ADEOS-II/GLI snow/ice products — Part II: Validation results using GLI and MODIS data

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Abstract

For the validation of snow/ice products of the Advanced Earth Observing Satellite-II/Global Imager (ADEOS-II/GLI), several field campaigns were performed for various types of snow conditions with the Moderate Resolution Imaging Spectroradiometer (MODIS) and GLI overpasses at four sites in Alaska and eastern Hokkaido, Japan from 2001 to 2005. The target satellite-derived snow parameters are snow surface temperature, mass fraction of soot, and two types of snow grain size retrieved from different spectral channels. The retrieved satellite products were compared with in-situ measured snow parameters based on snow pit work and snow sampling. The satellite-derived snow surface temperatures agreed well with in-situ measured values with a correlation coefficient (Rc) of 0.900 and a root-mean-square error (RMSE) of 1.1 K. The satellite-derived mass fractions of soot were close to in-situ measured mass fractions of snow impurities for the snow layer between the surface and down to 7 or 10 cm rather than between 0 and 2 cm, while the satellite-derived absolute values were lower than the in-situ measured ones (Rc=0.506 and RMSE=5.0 parts per million by weight (ppmw)). This discrepancy is due primarily to the difference in the composition of snow impurities assumed in the satellite algorithm (soot) and measured in-situ (mineral dust) suggesting that the satellite retrieval of soot is not producing soot concentrations in many cases but rather dust. Snow grain sizes retrieved from two satellite channels at λ=0.460 and 0.865 μm had better accuracy (Rc=0.840 and RMSE=125 μm) than those from a satellite channel at λ=1.64 μm (Rc=0.524 and RMSE=123 μm) from the comparison with simply depth-averaged snow grain size. When similar comparisons are made with the depth-averaged measured grain size by a 1/e weighting using flux transmittance, the results for Rc and RMSE are not improved due to some difficulties in calculating the depth-averaging by a 1/e weighting. For all our satellite products, the possible causes of errors are (1) satellite sensor calibration and (2) the bidirectional reflectance model (directional emissivity model for surface temperature) used in the algorithm together with the atmospheric correction. Two ways to improve the in-situ measurements are (1) the representativeness of the measured values and (2) the measuring methods. Field measurements also indicated that the
increased reflectance due to “sun crust” observed at wet snow surfaces under clear sky could cause an underestimation of satellite-derived snow grain size. This problem will be more severe for the grain size retrieved from the channel at $\lambda = 1.64\, \mu m$.

Keywords: Snow grain size; Snow impurity; Snow surface temperature; GLI; MODIS

1. Introduction

Parts I and III of this series of papers on ADEOS-II/GLI snow/ice products (Hori et al., 2007-this issue; Stamnes et al., 2007-this issue) presented the theoretical basis and the retrieved results for our satellite products including: cloud masking over snow/ice surface, surface classification, snow surface temperature, mass fraction of soot contained in the snow, and two types of snow grain sizes. The first two products are generally necessary procedures whenever snow/ice extent and snow albedo are retrieved from satellite data. The two types of snow grain sizes are retrieved from different spectral channels at which the penetration depths of light are different and thus the weighting functions on vertical profiles of snow grains are also different. We report on the results of field campaigns for the four snow products retrieved from the GLI data except cloud masking and surface classification.

Snow surface temperature is a basic and important parameter, which is used for monitoring of global change in the cryosphere in addition to its general use as an initial value in numerical weather prediction models together with other meteorological factors. The mass fraction of soot, which is the most absorptive natural material contained in snow, is our satellite product, but other types of snow impurities (water-insoluble solid particles) are potentially contained in natural snow and thus we generally refer to this parameter as “mass fraction of snow impurities.” Typical snow impurities are soot (black carbon) and mineral dust. The difference in the imaginary part of the refractive index between them is a factor of one hundred (Hess et al., 1998; Wiscombe & Warren, 1980). The snow impurity concentration is one of the most important snow parameters, which mainly affects the visible albedo (Warren & Wiscombe, 1980). Snow grain size is another very important snow physical parameter that mainly affects the near-infrared albedo (Wiscombe & Warren, 1980). Therefore, these two snow parameters essentially control the snow albedo for a snow layer of sufficient physical thickness to be optically opaque, although the albedo, in contrast to the bidirectional reflectance, is an apparent optical property that also depends on atmospheric effects (Aoki et al., 1999) and solar illumination angle. The effects of these snow physical parameters on albedo were quantitatively confirmed from spectral snow albedo measurements and snow pit work in Barrow, Alaska (Aoki et al., 1998) and in eastern Hokkaido, Japan (Aoki et al., 2000) and from long-term radiation budget measurements with frequent snow pit work in different types of snow-covered areas in Japan (Aoki et al., 2003, 2006; Motoyoshi et al., 2005). Liquid water in the snow can also affect the albedo. However, the enhanced liquid water content would increase the optically equivalent snow grain size because water is optically very similar to ice in the visible region (Wiscombe & Warren, 1980). Warren (1982) predicted the possibility of remote sensing of these snow parameters (surface temperature, snow impurity, and grain size) from their impact on snow albedo (reflectance). These snow parameters could be used not only to monitor climate change in the cryosphere but also to improve snow albedo sub-models in land-surface process modules of climate models (Aoki et al., 2003).

Remote sensing of snow grain size and snow impurities was started around 1980 and validation experiments have been performed since the 1990s. Using Kubelka-Munk theory, Sydor et al. (1979) attempted to detect the dust content of snow from Landsat-1 and 2 images over lake ice from the snow reflectance reduction due to dust contained in the snow. However, no direct validation was made for dust content in terms of its mass fraction. Dozier et al. (1981) concluded that it is potentially possible to retrieve snow grain size from satellite data using the Wiscombe and Warren (1980) snow reflectance model and NOAA/AVHRR images. Dozier and Marks (1987) developed a snow classification method based on a “two-stream” radiative transfer model together with an empirical relationship between snow grain size and single scattering parameters. They classified the snow surface into new snow, fine-grained snow, older metamorphosed snow, and vegetation over the snow cover from Landsat Thematic Mapper (TM) data over the southern Sierra Nevada, but they made no ground truth observations with the satellite overpass. Carlson et al. (1992) used the ratio of bidirectional reflectance at the wavelengths $\lambda = 1.5$ and 1.7 $\mu m$ to map the snow grain size over Antarctica from Galileo spacecraft flyby data. They reported that the retrieved results were consistent with ground truth data obtained at South Pole and Vostok, although the spatial resolution of the retrieved result was low. Bourdelle and Fily (1993) also used the Wiscombe and Warren (1980) model to retrieve grain size over Terre Adélie, Antarctica from Landsat TM4, TM5, and TM7. Although they made no ground truth observations, they reported that TM7 at longer wavelengths (TM4 at shorter wavelengths) is sensitive to the upper surface grain effective size (grain size in a deeper 10 cm snow layer).

Nolin and Dozier (1993) developed an algorithm for grain size retrieval using an exponential relationship between spherical snow grain size and bidirectional reflectance calculated with a discrete ordinate radiative transfer (DISORT) model (Stamnes et al., 1988, 2000) and Mie theory. Applying this algorithm to Airborne Visible/Infrared Imaging Spectrometer (AVIRIS) data at $\lambda = 1.04\, \mu m$ over Tioga Pass area and Mammoth Mountain, they found good agreement between retrieved grain sizes and those measured from snow samples for solar zenith angles to sloped snow surface less than 30°. Fily et al. (1997) also used a bidirectional reflectance model based on DISORT and the Mie theory to retrieve snow grain size from Landsat data over the
French Alps. They showed that the retrieved snow grain size from TM4 was close to the in-situ measured one, while those from TM5 and TM7 were much smaller than the field measurements. Using ratios between channels of a single channel gave only a qualitative estimate of the grain size. Nolin and Dozier (2000) developed a new algorithm for grain size retrieval based on the idea that in spectral snow reflectance the scaled band area of absorption by ice centered at $\lambda = 1.03 \mu m$ depends on snow grain size. They stated that this method is insensitive to illumination effects from topography and solar geometry although there are some limitations such as the needs for clean snow and hyperspectral data. They applied this technique to AVIRIS data but also to field spectrometer data of snowpacks in California, Greenland, Wisconsin and Colorado. From the field data, they obtained a very high correlation coefficient of 0.997 between model-derived grain radii and in-situ measured values for the range of grain size from 100 to 700 $\mu$m. Hori et al. (2001) showed preliminary results of retrieved snow grain size and mass fraction of soot in the Arctic from multiple passes of MODIS data using a two-channel method at visible ($\lambda = 0.469 \mu m$) and near-infrared ($\lambda = 0.858 \mu m$) wavelengths. Their algorithm is basically the same as the one used in the present study, which is based on DISORT and Mie theory for spherical ice particles (Stamnes et al., 2007-this issue).

