Analysis of the Impact of Snow on Daily Weather Variability in Mountainous Regions Using MM5

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ABSTRACT

The impacts of snow on daily weather variability, as well as the mechanisms of snowmelt over the Sierra Nevada, California–Nevada, mountainous region, were studied using the fifth-generation Pennsylvania State University–National Center for Atmospheric Research Mesoscale Model (MM5) forced by 6-h reanalysis data from the National Centers for Environmental Prediction. The analysis of two-way nested 36–12-km MM5 simulations during the 1998 snowmelt season (April–June) shows that the snow water equivalent (SWE) is underestimated when there are conditions with higher temperature and greater precipitation than observations. An observed daily SWE dataset derived from the snow telemetry network was assimilated into the Noah land surface model within MM5. This SWE assimilation reduces the warm bias. The reduction of the warm bias results from suppressed upward sensible heat flux caused by the decreased skin temperature. This skin temperature reduction is the result of the longer assimilated snow duration than in the model run without SWE assimilation. Meanwhile, the cooled surface leads to a more stable atmosphere, resulting in a decrease in the exaggerated precipitation. Additionally, the detailed analysis of the snowmelt indicates that the absence of vegetation fraction in the most sophisticated land surface model (Noah) in the MM5 package results in an overestimation of solar radiation reaching the snow surface, giving rise to heavier snowmelt. An underestimated surface albedo also weakly contributes to the stronger snowmelt. The roles of the vegetation fraction and albedo in snowmelt are further verified by an additional offline simulation from a more realistic land surface model with advanced snow and vegetation schemes forced by the MM5 output. An improvement in SWE description is clearly seen in this offline simulation over the Sierra Nevada region.

1. Introduction

Snowmelt runoff is a major water resource in the western United States, as 50%–70% of the annual precipitation in the mountainous regions (Serreze et al. 1999) is snowfall, and 75%–85% of the annual streamflow required for agricultural irrigation and other human activities is from snowmelt runoff (Grant and Kahan 1974). Recent studies using general circulation models (GCMs) that force hydrological streamflow models (e.g., Hamlet and Lettenmaier 1999; Miller et al. 1999; McCabe and Wolock 1999; Knowles and Cayan 2002; Miller et al. 2003) indicate that global warming is likely to result in significantly decreased snow mass in the western United States. The GCM-projected temperature increase would shift runoff peaks from spring to winter, decreasing the spring and summer streamflow amounts. Rain falling on snow and spring warming can also bring about heavy snowmelt runoff over very short time periods, potentially resulting in severe flood events. This was seen in the Sierra Nevada, California, in February 1998, when a warm storm caused the Central Valley to flood (Bowers 2001). Thus, forecasting snowmelt amounts and rates are essential to managing water resources and reducing flood risks in this region.

Climate and weather variations strongly affect the snow water equivalent (SWE) (Aguado 1990; Aguado et al. 1992; Cayan 1996; Neale and Fitzharris 1997; Aizen et al. 2000). Cayan (1996) found that SWE anomalies and the resulting streamflow for different areas in the western United States are highly correlated to atmospheric circulation patterns and sea surface temperature (SST) anomalies. Historical records indicate that early snowmelt often resulted from a warmer win-
ter, partly due to a shift in the atmospheric circulation over the north-central Pacific Ocean and the west coast of North America (Dettinger and Cayan 1995). Clark et al. (2001) investigated historic effects of El Niño–Southern Oscillation (ENSO) episodes on seasonal SWE evolution in the Columbia and Colorado River basins, and found that ENSO-related midlatitude atmospheric circulation anomalies influence the SWE in both basins. Johnson et al. (1984) found that mesoscale wind circulation could be generated in an area that has both snow and bare soil present. Snow cover area (SCA) variations have also been used to predict the strength of the North American monsoon circulation (Gutzler and Preston 1997). Furthermore, snow cover causes a temperature depression in the lower troposphere due to the high solar reflectivity of the snow surface and the large energy consumption of snowmelt (Groisman et al. 1994). These processes stabilize the lower atmosphere, which may lead to temperature inversions in the boundary layer.

