9 D1





TO Harold Oseroff
Code 902
NASA Goddard Spaceflight Center
Greenbelt, MD 20771

FROM Jeff Dozier
Department of Geography
University of California
Santa Barbara, CA 93106

CC Contracting Officer, NASA/GSFC
Publications Section, NASA/GSFC
Patent Counsel, NASA/GSFC
James Foster, NASA/GSFC
Carlena Leufroy, ONR, Pasadena

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**Landsat-D Investigations in Snow Hydrology**

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*Expenditures for period:*

Salaries
Benefits
Travel
Computer Time
Telephone
Xeroxing
Mail
Supplies & Materials (petty cash, instrumentation, etc.)
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TOTAL

D.1 **Atmospheric Model Development**

(paper prepared for LARS Symposium, June 1984)

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 $NE\Delta L$ 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 (≤1%) of the pixels in a snow covered scene.

Table 1. Landsat-4 TM Radiometric Characteristics

| band | $\mu m$ | | radiances ($W\ m^{-2} \mu m^{-1} sr^{-1}$) | | | |
|---|---|---|---|---|---|---|
| | | | NE$\Delta L$ | 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) | | | |

## 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 $\mu m$ 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 $\mu m$ corresponds to the highest snow reflectances measured, for fine, new snow in Antarctica. An optical grain size of

1000 $\mu m$ 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.

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

| | clean semi-infinite snow | | | | |
|---|---|---|---|---|---|
| | optical grain radius ($\mu m$) | | | | |
| band | 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 |
| | water cloud, 1$mm$ water | | | | |
| | optical droplet radius ($\mu m$) | | | | |
| band | 1 | 2 | 5 | 10 | 20 |
| 5 | .891 | .866 | .769 | .661 | .547 |
| 7 | .784 | .750 | .650 | .481 | .345 |
| | ice cloud, 1$mm$ water equivalent | | | | |
| | optical crystal radius ($\mu m$) | | | | |
| band | 1 | 2 | 5 | 10 | 20 |
| 5 | .817 | .780 | .665 | .513 | .383 |
| 7 | .765 | .730 | .642 | .478 | .341 |

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

$$\rho_p = \frac{L}{\mu_0 \, S_0 \, R^{-2}}$$

$L$ is radiance at the satellite, integrated over the wavelength band; $\mu_o$ is the solar zenith cosine on a horizontal surface; $\pi S_o$ 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 $\rho_p$, treating the atmosphere as a homogeneous layer and using the surface illumination angle and surface reflectance $\rho_o$ 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 $0 \le \tau \le \tau_o$, radiance is separated into downward $L_\downarrow$ and upward $L_\uparrow$ components.

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

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

$\omega_o$ is the single-scattering albedo. The $\gamma$'s are chosen according to the approximation used for the phase function, and depend on $\omega_o$, the phase asymmetry parameter $g$, and $\mu_o$. Meador and Weaver (1980) derive $\gamma$'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 $\mu_o$, 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 $\rho_p$. 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 $\rho_p$, $\mu_o$, $\mu_s$, and the $V$'s are known. If the scattering properties of the atmosphere, but not the density of the scattering elements, are known, then $\omega_o$ and $g$ are also known. The only unknowns are therefore $\rho_o$ and $\tau_o$, 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 $\mu_s$ over areas where $\rho_o$ is the same, the equation can be solved for $\rho_o$ for those pixels and $\tau_o$ at that elevation. Generally $\tau_o$ varies with elevation in an exponential way, i.e.

$$\frac{\tau_0(z)}{\tau_0(z_0)} = e^{-(z-z_0)/H}$$

$H$, the scaling height, is determined from values of $\tau_0$ at two different elevations. Once this relationship is established, so that $\tau_0$ can be estimated for all elevations, then spectral albedo $\rho_0$ 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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D2

**Registration of TM Data to Digital Elevation Models**

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ABSTRACT

Several problems arise when attempting to register Landsat Thematic Mapper (TM) data to U.S. Geological Survey digital elevation models (DEMs). Chief among these are:

- TM data are currently available only in a rotated variant of the Space Oblique Mercator (SOM) map projection. Geometric transforms are thus required to access TM data in the geodetic coordinates used by the DEMs. Due to positional errors in the TM data, these transforms require some sort of external control.
- The spatial resolution of TM data exceeds that of the most commonly available DEM data. Oversampling DEM data to TM resolution introduces systematic noise. Common terrain processing algorithms (e.g., slope computation) compound this problem by acting as high-pass filters.