The vertical variation of snow grain size was investigated by using multiple channels of near and middle infrared wavelengths of AVIRIS data obtained over the Arctic Ocean (Li et al., 2001) and from Airborne MultiSpectral Scanner (AMSS) data over the flat snow-covered farmland in eastern Hokkaido, Japan (Tanikawa et al., 2002). Both studies concluded that the shorter wavelength channel conveys snow grain size information from deeper layers than the longer wavelength channels due to the spectral variation of the light penetration depth as demonstrated by Wiscombe and Warren (1980), Tanikawa et al. (2002) also retrieved the mass fraction of snow impurities and compared it with ground truth data. Their retrieved results, assuming mineral dust as the snow impurities, were consistent with in-situ measured mass fraction of snow impurities. Painter et al. (2003) used a bidirectional reflectance model based on DISORT and Mie theory, but their approach to retrieving snow grain size is different from the other works, because it consists of a spectral mixture analysis based on endmembers of reflectance for different snow grain sizes. The value of the root-mean-square error between AVIRIS data over Mammoth Mountain and ground truth data was between 48 and 74 $\mu$m for the range of grain size from 100 to 700 $\mu$m. Kay et al. (2003) used the multispectral MODIS/ASTER Airborne Simulator (MASTER) data over Mt. Rainer in summer to retrieve the snow grain size, snow impurities and surface temperature. The snow grain size was estimated by comparing the MASTER spectra with modeled hemispherical-directional reflectance factor (HDFR) calculated by DISORT and Mie theory, and with the laboratory measured relationship between reflectance and snow grain size. The MASTER-derived snow grain sizes were biased from the ground truth data, but clear relationships among the snow parameters were reported. Green et al. (2006) estimated, from imaging spectrometer data, the abundance of the three phases of water in an environment that includes melting snow. Their analysis was based on the spectral shift in the absorption coefficient between water vapor, liquid water, and ice at 940, 980, and 1030 nm respectively. They applied a spectral fitting algorithm that infers the abundance of the three phases of water from AVIRIS data over Mount Rainier, and they showed that the water and ice abundances are related to the amount of liquid water and the sizes of the ice grains in the near-surface layer.

The previous studies mentioned above have many points in common. Many of them employed DISORT as the radiative transfer model and Mie theory for single scattering by assuming spherical shapes of the snow grains. Some of the satellite data or airborne data to which the algorithms were applied were taken over mountainous regions and thus topographic features might affect the retrieved results. The snow grain size retrieved from channels at wavelengths $\lambda < 1.4 \mu m$ would be affected by snow impurities if this effect was ignored in the algorithm. Our algorithms for retrieval of grain size and snow impurity concentrations also employ DISORT and Mie theory. It is thus necessary to validate the accuracy of our products on a flat snowfield. We conducted several field validation (ground truth) campaigns at four sites in Alaska and eastern Hokkaido, Japan from 2001 to 2005 mainly for snow temperature, mass fraction of snow impurities, and snow grain size in conjunction with MODIS and GLI overpasses. In this paper we report on the results of these field campaigns for satellite-derived snow surface temperature, mass fraction of soot, and two types of snow grain sizes.

2. Validation experiments

2.1. In-situ measurements

Snow parameters were obtained in-situ from snow pit work dug for each measurement, in which snow type, temperature, density, and snow grain size in each snow layer were measured together with snow sampling for the measurements of snow impurities. Snow surface temperatures were measured with a thermistor thermometer directly in the topmost layer. To avoid solar heating, we shaded only the detector part of the thermistor with snow cutter in the shortest possible time. As a definition of snow grain size, Grenfell and Warren (1999) proposed a radius of equal volume-to-surface-area sphere to represent nonspherical particle size. However, it was very difficult to measure both volume and area of snow particles in the short time. We operationally measured three types of dimensions for snow grain size with a handheld lens by the method described by Aoki et al. (1998, 2000, 2003). Fig. 1 indicates those three types of dimensions for the definition of snow grain size: $d_1$, the length of the major axis of crystals or dendrites; $d_2$, the branch width of dendrites or the width of a narrower portion of broken crystals; and $d_3$, the thickness of plate-like crystals. For aggregated grains, the dimension of the cluster and each grain’s diameter were measured as $d_1$ and $d_2$. In this study, we used the parameter $r_1$, which is one-half the value of $d_j$, where $j=1, 2$, and 3. Aoki et al. (1998, 2000) concluded that the optically equivalent snow grain size was $r_2$ for new snow or faceted crystals (lightly compacted snow) from the spectrally detailed albedo...
measurements together with snow pit work. In the case of granular snow, the optically equivalent snow grain size can be recognized as \( r_2 \) from the relationships between broadband albedo and snow grain size presented in Aoki et al. (2003). We thus use the dimension \( r_2 \) as in-situ measured snow grain size for comparison with the satellite-derived snow grain size. In our snow pit work, the values of \( r_2 \) were recorded for the maximum value \( (r_{2\max}) \), minimum value \( (r_{2\min}) \), and mean value \( (r_{2\mean}) \) among the snow grains contained in the observed snow layers. We define \( r_2 = r_{2\mean} \) if not otherwise specified. Furthermore, the averaged value of \( r_2 \) over the snow layer from surface to the specified depth (distance from surface) was also calculated and it is shown by \( r_2 \). Similarly the averaged values for \( r_{2\max}, r_{2\min}, \) and \( r_{2\mean} \) are shown by \( r_{2\max}, r_{2\min}, \) and \( r_{2\mean} \). In addition to \( r_2 \), depth-averaging of snow grain size \( \left(\overline{r_2}\right) \) was made by a 1/\( e \) weighting using flux transmittance in each snow layer. This method is presented in Appendix A.

Snow impurities were filtered using a Nuclepore filter after melting of the snow samples by the same method as described by Aoki et al. (2000, 2003). The concentration of snow impurities was estimated by direct measurement of the weights of Nuclepore filters, before and after filtering, using a balance. After the filtering the Nuclepore filters were dried in a desiccator. We adjusted an amount of melted snow sample for filtration so as to ensure two digits in weighing accuracy. The accuracy of the measured mass fraction is estimated to be 0.1 ppmw. In this procedure we used a two-stage filtering system of Nuclepore filters with different pore sizes of 0.2 \( \mu \)m and 5.0 \( \mu \)m for snow samples from the surface to two snow depths (shallower layer: 0 to 2 cm and deeper layer: 0 to 7 cm or 10 cm). A two-stage filtering system could produce rough information regarding the impurity types (Aoki et al., 2003). For most snow samples, the majority of the impurities were collected on the filter with a pore size of 5.0 \( \mu \)m, which suggests that the main constituent was mineral dust particles. Since our observation sites were located near or on the land, local emission of mineral dust from a road or a snow-free area would contribute to this constituent of snow impurities. The other snow parameters obtained for snow pit work were snow temperatures, snow density, and snow hardness, which were identified for each snow layer (= same snow type).

