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REMOTE SENSING OF SNOWPACK DENSITY USING SHORTWAVE RADIATION

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ABSTRACT

Albedo or satellite radiance measurements can be used to estimate average snowpack density by means of a multiple linear equation. The in situ data equation predicted density with a correlation (r^2) of 0.79 and a standard error of 0.027 gm cm^{-3}. The data from LANDSAT-1 were not as significant in a similar equation, possibly because of the large field of view.

INTRODUCTION

Current research on remote sensing of snowpack parameters, other than of areal extent, is largely centered in the microwave and gamma-ray portions of the electromagnetic spectrum. It appears possible, however, to use measurements in the shortwave region to estimate snowpack density statistically. In situ measurements at the Central Sierra Snow Laboratory and remote observations from the LANDSAT-1 spacecraft show the possibility of estimating average snowpack density with albedo or radiance measurements, respectively.

Snow courses and related instruments are the current source of snow water equivalent data used in streamflow and flood prediction equations. Snow water equivalent may be remotely obtained from pressure pillows (Beaumont, 1965) and from the horizontal-type radioisotope snow gage (Peck, 1972). Depth, density, and water equivalent may be obtained from a profiling nuclear snow gage (Smith, Halverson, and Jones, 1972). Only a limited number of these remotely-operated snow sensors can be deployed in an area because of the high cost of installation and operation.

To meet the increasing demands on the snow water resource requires more accurate streamflow estimates than those now being obtained through conventional snow survey methods. Streamflow estimates can be improved by increasing the sampling frequency temporally and spatially at a more reasonable cost than that associated with remote operation of snow sensors. This paper suggests a method for accomplishing this task. The method is based upon measurement of snow albedo from either aircraft-mounted or spacecraft-mounted radiometers, and the relation of these readings to point estimates of snowpack average density at certain sites. The necessary additional data would be obtained from the already established network of hydrometeorological stations where such data as temperature, precipitation, and snow depth, density, and water equivalent are currently monitored, or from a small number of new instrumented sites.

SELECTED LITERATURE REVIEW

The use of albedo measurements to monitor the physical condition of snow has been suggested in several previous studies. In 1956, Dunkle and Bevans wrote that their analysis of the reflectance and transmittance of snow would be of value in correlating albedo measurements with the physical characteristics of snow. Mil'kis (1956) published equations and resulting nomograms for finding snowpack density and albedo knowing elevation and date. Although Mil'kis emphasized that the density was given merely to characterize the state of the snow, his equations could be combined to estimate density knowing albedo. Giddings and LaChapelle (1961) found reflectance depends on the area of ice-to-air interface per unit volume of snow. They suggested "the possibility of following metamorphic changes in snow by optical methods without disturbing the naturally occurring processes." Experiments on the extinction coefficients of visible light (.4-.7μm) through snow as a function of density, grain size, and wavelength are presented in Mellor, **1965**. One of his conclusions was remote sensing might be used to detect subtle differences within the snowpack since the magnitude and wavelength of reflectance vary with snow type. Bergen (1970, 1971) also found a relation between the extinction coefficient and snow density and grain size. His experiments with CdS cells (.45-.90μm) settled within the snowpack agree with the correlation between transparency of snow and its air permeability and density suggested by the model of Dunkle and Bevans (1956). In 1974, Bohren and Barkstrom published a study on the theoretical optical properties of snow. Their calculations indicate very little (less than 5%) reflection takes place at an ice grain's surface; most radiation is internally transmitted. Hence, the results of remote observations of the snowpack would be dependent upon the snow's internal structure. Bohren and Barkstrom's calculations also indicate albedo under diffuse illumination should be independent of density and proportional to the square root of grain size. Snowpack remote sensing using shortwave radiation from spacecraft, however, is restricted to the non-diffuse condition of cloud-free weather. Further, an empirical relationship should exist between grain size and density since the two are related. Recently, O'Brien and Munis (1975) studied the spectral reflectance of snow from 0.6 to 2.5μm. Their measurements indicate a reduction in the reflectance results from the combination of densification and increased particle size that occurs with aging. In addition, O'Brien and Munis related various snow sample densities at wavelengths of 1.0, 1.1, 1.3, 1.8, and 2.4μm to their corresponding reflectances. They reported, "In each case, the decrease in reflectance with density showed about a 0.8 correlation at the

0.01 confidence level."

METHOD

In Situ Study

The relationship between albedo and snowpack density was investigated at the Central Sierra Snow Laboratory at Soda Springs, California. The Laboratory is a field station of the U.S. Forest Service, Pacific Southwest Forest and Range Experiment Station, and is located just to the west of the crest of the Sierra Nevadas at an elevation of 2100 m.

