Shortwave Radiative Fluxes on Slopes

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ABSTRACT

Snow-covered mountain ranges are a major source of water supply for runoff and groundwater recharge. Snowmelt supplies as much as 75% of the surface water in basins of the western United States. Net radiative fluxes make up about 80% of the energy balance over snow-covered surfaces. Because of the large extent of snow cover and the scarcity of ground observations, use of remotely sensed data is an attractive option for estimating radiative fluxes. Most of the available methods have been applied to low-spatial-resolution satellite observations that do not capture the spatial variability of snow cover, clouds, or aerosols, all of which need to be accounted for to achieve accurate estimates of surface radiative fluxes. The objective of this study is to use high-spatial-resolution observations that are available from the Moderate Resolution Imaging Spectroradiometer (MODIS) to derive surface shortwave (0.2–4.0 \( \mu \text{m} \)) downward radiative fluxes in complex terrain, with attention on the effect of topography (e.g., shadowing or limited sky view) on the amount of radiation received. The developed method has been applied to several typical melt seasons (January–July during 2003, 2004, 2005, and 2009) over the western part of the United States, and the available information was used to derive metrics on spatial and temporal variability of shortwave fluxes. Issues of scale in both the satellite and ground observations are also addressed to illuminate difficulties in the validation process of satellite-derived quantities. It is planned to apply the findings from this study to test improvements in estimation of snow water equivalent.

1. Introduction

a. Background

Snow-covered mountain ranges are a major source of water supply for runoff and groundwater recharge. Snowmelt supplies as much as 75% of the surface water in basins of the western United States (Beniston 2006). Factors that affect the rate of snowmelt include incoming shortwave (SW) and longwave radiation; surface albedo; snow emissivity and temperature; sensible, latent, and ground heat fluxes; and energy transferred to the snowpack from deposited snow or rain (Gray and Prowse 1992; Pomeroy et al. 2003). The net radiation generally makes up about 80% of the energy balance (Male and Granger 1981; Marks and Dozier 1992; Cline 1997). Therefore, the greatest potential sources of error in simulating snowmelt rates and timing are errors in radiation inputs.

Complex terrain poses a great challenge for obtaining needed information on radiative fluxes from satellites because of elevation issues, spatially variable cloud cover, rapidly changing surface conditions during snowfall, and melt. The situation in the region of interest here is unique because of the orographic effects of the mountains on cloud formation, affecting their distribution on the two sides of the mountains. Moreover, the Great Central Valley of California is frequented by the “tule” radiation fog that forms from late autumn through early spring, with the official period being from 1 November to 31 March. It occurs when the relative humidity is high during rapid night cooling and is usually confined to below 2000 ft (1 ft \( \approx 0.305 \text{ m} \)). The moisture for the fog formation is supplied by the ocean, which is also a source of nuclei for the condensation of the water
vapor. As will be shown in the results section, only high-resolution satellite observations can capture the unique and challenging complex-terrain radiation/cloud/fog conditions.

Observations from very-high-resolution satellites like those in the Landsat series (e.g., Rosenthal and Dozier 1996) are very useful for determining fractional snow cover; they are at low temporal resolution (about once in 16 days), and, as such, their applicability is restricted. Our approach utilizes routinely available observations from current satellites [e.g., the Moderate Resolution Imaging Spectroradiometer (MODIS)] and can serve as a prototype for use with future operational satellites, both polar orbiting and geostationary (further discussion is given in section 3).

b. Study objectives

The modeling and prediction of snowmelt in the western United States suffer from a lack of systematic and consistent information on energy-balance components. Most available information on surface radiative fluxes from satellites at relevant time scales is at spatial resolutions between 0.5° and 2.5° (Zhang et al. 2004; Hinkelman et al. 2009; Ma and Pinker 2012; Loeb et al. 2012). The objective of this study is to derive broadband SW radiative fluxes on slopes at the highest readily available resolution (5 km), using observations from MODIS (King et al. 1992; Platnick et al. 2003) to meet demands for such information [5 km is the highest resolution at which information on the auxiliary data on atmospheric and surface properties required for driving the methodology developed (or inference scheme) is also available from MODIS]. This information should facilitate improvements of snowpack modeling by realistically representing the spatial variability of the energy that drives the melting. We will address the intrinsic problems in evaluating radiative fluxes at 5-km resolution that are due to differences between the satellite and ground instrument view, we will address issues that are related to the use of routinely available ground observations (usually averaged over an hourly interval), and we will use an indirect approach to document that the higher-resolution fluxes are an improvement over those derived at lower resolution. An approach to derive radiative fluxes on slopes is also developed, and the quality of the slope-adjusted radiative fluxes is evaluated. The method for computing the SW fluxes is described in section 2, the resulting product and its evaluation are discussed in section 3, the improved representation of spatial and temporal variability of V2.0 (defined in section 2c) is described in section 4, the method used to distribute the SW fluxes on slopes is described in section 5, and a summary and discussion are given in section 6.

2. Approach

A brief description of the original model used to derive surface radiative fluxes from MODIS and improvements introduced for implementation at 5-km scale will be provided. Discussion of issues related to evaluation of high-spatial-resolution satellite products that are due to incompatibility between the scales of satellite and ground observations will follow. Since evaluation at mountain slopes adds additional complexity, data of the highest quality from conventional sites will initially be used as a benchmark for evaluation, to be followed by mountain sites and assuming that the instruments are mounted horizontally. A model will subsequently be developed to distribute the radiative fluxes on slopes and the evaluation will be repeated.

a. Model for shortwave fluxes, version 1.0

Observations from the MODIS instruments on the Terra and Aqua satellites (King et al. 2013) are used to produce surface radiative fluxes as needed for modeling various land surface processes in complex terrain, such as snowmelt, potential evapotranspiration (ET), or net primary productivity (NPP). The basic inference scheme implemented for the retrieval of SW fluxes is described in Wang and Pinker (2009) (version 1.0), where it was evaluated at 1° (~111 km) spatial resolution. Evaluation of version-1.0 products is also presented in Pinker et al. (2009), Niu et al. (2010), and Niu and Pinker (2011); evaluation of the radiative transfer component of the method is also presented in Randles et al. (2013). The feasibility of implementing the method at 5 km and application of the results to address scale issues over flat terrain are discussed in Su et al. (2008). In brief, SW radiative fluxes are computed in seven spectral intervals (0.2–0.4, 0.4–0.5, 0.5–0.6, 0.6–0.7, 0.7–1.19, 1.19–2.38, and 2.38–4.0 μm) while assuming a plane-parallel, vertically inhomogeneous, scattering and absorbing atmosphere. Water vapor absorption is parameterized following the methods of Ramaswamy and Freidenreich (1992) and Chou and Suarez (1999). Ozone absorption in the ultraviolet wavelengths and in the visible wavelengths is computed following the approach of Lacis and Hansen (1974). The single-scattering properties and vertical profiles of aerosols were derived from the Optical Properties of Aerosols and Clouds (OPAC) software package (Hess et al. 1998). Five atmospheric aerosol vertical profiles (continental, desert, maritime, Arctic, and Antarctic) are used with the inference scheme. Cloud extinction coefficients, single-scattering albedos, and asymmetry factors are computed from the parameterizations of Edwards and Slingo (1996) for water clouds and from Chou et al. (2002) for ice
clouds. Multiple scattering is dealt with by using the delta-Eddington approximation following the method of Joseph et al. (1976). Vertical atmospheric profiles are from the standard atmospheres of Kneizys et al. (1980). Top-of-atmosphere solar spectral irradiance data are from MODTRAN 3. The data sources for model input are described in Table 1.

