Noah land surface model modifications to improve snowpack prediction in the Colorado Rocky Mountains

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Simulated snowpack by the Noah land surface model (LSM) shows an early depletion due to excessive sublimation and too early onset of snowmelt. To mitigate these deficiencies, five model modifications are tested to improve snowpack performance: (1) time-varying snow albedo, (2) solar radiation adjustment for terrain slope and orientation, (3) reducing the surface exchange coefficient for stable boundary layers, (4) increase of fresh snow albedo, and (5) adjusting surface roughness length when snow is present. The Noah LSM is executed from 1 November 2007 to 1 August 2008 for the headwater region in the Colorado Rocky Mountains with complex terrain, and its results are evaluated against 1 km Snow Data Assimilation System (SNODAS) output and individual Natural Resources Conservation Service Snowpack Telemetry (SNOTEL) sites. The most effective way to improve magnitude and timing of seasonal maximum snow water equivalent (SWE) is the introduction of the time-varying albedo formulation and the increase in fresh snow albedo. Minor improvement is obtained by reducing nighttime sublimation through adjusting the stable boundary layer surface exchange coefficient. Modifying the surface roughness length over snow surfaces and adding a terrain slope and orientation adjustment for radiation has little effect on average SWE conditions for the entire modeling domain, though it can have a significant effect in certain regions. The net effect of all changes is to improve the magnitude and timing of seasonal maximum SWE, but the snow period end is now somewhat too long. Adding the terrain slope and orientation effects does have an effect on local surface energy flux components depending on the cell slope and orientation.


1. Introduction

The snowpack evolution in the Rocky Mountains of the Western United States directly affects the hydrologic resources of the millions of people who rely on the Colorado River. According to Edwards and Redmond [2005], on average, 85% of the streamflow in the Colorado River comes from snowmelt in high mountain regions in the upper portion of the basin. However, the seasonal evolution of snowpack in the Colorado River basin (i.e., accumulation and ablation) exhibits significant interannual variability. Furthermore, several portions of the western U.S. are presently exhibiting significant decreases in peak snowpack accumulation, earlier dates of peak runoff, longer frost-free periods and lower summertime streamflow values, presumably in response to long-term climate change [e.g., Easterling, 2002; Milly et al., 2005; Hamlet et al., 2007; Mote, 2006; Barnett et al., 2008; Hidalgo et al., 2009]. Although intermodel differences are substantial, the consensus of climate models participating in the Fourth Assessment Report of the Intergovernmental Panel on Climate Change [2007] climate model projections suggest that much of the western U.S. will experience increased warm season aridity due to higher temperatures and evaporative demand [Hoerling et al., 2007; Seager et al., 2007; Hoerling and Eischeid, 2008]. Taking into account rapidly increasing demand for Colorado River water from fast-growing population centers in the southwest, improving predictive capabilities of snowpack processes in models is imperative for comprehensive water resource management.

To assess the impact of future climate change on snowpack in the headwater region of the Colorado Rockies, a high-resolution (2 km) regional climate model, based on the Weather Research and Forecast (WRF) model coupled to the Noah land surface model (LSM), was recently applied for multiyear snow season simulations [Ikeda et al., 2010]. While the WRF/Noah model simulated winter precipitation...
verified well against Natural Resources Conservation Service Snowpack Telemetry (SNOTEL) observations, the snowpack characteristics modeled by the Noah LSM exhibited a too early snow period end, and too low and too early seasonal maximum snow water equivalent. Past studies have found a negative bias in the Noah representation of total snow water equivalent (SWE) and snow cover extent, and in the timing of spring snow depletion. *Sheffield et al.* [2003] found that over the United States, Noah tended to underestimate snow cover extent by about 22% in 12 km simulations. For mountainous regions, the under prediction was as much as 65%. Although Noah usually under predicted throughout the season, the largest under prediction occurred in spring. Noah also tended to have a shorter annual snow period (later start and earlier end) than observations. *Lynch et al.* [2009] also showed that compared to satellite and ground-based observations Noah underestimates SWE and snow cover extent throughout the United States. This low bias in snow cover and SWE seems a common weakness in other land surface models as well. For instance, the four LSMs in NLDAS (Noah, MOSAIC, VIC, and SAC) were found to under predict seasonal maximum SWE during a 3 year period [Pan et al., 2003]. In their pan‐Arctic hydrology study, *Slater et al.* [2007] found that Noah had similar performance to other models (VIC, CHASM, CLM, and ECMWF) in simulated seasonal SWE mean square error. Even the best models in SWE bias had problems with snowpack timing; being too high in the early snow season and too low in melt phase. *Slater et al.* [2007] also found that the Noah prediction of the mean end day of seasonal snowpack had a slight high bias, but performed as well or better than the other LSMs in their study.

[4] The community Noah LSM has been widely used in weather and regional climate models [Chen et al., 1996; Chen and Dudhia, 2001; Ek et al., 2003; Leung et al., 2005, 2006; Ji et al., 2008] and is currently used in the prediction models of the National Centers for Environmental Prediction and in land data assimilation systems such as the North America Land Data Assimilation System [Mitchell et al., 2004] and the Land Information System [Peters-Lidard et al., 2007]. Therefore, the goal of this paper is to address the early snowmelt problem in Noah over the mountainous Colorado Rocky regions by incrementally adding modifications to the Noah land surface model. In general, models used for snowpack prediction vary substantially from complex model such as SNTherm [Jordan, 1991], which considers multiple snow layers, solar radiation penetration, liquid water retention and snow grain growth, to simple single‐layer models with minimal parameters [e.g., Essery and Etchevers, 2004]. Noah can be considered on the low‐complexity side of the snow model spectrum, and a community effort to include a multilayer snowpack model within Noah is underway (G. Y. Niu et al., The community Noah land surface model with multiphysics options, submitted to Journal of Geophysical Research, 2010). However, this study will only focus on improving the publicly available Noah V3.0, whose inherent limitations include its single snow layer, lack of snow water retention, and snowmelt based on residual energy from the surface energy balance. Noah does consider patchy fractional snow cover, snowpack heat flux, and spatially varying snow albedo. One important purpose of this study is to understand the extent to which a simple vegetation‐snow blended model can be improved to represent seasonal evolution of snowpack, because this type of snow model is commonly used in weather and regional climate models. We also seek to understand the physical processes and model parameters that significantly control partitioning of incoming energy at the snow surface through surface‐atmospheric exchange. Model results are compared to available observation‐based data assimilation output and ground‐based observations.

