Antarctic ice sheet and sea ice regional albedo and temperature change, 1981–2000, from AVHRR Polar Pathfinder data

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Received 9 January 2007; received in revised form 1 June 2007; accepted 2 June 2007

Abstract

Spring–summer (November, December, January) ice sheet and sea ice regional surface albedo, surface temperature, sea ice concentration and sea ice extent averages and trends from 1981 to 2000 have been calculated for the Antarctic area. In this research the AVHRR Polar Pathfinder 5-km EASE-Grid Composites and the combined SMMR and SSMI data sets from the National Snow and Ice Data Center (NSIDC), Boulder, Colorado have been employed. A regional analysis has been made for five longitudinal sectors around Antarctica: the Weddell Sea (WS), the Indian Ocean (IO), the Pacific Ocean (PO), the Ross Sea (RS) and the Bellingshausen–Amundsen Sea (BS). The IO and PO sectors show ice sheet albedos of 0.85 and temperatures of −25 °C. The corresponding values in the RS and BS sectors are 0.80 and −16 °C respectively. The sea ice albedo is about 0.60 in the RS, BS and WS sectors and 0.55 in the IO and PO sectors. The average sea ice temperature varies around −12 °C. All the sectors show slight increasing spring–summer albedo trends and decreasing spring–summer temperature trends and similar interannual variability in albedo and surface temperature. The steepest ice sheet albedo trend of 0.0019±0.0009/yr is found in the RS sector. The steepest sea ice albedo trend of 0.0044±0.0017 /yr occurs in the PO sector. The steepest temperature trends for both the ice sheet and sea ice occur in the BS sector, having values of −0.075±0.040 °C/yr and −0.107±0.027 °C/yr respectively. The sea ice concentration shows slight increasing trends, the highest being in the PO sector (0.3±0.12%/yr), whereas the sea ice extent trends are near zero with the exception of the RS sector (14,700±8600 km²/yr) and the BS sector (−13,000±6400 km²/yr).

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Keywords: Antarctica; Ice sheet albedo; Sea ice albedo; Ice sheet temperature; Sea ice temperature

1. Introduction

The albedo of the Arctic and Antarctic ice sheets and their sea ice region is one of the most important parameters affecting the radiation budget of the earth–atmosphere system. Snow and ice have the highest albedo of all terrestrial surface types. However, changes in snow properties and snow cover, as well as synoptic weather events, such as snowfall or snow drifting by the wind can cause large changes in the surface albedo. An albedo increase from 0.54 (bare ice) to 0.89 after the occurrence of snowfall, has been observed by Pirazzini (2004). Also, changes in the sea ice concentration can strongly affect the sea ice regional albedo. Observations and realistic estimates of the variations and trends in the Arctic and Antarctic ice sheet and sea ice albedo are a prerequisite for estimating radiation balance changes in the polar regions and their influence on the global climate system. At the present time the long-term trend in the Antarctic ice sheet and sea ice albedo is insufficiently well known.

Stroeve (2001) studied the variability of the surface albedo over Greenland from 1981 to 1998 using AVHRR Polar Pathfinder data. Surface melting on the ice sheet decreased the surface albedo from 0.8 to values below 0.7 for melting snow. According to Stroeve’s results, anomalously low surface albedos were found in 1995 and 1998 as compared with the other years in the time series. As an explanation for the low albedo of these years, it has been suggested that the ice sheet experienced considerable melting. High albedos were found in 1992 due to lower temperatures and hence less melting following the eruption of Mount Pinatubo. The time series showed slight negative trends for surface albedo during the 18-year period. However, the overall negative trend in albedo was not statistically significant.

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doi:10.1016/j.rse.2007.06.005
Most recently, in three ship-based field experiments in the Antarctic sea ice region, spectral albedos were measured at ultraviolet, visible, near infrared and broadband wavelengths for open water and different ice types, both with and without snow cover (Bradt et al., 2005). The broadband albedos varied from 0.07 for open water to 0.87 for thick snow-covered ice under cloud. In the study of Bradt et al. (2005), the sea ice distribution observations for the years 1980 to 2000 from the Antarctic Sea Ice Process and Climate (ASPeCt) project were also available. Data of the summer surface albedo over four Antarctic sites were compared in order to estimate the spatial and temporal variability of the albedo (Pirazzini, 2004). For solar zenith angles less than 80° the albedo steadily decreased during the day. Snow metamorphism, sublimation during the day, and refreezing and/or crystal formation/precipitation during the night can explain the observed trend particularly in the coastal regions. The albedo variability associated with snow/ice metamorphism was larger than the variability due to changes in sky conditions. The daily mean albedo was found to vary between 0.58 and 0.82.

Grenfell et al. (1994) studied the variation of the spectral snow albedo with wavelength across the solar spectrum from 0.3 μm to 2.5 μm in the near infrared at the Amundsen–Scott South Pole Station during the Antarctic summers of 1985–1986 and 1990–1991. The albedo had a uniformly high value of 0.96–0.98 across the UV and visible spectrum, nearly independent of snow grain size and solar zenith angle. The albedo in the near IR was lower, dropping below 0.15 in the strong absorption bands at 1.5 and 2.0 μm, and it was found to be quite sensitive to grain size and somewhat sensitive to zenith angle. The spectrally-averaged albedos ranged from 0.80 to 0.85 for both overcast and clear skies.

Recently, Laine (2004) estimated the Arctic sea ice albedo variability and trends during 1982–1998. Whole-year and monthly sea ice regional albedo averages for June, July and August from 1982 to 1998 were processed from the AVHRR Polar Pathfinder data set. Time series for albedo, sea ice concentration, sea ice extent and surface air temperature were calculated for the sea ice cover for the Northern Hemisphere as a whole, and for six sub-regions: the Arctic Ocean, the Kara and Barents Seas, the Greenland Sea, the Labrador Sea, Hudson Bay, and the Canadian Archipelago. In general, the sea ice albedo in the central Arctic was between 0.5 and 0.7. The highest albedo values were mainly found in the Arctic Ocean north of Greenland. The lowest albedo (0.2–0.3) occurred in the fringe area of the Arctic Ocean, e.g., on the coasts of Alaska and Siberia. Generally, the sea ice albedo, concentration and extent trends were negative in Arctic regions, whereas the temperature trends there were positive (Laine, 2004).