b. Model for shortwave fluxes, version 2.0

To calculate the shortwave fluxes over the western United States at 5 km, the cloud fraction, cloud optical depth, and cloud drop size are first regridded to 5 km from 1-km data (at this resolution, all additional input parameters for the SW inference scheme are available). The 5-km cloud drop size is calculated as the mean of all pixel values within the $5 \times 5$ pixel block. The 5-km cloud optical depth $\tau_{5km}$ is calculated using logarithmic averaging:

$$\tau_{5km} = \exp[\ln(\tau_{1km})]. \quad (1)$$

For the MODIS level-2 SW flux products, the cloud-phase information (water or ice clouds) is included in the parameter “Quality_Assurance_1 km.” Aside from quality-control information, it also provides a flag to indicate the cloud phase and layer properties for each 1-km pixel. The 5-km cloud fraction and phase are aggregated from each $5 \times 5$ pixel block. Pixels flagged as “Undetermined” are treated as missing; pixels with layer information determined (single- or multilayer clouds) but with phase labeled as “Unknown” are classified as water clouds. Surface-albedo information is needed to account for multiple reflection of shortwave radiation between the surface and the atmosphere. In this study, the surface albedo is derived from the MODIS daily and weekly snow products (MOD10C1, MYD10C1, MOD10C2, and MYD10C2). The “Filled Land Surface Albedo” product is developed from the MODIS/Terra Land Surface Albedo Product (MOD43B3) by the MODIS atmosphere team. It gives spatially complete albedo maps for snow-free land surfaces at 1’ spatial resolution. To couple it with the 5-km radiative transfer calculations, the 1’ (of a degree) snow-free albedo data are resampled to 5-km grids. For deriving albedo of snow-covered areas, snow information at a daily time scale and at 0.05’ ($\approx$5 km) spatial resolution is used as available from both Terra (MOD10C1) and Aqua (MYD10C1). If both are available, an average is used. If no information is available at the daily time scale, the 8-day composite is used. In the original MODIS inference scheme (Wang and Pinker 2009), the spectral reflectance for snow was assumed to be 0.9 and 0.6 for the visible and near-infrared parts of the spectrum, respectively. In the updated version, the surface spectral reflectance in the presence of snow is derived from a combination of snow-cover percentage and the MODIS surface spectral reflectance products, which are provided as 5-yr (2000–04) climatological statistics (the underlying surface types are aggregated according to the International Geosphere–Biosphere Program classification; Moody et al. 2007).

c. Data used in evaluation

1) MODIS V2.0 SW ↓ FLUX PRODUCTS

The improved version (V2.0) of the SW inference scheme is used to produce instantaneous pixel-level (5 km) SW downward (SW ↓) fluxes from Terra and Aqua spectral observations. The product that is generated is provided swath by swath corresponding to the MODIS level-2 data; each pixel is identified by its...
coordinates and time of observation and covers an area of the western United States from the eastern foothills of the Rocky Mountains to the Pacific Ocean bounded by 35°–50°N and 100°–125°W from January through July of 2003, 2004, 2005, and 2009. A gridded instantaneous product at 0.05° resolution is also derived by averaging all pixels falling into a grid cell; the gridded product will be used in this study.

Instantaneous values from all Terra and Aqua observations are first normalized to the daily mean solar zenith angle and are then averaged to give an estimate of a daily flux. Because of the swath overlap at higher latitudes, for some regions there are as many as six observations available per day depending on the day of the year and location. In the following sections, the instantaneous pixel-level data within a 25-km radius of the station are averaged and the derived fluxes are evaluated against ground observations. The daily fluxes are used to investigate spatial and temporal variability of the SW↓ fluxes in the study area. In section 5, an approach to apply topographic correction is described; it is subsequently applied to the instantaneous gridded values for evaluation of improvements.

2) GROUND OBSERVATIONS

In the ideal situation, ground observations from mountain sites during the snow season should be used in evaluating satellite products that are intended for application in snowmelt modeling. Such observations are difficult to make because of limited access to the sites during the winter while it is snowing, and, therefore, the data are often not of the highest quality. In the evaluation process, we used observations from quality-controlled mountain sites (Fig. 1), as detailed in Table 2. To serve as a benchmark for evaluation of model performance under “ideal” maintenance conditions, we have augmented the mountain sites with observations from the “SURFRAD” stations (Augustine et al. 2005; http://www.esrl.noaa.gov/gmd/grad/surfrad/)

![Figure 1](http://journals.ametsoc.org/jamc/article-pdf/55/7/1513/3586661/jamc-d-15-0178_1.pdf)

**Fig. 1.** Study area with locations of various observational sites. The blue dots denote the mountain stations, and the magenta dots are the Oregon stations (as detailed in Table 2). The boxes indicate subdomains for further analysis.

