Numerical Simulation of the Atmospheric Effects on Snow Albedo with a Multiple Scattering Radiative Transfer Model for the Atmosphere-Snow System

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Abstract

The atmospheric effects on the spectral and spectrally integrated snow albedos at the snow surface and top of the atmosphere (TOA) are investigated. A multiple scattering radiative transfer model based on the “doubling and adding” method, combined with the Mie theory is applied to estimate the effects of absorption and scattering by atmospheric molecules, absorptive gases, aerosols and clouds.

It is shown that the spectral surface albedo is reduced by the atmospheric absorptive gases at large solar zenith angles. The solar zenith angle dependence is weakened in the wavelength region shorter than $0.5\ \mu\text{m}$ by the Rayleigh scattering, and at almost all wavelengths by the atmospheric aerosols and cloud cover. H_2O rich atmosphere decreases the spectral surface albedo at large solar zenith angles in the H_2O bands, while the additional reduction of downward solar flux in the near infrared region by H_2O absorption causes the spectrally integrated surface albedo to increase by several percent. Aerosols increase the spectrally integrated surface albedo at small solar zenith angles and reduce it at large solar zenith angles, however they reduce the spectrally integrated planetary albedo, except at large solar zenith angles. Optically-thick cloud cover increases both the spectrally integrated surface and planetary albedos at any solar zenith angle.

In the visible region at small solar zenith angles, the downward solar flux on the snow surface under cloudy sky can exceed that for the clear case, and both further exceed the extraterrestrial solar flux, resulting from the multiple reflection between snow surface and the atmosphere (cloud cover). The global solar radiation on the snow surface under cloudy sky, however, never exceeds that for the clear case and that at TOA.

1. Introduction

The cryosphere plays an important role for the earth energy budget because of its high reflectance, and the large variation of its areal extent. The snow-covered region of the earth's surface, in particular, has a large seasonal variation. Its albedo also varies greatly from that of freshly fallen snow to that of melting snow. Therefore, snow cover may constitute a sensitive mechanism for influencing global climate change. The melting of polar snow and ice by global warming, and the resultant sea level rise are matters currently causing much public anxiety (Warrick *et al.*, 1995). Recently, it is reported that the variation of the snow-covered region in the Eurasian Continent has an influence on annual variation of the East Asian climate (Yasunari *et al.*, 1991; Ose, 1996), and

many Himalayan glaciers are retreating (Higuchi *et al.*, 1980; Yamada *et al.*, 1992). An understanding of the snow radiative process is important for two general applications. The first is the calculation of the radiation budget of snowpack and the planetary radiation budget over snow-covered surfaces. This is important both for hydrology — because radiation is usually the dominant component in the surface energy budget of snow — and for global climate modeling. The second application is for planning the remote sensing of snowpack properties. This requires modeling of the optical properties at a high spectral detail (Warren, 1982). The programs of high spectral resolutional satellite sensors MODIS (Moderate Resolution Imaging Spectroradiometer) and GLI (Global Imager), both having 36 channels in the ultraviolet, visible, near infrared and infrared regions, will provide us the possibilities to estimate snow parameters such as grain size, impurities and layer structure from space. We, therefore, developed

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an accurate and realistic spectral albedo model of snow in the atmosphere-snow system.

Over the past four decades, a considerable number of studies have been made on the modeling of snow surface albedo. The first attempt for calculating the spectral albedo of snow with a multiple scattering radiative transfer in the visible and near infrared regions was done by Dunkle and Bevans (1956), who used the Schuster's two-stream approximation. Giddings and LaChapelle (1961) calculated snow albedo using a diffusion model. These two models were actually equivalent and valid only for diffuse incidence and high albedo (Warren, 1982). They did not start from the single scattering by each snow grain. In the 1970's, many multiple scattering models for snow albedo considering the single scattering by each snow grain have appeared (Barkstrom, 1972; Bohren and Barkstrom, 1974; Barkstrom and Quenfeld, 1975; Berger, 1979; Choudhury and Chang, 1979a,b). Bergen (1970, 1971) calculated the transmittance of snow layer with radiative transfer. However, these models were not accurate enough for the wide range of wavelength, grain size and solar zenith angle. Wiscombe and Warren (1980a) and Warren and Wiscombe (1980) simulated the spectral albedo of snow by the delta-Eddington approximation for multiple scattering, and the Mie theory for single scattering based on realistic physical parameters. They studied the dependence of spectral albedo on solar zenith angle, snow grain size, illuminating conditions and snow impurities, and compared with observations. They also discussed the effect of close packing and nonsphericity of each snow grain. Choudhury and Chang (1981a,b) and Choudhury (1981) developed a two-stream model considering the effects of the atmospheric gaseous absorption and surface reflection.

Various snow studies in the field of dynamics, chemistry and optics were reviewed by Mellor (1977) and detailed optical properties were reviewed by Warren (1982). The latter made an excellent review of snow albedo models and observations. Li (1982) simulated the bi-directional reflection distribution function (BRDF) of snow with the Mie theory and doubling method. This was the first model for BRDF of snow. Chýlek *et al.* (1983) extended the model of Wiscombe and Warren (1980a) and calculated the albedo of soot-contaminated snow. Carroll (1982) and Wendler and Kelly (1988) calculated the effects of snow surface striations and sastrugi on snow albedo. Blanchet and List (1987) examined the effect of anthropogenic aerosols in Arctic haze and snow on the radiation budget, using the atmospheric radiative transfer model with an interactive snow layer based on the delta-Eddington approximation. Warren *et al.* (1990) extended their model for calculations of spectral albedo and emissivity of CO₂ in Martian polar caps. Aoki (1992)

developed a multiple scattering model based on the Mie theory and "doubling and adding" method, applied to the Rayleigh atmosphere-snow system that includes realistic clouds. Aoki *et al.* (1993) simulated the surface albedo of two-layer snow with the same model. Grenfell and Warren (1994) used a two-layer snow model extended from the Wiscombe and Warren (1980a) model and obtained a good agreement of snow albedo with observation under cloudy condition in Antarctica. However, some discrepancies remained at near infrared wavelengths under the clear sky. They quoted three possible reasons for these discrepancies; underestimation of snow grain size, inaccurate correction of the instrument, and a fault in the model. Aoki *et al.* (1998) found that for new snow the optically effective grain size is the order of branch size of dendrites, from the observed spectral albedo at Barrow, Alaska and theoretically calculated albedo with the model used in this study.

The Wiscombe and Warren (1980a) model has been used operationally since the 1980's (Chýlek *et al.*, 1983; Marshall and Warren, 1986). Since this model is based on the delta-Eddington approximation, it is not applicable for large solar zenith angles and cannot be extended to the radiance model. With regard to the atmospheric effects, Wiscombe and Warren (1980b) calculated the spectral albedo at the snow surface and TOA, and the spectrally integrated planetary albedos (the latter was shown in Warren (1982)) with the summer atmospheric condition of the Antarctic Plateau. They reported that the differences in albedo between the snow surface and TOA were due to the Rayleigh scattering and gaseous absorption, and varied with the solar zenith angle. Choudhury and Chang (1981a,b) and Choudhury (1981) calculated the spectral albedo taking account of atmospheric effects by parameterizing the effect of aerosols, cloud cover and atmospheric gases, but the radiative interaction between atmosphere and snow was not incorporated. Blanchet and List (1987) calculated the effect of aerosols on the radiation budget considering this interaction and the atmospheric absorption, but they also used the delta-Eddington approximation.

It is necessary to investigate the atmospheric effects quantitatively with a more accurate multiple scattering model in the atmosphere-snow coupled system for the applications, such as remote sensing of snowpack properties, and the studies of aerosol and cloud radiative forcing over snow surface. It is also important to estimate the variability of snow albedo due to the atmospheric effects for experimental study of snow optical properties in which the observed albedos are compared with those calculated theoretically. The purpose of this study is to investigate the atmospheric effect on spectral and spectrally integrated snow albedos at the snow surface and TOA for the realistic atmospheres containing

absorptive gases, aerosols and cloud cover.

2. Radiative transfer model for the atmosphere-snow system

Basic behavior of the spectral surface albedo of snow has been clarified by many studies mentioned in the preceding section. The surface albedo of snow is essentially determined by a multiple scattering by snow grains, and the radiative interaction between snowpack and atmosphere. The albedo depends not only on the snow physical parameters but also on the external parameters, such as atmospheric conditions and solar zenith angle. The physical parameters can be placed in two groups: those related to single scattering (snow grain size, morphology and impurities), and those related to multiple scattering (snow depth, layer structure, density, water contents, surface condition and impurities). The external parameters contain the atmospheric condition (cloud cover, aerosols, air pressure and atmospheric gases) and solar zenith angle. The atmospheric effects on snow albedo examined in this study belong to the external parameters. The amounts of these effects, however, vary with the physical parameters such as snow grain size. It should be noticed that the effects of external parameters on snow albedo are related to the physical parameters.