The objective of this study is to extend investigation of long-term annual ice albedo changes to the Antarctic region. In the present study, spring–summer ice sheet and sea ice regional albedo averages for the months of November, December and January from 1981 to 2000 have been calculated for the entire Antarctic area. In addition, a regional analysis has been made both of the ice sheet and sea ice albedo variations of five longitudinal sectors around Antarctica: the Weddell Sea, the Indian Ocean, the Pacific Ocean, the Ross Sea and the Bellingshausen–Amundsen Sea. The sector definitions are the same as in the study of Bradt et al. (2005). The albedo estimation has been limited to the above-mentioned months of these years to ensure adequate sun elevations. A higher sun elevation allows more accurate satellite-based albedo estimates because of the reduced atmospheric disturbance.

Sea ice is thought to provide a positive ice albedo feedback, especially in the Arctic sea ice region. The melting surface induces sea ice retreat and darkens the bare ice, resulting in a lower surface albedo and thus increased solar radiation absorption at the ice surface, which in turn produces further melting and increases the area of open water in the sea ice area. In the reverse direction, a cold anomaly may produce a drier snow surface or allow a deeper snow cover and more ice to form, increasing the surface albedo. This may in turn cause a further cooling of the surface ice area (Perovich et al., 2003). In the Antarctic the feedback process is more complex. In the Antarctic sea ice region the ice is mainly melting from the bottom up (Bradt et al., 2005). This results in the melting ice retaining its snow cover much longer and thus retaining a high albedo. A warming climate may also generate more snow in both the sea ice and ice sheet regions, thus increasing the albedo. It is therefore necessary to study the potential relationship between the concurrent trends and anomalies of ice sheet and sea ice albedos, surface temperatures, sea ice concentration and sea ice extent in the Antarctic region.

2. Data

The Advanced Very High Resolution (AVHRR) Polar Pathfinder (APP) data provide long-time series (1981–2000) of calibrated surface albedo and surface temperature data for the polar regions. In this research the AVHRR Polar Pathfinder 5-km EASE-Grid Composites from the National Snow and Ice Data Center (NSIDC), Boulder, Colorado were employed. Data products used in this paper include the broadband albedo corrected for bi-directional reflectance and atmospheric effects, skin temperature, surface type mask and cloud mask. The albedo, temperature and cloud masking are all derived from the Cloud and Surface Parameter Retrieval (CASPR) system (Key, 2002). The general methodology for the broadband albedo retrieval can be found in Csiszar and Gutman (1999). The sensor drift for the visible AVHRR channels 1 and 2 (used for the EASE-Grid broadband albedo conversion) has been corrected using a set of time-varying coefficients by Rao and Chen (1994) and Rao and Chen (1999). The thermal AVHRR channels 4 and 5 (used for EASE-Grid skin temperature conversion) have been corrected by the method of Walton et al. (1998). Details of the data description and processing can be found in Fowler et al. (2000). The Polar Pathfinder products on grids of 5 km have been generated from AVHRR global area coverage (GAC) data. The data set consists of twice-daily composites (at approximately 02.00 and 14.00 Local Time (LT) for the Southern Hemisphere), of which only the afternoon data (14.00 LT) have been used here to ensure uniformity of illumination. Currently, the data for November and December in 1994, January in 1995 and January in 2001 (needed for this study) are missing from the archive.
The Scanning Multichannel Microwave Radiometer (SMMR) and Special Sensor Microwave Imagers (SSMI) provide long-time series of brightness temperature data for investigating the surface characteristics of the sea ice cover. For this study the sea ice concentrations have been extracted from the NSIDC data set, i.e. the Bootstrap Sea Ice Concentrations from Nimbus-7 SMMR and DMSP SSM/I. Daily and monthly data, from 1978 to 2003, are available for both the south and north polar regions. A detailed description of the data and its accuracy can be found in Comiso (1999).

The sea ice extents have been calculated from the combined SMMR and SSMI data sets using the NASA Team algorithms. The data set is developed by Cavalieri et al. (1997), Parkinson et al. (1999), Gloersen et al. (1999), Gloersen et al. (1992), Parkinson and Cavalieri (2002), and Zwally et al. (2002). In the data set the SMMR/SSMI-derived sea ice concentrations have been used to determine extents for sea ice having a concentration of at least 15%. The data set can be downloaded from http://polynya.gsfc.nasa.gov/seaice_datasets.html.

2.1. Cloud masking

To obtain the clear-sky surface albedo by the current method, cloud-free pixels need to be separated from cloudy ones. The success in carrying out cloud detection is one of the critical points in obtaining the clear-sky surface albedo and temperature. In polar regions, over snow and ice, cloud screening is especially complicated.

The AVHRR Polar Pathfinder 5-km EASE-Grid Composites provide three different cloud masks (Fowler et al., 2000; Stroeve, 2001). The first cloud mask is based on the CASPR single-day algorithm. The algorithm operates on data from a single day using a combination of spectral tests and thresholding of different AVHRR channels. The second cloud masking method referred to as the temporal thermal filtering method makes use of a long-time series of channel 4 data. Beginning with at least 120 days of channel 4 brightness temperatures, a combination of median and maximum filtering is applied to 20-day increments of the brightness temperatures for each pixel. The filtering selects the most likely clear-sky temperature for each pixel, and interpolates in time to provide an estimated temperature during cloudy periods. This is assumed to be close to a clear-sky time series of channel 4 brightness temperatures at each pixel location. A cloud mask is generated by thresholding the actual channel 4 with the time series. The third CASPR multi-day cloud mask is similar to the first, except that the channel 4 statistical mean is replaced with the channel 4 values from the long-time series of channel 4 data described above.

Stroeve (2001) has made a detailed analysis of the characteristics of the three cloud masks. According to her, the procedure based on the above-described temporal thermal filtering of AVHRR channel 4 proved to be the most reliable. However, in this study the cloud masking has been made using a combination of the temporal thermal filtering and the CASPR methods: cloudy areas were defined as those where the temporal filtering or either of the CASPR algorithms detected cloud. This combination most likely has a tendency to over-estimate the number of cloudy events but ensure the clear-sky surface pixels that can give valid albedo and temperature estimates. Even in this method, some uncertainties may still remain, but for lack of a better method this procedure is employed in this study.