In the western United States, SWE forecast accuracy remains a challenging problem due to the lack of sufficient observations and the strong heterogeneity of SWE distributions resulting from complex topography. Numerical models are promising tools for SWE forecasting, and efforts have been made using GCMs to simulate and predict the SWE in this region (Marshall and Oglesby 2003). However, GCMs are unable to adequately capture snow-related atmospheric processes in mountainous areas because of overparameterized physics, coarse spatial resolution, and in many models, simplified land surface schemes. Improvements in the atmospheric water and energy transport were introduced to regional climate models (RCMs) by Grell et al. (1994) to better understand the physical processes affecting SWE evolution in mountainous regions (Liston et al. 1999; Kim et al. 2000). Nevertheless, coarse treatment of land surface snow physics in these models results in poorly simulated SWE and related processes (Leung and Qian 2003).

The objectives of this study are to quantify the impact of snow on temperature and precipitation in the Sierra Nevada region during the snowmelt season by applying an assimilation methodology to an RCM. This will lead to a better understanding of the reasons for the poorly simulated SWE and associated processes, and provide new insight into how to improve SWE simulations. Liston et al. (1999) applied similar assimilation techniques to an RCM and evaluated the importance of SWE to regional climate and water resources over the Colorado River basin.

In the present study, we use an assimilation approach described in Liston et al. (1999) and documented in Sun et al. (2004) to explore the detailed physical processes in which SWE affects daily weather variability, identify the deficiencies of the RCM in the simulation of SWE, and detect the mechanisms of snowmelt. Here a two-way 36–12-km simulation was generated and analyzed using the fifth-generation Pennsylvania State University–National Center for Atmospheric Research (Penn State–NCAR) Mesoscale Model (MM5, version 3.6) (Grell et al. 1994). An additional simulation was performed and analyzed, in which the observed SWE for the Sierra Nevada region was assimilated into MM5. Through the SWE assimilation process, the impact of snow on the 2-m height air temperature ($T_a$) and precipitation ($P$) is explored, and the deficiencies of the Noah land surface model (NoahLSM) (Ek and Mahrt 1991) are identified. In this study, the version 2.6 of NoahLSM coupled with MM5 was used. However, significant efforts have been made by Mitchell et al. (2004) to improve the offline snow simulation of NoahLSM. In their version (2.7) of NoahLSM, the albedo and snow sublimation treatments have been greatly improved. NoahLSM-2.7 has been coupled to theEta operational mesoscale model (Ek et al. 2003), resulting in an improvement in regional weather simulation. However, NoahLSM-2.7 was not used in this study, because a complete evaluation of its performance in MM5 is needed. In addition, part of our study focuses on identifying the deficiencies of the structure of NoahLSM and how the vegetation fraction affects snowmelt, which is independent of the studies included in Mitchell et al. (2004) and Ek et al. (2003). The study period is the snowmelt season, where the snow–atmosphere feedback is strongest due to high solar radiation (Groisman et al. 1994). The snow–atmosphere–soil transfer (SAST) model with more sophisticated snow and vegetation physics (Jin et al. 1999) is forced with the MM5 output to further understand snowmelt processes. In this paper, section 2 provides a description of the model and data used. The assimilation of SWE is discussed in section 3, and the temperature and precipitation simulations are included in sections 4 and 5, respectively. In section 6, the early snowmelt issue in the Noah model is investigated, and the conclusions are included in section 7.

2. Model and data

The nonhydrostatic version of MM5 is used in this study, where the Grell convection scheme is adopted to parameterize cumulus clouds (Grell 1993), and the Medium-Range Forecast (MRF) planetary boundary layer (PBL) scheme is applied to solve boundary layer processes (Hong and Pan 1996). The microphysics scheme
selected is the simple ice scheme developed by Dudhia and Bresch (2002). The radiation scheme chosen is the cloud–radiation scheme that accounts for longwave and shortwave radiation interactions with cloud and clear air (Dudhia and Bresch 2002). Among several land surface models in MM5, the advanced NoahLSM is selected to characterize land surface processes. The NoahLSM has four soil layers with a total depth of 2 m and a single snow layer lumped with the top soil layer. The vegetation scheme in the NoahLSM was advanced by Chen et al. (1996), with the canopy resistance approach of Noilhan and Planton (1989).

