A Novel Parameterization of Snow Albedo Based on a Two-Layer Snow Model with a Mixture of Grain Habits

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(Manuscript received 12 October 2018, in final form 25 February 2019)

ABSTRACT

Snow albedo plays a critical role in the surface energy budget in snow-covered regions and is subject to large uncertainty due to variable physical and optical characteristics of snow. We develop an optically and microphysically consistent snow grain habit mixture (SGHM) model, aiming at an improved representation of bulk snow properties in conjunction with considering the particle size distribution, particle shape, and internally mixed black carbon (BC). Spectral snow albedos computed with two snow layers with the SGHM model implemented in an adding–doubling radiative transfer model agree with observations. Top-snow-layer optical properties essentially determine spectral snow albedo when the top-layer snow water equivalent (SWE) is large. When the top-layer SWE is less than 1 mm, the second-snow-layer optical properties have nonnegligible impacts on the albedo of the snow surface. Snow albedo enhancement with increasing solar zenith angle (SZA) largely depends on snow particle effective radius and also internally mixed BC. Based on the SGHM model and various sensitivity studies, single- and two-layer snow albedos are parameterized for six spectral bands used in NASA Langley Research Center’s modified Fu–Liou broadband radiative transfer model. Parameterized albedo is expressed as a function of snow particle effective radii of the two layers, SWE in the top layer, internally mixed BC mass fraction in both layers, and SZA. Both single-layer and two-layer parameterizations provide band-mean snow albedo consistent with rigorous calculations, achieving correlation coefficients close to 0.99 for all bands.

1. Introduction

Snow plays an essential role in the surface radiation balance as well as the hydrological cycle through complex snow–atmosphere feedbacks (Randall et al. 1994). In particular, surface albedo is one of the main regulators of the radiative balance over snow-covered areas. Even with a persisting snow cover, changes in snow albedo affect air temperature, snow-cover extent, and the melting of snow. Therefore, an accurate snow albedo model in conjunction with a physical snow-layer model should be incorporated into general circulation models (GCMs) to obtain more accurate climate projections. Surface albedo of snow is generally much higher than that of land or ocean, and has significant variability associated with snow particle size (e.g., Wiscombe and

Warren 1980), shape (e.g., Jin et al. 2008; He et al. 2017b), impurity (e.g., Warren and Wiscombe 1980; Aoki et al. 2000, hereafter A00), and solar zenith angle (SZA; e.g., Grenfell et al. 1994). Among these factors, snow particle size is the primary factor that determines snow albedo, particularly in the near-infrared regime (Warren 1982; Aoki et al. 2003).

Warren and Wiscombe (1980) found that absorptive impurities in snow such as mineral dust, organic carbon, and soot [i.e., black carbon (BC) particles] are the primary factors determining snow albedo at visible wavelengths. Several previous studies revealed that, among the three types of absorbing impurities, BC usually accounts for 50%–80% of the total light absorption by snow at visible wavelengths, with the highest proportion in areas with substantial industrial pollution (e.g., Doherty et al. 2010; Wang et al. 2013; Dang et al. 2017). Therefore, BC is considered to be the primary source of light absorption in

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most snow-covered regions. In addition, BC in snow significantly affects the local surface radiation budget, especially in the Northern Hemisphere. Hansen and Nazarenko (2004) estimate that BC in snow yields a climate forcing of $+0.3 \text{ W m}^{-2}$ in the Northern Hemisphere, and this forcing is twice as efficient as that associated with CO_2 . Flanner et al. (2012) point out that internally mixed BC enhances snow absorption efficiency at visible wavelengths by a factor of 1.8–2.1 compared with externally mixed BCs. Since internal mixing of BC in snow particles can affect many atmospheric processes such as ice nucleation, scavenging, and a combination of dry deposition over surface snow and melting and refreezing of snow, absorbing particles including BC should be taken into account in snow albedo simulations.

Snow particle shape variation also leads to uncertainty in snow albedo. In previous studies, snow particles were often assumed to be spherical particles for computational simplicity. However, many accumulated snow-sampling studies demonstrate that accumulated snow particles are typically nonspherical. Moreover, the particle shapes and sizes vary over time due to snow metamorphosis (e.g., Nakamura et al. 2001, hereafter N01). As snow cover persists, snow particle edges are gradually smoothed and the particles become more compact but are still in irregular shapes (Ishimoto et al. 2018, hereafter I18). Snow particle shape is an important factor determining albedo. Different particle shapes with the same effective radius have different mean particle volumes, causing different particle absorption. The asymmetry parameters of typical snow particles range from 0.75 to 0.9, depending on the particle shape. Jin et al. (2008) demonstrate a non-negligible effect of snow particle shape on visible snow albedo due to the variation of the asymmetry parameter. A snow particle shape with a low asymmetry parameter provides high visible snow albedo (Dang et al. 2016). Furthermore, snow albedo based on nonspherical snow particles with internally mixed BC has moderate sensitivity to snow particle shape (He et al. 2018c).

In addition to snow particle morphological characteristics, the vertically inhomogeneous variation of snow particle size affects spectral snow albedo. The snow cover may contain many layers accumulated over several months, with varying factors affecting snow layers between snowfalls such as melting, refreezing, and sublimation. With a snowfall over a snow cover, snow contains at least two layers: newly accumulated “fresh” snow in the top layer and preaccumulated “aged” snow in the bottom layer. Grenfell et al. (1994) demonstrate that simulated spectral snow albedo based on a two-layer assumption is more consistent with observations than the counterpart based on a single-layer

snow assumption. This is fundamentally due to the variation of the penetration depth of the incident solar radiation with wavelength (Zhou et al. 2003; Aoki et al. 2011). Based on these facts, a two-layer snow model was used to remote sensing applications regarding characterizing snow microphysical properties (Jin et al. 2008).

To efficiently compute snow albedo in GCMs with a snow forecast model, numerous snow albedo parameterizations have been developed (e.g., Marshall and Oglesby 1994; Yasunari et al. 2011). Although most snow albedo parameterizations consider snow particle size, and some consider snow impurity (Dang et al. 2015), particle shape (He et al. 2018c), or SZA (Gardner and Sharp 2010), none of the previous snow albedo parameterizations accounts for all these variables under the condition of two snow layers. In particular, simple snow albedo parameterizations assume a single snow layer to minimize the computational demand in dealing with snow albedo in GCMs, regardless of the fact that sophisticated land surface schemes in climate models do consider multiple snow layers (e.g., Molod et al. 2015; Saha et al. 2017). For rigorous snow albedo calculations for GCMs, a two-layer snow albedo parameterization is necessary to further improve the accuracy of snow albedo simulations as well as surface energy budget estimates over snow-covered regions.

Because of the need for a better representation of snow albedo, this study first describes a snow grain habit mixture (SGHM) model, a newly developed snow particle model based on in situ accumulated snow measurements. Using the SGHM, we demonstrate improvement in calculating rigorous spectral snow albedo due to a two-layer snow assumption. Finally, we also develop a two-layer snow albedo parameterization that is useful for rapid broadband shortwave radiative transfer calculation in weather forecasting and climate models.

