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

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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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Yamanouchi and Charlock [1997] (hereafter YC97) calculated the radiative effectiveness (RE) of sea ice with respect to the TOA albedo as the difference between albedo values corresponding to 100% and 0% of the sea ice concentration (SIC) using a linear regression of the TOA albedo on the SIC. Their results were based on one year of the Earth Radiation Budget Experiment (ERBE) data (February 1987 to January 1988) for the Southern Ocean. They obtained the lowest RE for January 1988 (10-14%) and the highest RE in October 1987 (up to 22%).

The seasonal and interannual variations in the Northern Hemisphere (NH) land snow cover and their impacts on the TOA radiative balance have been studied in detail by Groisman et al. [1994]. According to their results, the snow-albedo feedback plays an important role in spring, when it contributes significantly to the surface air warming. They used a linear regression as well to estimate RE as a function of air temperature.

This paper quantifies the RE of sea ice and land snow cover with respect to the TOA albedo using the satellite data of the entire ERBE period (November 1984 - February 1990) for both hemispheres. The analysis is based on the whole time-space domain where the sea ice and NH land snow appear, and reveals a remarkable similarity of the ice and snow RE with regards to the TOA albedo, despite the varying nature of the surface cover, different seasons, and locations. Section 2 describes the data. Section 3 presents the results, which are discussed in Section 4. The conclusions are given in Section 5.

2. Data

2.1. Albedo

We are using spectrally integrated (over the 0.2-5.0 μm band) monthly total-sky TOA albedo values from the ERBE data set gridded with a spatial resolution of 2.5°. Because the ERBE product is derived from three satellite missions (ERBS, NOAA-9 and -10), the data are available equatorward of 67.5°N and 67.5°S for the time period from November 1984 to February 1990 (64 months). From February 1985 to May 1989 (up to 52 months, depending on the length of the polar night) the data are also available poleward of those latitudes. In the ERBE data set, measured instantaneous radiances are converted to monthly TOA fluxes by applying scene-dependent angular models and averaging over time and space [Barkstrom 1989].

2.2. Presence of ice and snow

The presence of ice or snow on the surface is characterized by the gridded values representing monthly averages of the snow- or ice-covered area of a grid box in percentage of its entire area. These gridded values are obtained by averaging the grids of the data sets of SIC and snow cover onto the 2.5° monthly grid used by the ERBE product.

SICs are taken from the HadISST1 data set [Rayner et al. 2003], which for the period after 1978 derives them from the Special Sensor Microwave/Imager and the Scanning Multichannel Microwave Radiometer data using the Team algorithm [Gloersen et al. 1992]. These are then corrected using

