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ABSTRACT

The albedo of snow for different cloudiness conditions is an important parameter in the Earth's radiation budget analysis and in the study of snowpack thermal conditions. An efficient method is presented for approximate calculation of incident spectral solar flux and snowcover albedo in terms of atmospheric, cloud, and snow parameters. The global flux under partially cloudy skies is expressed in terms of the clear sky flux and a coefficient which models the effect of scattering and absorption by cloud patches and multiple reflections between the cloud base and snowcover. The direct and diffuse components of the clear sky flux are obtained using the spectral flux outside the atmosphere and the spectral transmission coefficients for absorption and scattering by molecules and aerosols. The high reflectivity of snow is considered when calculating the diffuse component. An approximate solution of the radiative transfer equation and parameterized water vapor attenuation coefficients are used for scattering and absorption by clouds. The spectral snow reflectance model consideres specular surface reflection and volumetric multiple scattering. The surface reflection is calculated using a crystal shape dependent bidirectional reflectance distribution function; volumetric multiple scattering is calculated using crystal size dependent approximate solution of the radiative transfer equation. The input parameters of the model are atmospheric precipitable water, ozone content, turbidity, cloud optical thickness, size and shape of ice crystals of snow and surface pressure. The model outputs spectral and integrated solar flux and snow reflectance as a function of solar elevation and fractional cloudcover.

The model is illustrated using representative parameters for the Antarctic coastal regions. The albedo for a clear sky depends inversely on the solar elevation. At high elevations the albedo depends primarily upon the grain size; at low elevation this dependence is on grain size and shape. The gradient of the albedo-elevation curve increases as the grains get larger and faceted. The albedo for a dense overcaste is few percent higher than the clear sky albedo at high elevations. A simple relation between the grain size and the overcaste albedo is obtained. For a set of grain size and shape, the albedo matrices (the albedo as a function of solar elevation and fractional cloudcover) are tabulated.

CONTENTS

THE ALBEDO OF SNOW FOR PARTIALLY CLOUDY SKIES

1. INTRODUCTION

The albedo of snow is an important parameter in hydrology (Rango, 1975 and Colbeck et al. 1979) and climatology (Kukla and Kukla, 1974; Williams, 1975; Kellogg, 1975 and Hahn and Shukla, 1976). The albedo of snow determines the total short wave radiation absorbed and reflected by snow. The melting of snow as a result of solar heating is studied in hydrology for water resources management. Runoff from snow provides greater than 65% of the streamflow in the Western United States. The short wave radiation reflected by snow is needed for the Earth's radiation balance analysis (Curran et. al. 1978). Predictions of global circulation models are sensitive to the extent of snowcover because of its high reflectivity. A model for the albedo of snow can supplement airborne and spacecraft remote sensing observations using visible and near-infrared radiometers.

The albedo of snow is defined as the ratio of reflected to incident solar energy. Calculation of the reflected energy requires using models for the incident spectral flux and the spectral reflectance of snow. Incident spectral flux varies with solar elevation, atmospheric parameters (e.g., precipitable water and turbidity etc.), the cloud characteristics (e.g., thickness and distribution of patches) and the reflectivity of snow. The spectral reflectivity of snow depends upon solar elevation, the grain size and grain shape, and the thickness of the snowcover.

An approximate model for calculating the incident spectral flux for partially cloudy skies is discussed in Section 2. The spectral integrated values (the insolation) are obtained and compared with observations. Section 3 discusses the spectral reflectance model of snow. In Section 4 the albedo values are obtained and compared with observations.

2. PARAMETERIZATION OF INCIDENT SPECTRAL SOLAR FLUX

A partially cloudy sky which includes the limiting cases of a clear or a totally overcast sky, is an inevitable reality of our weather systems. Clouds which are aerosols with diverse water phases, affect the spectral and directional characteristics of incident solar radiation. Clouds of different physical characteristics (e.g. density and thickness), however, affect the incident radiation differently. The incident radiation for a totally cloudy sky is partially diffuse for high altitude thin clouds but is generally totally diffuse for low altitude thick clouds. Also, near-infrared radiation is generally attenuated due to water vapor absorption. In a partially cloudy sky, the distribution of cloud patches is another factor which affects the incident radiation.

In a clear sky the incident radiation consists of direct and atmospherically scattered (diffuse) components. Snow being highly reflective, a significant portion of this radiation is reflected back into space. Clouds over snowfields entrap this radiation, causing multiple reflection between the cloudbase and the snowcover. Because of spectral variation in the reflectivity of snow, this multiple reflection modifies the spectral characteristics of the incident flux.

Accurate calculation of the incident flux is difficult for totally cloudy skies (Dave and Braslau, 1975) and it is significantly more difficult for a partially cloudy sky (McKee and Cox, 1974). In many hydrologic and climatic studies, such accuracy is generally compromised for the sake of simplicity and efficiency by using parameterized equations of observed or accurately calculated insolations in terms of fractional cloudcover (Sivkov, 1971; Lacis and Hansen, 1974). In the spirit of these parameterization equations, this section outlines a model for approximate calculations of the spectral fluxes, from which the insolation values can be obtained. The insolation results are compared with observations. Note that an understanding of the optical properties of snow (reflection, transmission and radiative heating) should be based upon the spectral values.

2-1. Parameterization Equations

For a partially cloudy sky, the incident flux may be written as (Wiscombe, 1975; Schneider and Dickinson, 1976):

$$G_c(a) = \frac{G_o(a)\,[1 - c + c\tau_{cc}(a)]}{1 - cA_{cc}A_s\tau_A^2} \qquad (2\text{-}1)$$

where

$G_o(a)$ = clear sky global flux for solar elevation a

c = fractional cloudcover

$\tau_{c\ell}(a)$ = transmission coefficient of cloud

$A_{c\ell}$ = diffuse reflectance of cloudbase

A_s = diffuse reflectance of snow

τ_A = transmission coefficient between cloudbase and snowcover.

Noting that the clear sky fraction is $(1 - c)$, the numerator of Equation 2-1 represents the sum of fluxes from both clear and cloudy portions of the sky. This numerator is essentially the basis of the Angstrom-Savinov parameterization of observed insolation in terms of fractional cloudcover (Sivkov, 1971). The denominator of this equation represents the enhancement of intensity due to multiple reflection between cloudbase and the snowcover (see Liljequist, 1956; Vowinckel and Orvig, 1962).

For a partially cloudy sky, the instantaneous values of the flux are quite variable due to the dynamical nature of the position of solar disk with respect to the cloud patches. Therefore in data analysis and parameterizations time averaged (hourly, daily, or monthly) flux values are considered (Sivkov, 1971). The fractional cloudcover and the solar elevation are the average values corresponding to this time duration. The clear sky flux (G_o) is the sum of direct and atmospherically scattered components. The direct flux (F_1) is calculated by multiplying the spectral flux outside the atmosphere (F_o) by the parameterized spectral transmission coefficients for absorption by ozone (τ_{o_3}), water vapor (τ_{H_2O}), mixed gases (τ_g) and scattering by molecules (τ_R) and aerosols (τ_a), as outlined by Leckner (1978), from the equation

$$F_1 = F_o \, \tau_{o_3} \, \tau_{H_2O} \, \tau_g \, \tau_R \, \tau_a \sin a \tag{2-2}$$

where a is the solar elevation angle. The diffuse flux incident over snowcover arises from atmospheric scattering of direct flux and atmospheric backscattering of radiation reflected by snow. Because of the high reflectivity of snow, the diffuse flux incident over snowcover is generally larger than over other surfaces for similar atmospheric conditions. The diffuse flux (F_2) is calculated by the method suggested by Robinson (1966) from the equation