<table>
<thead>
<tr>
<th>Site ID</th>
<th>Site name</th>
<th>State</th>
<th>Lat (°N)</th>
<th>Lon (°W)</th>
<th>Elev (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>USFmf</td>
<td>Flagstaff managed forest</td>
<td>AZ</td>
<td>35.143</td>
<td>111.727</td>
<td>2160</td>
</tr>
<tr>
<td>USFuf</td>
<td>Flagstaff unmanaged forest</td>
<td>AZ</td>
<td>35.089</td>
<td>111.762</td>
<td>2180</td>
</tr>
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<td>USFwf</td>
<td>Flagstaff wildfire</td>
<td>AZ</td>
<td>35.445</td>
<td>111.772</td>
<td>2270</td>
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<td>USMe2</td>
<td>Metolius intermediate pine</td>
<td>OR</td>
<td>44.452</td>
<td>121.557</td>
<td>1253</td>
</tr>
<tr>
<td>DAN</td>
<td>Dana Meadows</td>
<td>CA</td>
<td>37.897</td>
<td>119.257</td>
<td>2987</td>
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<td>TES</td>
<td>Tioga Entry Station</td>
<td>CA</td>
<td>37.910</td>
<td>119.260</td>
<td>3031</td>
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<td>TUM</td>
<td>Tuolumne Meadows</td>
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<td>2621</td>
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<tr>
<td>GLEES</td>
<td>Glacier Lakes Ecosystems Study Site</td>
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<td>41.364</td>
<td>106.239</td>
<td>3190</td>
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<td>USNR</td>
<td>Niwot Ridge</td>
<td>CO</td>
<td>40.033</td>
<td>105.546</td>
<td>3050</td>
</tr>
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<td>Reynolds Creek sheltered site</td>
<td>ID</td>
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<td>116.783</td>
<td>2100</td>
</tr>
<tr>
<td>RXP</td>
<td>Reynolds Creek ridge site</td>
<td>ID</td>
<td>43.186</td>
<td>116.783</td>
<td>2100</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Site ID</th>
<th>Site name</th>
<th>State</th>
<th>Lat (°N)</th>
<th>Lon (°W)</th>
<th>Elev (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>BON</td>
<td>Bondville</td>
<td>IL</td>
<td>40.125</td>
<td>88.367</td>
<td>213</td>
</tr>
<tr>
<td>FPK</td>
<td>Fort Peck</td>
<td>MT</td>
<td>48.317</td>
<td>105.100</td>
<td>634</td>
</tr>
<tr>
<td>TBL</td>
<td>Table Mountain</td>
<td>CO</td>
<td>40.130</td>
<td>105.240</td>
<td>1689</td>
</tr>
<tr>
<td>BURNS</td>
<td>Burns</td>
<td>OR</td>
<td>43.580</td>
<td>119.050</td>
<td>1295</td>
</tr>
<tr>
<td>EUINE</td>
<td>Eugene</td>
<td>OR</td>
<td>44.500</td>
<td>123.070</td>
<td>150</td>
</tr>
<tr>
<td>HERM</td>
<td>Hermiston</td>
<td>OR</td>
<td>45.820</td>
<td>119.280</td>
<td>180</td>
</tr>
</tbody>
</table>

Table 2. Locations and site identifiers (ID) for selected mountain observational stations used in this study. Note that RSH and RXP are two distinct sites that are located close together.
that are the continental U.S. contingent of the International Baseline Surface Radiation Network (BSRN; http://bsrn.awi.de/) (Ohmura et al. 1998), with an emphasis on stations with winter snow cover and located mostly in the western part of the United States. We used the Fort Peck, Montana (FTP; 48.1°N, 105.1°W); Bondville, Illinois (BON; 40°N, 88.4°W); and Table Mountain near Boulder, Colorado (TBL; 40.1°N, 105.2°W), sites, which are maintained according to the requirements of the BSRN. The BSRN sites provide data at 1- or 3-min frequency, whereas numerous sites at which information on radiative fluxes is of secondary interest tend to provide hourly averaged data (also the case with the mountain sites).

To facilitate stratification according to snow conditions (snow free and snow covered), information on snow conditions at each site is based on the Interactive Multisensor Snow and Ice Mapping System (IMS) daily Northern Hemisphere snow and ice cover data at 4-km resolution, as provided online (ftp://sidads.colorado.edu/pub/DATASETS/NOAA/G02156/4km). In addition, data from three high-quality West Coast stations [Burns, Eugene, and Hermiston, Oregon; Vignola et al. (2007)], as available online (http://solardat.uoregon.edu/SolarData.html), were used to strengthen the statistical significance of the evaluation results. The instantaneous pixel-level satellite products are used in these comparisons. First, we present results of evaluation at the optimal sites (SURFRAD/BSRN and the Oregon sites).

3. Evaluation of MODIS V2.0 SW↓ flux products
   a. Scale issues

   There is an inherent inconsistency of scale when matching satellite estimates at 5 km with ground observations. Ground instruments (radiometers) usually have a field of view (FOV) of ~170° looking upward, and, therefore, the actual sky area that contributes to the measured downward flux is generally much larger than 5 km. Observed fluxes integrate the entire FOV of the instrument and incorporate the distribution of aerosol and cloud conditions within the FOV. The 5-km footprint fills only part of this FOV. Under the assumptions used in the radiative transfer computations of the satellite algorithm, the calculated 5-km SW↓ flux is equivalent to the flux under a plane-parallel sky with the optical depth of the cloud and/or aerosol slab specified by the values retrieved for the 5-km pixel. Thus, the more representative the 5-km footprint is of the scale viewed by the radiometer, the closer is the agreement that can be expected between the satellite retrievals and the ground observations. One can expect better evaluation results under relatively uniform stratus clouds or clear-sky conditions than under broken cumulus clouds. Experiments to illustrate these issues are described in the next section.

   b. Optimal time and space scales for evaluation of satellite estimates of SW↓ fluxes

   Almost all evaluations of satellite retrievals against ground observations rely on the assumption that the temporal averaging at a fixed ground point is equivalent to the spatial average derived from the satellite observation. The validity of this assumption for atmospheric radiation fields is still under active investigation (Hakuba et al. 2013). Since the satellite estimates of SW↓ radiative fluxes are of relatively high spatial resolution and the ground measurements from the SURFRAD/BSRN and University of Oregon stations are available at 1-min temporal resolution, it is possible to investigate the impact of scale on the comparison results.

   Instantaneous MODIS-based SW↓ fluxes from the pixel-level products are compared with ground observations from SURFRAD sites averaged over a range of time intervals (10, 20, 30, and 60 min) and spatial scales (5-, 10-, 25-, and 50-km radii) (Table 3). The combination of 60-min averages of ground observations and 50-km-radius spatial averages of satellite estimates yielded the best results as measured by bias, standard deviation (STD), and correlation coefficient (CC) [a circle with a radius of 50 km (diameter 100 km) is about the size of a 1° grid box]. This combination will be used in the selected set of experiments at the SURFRAD and Oregon sites.

   c. Evaluation of SW↓ fluxes

   1) SURFRAD AND OREGON SITES

   Evaluation of MODIS broadband SW↓ fluxes at the FPK and BON SURFRAD sites was performed at
instantaneous and daily time scales using the pixel-level 5-km observations from each overpass (“swath”) averaged over a radius of 50 km and matched with ground observations averaged over an hour; snow and no-snow conditions as determined from the IMS data were evaluated independently (Fig. 2). For no-snow conditions, observations made during 2005 were used. Because the number of cases with snow conditions is lower than the number of cases with no snow, observations from 2004 were also added to the snow cases. The bias between the observations and the satellite-based SW\(\downarrow\) fluxes is larger for the snow cases. It is notoriously difficult to distinguish between snow and clouds when using satellite radiances (Wang and Key 2003; Curry et al. 1996), and radiometers during snowfall are vulnerable to snow on the domes and possible mounting problems even when the best possible maintenance procedures are followed. In networks that employ a high level of quality assurance, up-looking pyranometers (and pyrgeometers) are ventilated and thus are not vulnerable to snow on the domes; yet, snow can build up on the sunshield that surrounds the dome and introduce errors. Moreover, in the cloud-detection algorithms, snow can be interpreted as cloud, which would reduce the flux estimate from the satellite method (as also seen in Fig. 2). Therefore, it is difficult to determine the source of error in these comparisons. Evaluation of instantaneous and daily SW\(\downarrow\) for the three Oregon sites combined (Burns, Eugene, and Hermiston; http://solardat.uoregon.edu/SolarData.html) during 1 March–31 July 2005 is shown in Fig. 3 (no distinction was made between conditions with and without snow).