Table 1 lists acronyms. Section 2 describes the methods for single-scattering calculation, snow albedo simulation, and definitions of variables to be used for parameterization. Section 3 discusses and validates the SGHM model. Section 4 provides rigorous spectral surface albedo simulations and sensitivity studies based on the two-layer snow-cover assumption. Section 5 presents development and evaluation of the present single-layer and two-layer snow albedo parameterizations. Section 6 summarizes the findings and conclusions of this study.

2. Methods

a. Optical property calculations of snow particles

The single-scattering properties of snow particles are fundamental to the computation of snow albedo. Realistic snow and ice particles are nonspherical and have

TABLE 1. The list of acronyms.

Acronyms	Meaning
A00	Aoki et al. (2000)
N01	Nakamura et al. (2001)
I18	Ishimoto et al. (2018)
BC	Black carbon particles
PSD	Particle size distribution
SSA	Specific surface area
SWE	Snow water equivalent
SZA	Solar zenith angle
DROX	Droxtal
SBUL	Solid bullet rosette
SCOL	Solid hexagonal column
SGHM	Snow grain habit mixture
GCM	General circulation model
LaRC-FL RTM	NASA Langley Research Center's modified Fu-Liou radiative transfer model
SNICAR	Snow, Ice, and Aerosol Radiative model

inhomogeneous internal structures, and tremendous efforts have been made to calculate the single-scattering properties of ice crystals (e.g., Macke et al. 1996; Liou et al. 2014). The physical geometric optics method was developed and refined to obtain reliable single-scattering properties of nonspherical ice crystals (Yang et al. 2019, and references cited therein). Tang et al. (2017) consider nonspherical particles with spherical particle inclusions using a stochastic approach that is applicable to calculating the single-scattering properties of nonspherical snow particles with internal BC mixing based on an assumption that internally mixed BCs are spherical. In this study, BC impurities on the surface of snow particles are also represented by internally mixed BC because internally mixed BC is the primary cause of reduced visible snow albedo (Warren and Wiscombe 1980; He et al. 2017b). The refractive indices of ice and BC used in this study are based on Warren and Brandt (2008) and Stegmann and Yang (2017), respectively. The volume densities are assumed to be 0.917 g cm^{-3} for ice ρ_{ice} (Wiscombe and Warren 1980) and 1.7 g cm^{-3} for BC (Bond and Bergstrom 2006). Schwarz et al. (2013) show that BC impurity is larger in snow particles on the surface than in the atmosphere based on direct snow in situ measurements, and the BC PSD has a long tail and deviates from a lognormal distribution. Therefore, the PSD of BC inside snow particles is assumed to be the gamma distribution with an effective radius of $0.1 \mu\text{m}$ and an effective variance of 0.1. In this study, the single-scattering properties of snow particles are computed for pure snow particles and four levels of impurities (BC mass fractions of 50, 100, 500, and 1000 ng g^{-1}), with snow particle maximum dimensions ranging from 2 to $10000 \mu\text{m}$. The assumed snow particle shape distribution for snow on the surface is discussed in section 3a.

b. Rigorous snow albedo calculations

To rigorously calculate spectral snow albedo α_λ , we use an adding–doubling radiative transfer model developed by Huang et al. (2015). The wavelength range for spectral snow albedo calculations extends from 0.2 to $5 \mu\text{m}$. The bulk optical properties of snow (section 3a), which are precomputed with the single-scattering properties of individual snow particles, and optical thickness are assigned to each plane-parallel homogeneous snow layer. Note that we assume that the single scattering by particles in densely packed snow is independent. This assumption is widely used for previous snow albedo studies, although it might cause biases in radiative transfer, partially in the backscattering direction due to coherent scattering (Mishchenko et al. 2011) causing an overestimated snow albedo (He et al. 2017a). The bottom snow layer is assumed to be optically thick with an optical thickness of 960, that is, practically semi-infinite. This assumption is valid when the fresh snow layer is deeper than 30 cm, and is more plausible for the snow layer, which includes a considerable amount of impurities (Dang et al. 2016). In the calculation, we assume the surface underneath the snow layer to be black. A broadband radiative transfer model for flux calculations has a finite number of bands in a spectral region, and a radiative transfer equation is solved for each of the bands. To provide broadband snow albedo α , we convolute the spectral snow albedo with an approximate mean clear-sky solar spectrum S_λ for the Northern Hemisphere, obtained from the American Society for Testing and Materials reference dataset (available at <https://rredc.nrel.gov/solar/spectra/am1.5/>), expressed by

$$\alpha = \frac{\int_{\lambda_{\min.}}^{\lambda_{\max}} \alpha_\lambda S_\lambda d\lambda}{\int_{\lambda_{\min}}^{\lambda_{\max}} S_\lambda d\lambda}, \quad (1)$$

where $\lambda_{\max}$ and $\lambda_{\min}$ respectively indicate the lower and upper wavelengths of a spectral band. This study does not consider the solar spectrum of a cloudy sky. Note that the effect of cloudiness on snow albedo is insignificant in the visible band, but is substantial in near-infrared bands that include water or ice absorption (Aoki et al. 1999).

c. Snow albedo parameterizations

After producing the rigorously calculated snow albedo for various scenarios of snow properties, we develop a two-layer snow albedo parameterization for computational efficiency in GCM applications. The top layer is “fresh” snow with a variable thickness specified in terms

of SWE, and the bottom layer is “aged” snow. The model variables contain SWE in the top snow layer, snow particle effective radii in the top and bottom snow layers, the cosine of SZA, and the BC mass fraction that is the same in the two snow layers.

The snow particle effective radius R_e is defined as

$$R_e = \frac{3V}{4A_{\text{proj}}}, \quad (2)$$

where V and A_{proj} are the particle volume and projected area, respectively. In the literature, other size definitions such as the volume-equivalent sphere radius R_V and projected-area-equivalent sphere radius R_A are also widely used for snow particle studies, given by

$$R_V = \left(\frac{3V}{4\pi} \right)^{1/3}, \quad (3)$$

and

$$R_A = \left(\frac{A_{\text{proj}}}{\pi} \right)^{1/2}. \quad (4)$$

The parameter R_V can be obtained from particle mass in a straightforward manner (Brandes et al. 2007), and R_A is derived from photographic images of snow particles with Eq. (4) (A00). The SSA, defined as the ratio of the total surface area to the mass of a snow particle, is widely used and satisfies $R_e = 3/(\rho_{\text{ice}}\text{SSA})$ for spherical or convex particles (e.g., spheroids, hexagonal columns, and droxtals). It is not technically easy to directly measure SSA. Therefore, this study does not consider a conversion from SSA to R_e but provides conversions from R_V and R_A to R_e to enhance usability of these parameterizations. In this study, we consider the six spectral bands involved in LaRC-FL RTM (Fu and Liou 1993; Kratz and Rose 1999; Kato et al. 1999, 2005), which spectrally covers 0.2–4 μm .