[5] This paper is organized as follows: section 2 describes the modeling setup and observations used for comparison; section 3 presents the results of the control simulation and identifies the deficiencies addressed in the study; section 4 describes each of the model modifications; section 5 compares the incremental model modification results to observations; and the conclusions are presented in section 6.

## 2. Model Setup and Observations

### 2.1. Noah Land Surface Model

[6] Surface processes are simulated using the Noah land surface model [Chen et al., 1996; Chen and Dudhia, 2001; Ek et al., 2003], which is based on a diurnally dependent Penman potential evaporation approach, a multilayer soil model, and a primitive canopy model. Noah has been extended to include a modestly complex canopy resistance and frozen ground physics. When snow is present, Noah considers a blended snow‐vegetation‐soil layer and simulates the snow accumulation, sublimation, melting, and heat exchange at snow‐atmosphere and snow‐soil interfaces, based on a simple snow parameterization of Koren et al. [1999]. Noah does not have independent snow layers, nor does Noah have a canopy snow interception component.

[7] Albedo for a grid containing snow is calculated using a linear weighting between the nonsnow background albedo on the snow‐free fraction and a blended snow‐vegetation albedo on the snow‐covered fraction. Snow‐covered fraction ($\bar{f}$) is determined as a function of SWE:

$$\bar{f} = 1 - \left[ \exp \left( \frac{-2.6}{SWE_{\max}} \right) - \frac{SWE}{SWE_{\max}} \exp \left( -2.6 \right) \right].$$

(1)

when SWE is below the vegetation type‐dependent $SWE_{\max}$ (40 cm for grass/crop/shrub and 80 cm for forests). When SWE is greater than $SWE_{\max}$, $\bar{f}$ equals 1. The fraction of the grid covered by snow is parameterized differently between land surface models [e.g., Sheffield et al., 2003] and can be a very important determinant of model performance. Different $\bar{f}$ formulations are not considered this study because equation (1) was found to alleviate near‐surface temperature biases in coupled versions of Noah [Ek et al., 2003].

[8] The Noah LSM has been implemented and tested extensively in both offline modes and land data assimilation systems [e.g., Chen et al., 1996; Chen and Mitchell, 1999; Mitchell et al., 2004; Wood et al., 1998; Boone et al., 2004; Chen, 2005], and coupled modes such as the MM5 and Weather Research and Forecasting (WRF) [Skamarock et al., 2008] model and several National Centers for Environmental Prediction (NCEP) weather and climate models [Chen et al., 1997; Chen and Dudhia, 2001; Yucel et al., 1998; Marshall et al., 2003; Ek et al., 2003]. Noah can be executed in offline or coupled mode and provides fluxes of energy and water, including snow sublimation and snowmelt, from the land surface, while also maintaining stores of water, including
For this study, Noah model simulations are conducted using the High Resolution Land Data Assimilation System (HRLDAS) [Chen et al., 2007]. The HRLDAS was developed to provide consistent land-surface input fields for WRF (or any mesoscale or global model) that are on the same model grid and have had time to develop their equilibrium climatology, a process that takes up to several years and thus cannot be reasonably handled within the much more computationally expensive WRF framework [Chen et al., 2007].

To run the HRLDAS, land surface properties and atmospheric forcing data are required. The HRLDAS is flexible enough to use a wide variety of satellite, radar, model, and in situ inputs and can also be used, as it is in this study, simply as a driver for Noah.

The simulation domain has also been used for coupled WRF/Noah regional climate simulation for the Colorado Headwater region [Ikeda et al., 2010], which is approximately 1200 km × 1000 km and centered on Colorado (shown in Figure 1). The central and western portions of the domain are dominated by mountainous terrain while the eastern portion slopes gently toward the east. All simulations are completed at a 2 km horizontal resolution on a grid with 601 × 501 cells. The model integrates forward in time using a 1 hour time step. Most atmospheric forcing data used to drive the HRLDAS are taken from a WRF simulation over the same domain (see Ikeda et al. [2010] for a description and validation of the WRF simulation). For this study, the HRLDAS is modified to use model level output directly from 2 km WRF simulations. Since the model forcing fields are taken from WRF before any modifications are done to the Noah model, there may be biases in the surface fields due to poorly modeled snowpack. Using the first model level (~20 m) instead of the 2 m diagnostic variables should alleviate the potential feedback. Model-generated output being used as forcing are the lowest model layer temperature, humidity, and winds along with pressure, downward longwave radiation, and precipitation at the surface. Precipitation partitioning is done using the WRF model temperature (≤ 0°C implies snow) and Noah model skin temperature (≤ 0°C implies surface freezing of liquid temperature).

All of the above variables are difficult to obtain over such a large area and at a high spatial resolution. However, it is possible to get accurate observations of solar radiation through GOES satellite measurements [Pinker and Laszlo, 1992]. Since the HRLDAS can readily use these observations when available, downward solar forcing is taken from the GOES surface downward flux historical database (available at http://www.atmos.umd.edu/~srh). These data are available hourly and at a horizontal resolution of 0.5°, which is then interpolated to the HRLDAS 2 km grid.

Model state variables, including snow water equivalent and snow depth, are initialized using the WRF initialization on 1 November 2008, except for the SNOTEL locations, where snow states are initialized with observed SNOTEL fields.

As a validation of the WRF precipitation, Figure 2 shows a scatterplot of total accumulated WRF precipitation versus accumulated SNOTEL precipitation for the 9 month simulation period. Although there is spread in the data, no obvious bias or dependence on terrain height is displayed except for a low model bias for SNOTEL locations receiving greater than 1000 mm of precipitation, which may be due to sub-2-km terrain effects [see Ikeda et al., 2010].