2) MOUNTAIN SITES AT 25-KM-RADIUS AVERAGING

Several evaluation experiments for the snow season of January–July for 2003, 2004, 2005, and 2009 have been performed at the mountain sites listed in Table 2. The following method was followed. Instantaneous 25-km-radius area-averaged MODIS observations for both Terra and Aqua are compared with ground observations at each mountain site individually to identify possible outliers at the ground station. The ground observations
at the mountain sites are hourly averages with the time
stamp corresponding to the end time of the observa-
tions. For example, an average from 0900 to 1000 would
be denoted as 1000. Since optimal matching of satellite
observations with ground truth is an issue, we have
tested two different approaches. In the first experiment
the MODIS observation times are rounded and the
satellite observation is matched with the ground average
ending at the rounded-off hour. For example, a MODIS
observation time of 0950 is rounded to 1000 and is
compared with the ground observation for 1000, which is
the average from 0900 to 1000. A MODIS observation
time of 0915 is rounded to 0900 and is compared with the
ground observation for 0900, which is the average from
0800 to 0900. This approach is referred to as method 1 in
Tables 4–7. In the second experiment, satellite obser-
vations were always matched with the average of the
interval in which they fell, regardless of closeness to the
beginning or end of the time-averaging interval. This
approach is referred to as method 2 in Tables 4–7. An-
alyses of the results for individual stations indicate that
the ground observations at both TUM and TES agree
poorly with MODIS estimates in 2003, 2004, and 2005
and that TES has poor agreement in 2009. Results for
each individual station stratified by year are presented in
Tables 4–7 for each experiment. For the first experi-
ment, points outside 3 STD were eliminated; the elimi-
nated points amounted to fewer than 1% of data points.
In the second experiment, data points outside 2 STD
were eliminated; the exact percentage of eliminated
data points is given in the tables. A combined analysis of
all stations used in the first experiment was also con-
ducted and is presented in Fig. 4. In this combined
analysis, the suspicious stations have been eliminated.
As is evident from these results, there is no clear evi-
dence that when more outlier points were eliminated in
the experiments the evaluation indicates better agree-
ment with the observations. Possibly this is due to the
way the matching of the satellite observations to the
ground truth has been done, which in the first case
consisted of using hourly averages that were closest to the
satellite overpass. Note that the MODIS overpass
times, particularly for Aqua, occur when the sun is high,
meaning that the fluxes are large and therefore the er-
rors can also be large. Yet, the relative values are fairly
small, as is evident from Table 8, which shows the

![Graph](http://example.com/graph.png)

**TABLE 4.** Individual evaluation for mountain stations for 2003, using methods 1 and 2 defined in section 3c(2). Here, “Elim” gives the exact percentage of eliminated data points.

<table>
<thead>
<tr>
<th>Station</th>
<th>RMSE</th>
<th>Bias</th>
<th>CC</th>
<th>N</th>
<th>Method 1</th>
<th>RMSE</th>
<th>Bias</th>
<th>CC</th>
<th>N</th>
<th>Elim (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>CA_DAN</td>
<td>159.4</td>
<td>60.7</td>
<td>0.85</td>
<td>509</td>
<td></td>
<td>166.3</td>
<td>69.3</td>
<td>0.79</td>
<td>478</td>
<td>6.5</td>
</tr>
<tr>
<td>CA_TES</td>
<td>190.4</td>
<td>84.8</td>
<td>0.84</td>
<td>518</td>
<td></td>
<td>199.5</td>
<td>97.0</td>
<td>0.81</td>
<td>489</td>
<td>5.0</td>
</tr>
<tr>
<td>CA_TUM</td>
<td>211.8</td>
<td>96.2</td>
<td>0.74</td>
<td>534</td>
<td></td>
<td>253.8</td>
<td>128.7</td>
<td>0.60</td>
<td>509</td>
<td>5.0</td>
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<tr>
<td>CO_USNR</td>
<td>175.8</td>
<td>38.0</td>
<td>0.83</td>
<td>375</td>
<td></td>
<td>184.3</td>
<td>46.5</td>
<td>0.81</td>
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<td>ID_RSH</td>
<td>112.4</td>
<td>12.1</td>
<td>0.93</td>
<td>499</td>
<td></td>
<td>110.0</td>
<td>5.8</td>
<td>0.92</td>
<td>489</td>
<td>1.4</td>
</tr>
<tr>
<td>ID_RXP</td>
<td>101.0</td>
<td>11.2</td>
<td>0.94</td>
<td>540</td>
<td></td>
<td>101.8</td>
<td>14.6</td>
<td>0.93</td>
<td>533</td>
<td>1.5</td>
</tr>
</tbody>
</table>

**FIG. 3.** Evaluation of MODIS swath SW1 fluxes during 1 Mar–31 Jul 2005 for (a) instantaneous and (b) daily fluxes over three sites in Oregon (Burns, Eugene, and Hermiston) combined (http://solardat.uoregon.edu/ SolarData.html). Ground observations were averaged over an hour for (a) and a day for (b).
statistics for all mountain stations combined (excluding the suspect sites) for each year for both evaluation methods. For bias, they range from 2.1% to 5.5%; for STD, they are from 19% to 22% of the mean. The typical evaluations of SW fluxes from satellites are done on a daily basis, which represents a daily average value (over 24 h).