3. Bulk optical properties of the SGHM model

This section presents the SGHM model developed based on previous in situ measurements of snow on the surface. In addition, detailed descriptions of the SGHM model treatment of particle shape (section 3a), PSD (section 3b), and internally mixed BC (section 3c) are presented. The bulk optical properties of the SGHM model are evaluated in section 3d.

a. Snow particle habit mixtures

Many snow albedo schemes and parameterizations assume spherical snow particles (e.g., Dang et al. 2015)

FIG. 1. Habit fraction diagram of the snow grain habit mixture (SGHM) model. The particle habits include the solid hexagonal column (SCOL), solid bullet rosette (SBUL), and droxtal (DROX).

or a single nonspherical shape over the entire snow particle size range (e.g., He et al. 2018c). The snow albedo is larger with nonspherical than spherical snow particles (Picard et al. 2009). Kikuchi et al. (2013) develop the “global classification of snow crystals” from snow in situ measurements reported in the literature from midlatitudes to polar regions. In general, the representative snow particle shape varies with the particle size. On the basis of several studies including direct snow particle in situ measurements, we develop the SGHM model shown in Fig. 1. Snow particle habit fractions vary linearly with respect to the logarithmic maximum dimension ($\log D$) between particular size division points at $D = 100, 300, 1000$, and $3000 \mu\text{m}$.

In falling (i.e., fresh) snow in midlatitudes, needlelike dendrites are predominant (Kikuchi et al. 2013). From snow particle samples, newly accumulated snow has pristine needlelike and faceted particle shapes (A00; I18) or dendritic shapes (N01).

In the polar regions, fresh snow on the ground has a mixture of mostly column and platelike crystals with large aspect ratios and a smaller proportion of bullet cluster particles. The maximum dimension of columns and plates ranges from 2 to $1000 \mu\text{m}$, and the sizes of bullet clusters may be around several thousand microns (Ohtake and Yogi 1979; Walden et al. 2003). However, Walden et al. (2003) found some samples in Antarctica where quasi-spherical snow particles predominate with a size range from a few to several tens of microns. Under a condition of low snowfall intensity, snowflakes are well grown and have a dendritic shape with needlelike branches (I18).

As we described in section 1, the snow metamorphosis process plays a dominant role in the variations of snow

particle size and shape, after snow particles are accumulated on the surface (N01). In all regions, aged snow compared to fresh snow has significantly larger particle sizes and the former has faceted and granular particle shapes due to snow metamorphosis (I18).

From these reports, we assume snow particle habits to be mixtures of 1) SCOL with an aspect ratio of 0.25 (defined as the ratio of hexagonal column width $2a$ to length L ; i.e., $0 < 2a/L \leq 1$), 2) SBUL (Yang et al. 2013), and 3) DROX (Yang et al. 2003; Zhang et al. 2004). The single-scattering properties of plate particles with arbitrary inverse aspect ratios (i.e., $0 < L/2a \leq 1$) are similar to those of column-like particles with the corresponding aspect ratios (i.e., $2a/L$). Depending on the aspect ratio, SCOL can be a surrogate of a plate particle, which we do not consider in the SGHM model. To mimic the observed irregularity and variability of snow particles on the surface, we assume that the particles are severely roughened using the degree of surface roughness ($\sigma^2 = 0.5$) defined by Yang and Liou (1998). When the maximum dimension is less than $100 \mu\text{m}$, we consider a mixture of snow particles that has 90% SCOL and 10% DROX, which mimics a semispherical granular particle. From 100 to $300 \mu\text{m}$, the fraction of SCOL linearly decreases to 40%, and the fraction of SBUL increases to 20%. From 300 to $1000 \mu\text{m}$, the fraction of SBUL increases to 50%, while the fraction of SCOL decreases to 0%. From 1000 to $3000 \mu\text{m}$, the fraction of DROX increases to 100%, which mimics the snow particle metamorphosis that produces granular crystals. Thus, this newly developed SGHM model is microphysically consistent with snow particle observations.

b. Snow particle size distribution

The snow particle metamorphosis process causes a significant increase in snow particle size. It is a challenge to specify the PSD of aged snow particles throughout this process. Therefore, we empirically estimate the snow PSD from in situ snow measurements. A00 found that both the snow PSD and particle shape change to larger and more compact particles in 4 days after a snowfall. N01 investigated snow particle metamorphosis with laboratory experiments, demonstrating that needle-shaped snow particles became compact and granular particles, and the PSD is shifted to a larger size and with a broader PSD width, after exposure to solar radiation. I18 show accurate particle shapes of snow particles from natural accumulated snow and artificially generated snow including both new and aged snow samples, using an X-ray microtomography technique. Fresh snow particles have needlelike dendritic shapes, while aged snow particles have aggregated and granular

shapes. Thus, these studies suggest that snow particle metamorphosis causes the snow PSD to be broader with time.

To specify the PSD of snow particles, we first fit the PSDs to the gamma distribution using the 14 in situ measurement datasets from A00, N01, and I18. Those studies use several definitions of particle size bins, and we unify the size definitions by converting them to the maximum dimension D and effective radius r . The appendix presents a detailed description of these snow in situ measurement datasets and the gamma fitting method. In each sample of snow particles for each size definition, the gamma distribution with respect to the maximum dimension is

$$n(D) = N_0 \left(\frac{D}{v} \right)^{\kappa-1} \exp \left(-\frac{D}{v} \right), \quad (5)$$

where κ and v are the shape and scale parameters (Iwabuchi et al. 2012; Bi et al. 2014), and the gamma distribution with respect to the effective radius is

$$n(r) = N_0 r^{(1-3v_{\text{eff}})/v_{\text{eff}}} \exp \left(-\frac{r}{r_{\text{eff}} v_{\text{eff}}} \right), \quad (6)$$

where r_{eff} and v_{eff} are the effective radius and variance of the snow PSDs (Hansen and Travis 1974).

For the 14 datasets in A00, N01, and I18, the left and right panels of Fig. 2 show best-fit gamma distribution parameters in Eqs. (5) and (6), respectively. In each panel, the solid line depicts the linear fit to the 14 sets of parameters, and each panel shows a linear relationship for the PSDs of the available samples with respect to either the maximum dimension or the effective radius. Note that these snow PSDs have a PSD shape that is quite different from those associated with ice clouds. Specifically, the values of the parameters κ and v for contrails (Iwabuchi et al. 2012) and cirrus clouds (Bi et al. 2014) deviate from those for snow particles on the surface. The snow PSD empirical fitting line indicates a broader PSD width than in contrails or cirrus clouds, because the solid line has a higher value (a larger shape parameter κ) than either of the fitted values for cloud ice particles. This difference is obviously caused by different microphysical processes to form microphysical structures of ice crystals in clouds and in accumulated snow. The ice cloud PSD is determined by diffusional growth, collision, and sublimation in ice clouds, while the accumulated snow PSD is determined by diffusional growth, and metamorphosis associated with radiation, snowmelt, and refreezing. Note that a larger effective radius has a larger effective variance.