$$F_2 = \xi \left[F_o \, \tau_{o_3} \, \tau_{H_2O} \, \tau_g - \frac{F_1}{\sin a} \right] \sin^{4/3} a \qquad (2\text{-}3)$$

where the coefficient ξ depends primarily upon the albedo of snow. For a clean fine-grained snow of high latitudes with albedo generally larger than 0.80 (e.g., Antarctica) this coefficient is

$$\xi = 0.75 \text{ for } a < 10°$$

$$0.70 + 0.005a \text{ for } 10° < a < 40°$$

The sum of fluxes (2-2) and (2-3) gives the clear sky global flux G_o. All transmission coefficients in these fluxes are tabulated by Leckner (1978). The cloud transmission coefficient ($\tau_{c\ell}$) can be written due to scattering by cloud drops (T_s), and absorption by ambient water vapor (T_A) (Feigelson, 1978). For visible and near-infrared radiations, the cloud drops are almost pure scatterer of radiation with strongly asymmetric phase function. The delta-Eddington approximation of the radiative transfer equation (Joseph et al. 1976), which is designed for strongly asymmetric scattering medium, gives the scattering transmission coefficient as

$$T_s = e^{-\tau^*/\mu_o} + \frac{(2 + 3\mu_o)(1 - e^{-\tau^*/\mu_o}) - 3\tau^*(1 - g^*)\, e^{-\tau^*/\mu_o}}{4 + 3(1 - g^*)\tau^*} \qquad (2\text{-}4)$$

where

$\tau^* = (1 - g^2)\,\tau$

$g^* = g/(1 + g)$

$\mu_o = \sin a$

τ = optical thickness of cloud

g = scattering asymmetry factor (=0.8; Feigelson, 1978)

Apart from scattering, a spectrally selective absorption occurs due to ambient water vapor. It is this absorption which produces meteorologically important radiative heating of clouds. In this study the following angular averaged transmission coefficient derived from Leckner (1978) is used:

$$T_A = \exp\left\{ - \frac{0.498 \times k}{(1 + 41.915 \times k)^{0.45}} \right\} \qquad (2\text{-}5)$$

where

X = water vapor path length within cloud

k = spectral absorption coefficient of water vapor (Leckner, 1978).

Although cloud drops are not pure scatterers (Zudnkowski & Strand, 1969; Herman, 1977), absorption by liquid water takes place roughly in the same spectral interval as for the gas. It is thus a good approximation to include the effect of absorption by cloud drops by adjusting the water vapor path length (Lacis and Hansen, 1974). A mean water vapor path length for stratiform clouds is about 0.3 cm (Lacis and Hansen, 1974; Wiscombe, 1975).

During the downward pass through the cloud, the strongly absorbing frequencies of the radiation are depleted by the cloud. As a result, the cloud is relatively transparent for the radiation which will be reflected by snow. For calculating the albedo of cloud ($A_{c\ell}$) for this upwelling diffuse radiation, it is a good approximation to treat the cloud as a pure scattering medium (Lacis and Hansen, 1974). The diffuse reflectance of cloud can be obtained from the scattering transmission coefficient (2-4) as:

$$A_{c\ell} = 1 - 2 \int_0^1 T_s \, \mu_o \, d\mu_o$$

$$\approx \frac{2\left(1 - \frac{2}{\sqrt{3}}\right) e^{-\sqrt{3}\tau^*} + 3(1 - g^*)\,\tau^*}{4 + 3(1 - g^*)\,\tau^*} \tag{2-6}$$

For transmission coefficient between cloudbase and snowcover (τ_A), one should consider aerosols and water vapor attenuations because of their low scale heights. Assuming aerosols to be pure scatterers, its effect will primarily be to diffuse the radiation during the multiple reflection process. Since clouds are effective diffusor of the radiation, further scattering by the aerosol may be neglected in approximate calculations. Therefore, in this model only the selective spectral attenuation by water vapor within the cloudbase and snowcover is considered, i.e., the transmission coefficient (2-5) is substituted for τ_A. The water vapor path length is taken to be one fourth the precipitable water in the clear atmosphere, which is typical for stratiform clouds (Wiscombe, 1975; Schmetz and Raschke, 1979).

Note that in considering the effect of clouds on the incident radiation, the most important parameter is cloud's optical thickness because the reflection and transmission coefficients depend strongly upon it. Comparatively, the effect of moderate variations from the mean values of the water vapor path length within the cloud, and between cloudbase and snowcover is weak.

The reflectance of snow shows strong spectral dependence, and the magnitude at different wavelength vary with the metamorphic changes (Mellor, 1977). A recent study on the diffuse reflectance of snow based on the two-stream approximation of the radiative transfer equation (Choudhury and Chang, 1979) which provided qualitative understanding of these spectral and metamorphic variations, gave the following equation for a deep homogeneous snowcover:

$$A_s = 1 - \frac{2(1-\omega)^{1/2}}{(1-\omega a)^{1/2} + (1-\omega)^{1/2}} \tag{2-7}$$

where

$\omega = \frac{1}{2} + \frac{1}{2}\exp(-1.67\,k_i r)$ = single scattering albedo

$a = 0.87\exp(-2k_i r) + 0.97\,[1 - \exp(-2k_i r)]$ = asymmetry factor

k_i = spectral absorption coefficient of ice (Irvine and Pollack, 1968; Hobbs, 1974)

r = radius of ice grains or the circular area equal to the shadow area of ice grains (Asano et al. 1979).

Above equations constitute the model for the spectral incident flux under partially cloudy skies. The input parameters of the model are the clear sky atmospheric parameters (ozone path length, precipitable water, turbidity, and surface pressure), the cloud optical thickness and the grain size of snow. Illustrative results and comparison with observations follows.

2-2. Illustrative Flux Results

Since the emphasis in this report is on the radiation characteristics over extensive snowfields, the atmospheric parameters representative of Antarctic coastal regions (such stations as Maudheim – 71° 03′ S 10° 56′ W (Liljequist, 1956) and Mirny – 66° 33′ S 93° 01′ E (Rusin, 1964)) was chosen: ozone path length 0.35 cm NTP (Robinson, 1966), Angstrom turbidity 0.017 (Liljequist, 1956), precipitable water 0.25 cm (Nieman, 1977) and the surface pressure 1000 millibar. The ice grain size of snow is chosen to be 0.4 millimeters (Liljequist, 1956). The insolation (spectral integration from 0.3 to 2.5 μm) and the spectral flux values for a clear sky are shown in Figures 2-1 and 2-2 respectively. The observed range of insolation shown in Figure 2-1 is for Mirny during 1956 to 1957 period (Rusin, 1964). The mean values of observed insolation at all solar elevations are same for Mirny and Maudheim. The calculated insolation values differ from mean observations by about 10% at low elevations, and about 6% at high elevations. These differences may be attributed to the inaccuracies in the parameterized transmission coefficients (Leckner, 1978) and in the estimation of atmospheric parameters. The calculated values are definitely within the range of observations.