2.2. Data validation

Besides the uncertainties in the cloud masking, reasons for inaccuracies in the APP data can be due to measurement errors like the calibration of the AVHRR instrument as well as errors in the APP algorithm used to derive surface albedo and temperature from the satellite measurements. The revised calibration formulae presented in Rao and Chen (1999) yield relative albedo values in adjacent grid cells to within 5%. The calibration procedure of Walton et al. (1998) reproduces temperature measurements with an accuracy of 0.1–0.2 K. Algorithm errors may arise from the bidirectional correction from reflectance to hemispherical albedo, narrow-to-broadband conversion and atmospheric correction. The absolute error for the bidirectional correction in the global mean albedo has been estimated to be nearly 0.02 (Stroeve et al., 2001 and references therein). The narrow-to-broadband conversion based on the conversion factors is dependent on the atmospheric conditions. According to Stroeve et al. (2001), the absolute error is estimated to be about 0.05 in albedo at solar zenith angles lower than 65°. Analyzing the errors in the atmospheric correction, Stroeve et al. (2001) made a simple test using the 6 S radiative transfer model for a top-of-atmosphere albedo of 0.8. Using constant values of ozone, water vapour and aerosols, a 6% difference between surface and top-of-atmosphere were found. Most of the differences between the surface and the top of the atmosphere were due to atmospheric water vapour. In the atmospheric correction algorithm used for APP data, the different amounts of water vapour at lower and higher elevations are somewhat taken into account using information from AVHRR channels 4 and 5.

The APP products have been validated in several validation studies e.g. Wang and Key (2005), Maslanik et al. (2001) and Stroeve et al. (2001). Wang and Key as well as Maslanik et al. validated the products with data collected during the Surface Heat Balance of the Arctic Ocean (SHEBA) field experiment in the western Arctic. In addition Wang and Key used data from two Antarctic meteorological stations: South Pole and Neumayer. Stroeve et al. (2001) compared the APP surface albedo 5 km EASE-Grid Composites with the albedo measured at 14 automatic weather stations (AWS) around the Greenland Ice Sheet from January 1997 to August 1998. According to the results of Wang and Key, the bias for albedo (in absolute units) was −0.05 and rmse 0.10. For surface temperature the values of bias and rmse were 0.20 K and 1.98 K respectively. According to Maslanik et al. the bias for albedo was −0.03 and rmse 0.08. The results of Stroeve et al. (2001) showed that the APP surface albedo values were, on average, 10% less than those measured by the AWS stations. The stations employ a LI-COR 200SZ photoelectric diode to measure incoming and reflected solar radiation. The LI-COR instruments were calibrated against Eppley Precision Spectral Pyranometers: a tendency was found for the LI-CORs to be positively biased by about 4%. The
Fig. 1. Comparison of surface temperature monthly means observed in November, December and January at three Antarctic stations (Marble Point, Vostok and South Pole) with those derived from the APP data. The trends are calculated from spring–summer means.
differences in absolute albedo between the APP and stations measurements may thus be about 6%.

An approximation for the general accuracies for APP albedo and temperature, based on the above-mentioned studies among others, is given by Fowler et al. (2000). The accuracies for albedo and temperature, including the uncertainties of cloud masking, are approximated to be ±0.05 and ±2 K, respectively.

In the present study the APP-derived monthly mean surface temperatures in November, December and January from 1981 to 2000 are validated with those of three Antarctic meteorological stations, Marble Point, Vostok and South Pole (READER data by Turner et al., 2004). The correlation coefficients were 0.78, 0.99 and 0.99 with standard errors of 2.03 °C, 0.82 °C, 0.93 °C, and biases of 0.05 °C, −0.86 °C and −1.81 °C respectively. In the light of these results, the temperature accuracy seems to be consistent with that estimated by Fowler et al. (2000). The scatter plots and trends can be seen in Fig. 1.

2.3. Sea ice area

For the analysis of sea ice albedo and surface temperature, as well as that for sea ice extent, it is necessary to define the “ice area”. The sea surface in the study region varies from open water to a 100% sea ice concentration. The concentration classification for computing the albedo has been made using the EASE-Grid surface type mask based on SMMR and SSM/I data. The EASE-Grid classification has been constructed with intervals of 10%.

The aim here was to use the same 15% ice concentration cutoff as in the above-mentioned data set of sea ice extent. The lowest concentration interval was therefore chosen to be from 10 to 20%. This means that all the pixels covered by first-year ice and multiyear ice having an ice concentration between 10% and 100% have been included in the sea ice albedo and temperature estimations. Therefore, in this research, the term “sea ice albedo” and “sea ice surface temperature” mean the albedo and temperature of the sea ice region, not only the albedo and temperature of the actual ice floes (see Laine, 2004). In addition, for the albedo and temperature analysis, only those pixels have been included that have measurable values in every one of the years 1981–2000. This means that the ice area used for these calculations has been limited to be the same for every year.

3. Time series and trends

The variability and trends in the Antarctic ice albedo as well as those for the surface temperature, sea ice concentration and the sea ice extent are presented in the form of time series of spring–summer averages for the years 1981–2000 (the spring–summer averages are calculated from the beginning of November for the year in question to the end of January of the next year). It should be noted that the spring–summer average for the year 2000 consist of November and December data only (January 2001 is missing). The albedos and surface temperatures are calculated for the total ice cover, ice sheet and sea ice separately.
Fig. 3. Spring–summer-averaged Antarctic and Weddell Sea sector ice sheet and sea ice albedos, ice sheet and sea ice temperatures, sea ice concentrations and sea ice extents from 1981 to 2000.
Fig. 4. Spring–summer-averaged Indian Ocean and Pacific Ocean sector ice sheet and sea ice albedos, ice sheet and sea ice temperatures, sea ice concentrations and sea ice extents from 1981 to 2000. S indicates statistical significance at the 95% and 98% confidence levels, using a standard $F$-test.
Fig. 5. Spring–summer-averaged Ross Sea and Bellingshausen–Amundsen Sea sector ice sheet and sea ice albedos, ice sheet and sea ice temperatures, sea ice concentrations and sea ice extents from 1981 to 2000. S indicates statistical significance at the 95% and 98% confidence levels, using a standard F-test.
The time series are calculated for the Antarctic region as a whole, and for five longitudinal sectors: the Weddell Sea (WS), the Indian Ocean (IO), the Pacific Ocean (PO), the Ross Sea (RS) and the Bellingshausen–Amundsen Sea (BS) (Fig. 2). The albedos and the surface temperatures for these sectors are calculated by time- and area-averaging all the values from pixels identified as ice for each region. The corresponding time series of sea ice concentration and sea ice extent for annual spring–summer seasons have been calculated by averaging the monthly sea ice concentrations and extents for the same months, i.e., November, December and January.