d. Indirect evaluation of SW↓ products

The evaluations presented above (section 3c) address the need to match satellite and ground observations meaningfully. As explained, this required the use of satellite footprints averaged over a 25-km radius; as such, this does not provide a direct proof of whether there is a benefit in the higher-spatial-resolution information as compared with using a lower resolution. Additional insight on this issue can be gained from Fig. 5. The top-left panel in Fig. 5 shows the 5-km SW↓ product as obtained from MODIS/Terra level-2 observations. The right-hand panels in Fig. 5 show irradiances at 1° resolution computed using two different methods. The top-right panel in Fig. 5 shows results after the 5-km radiative flux data were aggregated into 1° grid boxes. To obtain the 1° data from 5-km MODIS input, the 5-km swath data were first gridded to 0.05° resolution and were then upscaled to a 1° grid by taking a 20 × 20 gridcell average. The bottom-right panel in Fig. 5 shows the SW↓ as derived from MODIS/Terra level-3 information that is provided at 1° resolution using the same algorithm as for the higher-resolution data. The bottom-left panel in Fig. 5 shows the difference between the two 1° products. As is evident, the differences are in the range from −15 to 25 W m⁻². In Fig. 6, the two types of 1° satellite retrievals from MODIS on the Terra and Aqua satellites were evaluated against ground observations from six of the mountain stations listed in Table 2. Hourly averaged data from these sites for the time period 1 December 2004–31 May 2005 are used. In Fig. 6, the left and right columns show results from Terra and Aqua, respectively. The top panel in Fig. 6 shows results for the 1° product upscaled from the 5-km data, and the bottom panel shows results for the 1° product inferred from the MODIS level-3 data. The preliminary results indicate that the agreement with ground measurements is better for the 1° product aggregated from the 5-km estimates for the MODIS/Terra product. For the MODIS/Aqua product, the aggregated 1° fluxes from the 5-km data yield a higher correlation and smaller STD but a somewhat larger bias than do those computed using 1° inputs from level 2. An additional evaluation for longer time periods needs to be undertaken.

4. Improved representation of spatial and temporal variability of MODIS V2.0

A better measure of the spatial and temporal variability of SW↓ and snow cover is of interest in formulating parameterizations of the surface energy budget and snowmelt. Snow information is also important so as to account correctly for multiple reflections between the surface and atmosphere in the radiative transfer computations that estimate surface SW↓ fluxes. Here, we

<table>
<thead>
<tr>
<th>Method 1</th>
<th>Method 2</th>
</tr>
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<tbody>
<tr>
<td>Station</td>
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<td>CA_DAN</td>
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<tr>
<td>CA_TES</td>
<td>201.9</td>
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<tr>
<td>CA_TUM</td>
<td>210.4</td>
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<td>CO_USNR</td>
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<td>ID_RSH</td>
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<td>ID_RXP</td>
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<table>
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<th>Method 2</th>
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</thead>
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<tr>
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<td>RMSE</td>
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<tr>
<td>WY_GLEES</td>
<td>153.4</td>
</tr>
</tbody>
</table>
use the 5-km radiative flux products developed in this study and available information on snow distribution to examine their spatial and temporal variability.

a. Spatial and temporal variability of snow cover

We focus on the magenta boxes that are shown in Fig. 7 and that were labeled as WA, ID, and CA in Fig. 1. They are defined as follows:

1) subarea CA (bounded by 121.4°W, 118.4°W, 36.5°N, and 39.5°N) is around the locations of sites DAN and TUM (see Table 2),
2) subarea ID (bounded by 117.3°W, 116.3°W, 42.5°N, and 43.5°N) is in Idaho around the location of site RME, and
3) subarea WA (bounded by 122.5°W, 120.5°W, 45.9°N, and 47.9°N) is in the Cascades from the Washington–Oregon border to just north of Snoqualmie Pass.

---

**Table 7.** As in Table 4, but for 2009.

<table>
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<th></th>
<th>Method 2</th>
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<td>CC</td>
<td>N</td>
<td>RMSE</td>
<td>Bias</td>
<td>CC</td>
<td>N</td>
<td>Elim (%)</td>
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<td>0.76</td>
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<td>CA_TES</td>
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<td>100.7</td>
<td>0.86</td>
<td>333</td>
<td>211.9</td>
<td>115.4</td>
<td>0.85</td>
<td>325</td>
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<tr>
<td>CO_USNR</td>
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<td>0.80</td>
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<td>175.7</td>
<td>26.7</td>
<td>0.81</td>
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<tr>
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<td>128.5</td>
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<tr>
<td>WY_GLEES</td>
<td>149.0</td>
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**Fig. 4.** Evaluation of MODIS swath instantaneous SW$_{\text{f}}$ for (a) 1 Jan–31 Jul 2003 at DAN, USNR, RSH, and RXP (from Table 4); (b) 1 Jan–31 Jul 2004 at DAN, USNR, RSH, and RXP (from Table 5); (c) 1 Jan–31 Jul 2005 at DAN, USNR, RSH, RXP, and GLEES (from Table 6); and (d) 1 Jan–31 Jul 2009 at the USFmf, USFu, USFwf, DAN, USNR, USMe2, and GLEES mountain sites (from Table 7).
In brief, the sizes of these subdomains are $3^\circ \times 3^\circ$ (CA), $2^\circ \times 2^\circ$ (WA), and $1^\circ \times 1^\circ$ (ID). Areas of different sizes were selected because of the distribution of measurement stations in the three regions and for climate diversity. The left panel of Fig. 7 shows the monthly mean snow cover for January 2005 in units of percent combined from the MODIS instantaneous snow products from *Terra* and *Aqua*, and the right panel shows the day-to-day variability in snow-covered area (%). MODIS snow products give daily snow fraction in percent; monthly mean snow cover is computed by taking a time average over the days in the month. Missing daily snow-cover values are filled with MODIS 8-day snow products. The day-to-day variability is calculated as the STD of daily values for January 2005 at each pixel. Variability is expressed as the STD of the snow coverage. Because the snow coverage is expressed in percentage, the variability (STD) is also in units of percentage. Even at the

![Image](https://example.com/image.png)

**Fig. 5.** An example of instantaneous SW$_\downarrow$ fluxes at different grid resolutions in the subarea CA ($121.4^\circ$–$118.4^\circ$W, $36.5^\circ$–$39.5^\circ$N) for 1 Jan 2005 from the MODIS/Terra satellite: (top left) 0.05° results from the MODIS level-2 5-km-resolution swath input, (top right) 1° results aggregated from the 0.05° results, (bottom right) 1° results from MODIS level-3 1° data, and (bottom left) the difference between the two 1° results. The red box indicates the location of the subarea CA. The white lines are the coastal and state boundaries. To obtain the 1° data from 5-km MODIS input, the 5-km swath data were first gridded to 0.05° resolution and then upscaled to a 1° grid by taking a $20 \times 20$ grid average.

