Remote sensing of snow characteristics in the southern Sierra Nevada

JEFF DOZIER
Center for Remote Sensing and Environmental Optics
University of California
Santa Barbara, CA 93106, U.S.A.
and
Jet Propulsion Laboratory
California Institute of Technology
Pasadena, CA 91109, U.S.A.

ABSTRACT Estimation of snow characteristics from satellite remote sensing data requires that we distinguish snow from other surface cover and from clouds, compensate for the effects of the atmosphere and rugged terrain, and interpolate snow albedo over the entire solar spectrum from measurements at a few wavelengths. We also need to account for topographic effects without requiring that satellite data be precisely registered to digital elevation data, because the poor quality of most digital elevation data introduces considerable noise into calculations of slope and azimuth. From simulation of a range of snow types and various atmospheric profiles, over possible illumination conditions, we can develop typical spectral signatures above the atmosphere over mountaneous terrain, which allow us to distinguish several classes of snow from other surface covers.

Télédétection des caractéristiques de la neige dans le sud de la Sierra Nevada

RÉSUMÉ L'estimation des caractéristiques de la neige d'après des données satellitaires de télédétection nécessite la distinction entre la neige et d'autres couvertures de surface et les nuages, la compensation des effets de l'atmosphère et du terrain accidenté et l'interpolation de l'albédo de la neige dans l'ensemble du spectre solaire d'après les mesures à quelques longueurs d'ondes. Il faut aussi rendre compte des effets topographiques sans exiger un
enregistrement précis des données satellitaires aux données numériques d'altitude. En effet, la médiocrité de la plupart des données numériques d'altitude introduit de nombreuses possibilités d'erreurs dans les calculs de pente et d'azimut. A partir de la simulation d'une gamme de types de neige et de divers profils atmosphériques pour des conditions éventuelles d'éclairage, nous pouvons concevoir des signatures types de spectres au-dessus de l'atmosphère qui surmonte les terrains montagneux. On peut ainsi distinguer plusieurs classes de neige des autres couvertures de surface.

INTRODUCTION

In attempting to use remote sensing to map snow and estimate snow characteristics over remote and inaccessible areas, we are faced with several problems: (1) We must distinguish snow from other surface cover and from clouds. (2) We must compensate for the effects of the atmosphere and rugged terrain. (3) Our space-borne radiometers typically measure reflectance in a few wavelength bands, but for climate modelling we are interested in snow's reflectance throughout the solar spectrum.

USE OF DIGITAL ELEVATION MODELS IN RADIATION CALCULATIONS

In all but very gentle terrain, significant variation in the surface climate results from local topographic effects. The major contributors to this variation are solar and longwave (thermal) irradiance, although there are also important topographic variations in wind speed and soil moisture. The topographic effects on solar irradiance are mainly variation in illumination angle and shadowing from local horizons. In the thermal part of the electromagnetic spectrum the emission from surrounding slopes usually causes valley bottoms to receive more thermal irradiance than unobstructed areas.

Various problems in calculating radiation over mountainous areas have been addressed by a stream of papers that have appeared in the last two decades (e.g. Garnier & Ohmura, 1968; Williams et al., 1972; Brazel & Outcalt, 1973; Unsworth, 1975; Dozier & Outcalt, 1979; Klucher, 1979; Marks & Dozier, 1979; Dozier, 1980; Arnfield, 1982; Dave & Bernstein, 1982; Olyphant, 1984; Anderson, 1985; Biber, 1986; Olyphant, 1986; Proy, 1986).

Most radiation calculations over terrain are made with the aid of digital elevation grids, whereby elevation data are represented by a matrix. In the United States these are available as "Digital Elevation Models" (DEM) from the U.S. Geological Survey (Elassal & Caruso, 1983). The 1:250,000 quadrangles for the entire United States are available at 63.5 m grid resolution (0.01 inch at map scale), and the 1:24,000 quadrangles are available at 30 m grid
resolution. Figure 1 is a gray-scale rendition of the 30 m DEM for the Mt. Tom area in the southern Sierra Nevada.

Dozier and Marks (1987) describe how to use digital elevation models to provide a topographic boundary description for an atmospheric radiative transfer calculation. They modify a two-stream model (Meador & Weaver, 1980) to account for local illumination angles, shadowing by local horizons, reflection from adjacent terrain, and masking of the sky hemisphere by terrain.

PROBLEMS WITH DIGITAL ELEVATION MODELS

Because digital elevation data in mountainous areas often do not exist, or, if they do, are often of poor quality, any mapping or
classification method that requires that satellite data be registered to digital elevation data is severely constrained and will probably fail. In particular, an algorithm that needs the slope and azimuth of a given pixel in order to interpret its multispectral radiometric signal imposes an impossible requirement, because any noise in the original elevation data is amplified by the differencing operations needed to calculate slope and azimuth.

FIG. 2  Slopes corresponding to elevations in Figure 1. The striping is caused by noise in the DEM

The available digital elevation models are badly contaminated with noise. The noise is not always apparent in images of the DEM itself, or in contour maps made from it, but images of variables derived from the DEM by some sort of differencing operation usually show considerable noise. Figure 2 shows a map of the magnitude of the slopes corresponding to the elevation grid in Figure 1, and Figure 3 shows a solar illumination image, made from Figure 2, a corresponding image of slope azimuths, at a solar zenith angle of 65° and a solar azimuth of 32° east of south. These angles
correspond to those of a Landsat TM image in mid-January. It is clearly not possible to register Figure 3 to the corresponding satellite image (Fig. 4). Any interpretation of the satellite data in Figure 4 that depends on a knowledge of the pixel-by-pixel illumination angle will not work.

