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UV albedo of arctic snow in spring

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Abstract

The relevance of snow for climate studies is based on its physical properties, such as high surface reflectivity. Surface ultraviolet (UV) albedo is an essential parameter for various applications based on radiative transfer modeling. Here, new continuous measurements of the local UV albedo of natural Arctic snow were made at Sodankylä (67.37° N, 26.63° E, 179 m a.s.l.) during the spring of 2007. The data were logged at 1-min intervals. The accumulation of snow was up to 68 cm. The surface layer thickness varied from 0.5 to 35 cm with the snow grain size between 0.2 and 2.5 mm. The midday erythemally weighted UV albedo ranged from 0.6 to 0.8 in the accumulation period and 0.5–0.7 during melting. During the snow melt period, under cases of an almost clear sky and variable cloudiness, an unexpected diurnal decrease of 0.05 in albedo soon after midday, and recovery thereafter, was detected. This diurnal decrease in albedo was found to be asymmetric with respect to solar midday, thus indicating a change in the properties of the snow. Independent UV albedo results with two different types of instruments confirm these findings. The measured temperature of the snow surface was below 0°C on the following mornings. Hence, the reversible diurnal change, evident for ∼1–2 h, could be explained by the daily metamorphosis of the surface of the snowpack, in which the temperature of the surface increases, melting some of the snow to liquid water, after which the surface freezes again.

1 Introduction

The relevance of snow for climate variability and change is based on its physical properties, such as high surface reflectivity, i.e., albedo (IPCC, 2007). High albedo has an important influence on the surface energy budget and on Earth's radiative balance (e.g., Forster et al., 2007). Snow albedo varies with wavelength, and therefore the strength of the feedback depends on a number of factors, such as the depth and age of the snow cover, and the amount of incoming solar radiation and cloud cover. The albedo of snow may decrease because of anthropogenic soot (Wiscombe and Warren, 1980; Warren and Wiscombe, 1980) and aging and melting (Wiscombe and Warren, 1980; Blumthaler and Ambach, 1988; Wuttke et al., 2006). In the melting process, often initiated in springtime by the increase of shortwave irradiance, the snow grains are filled with liquid water and the density of the snow may increase, e.g., in the open sites of Arctic Finland, to 350 kg/m^3 (Kuusisto, 1984). When the albedo decreases, more radiation is absorbed, and the melting of the snow may increase due to this albedo feedback mechanism (e.g., Bony et al., 2006). In addition to its climate connections, ultraviolet radiation (UV) reflected from snow and ice may cause unprotected eyes the painful condition of snow blindness (UNEP, 2002). The UV albedo for a surface with snow is high, and also due to multiple reflections affects downwelling radiation (Bais and Lubin, 2007). Moreover, surface UV albedo is an essential parameter for various applications based on radiative transfer (RT) modeling, including various satellite retrieval algorithms. For example, current satellite UV algorithms demand better information on UV albedo, especially for land when covered by snow (e.g., Arola et al., 2003; Tanskanen and Manninen, 2007).

Accurate ground-based long-term albedo measurements of snow are somewhat sparse due to the harsh conditions with snow. Broadband albedo, measured with pyranometers, has been more widely studied (Pirazzini, 2004; Pirazzini et al., 2006; Wuttke et al., 2006), and spectral studies are less (Perovich et al., 2002; Wuttke et al., 2006). Only a few studies on UV albedo of snow have been published, some of them for Antarctic (e.g., Smolskaia et al., 1999; Wuttke, et al., 2006) and few for Arctic snow (e.g., Perovich et al., 2002), and moreover, most of them are campaigns, not continuous high temporal resolution measurements. As far as albedo is concerned, Arctic and Antarctic snow differ from each other especially in two ways: Antarctic snow has smaller snow grain sizes, and has more pure snow unaffected by impurities. Grain sizes of up to approx. 3 mm have been reported for the Arctic snow (Pirazzini et al., 2006). For these reasons, a lower albedo is expected for Arctic snow.

On the basis of earlier studies by others, it was our hypothesis that snow melt will decrease the UV albedo of Arctic snow. Moreover, it was our goal to study, using continuous high temporal resolution 1-min measurements over the whole melt season, how the decrease actually happens: in a single dramatic change, step-by-step or decreasing more or less linearily, little by little. Also, our aim was to study how the temperature of the air and ground, grain size of the snow, and amount of snow, as well as incoming irradiance and cloudiness possibly affect the albedo. To avoid misinterpretation of the experimental data, the error sources have to be considered. A detailed analysis of the uncertainties in UV measurements is available in Bernhard & Seckmeyer (1999), and azimuthal errors in spectral UV data have been explained in Meinander et al. (2006). The effects of instrumental uncertainties, such as calibration and cosine error, atmospheric parameters, solar zenith angle, and geometric aspects, like slopes and shadows, are considered in Sects. 2 and 4, and discussed further in Sect. 5.

2 Materials and methods

2.1 UVB albedo measurements

New polar Arctic measurements on the local UV albedo of snow were planned and carried out in 2007 at Sodankylä, (67.37° N, 26.63° E, 179 m a.s.l.), Finland. For UV albedo measurements, two sensors of the UVBiometer Model 501 from Solar Light Co. (SL501) with similar spectral and cosine responses (Figs. 1 and 2) were used, one facing upwards and the other downwards at a height of 2 m. The SL501 spectral response resembles the action spectrum for erythema, wavelengths in the UVB (280–310 nm) being most weighted. For the albedo measurements, a fixed device for the setting-up and support of the two sensors, including independent levelling possibilities for the upward and downward SL501s, a blower to keep the sensors defrosted, and a data logger system, was planned and constructed at FMI (Fig. 3). Data were logged into the data base from the 25 Feb till 15 May 2007 at 1-min-intervals. This period included various phases, including both the accumulation and melting of snow. The albedo of snow (A) was calculated from the ratio of downwelling UV irradiance on to upwelling irradiance ($UV_{ery}\uparrow/UV_{ery}\downarrow$) measured at 2π.

2.2 Multiband data

In order to gain wavelength-dependent snow albedo information, the reflected irradiance was also measured at 1-min intervals with an multibandfilter radiometer (MBFR, NILU-UV type) radiometer, placed facing downwards close (3 m) to the SL501 albedo sensors. These multichannel measurements were made from 6 February till end of May 2007. In addition, one NILU-UV radiometer facing upwards was situated close by (30 m), on the roof of the Sodankylä Observatory. The NILU-UV radiometer measures UV in five channels with central wavelengths around 305, 312, 320, 340 and 380 nm, and bandwidths of around 10 nm at FWHM. A sixth channel measures photosynthetically active radiation (PAR) in the range of 400–700 nm. With these channels, UVA and UVB, and erythemally-weighted UV albedo can be calculated from the ratio of downwelling irradiance to upwelling irradiance, $UVA\uparrow/UVA\downarrow$, $UVB\uparrow/UVB\downarrow$, and $UV_{ery}\uparrow/UV_{ery}\downarrow$, measured at 2π. The characteristics of the instruments are described more in detail in Hoiskar et al. (2003).

2.3 Empirical calibration

The aim of this work was to have an understanding of this valuable empirical data set on albedo, measured under the hard conditions of Arctic snow, without introducing any additional uncertainties or errors to the data due to imperfect correction procedures, nor to let the uncertainties and errors in the original data to affect the final results.

During the winter months, the sun does not rise at all at Sodankylä. Even at the beginning of the measurement period in February, the sun is still very low. On 15 March the midday SZA falls less than 70.0 degrees for the first time (Table 1). With an SZA larger than 70 degrees, the cosine error increases dramatically (Fig. 2), but as most of the irradiance is then diffuse (at 300 nm more than 90%, e.g. Madronich, 1993), this has only little impact on the measurement results. However, the amount of radiation reaching the Earth is then minimal (UVI<1), increasing the uncertainties in the measurements. On the other hand, during the measurement period, the midday SZA significantly decreases (SZA<55 degrees, UVI>3) as the length of the day grows. To avoid misinterpretation of data, knowledge of the SZA is thus essential.