FIG. 2. Scatterplots of (left) shape and scale parameters for snow particle size distributions (PSD) with respect to the maximum dimension and (right) the effective radius and variance parameters for the snow PSD with respect to the effective radius. Solid lines depict the best fits of the parameters, and other lines depict previously developed PSD gamma distribution fits for contrails (dotted line; Iwabuchi et al. 2012) and cirrus clouds (dash-dotted line; Bi et al. 2014). Color plots indicate snow in situ measurements from N01 (pink), A00 (green), and I18 (purple).

c. Internally mixed BC in snow particles

The BC mass fraction in snow particles varies by several orders of magnitude among locations from usually $<1.0 \text{ ng g}^{-1}$ in Antarctica to sometimes $>1000 \text{ ng g}^{-1}$ in a polluted region such as the Tibetan Plateau (Zhang et al. 2018). The snow albedo reduction is defined as the difference between the snow albedo for pure snow particles and that for BC internally mixed snow. Figure 3a shows snow albedo reductions associated with internally mixed BC over visible and near-infrared wavelengths calculated with an adding-doubling radiative transfer model, assuming single-scattering properties of droxtals and a single snow layer.

The snow albedo reduction largely depends on snow particle size, and also on the wavelength. Figure 3b

demonstrates the effects of snow particle shape on snow albedo reduction for the same volume-equivalent sphere radius. This size definition assures the same amount of BC mixed internally in snow particles with different shapes. The snow albedo reduction depends on snow particle shape, which is primarily due to the difference in the single-scattering properties.

d. Bulk optical properties of the SGHM model

The microphysically consistent SGHM model is consistent with sampled snow particles. Figure 4 shows the bulk optical properties of the SGHM model. The single-scattering coalbedo is mainly determined by the internally mixed BC amount at visible wavelengths and by snow particle size at near-infrared wavelengths.

FIG. 3. Variation of snow albedo reductions associated with internal mixing of black carbon due to (a) snow particle size by assuming droxtal snow particles and (b) particle shape by assuming a volume-equivalent snow particle radius R_V of $100 \mu\text{m}$ and a BC mixing ratio of 1000 ng g^{-1} . Each calculation is made for monodispersed particles (i.e., uniform particle size).

FIG. 4. The optical properties of the droxtal (DROX), solid hexagonal column (SCOL), and snow grain habit mixture (SGHM) model. (left) Single-scattering coalbedo and (right) asymmetry parameter for (top) a visible wavelength ($0.5 \mu\text{m}$) and (bottom) a near-infrared wavelength ($1.25 \mu\text{m}$).

The particle shape (with the same particle volume) has an insignificant impact on the single-scattering coalbedo. In other words, the impacts of particle shape on the single-scattering coalbedo are not negligible under the condition with the same snow particle effective radius.

I18 demonstrates that the asymmetry parameter of snow particles decreases with increasing volume-equivalent spherical radius at both wavelengths 0.532 and $1.242 \mu\text{m}$. The SGHM computes similar changes in the asymmetry parameter at a visible wavelength ($\lambda = 0.5 \mu\text{m}$). However, single snow particle shape assumptions (i.e., all DROX or all SCOL) do not lead to a similar decrease. Therefore, the mixture of multiple snow particle shapes is necessary to optimally simulate the bulk optical properties of natural snow particles. For a near-infrared wavelength ($\lambda = 1.25 \mu\text{m}$), the SGHM does not compute a similar decrease with a volume-equivalent sphere radius larger than $200 \mu\text{m}$. This implies that the particle shape assumed in the corresponding size domain (i.e., mostly DROX) may not be appropriate, or the near-field interaction among snow particles may be important for the single-scattering properties in densely packed snow. In these

aspects, further investigations are needed and left for future work.

4. Rigorous snow albedo simulations

This section first compares the spectral snow albedo model based on the SGHM with an existing snow albedo model and direct snow albedo observations, and then presents sensitivity studies.

a. Comparisons with reliable simulations and observations

To evaluate the performance of the spectral snow albedo model based on the SGHM, we compare the spectral snow albedo simulations with those from the single-layer SNICAR model (Flanner et al. 2007) and observations in Antarctica (Grenfell et al. 1994). The SNICAR (<http://snow.engin.umich.edu/info.html>) model computes snow albedo based on the two-stream approximation and a spherical particle assumption. Figure 5 compares spectral snow albedo for a BC mass fraction of 0 ng g^{-1} (pure snow) and an SZA of 50° . The SNICAR calculation assumes a snow particle effective radius of

FIG. 5. Comparison of the two-layer-based spectral snow albedo simulations in this study with the SNICAR model (Flanner et al. 2007) and observations in Antarctica (Grenfell et al. 1994). The inset is a magnification of the top-left portion of the main panel.

80 μm , which achieves the best fit to the Grenfell et al. (1994) observations based on a single-layer assumption. Grenfell et al. (1994) provide particle size statistics for several snow depth bins from snow in situ measurements, indicating that average snow particle radius just below the surface snow is 52 μm , and the size sharply increases toward a snow depth of 10–100 mm, ranging from 122 to 191 μm [see Table 4 in Grenfell et al. (1994) for more details]. Based on their datasets, our model assumes SWE in the top layer to be 0.05 mm, and the top and second snow layer have effective particle radii of 52 and 160 μm , respectively. The results show that the present snow albedo calculation based on SGHM is consistent with both the single-layer SNICAR model and observations in a spectral region

from the visible to 1.4 μm , and outperforms SNICAR with $\lambda > 2.0 \mu\text{m}$ due to taking account of the two-layer snow surface and variation of snow particle habits in this study. These are promising results for the SGHM because spectral snow albedo simulations are consistent with observed snow particle size and snow-layer characteristics. Note that underestimated snow albedo by the SGHM compared to observations for $\lambda = 1.5\text{--}1.9 \mu\text{m}$ may come from uncertainty in the refractive index of ice compiled by Warren and Brandt (2008), which is further discussed in Carmagnola et al. (2013).

b. Sensitivity of snow characteristics to snow albedo

Figure 6 shows the sensitivity of the snow albedo to snow particle effective radius and internally mixed BC mass fractions, assuming a single snow layer and the SGHM bulk optical properties. As demonstrated previously (e.g., A00; Flanner et al. 2007; He et al. 2018c), a larger snow particle effective radius reduces the snow albedo in the infrared spectral region, while BC internal mixing reduces the snow albedo in the visible region. The magnitude of snow albedo reduction due to BC internal mixing becomes larger with a larger snow particle effective radius.

Figure 7 demonstrates the impact of SWE in the top snow layer on snow albedo under various two-layer snow conditions. The snow particle effective radius is assumed to be 30 μm for the top layer and 500 μm for the second layer, and the second layer is thick enough to be opaque. In the spectral region of 0.2–0.8 μm , the snow albedo does not change with SWE in the top layer of pure snow due to weak ice absorption, but increases as the top-layer SWE increases with polluted snow

FIG. 6. Snow albedo variations associated with (a) snow particle size and (b) BC mass fraction. All calculations assume a single opaque snow layer and an SZA of 45°.

FIG. 7. (a) Pure snow albedo variations associated with SWE by assuming a snow particle effective radius of $30\text{ }\mu\text{m}$ for the top snow layer and $500\text{ }\mu\text{m}$ for the second snow layer. The second layer is thick enough to be opaque. (b) As in (a), but comparing pure snow (solid lines) and snow with 1000 ng g^{-1} of BC (dashed lines). All calculations assume a solar zenith angle of 45° .