The trends have been estimated using the method of least squares for determining the best linear fit for the data. The values for the slopes thus found with their estimated standard errors for albedo, surface temperature, sea ice concentration and sea ice extent, as well as the correlation coefficients \( r \) between albedo, temperature, concentration and extent for the sectors mentioned above are indicated in Figs. 3–5. The statistical significance of each trend was tested using a standard \( F \)-test at confidence levels of 95% and 98%. The letter S marks the statistical significance of those trends for which the \( F \)-observed statistics are greater than the \( F \)-critical values. When forming averages over time and area, it can be assumed that the effects of individual measurement errors will be considerably reduced. For this reason the statistical trend analysis here has been carried out assuming zero measurement error.

The Antarctic region as a whole and all the sectors separately show slightly positive spring–summer albedo trends. However, most of these trends are not statistically significant. The albedo trends are statistically significant in the PO sector for total ice and sea ice, and in the RS sector for the ice sheet. All the regions show negative spring–summer surface temperature trends for the study period. Statistically significant temperature trends can be found in the BS sector for total ice and in the PO and BS sectors for sea ice. The spring–summer sea ice concentration trends are positive except in the IO and BS sectors. The PO sector shows the steepest positive sea ice concentration trend; it is also statistically significant. The sea ice extent trends are slightly negative in each region, except for the steeper negative trend in the BS sector and the relatively steep positive extent trend in the RS sector. However, none of these trends is statistically significant even at a confidence level of 95%.

4. Spatial distribution

To illustrate the average spatial distribution of the albedo and temperature, maps have been prepared of the average spring–summer albedo and surface temperature of the whole Antarctic ice cover for the entire 20 years (Figs. 6 and 7). In addition, a map of the average sea ice concentration distribution has been generated for the same time period (Fig. 8). As mentioned in Section 2.3, a sea ice concentration cutoff of 15% has been employed for the analysis. This means that open water pixels have been rejected from the averaging. Because the sea ice extent decreases substantially from November to January, the spring–summer sea ice averages shown in the spatial maps consist mainly of the November and December ice.

The average albedo for the total Antarctic ice area (including the ice sheet and sea ice regions) is 0.71 (Fig. 6). The ice sheet albedo varies between 0.75 and 0.90, having an average value of 0.83. Lower albedo values (0.60–0.70) occur over the Amery Ice Shelf on the eastern coast of Antarctica and in the Transantarctic Mountains south of the Ross Ice Shelf. The sea ice albedo shows a larger spatial variation ranging from 0.25 to 0.75. The average value for the spring–summer sea ice albedo over the whole Antarctica is about 0.60. Lower albedo values (0.25–0.50) can be found in the continent’s coastal region and at the outer edge of the sea ice. The highest sea ice albedo values (0.70–0.75) occur in the Weddell Sea near the continental coast.

The average spring–summer surface temperature for the total Antarctic ice area is about −17 °C. The ice sheet surface temperature varies strongly with topography, having a minimum of below −35 °C in East Antarctica, where the ice dome rises to over 4000 m (see Figs. 2 and 7). Especially in the IO and PO sectors the ice sheet temperature is low, below −30 °C over large areas. In western Antarctica the spring–summer temperature of the ice sheet is much higher, varying from −10 to −20 °C in the Ross and Ronne Ice Shelf regions. In the coastal regions around Antarctica the spring–summer temperature varies around −10 °C. The sea ice temperature is significantly higher than the ice sheet temperature, varying from about −5 °C to −13 °C.

The spring–summer sea ice concentration varies strongly from the sea ice margins to the continent’s edge, having an average of about 60%. Remarkably high concentrations (95–100%) occur in the Weddell and Bellingshausen–Amundsen Seas. High concentrations are also to be found in the eastern coastal regions.

For each longitudinal sector, the average albedos, temperatures, sea ice concentrations and sea ice extents with the corresponding standard deviations are to be found in Table 1. The highest ice sheet albedos (0.85) and lowest surface temperatures (−25.0 °C) occur in the IO and PO sectors. A very similar ice sheet albedo (0.84) can be found in the WS sector, at, however, a significantly higher average temperature (−20.1 °C). The ice sheet albedo is lowest in the RS and BS sectors, where...
the average albedo values are 0.80 and 0.78 respectively. In these sectors the average temperature is much higher having values of −16.3 °C and −16.1 °C. The average sea ice albedo is much lower than that of the ice sheet, having values of about 0.60 in the WS, RS and BS sectors. Even lower values (0.56 and 0.54) occur in the IO and PO sectors. The temperature variability between the sectors is notably smaller in the sea ice area than in the ice sheet area, ranging from −10.3 °C (PO) to −12.8 °C (WS). The sea ice extent varies considerably between the sectors. The largest sea ice extent occurs in the WS sector (3.8×10^6 km^2) and the smallest in the PO sector (0.9×10^6 km^2).

The standard deviation describes the spatial variability inside each sector. The standard deviations of albedo for the ice sheet are about 0.02 in each sector. The corresponding values for sea ice are twice as large, being about 0.05. In the case of surface temperature the situation is the opposite. The standard deviations of surface temperature in the ice sheet sectors are significantly larger than those in the sea ice sectors. In the case of sea ice concentration, the standard deviation of the BS sector is notably larger (4.7%) than those (2.3–3.1%) of the other sectors. The standard deviations of the sea ice extent range from 0.1 to 0.3×10^6 km^2, being lowest (0.1×10^6 km^2) in the regionally smallest PO sector and highest (0.3×10^6 km^2) in the largest WS sector.

The average spring–summer sea ice albedo values for each longitudinal sector from this APP data set have been compared to those of ship-based field experiments from the Antarctic Sea Ice Processes and Climate (ASPeCt) data set by Bradt et al. (2005) (Table 2). The ASPeCt data set includes more than 10,000 observations from ships for the years 1980–2000. To be more comparable with the APP spring–summer albedos that are averaged for the months of November, December and January, the ASPeCt albedo values are re-averaged from the spring (SON) and summer (DJF) seasons for the same latitudinal ice areas (ice concentration at least 15%) as in the APP albedo estimates. As seen from Table 2, the APP derived albedos are somewhat higher (about 0.1) than those of ASPeCt in every longitudinal sector.