2.3. SNOTEL Observation Sites

SNOTEL sites are an extensive, automated system to collect snowpack and related climatic data in the Western United States. A typical site has a pressure sensing snow pillow, storage precipitation gage, and an air temperature sensor and these measurements have been found to compare favorably to snow course measurements [Serreze et al., 1999]. SNOTEL data have been used as validation data in previous Noah studies [e.g., Livneh et al., 2009; Pan et al., 2003]. Snow water equivalent (SWE) observations at 112 SNOTEL sites, mostly in Colorado, are used to verify the Noah model output (see Figure 1). SNOTEL precipitation is also used to justify the use of WRF model output and as forcing data to test model sensitivity. To compare model output to SNOTEL observations, the four nearest WRF grid points to the observation location are averaged using a distance weighting method. Results are found to be insensitive to different averaging techniques. A majority of the locations have average terrain differences compared to WRF that are less than 100 m and nearly 80% are less than 200 m. For the 448 WRF grid points used to compare with SNOTEL, land cover is dominated by evergreen needleleaf forest (41%), savannah (37%), and mixed forest (13%).
2.4. SNODAS Output

There are no large-scale observations of SWE. The National Operational Hydrologic Remote Sensing Center produces a gridded SWE product output from their Snow Data Assimilation System (SNODAS) [Barrett, 2003]. Although not an observational product, SNODAS does integrate a physically based, spatially distributed energy and mass balance model with observed snow data from satellite and airborne platforms, and ground stations. SNODAS output has high spatial (30 arcsec or ∼1 km) and temporal (1 h) resolutions over the United States. In this study, SNODAS output is interpolated to the HRLDAS grid using bilinear interpolation and is used to compare against domain-wide simulations.

3. Control Simulation

The control simulation consists of the HRLDAS as the driver for the standard release Noah model version 3.0. All simulations are done on the 2 km domain shown in Figure 1. The simulation period is 1 November 2007 to 1 August 2008, the same period for evaluating coupled WRF/Noah simulations, and is executed at a 1 hour time step. Simulation description labels consisting of two-letter codes indicate different model changes or forcing data. In the control simulation, GOES solar (GS) and WRF model level (ML) forcing are used. Therefore, the control simulation has the code name GSML. The two-letter codes for all simulations are summarized in Table 1.

Three metrics are used for model performance evaluation: seasonal maximum SWE, date of seasonal maximum SWE, and date of last significant SWE. Similar statistics have been used in other studies using Noah [Pan et al., 2003; Slater et al., 2007]. Using these three metrics, a majority of important characteristics for snowpack hydrology, including the shape and timing of the SWE curve, can be determined. Seasonal maximum SWE and its date of occurrence are simply found by searching the model SWE time series during the 9 month simulation. Date of last significant SWE is found by searching from the simulation end until a time when greater than 5 mm SWE exists at the grid point.

As discussed previously, the control simulation produces too little peak SWE, the peak is too early in the snow season and the snowmelts too fast in spring. These results are similar to those found by other studies using the Noah model [Livneh et al., 2009; Pan et al., 2003; Sheffield et al., 2003; Niu et al., submitted manuscript, 2010]. Tables 2–4 sum-

Table 1. Two-Letter Codes Used for Simulation Labels

<table>
<thead>
<tr>
<th>Code</th>
<th>Definition</th>
</tr>
</thead>
<tbody>
<tr>
<td>GS</td>
<td>use GOES solar forcing</td>
</tr>
<tr>
<td>ML</td>
<td>use WRF model level forcing</td>
</tr>
<tr>
<td>LV</td>
<td>Livneh et al. [2009] time-decaying snow albedo</td>
</tr>
<tr>
<td>TA</td>
<td>terrain adjustment for grid point slope and aspect</td>
</tr>
<tr>
<td>CH</td>
<td>use WRF stability adjustment</td>
</tr>
<tr>
<td>85</td>
<td>set maximum albedo to 0.85</td>
</tr>
<tr>
<td>ZE</td>
<td>use $z_e$ based on exposed vegetation</td>
</tr>
<tr>
<td>OP</td>
<td>use SNOTEL observed precipitation</td>
</tr>
</tbody>
</table>
marize the results of the full HRLDAS domain and the SNODAS output. Table 2 shows that the control simulation predicts maximum SWE 60% lower than SNODAS for low-elevation locations (less than 2000 m) and varies between 45% and 28% lower for elevations between 2000 and 3500 m. Table 3 shows that the control simulation peak SWE is about 20–30 days too early compared to SNODAS for grids points lower than 3500 m. For the same elevations, Table 4 shows the disappearance of the snowpack about 15–30 days too early. Although a relatively small fraction of the domain, very high elevation grid points above 3500 m compare reasonably well for all three metrics.

Table 2. Domain-Averaged Maximum SWE for Each HRLDAS Simulation and SNODAS Output

<table>
<thead>
<tr>
<th>Domain</th>
<th>0–1000 m</th>
<th>1000–1500 m</th>
<th>1500–2000 m</th>
<th>2000–2500 m</th>
<th>2500–3000 m</th>
<th>3000–3500 m</th>
<th>3500–4500 m</th>
</tr>
</thead>
<tbody>
<tr>
<td>Number of grid points</td>
<td>35918</td>
<td>79639</td>
<td>90941</td>
<td>63694</td>
<td>20463</td>
<td>8523</td>
<td>1923</td>
</tr>
<tr>
<td>Control</td>
<td>10</td>
<td>11</td>
<td>16</td>
<td>66</td>
<td>190</td>
<td>355</td>
<td>481</td>
</tr>
<tr>
<td>+ LV</td>
<td>10</td>
<td>12</td>
<td>17</td>
<td>73</td>
<td>213</td>
<td>394</td>
<td>517</td>
</tr>
<tr>
<td>+ TA</td>
<td>10</td>
<td>12</td>
<td>17</td>
<td>73</td>
<td>213</td>
<td>394</td>
<td>517</td>
</tr>
<tr>
<td>+ CH</td>
<td>11</td>
<td>14</td>
<td>23</td>
<td>87</td>
<td>225</td>
<td>407</td>
<td>538</td>
</tr>
<tr>
<td>+ 8S</td>
<td>12</td>
<td>16</td>
<td>30</td>
<td>111</td>
<td>280</td>
<td>478</td>
<td>613</td>
</tr>
<tr>
<td>+ ZE</td>
<td>12</td>
<td>16</td>
<td>29</td>
<td>106</td>
<td>274</td>
<td>470</td>
<td>604</td>
</tr>
<tr>
<td>SNODAS</td>
<td>22</td>
<td>26</td>
<td>45</td>
<td>119</td>
<td>306</td>
<td>491</td>
<td>498</td>
</tr>
<tr>
<td>( \Delta \text{precipitation} )</td>
<td>23</td>
<td>–76</td>
<td>–149</td>
<td>–116</td>
<td>–78</td>
<td>18</td>
<td>76</td>
</tr>
</tbody>
</table>