Measurements were compiled for the 1969-1970 through 1971-1972 snow seasons. For the three seasons, 553 pairs of average density and concurrent snow albedo measurements were taken. Some days during each season were excluded. These included days on which storms were in progress, days following precipitation events of less than 7 mm water equivalent, and days when snow density measurements were not taken.

The dependent variable, average snowpack density, was measured with the isotopic profiling snow gage (Smith, Halverson, and Jones, 1972). Average snowpack density is the average of 1 cm incremental densities from the soil surface to the snow-air interface. Density determinations by this method have a standard error near .01 gm cm^{-3}.

Snow surface albedo measurements were taken with two pyrheliometers (0.3-4.0μm) situated 1 m above the snow surface, one mounted in the normal position and the second inverted. Decimal albedo determinations were made at half-hour intervals for three hours before or after the snow density measurements.

Developing a relationship to predict average snowpack density requires some information beyond albedo. Additional variables included the date of measurement, number of days since the last storm, and the type of precipitation in the last storm.

The rate of new snow densification and the density of the existing snowpack vary throughout the snow season. Late season snowpacks are denser than early season packs and new snow metamorphosis proceeds faster in late season. The date was expressed in the equations as solar declination, the angular displacement of the Sun north (+) or south (-) of the celestrial equator. Solar declination can be found in annual tables provided in several Naval Observatory publications (1975a or 1975b).

The type of precipitation affects the average density of the resulting snowpack. Winter precipitation may not be snowfall; it may be all rain or rain-and-snow mixed. A rain-on-snow event results in a rapid decrease in albedo and a rapid increase in density. Storms were coded into three groups: snow only, mixed snow and rain, and rain only. The precipitation type was then assigned the value 0, 0.5, or 1.0, respectively.

The several variables which appeared important were combined into a functional expression

$$RHO = f(A, SD, D, R)$$

in which

RHO = average snowpack density (decimal)
A = albedo (decimal)
SD = solar declination (degrees)
D = days since cessation of storm (integer)
R = proportion of rain to snow in last storm (0, 0.5, 1.0)

For a more complete description of the initial study as well as a study on areal snowpack variation, see Halverson and Smith (in process).

Cloud cover can significantly affect the shortwave radiation balance over snow (Ambach, 1974). The diffuse radiation normally present under overcast skies becomes even more uniform when over a snow surface (Fritz, 1955). To investigate the influence of cloudiness on the albedo correlation, the data were divided into three sets. The classification was based on average daily cloud cover as reported by the nearby Blue Canyon weather service office (NOAA, 1972). Days on which 0 to 2 tenths cloud cover were reported were considered clear days, 3 to 7 tenths were considered partly cloudy, and 8 to 10 tenths cloud cover days were classified as overcast.

LANDSAT Study

Regression methods were also applied to the spacecraft data (McMillan, 1975). Terrestrial radiance values over snow courses in the American River Basin, California, were obtained from the multispectral scanner subsystem (MSS) onboard the LANDSAT-1 satellite. Data from MSS band 7 (0.8-1.1μm) were chosen to avoid the problem of saturation of the sensor. A description of the satellite sensors and orbital parameters can be found in General Electric Corporation, 1972.

Digital radiance data from the MSS were obtained from LANDSAT computer tapes in unrectified, contoured 32 by 32 pixel arrays. These arrays were combined into a large map of the study area. Topographic maps (1:24,000 scale), with snow-course locations indicated, were optically enlarged, superimposed, and registered over the data arrays. The highest pixel value corresponding to the snow course location was recorded.

A variable to normalize the satellite radiance values for differing irradiance levels was needed. The potential solar insolation at each snow course was computed using the method of Frank and Lee (1966). Slope and aspect data were determined from the topographic maps.

Snowpack density is affected by temperature (Corps of Engineers, 1950). The sum of average daily air temperature above freezing since snowfall was included as a variable in this study, and was computed from data given in the monthly Climatological Data Summaries (NOAA, 1973). Data from the Tahoe Valley airport weather station, located near the snow courses, were used.

Snowpack density measurements were provided by the California Cooperative Snow Survey. Data from eight snow courses in the southeastern portion of the American River Basin were used. Only those locations and dates having a minimum of 60 cm of snow were considered. Table 1 lists the snow courses and their elevations.

Since density measurements were normally taken during the first part of the month and the spacecraft has an 18-day revisit cycle, the density had to be adjusted to the date of satellite overpass. This was accomplished by interpolation weighted according to the daily average temperature.