The normal distribution of the standard deviation of albedo in each ice sector, as well as the standard deviation of sea ice extent in each sector, is shown in Table 1.
In addition, APP albedos have been compared with those from selected data-sets of three Italian Antarctic campaigns and from the German Neumayer Station (Pirazzini, 2004). The comparison has been made employing clear-sky albedo measurements from three stations: Neumeyer Station, Hells Gate on the ice shelf near the Ross Sea and Dome Concordia situated on the Antarctic Plateau. The corresponding APP albedos for the same months have been averaged over about 15 km × 15 km regions centered on the station locations.

As seen from Table 3, the APP-derived albedos are consistent with those of the stations except at Dome C, where the APP albedo is a little higher.

Spring–summer albedo, temperature and concentration anomaly maps provide an illustrative way of estimating overall changes from year to year (Figs. 9, 10 and 11). These maps are generated by taking the difference between the spring–summer albedo, temperature and concentration for each year and the corresponding averages for the entire 20 years. The anomaly maps show that albedo, temperature and concentration all vary significantly from year to year during the study period. In addition, the sea ice margin variations can be visually estimated from the concentration anomaly maps.

Relatively strong positive albedo anomalies for both the ice sheet and sea ice regions are to be seen in the years 1983, 1988 (except the Ross Sea), 1993, and 1999. In addition, a strong positive sea ice albedo anomaly occurs in 1984, and especially strong positive sea ice albedo anomalies can be found in the years 1993 and 2000 in the Weddell Sea, Bellingshausen–Amundsen Sea and Ross Sea. Negative ice sheet and sea ice albedo anomalies occur in the years 1981, 1982, 1986 and from 1995 to 1997. A regionally small, but strong negative sea ice albedo anomaly in the Weddell Sea seaward of the Ronne Ice Shelf in 1997 is noteworthy. Also, it is interesting to contrast the strong negative ice sheet albedo anomaly in eastern Antarctica with the strong positive sea ice albedo anomalies in the above-mentioned sea ice regions.


The anomaly maps of sea ice concentration show a sea ice concentration change from year to year similar to the change appearing in the albedo anomaly maps. However, the connection is not evident in every spring–summer period studied. For example, in the spring–summer of 1983 the albedo is high in most sectors, whereas the concentration is relatively low. A close connection between albedo and concentration can be most clearly noticed in the spring–summers of 1997 and 2000, especially in the western sea ice regions. Also a correspondence between cold regions and high albedos, as well as a correspondence between warm areas and low albedos, is to be seen when comparing the temperature and albedo anomaly maps. The relationship between albedo and temperature anomalies appears most obviously in the period from 1992 to 2000.

5. Spatial distribution of the albedo, temperature and concentration trends

To provide a detailed view of the 20-year albedo, temperature and concentration trends, pixel-based spatial trend distribution maps were calculated for the entire Antarctic ice area (Figs. 12a, 13a and 14a). The trends were calculated pixel-by-pixel using a multiple linear regression on the spring–summer data. The corresponding maps of F-values, representing the statistical significance distribution for the trends, were also calculated (Figs. 12b, 13b and 14b).

The statistical significance distribution maps of both albedo and temperature show that the trends are statistically strongly significant only in the Antarctic coastal region and along the Transantarctic Mountains. The sea ice concentration trends are mostly statistically significant in spotlight areas around Antarctica.

Examination of the albedo trend distribution map shows that the trends are positive for most of the Antarctic ice sheet and sea ice areas. In particular, the coastline region shows a strong increase in albedo, as well as do large areas of western Antarctica. Negative ice sheet albedo trends can be found mostly in the eastern part of Antarctica. In the sea ice region, negative albedo trends occur mainly in the WS and IO sectors. A smaller region of strong negative trend in albedo can be seen in the Ross Sea near to the Ross Ice Shelf. Large areas of both negative and positive temperature trends can be seen over the Antarctic ice sheet. In the sea ice region, positive sea ice temperature trends occur mostly in the IO sector. However, the strongest local positive trend occurs in the Ross Sea seaward of the Ross Ice Shelf. The coastline region shows strong negative temperature trends.

Correlation maps have been made to discover the spatial and temporal correspondence between surface albedo, temperature and sea ice concentration trends (Fig. 15). These maps are generated by calculating the temporal correlations pixel-by-pixel between albedo and surface temperature, albedo and sea ice concentration and also surface temperature and sea ice concentration for the entire 20 years. The albedo and temperature trends show a negative correlation over most of the sea ice and ice sheet areas (Fig. 15a). However, some regions of weak positive correlations in the IO and PO ice sheet sectors are to be seen. The strongest negative correlations occur in eastern Antarctica in both ice sheet and sea ice regions. The correlation between albedo and sea ice concentration

### Table 2
Average spring–summer albedos derived from the APP and ASPeCt data sets for the five longitudinal sectors

<table>
<thead>
<tr>
<th>Albedo</th>
<th>Data/sector</th>
<th>WS</th>
<th>IO</th>
<th>PO</th>
<th>RS</th>
<th>BS</th>
</tr>
</thead>
<tbody>
<tr>
<td>APP</td>
<td>0.61</td>
<td>0.56</td>
<td>0.54</td>
<td>0.60</td>
<td>0.60</td>
<td></td>
</tr>
<tr>
<td>ASPeCt</td>
<td>0.53</td>
<td>0.48</td>
<td>0.40</td>
<td>0.48</td>
<td>0.52</td>
<td></td>
</tr>
</tbody>
</table>

### Table 3
Albedo

<table>
<thead>
<tr>
<th>Albedo</th>
<th>Data/station</th>
<th>Neumayer</th>
<th>Hells Gate</th>
<th>Dome C</th>
</tr>
</thead>
<tbody>
<tr>
<td>APP</td>
<td>0.83</td>
<td>0.71</td>
<td>0.84</td>
<td></td>
</tr>
<tr>
<td>Pirazzini (2004)</td>
<td>0.82</td>
<td>0.73</td>
<td>0.80</td>
<td></td>
</tr>
</tbody>
</table>
Fig. 9. Anomalies of spring–summer ice albedo 1981–2000.
Fig. 10. Anomalies of spring–summer ice temperature 1981–2000.
Fig. 11. Anomalies of spring–summer sea ice concentration 1981–2000.
trends is mainly positive around the sea ice area (Fig. 15b). Relatively very strong correlations occur in the coastal region and in the WS sector near the Antarctic Peninsula. The correlation between surface temperature and sea ice concentration trends is somewhat unclear (Fig. 15c). Weak, both positive and negative spotlike correlations can be found all over the sea ice regions in every sector. Maybe the most homogeneous area of negative correlation is to be seen in the WS sector on the coast of the Peninsula.