[a] SWE is measured in millimeters. Data are binned based on WRF terrain. Also included are differences between WRF precipitation forcing and SNODAS precipitation.

4. Model Modifications

To address some of the deficiencies in the control simulation, several modifications are applied to the Noah model. Each modification is described below. The changes are added sequentially meaning each modification is added to a model containing all the modifications listed above it. Also note that because it is not practical to show detailed results are added sequentially meaning each modification is added to a model containing all the modifications listed above it. Also note that because it is not practical to show detailed results from all SNOTEL sites, at times detailed results are presented from a representative site referred to as “KILN” that is located in central Colorado (see Figure 1).

4.1. Livneh et al. [2009] Time-Varying Albedo

One of the most important processes is the correct partitioning of incident radiation into radiative, sensible heat and latent heat fluxes. Net radiation drives the terrestrial surface energy budget. Surface albedo, the ratio of reflected to incident solar radiation, is an important surface characteristic that determines energy transfer at the surface. One of the fastest modifications to surface albedo is snow cover. Because the albedo of snow-covered land compared to snow-free land is so extreme, correctly predicting the presence of snow can greatly affect the near-surface energy balance. Snow cover can change the albedo of grassland by a factor of \( 3 \)–4 and forested regions by a factor of \( 2 \)–3 [Bettis and Ball, 1997; Thomas and Rowntree, 1992; Jin et al., 2002]. In modeling studies, these changes have been shown to have large effects on the surface energy budget and near-surface temperature [Thomas and Rowntree, 1992; Viterbo and Betts, 1999]. Modeling studies have also shown that snow-covered albedo changes can greatly affect the timing of spring snowmelt and subsequent streamflow peaks [Thomas and Rowntree, 1992]. Therefore, albedo, and more generally, the surface energy budget are the main areas where model improvement is sought.

The Noah model uses a linear blending of background vegetation/soil albedo and spatially dependent snow-covered albedo [Robinson and Kukla, 1985] based on snow cover fraction (see equation (1)). If snow cover does not change, albedo does not change. Livneh et al. [2009] incorporated an albedo decay scheme adapted from the DHSVM [Wigmosta et al., 1994] and VIC [Cherkauer et al., 2003] models, which are based on the observed albedo decay rates initially made by the U.S. Army Corps of Engineers [1956]. The scheme imposes an exponential decay in albedo \( (\alpha_{\text{snow}}) \) as the time between snow events increases:

\[
\alpha_{\text{snow}} = \alpha_{\text{max}}A^{Bt},
\]

where \( A = 0.94 \) (0.82) and \( B = 0.58 \) (0.46) during the accumulation (ablation) phase and \( t \) is the time in days since the last snowfall. The albedo decay occurs at different rates.
depending on snow surface temperature, decaying slower during the accumulation phase and faster during ablation. The new snow albedo ($\alpha_{max}$) is taken to be some adjustable fraction (currently 0.5) of the difference between the Robinson and Kukla value and ground observations of fresh snow (~0.85) [Wiscombe and Warren, 1980].

[23] Figure 3 shows the albedo for the month of November at the KILN SNOTEL site using both the control simulation albedo formulation and that of Livneh et al. [2009]. The Robinson and Kukla [1985] value for snow-covered albedo at this site is 0.52. Even with substantial snow cover, the control simulation albedo rarely gets above 0.5. The Livneh et al. formulation shows the decay of albedo with time and also the sharp increase in albedo when fresh snow falls. The downward spikes in the early part of the month occur when the snowpack is assumed to have liquid water at the surface.

4.2. Adjustment for Terrain Slope and Aspect

[24] The slope and orientation of the surface can make substantial differences to the amount of incident shortwave radiation through modification of the incident solar zenith angle [Whiteman, 2000]. Specifically, for the northern hemisphere, the directional orientation (aspect) of the land surface determines if incident radiation will increase (south facing), decrease (north facing), or have a diurnal adjustment (east and west facing). Figure 4 shows the grid point aspect values binned to the cardinal directions. There are nearly equal number of north and south facing grids, but due to the expansive, eastward sloping eastern portion of the domain, there are nearly twice as many east facing grids as west. In addition to terrain aspect, terrain slope also plays an important role in modifying the incident solar zenith angle. Figure 5 shows the terrain slope values calculated from the 2 km grid along with time of day and day of year values. This formulation has been used recently by Zängl [2004, 2009] and is based on the original formulation of Garnier and Ohmura [1968]. The methodology also partitions incident shortwave radiation into direct and diffuse components in which, for a given daylight time and location, the diffuse component is a constant fraction of the top of atmosphere radiation values while the direct component is altered based on the local incident solar zenith angle.