Data from four satellite passes were obtained. The dates of these passes, covering the snowmelt period, were March 16, April 3 and 21, and May 9, 1973. Light, scattered cumulus clouds on April 21, prevented data retrieval from some snow courses.

Snowpack density is known to vary with elevation. High elevation snowpacks densify at a slower rate than low elevation snowpacks. A term representing snow-course elevation was included in the equation. These values also acted as a counter, having the same value per snow course on different dates.

These variables were combined into a functional expression

$$RHO = f(RAD, SD, DEG, I, E)$$

in which the undefined terms are

RAD = satellite radiance (integer)
DEG = sum of average daily air temperature above freezing since snowfall (degrees, F)
I = potential solar insolation (decimal)
E = snow course elevation (meters)

RESULTS

The analyses were performed using a stepwise linear regression (IBM, 1970), a statistical technique for analyzing a relationship between a dependent variable and a set of independent variables. The criterion of importance is based on the reduction of sum of squares. In the equations that follow, the independent variables are listed in their order of importance. Regression coefficients and their significance for all equations are given in Tables 2 through 6.

In Situ Study

Initial graphical analysis showed albedo to be a predictor of average snowpack density, but the relationship was not linear. The square of albedo showed a better fit, and it was adopted for future analysis.

The equation using in situ data for all cloud conditions is:

$$\begin{aligned} RHO = &.00323SD + .00193D + .0279R \\ &- .0756A^2 + .412 + error. \end{aligned} \qquad (1)$$

The multiple correlation coefficient (r^2) is 0.79, with a standard error of estimate of 0.027 gm cm^{-3}, for 548 degrees of freedom.

The equation using in situ data for overcast conditions is:

$$\begin{aligned} RHO = &.00325SD - .123A^2 + .0334R \\ &+ .00145D + .436 + error. \end{aligned} \qquad (2)$$

The multiple correlation coefficient is 0.79, with a standard error of estimate of 0.031 gm cm^{-3}, for 159 degrees freedom.

LANDSAT Study

Potential insolation term, I, was not significant in reducing the sum of squares. Although there is a slight dependence of density on snow course slope and aspect, this term may be more important in future equations by combining with satellite radiance in a suitable algorithm.

The equation using remote data is:

$$\begin{aligned} RHO = &.00125DEG + .00243SD + (2.93x10^{-6})E \\ &- (2.96x10^{-6})RAD + .339 + error. \end{aligned}$$

The multiple correlation coefficient is 0.92, with a standard error of estimate of 0.016 gm cm^{-3}, for 23 degrees of freedom.

The computed t-values indicated each variable in the in situ study was significant above the 99 percent level (t_b greater than $t_{.01}$). This means there is less than 1 chance in 100 that any variable could be dropped from the equation because it is not significantly different from zero. The t-values associated with the LANDSAT variables indicate significance at varying levels of confidence. Variables other than satellite radiance were significant at the 75 to 95 percent level. The radiance t-value fell slightly below the 60 percentile.

DISCUSSION

An empirical relationship among average snowpack density, albedo, and various other site and storm variables does exist. The average snowpack density can be estimated from albedo measurements when general site and storm information are available. Snow albedo can be measured from aircraft (Bauer and Dutton, 1962). Because of cost, 1 or 2 samples would be measured rather than the traditional 10 points on a snow course. Similarly, Leaf and Kovner (1972) recommend fewer sample points per snow course, 1 or 2 versus 10, and a greater distribution of snow sampling areas within a watershed. Increasing the number of snow courses is difficult and costly when it increases foot travel or telemetered equipment, but is relatively easy to do when the information can be obtained remotely from aircraft or spacecraft.

Equation 1 is mainly a predictive equation, the date and days since snowfall reducing the majority of the sum of squares. Under overcast conditions (equation 2), the albedo term is more significant than during either partly cloudy or clear conditions. On the basis of these equations, it would appear aircraft-borne radiometers should be flown under the diffuse conditions of overcast skies to maximize the contribution of the albedo term.

The results of the satellite data regression (equation 3) were mixed. Although the correlation and standard error were favorable, the satellite radiance term was significant at only the 60 percent level. This low significance could result from several factors. The instantaneous field of view of the MSS is 0.64 ha (1.6 acres). Undoubtedly, the area covered by each pixel integrated snow and non-snow objects (vegetation, shadows) which decreased the radiance from the true snow value. Also, if a concerted effort were made to obtain snow course data on the date of satellite overpass, the error associated with the snow course density interpolation could be removed.

