Review 1

This paper presents a method to retrieve the specific surface area (SSA) of snow from spectral albedo measured at Dome-C, Antarctica with an automatic spectrometer "Autosolex". To measure the spectral radiant solar flux with Autosolex many correction procedures are discussed in which error analyses for spectral albedo and SSA were conducted. The diurnal cycle and seasonal variation of SSA are also discussed with possible reasons of these variations.

This manuscript is well-written and addresses the issues on the possible error of SSA in case of SSA retrieval from spectral albedo measurements on the Antarctic Plateau. The descriptions of the instrument are careful and the error analyses are appropriate. The observed SSA variation is informative and this kind of continuous measurement under such hard circumstance is important to monitor the climate change in Antarctica. I recommend this paper to be published after technical and minor revisions of the following issues.

We are grateful to the reviewer for his positive remarks and have made modifications in the manuscript according to his recommendations.

Specific comments:
p. 3, L19: Typo "cop" -> "cope"

done

p. 4, L2: "The spectral albedo of the surface is measured" should be preterit.

done

p. 4, L2-5: The terms of "section" appear here. Do these "sections" differ from sections mentioned at the end of 1. Introduction? The term "subsection" is better.

done

p.4, L22: Does the sentence "the second one is dedicated to the ultraviolet" explain the right-hand side spectrometer in Fig.2? It is written as "Spectrometer 700-900 nm" in the figure.

Both were erroneous, ultraviolet is infrared and the range in Fig.2 is 700-1100nm. This is corrected.

p.5, L16-17: "This arises because both the light collector materials and the sky depend on scattering intensity which usually decreases with longer wavelengths." This sentence is a bit difficult to understand particular for "sky". What is it about the sky depend on scattering intensity?

We change to "This arises because scattering by the light collector materials and the atmosphere strongly decreases with longer wavelengths."

p.7, L10-13: "To estimate the stray light, we assume it affects all the pixels equally ..." Stray light could sometimes cause large error for this kind of instrument. Please show

the fraction of contribution from the stray light to "dark and stray light" here or at Fig. 7.

We have separated the two corrections in the plot below (similar to Fig.7) and found the curve after the stray light correction (violet) can not be distinguished from the curve (orange) after the dark correction which means the stray light is very small. Instead of adding a figure with overlapping curves, we first show and address the first step alone (dark correction) and then add a short paragraph on the stray light correction "The next step (stray light correction) is not shown in Figure 7 because it has a negligible effect."
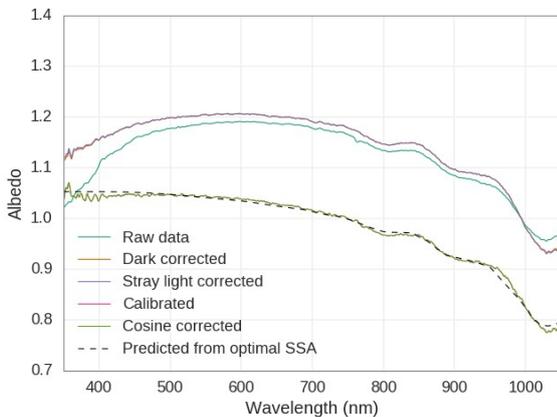
p.8, L17: Equation (6) differs from that of Grenfell et al. (1994) in which the correction for diffuse component is applied. I believe this correction would improve the accuracy of the measured diffuse component.

The equation (6) is indeed formally different from that given in Grenfell et al. (1994) because our absolute calibration is done with diffuse radiation and the imperfect cosine. It means the diffuse component is already calibrated after the "absolute calibration step". Despite this formal difference, considering two steps "absolute calibration"+"cosine correction" together is equivalent to Grenfell et al. We have added an explanation: "The (cosine) correction is applied only to the direct component as the diffuse component has been already calibrated at the absolute calibration step"

p.11, L1: "the small peaks due to damaged pixels like at 862 nm and 1069 nm." The small peak can be seen around 862 nm but not at 1069 nm in Fig. 7.

The x-axis scale of Figure 7 has been extended to show the peak around 1069 nm.

p.11 and Figure 7: The graph names "graph b)" (L11), "graphs (c)" (L17), "graph b" (L2), and "graph (c)" (L24) are used in the document. However, in Fig. 7 "a)" and "b)" are shown above the top panels, and "(c)" is not indicated. The "graph b)" in L13 seems to discuss on the middle panels.