In order to consider the radiative interaction between the atmosphere and snowpack (we will use the term "snow" to refer to "snowpack" hereafter), one snow layer is added below the atmospheric layers. The natural snow grains are nonspherical and packed close together. Wiscombe and Warren (1980) discussed these points and noted that a possible adjustment for near-field effects and nonsphericity of snow grains, are judged to produce a few percent reduction of albedo for all wavelengths. In our model, snow grains are assumed to be mutually independent spherical ice particles, and radiative transfer in snow is treated as in the multiple scattering model in the atmosphere containing aerosols or cloud particles. Table 1 shows the processes and parameters used in our radiative transfer model for the atmosphere-snow system. We used the Mie theory for single scattering and the "doubling and adding" method for multiple scattering without polarization. Accuracy of snow albedo at large solar zenith angles is improved by using the "doubling and adding" method compared with the delta-Eddington approximation. Snow grain is so large that the Mie phase function has a very sharp forward peak. This induces a large error in direct calculation of multiple scattering. Aoki *et al.* (1997) examined four kinds of approximations for phase function: Hansen's renormalization (Hansen, 1971); Grant's renormalization (Wiscombe, 1976); delta-M method (Wiscombe, 1977); and truncation method (Hansen, 1969; Potter, 1970). Aoki *et al.* (1997)

found that the operationally useful methods were the Grant's renormalization and truncation method for the phase function of snow particles. In the latter method, the result is not sensitive to the choice of the truncation angle between 5° and 20°. In this study, thus, we use the truncation method for the snow phase function with the truncation angle of 10° and the delta-M method for cloud and aerosol phase functions.

The atmospheric transmittance of gaseous absorption is calculated for H₂O, CO₂, O₂ and O₃. For the first three gases, we used the extended exponential-sum fitting of transmissions (ESFT) developed by Asano and Uchiyama (1987) with the spectral absorption coefficients calculated with the line-by-line algorithm developed by Uchiyama (1992). In ESFT, the number of weight in a sum of exponentials to calculate the transmission function (N) in Eq. (1) of Asano and Uchiyama (1987)) is 10 for each gas. For the atmosphere, three kinds of model atmospheres (Anderson *et al.*, 1986) are used. They are midlatitude winter (MW), subarctic winter (SW), and SW that does not include the lower layer less than 2 km. The last one is adopted to simulate the atmosphere over the Antarctic Plateau and in this study we call it as Antarctic summer (AS). The atmospheric effects on the surface albedo of snow are mainly examined for MW, because the column amounts of air and absorptive gases are richest and the largest atmospheric effects is expected. The precipitable water in MW, SW and AS is 0.86 g/cm², 0.42 g/cm² and 0.19 g/cm², respectively. The effects of aerosols and cloud cover are only examined for SW, because we intend to simulate the Antarctic summer coastal aerosols and high latitudinal middle clouds, respectively.

The radiative transfer and albedo calculations are made at 54 wavelengths (λ) in the range from 0.3 to 3.0 μm with the spectral resolution (bandwidth) of 0.05 μm . Although the wavelength region 0.3–3.0 μm covered in this model seems somewhat narrow for the calculation of the shortwave radiation budget, it is sufficient for the study of the atmospheric effects on snow albedo, because the snow surface albedo at $\lambda > 3.0 \mu\text{m}$ is almost zero and the solar energy that reaches to the surface in $\lambda < 0.3 \mu\text{m}$ and $\lambda > 0.3 \mu\text{m}$ is negligible. In calculating the spectrally integrated albedo and radiant flux, we used the data of extraterrestrial solar radiation compiled by Thekaekara (1974). According to this data, solar energy contained in the wavelength region of the present study is about 96.6 % at TOA. In the "doubling and adding" method, numerical integration over the scattering angle to calculate the radiant flux is conducted using a Gaussian quadrature of 16 points from 0° to 90° in zenith angle (32 streams), shown in Table 1, and thus albedo is only calculated for these 16 values of solar zenith angles

Table 1. Processes or parameters in radiative transfer model for the atmosphere-snow system.

Process or parameter	Method, value and references
Single scattering	Mie theory
Approximation for snow phase function	Truncation method with the truncation angle of 10° (Hansen, 1969; Potter, 1970)
Approximation for aerosol and cloud phase functions	Delta-M method (Wiscombe, 1977)
Multiple scattering	Doubling and adding method
Polarization	No
Atmospheric gaseous absorption	Exponential-sum fitting of transmissions with $N = 10$ (Asano and Uchiyama, 1987)
Atmospheric absorptive gases	H ₂ O, CO ₂ , O ₂ and O ₃
Model atmospheres	Midlatitude Winter (MW), Subarctic Winter (SW) and Antarctic Summer (AS = SW > 2 km), (Anderson <i>et al.</i> , 1986)
Atmospheric layers	Plane parallel, 13 layers for MW and SW, 11 layers for AS
Wavelengths (λ)	0.3–3.0 μm , 54 wavelengths
Wavelength resolution ($\Delta\lambda$)	0.05 μm
Extraterrestrial solar radiation	Thekaekara (1974)
Zenith angle resolution	16 points from 0° to 90° (= 32 streams)
Solar zenith angles (θ_0)	$89.7^\circ, 88.4^\circ, 86.1^\circ, 83.0^\circ, 79.0^\circ, 74.3^\circ, 68.9^\circ, 63.1^\circ, 56.8^\circ, 50.1^\circ, 43.2^\circ, 36.0^\circ, 28.6^\circ, 21.1^\circ, 13.5^\circ, 5.9^\circ$
Snow size distribution	Gamma-type size distribution, $v_{eff} = 0.125$ (Hansen, 1971)
Effective radius of snow grain (r_{eff})	50, 200 and 1000 μm
Snow optical depth (τ_s)	Semi-infinite (= 3200 at $\lambda = 0.5 \mu\text{m}$)
Refractive index of ice	Warren (1984)
Aerosol type	75 % sulfuric acid particles (SA) in troposphere and volcanic ash (VA) in stratosphere
Single scattering coalbedo at $\lambda = 0.5 \mu\text{m}$	8.81×10^{-8} for SA and 4.80×10^{-2} for VA
Size distribution of aerosols	Log-normal for both aerosol types, $v_{eff} = 0.25$ for SA and $v_{eff} = 0.15$ for VA
Effective radius of aerosol particles	0.1 μm for SA and 0.32 μm for VA
Optical depth of aerosols (τ_a)	0.02 (SA), 0.1 (SA+VA) and 0.3 (SA+VA) at $\lambda = 0.5 \mu\text{m}$
Refractive indices of SA and VA	AFGL (1985)
Cloud size distribution	Deirmendjian's cloud model (Deirmendjian, 1964), $v_{eff} = 0.11$
Cloud type	Spherical water droplet
Effective radius of cloud droplets	5 μm
Optical depth of cloud cover (τ_c)	2.5, 5 and 10 at $\lambda = 0.5 \mu\text{m}$
Refractive index of water	Hale and Querry (1973) for $\lambda < 0.7 \mu\text{m}$, Palmer and Williams (1974) for $0.7 \mu\text{m} \leq \lambda < 2.0 \mu\text{m}$, Downing and Williams (1975) for $\lambda \geq 2.0 \mu\text{m}$

(θ_0). Since we use plane parallel layers of the atmosphere and snow, the error in air mass calculation increases for large θ_0 . Kneizys *et al.* (1983) showed that the deviation of secant (*i.e.*, air mass of plane parallel atmosphere) from the spherical and refractive atmosphere, is within 1 % for $\theta_0 \leq 72^\circ$, and this error is kept up to $\theta_0 = 82^\circ$ for H₂O but only up to $\theta_0 = 60^\circ$ for O₃. The error is due mainly to the effect of the earth's curvature, and not atmospheric refraction. As a result, we just discuss the results for $\theta_0 \leq 79.0^\circ$ in this study and show the results at $\theta_0 = 83.0^\circ$ and 86.1° for reference.

We calculate the spectral surface albedo (α^s) and the spectral planetary albedo (α^p) of snow for three values of effective grain radii, $r_{eff} = 50, 200$ and $1000 \mu\text{m}$, following the gamma-type size distribution with effective variance $v_{eff} = 0.125$ as defined by Hansen (1971). According to Wiscombe and Warren (1980a), these three radii correspond to new snow, fine-grained older snow and old snow for the melting point. The snow layer is homogeneous with the optical depth $\tau_s = 3200$ at $\lambda = 0.5 \mu\text{m}$ for all r_{eff} being large enough (semi-infinite) for the calculation of surface albedo to converge at a constant value. For

the complex refractive index of ice, the data compiled by Warren (1984) is used.

