How Sensitive is the top of the Atmosphere Albedo to the Presence of Ice or Snow on the Surface?

Irina Gorodetskaya\textsuperscript{1,2}, Mark A. Cane\textsuperscript{1,2}, L.-Bruno Tremblay\textsuperscript{1,2}, and Alexey Kaplan\textsuperscript{1}

Abstract. Despite the wide attribution of the amplified surface warming to the ice/snow-albedo feedback, it is still not well known to what extent sea ice and land snow cover determine the top of the atmosphere (TOA) albedo. The radiative effectiveness (RE) of the ice/snow cover, i.e. changes in the TOA albedo corresponding to the ice/snow cover changes from 0\% to 100\%, is summarized using currently available satellite observations. The REs of the Northern Hemisphere sea ice, land snow, and Southern Hemisphere sea ice are 0.22, 0.23, and 0.16 respectively. The REs of ice and snow have similar magnitudes for seasonal means and are considerably lower than the observed changes in the surface albedo caused by the ice/snow cover to a large extent due to the reflecting effect of clouds in the absence of ice or snow.

1. Introduction

One of key questions in understanding long-term climate variability is the contribution of various feedbacks, which may amplify or dampen the climate system response to changes in radiative forcing [IPCC 2001]. The ice/snow-albedo feedback is believed to contribute significantly to the amplification of projected and observed warming at high latitudes [Curry et al. 1995; Groisman et al. 1994; Holland and Bitz 2003]. It is also one of the principal feedbacks in paleoclimate models, enhancing the creation of sea ice during glaciation [Budyko 1969; Gildor and Tziperman 2000].

A conventional view of the ice/snow-albedo feedback is that a reduction of the ice/snow extent due to warmer surface temperatures should decrease the surface albedo and thus increase the amount of the solar radiation absorbed at the surface. Conversely, a decrease in the surface temperatures would increase the surface albedo and thus reinforce the cooling. However, the strength of the ice/snow-albedo feedback is controlled not by surface albedo changes but by the changes in the albedo at the top of the atmosphere (TOA) [Ingram et al. 1989]. Changes in the cloud cover substantially modify the relationship between the ice/snow on the surface and the TOA albedo and lead to a large discrepancy in the magnitude of the ice/snow-albedo feedback among different models [e.g., Cess et al. 1994; Ingram et al. 1989; IPCC 2001]. Thus, despite the wide attribution of the amplified surface warming to the ice-albedo feedback, it is not well known to what extent changes in the sea ice and land snow cover affect the TOA albedo.

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other satellite and in situ sea ice concentration and extent data including corrections for the effects of surface melt water and wet snow on the passive microwave sensor signals. The biases in SICs derived from microwave radiiances using the Team algorithm [7-11%, according to Gloersen et al. 1992] are greatly reduced in the homogenization process used in the production of the 1° monthly HadISST1 data set [Rayner et al. 2003].

The NH snow cover data were obtained from the National Snow and Ice Data Center Weekly Snow Cover and Sea Ice Extent product (1971-1995). The original data set contained information about the presence (1) or absence (0) of snow in a given week for each grid cell of the 25 km equal-area grid based on the weekly snow charts revised by the Robinson et al. [1993] algorithm.

3. Results

Figure 1 shows the monthly TOA albedo plotted against sea ice and land snow cover percentages for every grid box.

Table 1. Area-weighted means of the TOA albedo and their standard deviations corresponding to 0% and 100% of ice/snow cover. Radiative effectiveness of ice/snow was computed both as the difference between the mean TOA albedo over 100% and 0% of ice/snow cover and using a linear regression. Correlation coefficients between ice/snow percentages and TOA albedo for all monthly data are in the last column.

<table>
<thead>
<tr>
<th></th>
<th>TOA albedo for ice/snow cover</th>
<th>Radiative effectiveness</th>
<th>Corr coeff</th>
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<tbody>
<tr>
<td></td>
<td>mean std mean std</td>
<td>mean std</td>
<td>Lin</td>
</tr>
<tr>
<td>NH sea ice</td>
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<td>0.21 0.21</td>
<td>0.81</td>
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<tr>
<td>SH sea ice</td>
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<td>0.16 0.15</td>
<td>0.82</td>
</tr>
<tr>
<td>NH land snow</td>
<td>0.34 0.08 0.57 0.09</td>
<td>0.23 0.22</td>
<td>0.77</td>
</tr>
</tbody>
</table>

where ice/snow was present at any month during the ERBE time period. The area-weighted mean TOA albedo corresponding to values of 0%, 100%, and all 10% bins of ice/snow cover are shown (solid black lines) along with standard deviations of these subsamples (dashes). Note the spikes in the spread of outliers at 0% of ice, as well as 0% and 100% of snow cover. However, these do not increase appreciably the values of standard deviations.