<table>
<thead>
<tr>
<th>Year</th>
<th>CC</th>
<th>STD (20%)</th>
<th>Bias (3.5%)</th>
<th>No. of points</th>
<th>CC</th>
<th>STD (20%)</th>
<th>Bias (3.5%)</th>
<th>No. of points</th>
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<td>23.5</td>
<td>1923</td>
<td>0.87</td>
<td>140.4</td>
<td>25.8</td>
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<td>0.86</td>
<td>145.1</td>
<td>37.0</td>
<td>1679</td>
<td>0.83</td>
<td>147.4</td>
<td>41.3</td>
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<td>2005</td>
<td>0.86</td>
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<td>2009</td>
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<td>2318</td>
<td>0.82</td>
<td>141.5</td>
<td>17.7</td>
<td>2142</td>
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| Method 1 | | | | | Method 2 | | | |
| No. of points | | | | | No. of points | | | |

Table 8. Summary of statistics related to Fig. 4.
5-km resolution, the spatial and temporal snow variabilities are significant. Therefore, the ability for snow models to utilize the variability in snow-cover distribution as well as in radiative fluxes (as represented by the 5-km MODIS observations) should improve estimation of snowmelt.

b. Spatial and temporal variability of SW fluxes

The variability of radiative fluxes in space and time is of importance for snow water equivalent (SWE) modeling. We investigate such variability in the three selected subdomains mentioned above. In Fig. 8, the left column shows the monthly mean values of $SW_{\downarrow}$ radiative fluxes for January 2005. The center column shows the frequency distribution of daily values for the corresponding month and subdomain. The right column shows the temporal STD for the subdomain for the month. The spatial variability of $SW_{\downarrow}$ in the three subdomains for January 2005 is presented in Fig. 9. The left column shows the monthly mean spatial variability. The spatial variability (as represented by the STD) is calculated within a $9 \times 9$ box around each grid point at a daily time scale, yielding the temporal average of the daily scale variability for the month. The center column shows the frequency distribution of the spatial variability for the whole month of January 2005. The right column shows the spatial variability of the monthly mean flux over an area of $9 \times 9$ grid cells. Each row is for a specified subdomain.

The representativeness of a location to an extended surrounding area depends on the spatial heterogeneity of the area. As seen in Fig. 9, where time-average spatial variability for three subareas (CA, ID, and WA) is shown (left column), there are locations where $SW_{\downarrow}$ radiation changes rapidly over a short distance, a feature that would not be captured at a lower resolution. For example, in the CA subarea, along the edges of the California basins, large flux gradients can be observed.
This corresponds to the transition from the tule-fog-covered California basin during winter season to the high-elevation mountain area. One would expect an observation located within this region to be less representative of its surroundings.

Spatial variability can be systematic or nonsystematic. The topographic pattern shown in the CA subarea is an example of systematic spatial variability. The south–north change of solar radiation caused by the sun–Earth geometry is another example of systematic spatial variability. Nonsystematic variability is mainly caused by processes such as cloud formation or aerosol pattern movement. Figure 10 shows the time series of mean fluxes and the mean spatial STD for the three subareas relative to the area mean. Calculations are based on daily mean values. The mean spatial variability changes with time. The relative variability in the ID subarea is less than in the other two subareas. The larger variability in CA than in ID is mainly due to the terrain of the region. A larger domain size, \( 3^\circ \times 3^\circ \) for CA versus \( 1^\circ \times 1^\circ \) for ID, may also contribute to the larger mean spatial variability.

As is evident, the mean values start to increase around February and somewhat earlier in the lower latitudes.

5. Topographic corrections

a. Model development

In mountainous regions, local topographic effects that are due to complex terrain may cause significant variation in the surface radiation budget. For \( SW \downarrow \), the variation is mainly induced by the changes of illumination angle, shadowing, and a limited sky-view factor along with reflection from surrounding terrain. The slope and aspect of the receiving surface determine the flux density of the incident radiation. Shadowing blocks the solar beam from reaching the surface and is especially important under low-sun conditions or in valleys. Sky-view restrictions limit the amount of diffuse radiation that reaches the surface. A common approach to computing the solar irradiance in mountainous terrain is to follow a two-step process. First, downwelling irradiance is computed for a flat lower boundary. Second, corrections are applied to the inferred \( SW \downarrow \) to derive the radiant flux received by the tilted surface within its surrounding topography. The resolution of the gridded terrain is usually higher than the grid size used in our radiative transfer calculation.

Treatment of topographic effects for retrieving \( SW \downarrow \) radiation fluxes from satellite observations has been investigated in several studies (Dozier and Frew 1990; Dubayah et al. 1990; Müller and Scherer 2005; Chen et al. 2006; Helbig et al. 2009; Lai et al. 2010; Lee et al. 2011). The approach for topographic correction applied in many studies follows the formulation

\[
F(x) = \chi F_B \cos(\gamma_0)\cos(\gamma_{0h}) + F_D f_{sky}(x) + \overline{\alpha}(F_D + F_B)[1 - f_{sky}(x)],
\]

(2)

Fig. 7. (left) Monthly mean snow-covered area (%) averaged from the MODIS daily snow products, and (right) the day-to-day variability in snow-cover area. The calculation is based on the 0.05\degree-resolution gridded data for January 2005. The topography of the region is shown in Fig. 1. The three focus sites (WA, ID, and CA) are marked by the boxes.
where $F(x)$ is the downwelling shortwave flux after the topographic correction. The first term on the right-hand side of Eq. (2) represents the solar beam radiation incident on a tilted plane. The second term is the downwelling diffuse radiation restricted by a sky-view factor (calculated as the percentage of the upper hemisphere that is not obstructed by surrounding terrain). The last term denotes the reflected radiation from the surrounding terrain. This last term is not present when a horizontal lower boundary is assumed. The $F_B$ and $F_D$ terms are the direct and diffuse components of the SW flux, respectively, without topographic correction. In addition, term $x$ is an indicator of blocking of the solar beam given as a binary number, $\gamma_0$ is the solar incidence angle measured from the tilted surface normal to the solar direction, $\gamma_{0h}$ is the solar incident angle to a horizontal receiving plane, $f_{\text{sky}}(x)$ is sky-view factor (the fraction of the upper hemisphere that is not blocked by the surrounding terrain), and $\bar{\sigma}$ is the surface albedo averaged over the grid box.

In our treatment of Eq. (2), the definition of the sky-view factor [Eq. (4), below] is the same as in Dozier and Frew (1990), but the terrain configuration factor $[1 - f_{\text{sky}}(x)]$ is different. Dozier and Frew (1990) approximate the terrain configuration factor as the difference between two sky-view factors, where one is the sky-view factor for an infinitely long slope and the other is the sky-view factor obtained by Eq. (4) below. The Dozier and Frew (1990) definition ignores the radiation coming from the terrain below the horizon.
As is, Eq. (2) is highly simplified, with the following approximations:

1) The diffuse sky radiation is isotropic.
2) The heterogeneity of the surface albedo and SW $Y$ in each individual grid cell is ignored at the satellite resolution.
3) The impact of complex terrain on multiple reflections between the surface and atmosphere is ignored.
4) Atmospheric scattering and absorption between slopes are ignored.
5) The surface is a Lambertian reflector.
6) Because of terrain reflection, there will be radiation exchange between neighboring grid cells. This is called the edge effect. The simplified approach assumes that the size of a grid cell is large enough that the edge effect can be neglected.