The effect of atmospheric aerosols is examined for three values of optical depths, $\tau_a = 0.02$, 0.1 and 0.3, at $\lambda = 0.5 \mu\text{m}$ in SW. The aerosol layer with $\tau_a = 0.02$ is the background level in the Antarctic coast in summer (Shaw, 1982; JMA, 1995) consisting of sulfuric acid (Shaw, 1982; Ito, 1989). The background aerosols are distributed in the troposphere below the height of 5 km for any case of τ_a and follow a log-normal size distribution with the effective radius of $0.1 \mu\text{m}$ (mode radius of $0.057 \mu\text{m}$). The aerosols with $\tau_a = 0.1$ and 0.3 consist of volcanic ash + background aerosols mentioned above. The volcanic ash aerosols are distributed in the stratosphere above the height of 10 km and follow a log-normal size distribution with the effective radius of $0.32 \mu\text{m}$ (mode radius of $0.23 \mu\text{m}$) (Herber *et al.*, 1996). High values of τ_a were observed in the Antarctic coastal stations after major volcanic eruptions: $\tau_a = 0.1$ for El Chichon in 1984 (Shiobara *et al.*, 1987); $\tau_a > 0.3$ for Cerro Hudson in 1991; and, $\tau_a = 0.2$ –0.3 for Mt. Pinatubo in 1992 (Kaneto *et al.*, 1994; Herber *et al.*, 1996). For complex refractive indices of sulfuric acid and volcanic ash, the data compiled by AFGL (1985) is used. In the study of cloud effect, we assumed the high latitudinal middle clouds (Ac and As), because the middle clouds are considered to have possible maximum effects on snow surface albedos in the clouds with frequent appearance in the polar region. Since the annual mean altostratus appears at 2–3 km height in the high latitude (Liou, 1992), the cloud layer is fixed at this height. The water vapor amount is assumed to be saturated in the cloud layer. This increases the precipitable water of SW by 8.5 percent. The effects of cloud cover are examined for three values of optical depths of spherical water clouds, $\tau_c = 2.5$, 5, and 10, at $\lambda = 0.5 \mu\text{m}$ based on the size distribution of Deirmendjian cloud model (Deirmendjian, 1964) with the effective radius $5 \mu\text{m}$. The solar disk cannot be seen through the cloud cover for $\tau_c \geq 5$, when the sun is at zenith direction.

3. Spectral surface albedo

Main characteristics that we notice about the spectral surface albedo of snow under the clear sky are: (1) the contrast between visible and near infrared wavelengths, (2) the snow grain size (r_{eff}) dependence, and (3) the solar zenith angle (θ_0) dependence, as shown in Fig. 1. In Fig. 1a the spectral surface albedos are shown not only for the three r_{eff} discussed in Section 2 but also for the other four values of r_{eff} , to see more details of the dependence of surface albedo on snow grain size. It can be seen that the surface albedo is high in the visible region, and low in the near infrared region. This is due mainly to the fact that the imaginary part of the

Fig. 1. (a) Spectral surface albedo of snow for various snow grain sizes r_{eff} , $\theta_0 = 63.1^\circ$ under clear atmosphere of MW. (b) Solar zenith angle dependence of the monochromatic surface albedo of snow under clear atmosphere (dashed lines) and cloudy case (solid lines) of MW at $\lambda = 1.0 \mu\text{m}$ for $r_{\text{eff}} = 50$, 200 and $1000 \mu\text{m}$.

refractive index of ice is low (less absorptive) in the visible region, and high (much absorptive) in the near infrared region. The surface albedo decreases with the increase of r_{eff} . This is significant especially in the near infrared region. The reason is that the single scattering albedo of a large snow grain is lower than that of a small grain, and its difference is large in the near infrared region. The monochromatic surface albedo at large θ_0 is higher than that at small θ_0 for clear sky as shown in Fig. 1b. The θ_0 dependence of surface albedo for clear sky is related to the direct solar beam and asymmetric phase function of the snow grain. Warren (1982) explained as follows: A photon on average undergoes its first scattering event closer the surface if it entered the snow at a grazing angle. If the scattering event sends it in an upward direction, its chance of escaping the snow without being absorbed is greater than it would be if it were scattered from deeper in the snow. This would be observed even if the ice par-

ticles scattered light equally in all directions. But the phenomenon is greatly enhanced by the extreme asymmetry of scattering function, whereby scattering within a few degrees of the forward direction is much more probable than scattering to other angles.

The snow surface is illuminated not only by the direct solar radiation but also the diffuse one. The θ_0 dependence of surface albedo is also related to these factors. In the case of no atmosphere the snow surface is illuminated only by the direct beam, and the albedo is higher for large θ_0 than that for small θ_0 . However, it is modified by the diffuse component under the atmosphere. In the clear sky for small θ_0 the surface albedo is increased by the additional contribution of diffuse components from larger zenith angles, and vice versa for large θ_0 . In the overcast case the direct solar beam is weakened or disappeared, and snow surface is illuminated mainly by diffuse radiation, and the surface albedo becomes constant — being independent of θ_0 as shown in the cloudy case of Fig. 1b. In Fig. 1b the value of θ_0 at which the cloudy sky albedo crosses the clear sky albedo is about 48° . Warren (1982) estimated this value as 50° from the albedo of snow surface illuminated by pure diffuse radiation. In the following subsections, we examine the effects of Rayleigh scattering, gaseous absorption, aerosols and cloud cover on the spectral surface albedo of snow via the theoretical calculations.

3.1 Effects of Rayleigh scattering and absorptive gases

Figure 2a shows the transmission function of gaseous absorption bands and Rayleigh scattering calculated by ESFT for direct solar radiation at the surface, where the values of $\theta_0 = 79.0^\circ$ have been used to clearly show the weak absorption bands. The fractions of diffuse and direct components in the downward solar flux vary depending on the gaseous absorption and Rayleigh scattering. Figure 2b shows the fraction of the diffuse component in the downward solar flux. The reason that the fraction of diffuse component is generally high at the shorter wavelengths, is the dominance of Rayleigh scattering. In addition to this, the diffuse component increases significantly at large θ_0 in the strong gaseous absorption bands.

In order to examine the clear atmospheric effects on the spectral surface albedo of snow, calculations have been made for the atmosphere of MW containing Rayleigh scattering and absorption by H_2O , CO_2 , O_2 and O_3 (we will denote this albedo by α_{4g}^s) and for no atmosphere (α_{na}^s). Figure 3 shows the difference between α_{4g}^s and α_{na}^s for three values of r_{eff} . The difference $|\alpha_{4g}^s - \alpha_{na}^s|$ is large in the region of strong Rayleigh scattering and strong gaseous absorption bands. In the region $\lambda < 0.5 \mu\text{m}$, the effects of Rayleigh scattering and O_3 absorption are

Fig. 2. (a) Spectral transmission function of the atmospheric absorption, and Rayleigh scattering at the surface for direct solar radiation of $\theta_0 = 79.0^\circ$ and clear atmosphere of MW. (b) Spectral distribution of the fraction of diffuse component in the downward solar flux for clear atmosphere of MW on the snow surface with $r_{eff} = 200 \mu\text{m}$. They are shown for fourteen solar zenith angles from 5.9° to 86.1° shown in Table 1. The values for $5.9^\circ \leq \theta_0 \leq 79.0^\circ$ are indicated by solid lines, and $\theta_0 = 83.0^\circ$ and 86.1° by dashed lines.

dominant, where it is found $\alpha_{4g}^s < \alpha_{na}^s$ at large θ_0 and $\alpha_{4g}^s > \alpha_{na}^s$ at small θ_0 . In this wavelength region, the fraction of diffuse component in the downward solar flux is large even at small θ_0 due to the Rayleigh scattering. This makes the θ_0 dependence of surface albedo faint (Fig. 4a). We also see large values of $|\alpha_{4g}^s - \alpha_{na}^s|$ for large r_{eff} in the same region. This is because the θ_0 dependence of α_{na}^s is significant for large r_{eff} .

Figure 3 also shows that there are large difference $|\alpha_{4g}^s - \alpha_{na}^s|$ in the three wavelength regions of strong gaseous absorption at large θ_0 . For example, the difference $|\alpha_{4g}^s - \alpha_{na}^s|$ increases significantly for $\theta_0 > 79^\circ$ at $\lambda = 1.375 \mu\text{m}$ (Fig. 4b). In the absorption bands the fraction of the diffuse component in downward solar flux becomes larger at large θ_0 (see

Fig. 3. Difference between the spectral surface albedo of snow for clear atmosphere of MW containing H_2O , CO_2 , O_2 and O_2 (α_{4g}^s) and that for no atmosphere (α_{na}^s) for (a) $r_{eff} = 50 \mu\text{m}$, (b) $r_{eff} = 200 \mu\text{m}$ and (c) $r_{eff} = 1000 \mu\text{m}$. The definitions of the solid line and dashed line are the same as Fig. 2.

Fig. 2b) since the direct solar beam suffers strong absorption along the long path compared with diffuse radiation. This reduces the value of α_{4g}^s small. On the contrary, at small θ_0 the diffuse fraction rather becomes small and makes α_{4g}^s close to α_{na}^s . Thus, the value of α_{4g}^s is lower than that of α_{na}^s only at large θ_0 in the gaseous absorption bands.

Since the calculation of radiative transfer is time

Fig. 4. Solar zenith angle dependence of the monochromatic surface albedo of snow with three values of r_{eff} for clear atmosphere of MW (solid line) and that for no atmosphere (dashed line) at (a) $\lambda = 0.325 \mu\text{m}$ and (b) $\lambda = 1.375 \mu\text{m}$.

consuming, it is worth noting the possibility to omit the absorptive gases and Rayleigh scattering in calculating the spectral surface albedo. Restricting our discussion to the average transmittance for $\Delta\lambda = 0.05 \mu\text{m}$, the absorption by the atmospheric gases can be omitted for all r_{eff} for the atmosphere of MW for $\theta_0 \leq 63.1^\circ$ within the error of 0.01. However, it should be remembered that this conclusion could be changed for different value of $\Delta\lambda$, since average transmittance varies greatly with band width.