(Fig. 7b). In the spectral region of $0.8\text{--}1.5\text{ }\mu\text{m}$, solar radiation penetrates into the top snow layer, and even a small SWE variation plays an important role in snow albedo when the top-layer SWE $< 1\text{ mm}$. Nolin and Dozier (2000) investigate penetration depth of solar light into a single snow layer. At wavelength $\lambda = 1.03\text{ }\mu\text{m}$, the penetration depth (defined by $1/e$ of solar intensity) is 1.2 mm of SWE for an effective radius of $50\text{ }\mu\text{m}$, and the penetration depth increases with increasing snow particle effective radius. Aoki et al. (2011) similarly shows that solar radiation penetration depth increases with larger snow particle effective radius, and decreases with increasing snow impurities. In

the spectral region of $\lambda > 1.5\text{ }\mu\text{m}$, solar radiation does not penetrate far into the top snow layer, and the snow albedo is almost identical to the albedo of the top layer when top-layer SWE $> 0.1\text{--}0.3\text{ mm}$. A typical snow cover has a vertically inhomogeneous structure with smaller snow particle effective radii in the upper layers. These results show that a two-layer snow-cover model (fresh and aged snow) is sufficient to simulate reliable spectral snow albedo at both visible and infrared wavelengths. This finding is consistent with the result of Dang et al. (2017).

Figure 8 illustrates snow albedo variations associated with SZA. In the infrared, the sensitivity of snow albedo

FIG. 8. (a) Pure snow albedo variations associated with SZAs for a single snow layer with a snow particle effective radius of $100\text{ }\mu\text{m}$. (b) As in (a), but comparing pure snow (solid lines) and snow with 1000 ng g^{-1} of BC (dashed lines). In both cases, the snow particle effective radius is $500\text{ }\mu\text{m}$.

FIG. 9. Snow albedo enhancement associated with SZAs for different SWE values in the top snow layer. All calculations assume a snow particle effective radius of $30\ \mu\text{m}$ for the top snow layer and $500\ \mu\text{m}$ for the second snow layer. The second snow layer is thick enough to be opaque.

to SZA is similar to the sensitivity to snow particle effective radius (Marshall 1989), but the former sensitivity is stronger at visible wavelengths. In this spectral region, a large SZA corresponds to a large snow albedo due to increased contribution by the side scattering of snow particles to the reflected solar radiation. Moreover, BC internal mixing enhances the snow albedo in the visible spectral region when SZA is large.

Figure 9 shows the snow albedo enhancement under two-layer snow conditions. Here snow albedo enhancement is defined as the snow albedo minus that with an SZA of 0° . The snow particle effective radius is $30\ \mu\text{m}$ for the top layer and $500\ \mu\text{m}$ for the second layer. The results show that the snow albedo enhancement is almost identical for the cases with top-layer SWEs of 0.01 and 1.0 mm in the infrared spectral region, and has a similar magnitude at visible to near-infrared wavelengths. This implies that snow particle optical properties in a thin top layer dominate the snow albedo enhancement when SZA is large.

5. Snow albedo parameterizations

This section describes mathematical formulas for single- and two-layer snow albedo parameterizations, and presents the evaluation of these parameterizations.

a. A single-layer snow albedo parameterization

This study first develops a snow albedo parameterization for a single snow layer in the six LaRC-FL RTM bands in the shortwave spectral region. Each parameterization is

TABLE 2. Size relationships (μm) among snow particle effective radius R_e , volume-equivalent sphere radius R_V , and projected-area-equivalent sphere radius R_A based on the SGHM model.

R_e	R_V	R_A
50	46.10	44.27
100	84.05	77.06
150	122.55	110.77
200	161.50	145.13
250	200.02	178.91
300	237.58	211.42
350	273.58	242.26
400	308.79	271.31
450	342.60	298.93
500	375.45	325.34
600	439.03	375.55
700	500.83	423.62
800	561.66	470.61
900	622.10	517.21
1000	682.52	563.86

based on a nonlinear regression fit to rigorously calculated snow albedo datasets using a large number of cases. The variables in the parameterization include snow particle effective radius R_e (30 – $1000\ \mu\text{m}$), internally mixed BC mass fraction C_{BC} (0 – $1000\ \text{ng g}^{-1}$), and the cosine of SZA μ_0 (0.2 – 1). The optical thickness of the snow layer is assumed to be 960 (i.e., semi-infinite). Some land-snow models use a different definition of snow particle size. Table 2 lists the relationships between R_e and those based on two other size definitions (R_V and R_A). Conversion of R_V or R_A to the equivalent R_e value may be necessary for implementing this proposed parameterization within the GCM framework. The snow albedo is nonlinear with respect to snow particle effective radius (e.g., Dang et al. 2015). Thus, the pure snow albedo $\alpha_{s,\text{pure}}$ is parameterized in the form

$$\alpha_{s,\text{pure}} = p_{s,0} + p_{s,1} \log_{10} R_n + p_{s,2} (\log_{10} R_n)^2, \quad (7)$$

where $p_{s,i}$ is a coefficient (see Table 3), R_n is defined as R_e/R_0 , and R_0 is assumed to be $30\ \mu\text{m}$. The first subscript s means a single snow layer, which could represent either fresh or aged snow. This parameterization provides a fairly accurate snow albedo, and the performance is well documented in the literature (Dang et al. 2015).

The snow albedo reduction associated with C_{BC} depends on the snow particle effective radius as shown in Fig. 8. Parameterization of snow albedo reduction is based on the following empirical equation, with coefficients $q_{s,i}$ similar to He et al. (2018c):

$$\Delta\alpha_s = q_{s,0} C_{\text{BC}}^{k_n}, \quad (8)$$

where

TABLE 3. Coefficients of the single-layer snow albedo parameterization for the six broad bands of the NASA Langley Research Center's modified Fu–Liou broadband radiative transfer model.