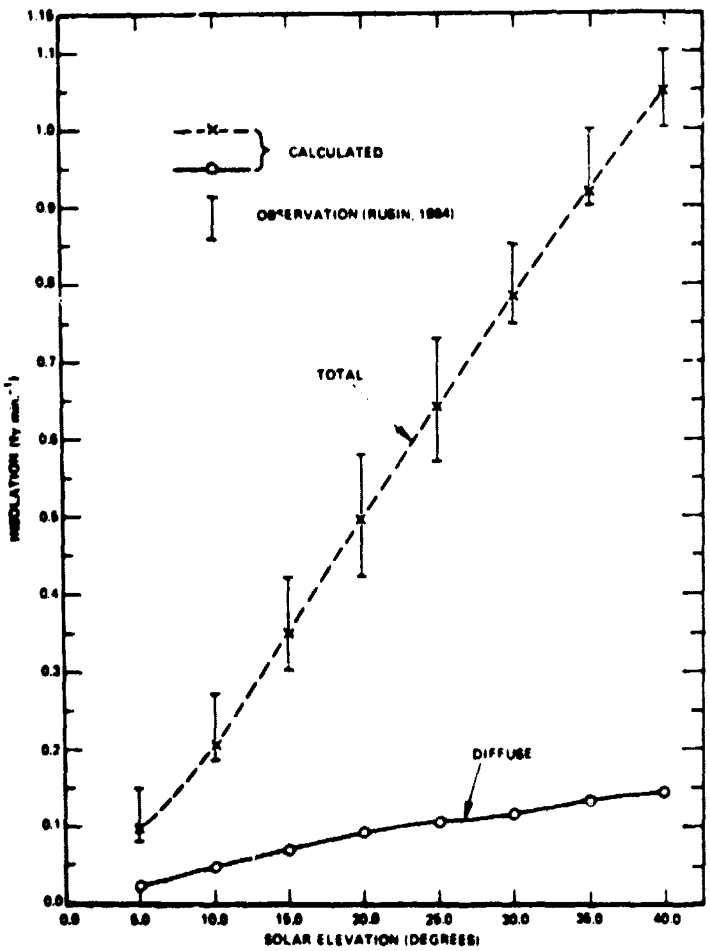

Figure 2-1. Clear Sky Insolation

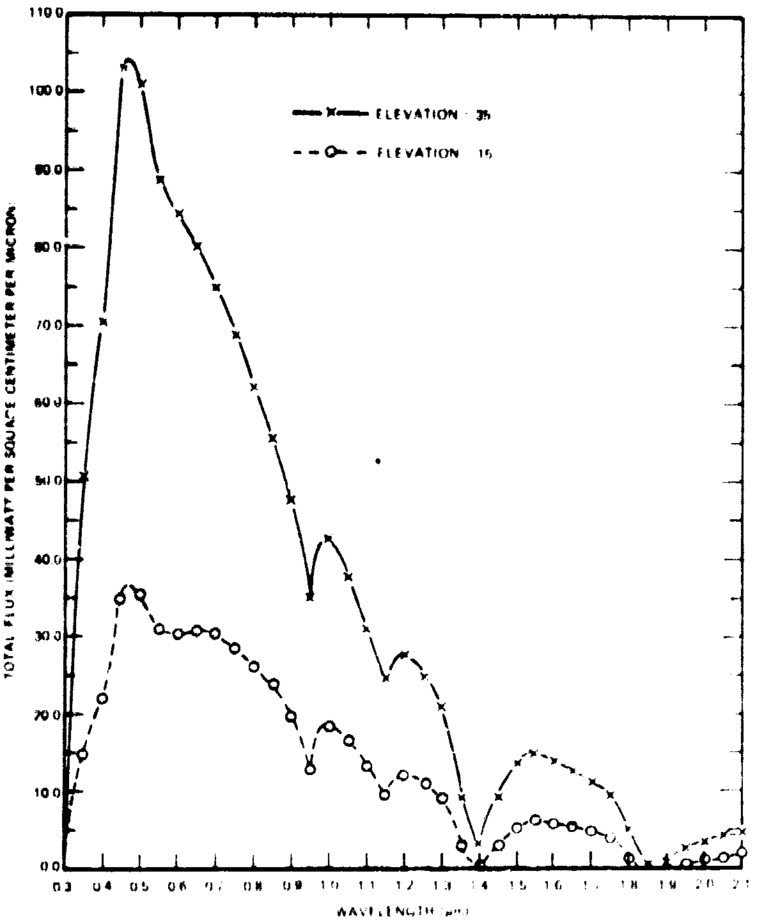

Figure 2-2. Clear Sky Spectral Flux at Different Solar Elevations

For totally cloudy skies (c = 1 in Equation 2-1) the insolation values for different cloud optical thicknesses, together with observations at Mirny (Rusin, 1964) are shown in Figure 2-3. The mean values of the observed insolations are within 5% of the calculated values for the cloud optical thickness of 20. This optical thickness is observed to the mean for stratiform clouds (Feigelson, 1978). The spectral flux values corresponding to this optical thickness is shown in Figure 2-4. A comparison of clear sky and totally cloudy sky spectral flux values, Figures 2-2 and 2-4, clearly illustrates the effect of multiple reflection between the cloudbase and snowcover: the intensity of visible flux diminishes very little compared to near infrared flux. Polar 'Whiteouts' (Catchpole and Moodie, 1971) is attributed to the intense and diffuse radiation which is observed over extensive snowfields under cloudy skies. Table 2-1 gives the ratio of cloudy and clear sky total insolations for varied cloud