We have corrected this: letters are used to designate the columns and "top, middle and bottom" are now used for the graphs within a column.

p.12, L13: The left term of Equation (11) is better "alfa(wavelength)" (as a function of wavelength).

Yes, we have also added the angle dependence.

p.14, L19: Equation (16) is not shown.

This blank equation was due to a latex typo,

p.15, L25: "is shown in gray in the background" It is written as "blue dots" in the caption of Fig. 13.

"gray" changed to "blue".

p.15, L27: "the geophysical features to (Libois et al., 2015)." Some terms may disappear before "(Libois et al., 2015)"?

The parenthesis have been removed and the sentence changed to :
" The interpretation of the geophysical features is addressed in Libois et al. 2015."

Review 2

General comments: The paper is a breakthrough in the observation of long-term surface properties of snow and spectrally resolved albedo over snow. In my view, the paper was not easy to read, as the technical details specific to albedometers and the snow measurements appear Methods, Results and Discussion - a structure which seems to me not very friendly for the reader. A structure where the instrument details, then the snow measurements, and then a discussion would be easier to follow. I suggest to re-organize the paper into two main sections:

- Theory, background, instrumental details

- Observations, interpretation and comparison.

We have reorganized the paper to address first the most generic and theoretical points and second the instrument details. After the introduction, a section addresses the theory with 1) a description of the SSA retrieval algorithm, 2) the vertical representativeness of the retrieved SSA and 3) the formulation to investigate the impact of instrument artifacts on SSA uncertainty. The next section is focused on the instrument description, processing description and illustration/uncertainties, and stability. The next section presents the time-series of SSA and the last one provides conclusions and future works. This follows the two points proposed by the reviewer except that the "Theory" and "instrumental details" are split in two sections because they are indeed different and represent 90% of the article.

I also found the title not very relevant, a title as "Development and long-term calibration of a new albedometer measuring vertically integrated snow surface SSA " would be much more clearly describe what you can expect in this paper.

We propose the following title which include the term "time-series" which is the main novelty of the instrument.

"Development and calibration of an automatic spectral albedometer to estimate near-surface snow SSA time-series."

A second important point concerns the "vertically integrated SSA" (my term). As is clearly shown in Fig. 15, the SSA determined via a spectrally resolved albedometer is in a complex way vertically integrated. As a snowpack is mostly horizontally layered, and often with very strong changes in surface SSA within millimeters (surface hoar, glaze, windcrusts, new snow,...) this point should be

considered already in section 2, not in the very end of the paper.

This has been moved in Section 2 as suggested.

I also found it a bit disturbing that no independent direct measurements of the SSA via stereology or micro-CT were performed. This would make the interpretation significantly more plausible and less speculative, especially concerning the role of the very surface.

We share this feeling but we do not have such independent measurements at Dome C and we believe it is very difficult to collect near surface snow and apply X-ray tomography (or stereology) techniques to obtain results with a sufficient accuracy in the typical Dome C conditions of high SSA and high vertical gradient in the topmost layer.

We have participated in several intercomparison campaigns over the past years where many measurements techniques have been combined. These intercomparisons took place in Alpine conditions where SSA is lower, snow is more homogeneous and the vertical gradient is probably much smaller than at Dome C. Only in such ideal conditions and after a few attempts, we were able to claim that a reasonable check (or "cross-validation") of the optically-based SSA measurements has been achieved (not published yet). We hope such attempts could be done at Dome C in the future.

As a very significant paper, I would like to suggest to the authors that they reconsider the structure.

As a final point, the peer-review will not be able to validate the method (see p20, L8), I can only check if the methods and procedures are reasonable!

Yes. The sentence is removed.

- Further points p 1 L10 The sentence "The comparison of the retrieved SSA with independent measurements made with an optical device operating at 1310 nm confirms the presence of a sharp and recurrent vertical gradient of SSA in the uppermost centimeter at Dome C, which challenges the assessment of the absolute accuracy from independent measurements." seems to me overstated. If the gradient is "sharp" can not be determined by the methods used: either near-infrared photography in a profile would be necessary, or micro-tomography. What is obvious and correct that the upper

snow layers are at times of higher SSA than the averaged SSA observed with Autosolex. This is not surprising concerning the calculated penetration depth! The same is valid for p17 L34ff.