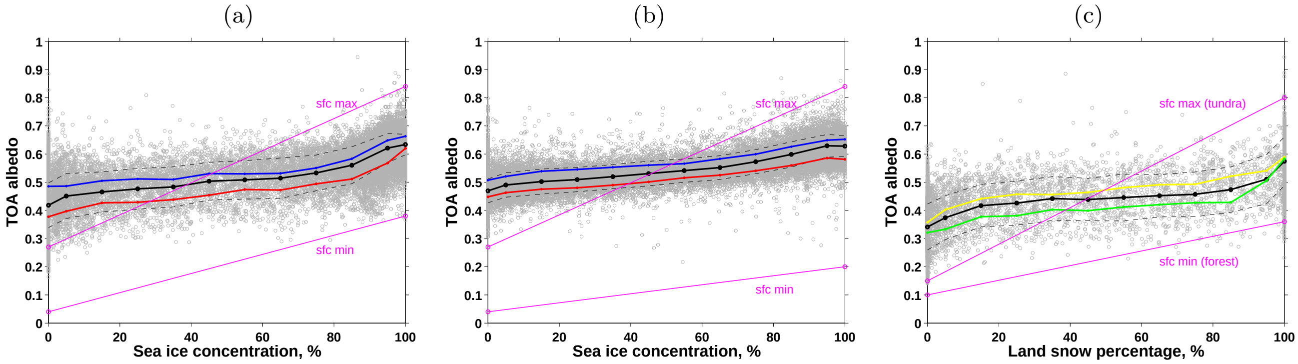

Figure 1. Monthly mean TOA albedo values (gray circles) plotted versus co-located sea ice concentrations in the (a) Northern Hemisphere, (b) Southern Hemisphere, and versus (c) NH land snow cover percentages. Included are data for all grid boxes where ice or snow was ever present during the ERBE time period. Area-weighted TOA albedo averages for 0%, 100%, and each 10% bin of ice/snow cover are shown for all data (solid black dots and line), for winter (blue), for summer (red), for fall (yellow), and for spring (green). The standard deviations for all data are shown by dashed lines. Winter (summer) is defined as December-February (June-August) in the NH and June-August (December-February) in the SH; the NH fall is September-November, and the spring is March-May. Thin magenta lines connect cited in literature maximum and minimum surface albedo values for the open ocean and sea ice, and for tundra and forest over land.

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 radiances 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 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 percentage 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.

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.

	TOA albedo for the ice/snow cover of				Radiative effectiveness		Corr coeff
	0%		100%		100-0%	Lin	
	mean	std	mean	std	diff	regr	
NH sea ice	0.42	0.08	0.63	0.04	0.21	0.21	0.81
SH sea ice	0.47	0.04	0.63	0.04	0.16	0.15	0.82
NH land snow	0.34	0.08	0.57	0.09	0.23	0.22	0.77

Two methods were used to estimate radiative effectiveness (RE) of the ice/snow cover with regards to the TOA albedo. The first approach computes RE = albedo(100%) - 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 $= A * P + B$, where P is the percentage of ice/snow cover. After A is computed, we estimate 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).

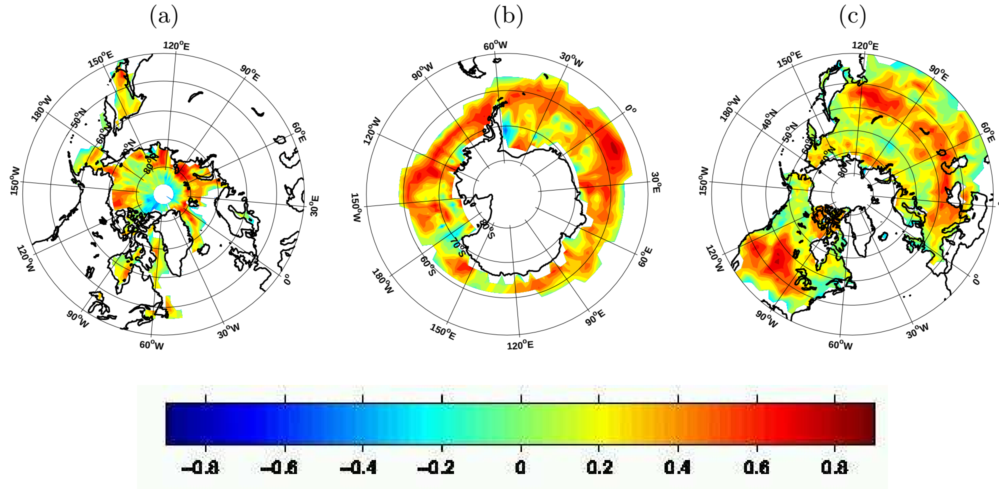

Figure 2. Correlation coefficients between monthly interannual anomalies of TOA albedo and sea ice concentrations in the (a) Northern Hemisphere, (b) Southern Hemisphere, and (c) snow cover percentages over the NH land, for November 1984 - February 1990 equatorward of 67.5^{o}N and 67.5^{o}S, and for February 1985 - May 1989 poleward of those latitudes.

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 are 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/m^2 of SW radiation incoming to the polar latitudes. Local reduction of SICs from 100% to 0% would then result in an 80 W/m^2 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 RE 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.

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