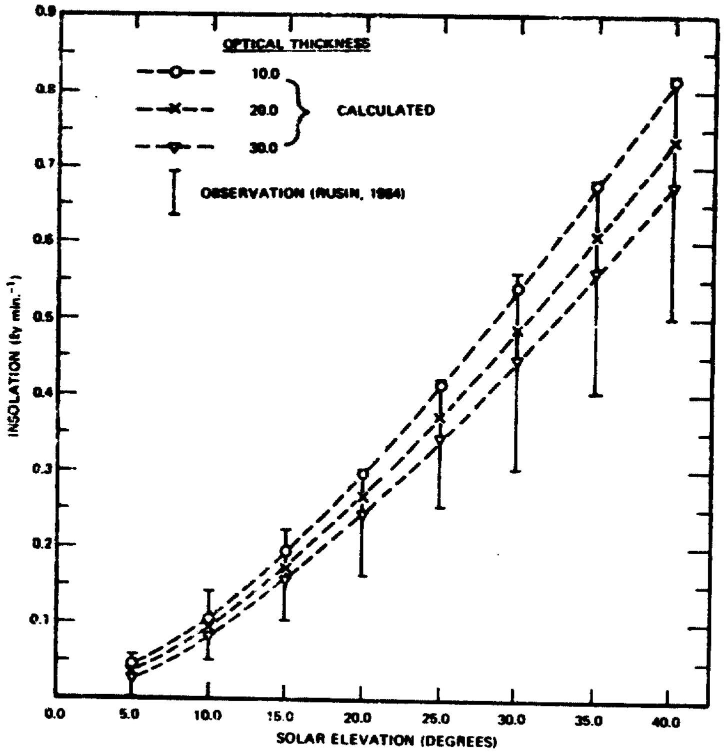

Figure 2-3. Totally Cloudy Sky Insolation

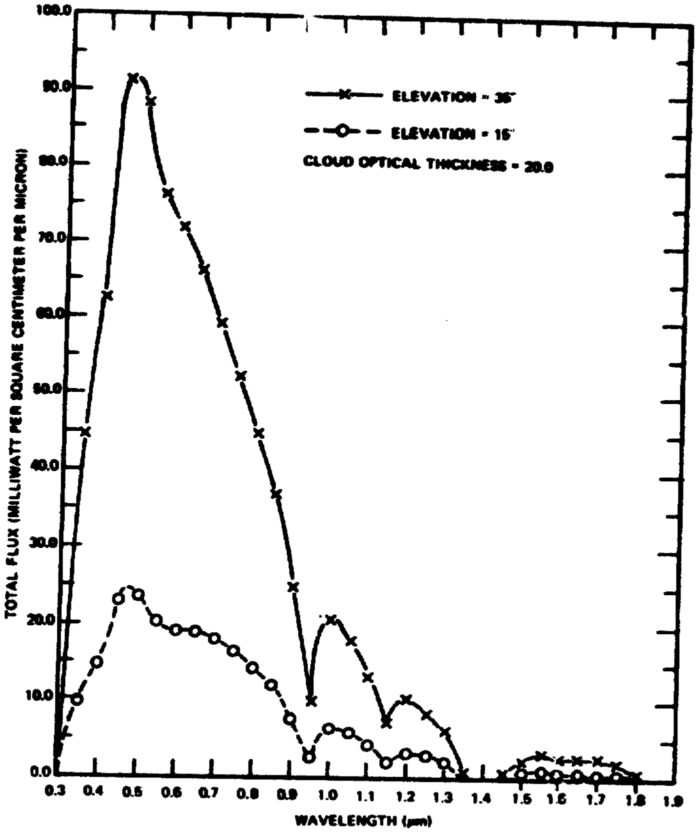

Figure 2-4. Totally Cloudy Sky Spectral Flux at Different Solar Elevations

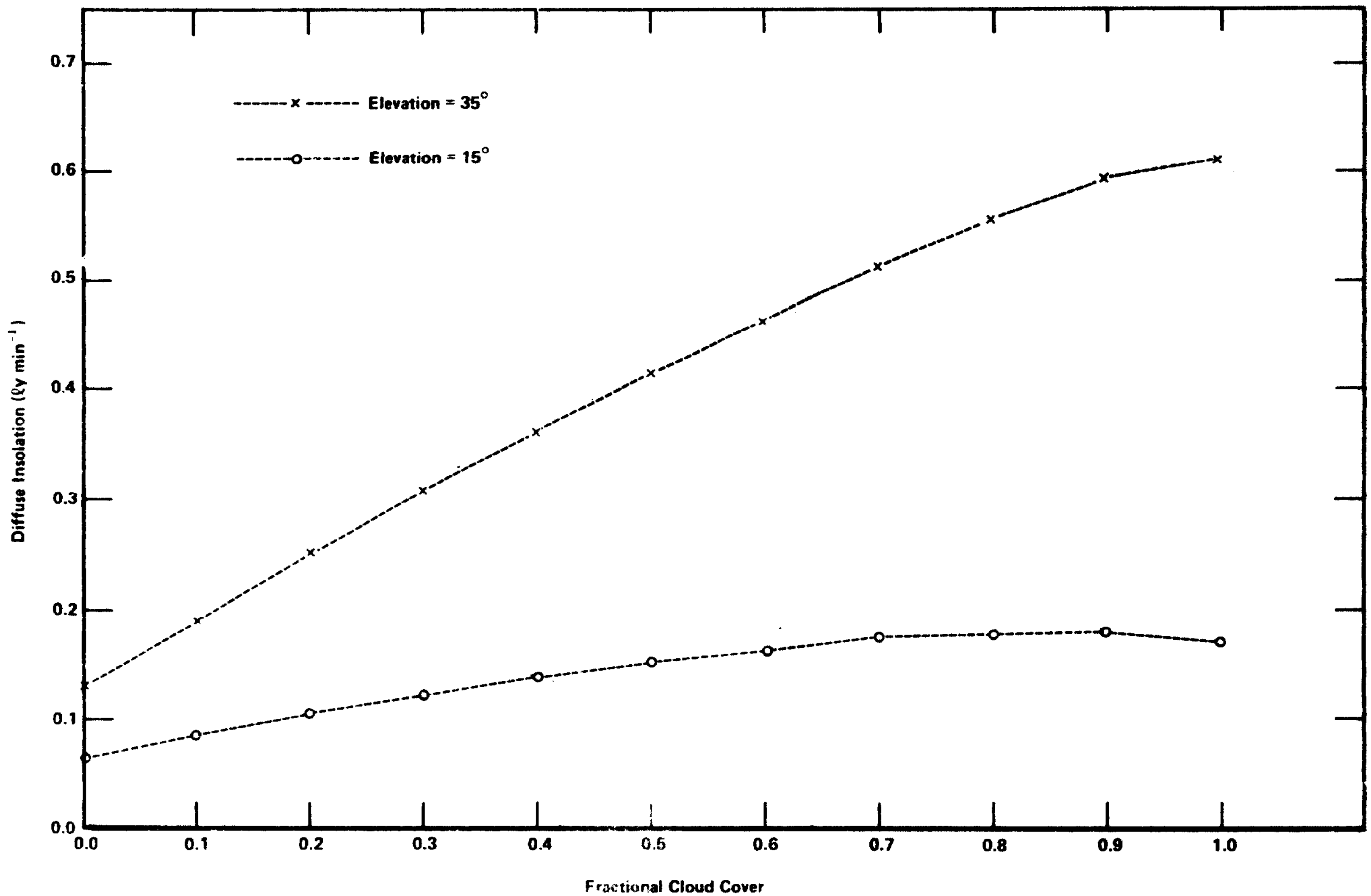

Figure 2-5: The Diffuse Insolation As a Function of Fractional Cloud Cover

Table 2-1: Ratio of Cloudy and Clear Sky Insolations

SOLAR ELEVATION (DEGREES)	FRACTIONAL CLOUD COVER										
	0.0	0.1	0.2	0.3	0.4	0.5	0.6	0.7	0.8	0.9	1.0
10.0	1.000	0.971	0.939	0.904	0.865	0.821	0.771	0.713	0.642	0.552	0.433
15.0	1.000	0.973	0.944	0.912	0.876	0.836	0.791	0.738	0.673	0.593	0.484
20.0	1.000	0.975	0.948	0.918	0.886	0.849	0.808	0.759	0.702	0.629	0.532
25.0	1.000	0.977	0.952	0.924	0.894	0.861	0.823	0.780	0.728	0.663	0.578
30.0	1.000	0.972	0.955	0.930	0.902	0.872	0.838	0.798	0.752	0.695	0.621
35.0	1.000	0.980	0.958	0.935	0.910	0.882	0.851	0.816	0.775	0.725	0.662
40.0	1.000	0.981	0.961	0.940	0.917	0.892	0.864	0.832	0.796	0.753	0.700

fraction and solar elevation. The daily average of this ratio for totally cloudy skies has been used in the parameterization of observed total insolation (Liljequist, 1956). The nonlinear variation of diffuse insolation as a function of fractional cloud cover (Figure 2-5) shows clearly the effect of multiple reflection (Schmetz and Raschke, 1979). Comparison between calculated and observed insolations, shown above, illustrates that reasonable estimates of the input parameters of the model (atmosphere, cloud and snow) can provide good results for the fluxes. Computationally present model is little more expensive than the parameterized equations for the insolation (Sivkov, 1971; Lacis and Hansen, 1974) but it has the flexibility of providing the direct and diffuse spectral fluxes.

The cloud model considered here is simplistic, partly because finite size of the cloud patches which leads to radiation leakage from the sides (McKee and Cox, 1974) is not considered. This refinement, albeit important in any definitive study is difficult to incorporate into a simple model considering the variability of cloud characteristics (shape and size of the cloud patches, and its distribution). Present one parameter cloud model (the parameter being the optical thickness) is approximate but simple to be useful for practical applications e.g. in hydrology where an estimation of average effects of cloudcover may suffice.