2.2. Validation sites

Our field validation campaigns for snow parameters were performed at four sites located in the Saroma Lagoon, Lake Abashiri and Nakashibetsu areas all in Hokkaido, Japan and in Barrow, Alaska (Fig. 2 and Table 1). All sites are topographically flat. At the Saroma and Barrow sites, field measurements were made on five cross-shaped grid points with a 1 km grid interval synchronized with each satellite overpass. The Saroma site was snow-covered lagoon ice (Fig. 2a) and the field experiments were made in February 2001 and 2002. The surface snow conditions were dry types such as new snow and faceted crystals, and the lower layers consisted of depth hoar and granular snow. The snow depth ranged from 7 to 31.5 cm and the underlying sea ice conditions were flat and stable lagoon ice. The Barrow site was snow-covered tundra (Fig. 2b) and the field measurements were made in April 2003. The surface snow conditions were very similar to those at the Saroma site, and the lower layers consisted of depth hoar. The snow depth ranged from 21 to 33 cm. The Abashiri site was snow-covered lake ice (Fig. 2a) and the field experiments were made in February 2001 and 2002. The surface snow conditions were dry types such as new snow and faceted crystals, and the lower layers consisted of depth hoar and granular snow. The snow depth ranged from 7 to 31.5 cm and the underlying sea ice conditions were flat and stable lagoon ice. The Barrow site was snow-covered tundra (Fig. 2b) and the field measurements were made in April 2003. The surface snow conditions were very similar to those at the Saroma site, and the lower layers consisted of depth hoar. The snow depth ranged from 21 to 33 cm. The Abashiri site was snow-covered lake ice (Fig. 2a) and the field experiment was conducted in March 2004. The snow condition was wet granular snow with a depth of 17 cm. The Nakashibetsu site was flat, snow-covered farmland (Fig. 2a) and the field measurements were made in March 2004 and 2005. In 2004 the snow conditions were wet granular snow with sun crust and a depth ranging from 92 to 106.5 cm. In 2005, the surface snow condition was compacted snow and the lower layers consisted of depth hoar and granular snow. The snow depth was 79 cm. Cross calibration among...
eight sensors including GLI and MODIS was carried out using spectral reflectance measurements taken during the present study at the Barrow site in April 2003 (Nieke et al., 2004). The calibration coefficients estimated from the field campaign were employed in the GLI channels. Snow crystal pictures are shown in Hori et al. (2006), in which the spectral directional emissivity of snow and ice in the atmospheric window was measured at seven sites including the validation sites in this study.

3. Satellite data and algorithms

GLI aboard ADEOS-II was launched on December 14, 2002. However, unfortunately ADEOS-II stopped operating on October 25, 2003 due to a malfunction of the power supply system. The available GLI data are limited to only 7 months from April to October in 2003. We thus used not only GLI data but also the data of Terra/MODIS and Aqua/MODIS. The channel specifications of GLI and MODIS used in the study are listed in Table 2. The orbits of satellites we used are a 1030 LT descending node for ADEOS-II and Terra, and a 1330 LT ascending node for Aqua. The target satellite-derived snow parameters are snow surface temperature ($T_s$) retrieved from two GLI channels 35 (10.8 $\mu$m) and 36 (12.0 $\mu$m) or the MODIS channels 31 ($\lambda = 11.0$ $\mu$m) and 32 ($\lambda = 12.0$ $\mu$m), snow grain radius ($R_{d,9}$) retrieved from two GLI channels 5 ($\lambda = 0.460$ $\mu$m) and 19 ($\lambda = 0.865$ $\mu$m) or the MODIS channels 3 ($\lambda = 0.469$ $\mu$m) and 2 ($\lambda = 0.858$ $\mu$m), snow grain radius ($R_{s,9}$) retrieved from the GLI channel 28 ($\lambda = 1.64$ $\mu$m) or the MODIS channel 6 ($\lambda = 1.64$ $\mu$m), and the mass fraction of soot contained in the snow ($C_s$) which is retrieved together with $R_{d,9}$. In this paper, satellite-derived snow parameters are expressed by capital letter variables and in-situ measured parameters by small letter variables. Although a correlation coefficient is usually indicated by ‘r’ or ‘R’, we use ‘Rc’ for a correlation coefficient between satellite-derived snow parameters and in-situ measurements to avoid confusion with snow grain radius.

Detailed descriptions of the algorithms for $T_s$, $C_s$, $R_{d,9}$, and $R_{s,9}$ are presented in Part I of this series of papers (Stamnes et al., 2007-this issue). However, we briefly describe for readers of this paper only the algorithms as well as the GLI wavelengths used in the algorithms. At first, cloud masking was performed with the GLI cloud masking algorithm (Stamnes et al., 2007-this issue), which is based on the brightness temperature difference between two channels at $\lambda = 3.715$ $\mu$m and 10.8 $\mu$m. Additionally, it uses the channel at $\lambda = 1.38$ $\mu$m for thin cirrus detection during daytime, four thermal channels ($\lambda = 3.715$, 8.60, 10.8, and 12.0 $\mu$m) for cloud detection during night, and two channels ($\lambda = 6.70$ and 10.8 $\mu$m) for surface temperature inversions to confirm clear sky condition in the polar regions. To the identified clear pixels a surface classification is made using the visible and infrared channels ($\lambda =$0.545, 0.678, 0.865, 1.05 and 10.8 $\mu$m). The approach to retrieve $T_s$ uses two “split-window” infrared channels at $\lambda = 10.8$ $\mu$m and 12.0 $\mu$m, which is commonly employed for sea surface temperature retrieval. The equation is expressed as a linear combination of measured brightness temperatures of those channels and satellite viewing angle. The mass fraction of soot and the snow grain sizes are retrieved with a lookup-table (LUT) method. To relate the reflectance of snow to grain size and soot fraction, the LUTs have been calculated with a radiative transfer model (DISORT) and Mie theory for single soot particles that are externally mixed with the snow particles. In these LUTs, simulated radiances at the top of the atmosphere (TOA) in the channels at $\lambda = 0.460$ $\mu$m and 0.865 $\mu$m have been calculated as a function of the snow grain radius between 50 and
2000 µm and the mass fraction of soot ranging from 0.02 to 2.5 ppmv. Since the channel at $\lambda=1.64$ µm is not sensitive to snow impurities, simulated radiances at TOA has been calculated as a function of the snow grain radius for the same range as mentioned above. Dependencies of the radiances on solar zenith angle ($\theta_s$), satellite zenith angle ($\theta_v$), and relative azimuth angle ($\Delta\phi$) between the sun and the sensor, are also taken into account in the tables. Retrievals of $C_s$ and $R_{g,0.9}$ are simultaneously made from the two satellite channels at $\lambda=0.460$ µm and 0.865 µm and the values of $R_{s,1.6}$ are retrieved only from the channel at $\lambda=1.64$ µm. Although an atmospheric correction is incorporated with each LUT by choosing the aerosol model from the GLI channel at $\lambda=0.38$ µm, we adopted a fixed rural aerosol model in this study for simplicity and because MODIS does not carry a MODIS channel at $\lambda=1.64$ µm. When the satellite-measured reflectance is outside the LUT, it is extrapolated within the extended ranges of $R_{s,1.6}$ and $R_{g,0.9}$, and $C_s$ data outside these extended ranges are masked (see Fig. 5 in Stamnes et al., 2007-this issue).

For the validation of these snow parameters, we have made 52 in-situ measurements synchronized with MODIS and GLI overpasses. However, we removed the cloudy data and data at times that deviated more than 1 h from satellite overpasses. As a result, 35 data points were selected (Table 1). Among these data, those employed for the validation were still selected for each snow parameter. For example, since the value of $T_s$ is measured with the infrared channels which see only the uppermost surface layer, ground data for any snow depth more than several centimeters can be used to compare with $T_s$. We used all of the 35 data points to validate $T_s$. For $R_{s,1.6}$, the ground data for any snow depth as well as $T_s$ can be used because the penetration depth at $\lambda=1.64$ µm is less than several millimeters (Li et al., 2001). However, a majority of the detectors in channel 6 ($\lambda=1.64$ µm) of Aqua/MODIS are non-functional (http://www.