MM5–NoahLSM was used to produce a two-way 36–12-km resolution nested simulation with domains D01 and D02, respectively (Fig. 1a), and is defined as the “control run.” The total number of the grid cells in the 36-km resolution domain is 9801 (99 × 99), and in the 12-km resolution domain it is 10 989 (99 × 111). In this study, the 12-km simulation (D02) is the focus of analysis. Twenty-three vertical sigma layers from the surface to the 100-mb level were configured in MM5-NoahLSM. To properly represent the PBL processes, the vertical sigma layers are closely spaced near the surface and coarsely spaced in the upper atmosphere. The National Centers for Environmental Prediction (NECP)–NCAR reanalysis data were used as MM5 initial and 6-h updated lateral boundary conditions for the period 1 April to 30 June 1998. This period is the snowmelt season in the Sierra Nevada region, as snow usually reaches its maximum amount around 1 April (Serreze et al. 1999) and then starts to melt. The MM5 output was saved every 6 h. The SST from the NCEP–NCAR reanalysis data was used to initialize MM5 and was also updated every 6 h. Because of the poor quality of the snow depth in the reanalysis data over our simulation domains, we used the 0.125° × 0.125° (~10 km at 38°N) SWE data produced by the National Aeronautics and Space Administration (NASA)/Land Data Assimilation System (LDAS) to initialize the model for both 36- and 12-km-resolution domains. Additionally, we used the observed snow telemetry (Snotel) in the Sierra Nevada area to replace those LDAS data at the Snotel locations for the model initial conditions. We quality-controlled the Snotel station data according to the method described in Clark et al. (2001) and selected the stations measuring daily SWE, 2-m air temperature ($T_a$), and precipitation ($P$). The quality-control process resulted in 26 Snotel stations chosen in the Sierra Nevada area (Fig. 1b). The maximum SWE at these 26 Snotel stations is 2319 mm in 1998, a strong El Niño year, which usually generates higher than normal SWEs in the Southwest (Jin et al. 2006). A second MM5 simulation was performed in which the model settings were exactly the same as in the control run, but the observed daily Snotel SWE data were assimilated into MM5 during the entire simulation period. The second simulation is defined as the “assimilation run.” There exists approximately 45%–70% vegetation coverage with the remainder bare soil and rock over the Snotel stations used here. Twenty-two grid cells out of 10 989 in the 12-km resolution domain are filled by the 26 Snotel stations that are all located in the Sierra Nevada area (Fig. 1b). The Snotel stations have an elevation range
from 1889 to 2865 m, with a mean elevation of 2386 m. In MM5, the elevation range at 12-km resolution for the 22 grid boxes filled with the Snotel data is from 1872 to 2678 m, with a mean height of 2208 m very close to the mean value for the Snotel stations. Weekly 4 km × 4 km satellite-derived vegetation fraction and surface albedo data from the Sea-viewing Wide Field-of-view Sensor (SeaWiFS) land products for 1 April to 30 June were used for evaluating the model simulations. The good reliability of the SeaWiFS data has been verified by McClain et al. (1998).

The NoahLSM in MM5 has a lumped soil–snow mass with SWE calculated using mass and energy balance equations. Different from the NoahLSM, the SAST model (Jin et al. 1999) separates the soil from the SWE by using three layers, each containing liquid water and solid ice, with both SWE and snow depth being computed. The soil temperature and moisture in the SAST model are solved numerically using 10 soil layers. The SAST vegetation scheme is based on Dickinson et al. (1993), and explicitly describes interactions between vegetation and snow that are ignored in the NoahLSM. A SAST offline simulation forced by the MM5 output averaged over the Snotel stations was compared with the 12-km resolution results from the NoahLSM in MM5 for 1 April to 30 June 1998.

3. Assimilation of SWE

In the western United States, water resources agencies measure the SWE on or around 1 April, a time when the SWE reaches its maximum amount as mentioned in section 2 (Serreze et al. 1999), to determine the spring and summer water allocations. Figure 2 shows the time series of observed and simulated (12-km resolution) daily SWE averaged over the 26 Snotel stations for the period 2 April to 30 June 1998. A higher snowmelt rate is seen in the control run during the simulation period (dashed line) when compared to the observations (solid line).

Here, we examine the reasons for the poorly simulated SWE in the control run and the significance of how snow affects temperature and precipitation. The Snotel SWE assimilation within the MM5 domains allow for an investigation of the SWE evolution and related processes. During the assimilation process, the modeled SWE was replaced by the Snotel data at 0000 UTC of each day in the model grid boxes where the Snotel stations exist (Fig. 1b). For cases with more than one Snotel SWE data per grid box, these data were geometrically averaged before updating. The sophisticated internal nudging system within MM5 was not activated during the assimilation process in this study. Based on the findings in Castro et al. (2005), the internal nudging is not necessary for fine resolution and small size domains (e.g., domains finer than 50-km resolution with a size smaller than 160 × 100 grid boxes). The domains used in this study apparently meet their criteria for fine resolution and small size domains.