Coef	Band 1	Band 2	Band 3	Band 4	Band 5	Band 6
λ (μm)	0.2–0.7	0.7–1.3	1.3–1.9	1.9–2.5	2.5–3.5	3.5–4.0
$p_{s,0}$	$9.740\,90 \times 10^{-1}$	$8.878\,89 \times 10^{-1}$	$3.093\,39 \times 10^{-1}$	$1.269\,79 \times 10^{-1}$	$1.435\,95 \times 10^{-2}$	$1.069\,46 \times 10^{-2}$
$p_{s,1}$	$-8.794\,90 \times 10^{-3}$	$-9.055\,19 \times 10^{-2}$	$-2.833\,09 \times 10^{-1}$	$-1.775\,34 \times 10^{-1}$	$-1.072\,86 \times 10^{-2}$	$-1.027\,71 \times 10^{-2}$
$p_{s,2}$	$-1.696\,30 \times 10^{-2}$	$-5.558\,51 \times 10^{-2}$	$7.061\,40 \times 10^{-2}$	$6.664\,92 \times 10^{-2}$	$4.349\,45 \times 10^{-3}$	$4.090\,57 \times 10^{-3}$
$q_{s,0}$	$2.834\,08 \times 10^{-3}$	$2.390\,82 \times 10^{-4}$	—	—	—	—
$q_{s,1}$	$4.362\,22 \times 10^{-1}$	$6.653\,30 \times 10^{-1}$	—	—	—	—
$q_{s,2}$	$1.153\,58 \times 10^{-1}$	$6.750\,72 \times 10^{-2}$	—	—	—	—
$s_{s,0}$	$7.127\,07 \times 10^{-1}$	$1.021\,29 \times 10^0$	$1.363\,20 \times 10^0$	$1.717\,68 \times 10^0$	$1.144\,13 \times 10^0$	$1.810\,24 \times 10^0$
$s_{s,1}$	$-6.791\,94 \times 10^{-1}$	$-2.792\,70 \times 10^{-1}$	$4.105\,01 \times 10^{-2}$	$-6.312\,57 \times 10^{-2}$	$2.584\,22 \times 10^{-3}$	$-4.959\,00 \times 10^{-2}$
$s_{s,2}$	$7.330\,10 \times 10^{-1}$	$4.030\,19 \times 10^{-1}$	$1.205\,94 \times 10^0$	$2.136\,34 \times 10^0$	$2.832\,19 \times 10^{-1}$	$1.454\,00 \times 10^{+1}$
$s_{s,3}$	$-7.570\,24 \times 10^{-1}$	$-7.540\,92 \times 10^{-1}$	$9.303\,24 \times 10^{-1}$	$5.361\,67 \times 10^{-1}$	$5.342\,67 \times 10^{-1}$	$9.521\,50 \times 10^{-1}$

$$k_n = q_{s,1} R_n^{q_{s,2}}. \quad (9)$$

From Eqs. (8) and (9), we obtain a representative snow albedo $\alpha_{s,0}$ by

$$\alpha_{s,0} = \alpha_{s,\text{pure}} - \Delta\alpha_s. \quad (10)$$

Gardner and Sharp (2010) consider SZA effects on snow albedo parameterization with SZA greater than 60° . However, the nonlinear dependence of snow albedo enhancement on both snow particle effective radius and BC internal mixing does not allow a simple parameterization (Figs. 8 and 9). Therefore, we empirically fit the snow albedo enhancement $\alpha_{s,\text{enh}}$ using coefficients $s_{s,i}$, the cosine of SZA μ_0 , and the following equation:

$$\alpha_{s,\text{enh}} = l_n \left(\frac{1 - \mu_0}{1 + \mu_0} \right)^{s_{s,0}}, \quad (11)$$

where

$$l_n = s_{s,1} + s_{s,2} \alpha_{s,0}^{s_{s,3}}. \quad (12)$$

We obtain the single-layer snow albedo as a function of snow particle effective radius, BC mass fraction, and SZA in the form

$$\alpha_s = \alpha_{s,0} + \alpha_{s,\text{enh}}. \quad (13)$$

Overall, this single-layer snow albedo parameterization considers nonlinear variations associated with snow particle effective radius, BC internal mixing in snow particles, and SZA.

Figure 10 shows the validation of the single-layer snow albedo parameterization in the six LaRC-FL RTM bands in comparison with rigorous simulations based on radiative transfer calculations described in section 4, for various snow and SZA cases such as snow particle effective radii of 100–500 μm , BC mass fractions

of 0–1000 ng g^{-1} , and cosines of SZA of 0.2–1. In Fig. 10, each parameterization represents the band-mean snow albedo with a correlation coefficient greater than 0.98. The comparisons also show that snow albedo changes associated with SZAs are important in visible to near-infrared bands (i.e., 1 and 2) and even play a major role in snow albedo in bands 3–6.

Figure 11 compares snow albedo parameterizations between this study and He et al. (2018c), which is a function of snow particle effective radius and BC internal mixing fraction assuming spherical and Koch snowflake particle models. Because the He et al. parameterization uses a constant SZA of 49.5° , this figure does not consider SZA dependence. The parameterization in this study more closely resembles the Koch snowflake parameterization than the spherical parameterization due to similar particle asphericity in the SGHM and Koch snowflake models. Both He et al. parameterizations in Fig. 11a tend to underestimate snow albedo due to differences in snow particle shape and in the internally mixed BC PSD. The He et al. parameterization is based on a monodispersed (i.e., a uniform particle size) BC PSD with a radius of 0.1 μm , while our study assumes an effective radius and variance of 0.1 μm and 0.1, providing weaker absorption based on the bulk optical properties of snow particles (He et al. 2018b). Therefore, our parameterization has a higher snow albedo than that computed by He et al. (2018c). In bands 3–6 (Fig. 11b), the bias is basically small and mainly due to particle shape differences and uncertainty in nonlinear regression fits.

b. A two-layer snow albedo parameterization

We expand the parameterization to a two-layer case. The subscript t in following equations indicates a two-layer parameterization. The two-layer snow albedo parameterization is similar to the single-layer parameterization based on Eqs. (7)–(13), but coefficients

FIG. 10. Comparisons of snow albedo for (a) band 1 to (f) band 6 of the NASA Langley Research Center's modified Fu–Liou broadband radiative transfer model, between rigorous snow albedo simulations (referred to as “Truth”) and the single-layer snow albedo parameterization (referred to as “Paramet.”). The color scale denotes the cosine of SZA associated with each data point.

$p_{s,i}$, $q_{s,i}$, and $s_{s,i}$ are replaced by counterparts $p_{t,i}$, $q_{t,i}$, and $s_{t,i}$ (Table 4). The variables in the parameterization include snow particle effective radii for the top R_{e1} (30–500 μm) and second layers R_{e2} (100–1000 μm), SWE in the top layer w (0.01–1 mm), C_{BC} , and μ_0 . In addition, two modifications take into account the bulk optical

properties of snow in each layer. First, we replace the single-layer R_n with

$$R_n = \{1 - e^{-t_0[w/(R_{e1}\mu_0)]}\} \left(\frac{R_{e1}}{R_0} \right) + e^{-t_0[w/(R_{e1}\mu_0)]} \left(\frac{R_{e2}}{R_0} \right), \quad (14)$$

FIG. 11. Comparisons of band-average snow albedo based on single-layer snow albedo parameterizations between this study and He et al. (2018c) for (a) bands 1 and 2 and (b) bands 3–6 of the NASA Langley Research Center's modified Fu–Liou broadband radiative transfer model. Orange and blue colors show spherical and Koch snowflake shape-based parameterizations from He et al. (2018c).

TABLE 4. Coefficients of the two-layer snow albedo parameterization for the six broad bands of the NASA Langley Research Center's modified Fu–Liou broadband radiative transfer model.