Table 1
Summary of field campaign sites, geometric conditions, and snow conditions

<table>
<thead>
<tr>
<th>Site</th>
<th>Longitude</th>
<th>Latitude</th>
<th>Month</th>
<th>Ground condition</th>
<th>$\theta_s$ (deg)</th>
<th>$\theta_v$ (deg)</th>
<th>Year</th>
<th>Snow depth</th>
<th>Snow type</th>
<th>No. data</th>
</tr>
</thead>
<tbody>
<tr>
<td>Saroma</td>
<td>144° 07' 09'' N&lt;sup&gt;c&lt;/sup&gt;</td>
<td>44° 07'</td>
<td>February</td>
<td>Lagoon ice</td>
<td>56.1–61.7</td>
<td>1.5–22.5</td>
<td>2001</td>
<td>8.5 cm</td>
<td>Dry</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td>143° 55' 46'' E&lt;sup&gt;c&lt;/sup&gt;</td>
<td>2001, 2002</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>2002</td>
<td>7–31.5 cm</td>
<td>Dry</td>
<td>7</td>
</tr>
<tr>
<td>Barrow</td>
<td>156° 37' 40'' W&lt;sup&gt;c&lt;/sup&gt;</td>
<td>43° 58'</td>
<td>April</td>
<td>Tundra</td>
<td>57.9–62.2</td>
<td>14.9–46.8</td>
<td>2003</td>
<td>21–33 cm</td>
<td>Dry</td>
<td>16</td>
</tr>
<tr>
<td>Abashiri</td>
<td>144° 11' 37'' E&lt;sup&gt;c&lt;/sup&gt;</td>
<td>43° 58'</td>
<td>March</td>
<td>Lake ice</td>
<td>47.1–47.6</td>
<td>12.9–20.8</td>
<td>2004</td>
<td>17 cm</td>
<td>Wet</td>
<td>2</td>
</tr>
<tr>
<td>Nakashibetsu</td>
<td>144° 42' 50'' E&lt;sup&gt;c&lt;/sup&gt;</td>
<td>43° 58'</td>
<td>March</td>
<td>Farmland</td>
<td>43.1–49.4</td>
<td>0.1–54.9</td>
<td>2004</td>
<td>92–106.5 cm</td>
<td>Wet&lt;sup&gt;f&lt;/sup&gt;</td>
<td>7</td>
</tr>
</tbody>
</table>

<sup>a</sup> Solar zenith angle.<br><sup>b</sup> Satellite zenith angle. Relative azimuth angles ($\Delta\phi$) between the sun and the sensor ranged in 42.3–117.9° or 211.5–307.1° among all sites.<br><sup>c</sup> The following “snow depth,” “snow type” and “No. data” are indicated for this year.<br><sup>d</sup> This column indicates the number of in-situ measurements at each site (total 35).<br><sup>e</sup> The indicated position is at the center point of a five cross-shaped grid points (south to north, east to west, and center) with a 1 km grid interval at which the field measurements were performed.<br><sup>f</sup> Sun crust was observed for all measurements at this site.

Table 2
GLI and MODIS channel specifications used in the present study

<table>
<thead>
<tr>
<th>GLI channel</th>
<th>Center wavelength (µm)</th>
<th>Band width (µm)</th>
<th>Spatial resolution (KM)</th>
<th>SNR&lt;sup&gt;b&lt;/sup&gt; or NEΔT(K)&lt;sup&gt;c&lt;/sup&gt;</th>
<th>MODIS channel</th>
<th>Center wavelength (µm)</th>
<th>Band width (µm)</th>
<th>Spatial resolution (km)</th>
<th>SNR&lt;sup&gt;b&lt;/sup&gt; or NEΔT(K)&lt;sup&gt;c&lt;/sup&gt;</th>
<th>Target</th>
</tr>
</thead>
<tbody>
<tr>
<td>5</td>
<td>0.460</td>
<td>0.01</td>
<td>1.0</td>
<td>880</td>
<td>3</td>
<td>0.469</td>
<td>0.02</td>
<td>0.5</td>
<td>243</td>
<td>$C_s$ and $R_{g,0.9}$</td>
</tr>
<tr>
<td>8</td>
<td>0.545</td>
<td>0.01</td>
<td>1.0</td>
<td>611</td>
<td>4</td>
<td>0.555</td>
<td>0.02</td>
<td>0.5</td>
<td>228</td>
<td>SC&lt;sup&gt;d&lt;/sup&gt;</td>
</tr>
<tr>
<td>13</td>
<td>0.678</td>
<td>0.01</td>
<td>1.0</td>
<td>235</td>
<td>1</td>
<td>0.645</td>
<td>0.05</td>
<td>0.25</td>
<td>128</td>
<td>SC</td>
</tr>
<tr>
<td>19</td>
<td>0.865</td>
<td>0.01</td>
<td>1.0</td>
<td>386</td>
<td>2</td>
<td>0.859</td>
<td>0.035</td>
<td>0.25</td>
<td>201</td>
<td>$C_s$, $R_{g,0.9}$, and SC</td>
</tr>
<tr>
<td>24</td>
<td>1.05</td>
<td>0.02</td>
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<td>0.03 at 285 K</td>
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</tbody>
</table>

<sup>a</sup> Values at nadir.<br><sup>b</sup> Signal-to-noise ratio for channels of $\lambda<3.0$ µm.<br><sup>c</sup> Noise-equivalent temperature difference at 300 K for channels of $\lambda>3.0$ µm.<br><sup>d</sup> Surface classification.<br><sup>e</sup> Cloud mask.
mcst.ssa.jil/macstweb/performance/aqua/aqua-nonfunct-dets.html). For the GLI channel 28 (l = 1.64 μm) whose spatial resolution is 250 m, the full resolution data could not be used operationally because the data were partially sampled only 1/64 of a 2 km by 2 km square due to a limitation in the data transmission. Since these satellite data (Aqua/MODIS channel 6 and GLI channel 28) could not be used for the reasons mentioned above, only the Terra/MODIS data for channel 6 were used for the retrieval of Rs,0.9. The parameters Rs,0.9 and Cs were estimated from the channels at l = 0.460 and 0.865 μm. Since the light absorption by ice is weak at these wavelengths, the light reflected from the snow cover is expected to contain information about snow parameters in the deep layers. According to Wiscombe and Warren (1980), the snow depth for which the snowpack becomes optically semi-infinite is 20 cm for new snow (effective grain radius r e = 50 μm and snow density ρ = 100 kg/m³), 20 cm for fine-grained old snow (r e = 200 μm and ρ = 400 kg/m³), and 50 cm for old melting snow (r e = 1000 μm and ρ = 400 kg/m³). We thus selected the ground data based on the following criterion: snow depth must be thicker than 25 cm for the snow with r e < 250 μm (dry snow) and thicker than 50 cm for the snow with r e ≥ 250 μm (wet snow). Zhou et al. (2003) presented the spectral variation of the minimum depth for a snow cover to be optically semi-infinite for grain radii larger than 200 μm. Our criterion is consistent with the model results of Zhou et al. (2003).

Fig. 3 shows examples of RGB-composite images (top) and satellite-retrieved snow parameters, T s (second row), Cs (third row), Rs,0.9 (fourth row), and Rs,1.6 (bottom row), in a domain of 100 km by 100 km centered at the Saroma site (left column), the Barrow site (second column), and the Nakashibetsu site (right column). The images around those three sites were made from MODIS, GLI, and MODIS data, respectively. There is no image for Rs,1.6 in Fig. 3b-5 because channel 28 (l = 1.64 μm) of the GLI could not be used as described in the previous paragraph. In Fig. 3a-2 to a-5, b-2 to b-4, and c-2 to c-5, the cloud areas are masked with light gray. Open ocean and non-snow-covered areas are indicated in dark gray. Satellite-derived parameters exceeding the extrapolated maximum boundary of the LUT used in the algorithm are indicated in dark red, and those smaller than the extrapolated minimum boundary are indicated in pink. In Fig. 3a-1, the white regions in the Sea of Okhotsk are covered with sea ice. The western part of Saroma Lagoon did not freeze in February 2003, while the eastern part was stable lagoon ice on which we carried out field measurements. In Fig. 3b-2, a striped structure of T s over the snow surface is seen. This is an artifact created by the GLI sensor (JAXA/EOC, 2003). The error from this striped noise was estimated to be less than 0.5 K. In Fig. 3a-1 and c-1, the inland dark-colored regions are areas with vegetation in the snow-cover, and thus Cs is estimated to be out of the extrapolated maximum boundary of the LUT (dark red-colored areas).