Therefore we must develop a method for snow mapping and classification from Landsat Thematic Mapper data that does not depend on registration to a digital elevation grid. We must interpret multispectral satellite data in a way that is sensitive to surface characteristics independent of topography. However, such simple methods as band-ratioing (dividing one band by another) implicitly assume that diffuse irradiance is the same in the two ratioed bands. This is usually not the case. If atmospheric scattering is
greater in one of the bands, the ratio for a surface in the shadow would be different than for the same surface in the sun. It is necessary to account for atmospheric scattering to interpret surface characteristics in shadowed areas.

FIG. 4 January 18, 1983 Landsat TM band 4 image of the area represented in the previous Figures

SPECTRAL FEATURES OF ALPINE SNOW

The spectral signature of snow depends on grain size and on contamination by absorbing impurities (Warren & Wiscombe, 1980; Wiscombe & Warren, 1980). The effective grain radius is approximately the spherical radius that corresponds to the volume/surface ratio of the actual grains. In the visible wavelengths snow albedo is not sensitive to grain size, but is sensitive to minor amounts of contamination by dust, pollen, etc. In the near-infrared wavelengths snow albedo is sensitive to grain size, but not to contamination.
Wavelength bands for the Landsat Thematic Mapper (TM) are given in Table 1; Markham & Barker (1986) give additional details on calibration, dynamic range, differences between Landsat-4 and Landsat-5, etc.

**TABLE 1  Landsat Thematic Mapper wavelength bands**

<table>
<thead>
<tr>
<th>Band No</th>
<th>Wavelengths (µm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>TM1</td>
<td>0.45 - 0.52</td>
</tr>
<tr>
<td>TM2</td>
<td>0.53 - 0.61</td>
</tr>
<tr>
<td>TM3</td>
<td>0.62 - 0.69</td>
</tr>
<tr>
<td>TM4</td>
<td>0.78 - 0.90</td>
</tr>
<tr>
<td>TM5</td>
<td>1.57 - 1.78</td>
</tr>
<tr>
<td>TM6</td>
<td>10.4 - 12.5</td>
</tr>
<tr>
<td>TM7</td>
<td>2.10 - 2.35</td>
</tr>
</tbody>
</table>

A measurement of snow reflectance in the near-infrared (for example TM4) allows us to estimate an effective radiative grain radius, while a measurement in the visible (for example TM2) gives an estimate of the extent to which snow albedo is degraded by contamination. Snow can be distinguished from both water and ice clouds in TM5, because water is less absorptive than ice in this band and the small crystals in cirrus clouds are more reflective than the larger snow grains.

By simulating top-of-atmosphere radiances for a variety of snow grain sizes, Dozier & Marks (1987) show that multispectral signatures result that can be recovered from satellite data over mountainous terrain, via pairwise TM band relationships. In Figure 5, for example, are pairwise clusters of points from TM bands 2, 4, and 5 from the Mt. Tom image (Figure 4). The values plotted are "planetary reflectances", defined by

\[
\rho = \frac{\pi r^2L}{\cos \theta_o S_0}
\]

where L is the satellite-measured radiance in the band, \( \theta_o \) is the solar zenith angle, \( S_0 \) is the exoatmospheric solar irradiance in the band, and r is the earth-sun radius vector. The small dispersion in the top graph results from the fact that the snow grains in this high-altitude winter scene have a narrow size distribution and small amounts of contaminants. In the two lower graphs, we find more dispersion because the reflectances in TM band of 5 for rock, soil and clouds are much greater than that for snow, relative to the reflectance in bands 2 or 4. Values above the diagonal line in the two bottom graphs are too bright in band 5 to be snow. Clouds have high reflectance in all bands, whereas soil and rock have high reflectances in band 5 but not in bands 2 or 4. TM band 1 can be used to distinguish snow in the shadows.
Based on these simulations and on examination of the data in Figure 4, we find that snow is discriminated by the following criteria:

a) $\rho_{TM1} \geq$ some threshold value, even for shadowed areas. In the Mt. Tom TM scene, this value is $\rho_{TM1} \geq 0.26$. 

and

b) $\rho_{TM5} \leq a + b_0\rho_{TM2}$, i.e. areas whose reflectance in TM5 is too large, relative to that in TM2, are not snow. The reason that $a < 0$ is that there is more diffuse irradiance in TM2 than in TM5.

These two criteria can be used to discriminate snow from other surfaces. Snow characteristics - grain size and contaminant amount - are inferred by examination of TM bands 2 and 4. Dispersion in
TM2 is caused by absorbing impurities in the snow; dispersion in TM4 is caused by variations in grain size.

CONCLUSION

The key to interpretation of satellite-measured snow reflectances in alpine terrain lies first in simulation of a variety of conditions using an atmospheric radiation model coupled to a lower boundary condition that depends on topography and on snow grain characteristics. From simulation of a range of snow types and various atmospheric profiles, over all possible illumination conditions, we can develop typical spectral signatures above the atmosphere over mountainous terrain. These allow us to distinguish several classes of snow from other surface covers, even though snow in the shadows is darker than other surfaces in sunlight. In particular, we discriminate snow from rocks, bare soil, and vegetation over soil, and we classify snow on the basis of grain size and contamination amounts. Thick clouds are easily distinguished from snow, and thin clouds over snow can be distinguished from thin clouds over soil.

REFERENCES

Markham, B.L. & Barker, J.L. (1986) Landsat MSS and TM post-
calibration dynamic ranges, exoatmospheric reflectances and at-satellite temperatures. *EOSAT Landsat Tech. Notes.* 1, 2-8, Lanham, MD.


