A Intercomparison and Validation of Snow and Sea-Ice Albedo Parameterization Schemes in Climate Models

C. A. Pedersen and J.-G. Winther, poster presentation at American Meteorological Society 85th Annual meeting, San Diego, USA, January 2005.

Motivation

The main motivation for this poster was similar to the snow only counterpart in Paper I, so here we will only review the sea-ice part: first to intercompare different sea-ice albedo parameterization schemes, and second to validate how good they described the sea-ice albedo compared to *in situ* measurements. We also identified the significant parameters for modeling the sea-ice albedo from a purely statistical point of view.

Results

Sea-ice albedo schemes from three models of different complexity is investigated in this study. The first two schemes has a linear dependency on surface temperature, while the last is a one dimensional thermodynamic sea-ice model (Ebert and Curry, 1993). The last scheme could not be used to its full potential here because of limitations in the validation data set. Data from three Russian drifting ice stations (North Pole Stations, covering 3, 3 and 8 years) are used for validation. The ice on the validation sites were relative thick, and snow covered the ice during winter and early spring. The sea-ice albedo is underestimated during the winter (similar as for the snow albedo schemes), which was a direct consequence of the ice being snow covered during winter. All albedo schemes are relative constant and fixed at the upper boundary values during winter and fluctuated between the thresholds in spring, summer and autumn. All schemes are found to have relative similar performance in terms of root mean square errors (RMSE) and correlation coefficients. The temperature dependent schemes has RMSE of 0.09-0.14, and correlation coefficients of 0.22-0.60, while the thermodynamic sea-ice model has RMSE of 0.09-0.12, and correlation coefficients of 0.24-0.46 for the three validation sites. The relative poor results for the thermodynamic sea-ice model are probably due to the validation datasets lacking important components like information about melt ponds and open water. The significant parameters for modeling sea-ice albedo from a statistical regression model was found to be temperature, snow depth, cloud cover and a dummy of snow depth.

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B Sea-Ice Albedo and Fractional Sea-Ice Types from Moderate Resolution Imaging Spectroradiometer (MODIS) Sensor

C. A. Pedersen, D. K. Hall, B. Hamre, E. Vermote, J.-G. Winther and S. Gerland, work in progress

Motivation

Recent research has shown a reduction of sea-ice cover in the Arctic since the satellite record began in 1979 (Comiso, 2006). The low albedo of open water as compared to sea ice, causes much of the incoming solar radiation to be absorbed in the ocean, which leads to further warming (Curry et al., 1995). Even a small change in albedo can have an important impact on the global energy balance, especially if sustained over a period of time. Satellite observations are the most effective means of monitoring the snow and ice, and the Moderate Resolution Imaging Spectroradiometer (MODIS) is particularly well suited as it image mid-to-high latitude snow and sea-ice covered areas with high spatial- (250-500 m) and fine spectral (7 bands in the visible and near-infrared part of the solar spectrum) resolution on a daily basis. We have developed algorithms for extracting daily, global sea-ice albedo and fractional sea-ice types from MODIS remote sensing data.

Methodology

The approach we are taking for sea-ice albedo is similar to the MODIS snow albedo prototype (Klein and Stroeve, 2002), with a few exceptions and supplements. The starting point of the procedure (Figure B.1) is the cloud free pixels from the surface directional reflectance MOD09 (Vermote and Vermeulen, 1999), corrected for atmospheric scattering and absorption. The satellite measures reflectance in the look direction of the sensor, and must be corrected for anisotropic effects of the surface in order to extract the narrowband albedo (reflectance integrated over all angles). Different anisotropic correction functions are calculated separately for snow, bare ice and open water, wavelengths and combinations of sensor and Solar angles, from the discrete radiative transfer model (DISORT; Hamre et al., 2004) by normalizing the bidirectional reflectance distribution function with the spectral albedo. The open water pixels are identified directly from MOD29 (Riggs et al., 2003), while a ratio-test of the normalized difference ice index (NDII; Stamnes, 2005), similar to the well-known normalized difference snow index, NDSI, is applied for separating between snow and bare ice. The narrowband albedos are converted to broadband albedos by weighting the MODIS band according to the