## I. INTRODUCTION

Many applications of Landsat Thematic Mapper (TM) data are contingent upon the image data being registered with a geographic database. In particular, the use of TM data to derive surface reflectance information requires knowledge of the terrain characteristics (elevation; attitude; relation to surrounding terrain) of each image pixel (Dozier, 1984). Digital terrain information for areas in the United States is readily obtained from the "digital elevation models" (DEMs) distributed by the National Cartographic Information Center (NCIC, 1982). This paper will therefore examine some of the problems inherent in coregistering TM and DEM data.

## II. BACKGROUND

### A. GEOMETRIC CHARACTERISTICS OF TM AND DEM DATA

The instantaneous-field-of-view (IFOV) of the TM, at a nominal spacecraft altitude of 705.3 $km$, yields a spatial resolution of 30 $m$ (120 $m$ for band 6) at the Earth's surface (Engel, 1980). The geometry of an unprocessed TM scene is quite complex (Beyer, 1980) and will not be discussed here, since almost all investigators utilize geometrically preprocessed ("P"-level) TM images (NASA, 1982; 1983). P-level TM data are resampled to a 28.5 $m$ pixel size and are cast into the Space Oblique Mercator (SOM) map projection, a cylindrical projection whose centerline is the satellite groundtrack (Snyder, 1981). Other map projections, notably Universal Transverse Mercator (UTM) and Polar Stereographic, are to be provided in the future.

NCIC DEM data are currently provided in two formats. The higher resolution data are 30 $m$ grids in the UTM projection, registered to standard U.S. Geological Survey 7.5 minute 1:24000 scale map quadrangles. The lower resolution data are 3 arc-second (approximately 90 $m$ at the equator) geodetic grids, derived from and registered to

USGS 1 degree by 2 degree 1:250000 scale map quadrangles. Low resolution DEM data are available for the entire United States. High resolution data are presently available only for selected areas.

B. DATA USED IN THIS INVESTIGATION

The DEM data source for this investigation was a 3 arc-second resolution geodetic grid corresponding to sheet NJ11-10 (Fresno, CA) of the USGS 1:250000 scale map series. A subset equivalent to the Mt Tom, CA quadrangle of the USGS 1:62500 scale map series was extracted for a study area. Higher resolution DEM data were not available for this area.

The TM data source was a P-level tape of scene E-40186-18024 (path 42, row 43, 18 January 1983), in the SOM projection with 28.5 $m$ pixels. A subscene encompassing the Mt Tom quadrangle was extracted.

Coregistering these data sets thus involved the following operations:

- a geometric transformation between SOM and geodetic coordinates
- resampling one of the data sets to the resolution of the other

## III. GEOMETRIC TRANSFORMATIONS

Image geometric transformations are generally run in reverse (Bernstein, 1975); that is, locations in a target image are mapped into a source image, from which pixel values are selected to be placed in the target image. This guarantees that every location in the target image will be assigned a value. For this investigation, the target image was selected to be the TM image; i.e., the DEM data were to be registered to the TM data. Selection of the TM image as the target was based on two considerations: retention of spatial resolution, and avoidance of double-resampling.

Mapping TM image locations into the DEM grid is a 3-step process:

[1]convert TM image coordinates (line$_T$, sample$_T$) to SOM coordinates $(x, y)$; then

[2]convert $(x, y)$ to geodetic coordinates $(\varphi, \lambda)$; then

[3]convert $(\varphi, \lambda)$ to DEM grid coordinates (line$_D$, sample$_D$)

Step [1] is complicated by the fact that the TM image grid is rotated and shifted with respect to the Landsat SOM coordinate system. Presumably, the image grid is rotated so as to align as closely as possible with the raw TM scan lines, and thus minimize the line-buffering required to construct a corrected TM scan line. The rotation and offset information necessary to perform step [1] is obtained from the HAAT (header, ancillary, annotation, and trailer) data file on the TM P-tape.