3. SPECTRAL REFLECTANCE OF SNOW

The reflectivity of snow like many other natural media such as water and sand depends upon surface and subsurface radiative scattering and absorption. Snow is a highly complex and variable medium. Due to wind action and heat exchange with the atmosphere, the shape and size of ice crystals undergo considerable variation. If melt-freeze process occurs then irregular globular mass of polycrystalline grains as well as surface crust and internal ice lenses can form. A gale apart from compaction will renew the surface with fine grained rounded crystals. Accompanying these changes in the physical characteristics, are the changes in the reflectivity of snow.

Observations with collimated radiation at moderate zenith angles (see Mellor, 1977) show that the reflectance has a characteristic and almost predictable spectral dependence: it is high (generally greater than 90%) for visible radiation; with exception of resonance like structures near 1.1, 1.85 and

2.25 μm the reflectance decreases to a few percent for near infrared radiation. In these observations, although the shape of reflectance curve did not change with time (except for the melt-freeze condition) the magnitude at different wavelengths did change. With the aging of snow, the visible reflectance changed little compared to near infrared reflectance.

For clear sky solar radiation, snow appeared to be a Lambertian reflector at high elevations but showed specular characteristics at low elevations (Knowles Middleton and Mungall, 1952; Dirmhirn and Eaton, 1975). For overcaste skies snow behaved like a Lambertian reflector (Liljequist, 1956).

Following previous studies on the reflectance of snow (Knowles Middleton and Mungall, 1952; Dunkle and Bevans, 1956; Liljequist, 1956; Bergen, 1975), a model for the spectral reflectance of snow was formulated in terms of specular surface reflection and volumetric multiple scattering (Choudhury, 1979c; Choudhury and Chang, 1980). For the surface reflection a bidirectional reflectance distribution function was defined which depended upon the crystal shape and topography. For volumetric multiple scattering solutions of the radiative transfer equation were used which, for a deep homogeneous snowcover, depended upon the crystal size. For clear sky solar radiation, this model for spectral reflectance agreed with observed spectral and angular reflection characteristics. This comparison for albedo will also be shown in the next section.

The spectral reflectance of snow can be calculated from the equation

$$R(\alpha) = \rho(\alpha)\, f(\alpha) + \rho(\alpha)\,[1 - f(\alpha)]\, R_1(\alpha) + [1 - \rho(\alpha)]\, R_2 \tag{3-1}$$

where

$\rho(\alpha)$ = ratio of direct and total spectral fluxes for solar elevation α

$f(\alpha)$ = directional surface reflectance

$$R_1(\alpha) = \frac{\omega^*}{2}\,\frac{2 - 3q\,\mu_o + (1 - \omega^*)\,3a^*\mu_o\,(2\mu_o - q)}{(1 + q)\,(1 - \sigma^2\mu_o^2)}$$

$$\omega^* = \frac{(1 - a^2)\,\omega}{(1 - a^2\,\omega)}$$

$$a^* = a/(1 + a)$$

$$\sigma = [3(1-\omega^*)(1-\omega^* a^*)]^{1/2}$$

$$q = \frac{2\sigma}{3(1-\omega^* a^*)}$$

$$\mu_o = \sin\alpha$$

and the remaining parameters are defined in Equation (2-7), which is used for the diffuse reflectance R_2. The expression for the collimated reflectance R_1, is obtained from the delta-Eddington approximation (Joseph et al. 1976; Sobolev, 1975, Choudhury, 1979a; Wiscombe and Warren, 1979).

The ratio of direct and total fluxes, $\rho(\alpha)$, is calculated from the incident flux model discussed in Section 2. The surface reflectance, $f(\alpha)$, calculated from the bidirectional reflectance distribution of a particulate medium (Trowbridge and Reitz, 1975) is given in Table 3-1 for spherical, oblong and faceted surface crystals. Note that the surface reflectance of Bergen (1975) is defined through Fresnel reflectivity and specific surface, the latter being dependent upon density and grain size.

Table 3-1
The Directional Surface Reflectance for Different Grain Shape (Choudhury, 1979a)

Grain Shape*	Solar Elevation (Degrees)							
	5.0	10.0	15.0	20.0	25.0	30.0	35.0	40.0
Spherical	0.337	0.150	0.089	0.060	0.044	0.034	0.027	0.023
Oblong	0.360	0.204	0.139	0.100	0.073	0.055	0.043	0.034
Faceted	0.660	0.302	0.194	0.129	0.086	0.060	0.044	0.034

*The grain shape is defined through an ellipsoidal representation. The ratio of major and minor axes for 'descriptive' shapes spherical, oblong and faceted grains are, respectively, 1.0, 4.5 and 31.5. The major axis is assumed parallel to the surface. The bidirectional reflectance distribution function is from Trowbridge and Reitz (1975).

Few pertinent features of Equation 3-1 are worth noting. The spectral variation of reflectance arises primarily from the absorption coefficient of ice, which affects the single scattering albedo, ω, and the asymmetry function of scattering, a (Choudhury and Chang, 1979). Changes in the grain size affect the reflectance also through ω and a (see Equation 2-7). The contribution of surface reflectance increases as the solar elevation decreases (see Table 3-1). At high elevations the spectral reflectance and hence the albedo will vary primarily with the grain size. These dependencies will be discussed further in the next section.

Note that equation (3-1) is valid for a deep homogeneous snowcover. The effect of impurities and other internal inhomogeneities on the reflectance cannot be studied from this equation. Caution should be exercised when comparing the theoretical predictions with field observations. The visible reflectance is expected to be affected most by impurities and internal inhomogeneities due to large single scattering albedo and penetration depth. Changes in the surface grain, however, will affect the near infrared reflectance.

4. ILLUSTRATIVE ALBEDO VALUES AND DISCUSSION

The albedo of snow is defined as the ratio of reflected and incident solar energy. From Equations (2-1) and (3-1) one obtains the albedo as:

$$A(a,c) = \frac{\int R(a)\, G_c(a)\, d\lambda}{\int G_c(a)\, d\lambda}$$

where the integration over wavelength (λ) is from 0.3 to 2.5 μm, which encompasses most of the solar spectrum. The atmospheric parameters for the Antarctic coastal region, and typical stratus cloud optical thickness, discussed in Section 2, are used in the calculation of the incident flux. The ice grain radii chosen for the calculation of the spectral reflectance of snow are 0.15 and 0.4 millimeters, which may be considered representative of spring and summer at the coastal regions (Liljequist, 1956). The calculated albedo matrices (the albedo as a function of solar elevation and cloud fraction) are given in Tables 4-1 to 4-4.