We remove this sentence from the abstract because the paper does not address the issue of the gradient with sufficient details.

p 3 L 27 ff references for the "manual devices"?

done

p18 L11 ff: Which spectral irradiance was used (W m-2 nm-1) for the calculation of the averaged penetration depth?

Albedo calculations from which we deduce the penetration depth are independent of the irradiance.

p22 L19 reference to Libous 2013 seems incomplete

p38 Fig. 14 The two datasets are from different depths (as shown in Fig. 15), so they are not really comparable. These data only show that the surface has almost always a higher SSA than the deeper "sampling" Autosolex.

This has been addressed with the reogranisation of the article. The vertical representativeness is first described in Section 2. Then the comparison with Asssap in Section 4 only concludes that the SSA is always higher at the surface than slightly deeper.

~~Estimation~~ Development and calibration of ~~superficial snow specific surface area from~~ an automatic spectral ~~albedo~~ albedometer to estimate near-surface snow SSA time-series~~at Dome C, Antarctica~~

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Abstract. Spectral albedo of the snow surface in the visible/near-infrared range has been measured for 3 years by an automatic spectral radiometer installed at Dome C (75°S, 123°E) in Antarctica in order to retrieve the specific surface area (SSA) of superficial snow. This study focuses on the uncertainties of the SSA retrieval due to ~~the instrument deficiencies and issues related to data processing~~ instrumental and data processing limitations. We find that when the solar zenith angle is high, the main source of ~~error~~ uncertainties is the imperfect angular response of the light collectors. This imperfection introduces a small spurious wavelength-dependent trend in the albedo spectra which greatly affects the SSA retrieval. By modeling this effect, we show that for typical snow and illumination conditions encountered at Dome C, retrieving SSA with an accuracy better than 15% (our target), requires the ~~slope of this trend not to exceed 2%~~ difference of response between 400 and 1100 nm not to exceed 2%. Such a small ~~slope~~ different can be achieved only by i) a careful design of the collectors, ii) an ~~ad hoc~~ *ad hoc* correction of the spectra using the actual measured angular response of the collectors, and iii) for solar zenith angles less than 75°. The ~~comparison of the retrieved SSA with independent measurements made with an optical device operating at 1310 confirms the presence of a sharp and recurrent vertical gradient of SSA in the uppermost centimeter at Dome C, which challenges the assessment of the absolute accuracy from independent measurements. Nevertheless, with~~ 3-year long time-series of retrieved SSA features a three-fold ~~variations of SSA during the summer seasons, we conclude that the retrieved SSA is accurate enough to provide a detailed picture of the October-March evolution of the surface snow~~ decrease every summer which is significantly larger than the estimated uncertainties. This highlights the high dynamics of near-surface SSA at Dome C.

1 Introduction

The summer surface energy budget on the ice sheets is largely controlled by the absorption of solar energy by snow (Van As et al., 2005; Ettema et al., 2010). The latter is determined by the downward irradiance reaching the surface and by snow

albedo. With solar irradiance of several hundreds of $W\,m^{-2}$ during daylight, a persistent change of 1% of the albedo represents an energy comparable with the ~~external~~ globally averaged radiative forcing caused by CO_2 concentration increase since pre-industrial time ($1.82\,W\,m^{-2}$, Myhre et al., 2009). While planetary albedo change is not expected to exceed a ~~tenth of this value~~ 0.1% (Myhre et al., 2009), local changes can be much larger owing to the dependence of snow albedo on multiple factors includ-

5 ing snow grain size and shape, surface roughness, snow depth, and the amount of light-absorbing impurities such as black carbon, dust and biological pigments ~~(Warren and Wiscombe, 1980; Warren et al., 1998; Aoki et al., 2000; Dumont et al., 2010; Zhuravleva a~~ These factors vary in space and time depending on the atmospheric conditions and are controlled by numerous processes giving rise to complex snow-albedo feedback loops between the snow cover and the atmosphere (Curry et al., 1995; Qu and Hall, 2007; Picard et al., 2012). In Antarctica, snow grain size is the main factor controlling the albedo along with the illumination

10 conditions (solar zenith angle and cloud cover) and surface roughness while the impurity content is usually low (Warren et al., 2006). Investigating the feedback loops thus requires long-term and accurate observations of albedo and snow properties.