6. Relationships with the Southern Oscillation

The Southern Oscillation is the result of a cyclic warming and cooling of the central and eastern Pacific Ocean that has significant effects on the sea level pressure. In particular, the difference between the pressure measured at Tahiti and that measured at Darwin can be used to define an index number called the Southern Oscillation Index (SOI). When the SOI is

Fig. 12. (a) 20-year change in the spring-summer ice surface albedo. (b) Statistical confidence levels of the spring–summer albedo trends. The green and red areas indicate 95% and 98% confidence levels respectively.

Fig. 13. (a) 20-year change in the spring-summer ice surface temperature. (b) Statistical confidence levels of the spring–summer temperature trends. The green and red areas indicate 95% and 98% confidence levels respectively.
positive, La-Niña (or tropical Pacific ocean cooling) is the case, whereas, when the value is negative El-Niño (or tropical Pacific ocean warming) is indicated. Monthly long-term SOI values have been calculated by the Australian Government Bureau of Meteorology. For this study the SOI values have been taken from:


The SOI is calculated as follows:

\[
\text{SOI} = 10 \frac{\text{Pdiff} - \text{Pdiffav}}{\text{SD}(\text{Pdiff})}
\]

where

- \( \text{Pdiff} \)  (average Tahiti MSLP for the month) \(-\) (average Darwin MSLP for the month).
- \( \text{Pdiffav} \)  Long-term average of \( \text{Pdiff} \) for the month.
- \( \text{SD}(\text{Pdiff}) \)  Long-term standard deviation of \( \text{Pdiff} \) for the month.
- \( \text{MSLP} \)  Mean Sea Level Pressure

Using this method, the SOI value ranges from about \(-35\) to \(+35\).

To find associations between the SOI yearly variations and the behaviour of the Antarctic sea ice in the spring–summer seasons, the SOI values for the spring–summer months November, December and January of each year have been averaged. These averages have then been compared with the spring–summer Antarctic ice sheet and sea ice albedos and temperatures, sea ice concentration and sea ice extent for every longitudinal sector. The curves illustrating the yearly relationships between these variations are shown in Figs. 16, 17 and 18.

The correlation between albedo and SOI is only at a moderate level (about \(0.30\)–\(0.35\)) for both ice sheet and sea ice sectors being, however, slightly stronger for the sea ice areas. The correlation is lowest between ice sheet albedo and SOI in the PO sector (0.18). The negative correlation between surface temperature and SOI is somewhat stronger than that between albedo and SOI. The strongest correlations are to be found between sea ice temperature and SOI in the WS sector (\(-0.47\)) and between ice sheet temperatures and SOI in the IO and PO sectors, being \(-0.50\) and \(-0.58\), respectively. The correlation between sea ice concentration and SOI is relatively strong only in the IO sector (0.48). In the WS and PO sectors the correlation is zero. Weak correlations (0.20–0.30) occur in the RS and BS sectors. The correlation between sea ice extent and SOI shows large variation among the sectors. The correlation is weak and negative in the WS and PO sectors (\(-0.20\)) and zero in the BS sector. Relatively strong positive correlation can be found in the IO sector (0.44) and especially strong correlation in the RS sector (0.69).

7. Discussion

Over the entire Antarctic ice region, as well as in every longitudinal sector, the albedo, the surface temperature, the sea ice concentration and the sea ice extent all show substantial annual variability. Increasing spring–summer albedo trends and decreasing temperature trends are generally to be seen. However, in trend analysis it is always important to remember that removing some points or changing the starting or ending times of the trend period can change the trend significance. In addition, any extrapolation of the trend lines beyond the 20-year study period could lead to incorrect long-term estimates.

The most important source of error affecting the accuracy of the APP data may be cloud masking. In this study, the cloud masking has been done using the CASPR procedure version that was found to be most reliable by Stroeve et al. (2001).
According to their study, this cloud masking method works reasonably well at lower elevations but has a tendency to overestimate the number of cloudy events at higher elevations. This tendency can be considered to be an advantage for ensuring the clear pixels for valid albedo and temperature estimates.

The measurement and algorithm accuracies for albedo and temperature are difficult to determine because of the small number of station observations. Combining the uncertainties from several validation studies, mentioned in Section 2.2, the general accuracies, including the inaccuracies in cloud masking, are given by Fowler et al. (2000) as being approximately ±0.05 and ±2 K for albedo and temperature, respectively. The temperature validation against the Marble Point, Vostok and South Pole stations, in the present study, results in the same or even better temperature accuracy, particularly in the cases of South Pole and Vostok. The larger temperature deviation between satellite and station measurements at Marble Point may be due to the more heterogeneous geography in the satellite footprint area. However, this validation may lend further credence to the cloud masking method. It may be argued that the albedo and temperature accuracy levels are insufficient for estimating the long-term trends and variability of the ice. However, when working with monthly and spring–summer means, the general temporal and spatial estimates of APP albedo and temperature may be considered to be adequate, particularly when analyzing the annual variability in the ice. Similar conclusions have been drawn in Stroeve et al. (2001), Stroeve (2001), Comiso (2000), Wang and Key (2005) and references therein. In trend analysis we must be more cautious in estimating the trend significance. However, all the trends, regardless of the sectors or ice type, show quite similar annual variability and trends during the study period. This may lend support to the view that trends estimated here do represent real trends in the Antarctic ice.

At SMMR and SSM/I frequencies the microwave emissivity is the same for melt ponds as for the open water between ice floes (Comiso & Kwok, 1996). In the Arctic regions the estimation of sea ice concentration from SMMR and SSM/I data can be open to interpretation during the spring–summer period because of the development of melt ponds. In the Antarctic melt ponds are very rare (Bradt et al., 2005). Sea ice concentration estimates from SMMR and SSM/I data in the Antarctic sea ice regions may therefore be considered to be more reliable, although the algorithms for ice concentration are somewhat sensitive to changes in snow properties and ice thickness.