The albedo of snow generally shows diurnal variations. At Maudheim, the albedo of new fallen snow on December 2, 1951 was 0.77 which, during 'warm' weather, decreased to 0.73 on December 8 but strong wind on December 11 gave 0.91 (Liljequist, 1956). It was argued that metamorphosis of the snow into larger rounded crystals during the day, and formation of crystals with more or less horizontal surfaces during the night occur. A gale, apart from compaction, renews the surface with fine grained and rounded crystals. At Mirny, the albedo decreased from 0.98 in July 1956 to 0.61 in

Table 4-1: Albedo Matrix for Spherical Surface Grains; Grain Radius = 0.15 millimeter

SOLAR ELEVATION (DEGREES)	FRACTIONAL CLOUD COVER										
	0.0	0.1	0.2	0.3	0.4	0.5	0.6	0.7	0.8	0.9	1.0
10.0	0.842	0.844	0.847	0.850	0.854	0.858	0.862	0.867	0.872	0.877	0.883
15.0	0.833	0.837	0.841	0.845	0.850	0.855	0.861	0.867	0.874	0.882	0.891
20.0	0.826	0.830	0.835	0.841	0.846	0.852	0.859	0.867	0.875	0.884	0.895
25.0	0.819	0.824	0.830	0.836	0.843	0.850	0.857	0.866	0.875	0.885	0.897
30.0	0.813	0.819	0.826	0.832	0.839	0.847	0.856	0.865	0.875	0.886	0.899
35.0	0.808	0.814	0.821	0.829	0.836	0.845	0.854	0.864	0.874	0.886	0.900
40.0	0.803	0.810	0.817	0.825	0.834	0.843	0.852	0.863	0.874	0.887	0.901

Table 4-2: Albedo Matrix for Faceted Surface Grains; Grain Radius = 0.15 millimeter

SOLAR ELEVATION (DEGREES)	FRACTIONAL CLOUD COVER										
	0.0	0.1	0.2	0.3	0.4	0.5	0.6	0.7	0.8	0.9	1.0
10.0	0.868	0.869	0.870	0.871	0.872	0.874	0.875	0.878	0.880	0.882	0.883
15.0	0.851	0.853	0.856	0.859	0.862	0.866	0.870	0.874	0.879	0.885	0.891
20.0	0.838	0.841	0.845	0.850	0.854	0.860	0.865	0.871	0.878	0.886	0.895
25.0	0.827	0.832	0.837	0.842	0.848	0.854	0.861	0.869	0.877	0.886	0.897
30.0	0.818	0.824	0.830	0.836	0.845	0.850	0.858	0.866	0.876	0.887	0.899
35.0	0.811	0.817	0.824	0.831	0.838	0.847	0.855	0.865	0.875	0.887	0.900
40.0	0.805	0.812	0.819	0.827	0.835	0.844	0.853	0.863	0.874	0.887	0.901

Table 4-3: Albedo Matrix for Spherical Surface Grains; Grain Radius = 0.4 millimeter

SOLAR ELEVATION (DEGREES)	FRACTIONAL CLOUD COVER										
	0.0	0.1	0.2	0.3	0.4	0.5	0.6	0.7	0.8	0.9	1.0
10.0	0.797	0.801	0.804	0.808	0.812	0.817	0.822	0.827	0.833	0.840	0.846
15.0	0.787	0.791	0.796	0.801	0.807	0.813	0.820	0.828	0.836	0.846	0.856
20.0	0.778	0.784	0.790	0.796	0.803	0.810	0.818	0.827	0.837	0.848	0.861
25.0	0.771	0.777	0.784	0.791	0.799	0.807	0.816	0.826	0.837	0.850	0.865
30.0	0.764	0.771	0.778	0.786	0.796	0.804	0.814	0.825	0.837	0.851	0.867
35.0	0.757	0.765	0.773	0.782	0.791	0.801	0.812	0.824	0.837	0.852	0.869
40.0	0.752	0.760	0.769	0.778	0.783	0.799	0.810	0.823	0.836	0.852	0.870

Table 4-4: Albedo Matrix for Faceted Surface Grains; Grain Radius = 0.4 millimeter

SOLAR ELEVATION (DEGREES)	FRACTIONAL CLOUD COVER										
	0.0	0.1	0.2	0.3	0.4	0.5	0.6	0.7	0.8	0.9	1.0
10.0	0.831	0.832	0.833	0.834	0.836	0.837	0.839	0.842	0.844	0.846	0.846
15.0	0.810	0.813	0.816	0.819	0.823	0.827	0.832	0.837	0.843	0.850	0.856
20.0	0.793	0.798	0.803	0.808	0.813	0.819	0.826	0.833	0.842	0.851	0.861
25.0	0.780	0.786	0.792	0.798	0.806	0.813	0.821	0.830	0.840	0.851	0.865
30.0	0.770	0.776	0.783	0.791	0.799	0.808	0.817	0.827	0.839	0.852	0.867
35.0	0.761	0.769	0.777	0.785	0.794	0.804	0.814	0.825	0.838	0.852	0.869
40.0	0.754	0.763	0.771	0.780	0.790	0.800	0.811	0.824	0.837	0.852	0.870

January 1957 (Rusin, 1964). A melt-freeze process leading to the formation of névé and ice was argued to be associated with this large change in albedo. Clearly these long-term changes in the albedo is associated with the metamorphism of snow i.e. changes in the crystal size and shape together with surface texture. At Mirny, the mean solar elevation during July is much less than that during January, which also has contributed to the change in albedo, as discussed below.

Apart from long-term diurnal variations the albedo depends upon solar elevation and cloudiness. With respect to solar elevation, the albedo generally decreases as the elevation increases (Liljequist, 1956; Rusin, 1964; Korff and Vonder Haar, 1972), but there is disagreement concerning the gradient of the albedo-elevation curve (see Korff and Vonder Haar, 1972, and Bryazgin and Koptev, 1969). Again due to metamorphic changes, the albedo is sometimes asymmetric with respect to the optimum elevation i.e. morning and afternoon values at the same elevation are not identical (Liljequist, 1956; Dirmhirn and Eaton, 1975). The albedo under cloudy skies seems to depend upon the cloud altitude. For high and middle altitude clouds, the albedo decreases as the solar elevation increases (Rusin, 1964), but for low altitude thick clouds the albedo tends to increase with the solar elevation (Rusin, 1964; Korff and Vonder Haar, 1972). Compared to clear sky albedo, however, the albedo for cloudy skies is generally a few percent higher.

The calculated albedo matrices show the observed daily trend under clear skies and the effect of dense cloudcover. For a clear sky, the solar elevation dependence of the albedo depends upon the grain size and the shape of surface crystals. At high elevations the albedo depends primarily upon the grain size. At low elevations the albedo increases as the grains become faceted. The gradient of albedo-elevation curve for snow which has undergone metamorphism under calm clear weather is expected to be higher than that for a wind blown snow. Associated with this, one may also observe that at high elevations, the albedo of a wind blown snow is higher. Thus the magnitude and the gradient of albedo will depend upon the history of the snowcover.

The spring (November) observations of albedo at Maudheim (Liljequist, 1956) compare well with the calculated values (Figure 4-1). The mean grain size during spring is measured to be about 0.3 millimeters. The observed values are between those predicted for rounded and faceted crystals. In