The optimum calibration of the raw signal requires calibration matrix with SZA and ozone (Webb et al., 2006):

$$E_{\mathrm{CIE}} = U - U_{\mathrm{offset}} * C * f_n(\mathrm{SZA}, \mathrm{TO_3}) * \varepsilon(T) * \mathrm{Coscor} \quad (1)$$

where E_{CIE} is the erythemal effective irradiance, U is the measured electrical signal from the radiometer, U_{offset} is the electrical offset for dark conditions, C is the calibration coefficient (a constant value determined for specific conditions like SZA=40° and O_3=300 DU), f_n is a function that can be expressed as a calibration matrix normalized at, e.g., SZA=40° and O_3=300 DU, $\varepsilon(T)$ is the temperature correction function, and Coscor is the cosine correction function.

Here, the calibration factors (C) of the sensors had been determined by the Finnish Radiation and Nucleation Safety Authority (STUK), before the albedo measurements (in 2005 and 2006), as an average for conditions with 38<SZA<60 degrees. The ratio of the calibration coefficients of the sensors was then C1/C2=1.13. After the albedo measurements, the calibration was carried out again, giving C1/C2=1.19. The sensor with the originally better response was installed to measure the upwelling reflected radiation.

Keeping in mind that i) the spectral responses of the sensors were measured to be similar (Fig. 1), and ii) the differences in the cosine responses were relatively small (Fig. 2), and iii) we are only interested in the relative signals of these two sensors, the calibration coefficients C1 and C2 might be sufficient for comparison of data representing a narrow SZA range, e.g., within 2–4 degrees. We also know that the upwelling irradiance measured by the SL501 is not as much affected by the cosine error, due to the missing direct component. The albedo derived from the SL501 data may therefore be an overestimation of the real albedo under a clear sky and high solar elevation angles, and uncertainties and errors minimize with increasing diffuse radiation under full cloudiness or lower sun.

However, due to the very low sun (SZA>60°), the use of the laboratory calibration as such was considered unsatisfactory. Several solutions existed: 1) New empirical calibration coefficients (C1 and C2) for the prevailing SZA conditions were produced on 27 March by turning both of the sensors upwards and calibrating them against each other; both of them were also calibrated against an SL501 placed on the roof of the observatory. The sensors were again similarly calibrated on 2 May, and also on 11 May after snow melt. 2) The data of the prior and subseqment calibrations as a function of SZA were available. These data could be used to produce a simple SZA correction. 3) It would be possible to use a radiative transfer model to calculate the calibration as in Eq. (1).

Here, the empirical calibration approach was used. The aim was to produce empirically calibrated data with error estimates for the prevailing SZAs, without introducing any additional uncertainty or error in the data, due to a simplified SZA correction, as will be discussed more detailed in Sect. 5.

The empirical calibration factors were determined independently using two different SL501 sensors as references. First, the roof SL501 was used as a reference for both of the two albedo measurement sensors. Then, one of the albedo sensors was used as a reference for the other. In addition, empirical calibration procedures were carried out after the snow melt (Table 2).

The empirical calibration factors were calculated on the basis of the measurements: C1=1.09, C2=0.71 for 22 March, and C1=1.29, C2=0.96 for both 2 May and 11 May. The ratio of the calibration factors C1/C2 in March was =1.54 for SZA 67–69 degrees. The ratio of the coefficients was the same whether the independent roof SL501 was used as a reference for both the sensors independently, or one of the two albedo SL501 sensors was used as a reference for the other. In May, the ratio C1/C2 for both occasions, on 2 May (60 deg<SZA<56 deg) and 11 May was 1.34. It seems that the ratio C1/C2 could possibly decline with the decline of SZA. The same fall in the C1/C2 ratio was evident when studying the data of the solar measurements made in 2007 by STUK for calibrating the same sensors (Fig. 4). It is possible, e.g., that the spectral response of one or both of the sensors had changed since its determination, as will be discussed more in Sect. 5. In any case, we can conclude that we have an SZA dependent-uncertainty in the data.

We can minimize or eliminate the SZA dependency effect by i) picking the corresponding SZA moments for each day, or ii) dividing albedo results into temporal subgroups based on SZA and snow conditions within which the daily variations are similar, and using the same empirical correction coefficient within the shorter period. Both of these approaches were used here, but the first was considered the better of the two, and was used for the more detailed studies for calculation of the empirically calibrated albedo A for SZA 56–60 degrees:

$$A = \frac{\mathrm{C2UV_{ery}} \uparrow}{\mathrm{C1UV_{ery}} \downarrow} \tag{2}$$

where C1 and C2 are the empirically determined calibration factors for the SZA 56–60 degrees, and $UV_{ery} \uparrow$ and $UV_{ery} \downarrow$ are the simultaneously measured upwelling and downwelling erythemally weighted irradiances.

The first occasion in 2007 on which a midday SZA<60 was achieved for a period of at least one hour was on 10 April. Hence, a continuous temporal data set including data measured with 56<SZA<60 degrees was obtained from 10 April until snow melting. These data were used as the core material for the current study.

The other subgroups used here for temporal data series, were based on SZA and snow, as follows: a) albedo during accumulation of snow in March and at the beginning of April with midday SZA>60 degrees using the 22 March empirical coefficients; b) albedo during the melting period from mid-April until snow melt and end of albedo measurements on 9 May with midday SZA <60 degrees. The difference caused by

ACPD

8, 4155–4198, 2008

these subgroup calibration differences was also estimated by combining the data sets produced with different calibration coefficients. This comparison reveals the error which had been in the data, if there had been one empirical calibration only, under too large SZA. These data were not used more in-depth studies but rather to understand the long-term variation of the albedo.

2.4 Ancillary data

At Sodankylä, the state of the atmosphere at a height of 2 m is measured once a minute by an automatic weather station (AWS). From these data, information on the beginning of rain, snow depth and cloud cover, e.g., can be gained.

Snow depth (h_s) and grain size (D) were measured at Sodankylä by one of the co-authors (Anna Kontu) from November 2006 until 14 May 2007, covering the whole albedo measurement period. The grain sizes of all the layers of snow were estimated visually regularily, approx. twice a week, by taking samples of snow on a screen with a 1-mm grid (Fig. 5). The sampling site was not exactly the same as that of the albedo measurements, but at a distance of 300 m under conditions that can be assumed to be similar. All other activities very close to the albedo measurement were forbidden; the diffusers were, however, cleaned if needed with as little disturbance to the surroundings as possible.

3 Ancillary results

3.1 Snow depth

In 2007, the maximum snow depth at Sodankylä was 68 cm, on 21 March (Fig. 6).

The second highest value of 67 cm occurred on the day before and the day after, but also on 12 April. After 12 April (julian day 102), the snow depth decreased monotonically until totally melted.

ACPD
8, 4155–4198, 2008

3.2 Temperature

On 25 February (julian day 56), the daily maximum air temperatures (Fig. 7) indicated a rapid jump, referred here as "a springthaw", from values below zero degrees Celsius to values above zero. Neither the daily minimum temperatures nor the minimum temperature of the ground showed any similar rapid change.

Between 1 March–15 May, the measured minimum temperature of the ground at 06:00 UTC rose above 0°C on only six days. Otherwise the temperature remained below zero. From this it follows that, apart from these few cases, the snow surface was always frozen in the evening, night and early morning hours, lacking solar warming. On the other hand, the "springthaw" was not enough to start the snow melt alone, thus indicating the importance of radiation as the starting force for the snow melt.

3.3 Snow grain size

In the UV range of wavelengths, the reflected signal comes from the very surface, and only the surface layer grain size data were used here.

In 2007, the measured thickness of the surface layer varied from 0.5 to 35 cm. The surface layer snow grain size results can be divided into two groups: before and after 16 April (Fig. 8). Before this date, the snow grain sizes were most often <0.5 mm. Thereafter, the grain size was most often 1.0–2.5 mm, indicating the beginning of the actual snow melt period. However, there are no grain size results available between 10–16 April, and so the snow metamorphosis, with increasing grain size, began sometime within that period.

Grain size data was then studied to determine whether grain size had a relationship to temperature, snow depth and time. Such a relationship would mean that an albedo model could possibly predict Arctic snow UV albedo as a function of time and temperature, rather than of grain diameter. Using the three variables with the highest linear correlations (Table 3), the empirical relationship, giving an 82% explanation (r^2=0.82)

ACPD
8, 4155–4198, 2008

for the Arctic snow grain size during the melting period was:

$$D = -0.04t + 0.19T_{max} - 0.03h_s + 4.98 \quad (3)$$

where D is grain diameter, t is time (in julian days) and T_{max} is the daily maximum air temperature at a height of 2 m (in Celsius) and h_s is the height of the snowpack.

