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REFLECTANCE MEASUREMENTS FROM LANDSAT THEMATIC MAPPER OVER RUGGED TERRAIN

J. DOZIER

University of California
Santa Barbara, California

ABSTRACT

Spectral albedo measurements from the Landsat-4/5 Thematic Mappers require that spacecraft upwelling radiances be corrected for atmospheric absorption and scattering and for local surface illumination. A two-stream model is developed, with a lower boundary condition that varies with incidence angle. TM data must be registered to digital terrain data. Reflectance from points in shadows can be used to estimate optical depth. Our primary application is determination of the spectral albedo of snow. The TM is better-suited for this purpose than the MSS because of its larger dynamic range.

I. INTRODUCTION

Satellite remote sensing has become increasingly important to study of the land surface climatology, because the data provide information on the spatial distribution of important parameters such as albedo, surface temperature, snow cover, vegetation index, etc. In snow and ice studies (my own particular interest) remote sensing has been used to improve the monitoring of existing conditions and has been incorporated into several runoff forecasting and management systems.

The most common operational use of remote sensing in snow studies is to monitor snow covered area (see papers by Rango in the REFERENCES), and satellite derived measurements of snow covered area are used as indices in snowmelt runoff models. The next step involves use of the radiometric characteristics of the satellite data. Measurements of snow reflectance from the Landsat-4/5 Thematic Mappers should lead to improved use of satellites in snow hydrology, because the data can be used in surface energy balance calculations. Basin-wide spectral albedo measurements from the TM could be used to better understand and predict the timing of the spring runoff, because these data can be combined with solar radiation calculations to estimate the net radiation balance.

II. TM RADIOMETRIC CHARACTERISTICS

Table 1 gives some radiometric characteristics of the Landsat-4 Thematic Mapper, launched in July 1982. Landsat-5 was launched March, 1984. Bands are listed in spectral order. In the radiance columns of the table, quantization and saturation radiances of the sensor bands are compared with the solar constant, integrated through the sensor response functions. Solar constant spectral distributions are from the NASA standard (Thekaekara, 1970) adjusted to fit the integrated values measured from the Nimbus-7 cavity radiometer of the earth radiation budget experiment (Hickey _et al._, 1980).

The last column in the table expresses the sensor saturation radiance as a percentage of the solar constant, integrated through the band response function. Except for band 1, these percentages are all significantly higher than comparable wavelength channels on the Landsat Multispectral Scanner. Moreover, the radiometric resolution NEΔL is better on the Thematic Mapper, because the signal is quantized to 8 bits instead of 6. Snow will frequently saturate TM1, but saturation in the other channels is usually confined to a small portion ($\leq$1%) of the pixels in a snow covered scene.

Table 1. Landsat-4 TM Radiometric Characteristics

band	μm		radiances ($W\ m^{-2} \mu m^{-1} sr^{-1}$) NEΔL	sat.	solar	%
TM1	.45	.52	.63	158	621	26
TM2	.53	.61	1.22	308	540	57
TM3	.62	.69	.92	235	468	50
TM4	.78	.90	.89	224	320	70
TM5	1.57	1.78	.13	32	66	49
TM7	2.10	2.35	.07	17	24	69
TM6	10.42	11.66	(thermal band)			

Table 2. TM Snow/Cloud Reflectance (60° illumination angle)

band	clean semi-infinite snow, optical grain radius (μm) 50	100	200	500	1000
1	.992	.988	.983	.974	.963
2	.988	.983	.977	.964	.949
3	.978	.969	.957	.932	.906
4	.934	.909	.873	.809	.741
5	.223	.130	.067	.024	.011
7	.197	.106	.056	.019	.010

band	water cloud, 1mm water, optical droplet radius (μm) 1	2	5	10	20
5	.891	.866	.769	.661	.547
7	.784	.750	.650	.481	.345

band	ice cloud, 1mm water equivalent, optical crystal radius (μm) 1	2	5	10	20
5	.817	.780	.665	.513	.383
7	.765	.730	.642	.478	.341

III. SNOW/CLOUD REFLECTANCE

Calculations of snow reflectance in all 6 TM reflective bands (i.e. 1, 2, 3, 4, 5, and 7), using a delta-Eddington model (Wiscombe and Warren, 1980; Choudhury and Chang, 1981) show that snow reflectance is sensitive to grain size in TM4 but not in TM1 or TM2. The same model can be used to calculate cloud reflectance. Table 2 shows calculations of integrated reflectance for snow of optical grain size 50-1000 μm over all reflective TM bands, and for water and ice clouds with thickness of 1 mm water equivalent over TM5 and TM7. An optical grain size of 50 μm corresponds to the highest snow reflectances measured, for fine, new snow in Antarctica. An optical grain size of 1000 μm is typical of snow that has undergone melt-freeze metamorphism. The cloud thickness of 1 mm was chosen to represent typical small, thin clouds that might obscure satellite observations of snow and that might not be evident in other wavelengths bands. Table 2 does not include any correction for atmospheric attenuation, the topic covered in the next section.

In the blue and green bands (1-2) snow reflectance is less sensitive to grain size, so measurements in these wavelengths will show the extent to which snow albedo is degraded by contamination from atmospheric aerosols, dust, pine pollen, etc. In the red and near-infrared bands (3-4), snow reflectance is sensitive to grain size but not to contaminants, so grain size estimates in these wavelengths can be used to spectrally extend albedo measurements. In both TM "shortwave infrared" bands, TM5 and TM7, snow is much darker than clouds, and water clouds are brighter than ice clouds in TM5. Warren (1982) and Dozier (1984) give physical explanations for these snow/cloud reflectance attributes.