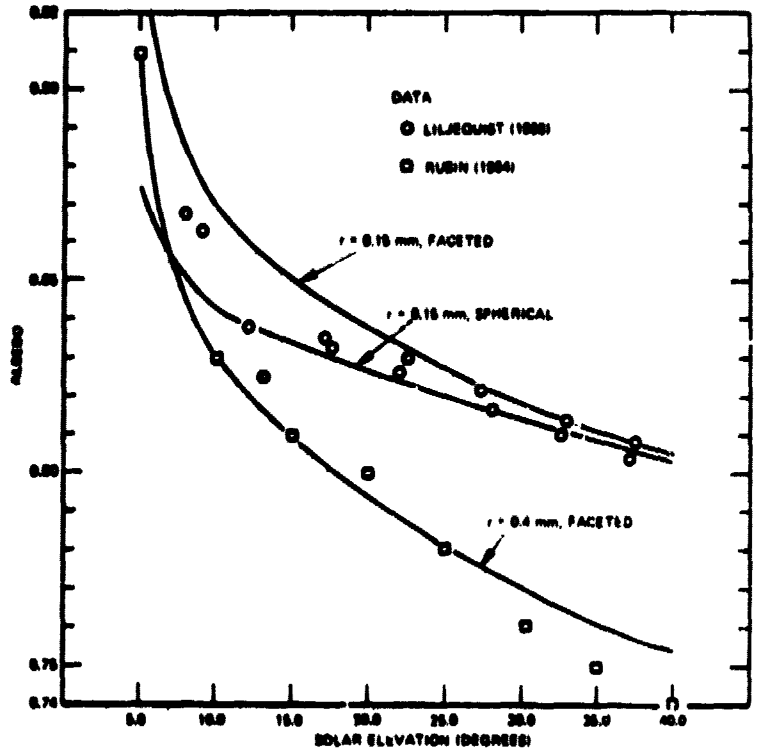

Figure 4-1. Comparison of Observed and Calculated Clear Sky Albedo

comparing the observations for December at Maudheim it was found that the grain size is about 0.4 millimeters. Indeed, the measured grain sizes during spring and summer were respectively 0.2-0.4 mm. and 0.4-0.6 mm (Liljequist, 1956). A comparison with observations at Mirny (Rusin, 1964) is also shown in Figure 4-1. The gradient of the albedo-elevation curve, both observed and calculated, is higher at Mirny compared to that at Maudheim. It is important to note that during observations at Maudheim real thawing was observed only once, but significant metamorphic changes was observed at Mirny. Thus the physical characteristics of snow at these two stations were probably different during the observations.

Barkstrom (1972), and Barkstrom and Querfeld (1975) compared theoretical results with these observations without considering the spectral reflectance and incident flux. Also these models do not consider the diffuse flux, which is of considerable importance at low elevations. Instead of calculating the albedo in terms of measured or expected grain size, an attempt was made to calculate it using other optical measurements (extinction coefficient and flux ratio). That being unsuccessful, theoretical parameters were simply adjusted to show "... the computed albedos agree with observed albedos to within 1%, except at $\mu_o = 0.08$." This approach to albedo is also taken by Wiscombe and Warren (1979).

In this paper the albedo is not obtained by adjusting theoretical parameters, but as the ratio of reflected and incident energy using the spectral reflectance and the incident flux. For totally cloudy skies, the albedo tends to increase with solar elevation and it is a few percent higher than for clear sky values at least at high elevations. Quantitatively, the calculated values are in better agreement with observations at Maudheim ('about 90%,' Liljequist 1956) than that at Mirny (about 91%, Rusin 1964). It was shown by Bohren and Barkstrom (1974) that the spectral reflectance of snow for diffuse incident radiation is proportional to square root of the grain size. Present study gives the following approximate relation between the albedo and the grain size.

$$A(a, c = 1) \approx 0.950 - 0.125 r^{1/2}$$

where r is the grain radius in millimeters. Because of inherent approximations, the formulation of Bohren and Barkstrom (1974) cannot be used to calculate the albedo.

A sense of dissatisfaction has been expressed regarding the ad hoc nature of separation of reflected radiation into surface and volumetrically scattered components (Barkstrom, 1972). The question was posed as to whether the radiative transfer in snow should be treated as a problem similar to that of a plane parallel atmosphere or a water body since for the atmosphere, as opposed to a water body, this separation is not done. It is, however, difficult to identify snow with either of these two media. Although for snow the surface is not a dielectric discontinuity like that of water or ice it is difficult to visualize the 'surface' of atmosphere glittering like a snow surface in clear sunny days. Realizing that this separation is a debatable point, calculations were also done treating snow as a plane parallel atmosphere i.e. $f(a) = 0$ in Equation 3-1 (Figure 4-2). It can be seen that surface reflection affect the albedo at low elevations, and inclusion of this reflection provides better agreement with [illegible]

Since an absolute accuracy of the delta-Eddington approximation for collimated flux [illegible] far, has not yet been established for snow, we compared the results of this approximation with [illegible] delta-two-stream (Schaller, 1979) and hybrid (Meador and Weaver, 1980) approximations. The results are shown in Figure 4-3. Difference between these approximations are quite noticeable for low solar elevations. Whereas the albedos for the delta-Eddington and the delta-two-stream fail to show observed inverse relationship with the solar elevation, the hybrid approximation does show this relation. Note that inclusion of the surface reflection in the delta-Eddington approximation essentially brings out this relation. The hybrid approximation, however, has been found to give inaccurate angular variation of albedo for thick perfect scattering media (see figure 4 in Meador and Weaver, 1980). In the visible region snow is an almost pure scattering medium. It is therefore difficult to assess how truthfully the hybrid approximation is predicting the solar elevation dependence of snow albedo. The qualitative agreement between this approximation and observations of Liljequist (figure 4-3) is to be noted, but without further study of this approximation it may not be used to draw conclusions regarding the surface reflection. By parameter fitting Wiscombe and Warren (1979) erroneously concluded that the delta-Eddington approximation without surface reflection can provide qualitative agreement with the observed solar elevation dependence of snow albedo.

The model of snow considered here is simplistic. The shape and size of ice grains are quite variable and the surface, being in direct contact with air, undergo metamorphism via wind action, vapor exchange and radiative processes. Observations led to stipulations that changes in surface texture and topography affect the albedo (Liljequist, 1956; Rusin, 1964; Hanson, 1960). Predictions from the radiative transfer equation will surely depend upon the shape and size of ice crystals (Cuzzi and Polalck, 1978), and the accuracy of the solution. This study is based upon approximate solutions of the radiative transfer equation. Obviously there is considerable room for further study. A ray tracing approach similar to Wolff (1975) should also be studied. For a better understanding of the reflective process, a careful set of spectral observations for wide elevation range is highly desirable.

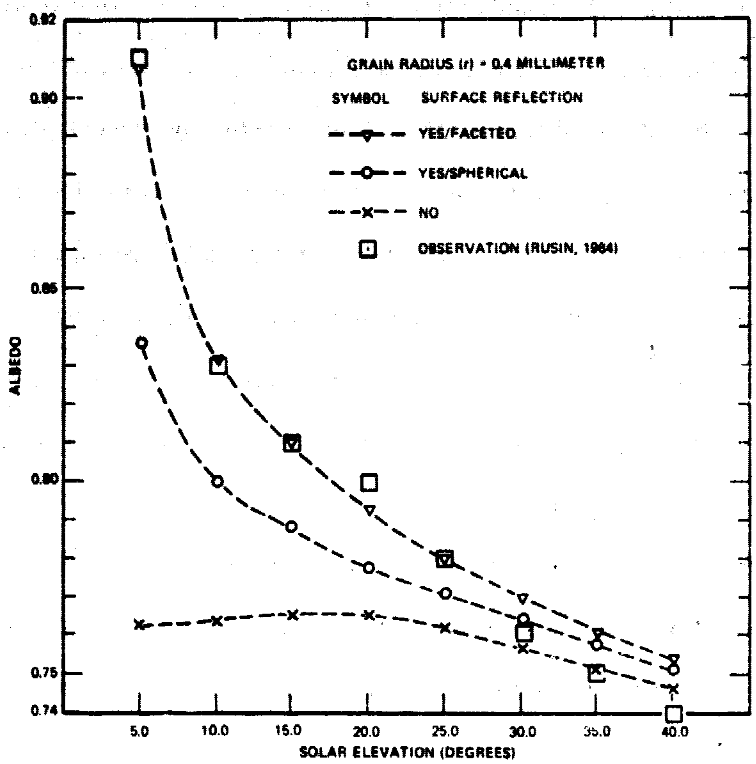

Figure 4-2. Solar Elevation Dependence of Clear Sky Albedo for Different Crystal Shape