Nevertheless, albedo measurements in Antarctica are scarce and subject to artifacts. Only a few Baseline Surface Radiation Network (BSRN) stations and some automatic weather stations deployed by the Institute of Marine and Atmospheric research (Utrecht University) provide time series of broadband albedo measured using upward and downward looking pyranometers.

15 ~~These~~ In addition, these sensors are subject to many artifacts such as imperfect response at high solar zenith angle, leveling and frost, which are difficult to correct ~~(van den Broeke et al., 2004; Nicolaus et al., 2010; ?)~~(van den Broeke et al., 2004; Bogren et al., 2016). Retrieving broadband albedo from remote sensing is not a straightforward alternative. It requires atmospheric correction, spectra extrapolation from a limited number of spectral bands, and angular extrapolation of the bi-conical measurements to get hemispherical surface reflectance (Stroeve et al., 2006). The latter is challenging because the bi-directional reflectance of snow

20 is much more sensitive to grain shape (Dumont et al., 2010) and surface roughness (Kuchiki et al., 2011) than hemispherical reflectance (i.e. albedo), thus introducing extra ~~uncertain~~ unknown variables that need to be estimated somehow. Moreover, satellite data can not be used under cloudy conditions or when the sun zenith angle is too high (Wang and Zender, 2010; Schaaf et al., 2011). ~~Even with a few overpasses of the satellites per day~~ Despite many satellite overpasses per day in polar regions, the time-series are discontinuous and snowfalls, which are periods of great albedo change, are systematically missed.

25 Measuring snow grain size is difficult as well. Estimating the maximal extent of the dominant crystals using hand-lens or binocular is the most widely used technique to measure grain size (Fierz et al., 2009; Aoki et al., 2000; Pirazzini et al., 2015). However, not only ~~it is~~ is it known to be imprecise and operator-dependent, but also it can not be automated for unmanned monitoring of grain size evolution. Recent advances in field measurement of the specific surface area (the surface area of the air-ice interface per unit of mass of snow) – which is equivalent to the optical radius commonly used in remote sensing

30 (Grenfell and Warren, 1999; Nolin and Dozier, 2000; Jin et al., 2008) – offer an attractive alternative. Examples of instruments and methods include contact spectroscopy (Painter et al., 2007), near-infrared photography (Matzl and Schneebeli, 2006), DUFISSS and IceCUBE (Gallet et al., 2009) and POSSSUM and ASSSAP (Arnaud et al., 2011). These techniques being based on optical measurements of the snow reflectance in the near or short wave infrared, it is arguable whether they provide data more related to albedo or a proper geometrical metric of the snow micro-structure. Several theoretical studies have indeed

35 shown the significant influence of the snow crystal shape on the albedo (Macke and Mishchenko, 1996; Neshyba et al., 2003;

Kokhanovsky and Zege, 2004; Picard et al., 2009) and on the e-folding depth (Libois et al., 2013) but no clear experimental evidence of this effect has been given yet for snow on the ground (Gallet et al., 2009; Libois et al., 2014b). In fact, these optical techniques have been shown to be accurate within 15% when compared with independent measurements using methane adsorption or micro-computed tomography techniques (Matzl and Schneebeli, 2006; Domine et al., 2006; Gallet et al., 2009).