3.2 Effects of aerosols and cloud cover

The fractions of diffuse component in downward solar flux are shown in Fig. 5a for the atmosphere containing aerosols and Fig. 5b for cloudy sky. It can be seen that aerosols and cloud cover increase the diffuse component in downward solar flux by comparing these figures with Fig. 2, where the diffuse component is shown for the clear sky of MW (basic behavior is same for SW). These figures will help to understand the behavior of the spectral surface albedo with aerosols or clouds shown in the following. Figure 6a shows the monochromatic surface albedos as a function of θ_0 for the atmosphere con-

Fig. 5. (a) Same as Fig. 2b, but for the atmosphere of SW containing aerosols with $\tau_a = 0.3$. (b) Same as Fig. 2b, but for the cloudy atmosphere of SW with three kinds of τ_c . Curves are successively displaced upward by 0.1 except for $\tau_c = 2.5$.

taining aerosols (α_{4g+a}^s), and for the aerosol free case (α_{4g}^s). The θ_0 dependence of α_{4g+a}^s becomes weak with the increase of the optical depth of aerosols τ_a . The difference between α_{4g+a}^s and α_{4g}^s is larger for large θ_0 than that for small θ_0 . Since the θ_0 dependence of each albedo varies with the wavelength, the difference between α_{4g+a}^s and α_{4g}^s also depends on the wavelength. Figure 6b shows $\alpha_{4g+a}^s - \alpha_{4g}^s$ as a function of wavelength at $\theta_0 = 5.9^\circ$ and 79.0° . The effect of background aerosols ($\tau_a = 0.02$) on the surface albedo is very small. However, the difference $|\alpha_{4g+a}^s - \alpha_{4g}^s|$ rises up to 0.14 in the absorption band ($\lambda = 1.375 \mu\text{m}$) for $\tau_a = 0.3$ and $\theta_0 = 79.0^\circ$. This is because the aerosols increase the fraction of diffuse component at large θ_0 in the absorption band as shown in Fig. 5a. At small θ_0 the diffuse component is rather reduced in the absorption bands compared with the non-absorbing region and results in very small values of $|\alpha_{4g+a}^s - \alpha_{4g}^s|$.

Figure 7a shows the monochromatic surface albedos of snow as a function of θ_0 for the cloudy atmosphere (α_{4g+c}^s) and clear case (α_{4g}^s). The θ_0 depen-

Fig. 6. (a) Monochromatic surface albedo of snow with $r_{\text{eff}} = 200 \mu\text{m}$ as a function of θ_0 for the atmosphere of SW containing aerosols (α_{4g+a}^s) with three kinds of τ_a and aerosol free case (α_{4g}^s) at $\lambda = 1.0 \mu\text{m}$. (b) Difference between α_{4g+a}^s and α_{4g}^s as a function of wavelength at $\theta_0 = 5.9^\circ$ (dashed line) and 79.0° (solid line).

dence of α_{4g+c}^s is completely lost for $\tau_c = 10$, where snow surface is illuminated only by diffuse radiation (see Fig. 5b). Figure 7b shows the spectral variation of $\alpha_{4g+c}^s - \alpha_{4g}^s$ at $\theta_0 = 5.9^\circ$ and 79.0° . The value of $|\alpha_{4g+c}^s - \alpha_{4g}^s|$ is small around $\lambda = 0.5, 1.5, 2.0$ and $2.8 \mu\text{m}$, where α_{4g}^s is close to unity or zero (see Fig. 1a). When the value of α_{4g}^s is close to unity, the light absorption by ice is so weak that a large part of the incident photons would escape from the snow surface almost independent of the incident angle. In the case of α_{4g}^s close to zero, the light absorption by ice is so strong that a large part of the incident photons would be absorbed in snow, almost independent of incident angle. Thus, the θ_0 dependence of surface albedo is small for clear sky in such wavelength regions, and the value of $|\alpha_{4g+c}^s - \alpha_{4g}^s|$ becomes small. Except for such wavelength regions, it can be seen that the effect of cloud cover on spectral surface albedo is quite large even for $\tau_c = 2.5$ at any value of θ_0 . This is because the fraction of

Fig. 7. (a) Same as Fig. 6a, but for the cloudy atmosphere of SW (α_{4g+c}^s) with three kinds of τ_c and clear case (α_{4g}^s). (b) Same as Fig. 6b, but for the cloudy atmosphere ($\alpha_{4g+c}^s - \alpha_{4g}^s$).

diffuse component is large enough at any value of θ_0 in the cloudy condition due to the large optical depth as shown in Fig. 5b.

3.3 Effect of difference in atmospheric type

We have seen the effects of clear atmosphere on surface albedo for MW in Subsection 3.1. The atmosphere, however, varies with place and time. In this subsection we examine the dependence of surface albedo on the atmospheric type. In three model atmospheres — MW, SW and AS — the column amounts of air and H_2O , are largest for MW, less for SW and smallest for AS. For SW and MW, the column amounts of gases, except H_2O , are almost the same. Hence, the atmospheric effect is generally largest for MW and smallest for AS. We calculated the difference in spectral surface albedos between MW (α_{MW}^s) and AS (α_{AS}^s) to examine the maximum effect of difference in the atmospheric type on the surface albedo.

Figure 8 shows the spectral variation of $\alpha_{MW}^s - \alpha_{AS}^s$. The value of α_{MW}^s is lower than that of α_{AS}^s in the gaseous (H_2O) absorption bands due to the reason shown in Subsection 3.1 for the case of $\alpha_{4g}^s - \alpha_{na}^s$.

Fig. 8. Difference in the spectral surface albedo of snow with $r_{eff} = 200 \mu\text{m}$ under the clear atmosphere between MW (α_{MW}^s) and AS (α_{AS}^s). The values for $5.9^\circ \leq \theta_0 \leq 79.0^\circ$ are indicated by solid lines on the left ordinate, and $\theta_0 = 83.0^\circ$ and 86.1° by dashed lines on the right ordinate.

At large θ_0 in the H_2O bands the fraction of diffuse component increases with H_2O amount. As a result the spectral surface albedo for H_2O rich atmosphere (α_{MW}^s) is lower than that for H_2O poor atmosphere (α_{AS}^s) at large θ_0 only in the H_2O absorption bands. The calculations were also done for snow with $r_{eff} = 50$ and $1000 \mu\text{m}$ and these results showed that the effect of difference in the atmospheric type on the surface albedo is very small at $\theta_0 \leq 63.1^\circ$, and α_{MW}^s agrees with α_{AS}^s within 0.01. For $\theta_0 > 63.1^\circ$, this effect gradually increases with θ_0 only in the H_2O absorption bands.

4. Spectral planetary albedo

4.1 Effect of the difference in atmospheric type

The effects of the gaseous absorption and Rayleigh scattering are too large to be neglected for the spectral planetary albedo. Wiscombe and Warren (1980b) calculated them over the snow surface in the Antarctic Plateau using the atmospheric radiative transfer model coupled with the snow albedo model of Wiscombe and Warren (1980a). They examined the atmospheric effects on the planetary albedo for the atmospheres in January and October, and pointed out that the planetary albedo is very slightly higher than the surface albedo only in the region $0.35 \leq \lambda \leq 0.43 \mu\text{m}$, due to the Rayleigh scattering. The reduction of the planetary albedo in the O_3 and H_2O absorption bands is more pronounced for October (large θ_0 and long slant path) than for January (midsummer).

We calculated the spectral planetary albedo (α_{4g}^p) over the snow for three kinds of model atmospheres (Fig. 9). It is easy to understand the atmospheric effects on α_{4g}^p by comparing these results with the

Fig. 9. Spectral planetary albedo for the atmosphere of MW, SW and AS at $\theta_0 = 63.1^\circ$ over the snow with (a) $r_{eff} = 50 \mu\text{m}$, (b) $r_{eff} = 200 \mu\text{m}$ and (c) $r_{eff} = 1000 \mu\text{m}$.

spectral surface albedo (α_{4g}^s) shown in Fig. 1a. The gaseous constituent that reduces the planetary albedo in ultraviolet and visible regions is O_3 . The degrees of these reductions are almost independent of the model atmospheres because amounts of O_3 are roughly same in three model atmospheres studied here. The value of α_{4g}^p around $\lambda = 0.4 \mu\text{m}$ is slightly higher than that of α_{4g}^s due to the Rayleigh scattering as noted by Wiscombe and Warren (1980b). In the near infrared region, the differences be-

Fig. 10. (a) Spectral planetary albedos for the atmosphere of SW containing aerosols (α_{4g+a}^p) with three kinds of τ_a and aerosol free case (α_{4g}^p) at $\theta_0 = 63.1^\circ$ over the snow with $r_{eff} = 200 \mu\text{m}$. (b) Difference between α_{4g+a}^p and α_{4g}^p as a function of wavelength at $\theta_0 = 5.9^\circ$ (left ordinate) and 79.0° (right ordinate).

tween model atmospheres become significant in the H_2O absorption bands, where the spectral planetary albedo for MW (α_{MW}^p) is lower than that for AS (α_{AS}^p). The reduction of planetary albedo by gaseous absorption is large for small r_{eff} . This is because the value of α_{4g}^s for snow with small r_{eff} is higher than those with larger r_{eff} . As an example, since the value of α_{4g}^s for $r_{eff} = 1000 \mu\text{m}$ is originally close to zero at $\lambda > 1.5 \mu\text{m}$ as shown in Fig. 1a, the apparent influence by gaseous absorption is very small. However, since the value of α_{4g}^s for $r_{eff} = 50 \mu\text{m}$ is in the range 0.05–0.4 for $1.5 < \lambda < 2.6 \mu\text{m}$, it can be reduced by a significant amount due to gaseous absorption.