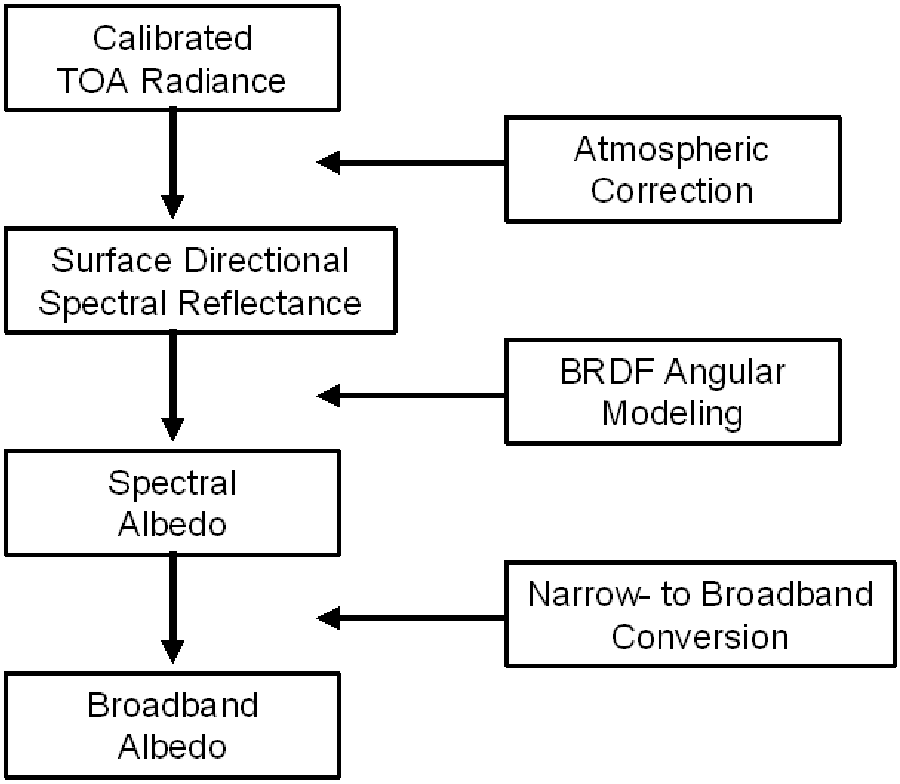

Figure B.1: Flow chart showing the design of the MODIS sea-ice albedo product.

incoming downward flux at the surface, by using simple regression convention formula (Liang, 2001).

The fractional sea-ice types are calculated from spectral unmixing (Vikhamar, 2003), which is an unsupervised classification technique that uses the method of least squares to model pixel spectra as a linear combination of "pure" reference spectra (endmembers). Endmembers for each of the six sea-ice types: dry snow, melting snow, bare sea-ice, thin ice, melt ponds and open water were calculated from previous field data. The spectral unmixing gives the fractional sea-ice type for every pixel in a MODIS scene.

The MODIS sea-ice albedo and fractional sea-ice type algorithms have been implemented and tested on several MODIS scenes with success. However, we have not been able to validate the albedo nor the fractional sea-ice types. As described in the outlook, optical sensors like MODIS require clear sky, while the Arctic is characterized by an extensive cloud cover, particularly during spring when our field campaigns normally take place. Therefore our attempts to collect appropriate validation data has failed. However, we hope to complete this work by validating the MODIS sea-ice albedo products in the near future.

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C Black Carbon in Arctic Snow and Implications for Albedo Changes

C. A. Pedersen, S. Gerland, S. Forsström, T. K. Berntsen, J. Ström, S. Warren, R. Brandt, A. Clarke, presentation at NySMAC meeting, Cambridge, UK, Oct. 2007, and work in progress

Motivation

Snow covered surfaces have a high albedo and most of the incoming solar radiation is reflected. Black carbon (BC) particles emitted by fossil fuel and incomplete biomass combustions, transported to the Arctic and deposited in the snow and ice, reduce the albedo of the surface and thus contribute to a warming of the climate. The climate forcing due to the direct albedo effect is +0.3 W/m^2 in northern hemisphere, if assuming BC concentrations giving an albedo effect of 1.5% in the Arctic and 3% in northern hemisphere land areas (Hansen and Nazarenko, 2003). Also, the efficacy of BC forcing is approximately 3 (Flanner et al., 2007), that is, for a given forcing, BC is three times as effective as CO_2 in changing global surface air temperatures.

The main motivation behind this ongoing work is to measure BC concentrations in Arctic snow and to quantify the corresponding alteration of the reflective properties of snow. So far, numbers for the albedo effect are basically derived from modeling.

Results

Samples of BC in snow together with snow optical properties and physical parameters have been collected at various locations in Svalbard in the years 2005-2007. Average elemental carbon (which is a proxy for BC) concentrations measured at Svalbard in spring 2007 was 12.7 ng of elemental carbon per g of snow (ng/g), ranging from 2-68 ng/g. This excludes one site close to the coal mines of Barentsburg with concentrations in the range 240-520 ng/g, and clearly contaminated by local sources. Previous measurements for Svalbard from 1983 show average values of 31 ng/g (Clarke and Noone, 1985), thus substantial higher concentrations than we measured. However, these concentrations can not be compared directly, as, first of all, strong emission control took place in the 1990ies, thus reducing the BC amounts substantially, and secondly, the concentrations from the 1980ies were another type of proxies. BC measurements were also carried out in the Arctic Ocean during the SHEBA experiment in 1998 (Grenfell et al., 2002), where BC proxies ranged from 1-10 ng/g, with mean values of 4-5 ng/g, well comparable with our values.

The elemental carbon concentrations showe relative large variabilities, both spatially and from the analysis method. Snow samples of the same depth collected at nearby locations at the same time varied by up to a factor of 2-3, while the analysis uncertainty depends on the absolute amount of BC (which at Svalbard is relatively low). With limited sampling and analysis capacity, there is a trade of between collecting much snow for few sites (and reduce the analysis error) or collecting less snow for many sites (thus reducing the spatial variability).

Optical properties of snow depend on several snow physical and textural properties (snow depth, grain size *etc.*) and environmental factors (clouds, sun angle, aerosols/particles). However, aerosols like BC only affect the optical properties in the visible part of the spectrum (below 900 nm), however they may indirectly change the albedo at other wavelengths due to particle-related enhanced snow metamorphosis and melting. Due to the low BC concentrations in the Arctic, it proved to be difficult to separate the BC signal from the dependency of the other snow and environmental parameters. The lower limit for a detectable albedo effect is said to be 1% (Warren and Wiscombe, 1980), but it proved to be very difficult to resolve such small deviations in the field with all the other parameters varying. Model calculations by Warren and Wiscombe (1980) show that 10 ng/g BC in new snow or 4 ng/g BC in coarse grained old snow, will reduce the spectral albedo with 1% at the most sensitive wavelength (470 nm). The effect is less at other wavelengths. To reduce the broadband albedo 1% require a higher BC concentration of 40 ng/g in new snow or 10 ng/g in old snow. These calculations were for an external mixture of snow and BC, however, BC can be up to a factor of 2 more effective if the mixing is internal (coated particles).

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