Step [2] performs the transformation from UTM to geodetic coordinates. For this investigation, routines from the USGS General Map Projection Package were used (Thormodsgard and DeVries, 1982). Step [3] is trivial, since the DEM grid is aligned with the geodetic coordinate system.

Instead of evaluating steps [1-3] above for each point in the TM image, a regularly spaced mesh of 300 TM locations were transformed. The resulting list of line$_T$, sample$_T$, line$_D$, sample$_D$ was fed into a stepwise regression program, which generated the coefficients of two polynomial mapping functions. For the study area selected, all terms higher than first order were insignificant. A general-purpose image warping program evaluated these polynomials to assign each TM grid location a counterpart in the DEM grid. The revised geometric processing sequence is thus:

[a]Evaluate the complete TM-SOM-geodetic-DEM coordinate transform sequence for a sparse mesh of TM grid locations.

[b]Perform a stepwise regression on the location pairs generated in step [a] to determine a the coefficients of simple polynomial transforms.

[c]Substitute the polynomials from step [b] for the transforms in step [a], and compute the TM-DEM transform directly.

Steps a-c are computationally much faster than steps 1-3, since the SOM-geodetic transforms in particular require extensive trigonometric calculations.

The procedure outlined so far assumes that both the TM and DEM data are precisely located in their own coordinate systems. The DEM data have been planimetrically edited, but the TM data still contain gross linear positional errors, apparently due to uncertainties in the satellite position. That the errors are positional rather than attitudinal is inferred by the fact that a simple translation may bring the TM and transformed DEM grids into registration. In our investigation, this final translation was accomplished by displaying the instantaneous difference between the two images on a video display processor (IIS, 1979) while moving one of them under trackball control.

## IV. RESAMPLING

The DEM data must be resampled to assign elevation values to any non-integral DEM locations selected by the above transforms. Two resampling algorithms were tried: nearest neighbor (zero-order) and "cubic convolution" (Simon, 1975). To match the spatial resolution of the TM data, the DEM data must be oversampled - a single DEM pixel will map into several TM locations. For this reason, nearest neighbor resampling produced an unacceptably "jagged" output image (essentially, a zoom-by-replication operation), and cubic convolution became the preferred resampling method.

An unfortunate side effect of the scale change in the elevation data was apparent with both resampling methods. The oversampling caused the orientation of the raw DEM grid to be visible as a regular pattern in the "jagged" edges. The effect was naturally more noticeable with nearest neighbor resampling, but was also present when cubic convolution resampling was used.

The presence of regularities in the resampling-induced discontinuities had a severe impact on the generation of terrain slope information. Our standard terrain processing involves the generation of two gradient images, one representing the magnitude (slope), and the other the direction (exposure). Since the gradient operation acts as a high-pass filter, the resampling-induced grid pattern was accentuated in the slope and exposure images, essentially swamping them with high-frequency systematic noise.

Various approaches were tried to mitigate this problem. Cubic convolution produced less noise than nearest neighbor resampling. The slope image was less noisy when computed from the raw DEM data and then resampled to TM resolutions, than when computed from the resampled elevation data. This approach could not be used for the exposure image; since exposures are stored as azimuthal angles, values for which overflow or underflow are handled incorrectly.

## V. CONCLUSIONS

Three problems currently impede the integration of TM and DEM data:

- In most circumstances, TM and DEM data are not available in the same map projection, necessitating geometric transformation of one of the data types. This problem should be alleviated by the forthcoming general availability of TM and DEM data in the UTM projection.
- The TM data are not accurately located in their nominal projection. Human intervention is required to fine-tune image locations. Introduction of ground control points into the P-tape generation process, already in progress, should improve this situation.
- TM data have higher resolution than most DEM data, but oversampling the DEM data is not practical. The full resolution of TM data thus cannot be expoited over areas where high-resolution DEM data are unavailable.

### A. FURTHER WORK

When TM and DEM data which meet the above criteria are available for a mountainous region, relief displacement effects should be investigated. Simple "back-of-the-envelope" calculations indicate that relief displacements of up to 10 pixels could be expected in a study area like that used in this investigation.

Given TM and DEM data of equivalent spatial resolutions, the desirability of registering the TM data to the DEM grid, rather than the reverse, should be investigated. While this approach involves re-resampling the TM data, it allows the use of an unrotated grid as the geographic reference.

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