4. Simulation of temperature

Time series of observations and 12-km $T_a$ simulations with and without the SWE assimilation averaged over the Snotel stations for 2 April to 30 June 1998 are shown in Fig. 3. This figure indicates that without the
SWE assimilation, MM5-NoahLSM produced strong warm biases during the snowmelt season. The mean modeled warm bias for the control run is 4.7°C. With the SWE assimilation, the model-simulated $T_a$ agrees well with observations, and the bias decreases to 1.3°C (Fig. 3).

Figure 4a illustrates that the mean sensible heat flux ($H$) over the simulation period with the SWE assimilation is 10.1 W m$^{-2}$ compared to 34.1 W m$^{-2}$ without SWE assimilation. The decreased $H$ transmits less energy to the air and leads to the lower $T_a$ (Fig. 3). In addition, the decreased $H$ can be attributed to a decrease in the surface skin temperature ($T_s$) (Fig. 4b), which decreases by 6.3°C on average due to the SWE assimilation.

Figure 4b shows that the $T_s$'s are almost identical for the assimilation and control runs before 1 May, but the significant $T_s$ differences between the two runs occur in May and June, which give rise to the $H$ and $T_s$ differences as shown in Figs. 4a and 3. The mean value of the $T_s$ differences is 9.5°C (Assimilation–Control) for May and June. Examination of the snow coverages at the 26 Snotel stations studied indicates that the number of the snow-covered stations shows a faster decrease in the control run than in the assimilation run (Fig. 4c) after 1 May, from which the $T_s$’s for these two runs start to have significant differences. In the control run, snow completely melts out at all of the 26 Snotel stations on 10 June while there are still 22 snow-covered stations in the assimilation run, resulting in the lower $T_s$ due to energy consumption by snowmelt. In addition, although the number of the snow-covered stations in the assimilation run is significantly larger than that in the control run, the SWE assimilation increases the albedo by only 0.02 due to large forest coverage in this area (~45%–70% based on the SeaWiFS data during the snowmelt season, which is discussed in Fig. 8b). Thus, the impact of the minor albedo increase on $T_s$ is negligible, which is explained again later. Therefore, the longer snow duration in the assimilation run, where a stronger snowmelt occurs (Table 1), is the major reason for the lower $T_s$.

The $T_s$ reduction during May and June can be also quantitatively understood by analyzing variations of all components in the surface energy balance equation (Chen and Dudhia 2001):

$$ (1 - \alpha) S_\downarrow + L_{\text{net}} - H - LE - H_m - G = 0, \tag{1} $$

where $\alpha$ is the surface albedo, $S_\downarrow$ is downward solar radiation, $L_{\text{net}}$ is the net longwave radiation at the surface, $G$ is the ground heat flux, $L$ is the latent heat of vaporization (2.5e$^6$ J kg$^{-1}$), $E$ is evaporation (LE is the latent heat flux), and $H_m$ is the snowmelt heat flux. When snow is mixed with vegetation within a model grid box, the albedo is computed as an area-weighted average of vegetation and snow surface albedo. Examination of the components in Eq. (1) and Table 1 indicates that the change in net solar radiation with the SWE assimilation is only −0.3 W m$^{-2}$ and is ignored. The surface gains energy from the changes in $L_{\text{net}}, H, LE$, and $G$, and loses energy from the changes in $H_m$ (Table 1). The gained energy tends to increase $T_s$ and does not account for the $T_s$ reduction. Thus, only $H_m$ is responsible for the lowered $T_s$ in the assimilation run, where the $H_m$ is 136.9 W m$^{-2}$ greater than in the control run averaged over May and June. Therefore, the
lower $T_s$ in the assimilation run can be mainly attributed to more energy consumption by $H_m$ due to the longer snow duration, as mentioned earlier. In the control run, the faster snowmelt produces an earlier ending of the snow season at the Snotel stations, resulting in the faster increase in $T_s$ as shown in Fig. 4b.