IV. ATMOSPHERIC CORRECTION

A. PLANETARY ALBEDO

From the values in Table 1, digital satellite radiance numbers can be converted to radiances. At this stage we make the Lambertian assumption: upwelling radiance is independent of viewing direction. The apparent planetary albedo, derived directly from the satellite data with no corrections for terrain, is

$$\rho_p = \frac{L}{\mu_o S_o R^{-2}}$$

L is radiance at the satellite, integrated over the wavelength band; μ_o is the solar zenith cosine on a horizontal surface; πS_o is the spectral solar constant, integrated over the wavelength band (the "solar" values in Table 1); and R is the earth-sun radius vector (ratio of earth-sun distance to its mean value).

B. ATMOSPHERIC MODEL

Atmospheric correction over areas of mountainous terrain has only recently been examined in the literature (Dozier and Frew, 1981; Sjoberg and Horn, 1983). A new approach that appears more promising than previous algorithms is to calculate planetary albedo ρ_p, treating the atmosphere as a homogeneous layer and using the surface illumination angle and surface reflectance ρ_o for the lower boundary condition. The TM data must be registered to digital terrain data, so that we can correct for varying illumination angle and shadowing by adjacent terrain (Frew, 1984).

The following system of first-order ordinary differential equations approximates the radiative transfer equation for non-emission conditions with the phase function averaged over azimuth (Meador and Weaver, 1980). In this "two-stream" approximation for a homogeneous layer with optical depth $0 \leq \tau \leq \tau_o$, radiance is separated into downward $L_\downarrow$ and upward $L_\uparrow$ components.

$$\frac{dL_\uparrow}{d\tau} = \gamma_1 L_\uparrow - \gamma_2 L_\downarrow - S_o R^{-2} \omega_o \gamma_3 e^{-\tau/\mu_o}$$

$$\frac{dL_\downarrow}{d\tau} = \gamma_2 L_\uparrow - \gamma_1 L_\downarrow + S_o R^{-2} \omega_o \gamma_4 e^{-\tau/\mu_o}$$

ω_o is the single-scattering albedo. The γ's are chosen according to the approximation used for the phase function, and depend on ω_o, the phase asymmetry parameter g, and μ_o. Meador and Weaver (1980) derive γ's for 7 different approximations.

The common upper boundary condition is that there is no incoming diffuse radiation at the top of the atmosphere:

$$L_\downarrow(0) = 0$$

Over mountainous terrain the lower boundary condition is complicated, because the surface illumination angle $\arccos\mu_s$ is not necessarily the same as μ_o, and because a portion of the incoming radiation is reflected from adjacent terrain. The "view factors" V_d and V_s represent the portion of the overlying hemisphere obscured by terrain and corrected for angular effects. V_d is the view factor for incident diffuse irradiance; V_s is the view factor for incident direct irradiance. The lower boundary condition is

$$L_\uparrow(\tau_o) = \rho_o \{S_o e^{-\tau_o/\mu_o} [\rho_o V_s + \mu_s] + L_\downarrow(\tau_o) [1 - V_d(1-\rho_o)]\}$$

With the Lambertian assumption the satellite measures ρ_p. For near-nadir viewing satellites, we expect that an anisotropic correction can be applied empirically. Solution of the differential equations leads to a complicated expression of the form

$$f(\rho_p, \rho_o, \mu_o, \mu_s, V_d, V_s, \omega_o, g, \tau_o) = 0$$

Of these variables ρ_p, μ_o, μ_s, and the V's are known. If the scattering properties of the atmosphere, but not the density of the scattering elements, are known, then ω_o and g are also known. The only unknowns are therefore ρ_o and τ_o, the surface reflectance (which is what we want to measure) and the optical depth of the atmosphere in the wavelength band.

Now if we have a measurement at two different values of μ_s over areas where ρ_o is the same, the equation can be solved for ρ_o for those pixels and τ_o at that elevation. Generally τ_o varies with elevation in an exponential way, i.e.

$$\frac{\tau_o(z)}{\tau_o(z_o)} = e^{-(z-z_o)/H}$$

H, the scaling height, is determined from values of τ_o at two different elevations. Once this relationship is established, so that τ_o can be estimated for all elevations, then spectral albedo ρ_o can be estimated for all pixels.

C. FUTURE PLANS

The approach can be tested by comparison with a detailed atmospheric model, based on LOWTRAN6 (Kneizys *et al*., 1983) and ATRAD80 (Wiscombe, 1976), but with modifications to allow computation of azimuthally-dependent radiance instead of just azimuthally-averaged radiance. For a range of atmospheric profiles, we will compare the upwelling radiance at the satellite, over the range of viewing angles for the TM, with the values calculated for the simpler two-stream model described above. If the relationship is systematic, the simpler, invertible model can be used for atmospheric correction.

V. CONCLUSION

Landsat-4/5 Thematic Mapper data can be used to determine spectral albedo values over mountainous terrain. All TM channels except 1 have suitable dynamic ranges for snow albedo measurement. The atmospheric correction requires no correlative measurements but assumes that pixels in shadow near those in sunlight have the same albedo.

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Jeff Dozier is Associate Professor of Geography and Director of the Computer Systems Laboratory at the University of California, Santa Barbara. His research interests are in snow hydrology, remote sensing, atmospheric radiation, and computer software development. He received his Ph.D. from the University of Michigan in 1973.