Given the above assumptions, the topographical correction to SW $Y$ radiation is basically a geometric problem. For a receiving plane with tilt angle $\theta_r$ and azimuth angle $\phi_r$, the cosine of the solar zenith angle, which is the angle between the sunbeam and the receiving plane normal, is given as

$$\cos(\gamma_0) = \cos(\theta_0) \cos(\theta_r) + \sin(\theta_0) \sin(\theta_r) \cos(\phi_r - \phi_0).$$

(3)
If \(\cos(\gamma_0) < 0\), the receiving point is self-shaded.

The sky-view factor by definition (Dozier and Frew 1990) is

\[
f_{sk} = \frac{1}{\pi} \int_{\Omega} \cos(\gamma) \, d\Omega = \frac{1}{\pi} \int_{0}^{2\pi} \int_{0}^{h_\phi} \sin(\theta) \cos(\gamma) \, d\phi \, d\theta.
\]

(4)

Dozier and Frew (1990) expressed Eq. (4) as

\[
f_{sk} = \frac{1}{\pi} \int_{0}^{2\pi} \int_{0}^{h_\phi} \sin(\theta)[\cos(\theta) \cos(\theta_j) \sin(\phi - \phi_j)] \, d\phi \, d\theta
\]

\[
= \frac{1}{2\pi} \int_{0}^{2\pi} d\phi \left\{ \cos(\theta_r) \sin^2(h_\phi) + \sin(\theta_j) \cos(\phi - \phi_j)[h_\phi - \sin(h_\phi) \cos(h_\phi)] \right\}
\]

\[
\approx \frac{1}{2\pi} \sum_{i=1}^{n} \{ \cos(\theta_r) \sin^2(h_\phi) + \sin(\theta_j) \cos(\phi - \phi_j)[h_\phi - \sin(h_\phi) \cos(h_\phi)] \} \Delta \phi_i.
\]

(6)

Equation (6) must be evaluated numerically.

A digital elevation model (DEM) is needed to determine the shaded-or-unshaded indicator function \(\chi\), the solar incidence angle \(\gamma_0\), and the sky-view factor \(f_{sk}\).

The DEM used in this study is the "HYDRO1K" database developed by the National Center for Earth Resources Observation and Science (EROS) of the U.S. Geological Survey (USGS; http://webgis.wr.usgs.gov/globalgis/metadata_qr/metadata/hydro1k.htm; https://lta.cr.usgs.gov/HYDRO1K). It is based on the USGS global 30 arc s elevation dataset (GTOPO30). The HYDRO1K, on a continent-by-continent basis and at a resolution of 1 km, provides hydrologically corrected DEMs along with a suite of ancillary datasets including slope angle and azimuth angle.

To numerically calculate \(\chi, \gamma_0\), and \(f_{sk}\), we first obtain the local horizons for each 1-km terrain cell. The local horizons are determined within a rectangular box of 40 km \times 40 km size centered at each cell. Around the center of each cell, the 360° azimuth is divided into 64 sections. For each section, a maximum elevation angle is determined as the representative elevation angle of the section from all 1-km cells located within the section. For a central cell \(i\), the elevation angle between cell \(i\) and another cell \(j\) in the section is calculated as

\[
\tan(\beta_{ij}) = \frac{z_j - z_i}{|x_j - x_i|},
\]

(7)

where \(z_j\) and \(z_i\) denote the altitudes of the two cells, \(|x_j - x_i|\) is the distance between the two cells, and \(\beta_{ij}\) is the elevation angle from cell \(i\) to cell \(j\) given as \(\beta_{ij} = (\pi/2) - h_{\phi,ij}\).

The elevation angle is measured from the horizon as shown in Fig. 11; if cell \(j\) is lower than cell \(i\), \(z_j - z_i\) is set to zero. If a receiving cell is not tilted, obstruction of the sky is solely caused by its surrounding cells (\(j \neq i\)). For a tilted receiving cell, part of the sky view from the center of the receiving surface may be blocked because of its tilt relative to the mountain slope. This effect is also accounted for in the calculation.

The sky-view factor \(f_{sk}\) is calculated using Eq. (4) with \(n = 64\). For the direct solar beam, the solar azimuth direction is first determined. Then, the solar zenith angle is compared with the representative horizon angle of the corresponding direction. If the sun is lower than the horizon angle, the shaded or unshaded indicator function \(\chi\) is set to 0; otherwise, it is set to 1. Self-shadowing is determined by the cosine of the solar incidence angle, as shown in Eq. (3).

With all radiant and geometric parameters being available, the downwelling radiation flux onto a tilted receiving plane in mountainous terrain can be estimated by Eq. (2). For our study, \(F_B\) and \(F_D\) are obtained directly from the radiative transfer calculation without the need to estimate the two components from the global flux, as is typically done when only the total solar flux is available from parameterizations or measurements. Note that 5 km is the resolution at which the radiative fluxes are computed assuming average surface elevation.

The actual resolution of the radiative fluxes after the topographic correction depends on the resolution of the DEM used. Dozier and Frew (1990) use 100-m-resolution DEMs; the topographic-correction method described here can be applied to any resolution DEM independent of the resolution of the radiation fluxes.

The extent to which the radiation coming from below the horizon can be ignored depends on the surface type of the slope that reflects the radiation. If the below-horizon surface happens to be snow covered with high albedo,
ignoring the radiation from that part of the terrain may cause large differences in the computed fluxes. Because for most of the Earth surfaces the portion of received radiation that comes from the surrounding terrain is less than 3% (Chen et al. 2006), and even less for domain-average values, large differences should not be expected.

b. Impact of the topographic correction

Figure 12 shows an example of the impact of topography on SW$_\downarrow$ in subarea CA for 2025 UTC 14 January 2005. These images are at 1-km resolution; that is, the 5-km radiative fluxes were downscaled according to the DEM used (1-km resolution). The two stages of correction are considered separately. The top-left panel of Fig. 12 shows the original flux before the topographic correction. The top-center panel shows the result after topographic correction if the instrument is mounted horizontally. This is the typical situation for ground instruments that are set up to measure SW$_\downarrow$. In this situation, the topographic effects are mainly the shadowing from neighboring mountains and the sky-view restriction effect. The top-right panel (afternoon situation) shows the SW$_\downarrow$ received by a tilted slope, corrected for the topographic effects. Differences between the corrected and uncorrected fluxes for the horizontal receiving surface, as shown in the bottom-center panel, are less than 10 W m$^{-2}$ for most of the mountain region, although they range from $-15$ to $+20$ W m$^{-2}$. The bottom-right panel shows the differences after and before corrections for the mountain slope. In this situation, in which some sloped surfaces face the sun directly and receive incoming solar radiation with increased flux density, the difference can be as large as 400 W m$^{-2}$.