Even before the "springthaw" date, there were occasionally days with maximum air temperatures T_{max}>0 degrees. On such days, e.g. on 26 and 29 March, the snow grain size was measured to be 2 mm.

4 UV albedo of snow

In addition to SZA information, the results were grouped according to cloudiness, as well as accumulation and melting of snow. We studied first the almost clear sky cases for SZA<60 degrees, followed by cases of variable cloudiness, thereafter cases prior and after melt period, as well as averaged daily albedo during melting. Finally, long term variation in albedo was studied. The most important results were findings of an unexpected diurnal change in albedo during melting.

An overview of the albedo results for SZA<60 degrees conditions in April is shown in Table 4. These results represent all cases of cloudiness. The minimum and maximum of the daily SL501 mean albedo, for SZA 56–60 degrees, are included, indicating the diurnal changes. In some cases the results also suggested an asymmetric albedo, i.e. albedo decreasing from morning to afternoon. The results of the NILU-UV albedo are not included in Table 4, due to their use as complementary data for the current study.

4.1 UV albedo under almost cloudless sky

The signal of the upward sensor was used to study the cloudiness of the sky at exactly the same location and moment at which the albedo measurements were performed.

ACPD
8, 4155–4198, 2008

For 56<SZA<60, i.e., from 10 April onward, the almost-clear or clear-sky cases in April were: the 10, 15, 18, and 22 of the month (Table 4). Of these four cases, the first one occurred at the end of the snow accumulation period, and the last three during the melting season.

4.1.1 Stable albedo during accumulation period

The first case, on 10 April (julian day 100), was an almost clear day during the snow accumulation period. The snow UVB albedo at midday was 0.64–0.66 (Fig. 9), slightly increasing as the sun reached its highest elevation. The slight variation in the incoming irradiance had no effect on the albedo. Similarily, the NILU-UV albedo showed stable UVA and UVB reflectance. The surface layer snow grain size was small, 0.3 mm. The air temperature varied between −19.8–2.3°C. At 06:00 UTC the temperature on the surface of the snowpack was −22.9°C. Next, the clear sky cases of melt season were studied.

4.1.2 Diurnal change in albedo during melting

The second, almost clear sky case of 15th April occurred at the very beginning of the snow melt period. During the measurements from 09:00–12:00 UTC with SZA 56–60°, the albedo of the snow unexpectedly decreased by 0.05 from 0.6 to 0.55, and then recovered (Fig. 10). The drop in albedo happened soon after solar midday.

The next two almost clear sky cases came in the middle of the melting season, when almost 1/3 of the accumulated snow had melted. On 18 April, a similar behaviour in snow albedo occurred: a clear change from 0.6 to 0.55 and a recovery back to 0.6. On 22 April, the same occurred again, but this time the albedo varied at a slightly lower level, changing from 0.58 to 0.53.

Hence, in all three cases at the snow melt period a slight drop in the general albedo level (Fig. 10), asymmetrical to the solar zenith angle, was observed to be superimposed on the general albedo level.

For a more detailed study, data on temperature, incoming irradiance, and multichannel NILU-UV albedo in the UVA and UVB channels were available:

- 15 April (julian day 105): the snow depth was 58 cm, and the measured air temperatures were between 2.2–12.2°C. At 06:00 UTC Tmin on the ground was close to zero (0.5°C), and in the next morning it was below the zero (−4.1°C); during that next day the snow depth decreased by 3 cm, with a very thick surface layer (35 cm) and large snow grain size (2.5 mm), indicating a major snow metamorphosism. In the NILU-UV data, both the UVA and UVB decreased, similarily to SL501 (Table 5).

- 18 April (julian day 108): air temperature was between −7.8–4.2°C, and the snow depth was 45 cm. The next morning, Tmin on of the ground was −4.9°C. In the NILU-UV data, both the UVA and UVB decreased, too. Two days later, a snow grain size of 2.0 mm was measured, while the surface layer depth was 17 cm, both indicating snow metamorphose and melting continuing intensively.

- 22 April (julian day 112): air temperatures ranged from −9.9 to 7.9°C, and snow depth was 42 cm. In the NILU-UV data, both UVA and UVB decreased again (Fig. 11).

Therefore, we know that the temperature on the snow surface was below zero in the next morning at 06:00 UTC in two cases, but in the first case did temperature remain slightly above zero (0.5°C). The largest grain size of 2.5 mm was measured in this case, too. In all cases, the NILU-UV data confirmed the unexpected discovery using SL501 data: a diurnal change of UV albedo asymmetrically to solar midday was detected using two independent measurement devices.

ACPD

8, 4155–4198, 2008

4.2 UV albedo under variable cloudiness

4.2.1 Stable albedo during accumulating snow

The smallest grain sizes of 0.2–0.3 mm, indicating new snow or snow below-zero conditions during the period of accumulating snow, were observed on four days. Those days were: 20 March, 3, 5, and 10 April (julian days of 79, 93, 95, and 100). The most reliable calibration is for 10th April, which was presented with the clear sky results. The next most reliable cases were on the 3 April: a midday, and stable, albedo of 0.68. On the 5 April the albedo was 0.72 and stable. These cases thus confirm the clear sky case: during the snow accumulation period, the albedo remains stable, increasing slightly, if at all, increasing in the midday period.

4.2.2 Diurnal albedo of melting snow under variable cloudiness

There were two days on which were measured the maximum snow grain size of 2.5 mm in the surface layer, i.e., 16 April and 8 May (julian days 106 and 128). On 17 April the grain size was measured to be 2 mm, with a 17 cm top layer.

On the basis of the AWS snow depth measurement, the snow pack decreased monotonically since 12 April (day 102). We therefore assume the period of rapid melting, with the largest grain sizes, to have taken place during, at least, the days of 16–20 April (julian days 106–110). 8 May, day 128, is at the end of the measurement period with the snow partly melted, revealing the ground, too. For this reason these data were not included cases presented here:

– 16 April: the thickness of the surface layer was 35 cm, whereas during the three earlier measurements, during the whole accumulation period, it had varied between 1–6 cm. Hence, a major change in the snow, with a deep metamorphosized and homogenized surface layer, took place on 16 April. The cloudiness was highly variable, yet a drop in the SL501 albedo after midday from 0.6 to 0.55 took place.

– 17 April: the snow depth was 50 cm at 06:00 UTC and 46 cm at 18:00 UTC, and the next day it was 45 cm. From this we can conclude that conditions similar to those on the clear sky day of 15 April, continued on days 106 and 107, despite the variable cloudiness. The SL501 albedo dropped from 0.6 to 0.55 after midday, and then recovered.

– 19 April: the results showed a similar diurnal decrease, too.

– 20 April: the cloudiness was highly variable, but the SL501 albedo decreased from approx. 0.6 to 0.55, as had been found in the clear sky cases earlier presented.

Hence, these cases with variable cloudiness confirm the clear sky SL501 cases during the melting season: an albedo slightly decreasing by 0.05 soon after midday, and then recovering after that.

4.3 Albedo before and after the period of diurnal change

The period of variable albedo began on the 15 and ended on 25 April. The cases before and after these dates show a flat albedo signal (Table 4). Prior to the diurnal albedo change, a flat signal of $A>0.6$ was detected. On the basis of the snow depth data, this was the period before the snow melt. Following the variable diurnal albedo, there was a diurnally-stable midday albedo, of $A<0.5$. These results would suggest that, after a drop in the albedo to a level of 0.5, the diurnal change in the albedo disappears, possibly signaling the end of some stage in the melting process.

4.4 Average daily UV albedo of melting snow

The midday erythemally weighted UV albedo ranged from 0.6 to 0.8 in the accumulation period and 0.5–0.7 during melting. The averaged daily UV albedo of snow for 56<SZA<60 degrees during the melt was a second-order polynomial as a function of snow height:

$$A = -6E - 05h_s^2 + 0.0114h_s + 0.1809 \quad (4)$$

where h_s is the snow height [cm]. This formulation, with r^2=0.86, is adjusted for the melt period only (data of julian days 100–120, Fig. 12).