As discussed above, the surface albedo increases by only 0.02 when the assimilated SWE is included (Table 1), because the dense forest coverage in areas of the Sierra Nevada region (around 58% on average during the snowmelt season) conceals a large part of the snow and greatly restricts the effect of snow on the surface albedo. The mild increase of the albedo-induced insolation loss (6.1 W m$^{-2}$ on average) is almost balanced by the insolation gain due to a decrease in cloudiness in the assimilation run (5.8 W m$^{-2}$ on average) (figure not shown). The decreased cloudiness results from a drier lower atmosphere due to the weaker evaporation (Table 1) and diminished vertical convection induced by the more stable atmospheric column. Details regarding atmospheric stability are given in section 5b.

5. Precipitation

Figure 5 represents the time series of the daily $P$ averaged over the 26 Snotel stations for 2 April through

<table>
<thead>
<tr>
<th>$\alpha$</th>
<th>$(1-\alpha)S$</th>
<th>$L_{net}$</th>
<th>$H$</th>
<th>LE</th>
<th>$H_m$</th>
<th>$G$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ctr</td>
<td>0.16</td>
<td>251.7</td>
<td>−57.5</td>
<td>34.1</td>
<td>59.3</td>
<td>95.9</td>
</tr>
<tr>
<td>Assim</td>
<td>0.18</td>
<td>251.4</td>
<td>−28.7</td>
<td>−10.1</td>
<td>43.3</td>
<td>186.2</td>
</tr>
<tr>
<td>Diff</td>
<td>0.02</td>
<td>−0.3</td>
<td>28.8</td>
<td>−44.2</td>
<td>−16</td>
<td>90.3</td>
</tr>
</tbody>
</table>

(Assim-Ctr)
30 June 1998. For the control run, the bias for the total simulated $P$ is 53 mm, but the bias decreases to 9 mm with SWE assimilation, indicating that snow has a significant influence on precipitation processes. MM5-simulated precipitation can be divided into convective and nonconvective precipitation:

$$P = P_{nc} + P_c,$$

(2)

where $P_{nc}$ is the nonconvective precipitation, and $P_c$ is the convective precipitation. In general, $P_c$ occurs when the atmosphere is unstable, while $P_{nc}$ occurs when the atmospheric column is stable, and air masses are cooled. Figures 5b and 5c indicate that more than 80% of $P$ is $P_{nc}$ during our study period, because cold near-surface air temperature over the mountainous region suppresses atmospheric convective activity during the spring, where the observed mean $T_a$ for April and May is 0.4°C, and the simulated mean $T_a$ with SWE assimilation is 1.9°C for the same period. Figure 5b shows that the SWE assimilation procedure results in a 16-mm $P_{nc}$ reduction (218 minus 202 mm). Meanwhile, the major $P_c$ events occurred in June 1998 when the weather became warmer. In this month, the mean $T_a$ is about 7°C higher than that for April and May for both observations and simulations with SWE assimilation. Figure 5c shows that the SWE assimilation diminishes the $P_c$ by

Fig. 5. Comparison of the observed (solid line) and simulated precipitation from the control run (dashed line) and the assimilation run (dotted line) averaged over the 26 Snotel stations for 2 Apr–30 Jun 1998. (a) Total precipitation; (b) convective precipitation; (c) nonconvective precipitation (unit: mm).
29 mm (from 55 to 26 mm) during the 3-month simulation period, contributing to 66% of the total P bias reduction.

a. Impact of cumulus parameterization on precipitation

To determine if the precipitation simulation can be further improved, three additional runs with SWE assimilation under different cumulus schemes were performed. In Table 2, all simulations, except for the Grell-Ctr run, have the same SWE assimilation with different cumulus schemes. Table 2 shows that only the Kuo cumulus convection scheme produces less P (197 mm) than observations (219 mm) and more Pc than Pnc. All other simulations overestimate P with Pnc dominant. Table 2 also shows that the Grell cumulus convection scheme with SWE assimilation generated the best results among all simulations evaluated here, but the Grell scheme without SWE assimilation produced the worst results. The comparison in Table 2 further stresses the importance of snow to the precipitation simulation in our study area.