4.2 Effects of aerosols

The effect of aerosols on the planetary albedo varies with the geometric condition such as θ_0 , and the properties of aerosols such as absorptivity, particle size and vertical profile. In this study, we assumed the sulfuric acid aerosols in the troposphere

were the background aerosols and the volcanic ash aerosols were in the stratosphere. Figure 10a shows the spectral planetary albedo for the atmosphere containing aerosols (α_{4g+a}^p), and the aerosol free case (α_{4g}^p). There is no significant change in the planetary albedo caused by the background aerosols, but the volcanic ash reduces the planetary albedo in the shorter wavelength region, and increases it in the longer wavelength region. To see the details of the effect of aerosols, the differences between α_{4g+a}^p and α_{4g}^p are calculated for $\theta_0 = 5.9^\circ$ and 79.0° (Fig. 10b). The effect of aerosols is very small for $\tau_a = 0.02$ at both values of $\theta_0 = 5.9$ and 79.0° . In the case of $\tau_a \geq 0.1$, the value of α_{4g+a}^p is higher than that of α_{4g}^p in the longer wavelength region. In the region $\lambda > 1.4 \mu\text{m}$ the single scattering albedo of volcanic ash particle is higher than that of snow grain. This is the reason for $\alpha_{4g+a}^p > \alpha_{4g}^p$ for $\lambda > 1.4 \mu\text{m}$. In the region $\lambda \leq 1.4 \mu\text{m}$, if there is no absorption by the atmospheric gases, it would be $\alpha_{4g+a}^p < \alpha_{4g}^p$ due to the difference in single scattering albedo between the volcanic ash and snow grain. Therefore, in the wavelength region with weak absorption or without absorption, the value of α_{4g+a}^p is lower than that of α_{4g}^p . In the wavelength regions with gaseous absorption (including the weak absorption bands for $\theta_0 = 79.0^\circ$, because the atmospheric absorption is intensified along the long slant path), the planetary albedo is reduced by the atmospheric absorption for the aerosol free case. Since the volcanic ash distributes above most of the absorptive gases, the direct solar beam is scattered upward by the volcanic ash before suffering the atmospheric absorption. This is the reason of $\alpha_{4g+a}^p > \alpha_{4g}^p$ in the wavelength regions with gaseous absorption of $\lambda \leq 1.4 \mu\text{m}$. Thus, the spectral distribution of the effect of aerosols on the planetary albedo varies according to θ_0 .

4.3 Effect of cloud cover

Figure 11a shows the spectral planetary albedo for the cloudy atmosphere (α_{4g+c}^p) and for the clear case (α_{4g}^p), where the cloud cover basically increases the planetary albedo, especially in the near infrared region. This is because a cloud droplet is smaller than a snow grain, and the single scattering albedo of a cloud droplet is higher than that of a snow grain. Valovcin (1978) noted that reflectance in the region $1.4 < \lambda < 1.8 \mu\text{m}$ can be used in discriminating clouds from snow by aircraft observation. This method uses the difference in albedo between cloud cover and snow surface. An example of the θ_0 dependence of the monochromatic planetary albedo for the cloudy atmosphere is shown in Fig. 11b. The cloud cover increases the planetary albedo throughout the range of θ_0 . However, it should be noted that although it is not shown here, this situation could be reversed in some cases. For example, the value of

Fig. 11. (a) Same as Fig. 10a, but for the cloudy atmosphere of SW (α_{4g+c}^p) with three kinds of τ_c and clear case (α_{4g}^p). (b) Monochromatic planetary albedo as a function of θ_0 for the cloudy atmosphere and clear case at $\lambda = 1.0 \mu\text{m}$ over the snow with $r_{eff} = 200 \mu\text{m}$.

α_{4g+c}^p is slightly lower than that of α_{4g}^p for $\theta_0 = 5.9^\circ$, $0.5 < \lambda < 0.8 \mu\text{m}$, $\tau_c = 2.5$ and $r_{eff} = 200 \mu\text{m}$. This is because the multiple scattering by the cloud droplets increases the effective path length of diffuse radiation and thus increases the gaseous absorption.

5. Spectrally integrated surface and planetary albedos

5.1 Effect of the difference in atmospheric type

The "surface albedo" observed with a pyranometer is the spectrally integrated surface albedo ($\bar{\alpha}^s$). Its θ_0 dependence has been investigated in many observational studies (e.g., Carroll and Fitch, 1981; Yamanouchi, 1983) and theoretical studies (Wiscombe and Warren, 1980a, Choudhury and Chang, 1981a). The argument whether surface specular reflection should be considered for the study of θ_0 dependence of the spectral surface albedo has been made between Warren and Wiscombe (1981) and Choudhury (1981). This controversy arises from the problem that the θ_0 dependence of observed albedo did not agree with the result of theoretical

calculation made on the assumption of spherical ice particles for snow grains. Warren (1982) argued that the reasons for the confusion are: (1) modeling error, (2) instrument error, (3) inadequate observation of snow grain size, impurity content, and surface roughness — all of which should affect the slope of surface albedo — and, (4) inadequate knowledge of the spectral distribution of the incident radiation. Yamanouchi (1983) compared the observation at Mizuho Station in Antarctica with the theoretical model of Wiscombe and Warren (1980a) without surface reflection, and found that the theoretical model can explain the measured snow albedo for a renewed snow surface with fine-grained rounded particles. However, the model cannot explain the snow albedo when the surface grains become more faced as mentioned by Choudhury (1981) or when sintering of particles occurred and surface specular reflection comes more pronounced. All of these effects have to be considered when observed surface albedos are compared with the theoretical calculations. In the followings, we would like to take a notice to another important factor, the atmospheric effect, that affects on $\bar{\alpha}^s$ of snow:

Figure 12a shows $\bar{\alpha}^s$ of snow as a function of θ_0 for three model atmospheres. It has been shown in Subsection 3.3 that the effect of difference in atmospheric type on the spectral surface albedo is very small ($|\alpha_{MW}^s - \alpha_{AS}^s| < 0.01$) for $\theta_0 \leq 63.1^\circ$ and α_{MW}^s is lower than α_{AS}^s only in gaseous absorption bands for $\theta_0 > 63.1^\circ$. However, the value of $\bar{\alpha}^s$ varies depending on the model atmospheres and $\bar{\alpha}^s$ for MW ($\bar{\alpha}_{MW}^s$) is higher than that for AS ($\bar{\alpha}_{AS}^s$) at any θ_0 . Namely, the cause of differences in $\bar{\alpha}^s$ between model atmospheres is not attributed to the differences in spectral surface albedo in the gaseous absorption bands.

The spectrally integrated surface albedo $\bar{\alpha}^s$ is given by

$$\bar{\alpha}^s = \frac{\int_0^\infty \alpha^s(\lambda) F^\downarrow(\lambda) d\lambda}{\int_0^\infty F^\downarrow(\lambda) d\lambda}, \quad (1)$$

where $F^\downarrow(\lambda)$ is the downward solar flux at snow surface, and $\alpha^s(\lambda)$ is the spectral surface albedo of snow. The value of $F^\downarrow(\lambda)$ varies depending on the atmospheric condition, especially on H_2O amount. The Eq. (1) means that $\bar{\alpha}^s$ is the weighted mean of $\alpha^s(\lambda)$ with the weight of normalized downward flux. Even if $\alpha^s(\lambda)$ does not change with the atmospheric condition, $\bar{\alpha}^s$ could change depending on the atmosphere through $F^\downarrow(\lambda)$. The major H_2O bands are located in the near infrared region. Hence, $F^\downarrow(\lambda)$ for MW (H_2O rich) is smaller than that for AS (H_2O poor) in the near infrared region, but both are almost the same in the visible region. On the other hand, $\alpha^s(\lambda)$ is high in the visible region and low in the near infrared region. Since the weight ($F^\downarrow(\lambda)$) for low value of $\alpha^s(\lambda)$ in the near infrared region is

Fig. 12. Spectrally integrated albedos as a function of θ_0 at (a) snow surface ($\bar{\alpha}^s$) and (b) TOA ($\bar{\alpha}^p$) over the snow with $r_{eff} = 50, 200$ and $1000 \mu m$ for the clear atmospheres of MW, SW and AS.

small for MW and large for AS, the value of $\bar{\alpha}_{MW}^s$ becomes higher than that of $\bar{\alpha}_{AS}^s$. The maximum difference between $\bar{\alpha}_{MW}^s$ and $\bar{\alpha}_{AS}^s$ amounts to 0.015 for $r_{eff} = 50 \mu m$ and 0.021 for $r_{eff} = 1000 \mu m$ at $\theta_0 = 5.9^\circ$. Thus, several percent of spectrally integrated surface albedo could vary with the H_2O amount, even for the same spectral surface albedo.