Since the spatial resolutions of the GLI channels we used were 1 km viewed at nadir, those of the snow products retrieved from GLI data were 1 km. However, since the spatial resolutions of the MODIS channels that we used (see Hori et al., 2007-this issue; Stamnes et al., 2007-this issue) except for T s were 0.5 km, those of Cs, Rs,0.9, and Rs,1.6 were also 0.5 km. The in-situ measured snow parameters are compared with the satellite-derived parameters, which were averaged in four different ground areas of circles with a diameter of D = 1, 2, 4, and 8 km. The corresponding satellite pixels are extracted when the center position of the ground instantaneous field of view (GIFOV) of the satellite pixel is located within the circles centered at the location of the snow pit work. In the case of D = 1 km the number of extracted pixels (N) is usually one for the satellite data with spatial resolutions of 1 km, and N = 1 to 4 for those with spatial resolutions of 0.5 km. When N is 3 or 4, that is MODIS data, the standard deviation could be calculated for Cs, Rs,0.9, and Rs,1.6. Comparison results between satellite-derived snow parameters and in-situ measurements are shown in the next section.

4. Results and discussion

4.1. Snow surface temperature

Fig. 4 illustrates the relationship between the satellite-derived snow surface temperatures (T s) for the area with D = 1 km and in-situ measured ones. The values agree well with each other around the melting point, while the values of T s deviate somewhat from the 1:1 line of both parameters at lower surface temperatures. Each value of in-situ measured snow surface temperature plotted in Fig. 4 was obtained from snow pit work performed at one location (corresponding to each dot in the figure), while T s represents the spatial average over the GIFOVs for the area with D = 1 km. When the snow surface temperature was close to the melting point (wet snow), the surface temperatures at any position within a satellite pixel, including ground snow pit point, would be very close to the melting point. This tendency can be seen spatially in Fig. 3c-2 around the Nakashibetsu site, where T s around the melting point is prevalent in the snow-covered areas. However, in the case of lower surface temperature (dry snow), the surface temperatures over the satellite looking GIFOV would have a spatial variation larger than those around the melting point. Then one needs to measure the snow surface temperatures at several points on the snow surface. The positive bias seen in T s for dry snow is attributable to both satellite algorithm and in-situ measurements. The possible reason of the former is the errors of the coefficients in multilinear regression, which are related to atmospheric correction and snow emissivity, used in the GLI snow/ice surface temperature algorithm (Stamnes et al., 2007-this issue). The latter, if any, is that the detector head of the thermistor thermometer, which was shaded to avoid solar heating, might be over-shaded with ambient snow surface. However, the values of Rs and RMSE between T s and in-situ measured snow surface temperature were 0.900 and 1.1 K, respectively. The second column in Table 3 presents the values of Rs and RMSE for T s calculated from different size of sampling areas with D = 1, 2, 4, and 8 km in satellite image. The values of Rs and RMSE slowly degrade with D.

The MODIS land surface temperature (LST) product, which was produced by the generalized “split-window” algorithm

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Fig. 3. RGB-composite image (top) and satellite-retrieved snow parameters, $T_s$ (second), $C_s$ (third), $R_{s,0.9}$ (fourth), and $R_{s,1.6}$ (bottom), in domains of 100 km by 100 km centered at the Saroma site (Fig. 3a-1 to a-5), the Barrow site (Fig. 3b-1 to b-5), and the Nakashibetsu site (Fig. 3c-1 to c-5). The images were made from MODIS data at 1025 LT on 24 February 2002 for Saroma site, GLI data at 1357 LT on 14 April 2003 for the Barrow site, and from MODIS data at 1040 LT on 22 April 2004 for the Nakashibetsu site. In Fig. 3a-2 to a-5, b-2 to b-4, and c-2 to c-5, the light gray-colored areas indicate cloud cover and the dark gray-colored areas indicate open ocean and non-snow-covered areas. Satellite-retrieved parameters exceeding the extrapolated maximum boundary of lookup table used in the algorithm are indicated in dark red, and those smaller than the extrapolated minimum boundary of the lookup table are indicated in pink.
(Wan & Dozier, 1996) similar to our algorithm, were validated with MODIS Airborne Simulator (MAS) data collected in 11 clear-sky cases in several field campaigns over an unfrozen lake, snow-free/snow-covered grassland, and a snow-free rice field (Wan et al., 2002). They reported that the accuracy of the daily MODIS LST product at 1-km resolution is better than 1 K in the range from 263 to 300 K. Our RMSE has an accuracy comparable to that of the MODIS LST product. Wan et al. (2002) also stated that the major sources of uncertainty in the LST validation are the spatial variations in surface temperature and emissivity within a MAS or MODIS pixel. Hori et al. (2007-this issue) compared the GLI-derived $T_s$ with MODIS daily ice surface temperature (IST) (MOD29E1D, L3 Global 4 km EASE-Grid V004, Hall et al., 2003). They reported that the GLI-derived $T_s$ were well correlated ($R_c=0.952$, see Fig 5 in Hori et al., 2007-this issue) with the MODIS IST with slight negative biases. Hori et al. (2006) recently determined the spectral directional emissivity in the 8–14 μm atmospheric window for various types of snow from field measurements, and showed that measured emissivity reductions were significant for large snow grain sizes and large viewing angles. Such information would be useful for the improvement of the surface temperature algorithm.

4.2. Mass fraction of snow impurities

For a ground-based estimate of the mass fraction of snow impurities, we used a sum of mass fractions measured by a two-stage filtering system with pore sizes of 0.2 μm and 5.0 μm of Nuclepore filter as mentioned in Section 2.1. Fig. 5 shows the relationship between the satellite-derived mass fraction of soot ($C_s$) for the area with $D=1$ km and the in-situ measured mass fraction of snow impurities for the shallower snow layers (Fig. 5a: 0 to 2 cm) and the deeper snow layers (Fig. 5b: 0 to 7 cm or 10 cm). Since a significant source of snow impurity is from atmospheric aerosols, the mass fraction of snow impurities at the snow surface is generally (not always) higher than in the lower layers (Aoki et al., 2003, 2006). This tendency is consistently observed for in-situ measured data higher than 10 ppmw, which were all wet snow, in both figures.

In both Fig. 5a and b, the satellite-derived mass fractions of soot are lower than the in-situ measured mass fractions of snow impurities. This is probably because the snow impurity was assumed to be soot in the satellite algorithm, whereas the main composition of the in-situ measured impurities at our sites was mineral dust, as mentioned at the end of Section 2.1. Comparing the results of Fig. 5a and b, we note that the values of $R_c$ and RMSE for the deeper snow layer (Fig. 5b) are better than those for the shallower layer (Fig. 5a). The value of $C_s$ is retrieved from a combination of two channels at $\lambda=0.460$ and 0.865 μm. Although snow impurity information is contained in both channels, it is more pronounced in the former channel, in which the light absorption by ice is weaker than in the other channel. The channel at $\lambda=0.865$ μm was used for soot retrieval because the reduction rate of snow reflectance at $\lambda=0.460$ μm per unit mass fraction of soot depends on snow grain size (Wiscombe & Warren, 1980), and the channel at $\lambda=0.865$ μm is more sensitive to change in snow grain size (see Stamnes et al., 2007-this issue for more details about the algorithm). Thus, the snow grain size can be retrieved together with the snow impurities (soot) from these two channels. Our validation results, showing that the values of $C_s$ agree better with in-situ measurements for the deeper snow layer (Fig. 5a) than for the shallower layer (Fig. 5b), are consistent with the snow reflectance at $\lambda=0.460$ μm being influenced by the deeper snow layers.