b. Impact of SWE on precipitation

Convective precipitation Pc results from an unstable atmosphere, and equivalent potential temperature (θe) is a useful tool for measuring atmospheric stability. Figure 6a illustrates the mean profiles of the MM5-simulated θe for 1–15 June 1998, a period with significant Pc, averaged over the Snotel stations for the control and assimilation runs. Here θe can be defined as a function of moisture, temperature, and pressure:

$$\theta_e = T[1 + Lq/c_p T] (1000/P_r)_{R_g C_p}$$

where q is the mixing ratio (kg kg\(^{-1}\)), T is the air temperature, Pr is the pressure, and \(R_g\) is the gas constant for dry air (287 J K\(^{-1}\) kg\(^{-1}\)). Figure 6a shows that θe (averaged over the Snotel stations) decreases with increasing height from the surface to 600 hPa for each run, indicating that this part of the modeled atmosphere is unstable. The SWE assimilation produces a lower θe below 600 hPa. Based on Table 1, it is understood that the weaker surface evaporation and sensible heat flux in the assimilation run decrease the moisture and temperature in the lower atmosphere, and the drier and cooler air mass reduces θe, leading to a more stable atmosphere with decreasing Pc.

When q in Eq. (3) is set to 0, θe is equivalent to potential temperature \([T_n = T(1000/P_r)_{R_g C_p}]\), which describes atmospheric stability under the dry condition. Table 3 shows that \(T_n\) decreases from the 800-hPa level to the 600-hPa level in both assimilation and control runs (\(\Delta T_n\) is equal to −8.41° and −8.29°C, respectively), indicating that the lower part of the atmosphere is stable under the dry condition. However, \(\Delta \theta_e\) (defined in Table 3) increases by 1.73° and 2.22°C in the assimilation and control runs, respectively, which turns the stable atmosphere into an unstable one. Thus, the air moisture has a dominant impact on the atmospheric stability during the period of 1–15 June 1998. The SWE assimilation decreases the \(\Delta T_n\) by 0.12°C and \(\Delta \theta_e\) by 0.49°C, both tending to increase the atmospheric stability. The decrease in \(\Delta T_n\) and \(\Delta \theta_e\) indicates that 24% of the stability increase (\(\Delta T_n/\Delta \theta_e = 0.12/0.49 \times 100/\% = 24/\%\)) comes from the colder surface in the assimilation run, and 76% of it results from the weaker surface evaporation as shown in Table 1.

Figure 6b shows the vertical wind velocities from the control and assimilation runs. With SWE assimilation, the vertical velocity decreases by about 0.01 m s\(^{-1}\) between 800 and 400 hPa, as compared to the control run, and the SWE assimilation even leads to a downdraft below the 750-hPa level. The decrease in vertical velocity is attributed to the increased atmospheric stability and the colder air mass, resulting in the weaker Pc. The descending air mass may also cause moist air divergence to occur in the lower atmosphere (figure not shown), and partially contribute to the drier atmosphere in our study region. The drier atmosphere resulting from the weaker surface evaporation and the airmass divergence apparently reduces the magnitude of Pnc in the assimilation run under a stable atmosphere (Fig. 5b).

6. Simulation of the snowmelt

The accurate simulation of snowmelt is crucial to improving the forecasts of available water resources, especially in the Sierra Nevada (Grant and Kahan 1974). To evaluate the simulation of snowmelt, observed daily
SWE, $P$, and $T_a$ for the 26 Snotel stations were used to estimate snowmelt amount following:

$$H_m^t = L_i (SWE^{t-1} + P_s - SWE^t),$$

(4)

where $P_s$ is the snowfall, and is 0 when $T_a \geq 2.2^\circ C$, or $P_s$ is equal to the observed $P$ when $T_a < 2.2^\circ C$. The superscript $t$ represents the current day, and $t-1$ represents the previous day; $H_m^t$ is the estimated snowmelt heat flux at a Snotel station; and $L_i$ is the latent heat of fusion ($3.3e^5$ J kg$^{-1}$). Equation (4) is an approximation, as some factors such as the effect of rainfall on snowmelt are not included. A rain versus snow threshold temperature of 2.2$^\circ C$ was first introduced by Auer (1974) and was later adopted by Dickinson et al. (1993). Figure 7 shows that MM5-NoahLSM overestimates $H_m$ by 147.5 W m$^{-2}$ with SWE assimilation (the thin solid line) compared with that estimated using Eq. (4).

As discussed above, the SWE assimilation improves the $T_a$ simulation in comparison with that from the control run without SWE assimilation (Fig. 3). In addition, when snow is melting, the maximum $T_a$ is restricted to the melting point (0$^\circ$C) in MM5-NoahLSM no matter how much snow mass melts, indicating that $T_a$ is not directly related to, or is decoupled from, the snowmelt amount. Thus, relatively accurate temperatures can be simulated even though there are large errors in the snowmelt simulation. To identify the reasons for the snowmelt bias, the land surface features are examined in the following section.