Figure 12b shows the spectrally integrated planetary albedo ($\bar{\alpha}^p$) as a function of θ_0 . In this case, $\bar{\alpha}^p$ has the same form as Eq. (1) with replacing $\alpha^s(\lambda)$ by $\alpha^p(\lambda)$ and $F^\downarrow(\lambda)$ by the downward solar flux at TOA. Since the downward solar flux at TOA does not depend on the atmospheric condition, the behavior of $\alpha^p(\lambda)$ would be directly reflected on $\bar{\alpha}^p$. Contrary to the case of $\bar{\alpha}^s$, the value of $\bar{\alpha}_{MW}^p$ is lower than that of $\bar{\alpha}_{AS}^p$. This is because $\alpha_{MW}^p < \alpha_{AS}^p$ mainly in the H_2O bands (see Fig. 9). The difference between $\bar{\alpha}_{MW}^p$ and $\bar{\alpha}_{AS}^p$ increases with θ_0 , and amounts to 0.043 for $r_{eff} = 50 \mu m$ and 0.028 for $r_{eff} = 1000 \mu m$ at $\theta_0 = 79.0^\circ$.

Although $\bar{\alpha}^s$ gradually increases with θ_0 , $\bar{\alpha}^p$ rapidly decreases at around $\theta_0 = 75^\circ$. The former reflects the property of θ_0 dependence of $\alpha^s(\lambda)$. The reason of the latter phenomenon is that the absolute

values of the downward solar flux on the snow surface is reduced by the strong absorption of the direct solar beam propagating along the long slant path at large θ_0 and the resultant upward solar flux at TOA is also small. The value of $\bar{\alpha}^p$ is lower than that of $\bar{\alpha}^s$ due to the atmospheric absorption, and the difference between $\bar{\alpha}^s$ and $\bar{\alpha}^p$ becomes roughly 0.1–0.3, although it depends on r_{eff} , θ_0 and the model atmospheres.

5.2 Effect of aerosols

The θ_0 dependence of spectrally integrated surface albedo for the atmosphere containing aerosols ($\bar{\alpha}_{4g+a}^s$) is shown in Fig. 13a. The effect of background aerosols $\tau_a = 0.02$ on $\bar{\alpha}^s$ is very small, and the curve is overlapped on that for aerosol free atmosphere ($\bar{\alpha}_{4g}^s$). When the optical depth of aerosols τ_a increases, the aerosols weaken the θ_0 dependence of $\bar{\alpha}^s$. This comes from the characteristic of θ_0 dependence of the monochromatic surface albedo, shown in Fig. 6a. The curve of $\bar{\alpha}_{4g+a}^s$ crosses that of $\bar{\alpha}_{4g}^s$ at around $\theta_0 = 55^\circ$. That is, the aerosols increase $\bar{\alpha}^s$ at small θ_0 , but reduce it at large θ_0 . Blanchet and List (1987) showed the reduction of $\bar{\alpha}^s$ by 0.006 caused by their Arctic aerosol model 1 with $\tau_a = 0.081$ for $\theta_0 \sim 70^\circ$ on the snow with the grain radius of $200 \mu\text{m}$. The estimate of $\bar{\alpha}^s$ reduction by our aerosol model is 0.004 at $\tau_a = 0.1$, $\theta_0 = 70^\circ$ and $r_{eff} = 200 \mu\text{m}$. According to Warren and Clark (1986) the aerosols, such as Arctic haze contained in snowfall and snowpack (snow impurities), could reduce $\bar{\alpha}^s$ up to 0.01–0.04 by absorbing the solar radiation in snow in the Arctic region. If the Antarctic snow is polluted by the absorptive impurities, the atmospheric aerosols could enhance its reduction of $\bar{\alpha}^s$ at $\theta_0 > 55^\circ$.

The atmospheric aerosols have the effect of reducing the insolation to the surface by shading the downward solar radiation. Figure 13b shows the change in global solar radiation ($\bar{F}$) by aerosols as a function of θ_0 . At small θ_0 the value of $\bar{F}$ is increased slightly by aerosols due to the reason that will be described in the next section. However, it is reduced by aerosols except for small θ_0 . This reduction is estimated by 3 % for $\tau_a = 0.1$, and 11 % for $\tau_a = 0.3$ at $\theta_0 = 70^\circ$. It is more significant than this for $\theta_0 > 70^\circ$. From the viewpoint of the surface radiation budget in the shortwave region, this reduction in $\bar{F}$ (cooling the surface) compensates the reduction of $\bar{\alpha}^s$ (heating the surface) by the aerosols and snow impurities mentioned above. Particularly in the case of volcanic ash aerosols, the effect of reduction in $\bar{F}$ exceeds that of $\bar{\alpha}^s$ at $\theta_0 > 55^\circ$.

Figure 13c shows the θ_0 dependence of spectrally integrated planetary albedo for the atmosphere containing aerosols ($\bar{\alpha}_{4g+a}^p$), where the curve of $\bar{\alpha}_{4g+a}^p$ with $\tau_a = 0.02$ overlaps with that for aerosol free atmosphere ($\bar{\alpha}_{4g}^p$). The effect of the background

Fig. 13. (a) Same as Fig. 12a, but for the atmosphere of SW containing aerosols ($\bar{\alpha}_{4g+a}^s$) with three kinds of τ_a and aerosol free case ($\bar{\alpha}_{4g}^s$). (b) Global solar radiation ($\bar{F}_{4g}$) for aerosol free atmosphere of SW over the snow with $r_{eff} = 200 \mu\text{m}$ (left ordinate) and the change in global solar radiation ($\Delta\bar{F}/\bar{F}_{4g}$) by the aerosols (right ordinate), where $\Delta\bar{F} = \bar{F}_{4g+a} - \bar{F}_{4g}$ and $\bar{F}_{4g+a}$ is the global solar radiation for the atmosphere of SW containing aerosols with three kinds of τ_a . (c) Same as Fig. 12b, but for the atmosphere of SW containing aerosols ($\bar{\alpha}_{4g+a}^p$) with three kinds of τ_a and aerosol free case ($\bar{\alpha}_{4g}^p$).

aerosols on $\bar{\alpha}^p$ is very small in the Antarctic, but this is due to small optical depth. In the cases of $\tau_a \geq 0.1$, the aerosols (mainly volcanic ash) reduce $\bar{\alpha}^p$ except at large θ_0 . This means that the effect of volcanic ash in the Antarctic is positive (heating) on the radiation budget in the shortwave region at TOA except at large θ_0 . Blanchet and List (1987) showed the reduction of $\bar{\alpha}^p$ by 0.021 caused by their Arctic aerosol model 1 with $\tau_a = 0.1$ for $\theta_0 \sim 70^\circ$ on the snow with the grain radius of 200 μm . In our case the reduction of $\bar{\alpha}^p$ is estimated by 0.003 caused by the aerosols with $\tau_a = 0.1$ for $\theta_0 = 70^\circ$ and $r_{eff} = 200 \mu\text{m}$. However, the reduction of $\bar{\alpha}^p$ amounts to 0.012 for $\tau_a = 0.3$. These results suggest that the thick volcanic ash aerosols over the Antarctic could have the effect of a same order as the Arctic haze at $\theta_0 = 70^\circ$ on the radiation budget at TOA. On the other hand, volcanic ash increases $\bar{\alpha}^p$ of snow at large θ_0 . Thus, it should be noticed that the radiative forcing of the Antarctic aerosols may vary with θ_0 .

5.3 Effect of cloud cover

It is the well known phenomenon that the value of $\bar{\alpha}^s$ of snow under cloudy sky is higher than that for the clear case (Liljequist, 1956; Grenfell and Maykut, 1977; Wiscombe and Warren, 1980a; Warren, 1982; Yamanouchi, 1983). To reproduce such a situation theoretically, the spectrally integrated surface albedos are calculated in the case of a cloudy atmosphere ($\bar{\alpha}_{4g+c}^s$) for $r_{eff} = 50$ and 200 μm , and the cases of cloudy atmosphere ($\bar{\alpha}_{4g+c}^s$) and clear atmosphere ($\bar{\alpha}_{4g}^s$) for $r_{eff} = 1000 \mu\text{m}$ (Fig. 14a). The value of $\bar{\alpha}_{4g+c}^s$ increases with τ_c and it becomes higher than $\bar{\alpha}_{4g}^s$ at any θ_0 for $\tau_c \geq 5$. This cannot be understood intuitively from the θ_0 dependence of the monochromatic surface albedo shown in Fig. 7a. Liljequist (1956) explained this phenomenon by the difference in spectral distribution of downward solar flux between the clear case and cloudy case. The details of this explanation could be given using Eq. (1) in the following: Under the cloudy atmosphere the downward solar flux $F^\downarrow(\lambda)$ is much smaller than that for the clear case in the near infrared region, whereas in the visible region both $F^\downarrow(\lambda)$ are not so much different or $F^\downarrow(\lambda)$ under the cloudy atmosphere is even larger than that for the clear case at small θ_0 , as will be shown in the next section. On the other hand, the spectral surface albedo $\alpha^s(\lambda)$ is high in the visible region and low in the near infrared region under both clear and cloudy conditions. In Eq. (1) the spectrally integrated surface albedo $\bar{\alpha}^s$ is the flux weighted mean of $\alpha^s(\lambda)$ as mentioned in Subsection 5.1. Since the weight ($F^\downarrow(\lambda)$) for low $\alpha^s(\lambda)$ in the near infrared region is large for the clear case and small for the cloudy case, the value of $\bar{\alpha}_{4g+c}^s$ becomes higher than that of $\bar{\alpha}_{4g}^s$. In Fig. 14a it is also shown that the θ_0 dependence of

Fig. 14. (a) Spectrally integrated surface albedos as a function of θ_0 in the case of cloudy atmosphere of SW ($\bar{\alpha}_{4g+c}^s$) with $\tau_c = 10$ for $r_{eff} = 50$ and 200 μm , and the cases of cloudy atmospheres ($\bar{\alpha}_{4g+c}^s$) with three kinds of τ_c and clear atmosphere ($\bar{\alpha}_{4g}^s$) for 1000 μm . (b) Same as Fig. 14a, but for TOA in the cloudy case ($\bar{\alpha}_{4g+c}^p$) and clear case ($\bar{\alpha}_{4g}^p$).