Two methods were used to estimate radiative effectiveness (RE) of the ice/snow cover with regards to the TOA albedo. The first approach computes \( \text{RE} = \text{albedo(100%)} - \text{albedo(0%)} \), using the averages of TOA albedo observations corresponding to 0% and 100% of ice/snow cover (shown in Fig. 1 and Table 1). The second approach follows YC97 and Groisman et al [1994] and is based on the linear regression albedo(100%) = A * P + B, where P is the percentage of ice/snow cover. After A is computed, we estimate \( \text{RE} = A * (100\% - 0\%) = 100 * A \). Both methods give very similar estimates of the RE for the hemispheric sets of ice and snow cover observations (Table 1). Because of the high number of data points and relatively high correlation coefficients, the errors in estimates of mean albedo and RE in Table 1 can only affect their respective third decimal places and thus are not shown.

Calculations of the RE estimates show a reasonable degree of robustness. Our estimates of the SH sea ice RE for October, January, and April obtained by linear regression for all years in the ERBE period, 0.18, 0.13, and 0.14 respectively, are close to the YC97 estimates for October 1987, January 1988 and April 1988, 0.19, 0.13, and 0.12 respectively. The latter estimates were calculated from Table 1 of YC97 as area weighted averages of zonal values excluding the lowest latitude values for January and October, which YC97 considered an artifact.

Our RE estimates of 0.22-0.23 for the NH land snow based on the whole time-space domain (Table 1) are in agreement with the RE estimates obtained by Groisman et al. [1997, Figure 9] for the coldest temperatures they considered (-8 to -14°C). Their RE estimates for warmer temperatures are smaller (decreasing to 0.05 as surface air temperature increases to 0°C).
The correlation coefficients between ice/snow percentages and TOA albedo, as shown in Figure 1, are quite high, about 0.8 (Table 1). The spatial patterns of corresponding correlations (not shown) have high values in most places where large seasonal changes in ice/snow cover are present (e.g., Nordic seas, Arctic peripheral seas, most of Southern Ocean and of the NH land). Low correlations were found in lower latitudes where the sea ice or land snow is infrequent and in areas with perennial sea ice (Central Arctic and Weddell Sea).

The seasonal change in the mean TOA albedo is almost constant over all ice and snow cover bins, resulting in small seasonal variations of the RE (Figure 1). (Relatively large month-to-month changes are reduced significantly by seasonal averaging). Winter and fall mean TOA albedo values generally exceed summer and spring TOA albedo values over all SIC bins in both hemispheres. Highest TOA albedo values over the ocean are in winter, and the lowest are in summer (Figures 1a,b, blue and red lines). Over the NH land, the largest (smallest) TOA values are found during fall (spring) (Figure 1c, yellow and green lines).

By subtracting five year climatological means (1985 - 1989) from each data set and computing correlation coefficients of resulting anomalies, we examined correlations between interannual variability in the ice/snow cover and TOA albedo (Figure 2). The correlation values higher than 0.45 are significant at the 99% level. High correlations are found in the Arctic peripheral seas and most of the Southern Ocean (Figures 2a,b). In the Central Arctic, interannual variations in SIC are small, thus the correlations are insignificant (Figure 2a). High correlations between snow concentration and TOA albedo anomalies (Figure 2c) are found in the 40-50°N latitude band and in the most northern areas of Eurasia and North America. The correlation coefficients are insignificant near the southern edge of the analysis domain where snow is infrequent.

4. Discussion

The scatter in the TOA albedo values over each SIC bin is due to the variety of factors: effects of sea ice age and properties, cloud cover, solar zenith angle, and errors in satellite data. Comparison of TOA albedo values with available surface albedo observations helps to identify the role of clouds in the RE of ice/snow.

Surface observations for the Arctic Ocean show that its surface albedo over ice can be as low as 0.38 during melt season and as high as 0.84 for the ice covered by dry snow (observations are only available from spring to fall) [Curry et al. 2001]. The surface albedo of the Southern Ocean sea ice varies from about 0.2 to 0.84 depending on the ice type and snow cover [Allison et al. 1993]. The open ocean surface albedo can increase from 0.03 for low solar zenith angles up to 0.27 for high solar zenith angles [Curry and Webster 1999]. These surface albedo values define an envelope of the surface albedo ranges shown by magenta lines in Figures 1a,b.