Effects of the land surface characteristics on snowmelt variability

The NoahLSM has a simple slab snow module, and the snowmelt is computed from the energy balance equation [Eq. (1)]. The surface albedo is the average of the vegetation and snow/soil albedos, weighted in terms of their areal percentages in the model grid box. Equation (1) indicates that when snow is present, all the net solar radiation is used to estimate snowmelt. This esti-

\begin{table}[h]
\centering
\begin{tabular}{|c|c|c|}
\hline
Assim (°C) & Control (°C) & Assim-Control (°C) \\
\hline
$\Delta T_h$ & -8.41 & -8.29 & -0.12 \\
$\Delta \theta_e$ & 1.73 & 2.22 & -0.49 \\
\hline
\end{tabular}
\caption{The comparison of $\Delta T_h$ and $\Delta \theta_e$ in the control and assimilation runs ($\Delta T_h = T_h$ at the 800-hPa level $- T_h$ at the 600-hPa level; $\Delta \theta_e = \theta_e$ at the 800-hPa level $- \theta_e$ at the 600-hPa level).}
\end{table}

Fig. 6. Simulated vertical profiles of (a) equivalent potential temperature (°C); (b) vertical wind velocity (m s$^{-1}$) (the positive is upward) from the control run (dashed line) and assimilation run (dotted line). The profiles are averaged over the 26 Snotel stations for the period 1–15 Jun 1998.
mation is accurate when a model grid box is completely covered by snow, where 100% of the insolation can reach the snow surface. However, if there is fractional vegetation cover, and the vegetation is mixed with snow in the grid box, then only part of the solar radiation is able to reach the snow surface, and the remaining solar radiation is absorbed by the vegetation. In this case, if the entire solar radiation is still used to estimate snowmelt without considering the influence of the vegetation fraction, a faster snowmelt will occur, as seen in the NoahLSM. The SeaWiFS satellite-derived data indicate that the Sierra Nevada are covered by high and dense forest, and the vegetation fraction varies from 45% to about 70%. Thus, we have shown that the NoahLSM overestimates the snowmelt over the Snotel stations (the solid line in Fig. 7). Although there is a 0.02 increase in the surface albedo when the SWE assimilation process is included in the NoahLSM, the surface albedo with the maximum value of 0.18 is still underestimated during a large part of the simulation period, when compared to the SeaWiFS satellite-derived data, where the highest surface albedo averaged over the Snotel stations is 0.43. To understand how the surface albedo impacts the snowmelt, the solar radiation absorption at the snow surface is recalculated using the observed surface albedo. The first term in Eq. (1) is changed to \((1-\alpha_{\text{obs}})S_\downarrow\) to recalculate \(H_{m}\) as illustrated in Fig. 7, indicating that the reestimated mean \(H_{m}\) decreases slightly from 186.5 to 158.5 W m\(^{-2}\) (Table 3). The exception to this reduction is toward the end of the simulation period (15–30 June), where the reestimated \(H_{m}\) becomes stronger than that from the SWE assimilation run, resulting from the observed lower albedo (0.14 in comparison with 0.18 from the assimilation run averaged over the last 16 days of June). Examination of the SeaWiFS data indicates that the observed vegetation coverage nears its maximum, approximately 70%, during the 16-day period (15–30 June), and is a possible reason for the lower albedo. Figure 7 indicates that the impact of albedo on snowmelt is marginal. However, the vegetation cover in reality apparently affects the solar radiation budget at the snow surface. Hence, there is a need to understand influences of vegetation cover on snowmelt.

Because of the formulation of Eq. (1) it is difficult to include the vegetation fraction and related processes in NoahLSM to describe the influences of vegetation on SWE. Therefore, the SAST model (Jin et al. 1999) with sophisticated snow and vegetation schemes was used to explore how vegetation fraction affects snowmelt. SAST was forced with the MM5 assimilation run output. The forcing data from the MM5 output over the Snotel stations were averaged as an area-mean input to SAST in single-column mode. The simulated \(H_{m}\) are presented in Fig. 7 (the dotted line) and Table 4, showing that the mean \(H_{m}\) decreases to 44.8 W m\(^{-2}\) for the simulation period, which is very close to the observed mean value of 38.7 W m\(^{-2}\) (Table 4). Furthermore, the SAST-simulated albedo and vegetation fraction are generally in good agreement with the observations (Fig. 8). However, the contribution of the vegetation fraction to this improvement is still not fully understood. To
isolate the role of the vegetation fraction within the snow simulation, an additional offline simulation was performed with the SAST model in which the vegetation fraction is set to zero. The SWE simulation in Fig. 9 shows that a significantly faster SWE loss occurs due to the absence of the vegetation fraction. This faster SWE loss gives rise to an earlier ending of the snow season (in May), which differs from the observations and simulations with the vegetation fraction included.