5. CONCLUSION

An approximate method for calculating the direct and diffuse spectral solar fluxes over extensive snowfields for partially cloudy skies is discussed. The input parameters for the calculation of these fluxes are atmospheric precipitable water, turbidity, ozone content, surface pressure, the optical thickness of clouds and the grain size of snow crystals. A spectral reflectance model of snow was outlined which requires the grain size and grain shape as input parameters. Illustrative insolation and albedo values are obtained from spectral reflectance and incident flux for representative atmospheric and snow parameters of Antarctic coastal regions. Although there are uncertainties regarding the spectral values due to a lack of experimental results, the spectrally integrated values (insolation and albedo) compare favorably with observations. Unlike previous calculations of the albedo by adjusting theoretical parameters, present study calculates it from the spectral values considering grain

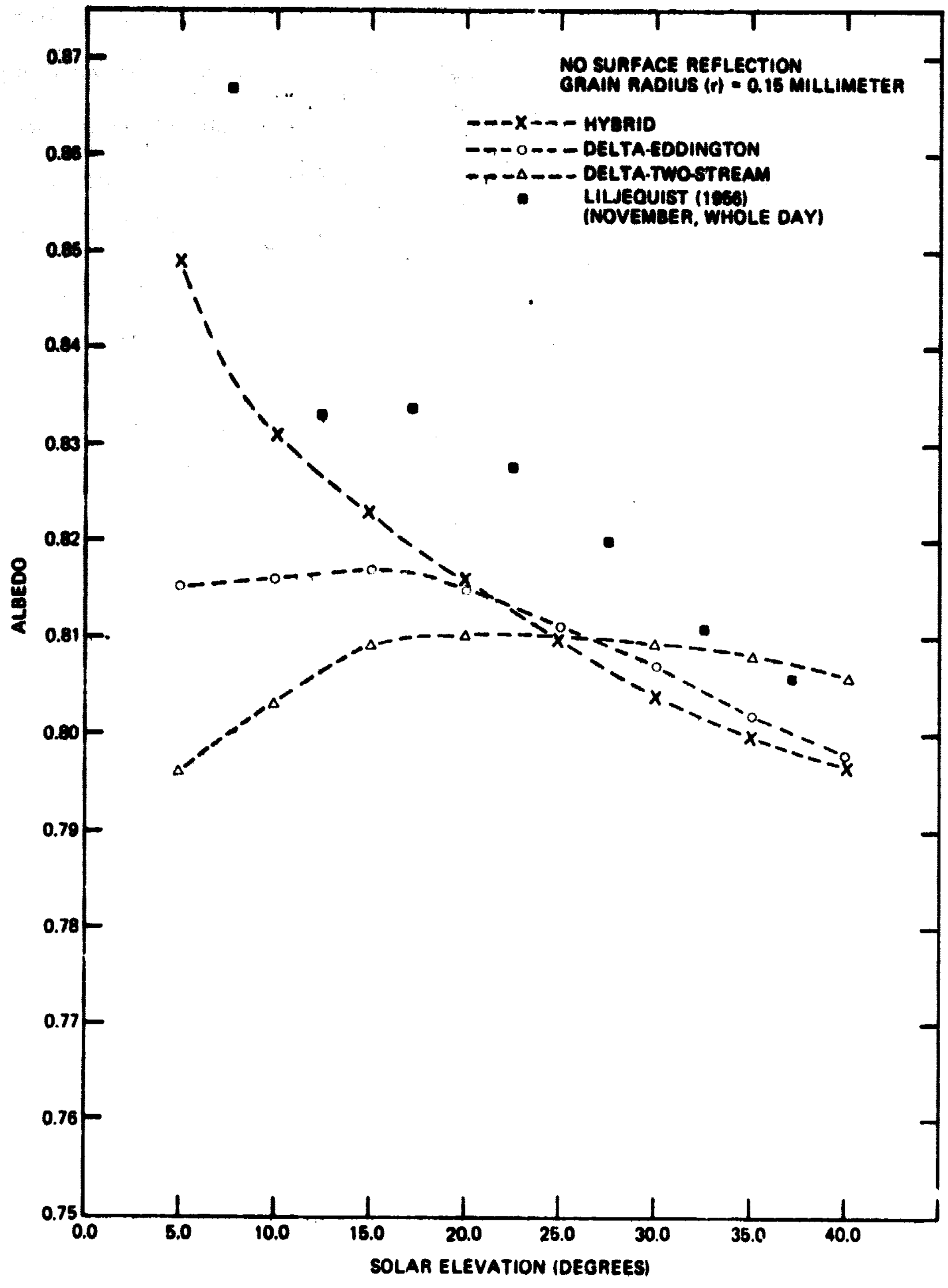

Figure 4-3. A Comparison of Radiative Transfer Solutions for Clear Sky Albedo

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size and shape of ice crystals of snow. This study failed to reproduce the observed solar elevation dependence of the albedo for clear sky conditions when radiative transfer in snow was treated as a problem similar to that of a plane parallel atmosphere. The conclusion and the agreement between theory and observations are, however, tentative because the two-stream and the delta-Eddington approximate solutions of the radiative transfer equation are used in this study. Further study of the radiative transfer equation for snow, and a careful set of spectral reflectance measurement for a wide solar elevation range are highly desirable.

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An abbreviation will appear in the Proceeding of the 37th Eastern Snow Conference.

16. Abstract

The albedo of snow for different cloudiness conditions is an important parameter in the Earth's radiation budget analysis and in the study of snowpack thermal conditions. An efficient method is presented for approximate calculation of incident spectral solar flux and snowcover albedo in terms of atmospheric, cloud, and snow parameters. The global flux under partially cloudy skies is expressed in terms of the clear sky flux and a coefficient which models the effect of scattering and absorption by cloud patches and multiple reflections between the cloud base and snowcover. The direct and diffuse components of the clear sky flux are obtained using the spectral flux outside the atmosphere and the spectral transmission coefficients for absorption and scattering by molecules and aerosols. The high reflectivity of snow is considered when calculating the diffuse component. An approximate solution of the radiative transfer equation and parameterized water vapor attenuation coefficients are used for scattering and absorption by clouds. The spectral snow reflectance model considers specular surface reflection and volumetric multiple scattering. The surface reflection is calculated using a crystal shape dependent bidirectional reflectance distribution function; volumetric multiple scattering is calculated using crystal size dependent approximate solution of the radiative transfer equation. The input parameters of the model are atmospheric precipitable water, ozone content, turbidity, cloud optical thickness, size and shape of ice crystals of snow and surface pressure. The model outputs spectral and integrated solar flux and snow reflectance as a function of solar elevation and fractional cloudcover.

The model is illustrated using representative parameters for the Antarctic coastal regions. The albedo for a clear sky depends inversely on the solar elevation. At high elevations the albedo depends primarily upon the grain size; at low elevation this dependence is on grain size and shape. The gradient of the albedo-elevation curve increases as the grains get larger and faceted. The albedo for a dense overcaste is a few percent higher than the clear sky albedo at high elevations. A simple relation between the grain size and the overcaste albedo is obtained. For a set of grain size and shape, the albedo matrices (the albedo as a function of solar elevation and fractional cloudcover) are tabulated.

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