4.5 Long term variation in SL501 UV albedo

Studying the time series of the continuous 1-minute albedo data, the SZA-dependency (U-dependency) of the differences between the responses of the sensors is evident (Fig. 13), as well as the occurrence of evening and night-time with no incoming irradiance. On closer inspection, when the midday data with SZA 56–60 degrees are used, the albedo signal is quite flat (e.g., Fig. 9).

An increase in albedo from 0.72 to 0.81 was detected on 11 April, possibly due to new snow (as evidenced from the AWS data for the same time), but further study is beyond the scope of the present work. More detailed study would require a detailed analysis focusing on the AWS rain data.

The SL501 midday albedo results for Arctic snow, can be divided into two groups: an albedo of 0.6–0.8 between 22 March–14 April (the accumulation period), and an albedo of 0.4–0.6 between 15 April–3 May (snow melt). These data were calibrated by using different empirical calibration factors for cases ZSA<60 degrees and for SZA>60. In studying the hypothetical effect of using the March coefficients (midday SZA>60) for May data, an error <6% was caluculated for 09:00–12:00 UTC. The use of different coefficients for the prevailing SZA conditions, as presented earlier, reduced this error in the long term data.

The albedo of snow was recorded to lower little by little as the snow melted (Fig. 13). The results would also suggest a possible SZA asymmetry in snow albedo (Table 4). The calibration procedures presented in this study have no effect on this finding, as it is a question of asymmetry according to noon: a SZA correction would produce SZA symmetric data if the prevailing conditions had not changed between noon and afternoon. Yet, to study this SZA asymmetry further would require knowledge on the factors affecting the SZA correction in the data (such as the ratio of direct-to-diffuse irradiance). After the beginning of May, the albedo results were characterized more by the amount of ground visible, than by the actual albedo properties of the snow.

5 Discussion

For a biological object, the UVB (erythemal) albedo is more important than information about the non-weighted UV albedo. Snow cover can increase erythemal irradiance by up to 60% compared to a snow-free case (e.g. Weatherhead et al., 2005). Here, use was made of erythemal UV albedo measurements by broadband SL501 radiometers with similar spectral responses, thus resulting in errors of less than 1% due to differences in the sensors (WMO, 1996). The calibration of the sensors was made before (in 2005 and 2006) and after (2007) the albedo measurements. Yet, in post-calibration measurements, it was observed that the responses had changed since the last spectral response measurements (2001 and 2004), which resulted in a U-dependency as a function of SZA in the albedo results. Basically, a drift in a sensor response might be nonlinear, either momentary or lasting, or even occasionally due to environmental conditions, such as the internal humidity of the sensor, temperature, total ozone, etc. Here, the problem was solved by empirical calibrations, as described, and by using only the data within a SZA-range of 56–60 degrees for the analysis. Thereafter, the error in the data due to the sensors can be determined on the basis of the post-calibration measurements (Fig. 4): within an SZA of 56–60 degrees, the SZA dependency, i.e., the *U*-shape due to difference in the spectral responses of the sensors, caused an error of less than 3%. Hence, on the basis of our experience, the spectral responses of SL501 sensors may change in time, and therefore the responses should be determined on a regular basis, preferably every year or every second year.

With these data, any possible dependency of albedo on solar zenith angle could only be compared for similar conditions of cloudiness and solar zenith angles. On the other hand, any albedo asymmetry close to midday could be reliably studied for SZA 56–60 degrees without introducing any additional uncertainty or error in the data due to a simplified SZA correction. For example, a SZA and ozone dependent calibration factor can not correct the data unless a proper spectral reponse funtion of the sensor is used. An example of such an error (an outlier) possibly introduced by a simple SZA correction in the albedo data is presented in Wuttke et al. (2006). On the basis of our experienc, an optimal 1-min SZA correction, for both the upward and downward broadband sensor independently, would require knowledge on the spectral and cosine response functions, temperature correction functions, and 1-min data on the radiation distribution, and the ratio of the diffuse-to-direct irradiance affected by clouds, ozone, aerosols, albedo and SZA.

An unexpected diurnal change in UV albedo, measured by the SL501 sensors, was detected during the melt period at SZA. The albedo decreased by approx. 0.05 soon after midday and then recovered to the same, or almost the same, level. If this diurnal change in albedo were to be due to any shadowing effects, rather than the properties of snow, the shadowing would be best seen under clear sky conditions, and not in cloudy situations. Here the diurnal change was evident for all states of sky. Furthermore, as the sensor facing upward was at the same height as the sensor measuring reflected radiation, any shadowing would necessarily be evident in the signal of the incoming irradiance at some time during the day (although not necessarily at the exactly same time as the downward sensor recorded the reflected radiation from the shadowed surface). The incoming irradiance did not indicate any such shadowing (Fig. 10). Also, when the site for the albedo measurements was chosen, the horizon towards the South was selected to be without shadowing. The frame supporting the sensors was also placed to the North to avoid shadows. Finally, the same diurnal decrease was evident in the data of the NILU-UV radiometer measuring the upwelling radiation close to the SL501.

Our findings on UV albedo of natural melting snow are in accordance with the speculations presented by Wiscombe and Warren (1980). They explain that snow albedo decrease due to liquid water content increase follows from the fact that liquid water replaces air between ice grains. As the refractive index of liquid water is close to ice

ACPD
8, 4155–4198, 2008

for wl<5000 nm, the replacement of of air by liquid water between ice grains could increase the effective grain size.

Furthermore, the results of diurnal variation of liquid water content at various snow layers (Fig. 14) by one of the authors (Martti Toikka) suggest that the temperature rise increases the liquid water content first on the surface layer of the snowpack. Then, as the temperature drops toward evening and night, the liquid water falls into deeper layers. Thus, liquid water on the snow surface, as well as the effective grain size due to liquid water, would increase only temporarily. When the temperatures drop, most of the liquid water is no longer in the surface layer, but in the layers below. Thus, these diurnal results of the liquid water in snow unticorrelate with the diurnal albedo results, offering empirical explanation to our albedo observations. Earlier, Kuusisto (1984) has also stated that the thick snow cover in northern Finland starts to release water from surface layers through percolation channels, while the deeper layers of the cover are still relatively unmetamorphosized. Opposite to this, in southern Finland, the thin snow cover is quickly metamorphosized all the way to the ground surface.

Similar results on diurnal decline of albedo were observed by Pirazzini in the Antarctica (personal communication, 2007), although not reported in the article by Pirazzini (2004). The same diurnal decline was observed by them in Arctic conditions, too (Pirazzini et al., 2006). In the Antarctic conditions, their observation for the diurnal decline was later in the afternoon. That can possibly be explained by the environmental conditions: in the Arctic the air temperature and the amount irradiance may be bigger than in the conditions of their study, thus the diurnal decrease in albedo due to liquid water on the surface of the snow would occur later afternoon in their study. Earlier, a minimum albedo has been detected by McGuffie and Henderson-Sellers (1985) in Canada, too. They suggested the albedo decrease to be due to snow grain metamorphism caused by heating of the surface.

According to Grenfell et al. (1994) the albedo of snow depends on its physical properties, and varies according to wavelength. Here, the spectral distribution of the snow albedo between UVA and UVB could be studied on the basis of a multiband filterradiometer facing downwards, and another upward MBFR radiometer nearby. These complementary results confirmed the findings of diurnal variability in the UVB albedo.

In the current study some indication on the asymmetric UV albedo was observed, too (Table 4). Also, McGuffie and Henderson-Sellers (1985) have reported of diurnal hysteresis of snow albedo, i.e. that the albedo is different for the same solar elevation angle at different times of day. They suggest the variation should be attributed to the diurnal deposition and evaporation of a hoar-frost coating on the snow surface. Pirazzini (2004) and Wuttke et al. (2006) have recorded more recently albedo results in the Antarctic sites with several instruments, which are opposite to the ones predeicted by theory. They both found a decline of albedo for increasing SZA. A possible levelling error was not consired the source for this observed diurnal cycle. Wuttke et al. (2006) speculate the reason for this opposite dependence of albedo on SZA to be rather due to changing snow conditions due to the steady solar insolation. Pirazzini et al. (2006) found similar results in Arctic conditions, too. Earlier, McKenzie et al. (1996) speculated that their UV albedo measurements of long grass (no snow) showing slightly higher albedos in the afternoon than in the morning might be due to either various error sources (e.g., leveling, angledependent reflections) or real changes in the surface, (e.g., morning dew evaporating or light dependent plant physiology). Hence, it is essential to understand and separate all the factors affecting the albedo results.