The launch of the next satellite in the LANDSAT-series (LANDSAT-C) will enable satellite data to be obtained with one-quarter the field of view of present LANDSATs. A Return Beam Vidicon camera system onboard this future satellite will have an instantaneous field of view of only 0.16 ha (0.4 acres). This should decrease the possibility of viewing snow and non-snow objects simultaneously.

There is an error associated with the predicted variable in any linear regression technique. The standard error in this study ranged from 0.016 gm cm^{-3} (1.6 percent density) to 0.031 gm cm^{-3} (3.1 percent density). According to Work et al. (1956), the slotted Federal Sampler overmeasures water equivalent by 7 to 12 percent. Although it is difficult to convert errors in snow water equivalent to errors in density because of the influence of depth, it appears that snow density measurements are high by about 0.040 gm cm^{-3}. An additional source of error is present with snow tube data; the Federal Sampler always overestimates density, so the errors are not normally distributed about the true snow density. In contrast, average density estimates based on the discussed equations are better distributed around the true snow density.

CONCLUDING REMARKS

Using albedo and general site and storm information to estimate average snowpack density shows promise as an alternative to establishment of additional conventional snow survey courses in supplying data for water supply forecasting. Errors associated with this statistical method are equivalent to those from the traditional snow survey measurement. Areal water equivalent can be more accurately estimated by this method since measurements are taken in more sections of the watershed.

The results of the in situ study at the Central Sierra Snow

Laboratory suggest the immediate application of aircraft-mounted radiometers to obtain albedo data for use in estimating average snowpack density. The results of the LANDSAT study, however, are not as acceptable for immediate use in the Sierra Nevadas. Satellite data may be applicable, though, in the broad flat regions of the Midwest. Similarly, data from the visible channel of the NOAA-series environmental satellites may be useful in equations for major river basins of the Midwest or Canadian Plains.

The regression equations presented are empirical relationships. Coefficients may need re-evaluation if they are applied to areas where the snow maturation process is different from that near the American River Basin, California.

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TABLE 1

Snow Courses used in LANDSAT Study

Snow Courses	Elevation (m)
Alpha	2320
Echo Summit	2270
Lake Audrain	2230
Darrington	2160
Wrights Lake	2100
Phillips	2070
Lyons Creek	2040
Tamarack Flat	2000

TABLE 2

Linear Regression Coefficients and t-Values
All Cloud Conditions

Variable	Value	t_b	$t_{.01}$
SD	.00323	24.57	2.58
D	.00193	11.03	2.58
R	.0279	9.50	2.58
A^2	-.0756	-7.57	2.58
intercept	.412		

degrees of freedom = 548 $r^2 = 0.79$

standard error of estimate = 0.027 gm cm^{-3}

TABLE 3

Linear Regression Coefficients and t-Values
Clear Conditions (0/10-2/10)

Variable	Value	t_b	$t_{.01}$
SD	.00321	12.02	2.58
D	.00254	7.95	2.58
R	.0257	4.67	2.58
A^2	-.0617	-3.31	2.58
intercept	.401		

degrees of freedom = 218 $r^2 = 0.75$

standard error of estimate = 0.028 gm cm^{-3}

TABLE 4

Linear Regression Coefficients and t-Values
Partly Cloudy Conditions (3/10-7/10)

Variable	Value	t_b	$t_{.01}$
SD	.00330	17.42	2.60
D	.00162	6.32	2.60
R	.0239	5.92	2.60
A^2	-.0412	-2.89	2.60
intercept	.402		

degrees of freedom = 161 $r^2 = 0.84$

standard error of estimate = 0.022 gm cm^{-3}

TABLE 5

Linear Regression Coefficients and t-Values
Overcast Conditions (8/10-10/10)

Variable	Value	t_b	$t_{.01}$
SD	.00325	12.67	2.60
A^2	-.123	-6.63	2.60
R	.0334	5.48	2.60
D	.00145	4.01	2.60
intercept	.436		

degrees of freedom = 159 $r^2 = .79$

standard error of estimate = 0.031 gm cm^{-3}

TABLE 6

Linear Regression Coefficients and t-Values
LANDSAT Study

Variable	Value	t_b		
DEG	1.25×10^{-3}	2.23	$t_{.05}$ =	2.07
SD	2.43×10^{-3}	1.25	$t_{.20}$ =	1.32
E	2.93×10^{-6}	1.13	$t_{.30}$ =	0.86
RAD	-2.96×10^{-6}	0.49	$t_{.40}$ =	0.53
intercept	.339			

degrees of freedom = 23 $r^2 = .92$

standard error of estimate = 0.016 gm cm^{-3}