Table 4. Domain-Averaged Last Julian Date of 5 mm SWE for Each HRLDAS Simulation and SNODAS Output

<table>
<thead>
<tr>
<th>Domain</th>
<th>0–1000 m</th>
<th>1000–1500 m</th>
<th>1500–2000 m</th>
<th>2000–2500 m</th>
<th>2500–3000 m</th>
<th>3000–3500 m</th>
<th>3500–4500 m</th>
</tr>
</thead>
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<tr>
<td>Control</td>
<td>33</td>
<td>32</td>
<td>35</td>
<td>75</td>
<td>115</td>
<td>143</td>
<td>167</td>
</tr>
<tr>
<td>+ LV</td>
<td>33</td>
<td>32</td>
<td>35</td>
<td>75</td>
<td>117</td>
<td>148</td>
<td>171</td>
</tr>
<tr>
<td>+ TA</td>
<td>33</td>
<td>32</td>
<td>35</td>
<td>75</td>
<td>117</td>
<td>148</td>
<td>171</td>
</tr>
<tr>
<td>+ CH</td>
<td>33</td>
<td>34</td>
<td>42</td>
<td>79</td>
<td>119</td>
<td>148</td>
<td>171</td>
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<td>50</td>
<td>90</td>
<td>133</td>
<td>167</td>
<td>191</td>
</tr>
<tr>
<td>+ ZE</td>
<td>34</td>
<td>37</td>
<td>47</td>
<td>87</td>
<td>131</td>
<td>165</td>
<td>189</td>
</tr>
<tr>
<td>SNODAS</td>
<td>63</td>
<td>75</td>
<td>70</td>
<td>98</td>
<td>132</td>
<td>158</td>
<td>165</td>
</tr>
</tbody>
</table>

4.3. Stability Adjustment for Stable Boundary Layers

[25] New in WRF version 3.0 is an adjustment to the surface exchange coefficient when the atmospheric boundary layer is stable. The adjustment effectively decouples the surface from the atmosphere by increasing the Zilitinkevich coefficient, which then decreases the exchange coefficient [see Chen et al., 1997]. If a surface inversion exists, the Zilitinkevich coefficient is increased by a factor of three. An additional increase is applied as a quadratic function of terrain height (h) for elevations greater than 580 m ($9.0 \times 10^{-6}$ h^6). The effect of the adjustment can easily be seen in Figure 6, which shows the surface exchange coefficient for the first 2 days of November over the KILN SNOTEL location both with and without the stability adjustment. During the stable nighttime periods, surface exchange is decreased by more than 80%, substantially reducing snow sublimation and downward sensible heat flux from the atmosphere.

4.4. Sensitivity to Maximum Snow Albedo Parameter

[26] The Livneh et al. [2009] albedo scheme described in section 4.1 uses a factor of 0.5 to weight the Robinson and Kukla [1985] maximum snow albedo and the albedo for fresh snow (0.85), and resulted in too low albedo for new snow.

Table 5. Average Maximum SWE for Each HRLDAS Simulation at SNOTEL Locations and SNOTEL Sites

<table>
<thead>
<tr>
<th>Number of sites</th>
<th>2000–2500 m</th>
<th>2500–3000 m</th>
<th>3000–3500 m</th>
<th>3500–4500 m</th>
<th>Correlation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Control</td>
<td>3</td>
<td>47</td>
<td>59</td>
<td>3</td>
<td>0.82</td>
</tr>
<tr>
<td>+ LV</td>
<td>262</td>
<td>320</td>
<td>445</td>
<td>444</td>
<td>0.82</td>
</tr>
<tr>
<td>+ TA</td>
<td>287</td>
<td>368</td>
<td>494</td>
<td>482</td>
<td>0.82</td>
</tr>
<tr>
<td>+ CH</td>
<td>285</td>
<td>360</td>
<td>495</td>
<td>482</td>
<td>0.82</td>
</tr>
<tr>
<td>+ 85</td>
<td>284</td>
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</tr>
<tr>
<td>+ ZE</td>
<td>330</td>
<td>465</td>
<td>576</td>
<td>575</td>
<td>0.79</td>
</tr>
<tr>
<td>+ OP</td>
<td>328</td>
<td>462</td>
<td>573</td>
<td>572</td>
<td>0.80</td>
</tr>
<tr>
<td>SNOTEL</td>
<td>333</td>
<td>476</td>
<td>563</td>
<td>544</td>
<td>0.92</td>
</tr>
</tbody>
</table>

aSWE is measured in millimeters. Data are binned based on SNOTEL terrain.
cover. This modification eliminates the Robinson and Kukla albedo and sets all fresh snow to have an albedo ($\alpha_{\text{max}}$ in equation (2)) of 0.85. The albedo continues to decay in time from this maximum fresh snow value using equation (2). The red curve shown in Figure 3 already has this 0.85 modification. This modification works in conjunction with the Livneh et al. albedo modification. Setting albedo to an unvarying constant of 0.85 is inappropriate because of processes such as snow aging and snow falling off canopy.

4.5. Roughness Length Adjustment for Vegetation With Snow

Currently, the Noah model uses a background vegetation roughness length that is independent of the presence of snow. A modification is added so that only the exposed part of the vegetation is used in the roughness length calculation. To calculate the new roughness length, the canopy height is first approximated by multiplying the background roughness length by 7 [see Arya, 1988]. Next, model snow depth is subtracted from the canopy height to approximate the exposed vegetation. This exposed canopy height is then divided by 7 to get the new exposed roughness length. The exposed roughness length is applied only over the snow-covered fraction of the grid.

5. Results

Results are given in two categories: (1) the entire domain compared with SNODAS output and (2) model grid points associated with the SNOTEL observation sites, which are usually located at high elevations.

5.1. Domain-Wide Effects

The domain-wide effects of each modification are summarized in Tables 2–4 for maximum SWE, date of maximum SWE and date of last 5 mm SWE, respectively. For maximum SWE shown in Table 2, the individual modifications have little effect for grid points lower than 1500 m. This may be because most of these points are in the “marginal” or “transient” snow zones in the south and southeast portions of the domain and have much more dependence on seasonal timing of precipitation forcing. For higher-elevation locations, it is clear that the major improvements are made by employing the Livneh et al. [2009] albedo scheme (LV), setting fresh snow albedo to 0.85(85), and using stability adjustment (CH). The net effect of all modifications in the three terrain bins between 2000 and 3500 m is to reduce the under prediction compared to SNODAS from 45%, 38%, and 28% to 11%, 10%, and 4%, respectively.