Satellite-based measurements using AVHRR data allow only clear-sky albedo estimations. The albedo of sea ice may be somewhat different under cloudy conditions, due to the different spectral distribution of the incoming short-wave radiation and changes in the ice surface scattering layer (Perovich et al., 2002). This research is, however, limited to consideration of clear-sky albedo measurements.

Several factors affect the clear-sky albedo in the Antarctic ice sheet and sea ice regions. On the Antarctic ice sheet, snow metamorphism is one of the main factors particularly in the coastal regions, where the summer temperatures are relatively high. In the interior of the continent, snow metamorphism is slower due to the much lower temperatures (Pirazzini, 2004). The albedo decreases when snow ages due to the increase in snow grain size. After a new snowfall the albedo is higher due to the small size of the snow grains. An albedo increase can also be

![Fig. 15. Correlation between: (a) albedo and temperature time series, (b) albedo and sea ice concentration time series, (c) temperature and sea ice concentration time series.](image-url)
due to snow drifting by wind. The smallest grains of the suspended drifting snow fall last on to the surface, causing an albedo increase (Grenfell et al., 1994; Pirazzini, 2004). On the Antarctic ice sheet the wind often generates irregular ridges of snow that lie parallel to the direction of the wind (sastrugi). The effect of sastrugi is to slightly decrease the albedo, because of shadowing especially at low solar elevations and light absorption by neighboring sastrugi (Pirazzini, 2004 and references therein). The Antarctic ice sheet topography can be seen in Fig. 2. The highland area of Antarctica is concentrated in Fig. 16. Trends of ice sheet and sea ice albedo, temperature, sea ice concentration, sea ice extent and the SOI for the Weddell dell Sea and the Indian Ocean.
the eastern part of the continent. The spatial average spring–summer temperature distribution follows the topography distribution, as would be expected (see Fig. 7). However, it is interesting to notice that the average spring–summer albedo is also distributed similarly to the topography (see Fig. 6). The highland area occurs mainly in the IO and PO sectors having average albedos of 0.85 (Table 1). Probable reasons for the higher albedo values in these highland sectors include the
deposition of finer snow grains and slower snow metamorphism in these colder regions, where the average spring–summer temperature is about \(-25\, ^\circ\text{C}\). In the Transantarctic Mountains, shadowing probably lowers the albedo. The solar radiation absorption by impurities in the snow is most probably insignificant in the high regions. The lower albedo values (about 0.75) in the coastal region, particularly in the Antarctic Peninsula and on the Ross and Ronne Ice Shelves in the RS and BS sectors and the Amery Ice Shelf in the IO sector could be explained by stronger snow metamorphism due to the higher average temperatures (about \(-16\, ^\circ\text{C}\)). In addition the snow wetting due to melting or summer rainfall may lower the albedo in the warmer coastal regions. It is also possible that the snow is not as pure in these regions as at higher elevations.

The snow-free sea ice albedo depends strongly on the ice thickness especially when the ice is thin (0–30 cm) (Lindsay, 2001). In regions of thin ice (low sea ice concentration), ice thinning or growth can be a significant cause of albedo change. However, in the Antarctic, bare ice is usually rapidly covered by snow. The albedo of the snow-covered sea ice depends primarily on the snow depth but also on the grain size and wetness. Flooding by seawater and wave overwashing are common phenomena in the Antarctic sea ice region throughout the year (Massom et al., 2001 and references therein). These processes lower the albedo by removing the snow cover and wetting the snow. Other factors affecting albedo, such as snowfall, drift snow transport and snow metamorphism may be significant causes of albedo changes also in the sea ice regions. In the WS sector very high sea ice concentrations occur adjacent to the Antarctic Peninsula and the Ronne Ice Shelf. The higher surface albedo follows the high concentration pattern. In fact, the same phenomenon can be seen in the sea ice regions around Antarctica. The low albedo values in the narrow coastal region around Antarctica are consistent with higher surface temperatures in that region. This may be connected with snow-covered sea ice melting and wetting, and thus a lowered albedo, induced by higher temperatures. A region having a significantly low albedo can be seen in the front of the Ross Ice Shelf, surrounded by sea ice with a much higher albedo. This has been explained as being a consequence of persistent synoptic winds off the Ross Ice Shelf and katabatic surge events causing a reduced sea ice concentration (see Fig. 8) or even the formation of a large coastal polynya (Zwally et al., 2002). In general, both albedo and sea ice concentration are lower in the eastern part of Antarctica. In the IO sector the low concentrations (and albedos) may be associated with the formation of eddies and the upwelling of warm water (Zwally et al., 2002 and references therein). The sea ice cover is lowest and may also be thinnest in the PO sector (Table 1), probably due to the northernmost-extending continental boundary (except for the Antarctic Peninsula). This leads to low sea ice concentrations and thus low albedo values, as can be seen in Figs. 6 and 8 and in Table 1.

The sea ice albedo is clearly higher in the western part in Antarctica (about 0.60 in the RS, BS and WS sectors) than in its eastern part (about 0.55 in the IO and PO sectors) (Fig. 6). The albedo observations from the ASPeCt experiment (Bradt et al., 2005) are parallel to these values, although a little lower (Table 2). The bias in the albedo values between the two may originate from several factors. One cause of the bias may be temporal and spatial differences between the ship- and satellite-based observations. Also, the methods for estimating the albedo are different. In contrast to direct APP-based albedo measurements, the ASPeCt...
albedos have been assigned using typical albedos for each ice type (Bradt et al., Table 1). The average ASPeCt albedos have then been obtained for each observation by weighting with the fractional coverage for each surface type. In addition, the seasonal ASPeCt albedo averages including both ice and water may also contain a higher proportion of open water than the APP albedos (sea ice concentration at least 15%). Some part of the bias might also be due to thin cirrus clouds. As shown in Table 3, the APP albedos compared with those selected from data sets of three Antarctic stations (Neumayer, Hells Gate and Dome C) are more consistent than the APP albedos compared with those of the ASPeCt. This may result from the direct albedo measurements by both stations and satellite. Also, the temporal and spatial differences are much smaller than between the satellite- and ship-based observations. One reason for the higher consistency may also be a more homogenous ice surface in the footprints of satellite centered on the station locations.