6. Summary

An inference scheme for deriving SW$_\downarrow$ fluxes from MODIS spectral observations was modified to allow implementation with MODIS 5-km observations, with special attention to snow conditions on the ground and the effect of mountain slopes on the radiation received. This required the development of procedures to incorporate highly variable surface conditions into the model, replacement of missing information in the MODIS database with independent observations (i.e., aerosols from the Multispectral Imaging Spectroradiometer (MISR) or water vapor from the NCEP–NCAR reanalysis) and the development of an approach to distribute the radiative fluxes on slopes. Challenging also was the evaluation process in itself because comparison between surface observations and satellite footprints is complicated, primarily because of the different fields of view of the instruments used (Long and Ackerman 1995; Hakuba et al. 2013). Within such limitations, the inferred fluxes were evaluated against well-maintained and calibrated ground stations that follow the rigorous requirements of the BSRN protocols as well as against dedicated observations in the mountainous region of interest, where SWE is observed. The method was implemented for several snow seasons, providing information that is the first of its kind. The newly available information is of sufficient duration to be useful for sensitivity testing of SWE estimations to said information.
The evaluation experiments identified difficulties in optimally matching ground stations with high-resolution satellite observations. As such, the results obtained do not provide absolute metrics on the quality of the satellite-based estimates. These experiments also indicate that there is a need for improvement in the design of data collection at mountain sites, possibly at higher temporal resolution, as well as on inclined surfaces. An indirect evaluation of the quality of the 5-km radiative flux estimates was also attempted. This was possible because relevant MODIS data for computing surface radiative fluxes are available at two resolutions: 1) gridded to 1° and 2) at a 5-km pixel level. Therefore, it is possible to use directly the 1° product to derive surface fluxes (e.g.,

![FIG. 11. Topographic correction parameters. The red line represents a terrain configuration. The blue lines represent the sky view. The factor \(1 - f_{sky}\) allows for contributions from the terrain part of the field of view, including the part below the horizon.]

![FIG. 12. Impact of surface topography on solar radiation for subdomain CA at 2025 UTC 14 Jan 2005: (top left) downwelling solar radiation before topographic correction, (top center) SW downwelling flux after topographic correction over a horizontal receiving surface, (top right) SW downwelling flux incident on the slope face, (bottom left) surface elevation of the domain, (bottom center) difference between after and before (after - before) topographic correction for horizontal receiving surface, and (bottom right) difference for a slope face.]

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Wang and Pinker 2009) or to use the 5-km product, produce the fluxes at that level, and then aggregate them to 1° resolution. These two resulting 1° products were then compared with ground observations at the mountain sites. As is evident from Fig. 6, the product that is based on the 5-km input data is in better agreement with the surface measurements and is an indirect indicator about the quality of the 5-km product. The merit of the high-resolution surface fluxes is also due to their ability to depict small-scale features of high spatial variability that impact many physical processes, such as snowmelt, evapotranspiration, and net primary productivity. As seen from Fig. 8, the monthly mean fluxes over the CA subdomain clearly show the locations of Lake Tahoe and Mammoth Lake, which, on average, receive less radiation than the surrounding area but have large spatial variability. Lake Tahoe in the Sierra Nevada is at a surface elevation of 6225 ft and is the largest alpine lake in North America. Mammoth Lake is also in the CA subdomain at an elevation of 7880 ft. It also clearly shows the reduced radiation over the fog-covered areas and the effect of clouds on the slopes. Such realistic depictions of these features, which are typical of complex terrain, are not seen in the 1° product shown in Fig. 5. The spatial variability of SW↓ fluxes is shown in Fig. 9 for January of 2005. The left column shows the monthly mean spatial variability calculated within a 9 × 9 box around each grid point at daily time scale, namely, the temporal average of the daily scale variability for the month. Again, Lake Tahoe and Mammoth Lake are distinctly seen, showing large variability in space and time corresponding to the variability in cloud formation over these lakes.

While MODIS observations are of relatively high spatial variability, the temporal frequency of the observations is limited. In mountainous regions affected by topography and temporal shading effects, additional attention needs to be given to such features in the process of evaluation on daily time scales. In particular, how well the diurnal cycle is represented when only two observations per day are available must be assessed. Experiments with the 5-km datasets will allow us to determine the combination of spatial and temporal sampling that provides optimum SWE model performance. We anticipate that the use of these satellite data will greatly improve the accuracy of snow models, providing substantial benefits to the snow-modeling stakeholder community as well as to other land-surface–related processes like evapotranspiration. We plan to make available an Internet site with the instantaneous data for each swath gridded to 0.05° and each pixel labeled with the time of orbit overpass. We will also provide gridded data at 0.05° for daily values that are based on observations from both the Terra and Aqua satellites.

The work presented here can serve as a precursor for possible operational products that can advance snowmelt modeling. For instance, the Joint Polar Satellite System (JPSS) of NOAA’s next-generation polar-orbiting operational environmental satellite system was launched in October of 2011, and the launch of the second satellite in the JPSS program is planned for early 2017 (the JPSS-1). JPSS will provide operational continuity of satellite-based observations and products for NOAA Polar-orbiting Operational Environmental Satellites (POES). The JPSS program includes a series of advanced spacecraft and sensitive instruments such as the Visible Infrared Imager Radiometer Suite (VIIRS) that combines the radiometric accuracy of the Advanced Very High Resolution Radiometer (AVHRR), which is currently flown on the NOAA polar orbiters, with the high spatial resolution of the Operational Line-scan System (OLS) flown on DMSP platforms. VIIRS will provide users with spectral coverage from 0.412 to 12.00 μm in 22 bands, imagery at ~375-m nadir resolution in five bands, and moderate-resolution (~750 m at nadir) radiometric quality data. The MODIS-based method is transferable to this system. Another relevant system is the Geostationary Operational Environmental Satellite-R Series (GOES-R), the first of which is scheduled to be launched in 2016. It is the next generation of geosynchronous environmental satellites, which will provide atmospheric and surface measurements of Earth’s Western Hemisphere. The GOES-R series of satellites (GOES-R, S, T, and U) will extend the availability of these systems through 2036 and provide information at hourly time scale at native 1-km spatial resolution relevant to snowmelt modeling in rugged terrain. The currently planned operational product for GOES-R will be at 0.5° only. As yet, the need for products of higher spatial resolution and on slopes has not been addressed.

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