Over land, vegetation has a substantial effect on surface albedo, reducing the impact of snow cover on the TOA albedo too. Note that no interannual correlations of the TOA albedo with the snow cover were found in the latitude band from about 55° to 70°N and in the north-west of North America (Figure 2c). In this band forests decrease the TOA albedo over the snow-covered areas in winter (while frequent frontal clouds increase the TOA albedo in summer, cf. Serreze et al. [2001]). Liou [1992, Table 6.4] finds surface albedo values of ~0.15 for summer tundra and ~0.1 for boreal forest. In winter surface albedo can range between 0.8 over the snow-covered polar tundra and 0.36 over the boreal forest [Robinson and Kukla 1985]. In Figure 1c the magenta lines connect the summer (snow free) and winter (snow covered) values of surface albedo over tundra (upper line) and over boreal forest (lower line).

In all three panels of Figure 1 the observed values of TOA albedo are outside of the range of surface albedo at low percentages of ice/snow cover and within the surface albedo range at high percentages of the ice/snow cover. This implies the important role of the non-surface effects, i.e. the atmosphere and clouds, in reducing the effects of the ice/snow cover on the TOA albedo values.

In an attempt to disentangle the effects of surface conditions and clouds on TOA albedo we tested the clear-sky TOA albedo product from the ERBE as well. Against our expectations, the correlation coefficients between the clear-sky TOA albedo and the ice/snow cover percentages were no higher than those for the all-sky TOA product. Apparently the high cloudiness in polar latitudes and errors in the ERBE's clear sky scene identification over ice or snow covered surfaces result in large undersampling (and consequently slow error reduction in averaging) of clear-sky values [Li and Leighton 1991].
5. Conclusions

Using more than 5 years of ERBE all-sky TOA albedo data from the 2.5°x2.5° grid boxes over polar oceans and the NH land, we found that when sea ice or land snow cover percentage varies from 0% to 100%, the TOA albedo changes by about 0.2. This value was shown to be robust and widely applicable, in generalization of the earlier studies, focused on either ice or snow, smaller regions or shorter periods. We identified the reflecting effect of clouds at the low percentage of ice/snow cover as the main culprit for the reduced RE of ice/snow with regards to the TOA albedo (in comparison to the ice/snow effect on surface albedo). Despite this reduction in the RE, the TOA albedo depends significantly on the ice/snow presence, as demonstrated by high seasonal and interannual correlations in all areas with large seasonal or interannual variability of the ice/snow cover and without tall vegetation.

The presence of a high amount of clouds during the whole year in the polar regions shields the effects of the surface on the reflected short wave (SW) radiative fluxes at the TOA [cf. Ingram et al. 1989; YC97]. If the sea ice retreat in response to a surface warming trend is associated with increased cloudiness, the latter will reduce the magnitude of the ice-albedo feedback and its ability to amplify the surface warming. However, during seasons when the amount of incoming solar radiation is high, the estimated 0.2 radiative effectiveness of ice and snow is enough to have a prominent influence on the reflected SW radiation. In summer, there is about 400 W/m² of SW radiation incoming to the polar latitudes. Local reduction of SICs from 100% to 0% would then result in an 80 W/m² decrease in reflected SW radiation. Since the observed negative trends in the Arctic sea ice extent during 1978-1996 [Parkinson et al. 1999; Serreze et al. 2000] have been largest in summer and early autumn in many areas where the correlations between the SIC and TOA albedo anomalies are high (e.g., Arctic peripheral seas, Fig. 2a), this decrease in reflected radiation can have a substantial impact on the local SW radiation balance.

These estimates of the TOA albedo response to sea ice and land snow cover percentage changes obtained for the present climate provide a useful constraint to test current climate models. Simplified models should calibrate the ice/snow-albedo feedback, using the ~0.2 value of TOA REF of sea ice and land snow cover rather than larger values based on surface albedo changes. These estimates can be also useful for validations of general circulation models used for long-term climate change studies. The ability of clouds to modify the ice/snow-albedo feedback should not be underestimated.

Acknowledgments. The authors are grateful to Hezi Gildor, Beate Liepert, Cecilia Bitz, Bill Rossow, Pavel Groisman and Judah Cohen for valuable comments and discussions; Seiji Kato, Norman Loeb, Bruce Wielicki, and Felix Tubiana for information about ERBE data; Richard Ianuzzi, Gerd Krahmann, and Felix Tubiana for providing satellite data; Nick Rayner for advice on sea ice data; Richard Ianuzzi, Gerd Krahmann, and Felix Tubiana for programming assistance. We thank NASA’s ERBE Data Management Team, National Snow and Ice Data Center, and Hadley Center for Climate Prediction and Research for providing satellite data. IG was supported by the NOAA grant NA06GP0567, NSF grant ATM0896515, and NASA Fellowship ESSF0400000163. MAC and AK were supported by NOAA grants NA06GP0567 and NA08GP0437. LBT was supported by the NSF grant OPP-0230264. This is LDEO contribution xxxx.

References


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