During the winter of 2006/2007, permanent snow fell in Sodankylä in the middle of October, but almost all the snow melted at the end of November (Kontu et al., 2007). Snow density during the winter months was determined to be between 0.18–0.21 g/cm^3 (Kontu et al., 2007). In addition to the temperature, other environmental factors, like rainfall, wind, humidity, cloudiness, as well as the properties of the snow and the ground under the snow, affect the process of snow melt. On the basis of AWS data on snow depth, 12 April was the date after which the snow amount only decreased from one day to the next. Hence, these data are in accordance with the start of diurnal albedo change detected from 15 April until 25 April. The prior and subsequent albedo was stable.

ACPD
8, 4155–4198, 2008

Title Page
Abstract Introduction
Conclusions References
Tables Figures
◂ ▸
Back Close
Full Screen / Esc
Printer-friendly Version
Interactive Discussion

The snow the grain size has been reported to vary generally by less than 50% on the topmost 10–20 cm of snow (Warren and Wiscombe, 1980). We found that in the Arctic conditions of Sodankylä, belonging to the global snow class of Taiga snow, the grain sizes of the top layer varied from 0.2–2.5 mm, containing variation by 125%. Thus the variability in grain size of the Arctic snow in Finland, and the maximum grain sizes were found to be extreme.

In a literature review presented in Warren & Wiscombe (1980) for their albedo model, it was found that some papers reported the albedo to increase with solar zenith angle. We found a slight increase at midday during the accumulation period with small grain size. This could possibly be physically explained by an increase in the specular component at midday compared to morning or afternoon. The midday increase would then be bigger the smaller the grain size.

In summary, our results suggest:

– a high and stable midday albedo for SZA 56–60 degrees during the snow accumulation period; albedo maximum in solar midday during accumulation period in clear sky and under variable cloudiness.

– possibly UV albedo asymmetric to solar zenith angles.

– a diurnal change of 0.05 in albedo during the melt period, in cases of clear sky and variable cloudiness with SZA 56–60 degrees.

– a little by little decrease in the general abedo level with the melting of snow.

– a stable lower albedo at the end of the melt period.

With the help of ancillary data on temperature, grain size and snow depth, these are explained by a high albedo induced by a small grain size during accumulation time, and a diurnal change in albedo by snow grain metamorphism caused by heating of the surface, melting some of the snow to liquid water, and the metamorphism ceasing by the end of the melt period. These findings were made possible only by continuous high temporal resolution measurements, and would not have been found by measurement campaigns.

The advantages of snow albedo measurements in Finland are the facts that here i) the snow cover melts every year, and ii) we have five out of the six global snow classes, only the alpine snow missing, and iii) the topography in Finland is flat, thus favorable to albedo studies, iv) clean snow can be found in the remote areas of the Finnish Lapland, vi) the snow grain size of the top layer varies greatly from small to extremely big grain sizes of 2.5 mm. In the future, we intend to continue ground-based UV albedo measurements under these conditions in Finland, and comparisons are planned to extend to Arctic-Antarctic, and between UV and broadband albedo. The liquid water content of the top layer of the snow from midday for two hours forward at intervals of approx. half an hour or even less should be measured in the melt period together with continuous high temporal resolution (1-min) albedo measurements and the ancillary data as presented here. Possibly some other parameters may also be included (including spectral albedo), or the temporal resolution of the parameters may be improved to study the diurnal decrease and the possible asymmetry. Wiscombe and Warren (1980) have reported that only a small number of albedo models had been put forward prior to their model, reflecting the lack of high-quality data against which to check such a model, and the fact that some of the data are contradictory. The albedo model introduced in their paper is in use in the commonly-applied radiative transfer (RT) model Libradtran (Mayer and Kylling, 2005), and currently highly referred to. Using our empirical data, a UV albedo parametrization for Arctic snow during accumulation and melt time periods could be elaborated, and the existing albedo models verified.

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Table 1. The midday SZA values [degrees] at Sodankylä (67.22° N, 26.37° E) in 2007.

Month	Feb		March		April		May
Date	1	15	1	15	1	15	1
SZA	85	80	75	70	63	58	52

ACPD
8, 4155–4198, 2008

UV albedo of arctic snow in spring

O. Meinander et al.

ACPD

8, 4155–4198, 2008

UV albedo of arctic snow in spring

O. Meinander et al.

Table 2. Empirical calibration procedures in 2007.

Date	Procedure
22.3.	Both sensors upward, 09:00–10:00 UTC, 67.2<SZA<68.6
2.5.	Both sensors upward (full cloudy, snow on the ground), 07:00–08:00 UTC 56.4<SZA<60.6
9.5.	Upward sensor turned downward and other turned up (snow almost melted)
11.5.	Both sensors turned upward
14.5.	Both sensors turned downward

ACPD

8, 4155–4198, 2008

UV albedo of arctic snow in spring

O. Meinander et al.

Title Page
Abstract | Introduction
Conclusions | References
Tables | Figures
Back | Close
Full Screen / Esc
Printer-friendly Version
Interactive Discussion

Table 3. The correlation coefficient (r) of snow grain size and other measured parameters. The number of cases was n=17 between the julian days of 64–132.

Parameter	r
Minimum ground temperature	0.37
Daily maximum temperature	0.79
Grain size and time (julian day)	0.55
Depth of total snow pack	0.47
Depth of the surface layer	0.22

Table 4. SL501 UV albedo of Arctic snow at Sodankylä for midday SZA<60 degrees in April 2007. The column of diurnal decline in albedo refers to temporary decline soon after midday (approx. duration), whereas asymmetric decline suggests a decline from morning towards afternoon.

Date	Snow conditions	Cloudiness	Albedo minimum and maximum	Albedo average	Diurnal decline
10.4.	Accumulation period	Almost clear	0.64–0.67 (stable, slightly highest at midday)	0.66	NO
11.4.	Accumulation period	Almost full cloudy	0.72–0.81 (stable, temporary high values due to snow fall)	0.74	NO
12.4	Snow melt period starts	Variable	0.66–0.70 (stable, slight indication of diurnal decline)	0.69	NO
13.4.	Melt	Variable	0.66–0.68	0.67	NO
14.4.	Melt	Variable	0.61–0.65 (slight indication of diurnal decline)	0.63	NO
15.4.	Melt	Almost clear	0.55–0.60	0.59	YES (60 min)
16.4.	Rapid melt	Variable	0.56–0.60	0.58	YES (20 min)
17.4.	Rapid melt	Variable	0.54–0.60	0.57	YES (60 min)
18.4.	Rapid melt	Almost clear	0.54–0.60	0.58	YES (100 min)
19.4.	Rapid melt	Variable	0.56–0.63	0.60	YES (60 min)
20.4.	Rapid melt	Variable	0.56–0.62	0.59	YES (60 min)
21.4.	Melt	Variable	0.58–0.61 (stable or slightly lowest at midday)	0.59	NO
22.4.	Melt	Almost clear	0.55–0.60	0.56	YES (100 min)
23.4.	Melt	Almost full cloudy	0.57–0.61	0.58	NO
24.4.	Melt	Variable	0.51–0.58 (suggesting asymmetric decline)	0.55	YES (80 min)
25.4.	Melt	Variable	0.50–0.54	0.52	YES (80 min)
26.4.	Melt	Variable	0.51–0.52	0.51	NO
27.4.	Melt	Variable	0.46–0.49 (suggesting asymmetric decline)	0.47	NO
28.4.	Melt	Variable	0.46–0.50 (stable or slightly lowest at midday)	0.47	NO
29.4.	Melt	Variable	0.44–0.5 (stable or slightly lowest at midday)	0.47	NO
30.4.	Melt	Variable	0.44–0.46	0.45	NO

ACPD
8, 4155–4198, 2008

UV albedo of arctic snow in spring

O. Meinander et al.

Table 5. The approximated decline in NILU-UV midday albedo in the UVA and UVB channels during the clear sky cases of the melt period in 2007.

