snow albedo field provided with the NOAH LSM. To represent the impact of changes in aerosol deposition (e.g., changes in local emission or transport) on snow albedo in a qualitative way, the snow albedo is decreased or increased by 5% and 10% of the default value in four sensitivity simulations. The ±10% alteration in the default snow albedo corresponds to ±0.06 in physical units that can occur with a moderate amount of BC deposition on snowpack. The decrease and increase of snow albedo can, for example, qualitatively represent the increase and decrease of local anthropogenic BC emissions, respectively. The simulations are performed for the cold season October 2050–April 2051 using the initial and lateral boundary forcing data from an NCAR-CCSM3 climate change projection generated with the SRES-A1B emission scenario.

The control simulation—using the default observation-based snow albedo values provided with the NOAH LSM— shows that the snowfall amount and SWE in the Sierra Nevada region vary significantly according to terrain elevation. The elevation dependence is caused mainly by the large low-level temperature variations associated with significant variations in terrain heights, the tendency of increasing precipitation with increasing terrain elevation (e.g., Kim 1997; Kim and Lee 2003), and the location of freezing level that appears between the 2000 m and 2250 m levels in the region during the winter. Previous studies over complex terrain (e.g., Giorgi et al.

1997; Kim 2001) also found significant differences in precipitation and snow accumulation across the seasonal mean freezing level altitude.

Decreased snow albedo (increased aerosol deposition) promotes early season snowmelt and runoff, resulting in smaller SWE in the early part of the cold season. The reduced early season SWE in turn results in reduced snowmelt and runoff in the later part of the season. Thus, reduced snow albedo that can occur by the increase in local emissions and subsequent BC deposition on the Sierra Nevada snowpack can enhance earlier snowpack depletion. This will further reduce warm season water resources and add to the effects of the low-level warming induced by increased greenhouse gases. An increase in snow albedo, possibly from the reduction in anthropogenic BC emissions, reduces snowmelt and runoff, and thus increases SWE, in the early part of the cold season. The resulting increase in early season SWE enhances late season snowmelt and runoff. This can partially alleviate one of the most important adverse impacts, the loss of snowpack earlier in the cold season to deplete the snowmelt runoff in spring and early summer, of the anthropogenic global warming on the Sierra Nevada snowpack and California's warm season water resources.

Examinations of the response of the major components in surface energy balance, the grid-mean surface albedo, the absorbed surface insolation and the sensible heat flux, show that the impact of the snow albedo changes is most pronounced in high elevation regions where snow cover is more significant. The grid-mean surface albedo that is calculated by a area-weighted average of the snow albedo and snow-free albedo, varies with snow albedo, most noticeably in higher elevation range. Decreases (increases) in snow albedo result in an increase (decrease) in the absorbed surface insolation, surface and 2-m temperatures, and sensible heat fluxes as well. The variations in SWE, surface energy fluxes, snowmelt, and the grid-mean surface albedo in response to the alterations in snow albedo support that the alterations in snow albedo is further amplified through local snow-albedo feedback.

This study's examination of the sensitivity of snowpack hydrology to changes in snow albedo is qualitative; however, it suggests that changes in anthropogenic emissions can influence the warm-season water supply in California via the snow albedo and, subsequently, the cold-season snowpack. The largest uncertainties in the current study are due to the lack of quantitative knowledge on two important factors: the amount of aerosol deposition on the Sierra Nevada snowpack and the relationship that links the amount and type of aerosol deposition and snow albedo. This study's results reveal that the modification of snow albedo, possibly due to anthropogenic aerosol deposition, can result in significant alterations in the Sierra snowpack, and as a result, changes in the warm season water supply in California. Quantifying the amount of anthropogenic aerosol deposition on the Sierra snowpack and the associated snow albedo changes is a topic of future research.

A comparison of the SWE fields simulated using the WRF-SSiB model with the single-layer and multi-layer snow models indicates that systematic errors in the SWE during a major snow ablation period simulated using a single-layer snow model could be alleviated significantly by using a multi-layer snow model. The latter typically includes more comprehensive treatment of vertical variations in the absorption of insolation and, subsequently, the temperature gradient within the snowpack, in addition to other important physical processes such as snow compaction, heat conduction, and snow grain growth. The results in the Pacific coastal regions are inconclusive, possibly due to the use of a horizontal resolution still too coarse to represent

the orographic variations in the region. However, the improvement in simulating snowpack is clearly shown in the Rocky Mountains region where the terrain is represented better than in the Pacific coastal region. Despite shortcomings due to the relatively coarse spatial resolution, this study shows that it is important to include a comprehensive snow model in assessing the climate change impact on water resources in California. Future works on this subject will focus on the Sierra Nevada region with a finer spatial resolution and a more detailed observational dataset in order to examine the model errors associated with the numerical structure and the representation of the underlying physical processes.

Development of the formulations to account for the impact of aerosol depositions on snow albedo and the treatment of physical processes, as well as overall snow modeling, has been difficult due to the lack of reliable observational data. Thus, comprehensive observational studies, particularly in conjunction with process/modeling studies, are necessary to improve snow modeling.

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