In general, the correlations between the averaged sea ice albedo and sea ice concentration time series, calculated for the whole study period, are somewhat lower in most sectors than the correlations between albedo and temperature. However, the averaged correlations represent the total areas of the sectors. In such cases some contrasting correlations inside a sector might somewhat obscure the correlation analysis. The correlation maps therefore provide a more detailed view of the spatial and temporal correlations between the albedo, temperature and sea ice concentration time series (Fig. 15). The correlation between the sea ice sheet and sea ice albedo and temperature is negative, especially in the sea ice area. This may be related to the albedo dependence on temperature changes in these areas. Smaller spotlight areas of positive correlations exist in the Weddell Sea and western Ross Sea. In the ice sheet area some extended locations with weak positive correlations occur in the IO and PO sectors (Fig. 15a). As seen in Fig. 12a, these patterns of zero or weak positive correlations in the ice sheet region are similar to the patterns of negative albedo trends. These patterns of negative albedo trends may be due to strong negative anomalies in the same locations in the years 1993 and 2000 (see Fig. 9). Corresponding patterns of temperature in those years cannot be found. However, by comparing the albedo trends (Fig. 12a) with those of temperature (Fig. 13a), the strongest increase in temperature occurs in the same highland region as some of the decreased albedo trends. As seen in Figs. 12b and 13b, the albedo and temperature trends are not statistically significant in these regions. Therefore, these correlation patterns cannot be explained only by changes in the surface temperature but partly by measurement inaccuracies, too.

The correlation between the sea ice albedo and concentration time series seems to be strongest in the narrow coastal region round the Antarctic, especially in the BS and WS sectors. Also the “tail” of high correlation eastward from the Antarctic Peninsula is noteworthy. It is interesting to see that strong correlations occur mainly in the coastal regions, including the “tail” where the average sea ice surface temperature is highest (see Fig. 7). This suggests that a sufficiently high sea surface temperature with some melt occurring is a prerequisite for strong correlations between the albedo and sea ice concentration.

Several studies have analyzed the relationships between Antarctic sea ice characteristics and the Southern Oscillation (SO) (Kwok & Comiso, 2002 and references therein, Liu et al., 2002, 2004; Turner, 2004; Yuan, 2004). A quasi-stationary wave in the sea ice linked to SO variability is called the Antarctic Dipole (ADP) (Yuan, 2004). The ADP consists of contrasting anomalies in the sea ice edge, sea surface pressure and sea surface temperature in the Bellingshausen–Weddell Sea region versus the Ross–Amundsen Sea region. To discover possible reasons for the Antarctic spring–summer ice variability, it is of value to also study here the potential relationship between the ADP and ice characteristics.

The possible indications of an ADP response in the spring–summer ice have been estimated by studying the variability and correlations of these with the Southern Oscillation Index (SOI) (Figs. 16, 17 and 18). The correlation between albedo and the SOI is at a moderate level. In the case of surface temperature, the variability of the correlation is a little larger, but no particular contrasts are to be seen between the sectors. Some attention must be paid to the relatively high correlation values between the ice sheet temperature and the SOI in the IO and PO sectors. In spite of low or moderate correlation values, some associations between the SOI and albedo and between the SOI and temperature are to be seen from the curves.

The sea ice concentration and the sea ice extent show much larger correlation variability between the sectors, both having relatively strong positive values as well as zero or even negative values. In the case of sea ice concentration, the correlation with the SOI is relatively strong only in the IO sector, being zero in the WS and PO sectors. The strongest correlation exists between the sea ice extent and the SOI in the RS sector, whereas it is reversed in the WS sector. In the light of these results, it seems that some indications of an ADP response in the spring–summer ice sheet and sea ice data occur mainly in the case of sea ice concentration and especially sea ice extent. The ADP response in the ice sheet and sea ice albedo seems to be smaller.

8. Conclusions

The principal results of the present study, are the increasing spring–summer albedo trends and the decreasing spring–summer temperature trends for Antarctica as a whole, as well as for all of the longitudinal sectors. The slight cooling trends seem to be parallel with the results of Comiso (2000), who studied Antarctic temperature trends using both satellite and station data. In the present study, all the regions show very similar interannual variability in albedo and surface temperature, especially the upward sea ice albedo trends in 1989–1993 and 1995–2000. The sea ice concentration shows slight increasing trends in most sectors, whereas the sea ice extent trends seem to be near zero, with the exception of the Ross Sea with a relatively strong positive trend and the Bellingshausen–Amundsen Sea with a steep negative trend. The pronounced trends of the Ross and Bellingshausen–Amundsen Seas are in agreement with the results of Liu et al. (2004) and Zwally et al. (2002).

Although the satellite data involves the before-mentioned uncertainties, they provide the only means of estimating the large
spatial variability in the albedo, temperature, sea ice concentration and sea ice extent. In the future, more station-based temperature and particularly albedo measurements will be needed for validation of the satellite data. Combined satellite and station data would provide the most reliable means for evaluating the variability of the ice sheet and sea ice in the Antarctic regions. More efforts are also required to improve cloud masking methods. The AVHRR, SMMR and SSM/I data provide useful series for investigating long-term changes in the sea ice cover. However, modern instrumentation is a prerequisite for further analysis of the sea ice cover. The Moderate Resolution Imaging Spectroradiometer (MODIS) aboard the Terra and Aqua satellites already provides higher spatial resolution for observing the ice surface albedo and temperature as well as the sea ice extent. One of the most interesting new satellites will be the CryoSat-2 that is due for launch in 2009. By monitoring precise changes in the thickness of the polar ice sheets and sea ice, this satellite will improve understanding of the effects of the global climate change on the polar ice caps. Future work will be focused on employing the data of these instruments.

Acknowledgments

I gratefully acknowledge the National Snow and Ice Data Center in Boulder, Colorado, for providing the AVHRR Polar Pathfinder 5-km EASE-Grid data. I am also deeply indebted to the Academy of Finland and the Finnish Funding Agency for Technology and Innovation for funding this study. The valuable comments of the two anonymous reviewers and my colleagues Dr. Terhiikki Manninen, and Robin King as well as Prof. Jarkko Koskinen are greatly appreciated.

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