Coef	Band 1	Band 2	Band 3	Band 4	Band 5	Band 6
λ (μm)	0.2–0.7	0.7–1.3	1.3–1.9	1.9–2.5	2.5–3.5	3.5–4.0
$p_{t,0}$	$9.667\,98 \times 10^{-1}$	$8.472\,33 \times 10^{-1}$	$3.173\,71 \times 10^{-1}$	$1.689\,31 \times 10^{-1}$	$1.674\,40 \times 10^{-2}$	$2.094\,81 \times 10^{-2}$
$p_{t,1}$	$1.463\,98 \times 10^{-2}$	$-3.691\,01 \times 10^{-2}$	$-2.801\,61 \times 10^{-1}$	$-2.329\,44 \times 10^{-1}$	$-1.464\,26 \times 10^{-2}$	$-2.933\,87 \times 10^{-2}$
$p_{t,2}$	$-2.069\,75 \times 10^{-2}$	$-6.607\,01 \times 10^{-2}$	$6.448\,11 \times 10^{-2}$	$8.155\,70 \times 10^{-2}$	$6.439\,31 \times 10^{-3}$	$1.271\,12 \times 10^{-2}$
$q_{t,0}$	$2.842\,75 \times 10^{-3}$	$2.228\,77 \times 10^{-4}$	—	—	—	—
$q_{t,1}$	$4.404\,91 \times 10^{-1}$	$6.465\,29 \times 10^{-1}$	—	—	—	—
$q_{t,2}$	$9.617\,21 \times 10^{-2}$	$6.693\,62 \times 10^{-2}$	—	—	—	—
$s_{t,0}$	$1.184\,22 \times 10^0$	$1.057\,79 \times 10^0$	$1.435\,39 \times 10^0$	$1.677\,88 \times 10^0$	$1.393\,09 \times 10^0$	$1.897\,28 \times 10^0$
$s_{t,1}$	$-1.957\,62 \times 10^0$	$-6.977\,43 \times 10^{-1}$	$3.487\,84 \times 10^{-2}$	$2.385\,03 \times 10^{-2}$	$-4.052\,57 \times 10^{-2}$	$1.360\,49 \times 10^{-2}$
$s_{t,2}$	$2.028\,00 \times 10^0$	$8.815\,07 \times 10^{-1}$	$8.771\,38 \times 10^{-1}$	$1.883\,33 \times 10^0$	$4.383\,28 \times 10^{-1}$	$1.003\,62 \times 10^4$
$s_{t,3}$	$-2.870\,68 \times 10^{-1}$	$-1.739\,58 \times 10^{-1}$	$8.350\,08 \times 10^{-1}$	$6.647\,71 \times 10^{-1}$	$3.934\,35 \times 10^{-1}$	$2.423\,50 \times 10^0$
$t_{t,0}$	$8.293\,27 \times 10^1$	$1.423\,19 \times 10^2$	$7.219\,56 \times 10^2$	$8.920\,34 \times 10^2$	$2.055\,85 \times 10^3$	$2.392\,00 \times 10^3$

where t_0 is a coefficient, R_{e1} and R_{e2} are the snow particle effective radii for the top and second snow layers, and w is SWE in the top layer. The second layer is opaque as in the previously discussed cases. The coefficient t_0 roughly represents absorptivity of the bulk snow optical properties in a given band, and therefore a larger t_0 value indicates a larger contribution of the snow optical properties of the top layer to the band snow albedo. Second, to calculate snow albedo enhancement associated with SZA, we replace Eq. (11) with

$$\alpha_{t,\text{enh}} = l_n \left(\frac{1 - \mu_0}{1 + \mu_0} \right)^{s_{t,0}}, \quad (15)$$

where

$$l_n = s_{t,1} + s_{t,2} \alpha_{s,0}^{s_{t,3}}. \quad (16)$$

Because the snow albedo enhancement associated with SZA is most sensitive to the optical properties of the top snow layer, we use the same representative snow albedo $\alpha_{s,0}$ [Eq. (10)] from the single-layer snow albedo parameterization for the top-layer optical properties (i.e., R_{e1} and C_{BC}). As in the single-layer snow albedo parameterization, the coefficients for the two-layer snow albedo parameterization are determined through nonlinear regression fits based on rigorous snow albedo calculations with various snow particle characteristics, SZA conditions, and top-layer SWEs.

Figure 12 compares snow albedo from the two-layer snow albedo parameterization with the rigorous calculations. Overall, the two-layer snow albedo parameterization agrees with rigorous calculations with a correlation coefficient of 0.9897.

It is also of interest to estimate the snow albedo bias when the two-layer snow structure is ignored. Figure 13 shows the snow albedo differences between the single- and

two-layer snow albedo parameterizations. We assume that the two-layer snow albedo parameterization provides an unbiased snow albedo, while the single-layer snow albedo parameterization provides two values of snow albedo models: one is based on snow optical properties from the top layer (i.e., without considering the second layer), and the other is based on those in the second snow layer (i.e., without considering the top layer). The snow accumulation depth d_{snow} of the top layer is assumed to be 2 cm with a typical fresh-snow density ($\rho_{\text{snow}} = 0.11 \text{ g cm}^{-3}$), which corresponds to SWE of 2.2 mm. Figures 13a and 13c indicate that

FIG. 12. Comparisons of band snow albedo for a two-layer snow surface between rigorous snow albedo simulations (referred to as “Truth”) and the two-layer snow albedo parameterization (referred to as “Paramet.”). The color of each data point denotes the band number of the NASA Langley Research Center’s modified Fu–Liou broadband radiative transfer model.

FIG. 13. Snow albedo differences associated with neglecting either the (left) top or (right) second snow layer, for each band of the NASA Langley Research Center's modified Fu–Liou broadband radiative transfer model, based on single-layer snow albedo parameterizations. Columns show cases with a snow particle effective radius of the top layer of (top) 80 and (bottom) 250 μm . The color of each data point indicates the snow particle effective radius of the second layer.

neglecting the top layer causes significant biases in the band snow albedo even for a 2-cm snow layer. However, neglecting the second snow layer provides less bias compared with neglecting the top layer, especially in infrared bands. In visible to near-infrared bands, the effects of neglecting the second layer is minimal for fresh snow since the corresponding optical thickness of the top snow layer with a small snow particle effective radius is large. However, as snow in the top layer metamorphoses into a larger snow particle effective radius, the corresponding optical thickness of the top layer decreases, and the optical properties of the second layer snow contribute more to the band snow albedo (Fig. 13d).

6. Summary and conclusions

This study developed a snow grain habit mixture (SGHM) model for snow particles in snow cover based on in situ snow measurements. The SGHM takes into account internally mixed BC in snow particles, snow particle habit fraction varying over the snow particle size domain, and snow PSDs depending

on snow particle effective radius. The SGHM-based snow optical and microphysical property simulations are consistent with observations at visible wavelengths. The asymmetry parameter decreases with an increase in snow particle effective radius (118) due to an increasing fraction of compact snow particles. Most previous studies assume a single snow particle shape over the entire size range, and are not capable of simulating the decrease of the asymmetry parameter with increasing particle sizes. The SGHM in conjunction with an adding–doubling radiative transfer model and observed snow microphysical property dataset (Grenfell et al. 1994) computes spectral snow albedos consistent with available observations.

Sensitivity studies reveal that when fresh snow covers an aged snow layer, top-snow-layer optical properties essentially determine the observed spectral snow albedo in the infrared spectral region. However, when the fresh snow is only approximately 1 cm deep or less ($\text{SWE} < 1 \text{ mm}$), small SWE variations in the top layer significantly change the snow albedo in the visible and near-infrared spectral region under polluted snow conditions. Spectral snow albedo is

TABLE A1. Snow in situ measurement datasets used for the gamma fitting. The size definition in the third column includes the projected-area-equivalent sphere radius R_A , maximum dimension D , or volume-equivalent sphere radius R_V . The particle shape in the fourth column includes a solid hexagonal column (SCOL) or droxtal (DROX).