The third and fourth columns in Table 3 present the values of $R_c$ and RMSE for $C_s$ calculated from different size of sampling areas in satellite image. RMSE does not depend on $D$ for both the shallow snow case (0–2 cm) and the deep snow case (0–7 cm or 0–10 cm). The values of $R_c$ are rather improved for $D=2$ km and 4 km in both cases and decrease for $D=8$ km. Since $C_s$ has a large negative bias compared with in-situ measurement, it is difficult to find a physical explanation from the dependence of $R_c$ and RMSE on $D$.

Even for the better case of Fig. 5b, the accuracy of $C_s$ indicated by the values of $R_c=0.506$ and RMSE=5.0 ppmw is insufficient. This inaccuracy is due primarily to the difference in composition of the snow impurities assumed in the satellite algorithm (soot) and measured in-situ (mineral dust). Particularly, the retrieved results of $C_s$ in Hokkaido shown in Fig. 3a-3 and c-3 are higher than those around Barrow (Fig. 3b-3). The background mass fractions of snow impurities in Hokkaido, where the main composition is dust, were estimated to be several ppmw by Aoki et al. (2003, 2006). The other causes of the error in $C_s$ are inappropriate bidirectional reflectance distribution function (BRDF) model used in the algorithm and satellite sensor calibration error, which will be discussed in the next section where we present the results of snow grain size retrieval. Since mineral dust could also reduce the visible reflectance of snow, it is necessary to distinguish the effects of
soot and mineral dust in the satellite algorithm. The spectral variations of the two components are different at the visible wavelengths, where the imaginary part of the refractive index of soot is almost constant with wavelength (Hess et al., 1998), while that of mineral dust decreases with wavelength (Aoki et al., 2005; Hess et al., 1998). Thus, at least two visible channels are necessary to discriminate the spectral response of the two snow impurities. However, the optical properties of mineral dust have a large uncertainty and current estimates of the imaginary part of the refractive index differ by one order of magnitude (Aoki et al., 2005). Snow impurity type, shape, and mixing condition with ice particles are also possible sources of error in $C_s$. Hansen and Nazarenko (2004) recently demonstrated that long-range transport of strongly absorptive aerosols such as black carbon could be a cause of snow albedo reduction. Remote sensing of the mass fraction of snow impurities (soot and mineral dust) is a very important yet very challenging task.

### 4.3. Snow grain size

The satellite-derived snow grain sizes are of two types. One is retrieved from a combination of the two channels at $\lambda = 0.460$ and 0.865 $\mu$m ($R_{0.9}$), and the other from the channel at $\lambda = 1.64$ $\mu$m ($R_{1.6}$). Fig. 6 shows comparisons of in-situ measured grain size ($\bar{r}_2$) with that retrieved from the satellite data $R_{0.9}$ (Fig. 6a) and $R_{1.6}$ (Fig. 6b) for the area with $D = 1$ km. The light absorption by ice is weakest at $\lambda = 0.47$ $\mu$m and it generally increases with wavelength with some spectrally local absorption bands. Thus, the $R_{0.9}$ is expected to contain snow grain information from snow layers deeper than $R_{1.6}$. Meanwhile, the reflected light from the snow generally contains more information of grain size in the topmost layer than in the lower layers. This information content (weighting function) contained in the reflected light will attenuate exponentially with depth if the snow is homogeneous. For comparison between the satellite-derived snow grain size and in-situ measured values, the exponentially weighted averaging of $R_2$ is preferable for the in-situ measured values. However, the natural snow cover is vertically inhomogeneous implying that the attenuation rate of the information content of grain size depends on the snow density and the grain size itself. We thus firstly compared $R_{0.9}$ and $R_{1.6}$ with in-situ measured snow grain size ($\bar{r}_2$) simply averaged over the snow layer from the surface to various snow depths with snow thicknesses of 0.5, 1, 2, ..., 10 cm for $R_{1.6}$ and 0.5, 1, 2, ..., 25 cm for $R_{0.9}$. The minimum RMSEs were obtained between $R_{0.9}$ and $\bar{r}_2$ for the snow layer from the surface to 5 cm (Fig. 6a), and between $R_{1.6}$ and $\bar{r}_2$ for the snow layer from the surface to 0.5 cm (Fig. 6b).

The satellite-derived $R_{0.9}$ values correlate comparatively well with in-situ measured $\bar{r}_2$, and the value of $R_2 = 0.840$ is better than that for $C_s$. The value of RMSE=125 $\mu$m is acceptable for granular snow with large grain size, but is insufficient for new snow and fine-grained snow with small grain size. There are both underestimates and overestimates of $R_{0.9}$ around $\bar{r}_2$=100 $\mu$m (Fig. 6a). The values of $R_{1.6}$ are underestimated as a whole (Fig. 6b). Particularly for large snow grains of $\bar{r}_2$=200–300 $\mu$m (wet snow), the underestimates of $R_{1.6}$ are significant, while $R_{0.9}$ agreed well with $\bar{r}_2$ for large snow grains. Similar underestimates of satellite-retrieved grain sizes from the middle infrared channels were reported by Fily et al. (1997) using Landsat TM5 ($\lambda = 1.55$–1.75 $\mu$m) and TM7 ($\lambda = 2.08$–2.35 $\mu$m) and by Kay et al. (2003) using MASTER channels 12 to 25 ($\lambda = 1.59$–2.42 $\mu$m). As a result, our result for $R_{1.6}$ is worse than for $R_{0.9}$.

The simple averaging of snow grain size ($\bar{r}_2$) may be an error source of satellite-derived snow grain radius. Nolin and Dozier (2000) emphasized the optical-equivalent snow grain size should be calculated by taking into account that light penetration has a 1/e relationship with depth. We thus made depth-averaging of measured snow grain size by a 1/e weighting using flux transmittance as shown in Appendix A. We secondly compared this depth-averaged measured snow grain radius ($\bar{r}_{2/e}$) with the satellite-derived grain radii. The comparison result with $R_{0.9}$ and $R_{1.6}$ for the area with $D = 1$ km is shown in Fig. 7. The $R_2$ and RMSE results of both $R_{0.9}$ and $R_{1.6}$ are almost similar to those for $\bar{r}_2$. Since the values of $\bar{r}_{2/e}$ are almost the same as $\bar{r}_2$ in the case for $R_{1.6}$ (Fig. 7b) because $\bar{r}_2$ compared with $R_{1.6}$ is the value at topmost layer (0–0.5 cm), the comparison result does not
change. Although in the case for $R_{0.9}$ the values of $\bar{r}_{1/e}$ in Fig. 7a somewhat change from Fig. 6a, $R_c$ and RMSE are not improved as a result. In the estimation of $\bar{r}_{1/e}$, the following problems are considered: To calculate $\bar{r}_{1/e}$ the flux transmittance at each snow layer is necessary as shown in Appendix A. We used SWE1 (see Appendix A) at $\lambda = 0.865$ and 1.64 $\mu$m, and the measured snow density in calculating flux transmittance for the comparisons with $R_{0.9}$ and $R_{1.6}$, respectively. Since $R_{0.9}$ is retrieved from two channels at $\lambda = 0.460$ and 0.865 $\mu$m, a $1/e$ weighting should also be made from both channels. However, the contribution from the two channels would change depending on snow grain size and snow impurities, although the contribution from the channel at $\lambda = 0.865$ $\mu$m should be essentially larger than that at the channels at $\lambda = 0.460$ $\mu$m. Furthermore, the effect of snow impurities should also be accounted for when calculating the flux transmittance, although it is very difficult because we only know the mass fractions of snow impurities for the limited snow layers of 0–2 cm and 0–7 cm or 0–10 cm. There are further uncertainties in the snow density. The vertical resolution of measured snow density with snow sampler (vertical size is 3 cm) is lower than that for snow grain size (1 cm). Thus, the accurate estimation of $\bar{r}_{1/e}$ is actually quite difficult except for a homogeneous snow layer without impurities.