$\bar{\alpha}_{4g+c}^s$ becomes weaker with the increase of τ_c . This comes from the characteristic of θ_0 dependence of the monochromatic surface albedo under the cloudy atmosphere shown in Fig. 7a. The cloud cover also has the effect to reduce the insolation to the surface by shading the downward solar radiation, as will be shown in the next section. Together with the effect of increasing $\bar{\alpha}^s$ mentioned above, the cloud cover has a cooling effect on the radiation budget in the shortwave region at snow surface.

Figure 14b shows the spectrally integrated planetary albedos for the cloudy atmosphere ($\bar{\alpha}_{4g+c}^p$), where $\bar{\alpha}_{4g+c}^p$ is higher than that for the clear case ($\bar{\alpha}_{4g}^p$) at any θ_0 . This is because a cloud droplet is smaller than a snow grain and the single scattering albedo of a cloud droplet is higher than that of a snow grain. Thus, the cloud cover has the so-called "albedo effect" (cooling effect) on the radiation budget in the shortwave region at TOA. These results are consistent with the study of Yamanouchi and

Charlock (1995), using satellite data and radiation budget observation at the snow surface.

6. Effect of multiple reflection between snow surface and cloud cover

Wendler *et al.* (1981) discussed the increase of irradiance on the surface accompanied by the change of surface optical properties from tundra (low albedo) to snow (high albedo) under Arctic stratus cloud at Barrow, Alaska. They argued that this is due to the multiple reflection between snow surface and cloud cover. Yamanouchi (1983) also noticed that both the global solar radiations for clear and cloudy skies are larger than the observations in mid latitudinal regions due to the multiple reflection between snow surface and atmosphere (cloud cover).

Figure 15a shows the spectral distributions of downward solar flux ($F^\downarrow$) at snow surface under clear and cloudy skies, and $F^\downarrow$ at TOA (extraterrestrial solar flux). In the near infrared region, the value of $F^\downarrow$ is large in order of TOA, clear atmosphere, and cloudy atmosphere at any θ_0 . In the visible region at $\theta_0 = 74.3^\circ$ the value of $F^\downarrow$ are large in the same order as the near infrared region. However, at $\theta_0 = 43.2^\circ$ the value of $F^\downarrow$ for the cloudy atmosphere is almost the same as that for the clear case, and both exceed the extraterrestrial solar flux in the region $0.375 \leq \lambda \leq 0.475 \mu\text{m}$. Furthermore, at $\theta_0 = 5.9^\circ$ the value of $F^\downarrow$ for clear atmosphere is larger than the extraterrestrial solar flux for the broader region of $0.375 \leq \lambda \leq 0.525 \mu\text{m}$. The value of $F^\downarrow$ for the cloudy atmosphere exceeds both of $F^\downarrow$ for the clear case and extraterrestrial solar flux at almost all visible wavelengths.

The reason for this phenomenon is somewhat complicated to explain. We shall consider a simple model of cloud-snow system where there is no Rayleigh atmosphere, and no absorption by the atmospheric gases but only cloud. Let us denote the monochromatic surface albedo of snow by R_s , and those of cloud against the direct solar flux and upward solar flux by R_{cd} and R_{cu} , respectively. When the direct solar flux of an amount of unity illuminates this system, the total amount of monochromatic downward solar flux $F^\downarrow$ at the snow surface becomes

$$F^\downarrow = \frac{(1 - R_{cd})}{(1 - R_s R_{cu})}, \quad (2)$$

after the multiple reflection between snow surface and cloud layer. If R_{cd} and R_{cu} are the same, $F^\downarrow$ never exceeds unity. But they could be different in the real field. To examine this we only consider the visible region ($\lambda = 0.45 \mu\text{m}$) and small θ_0 . The values of R_{cd} was calculated by one-layer cloud model with $\tau_c = 10$ (without underlying snow surface) illuminated by the direct solar beam, and we obtained $R_{cd} = 0.36$ at $\theta_0 = 5.9^\circ$. For the calculation of R_{cu} ,

Fig. 15. (a) Spectral distribution of downward solar flux ($F^\downarrow$) at snow surface with $r_{\text{eff}} = 200 \mu\text{m}$ under clear atmosphere of SW, and cloudy case with $\tau_c = 10$, and $F^\downarrow$ at TOA for three values of θ_0 . Curves are successively displaced upward by $0.5 \text{ kW/m}^2\mu\text{m}$ except for $\theta_0 = 74.3^\circ$. (b) Global solar radiation at snow surface as a function of θ_0 for clear atmosphere ($\bar{F}_{\text{clr}}^s$) and cloudy case ($\bar{F}_{\text{clr}}^c$), that at sea surface for clear case ($\bar{F}_{\text{clr}}^o$) and cloudy case ($\bar{F}_{\text{clr}}^o$), and that at TOA ($\bar{F}_{\text{TOA}}$).

we assumed the cloud base is illuminated by diffuse radiation, which is equivalent to the direct insolation of solar beam at $\theta_0 = 48^\circ$ as shown in Fig. 1b, and obtained $R_{cu} = 0.49$ from the one-layer cloud model described above. Similarly the snow surface is also illuminated by diffuse radiation, and we obtained $R_s = 0.98$ at $\theta_0 = 48^\circ$ from one-layer snow model with $r_{\text{eff}} = 200 \mu\text{m}$ illuminated by the direct solar beam. Using these parameters, we obtain $F^\downarrow = 1.23$, showing the downward solar flux on the snow surface exceeds the extraterrestrial solar flux.

For the case of clear atmosphere, R_{cd} and R_{cu} in Eq. (2) are replaced by the albedos of the atmosphere, R_{ad} and R_{au} , respectively. These values are calculated for the Rayleigh atmosphere illuminated by the direct solar beam without underlying snow surface, and we obtained $R_{ad} = 0.10$ at $\theta_0 = 5.9^\circ$

Table 2. Major characteristics of the atmospheric effects on snow albedos: spectral surface albedo (α^s), spectral planetary albedo (α^p), spectrally integrated surface albedo ($\bar{\alpha}^s$) and spectrally integrated planetary albedo ($\bar{\alpha}^p$). Subscripts means as follows; 4g: absorptive atmosphere with Rayleigh scattering, na: no atmosphere, AS: model atmosphere of Antarctic summer, MW: model atmosphere of midlatitude winter, 4g + a: aerosol contained atmosphere, 4g + c: cloudy atmosphere. Meanings of the other symbols are given in Table 1.

Albedo	Gaseous absorption and Rayleigh scattering	Difference of atmospheric type	Aerosols	Cloud cover
α^s	Gas: $\alpha_{4g}^s < \alpha_{na}^s$ in absorption bands at large θ_0 , Rayleigh: Weaken the θ_0 dependence at $\lambda < 0.5 \mu\text{m}$ (*1)	$\alpha_{MW}^s < \alpha_{AS}^s$ in H_2O bands at large θ_0	Weaken the θ_0 dependence (*1)	Weaken the θ_0 dependence (*1)
α^p	Gas: $\alpha_{4g}^p < \alpha_{na}^p$ in absorption bands, Significant for small r_{eff} , Rayleigh: Slightly $\alpha_{4g}^p > \alpha_{na}^p$ around $\lambda = 0.4 \mu\text{m}$	$\alpha_{MW}^p < \alpha_{AS}^p$ in H_2O bands	$\alpha_{4g+a}^p < \alpha_{4g}^p$ at shorter λ and $\alpha_{4g+a}^p > \alpha_{4g}^p$ at longer λ	$\alpha_{4g+c}^p > \alpha_{4g}^p$ in the near infrared (*2)
$\bar{\alpha}^s$	$\bar{\alpha}_{4g}^s > \bar{\alpha}_{4g}^p$, Significant for MW	$\bar{\alpha}_{MW}^s > \bar{\alpha}_{AS}^s$	Weaken the θ_0 dependence (*1)	Weaken the θ_0 dependence, $\bar{\alpha}_{4g+c}^s > \bar{\alpha}_{4g}^s$ at any θ_0 for large τ_c
$\bar{\alpha}^p$		$\bar{\alpha}_{MW}^p > \bar{\alpha}_{AS}^p$	$\bar{\alpha}_{4g+a}^p < \bar{\alpha}_{4g}^p$ except at large θ_0 and $\bar{\alpha}_{4g+a}^p > \bar{\alpha}_{4g}^p$ at large θ_0	$\bar{\alpha}_{4g+c}^p < \bar{\alpha}_{4g}^p$ at any θ_0

*1: Albedo is increased at small θ_0 and reduced at large θ_0 , *2: Slightly $\alpha_{4g+c}^p < \alpha_{4g}^p$ for the visible, small θ_0 and small τ_c .

and $r_{au} = 0.14$ at $\theta_0 = 48^\circ$ for $\lambda = 0.45 \mu\text{m}$. Using these values with $R_s = 0.97$ for direct illumination at $\theta_0 = 5.9^\circ$, we obtain $F^\downarrow = 1.04$ showing again the downward solar flux on the snow surface is larger than the extraterrestrial solar flux.