The timing of maximum SWE in Table 3 shows a similar improvement for high-elevation grid points with little change in the lower elevations. Again, the Livneh et al. [2009] albedo scheme and fresh snow albedo change appear to have the greatest positive impact increasing the maximum SWE date by 2–9 and 10–15 days, respectively, for high elevations.

For timing of snow disappearance, Table 4 shows similar results to the timing of maximum SWE. For low-elevation grid points, the snow still disappears 30 days earlier than SNODAS, while for grid points higher than 3000 m the snow is not melting early enough.

The terrain adjustment has very little effect on any of the three domain-averaged metrics since the net orientation of

Table 6. Average Julian Date of Maximum SWE for Each HRLDAS Simulation at SNOTEL Locations and SNOTEL Sites.a

<table>
<thead>
<tr>
<th></th>
<th>2000–2500 m</th>
<th>2500–3000 m</th>
<th>3000–3500 m</th>
<th>3500–4500 m</th>
</tr>
</thead>
<tbody>
<tr>
<td>Control</td>
<td>64</td>
<td>61</td>
<td>78</td>
<td>96</td>
</tr>
<tr>
<td>+ LV</td>
<td>65</td>
<td>74</td>
<td>85</td>
<td>102</td>
</tr>
<tr>
<td>+ TA</td>
<td>65</td>
<td>75</td>
<td>85</td>
<td>102</td>
</tr>
<tr>
<td>+ CH</td>
<td>63</td>
<td>74</td>
<td>84</td>
<td>101</td>
</tr>
<tr>
<td>+ 85</td>
<td>69</td>
<td>88</td>
<td>93</td>
<td>112</td>
</tr>
<tr>
<td>+ ZE</td>
<td>69</td>
<td>87</td>
<td>92</td>
<td>110</td>
</tr>
<tr>
<td>+ OP</td>
<td>69</td>
<td>92</td>
<td>101</td>
<td>118</td>
</tr>
<tr>
<td>SNOTEL</td>
<td>107</td>
<td>110</td>
<td>101</td>
<td>112</td>
</tr>
</tbody>
</table>

aData are binned based on SNOTEL terrain.

Table 7. Average Last Julian Date of 5 mm SWE for Each HRLDAS Simulation at SNOTEL Locations and SNOTEL Sites.a

<table>
<thead>
<tr>
<th></th>
<th>2000–2500 m</th>
<th>2500–3000 m</th>
<th>3000–3500 m</th>
<th>3500–4500 m</th>
<th>Correlation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Control</td>
<td>119</td>
<td>133</td>
<td>150</td>
<td>167</td>
<td>0.56</td>
</tr>
<tr>
<td>+ LV</td>
<td>121</td>
<td>137</td>
<td>155</td>
<td>172</td>
<td>0.60</td>
</tr>
<tr>
<td>+ TA</td>
<td>121</td>
<td>137</td>
<td>155</td>
<td>172</td>
<td>0.60</td>
</tr>
<tr>
<td>+ CH</td>
<td>121</td>
<td>139</td>
<td>155</td>
<td>172</td>
<td>0.63</td>
</tr>
<tr>
<td>+ 85</td>
<td>135</td>
<td>156</td>
<td>175</td>
<td>191</td>
<td>0.70</td>
</tr>
<tr>
<td>+ ZE</td>
<td>134</td>
<td>156</td>
<td>174</td>
<td>190</td>
<td>0.70</td>
</tr>
<tr>
<td>+ OP</td>
<td>140</td>
<td>159</td>
<td>175</td>
<td>192</td>
<td>0.75</td>
</tr>
<tr>
<td>SNOTEL</td>
<td>140</td>
<td>147</td>
<td>162</td>
<td>140</td>
<td></td>
</tr>
</tbody>
</table>

aData are binned based on SNOTEL terrain.
the high sloping terrain is almost neutral. The roughness length adjustment has little effect or makes some of the metrics worse because decreasing the roughness length effectively reduces the surface fluxes. The extra energy retained at the surface then goes into melting snow; therefore even though sublimation is reduced, net SWE impact is negative. Since the stability adjustment predominantly functions during the night when there is no solar input, sublimation is reduced. Periods of excessive downward nighttime sensible heat flux are also reduced and the snowpack is increased.

[33] Figure 7 shows the simulation success in snow cover comparison to SNODAS for all points and for terrain bins using all model modifications (GSMLVTACH85ZE). The blue lines show the percentage of points when both HRLDAS and SNODAS predict no snow and the green lines show when both predict snow. The black lines indicate the fraction where HRLDAS predicts snow and SNODAS does not and the red lines show when SNODAS predicts snow and HRLDAS does not. The sum of the black and red lines being zero would mean perfect correlation between the models. The black line being equal to the red line means the snow-covered area is the same, but is misplaced. This is not unexpected considering HRLDAS and SNODAS have difference precipitation forcing.

[34] For lower elevations (<2000 m), HRLDAS tends to overpredict snow extent in the early part of the snow season, peaking in January (black line). For higher elevations (>2500 m) where most SNOTEL sites are located, HRLDAS starts with too much snow. From January to March all high-elevation locations are snow covered and near the end of the season HRLDAS keeps snow too long as was seen in Table 4. Since most of the time the black line is higher than the red line, this implies the HRLDAS snow is, in general, too extensive compared to the SNODAS snow. Except for the lower elevation, midwinter peak in the HRLDAS snow relative to SNODAS, agreement between the models is usually greater than 80%.

[35] See the auxiliary material for a similar snow cover comparison with the Moderate Resolution Imaging Spectro-
Figure 7

(a) All points (299572 pts)
(b) Pts < 1000 meters (35875 pts)
(c) Pts 1000 - 1500 meters (78365 pts)
(d) Pts 1500 - 2000 meters (90772 pts)
(e) Pts 2000 - 2500 meters (63660 pts)
(f) Pts > 2500 meters (30900 pts)
radiometer (MODIS) snow cover product. The results of the comparison with MODIS snow cover are very similar.