Date	UVA albedo Diurnal decline	UVB albedo Diurnal decline
15.4.	0.04	0.06
18.4.	0.06	0.05
22.4.	0.04	0.04

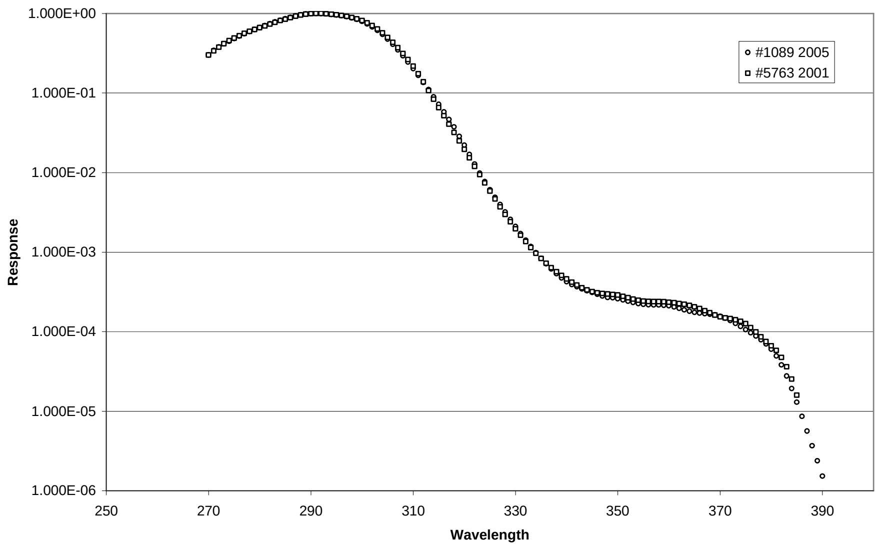

Fig. 1. Spectral responses of the SL501 sensors, in logaritmic scale showing the maximal differences.

ACPD
8, 4155–4198, 2008

Title Page
Abstract
Introduction
Conclusions
References
Tables
Figures
Back
Close
Full Screen / Esc
Printer-friendly Version
Interactive Discussion

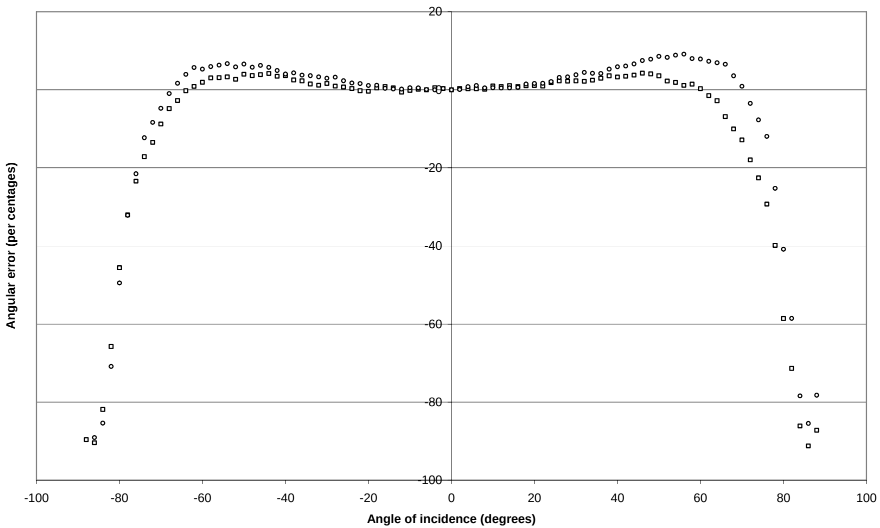

Fig. 2. Cosine responses of the SL501 sensors.

ACPD

8, 4155–4198, 2008

UV albedo of arctic snow in spring

O. Meinander et al.

Fig. 3. The holders designed for the albedo measurements.

ACPD
8, 4155–4198, 2008

UV albedo of arctic snow in spring

O. Meinander et al.

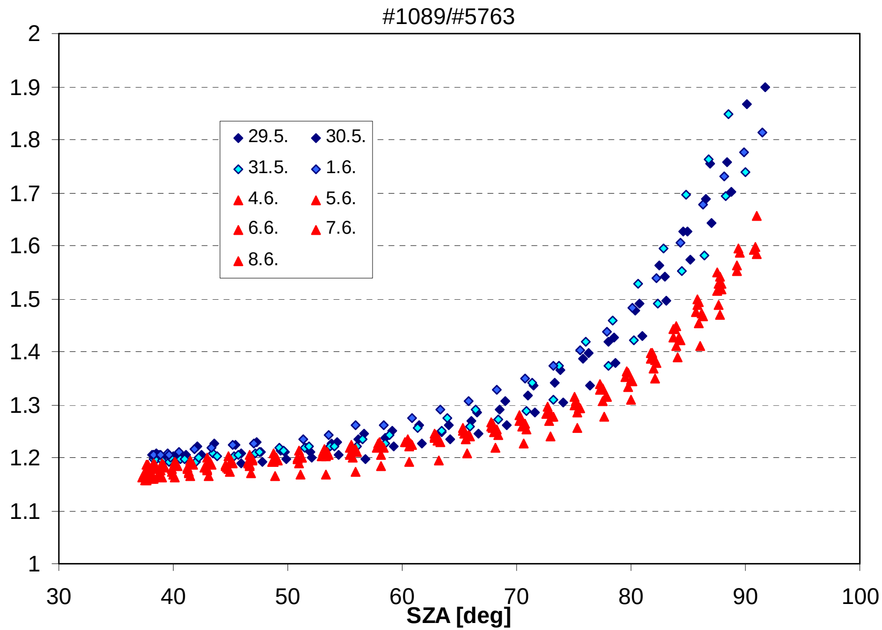

Fig. 4. Results of the post-calibration measurements of SL501s in 2007.

ACPD
8, 4155–4198, 2008

UV albedo of arctic snow in spring

O. Meinander et al.

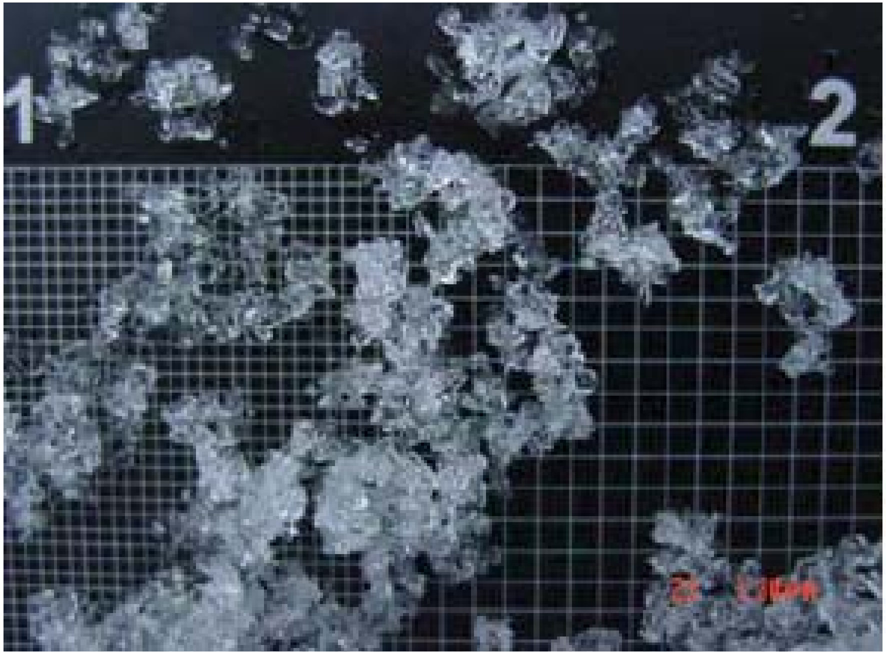

Fig. 5. The 1-mm grid used for the snow grain size measurements.