Mitchell et al. (2004) indicate that overrepresented snow surface sublimation in NoahLSM is another factor that results in the lower SWE based on the offline simulation results. However, in our control run, the latent heat flux is 59.3 W m$^{-2}$ (Table 1), which includes both sublimation and evaporation, and the snowmelt heat flux is 95.9 W m$^{-2}$. When these two fluxes are converted to mass unit, the snowmelt rate is 25 mm day$^{-1}$, but the combination of evaporation and sublimation is only 2 mm day$^{-1}$. The sublimation is significantly smaller than the snowmelt rate. Therefore, in our application, sublimation is not a dominant process that changes the SWE.

Therefore, realistic treatment of the land surface features is essential to SWE and snowmelt simulations. A sophisticated land surface model has been coupled into MM5 as part of our ongoing effort to improve the predictability of SWE and its impact on climate and water resources. These results will be addressed elsewhere.

7. Conclusions

This study evaluated snow water equivalent (SWE) simulated in MM5-NoahLSM to understand the impact of daily variations of SWE and snowmelt mechanisms
in the Sierra Nevada. Such an understanding will advance SWE and snowmelt simulations, providing a direct benefit to water resource management in this region. It was shown here that MM5-NoahLSM significantly underrepresents SWE in the Sierra Nevada region, which results in the warm bias in surface air temperature and stronger precipitation. The assimilation of the Snotel SWE was shown to have reduced the warm bias and stronger precipitation, and the modeled surface air temperature and precipitation are now significantly closer to observations.

First, the reduction of the warm bias in surface air temperature in the assimilation run results from the suppressed upward sensible heat flux caused by a decreased skin temperature. This skin temperature reduction is the result of the longer assimilated snow duration than in the control run (no SWE assimilation).

Second, the reduction of the overestimated precipitation is from the stronger stabilization of the lower atmosphere and the suppression of upward air motion caused by the SWE-induced lower near-surface temperature. Concurrent to these atmospheric processes, the lower surface evaporation in the assimilation run also has an impact on the precipitation reduction.

However, the modeled snowmelt rate with the SWE assimilation is still greater than observations. The resulting surface energy balance indicates that the lack of a vegetation fraction formulation in the MM5-NoahLSM results in extra solar radiation reaching the snow surface, which is the primary reason for the exaggerated snowmelt rate. The decreased surface albedo also contributes to the increased snowmelt. With realistic descriptions of vegetation fraction and surface albedo, an offline version of the SAST land surface model, forced by the assimilated MM5 output, produced SWE and snowmelt simulations that were closer to observations. These enhancements have further verified the importance of the treatment of the land surface features in MM5-NoahLSM. In addition, in MM5-NoahLSM, it is found that snow surface sublimation is a minor factor in affecting SWE during our simulation period, although its importance to SWE was indicated in Mitchell et al. (2004).

The analysis in this study shows that the assimilation method is an efficient way to understand how snow impacts regional weather and water resources, especially during the end of the snow accumulation season when water resource managers determine spring and summer allocations. The assimilation methods have been extensively used in weather, climate, and water resources modeling communities to generate high quality datasets and to understand weather, climate, and hydrological processes and phenomena. In the meantime, the assimilation methodology used here has helped identify the structure deficiencies in MM5-NoahLSM, and from which new model advancements can be developed. Essentially, the NoahLSM is structured in a point fashion at the surface, which is unable to account for the energy and water exchanges between the canopy and the snow–soil surface. In order for NoahLSM to improve snow simulation, this model needs to be changed from the point structure to one-dimensional structure.

In addition, the assimilation processes need to be
supported by observations, which are not always available at the needed spatial and/or temporal resolutions of study. To improve current numerical weather and climate forecasts, realistic model structures and physics must be incorporated.

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