5.2. Comparison With SNOTEL Observations

The control simulation shows the same pattern of under prediction of the three metrics for the 112 SNOTEL sites as in domain-wide SNODAS comparison (Tables 5–7). An additional simulation is completed over 448 WRF-SNOTEL locations (the 4 closest WRF points to the 112 SNOTEL sites) using the SNOTEL observed precipitation (OP). For statistical robustness, only results for the 47 sites between 2500 and 3000 m and 59 sites between 3000 and 3500 m are discussed. For each of these statistics, average of the four closest WRF points are used to compare to the observation site. Using the five model changes, maximum SWE simulation shown in Table 5 improves for the lower-elevation sites from an under prediction of 41% to an under prediction of 15% using WRF precipitation forcing (GSMLLVTACH85ZE) and 13% using SNOTEL forcing (GSOPMLLVTACH85ZE). Likewise, the higher-elevation sites improve maximum SWE from an under prediction of 34% to under predictions of 15% and 17% depending on the precipitation forcing. Regardless of the model modification added, correlation between model results and SNOTEL observations remains at about 0.8 (see Table 5), but the use of SNOTEL precipitation increases the correlation to 0.92.

Figure 8 shows a scatterplot of seasonal maximum SWE for the control simulation and simulation with all five model modifications. Figure 8 shows that the modified model reduces the low bias in SWE. However, the spread in each simulation is very similar, therefore, not affecting the correlation as seen in Table 5. The largest disagreement is at locations with SNOTEL maximum SWE greater than 900 mm, where HRLDAS greatly underestimates maximum SWE. This is not surprising considering the precipitation scatterplot in Figure 2 shows precipitation under prediction when SNOTEL precipitation increases the correlation to 0.92.

The date of peak SWE improves by 26 and 14 days for low- and high-elevation sites, respectively (Table 6).

Figure 7. Time series of HRLDAS snow cover with all model modifications compared to SNODAS snow cover, where snow cover is defined as grid points with greater than 1 mm SWE. Blue lines show locations where both HRLDAS and SNODAS indicate no snow cover, green lines show locations where both models indicated snow cover, black lines show locations where HRLDAS indicates snow and SNODAS does not, and red lines show locations where SNODAS indicates snow and HRLDAS does not. (a) Data for the entire domain. Separation of model grid points based on terrain heights of (b) less than 1000 m, (c) 1000–1500 m, (d) 1500–2000 m, (e) 2000–2500 m, and (f) greater than 2500 m.

Figure 8. Seasonal maximum SWE (mm) scatterplot for HRLDAS compared to SNOTEL at all SNOTEL stations. Each HRLDAS point is an average of the four closest points to the SNOTEL station. The control simulation is in black, and the simulation with all model changes is in red.
high-elevation sites get the timing of peak correct, the lower-elevation sites are still early by 18 days. Timing of the last date of 5 mm SWE go from under predictions of 12 days to over predictions of about 12 days (Table 7). Although correlation improves incrementally with model changes, the major impact occurs with maximum albedo, which results in snow remaining for too long.

[39] The seasonal SWE time series for each of the model simulations averaged over all SNOTEL sites are shown in Figure 9. The control simulation clearly has too early and too low peak SWE and melts the snow too early in spring. The Livneh et al. [2009] albedo formulation shows incremental improvement in all three metrics (note in Figure 9 that the black dashed line overlies the blue dashed line). Terrain and stability adjustments (TA and CH) do not have a large overall net effect. The largest incremental improvement in peak SWE amount and timing occurs with the adjustment of maximum albedo (85), although this change keeps snow on the ground too late in the spring. Finally, the roughness length ($z_o$) formulation (ZE) has very little effect on average.

[40] Although in practice the 85 simulation should be coupled to the LV simulation (see section 4.4), a test simulation where the snow albedo is permanently set at 0.85 produces a snowpack that remains for about a month longer than observations (SNOTEL). However, the simulation does not produce a large difference in timing of maximum SWE (<1 day difference) and magnitude of maximum SWE (~20–30 mm difference).

[41] To analyze why the SWE curve changes between the control simulation and simulation with all modifications, snow sublimation and snowmelt are added to the SWE curve in Figure 10. Figure 10 clearly shows that the model changes suppress sublimation throughout the simulation while keeping the snowmelt curves similar through April and May until they diverge in late May. In the control simulation, about 45% of the snow sublimes, while about 15% sublimes in the fully modified simulation.

[42] Small differences in precipitation can make large differences in SWE especially in the late spring melt season. Figure 11 shows the SWE time series for the KILN SNOTEL site using the fully modified model with both WRF and SNOTEL precipitation forcing. Total accumulated precipitation for both WRF and SNOTEL are also shown. At this location, SNOTEL precipitation is higher by between 10 and 70 mm during the simulation. For most of the simulation, SWE differences are comparable to precipitation differences. However, during the spring melt period SWE differences can be as much as 170 mm immediately before the spring snowpack collapse. Also shown in this plot are the four points closest to the SNOTEL site used to calculate the average. Even at 2 km resolution there is a substantial amount of spatial variability, which must be considered when addressing any direct comparison between SNOTEL site observations and model grid point output. This spatial variability is also one of the reasons the results presented focus on the SNOTEL average and not individual SNOTEL locations.

5.3. Terrain Slope and Orientation Effect

[43] The adjustment for terrain slope and orientation has very little effect on the domain-averaged and site-averaged
Figure 10. Time series of SWE (solid lines), snow sublimation (long-dashed lines), and snowmelt (short-dashed lines) for the 9 month simulation starting 1 November 2007 averaged over all SNOTEL locations. Control simulation is in red, and simulation with all model changes is in blue. Black dots are the SNOTEL SWE observations.