The four right columns in Table 3 present the values of $R_c$ and RMSE for $R_{0.9}$ and $R_{1.6}$, which are averaged for different size of sampling areas in the satellite image. The values of $R_c$ and RMSE generally degraded with $D$. The peaks of $R_c$ are

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Fig. 5. Scatter plot showing the correlation between the satellite-derived mass fraction of soot ($C_s$) for the area with $D=1$ km and the in-situ measured mass fraction of snow impurities for the snow layers of (a) 0 to 2 cm and (b) 0 to 7 cm or 10 cm. The solid line represents a 1:1 relationship. $R_c$ is the correlation coefficient calculated from the linear regression. The vertical error bar indicates the standard deviation of $C_s$ calculated from the 0.5 km-resolution MODIS data.

Fig. 6. Scatter plot showing the correlation between the satellite-derived snow grain radius ($\bar{r}_2$) for the area with $D=1$ km and the in-situ measured values ($\bar{r}_2$) averaged over the snow layers from the surface to 5 cm (Fig. 6a) and from the surface to 0.5 cm (Fig. 6b). The solid line represents a 1:1 relationship. $R_c$ is the correlation coefficient calculated from the linear regression. The vertical error bar indicates the standard deviation of $\bar{r}_2$ calculated from the 0.5 km-resolution MODIS data and horizontal bars indicate $\bar{r}_{2\text{max}}$ and $\bar{r}_{2\text{min}}$. 

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found at $D=2$ km for $R_{s0.9}$ and $D=1$ km for $R_{s0.9}$. Since the geolocation accuracy is likely on the order of one-half the GIFOV, the peaks of $R_c$ at $D=2$ km for $R_{s0.9}$ might be attributed to the difference between the satellite observed GIFOV and the location of a ground measurement. The values of RMSE quickly degraded with $D$ for $R_{s0.9}$ and do not change with $D$ for $R_{s1.6}$.

### 4.4. Possible error sources of $C_s$, $R_{s0.9}$, and $R_{s1.6}$

To estimate the error of satellite products due to the sensor calibration, we calculated the errors in satellite products caused by a 1% error in the TOA radiance for the GLI channels using our algorithms under different conditions of mass fraction of snow impurities and snow grain radius at $\theta_0=50^\circ$ and nadir viewing direction (Table 4). The estimated errors shown in Table 4 have large variations depending on the snow parameters themselves, because the BRDF changes non-linearly with snow parameters. Only a 1% error in the TOA radiance causes several % to several tens % of errors in the satellite products for $C_s$ and $R_{s0.9}$. Nieke et al. (2004) performed the cross-calibration for satellite sensors including GLI, Terra/MODIS and Aqua/MODIS using the spectral reflectance data in-situ measured by our field campaign in April 2003 at the Barrow site. They estimated that the satellite sensor errors ranged from $-4.5\%$ (GLI) to $+4\%$ (Aqua/MODIS) at $\lambda=0.460\,\mu m$ for GLI ($\lambda=0.469\,\mu m$ for MODIS), and from $+0.1\%$ (Terra/MODIS) to $+3.7\%$ (Aqua/MODIS) at $\lambda=0.869\,\mu m$ from the theoretically calculated TOA radiance. These sensor errors can possibly cause a maximum error of 100–200% for $C_s$ and 100% for $R_{s0.9}$. However, the differences between the satellite-derived parameters and the in-situ measurements for $C_s$ (Fig. 5) and $R_{s0.9}$ (Figs. 6 and 7) were sometimes beyond the estimated satellite sensor errors mentioned here.

The next probable common error source for $C_s$, $R_{s0.9}$, and $R_{s1.6}$ concerns the bidirectional reflectance of the snow surface. Many investigators doing remote sensing and atmospheric radiation have tried to measure the snow HDRF (e.g., Painter & Dozier, 2004a; Warren et al., 1998) and have compared such measurements with theoretically calculated values (e.g., Aoki et al., 2000; Leroux et al., 1998). From those results, it is commonly understood that HDRF (or alternatively BRDF) calculated using Mie theory for spherical ice particles is not the best choice (Kokhanovsky et al., 2005; Leroux et al., 1998), particularly for large solar zenith angles and/or large viewing angles (Painter & Dozier, 2004b). Painter and Dozier (2004a) compared the measured HDRF for different types of snow surfaces with HDRF models calculated with DISORT which used single-scattering parameters for ice spheres with radii that matched the surface-area-to-volume ratio derived from stereo-optical analysis of snow samples. They concluded that all HDRF models underestimated reflectance for $\lambda>1.3\,\mu m$ and had large variations.

![Figure 7](image_url)  
Fig. 7. Same as Fig. 6, but for depth-averaged measured snow grain radius ($\bar{r}_{1/e}$) by a $1/e$ weighting using flux transmittance. The horizontal bars indicate $\bar{r}_{1/e}_{\max}$ and $\bar{r}_{1/e}_{\min}$.

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<tr>
<td>28</td>
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<td>500</td>
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$a$ Center wavelengths of GLI channels 5, 19, and 28 are 0.460 $\mu m$, 0.865 $\mu m$, and 1.64 $\mu m$, respectively.

$b$ Errors of satellite products are calculated at $\theta_0=50^\circ$ and the satellite zenith angle $\theta_v=0^\circ$ (nadir).
absolute errors in the perpendicular plane. In our validation campaigns, the solar zenith angles ($\theta_0$) were 56.1 to 62.2° at the Saroma and Barrow sites, and $\theta_0=43.1$ to 49.4° at the Abashiri and Nakashibetsu sites (Table 1). Some of the satellite zenith angles ($\theta_\nu$) are beyond 45°. For such geometric conditions the accuracy of the BRDF model would be critical. Tanikawa et al. (2006) concluded that snow HDRF calculated using nonspherical ice particles with rough surfaces on snow crystals was more suitable than those using spherical particles, based on comparisons with in-situ measured HDRFs for new snow at $\theta_0=58°$ and granular snow at $\theta_0=46°$. Since the anisotropic reflectance of the snow surface is very strong at $\lambda=1.64 \mu m$ (Aoki et al., 2000), the accuracy of $R_{s1.6}$ is expected to be lower than that of $R_{s0.9}$. In Fig. 3a-5 showing $R_{s1.6}$, there were many pink-colored pixels in snow-covered areas, which indicate that the retrieved $R_{s1.6}$ was less than the extrapolated minimum boundary of the LUT. This discrepancy is attributed to two possible causes: (1) use of an inappropriate BRDF model in the algorithm including atmospheric correction and (2) satellite sensor calibration errors. These are common problems for the retrievals of $C_s$, $R_{s0.9}$, and $R_{s1.6}$, where for $T_s$ the term “BRDF” can be replaced by “directional emissivity” in cause (1) above.

Fig. 8. (a) Photograph of sun crust. The size of blue plate is 9.5 cm by 6.5 cm. (b) Sun glint (glinting region in the top two-thirds of snow surface in the picture) from sun crust in the solar direction ($\theta_0=51°$). In the bottom one-third of snow surface in the picture, the sun crust was removed and the second layer can be seen. These pictures were taken at 1336 LT on March 23, 2006 at Nakashibetsu site.
Furthermore, for all of our satellite products, improvements are also needed for in-situ measurements as: (1) the representativeness of the measured values and (2) the measuring methods. Horizontal heterogeneities of snow grain size can be estimated from the results for $\lambda^N_{2}$ measured at five cross-shaped grid points with a 1 km grid interval at the Barrow site. The ratio of each $\lambda^N_{2}$ to the averaged value for five $\lambda^N_{2}$ ranged from 47 to 170% for $\lambda=0.865 \mu m$ and from 41 to 265% for $\lambda=1.64 \mu m$ on April 14, and from 81 to 114% for $\lambda=0.865 \mu m$ and from 83 to 111% for $\lambda=1.64 \mu m$ on April 26. These variations in $\lambda^N_{2}$ are due to the horizontal heterogeneity. Such range of error could be contained in in-situ measurements of snow grain size, which is comparable to the errors due to the sensor calibration mentioned in the first part of this section. The representativeness is an issue such as the difference in area between in-situ measurement and the corresponding satellite pixel as mentioned for $T_e$ in Section 4.1.