Snow dataset ID	Information from original references Representative particle shape	Treatment in this study	
		Size definition	Particle shape
A00, Feb22	New snow + faceted crystal	R_A	SCOL
A00, Feb23	Faceted crystal + fine-grain old snow + depth hoar	R_A	SCOL
A00, Feb24	Faceted crystal + fine-grain old snow	R_A	DROX
A00, Feb25	Fine-grain old snow + granular snow	R_A	DROX
N01, Fig. 5	Dendritic new snow	D	SCOL
N01, Fig. 8	Compacted snow	R_A	DROX
N01, Fig. 11	Granular snow	R_A	DROX
I18, Sample A	Fresh snow	R_V	SCOL
I18, Sample B	Fresh snow	R_V	SCOL
I18, Sample C	Granular snow	R_V	DROX
I18, Sample D	Granular snow	R_V	DROX
I18, Sample E	New snow	R_V	SCOL
I18, Sample F	New snow	R_V	SCOL
I18, Sample G	Compacted snow	R_V	DROX

enhanced at large SZAs, and the magnitude of the snow albedo enhancement is a function of top-layer snow particle effective radius and internally mixed BC mass fraction.

We developed single-layer and two-layer snow albedo parameterizations for GCM applications, in six bands used in LaRC-FL RTM by regression fitting to rigorous snow albedo calculations based on the SGHM model. The single-layer snow albedo parameterization is a function of the snow particle effective radius, internally mixed BC mass fraction, and SZA. The two-layer snow albedo parameterization considers snow particle effective radii for a top (fresh snow) and second (aged snow) layer, SWE in the top layer, an internally mixed BC mass fraction in the two layers, and SZA. Both the single-layer and two-layer parameterizations provide consistent broadband snow albedo compared to those from rigorous calculations, achieving correlation coefficients close to 0.99 for all bands. Note that applying these parameterizations to an extreme snow condition that is not included in the development of these parameterizations (e.g., C_{BC} far larger than 1000 ng g^{-1}) may cause biases.

Comparing our single-layer snow albedo parameterization with the He et al. (2018c) parameterization based on sphere and Koch snowflake models, the present parameterization leads to higher snow albedos because the SGHM assumes that internally mixed BC and snow particles have a gamma PSD (note that He et al.'s BC and snow particles have uniform sizes), and the two parameterizations assume different snow particle shapes. Theoretical calculations reveal that with polluted snow, the snow albedo in visible wavelengths is sensitive to the PSD of

internally mixed BC (He et al. 2018b) and is also sensitive to particle shape (He et al. 2018a). There is still a large gap in the understanding of the mixing state of black carbon in snow. Although this study considers BC in snow as 100% internally mixed, actual polluted snow particles may have both internal and external BC mixing. The absorption of externally mixed BC in snow is 50% of that of internally mixed BCs (e.g., Flanner et al. 2012), and a two-layer snow albedo parameterization based on 100% external mixing may also be useful and should be developed in the future. To improve snow property parameterization, it is necessary to carry out extensive in situ measurements of snow particle morphological characteristics. In particular, more direct snow in situ measurements targeting the BC mixing state will be helpful.

Sophisticated GCM snow–land surface schemes can resolve up to several snow layers, such as the Simplified Simple Biosphere, version 3 (SSiB-3; Xue et al. 1991, 2003), and a modified snow scheme in Noah (Saha et al. 2017). These schemes have been incorporated into regional climate models (Oaida et al. 2015). As shown in Fig. 13, ignoring a snow layer of few centimeters can induce nonnegligible biases in computed snow albedo, particularly in visible to near-infrared bands from 0.2 to $1.3 \mu\text{m}$. The two-layer snow albedo parameterization in conjunction with a snow metamorphosis scheme (e.g., Niwano et al. 2012), or two-layer snow microphysical property retrievals based on satellite measurements (e.g., Jin et al. 2008) may be used to further reduce uncertainty in snow albedo associated with realistic snow layers and improve surface radiation budget estimations in GCMs.

Acknowledgments. This work was partly supported by the NASA Clouds and the Earth's Radiant Energy System Project (Grant NNX17AB03G) and partly by the endowment funds related to the David Bullock Harris Chair in Geoscience at the College of Geosciences, Texas A&M University (Grant 02-512231-00001). Light-scattering computations and radiative transfer calculations were conducted at Texas A&M Supercomputing Facilities, and the authors gratefully acknowledge facility staff for their help and assistance. The authors thank Dr. Guanglin Tang for help with computing the optical properties of snow particles.

APPENDIX

PSD Fitting with Snow In Situ Measurement Datasets

In the development of the PSD fit based on the gamma distribution, we use in situ measured PSD datasets from **A00** (four datasets; see their Fig. 4), **N01** (three datasets; their Figs. 5, 8, and 11), and **I18** (seven datasets; their Fig. 5a). We manually read the number of particles for each bin from these figures. Although manual reading causes some uncertainty in snow PSDs, we assume that the uncertainty is small and does not induce a significant error in fitting of snow PSDs with the gamma distribution. Second, we unify the definition of the size bins among the 14 samples. **A00** defines the projected-area-equivalent sphere radius as the snow particle size, and **I18** uses the volume-equivalent sphere radius. In **N01**, the size definition differs among samples. For dendritic snow (Fig. 5 in **N01**), the longest column branch is defined as the diameter, and we treat the size definition as the maximum dimension. For other compact and granular particles (Figs. 8 and 11 in **N01**), each grain is assumed to be spheroid and the particle size is defined as $(ab)^{1/2}$ where a and b respectively indicate the major and minor diameter of an spheroid. However, quantitative values a and b for each sample are unavailable in **N01**. Therefore, we treat the size definition based on the spheroid assumption as projected-area-equivalent sphere radius.

In the size conversion, the snow particle shape is assumed to be SCOL for new and fresh snow and DROX for compacted, granular, and aged snow, as diagnosed in **A00**, **N01**, and **I18**. Table A1 lists the 14 snow particle sample datasets. In sample “**A00**, Feb24,” we assume the particle shape to be DROX rather than SCOL because corresponding snow particle images suggest that the representative aspect ratio of particles seems to be close to unity. With these datasets and the method described in section 3b [Eqs. (5) and (6)], we perform a

FIG. A1. Snow particle size distributions in terms of the maximum dimension for each snow in situ measurement obtained manually from Fig. 4 in **A00**. The locations of the symbols give the number of snow particle samples for each size bin. The colors of the symbols indicate snow dataset ID (for each day). The solid lines are the gamma distribution fits according to Eq. (5).

PSD fitting with each snow in situ measurement dataset. Figure A1 shows example PSD fittings based on the maximum dimension size definition using four **A00** samples. The gamma distribution fits the snow PSD well. We perform the gamma distribution fitting for all snow in situ measurement datasets for each size definition (i.e., the maximum dimension and particle effective radius), which are used for snow PSD analysis as shown in Fig. 2.

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