ACPD
8, 4155–4198, 2008

UV albedo of arctic snow in spring

O. Meinander et al.

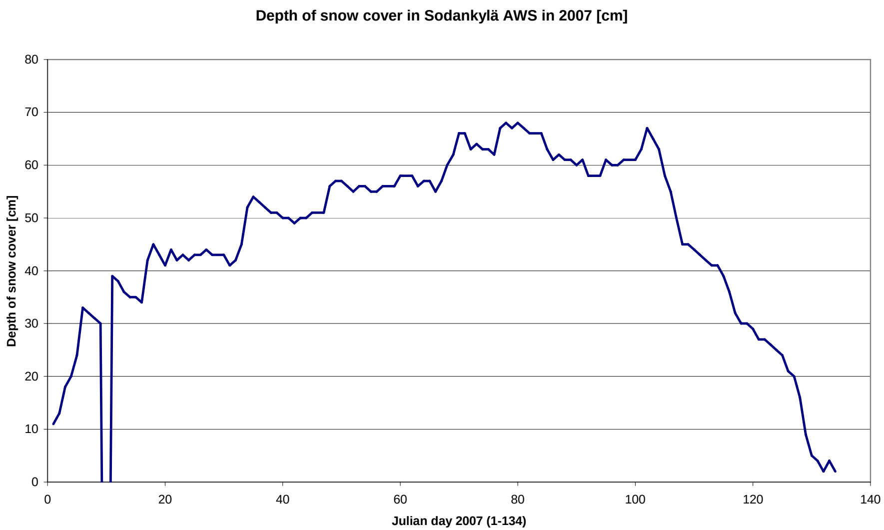

Fig. 6. Snow depth during winter and spring 2007.

ACPD

8, 4155–4198, 2008

UV albedo of arctic snow in spring

O. Meinander et al.

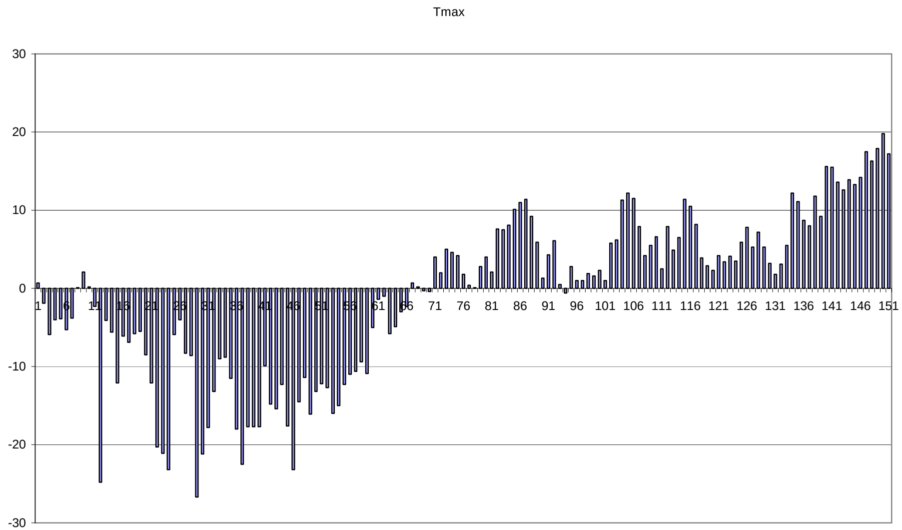

Fig. 7. Air temperature maximum showing the big jump, the "springthaw".

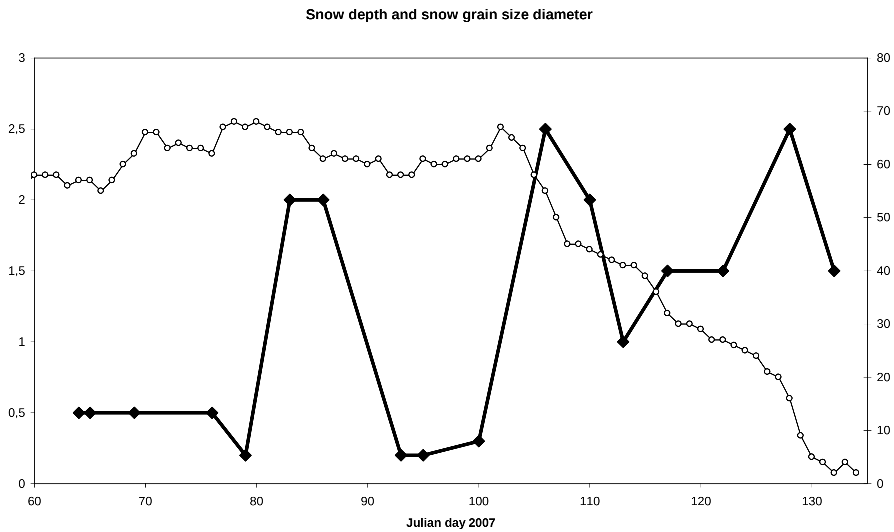

Fig. 8. Grain size and snow depth in spring.

ACPD

8, 4155–4198, 2008

UV albedo of arctic snow in spring

O. Meinander et al.

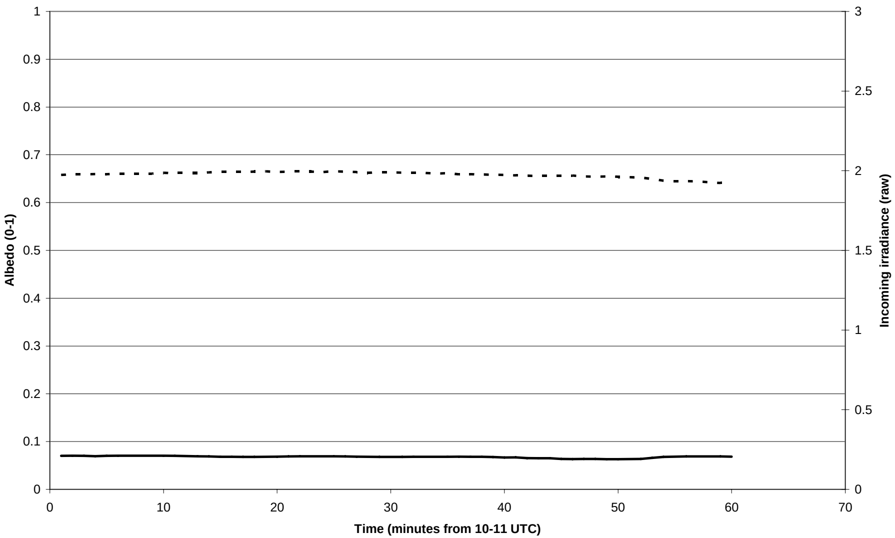

Fig. 9. Flat albedo signal of a clear sky case (10 April) during the accumulation period.

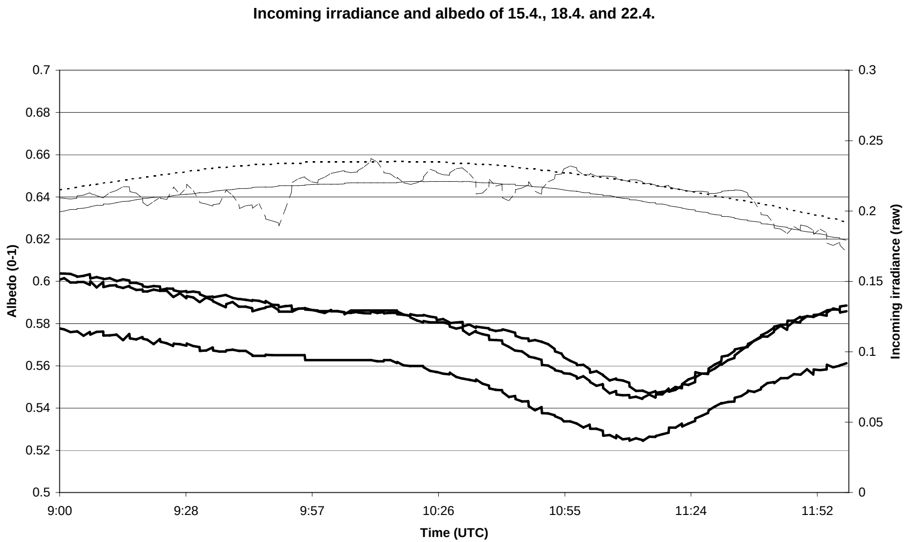

Fig. 10. Diurnal decrease under clear sky cases of 15, 18 and 22 April, marked with the solid lines. The incoming irradiances are marked with the dashed lines.