Figure 11. Time series of SWE at KILN using all model changes with precipitation forcing from SNOTEL (thick blue line) and WRF (thick red line). Their difference is shown by the short-dashed blue line. SWE at the four WRF grid points used for the average is shown by thin red and blue lines. SNOTEL and WRF precipitation forcing and their difference are shown by green solid, long-dashed, and short-dashed lines, respectively. The black line shows the KILN SWE observations.
Adjusting the surface roughness length when snow is present has a stable boundary layer adjustment to surface exchange. Albedo. Minor improvement was obtained by introducing varying albedo formulation and increase the fresh snow albedo are included in Noah (V3.2). The lack of overlying canopy and liquid water retention in the snow matrix (V3.0) tested in this study is not complex. The addition of dynamic albedo and a new exchange coefficient formulation have been added in Noah (V3.1). The roughness length and slope exchange coefficient formulation have been added in Noah (V3.0) as a reference and have in some form been introduced into subsequent versions of Noah. Dynamic albedo and a new exchange coefficient formulation have been added in Noah (V3.1). The roughness length and slope exchange coefficient during stable regime will slightly decrease snow sublimation, but significantly increase snowfall. Therefore, the increase in snow resulting from these two changes is due to the decrease of available energy at the surface. The improvement from the CH simulation decreases downward sensible heat flux (which brings warmer air aloft to the snow surface) and hence reduced sublimation during stable boundary layer conditions.

Adding the terrain slope and orientation effects into Noah does not have a net effect on average simulated SWE. However, the addition does have an effect on individual grid cells. North and south facing grid will receive less and more solar radiation, respectively. East and west facing slope will change the diurnal distribution of incoming radiation. The changes are magnified with increasing slope and may be important in boundary layer development and convective precipitation triggering when coupled to an atmospheric model.

Note that small differences in precipitation input for Noah made quite substantial differences in the accumulation and ablation of snow, particularly in the late spring snowmelt season. There is also an interesting interplay of sublimation and melting. For instance, aggressively reducing the surface exchange coefficient during stable regime will slightly decrease snow sublimation, but significantly increase snowmelt, leading to an early snow disappearance. Although not shown, the runoff predicted by the model has a very similar timing and magnitude to the snowmelt curve shown in Figure 10. This means that for water resources prediction, the model now predicts a much larger spring/summer runoff magnitude with the timing unchanged at the beginning of the melt period but extend for longer into the summer.

The modifications suggested in this work use Noah (V3.0) as a reference and have in some form been introduced into subsequent versions of Noah. Dynamic albedo and a new exchange coefficient formulation have been added in Noah (V3.1). The roughness length and slope-aspect modifications along with a user-defined option to adjust maximum snow albedo are included in Noah (V3.2).

The treatment of snow in the current version of Noah (V3.0) tested in this study is not complex. The lack of overlying canopy and liquid water retention in the snow matrix make it difficult to reproduce the natural evolution of the snow cover is too early in the season with the control simulation, but becomes too late with the increase of maximum albedo.

Although albedo is allowed to decrease with time, the LV simulation still uses a higher albedo for snow than the control simulation, and using the maximum snow equal to 0.85 (labeled as the 85 simulation) further increases the fresh snow albedo. Therefore, the increase in snow resulting from these two changes is due to the decrease of available energy at the surface. The magnitude of the increase or decrease depends directly on the magnitude of the slope.

Table 8. Difference in Maximum SWE Between Simulation With Terrain Orientation Adjustment (GSMLLVTA) and Without Adjustment (GSMLLV) Binned by Slope of Terrain and Terrain Orientation*

<table>
<thead>
<tr>
<th>Slope Range</th>
<th>North</th>
<th>East</th>
<th>South</th>
<th>West</th>
</tr>
</thead>
<tbody>
<tr>
<td>0°–2°</td>
<td>+1</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>2°–4°</td>
<td>+3</td>
<td>0</td>
<td>-4</td>
<td>0</td>
</tr>
<tr>
<td>4°–6°</td>
<td>+7</td>
<td>+1</td>
<td>-7</td>
<td>0</td>
</tr>
<tr>
<td>6°–8°</td>
<td>+12</td>
<td>+1</td>
<td>-12</td>
<td>-1</td>
</tr>
<tr>
<td>8°–10°</td>
<td>+15</td>
<td>+2</td>
<td>-16</td>
<td>-1</td>
</tr>
<tr>
<td>10°+</td>
<td>+13</td>
<td>0</td>
<td>-19</td>
<td>-2</td>
</tr>
</tbody>
</table>

*SWE is measured in millimeters.

Figure 12. April mean diurnal incident shortwave radiation difference between simulations with and without the surface terrain adjustment (black) at a grid point near KILN with a slope of 6.7° and orientation of 254°E (west facing). Sensible and latent heat flux differences are shown by red and blue, respectively.

Even though the net terrain effect is small, Table 8 summarizes the effect of terrain slope and orientation on maximum SWE. East and west facing slopes have little effect regardless of slope magnitude. North and south facing slopes have complimentary effects with north facing slopes increasing maximum SWE and south facing slopes decreasing maximum SWE. The magnitude of the increase or decrease depends directly on the magnitude of the slope.

6. Conclusions

Modifications are made to the Noah land surface model to try to improve snowpack simulation in the Colorado Rocky Mountains. The most effective way to improve magnitude and timing seasonal maximum SWE was to introduce a time-varying albedo formulation and increase the fresh snow albedo. Minor improvement was obtained by introducing a stable boundary layer adjustment to surface exchange. Adjusting the surface roughness length when snow is present and adding a terrain slope and orientation adjustment had little effect on average SWE simulation. The end date of snow melting is summarized the effect of terrain slope and orientation on maximum SWE. The magnitude of the increase or decrease has complimentary effects with north facing slopes increasing maximum SWE and south facing slopes decreasing maximum SWE. East and west facing slopes have little effect with the increase of maximum albedo.
Snowpack. This study shows that improvement can be made relative to observations in the Colorado Rocky Mountains without drastically changing the existing Noah model. Comparison to observations in other environments where, for example, snowpack liquid water retention is more important (e.g., the Pacific Northwest) may show that more detailed process representations are necessary. To completely capture the physical representation of natural snow phenomena like canopy intercepted snow sublimation and springtime collapse of the snow, more complex processes will need to be incorporated.

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