The keys to the condition which these phenomena happen are “snow surface” (high R_s), “visible wavelength” (weak atmospheric absorption), and “small solar zenith angle” ($R_{cd} < R_{cu}$, $R_{ad} < R_{au}$). When θ_0 is large, R_{cd} (or R_{ad}) is not less than R_{cu} (or R_{au}), and thus $F^\downarrow$ never exceeds unity. For the near infrared region, the value of R_s is low and thus $F^\downarrow$ never exceeds unity again.

Figure 15b shows the spectrally integrated values (global solar radiation) of the downward solar flux at the snow surface under clear atmosphere ($\bar{F}_{clr}^s$) and the cloudy case ($\bar{F}_{cld}^s$), and those at the sea surface under the clear case ($\bar{F}_{clr}^o$) and the cloudy case ($\bar{F}_{cld}^o$) as a function of θ_0 . For comparison, the corresponding value at TOA ($\bar{F}_{TOA}^o$) is also shown. In calculating the $\bar{F}_{clr}^o$ and $\bar{F}_{cld}^o$, the sea surface albedo was assumed to be Lambertian, with a value of 0.05 for all wavelengths. Although the monochromatic value of $F^\downarrow$ under the cloudy atmosphere exceeds the extraterrestrial solar flux at small θ_0 in the visible region as has been above, $\bar{F}_{cld}^s$ never exceeds both $\bar{F}_{clr}^s$ and $\bar{F}_{TOA}^o$, and $\bar{F}_{clr}^s$ also never exceeds $\bar{F}_{TOA}^o$. On the other hand, $\bar{F}_{clr}^s$ (or $\bar{F}_{cld}^s$) is larger than $\bar{F}_{clr}^o$ (or $\bar{F}_{cld}^o$), and the ratio $\bar{F}_{clr}^s/\bar{F}_{clr}^o$ (or $\bar{F}_{cld}^s/\bar{F}_{cld}^o$) becomes 1.06 (or 1.87) at $\theta_0 = 5.9^\circ$. These are due to the effect of multiple reflection between snow surface and the atmosphere (or cloud cover). In particular, the difference in global solar radiation between snow surface and sea surface is significant for the cloudy

case. This shows the interaction between snow surface and cloud cover is very important for the radiation budget in the cryosphere.

7. Summary

The multiple scattering radiative transfer model for the atmosphere-snow system has been developed to estimate the atmospheric effects on the spectral and spectrally integrated snow albedos at the snow surface and TOA. In this model the interaction between the snow surface and atmosphere is calculated using the “doubling and adding” method for multiple scattering, combined with the Mie theory for single scattering by snow grain, aerosol particles and cloud droplets in the wavelength region from 0.3 to 3.0 μm . The atmospheric transmittance of gaseous absorption is calculated by the extended exponential-sum fitting of transmissions for three kinds of model atmospheres.

The surface albedo varies with the fraction of the diffuse component in the downward solar flux. This fraction changes with the atmospheric composition. As a result, the behavior of surface albedo is influenced by the atmosphere. The major characteristics of the atmospheric effects on snow albedos clarified in this study are summarized in Table 2. The atmospheric absorptive gases reduce the surface albedo α_{4g}^s in the absorption bands at large θ_0 , compared with that for no atmosphere (α_{na}^s). In the H_2O absorption bands the surface albedo for H_2O rich atmosphere (α_{MW}^s) is lower than that for H_2O poor atmosphere (α_{AS}^s) for $\theta_0 > 63.1^\circ$. However, the spectrally integrated surface albedo for MW ($\bar{\alpha}_{MW}^s$) is higher than that for AS ($\bar{\alpha}_{AS}^s$) by up to 0.021.

This is because the downward solar flux $F^\downarrow(\lambda)$ in the near infrared region is lower for H₂O rich MW than AS. This point is important for the interpretation of the snow albedo observed with such spectrally integrated instruments as pyranometers. On the other hand, the spectrally integrated planetary albedo for MW ($\bar{\alpha}_{MW}^p$) is lower than that for AS ($\bar{\alpha}_{AS}^p$) by up to 0.043. At any θ_0 the value of $\bar{\alpha}^p$ is lower than $\bar{\alpha}^s$ due to the atmospheric absorption. The difference between $\bar{\alpha}^p$ and $\bar{\alpha}^s$ lies in the range of 0.1–0.3.

Antarctic coastal aerosols weaken both the θ_0 dependences of spectral surface albedo α_{4g+a}^s , and spectrally integrated surface albedo $\bar{\alpha}_{4g+a}^s$ at almost all wavelengths. However, these aerosols have the effect of reducing the insolation to surface by the shading effect. Background sulfuric acid aerosols in the Antarctic have no significant effect on the planetary albedo α_{4g+a}^p (thus on $\bar{\alpha}_{4g+a}^p$), due to the low optical depth. The volcanic ash, however, reduces α_{4g+a}^p in the shorter wavelength region, and increases it in the longer wavelength region. The resultant effect on $\bar{\alpha}_{4g+a}^p$ by the volcanic ash is the reduction, except at large θ_0 . Although these are basically due to the difference in single scattering albedo between the volcanic ash and snow grain, the difference in vertical profile between the volcanic ash and absorptive gases also affects the result.

High latitudinal middle clouds with large optical depth increase the spectrally integrated surface albedo $\bar{\alpha}_{4g+c}^s$ at any θ_0 . This is explained by the difference in spectral distribution of downward solar flux between clear and cloudy cases. Due to this effect and the shading effect of insolation, the cloud cover has a cooling effect on the radiation budget in the shortwave region at the snow surface. The cloud cover increases the spectral planetary albedo α_{4g+c}^p in the near infrared region (and thus increases $\bar{\alpha}_{4g+c}^p$) at any θ_0 due to the difference of single scattering albedo between cloud droplet and snow grain. Thus, the cloud cover studied here has a cooling effect at any θ_0 in the shortwave region at TOA.

The downward solar flux on snow surface under cloudy sky can exceed that for clear case in the visible region at small θ_0 , and both further exceed the extraterrestrial solar flux. This phenomenon is explained by the multiple reflection between snow surface and atmosphere (or cloud cover). Global solar radiation on snow surface under cloudy sky never exceeds that for clear case and that at TOA, but it is 1.87 times larger than that on the sea surface. Although the increase of downward solar flux by the presence of cloud cover over snow surface was observed in the past studies, the phenomenon that the downward solar flux exceeds the extraterrestrial solar flux has never been observed. This may be due to the lack of spectral observations under the condition of small θ_0 on snow surface. Such phenomenon may be observed at high mountain areas in the low

latitude.

It has been found that the effects of clouds on albedo and radiation budget are important on the snow surface. We did not study the effects of high cloud or diamond dust, which frequently appear in the polar region. The investigations for such ice clouds with small optical depth are also important items in the snow covered area. However, the amount of their effects on snow albedo will be smaller than those studied in the present work.

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大気—積雪系の多重散乱放射伝達モデルによる 積雪アルベドに及ぼす大気の効果の数値シミュレーション

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雪面及び大気上端 (TOA) における波長別及び波長積分した積雪アルベドに対する大気の影響を調べた。そこでは空気分子、吸収気体、エアロゾル、雲による吸収と散乱の影響を、doubling and adding 法と Mie 理論に基づいた大気—積雪系の多重散乱放射伝達モデルによって見積もった。

波長別雪面アルベドは太陽天頂角が大きいとき、大気中の吸収気体によって大気がないときに比べて減少することが示された。その太陽天頂角依存性は波長 $0.5\ \mu\text{m}$ 以下でレイリー散乱によって弱められ、ほとんどの波長でエアロゾル及び雲によって弱められた。水蒸気の豊富な大気は、太陽天頂角が大きいとき、水蒸気の吸収帯で波長別アルベドを減少させた。ところが近赤外域の下向きフラックスが水蒸気の吸収によって減少するため、波長積分したアルベドは数パーセント高くなった。エアロゾルは太陽天頂角が小さいとき波長積分した雪面アルベドを増加させ、太陽天頂角が大きいときには減少させた。しかし、エアロゾルは太陽天頂角が大きいときを除き、波長積分したプラネタリーアルベドを減少させた。光学的に厚い雲は太陽天頂角に依らず波長積分した雪面及びプラネタリーアルベドの両者を増加させた。

太陽天頂角が小さいとき可視域では、雪面上における曇天時の下向きフラックスが晴天時のそれを上回り、また両者はさらに大気外日射フラックスを上回り得ることがわかった。この現象は雪面と大気 (雲) の間の多重反射によって説明できる。しかし、雪面上における曇天時の全天日射量は、晴天時及び TOA におけるそれらを上回ることにはなかった。