We found another possible reason for the underestimate of $R_{\lambda,6}$ under wet snow conditions (and thus large grain size) based on the field measurements. It is “sun crust” created on the snow surface (Fig. 8). Sun crust is a thin, glittering ice layer, which sometimes forms on the snow surface on sunny days (Ozeki & Akitaya, 1996). The depth of the sun crust shown in Fig. 8a was several millimeters. However, the depth has a large variation, because the sun crust consisted of snow grains and ice-bridge bonding each grain to the others. The upper surface was flat compared with the bottom. Sun crust could create a sun glint (specular reflection of solar illumination) and spectral reflectance (thus albedo) is increased in the glint region as we can see in Fig. 8b. We frequently observed sun crust for conditions of wet granular snow under clear sky in spring in Japan. Fig. 9 plots the measured spectral albedo of a granular snow surface with sun crust on March 22, 2004 (synchronized with Fig. 3c) along with the theoretically predicted albedo for homogeneous snow without sun crust.

Although the photographs shown in Fig. 8 were taken on the next day of the measurement in Fig. 7, both sun crusts were very similar to each other. The albedo observing system and the spectrometer were the same as those used by Aoki et al. (2000). Theoretical spectral albedos were calculated for three kinds of layers below sun crust. In Fig. 8, the measured albedo agrees with the calculated albedo for $r_e=300$ to $500 \mu m$, which are values close to the measured snow grain radius (500 $\mu m$). On the other hand, at $\lambda>1.5 \mu m$ the light absorption by ice is strong and the penetration depth is very small (~less than several millimeter for ice plate). The albedos at these wavelengths would be determined primarily by reflection from sun crust. The measured albedo agrees with the calculated values for $r_e=100 \mu m$ or it is still higher than the calculated values in strong absorption bands by ice around $\lambda=1.5$ and 2.0 $\mu m$. This result suggests that the sun crust has a fine structure equivalent to spherical particles of $r_e=100 \mu m$, although the in-situ measured $r_e$ of sun crust was 250 $\mu m$. We do not know the cause of this discrepancy between the model-derived $r_e(100 \mu m)$ and in-situ measured $r_e(250 \mu m)$ at present. The positive bias of albedo at $\lambda=1.5$ and 2.0 $\mu m$ could be a contribution from sun glint reflection off the sun crust. This albedo bias was more pronounced with the growth of sun crust in the afternoon. Tanikawa et al. (2006) confirmed the very high reflectance in the forward scattering direction of sun crust from the HDRF measurements at the same site. This increased reflectance due to sun crust would cause an underestimate of the retrieved snow grain size. On this day, the MODIS-derived snow grain sizes $R_{\lambda,6}$ and $R_{\lambda,1.6}$ were 691 $\mu m$ and 63.1 $\mu m$, respectively. In Fig. 3c-4 and c-5, similar results showing that $R_{\lambda,0.9}>R_{\lambda,1.6}$, were seen in wide areas around the Nakashibetsu site. One should note that satellite-derived parameters of snow near the melting point using channels at $\lambda>1.5 \mu m$ could be affected by sun crust.

5. Conclusions

We conducted several field campaigns for various types of snow conditions at four sites in eastern Hokkaido, Japan and Barrow, Alaska from 2001 to 2005 to validate ADEOS-II/GLI snow/ice products. These products are snow surface temperature
(T), mass fraction of soot contained in the snow (C), and two types of snow grain sizes retrieved from different channels (R_{0.9} and R_{1.6}). These satellite products were compared with in-situ measured snow surface temperatures and vertical profiles of optically equivalent snow grain size based on snow pit work, and the mass fraction of snow impurities measured from snow samples. The satellite-retrieved values of T agree well with in-situ measured values for temperatures close to the melting point, while they deviated somewhat from in-situ measured values at lower surface temperatures. The resultant values of correlation coefficient (R) and root-mean-square error (RMSE) were 0.900 and 1.1 K, respectively. This was due primarily to the difference in composition of snow impurities assumed in the satellite algorithm (soot) and measured in-situ (mineral dust). Snow grain sizes retrieved from satellite channels at  = 0.460 and 0.865 μm (R_{0.9}) had a better accuracy (R_c = 0.840 and RMSE = 125 μm) than those from a satellite channel at  = 1.64 μm (R_{1.6}) (R_c = 0.524 and RMSE = 123 μm) based on a comparison with simply depth-averaged snow grain size. When similar comparisons are made with the depth-averaged measured grain size obtained by a 1/e weighting using flux transmittance, the results for R_c and RMSE are not improved due to some difficulties in calculating the depth-averaging by a 1/e weighting, the results for R_{0.9} and R_{1.6} were underestimated, especially for wet and large snow grains.

For all of our satellite products, the possible causes of errors are (1) the satellite sensor calibration and (2) the bidirectional reflectance model (directional emissivity model for T) used in the algorithm together with the atmospheric correction. The issues addressed with respect to the in-situ measurements are (1) the representativeness of the measured values and (2) the measuring methods. In addition, field measurements indicate that the increased reflectance due to “sun crust”, observed at wet snow surfaces under clear sky, could cause the underestimate of snow grain size. The impact of sun crust on the underestimate of satellite-derived snow grain size should be more enhanced for R_{1.6} than for R_{0.9}.

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Appendix A. Depth averaging of measured snow grain size by a 1/e weighting using flux transmittance

Depth-averaged snow grain radius $r^{1/e}$ at the wavelength $\lambda$ by a 1/e weighting using flux transmittance is calculated from the measured snow grain radius $r(l)$ at a distance from snow surface $l$, and flux transmittance $T(\lambda, r, c, l)$ where $c$ is mass fraction of snow impurities, as given by

$$r^{1/e}(\lambda) = \frac{\int_0^l r(l) T(\lambda, r, c, l) dl}{\int_0^l T(\lambda, r, c, l) dl}.$$  (1)

When the snow layer is homogeneous, $T(\lambda, r, c, l)$ decreases exponentially with $l$ except for close to the upper boundary (Bohren & Barkstrom, 1974). The value of snow water equivalent $T(\lambda, r, c, l)$ was calculated from SWE_{1/e}(\lambda, r, c) which depends on $\lambda$, $r$, and $c$. Fig. A1 presents SWE_{1/e}(\lambda, r, c) at $\lambda = 0.865$ μm and 1.64 μm as a function of $r$, calculated with a radiative transfer model (Aoki et al., 1999, 2000) assuming spherical snow particles with log-normal size distribution of standard deviation 1.6 (Grenfell & Warren, 1999). Flux transmittance $T(\lambda, r, c, l)$ is calculated from SWE_{1/e}(\lambda, r, c) and snow density $\rho_s(x)$ at a depth $x$:

$$T(\lambda, r, c, l) = \exp \left(- \int_0^l \frac{\rho_s(x)}{\text{SWE}_{1/e}(\lambda, r, c)} dx \right).$$  (2)

When the snowpack is inhomogeneous, the integral in Eq. (2) can be calculated from SWE_{1/e}(\lambda, r, c) and snow density in each homogeneous snow layer. Since we measured the values of $c$ for only snow layers of 0–2 cm and 0–7 cm or 0–10 cm from snow pit work in this study, no snow impurities ($c = 0$) are assumed in calculating SWE_{1/e}(\lambda, r, c). We finally estimated $T(\lambda, r, c)$ and $r^{1/e}(\lambda)$ using in-situ measured snow density for $\rho_s(x)$, in-situ measured snow measured $r_2$ for $r(l)$, and SWE_{1/e}(\lambda, r,

![Fig. A1. Snow water equivalent SWE_{1/e} for flux transmittance of 1/e at $\lambda = 0.865$ μm and 1.64 μm as a function of snow grain radius, which is calculated with radiative transfer model assuming spherical snow particles with log-normal size distribution of standard deviation 1.6. The values of SWE_{1/e}(\lambda, r, c = 0) at any value of $r$ are actually determined from regression line of power shown in the figure.](image-url)
$c=0$ at $\lambda = 0.865 \text{ mm}$ and $1.64 \text{ mm}$, where $SWE_{c}(\lambda, r, c=0)$ at any value of $r$ are actually determined from regression curves of power shown in Fig. A1.

References