ACPD
8, 4155–4198, 2008

UV albedo of arctic snow in spring

O. Meinander et al.

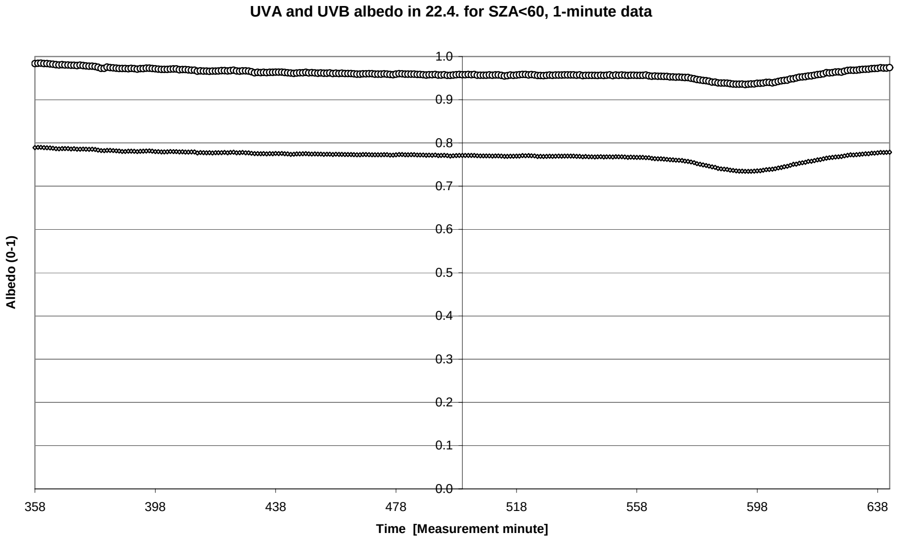

Fig. 11. Diurnal decrease in a NILU-UV measurement. UVA and UVB albedo in 22.4.2007 for cases with SZA<60 degrees; 1-min data measured with a NILU-UV radiometer. The midday is marked with the y-axis. Only relative changes in the UVA and UVB signals are to be considered.

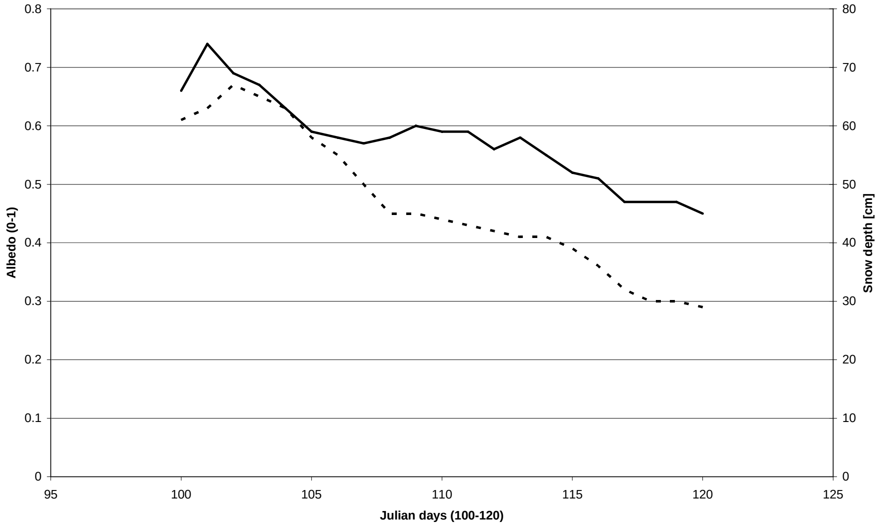

Fig. 12. Snow depth and albedo. Average of snow UV albedo for midday (solid line) and snow depth (dashed line) for each day during melt season from julian days 100 to 120 in spring 2007.

ACPD
8, 4155–4198, 2008

UV albedo of arctic snow in spring

O. Meinander et al.

ACPD

8, 4155–4198, 2008

UV albedo of arctic snow in spring

O. Meinander et al.

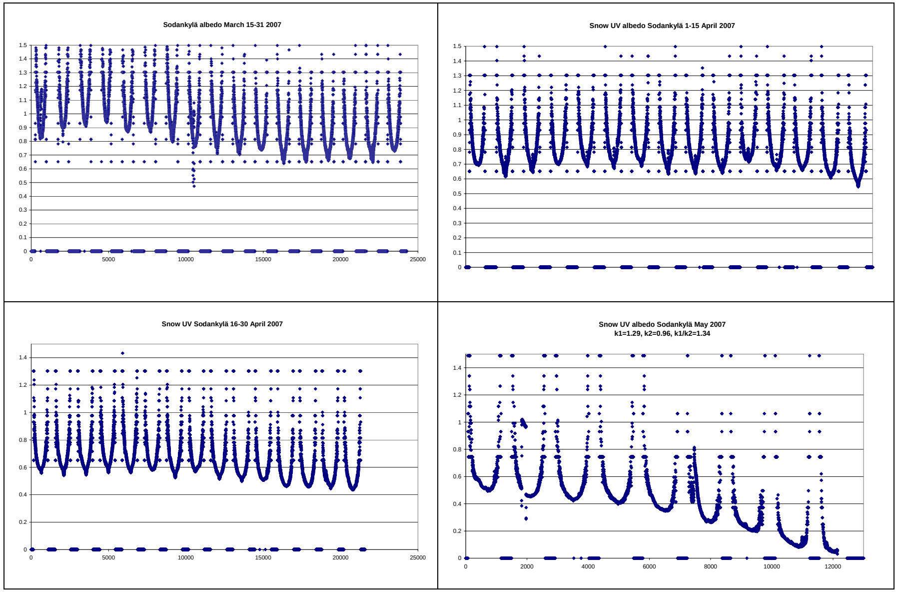

Fig. 13. Long-term albedo with the *U*-shape.

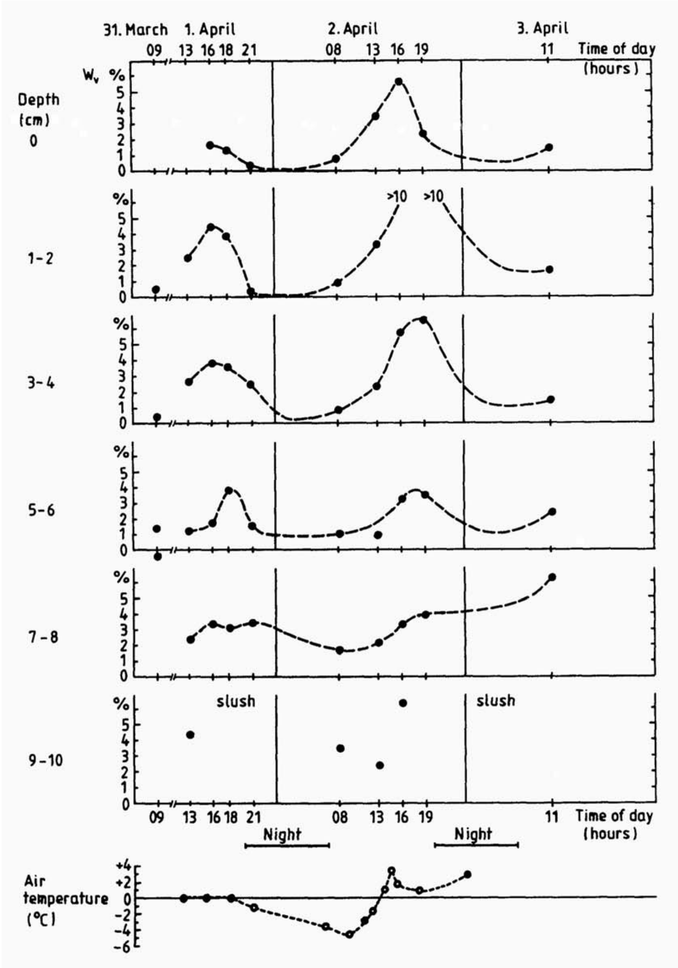

Fig. 14. Measured diurnal variation of liquid water content at different depth levels of snow pack.

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UV albedo of arctic snow in spring

O. Meinander et al.