5 This accuracy is sufficient to achieve a 1% accuracy of broadband albedo (Gardner and Sharp, 2010) which is considered to be adequate for climate study (?)(Bogren et al., 2016). Intensive campaigns of SSA measurements have provided new insight of the snow metamorphism in the Alps and in Antarctica over seasons and have allowed refined evaluation of detailed snow model predictions (Picard et al., 2012; Morin et al., 2013; Roy et al., 2013; Libois et al., 2015). However, these techniques still require an operator which is often a limitation, particularly in the Antarctic. Optical satellite remote sensing has been used

10 to retrieve optical radius (e.g. Nolin and Dozier, 2000; Painter et al., 2009; Negi et al., 2011; Mary et al., 2013) in particular in Antarctica (Scambos et al., 2007; Jin et al., 2008). Mary et al. (2013) and Tanikawa et al. (2015) give an overview of the numerous existing algorithms. Obtaining grain size on a large spatial extent and for many years is of great interest, but attaining an accuracy similar to that of field techniques requires to overcome several additional defects that are specific to space-borne sensors and are in fact very similar to those faced for albedo retrieval. Microwave remote sensing is not subject to these issues

15 but as far as albedo study is concerned, only the highest available frequencies could be able to provide superficial grain size (Picard et al., 2012). However, the technique proposed by Picard et al. (2012) is specific in many ways to the conditions of the Antarctic Plateau and requires additional assumptions about the density and temperature of superficial snow compared to the optic domain.Wuttke et al. (2006)

Ground-based observation of spectral albedo has several advantages compared to broadband measurement: it can provide
20 a finer understanding of the albedo variation and can be used to estimate SSA using similar principles to those used in remote sensing and by the optical techniques mentioned above. It is also easier to assess the data quality or develop corrections by exploiting the spectral signature. Many spectral albedo measurements have been conducted in Antarctica (e.g. Grenfell et al., 1994; Hudson et al., 2006; Hudson and Warren, 2007; Marks et al., 2015; Pirazzini et al., 2015) but they were limited to short term campaigns and the temporal evolution was not the priority. Long-term monitoring of spectral

25 albedo requires fully automated automatic instruments with a good robustness to cop cope with the Antarctic conditions (Wuttke et al., 2006; Salzano et al., 2016). With the advent of cheaper and compact spectrometers free of mechanical parts (Kantzas et al., 2009; Nicolaus et al., 2010), operating such instruments in Antarctic conditions is becoming easier.

The objective of the present study is to describe the automatic spectral radiometer that we developed and installed at Dome C (75°S, 123°E) and the data processing developed to estimate time-series of spectral albedo and SSA from the raw

30 irradiance measurements. We particularly focus on the uncertainties of the estimated albedo and SSA. For this, we analyze the improvement of the quality through the successive processing steps. We discuss the results in the light of a SSA target uncertainty of 15%, as claimed to be reached by manual devices (— near-infrared photography , DUFISSS, POSSSUM) (Matzl and Schneebeli, 2006), DUFISSS (Gallet et al., 2009), POSSSUM (Arnaud et al., 2011)) – and considering it is difficult to perform better than those man-operated tools. Theoretical modeling of the imperfections of the instrument are used to

35 help in this analysis and also to provide quantitative recommendations for the design of future spectral radiometers. At last

Simulations are also used to explore the representativeness of the ~~estimated SSAis explored by comparing the estimates with independent measurements and by using radiative transfer simulations~~retrieved SSA. The geophysical interpretation of the 3-year long SSA time-series obtained with our instrument and its comparison to snow model simulations are addressed in Libois et al. (2015).

5   Section ~~3 describes the instrument, the processing and method to retrieve SSA. Results are presented in section 4and discussion and conclusion is~~ 2 describes the retrieval of SSA from spectral albedo and uncertainties induced by instrumental artifacts. Section 3 presents the instrument deployed at Dome C and the data processing to obtain calibrated albedo. The 3-year long time-series of retrieved SSA is presented and analyzed in Section 4. Discussion and conclusion are provided in the last section.

10 2   ~~Materials and method~~Theory

~~The spectral albedo of the surface is measured~~ In this section, we present the algorithm to retrieve SSA from observed spectral albedo and evaluate the uncertainties resulting from working hypothesis and instrumental artifacts.

2.1   Retrieval of specific surface area from spectral albedo

Snow SSA at the surface is obtained by fitting a theoretical albedo model with SSA as the main unknown to observed
15 albedo spectra over some range of wavelengths. To keep the number of unknowns moderate in order to prevent unstable optimization, the following assumptions are made: 1) The snowpack is horizontally and vertically homogeneous which means only one SSA value is retrieved. 2) The surface is flat. Roughness which tends to weaken the solar zenith angular response of the snow is neglected (Warren et al., 1998). 3) The surface and the sensor are perfectly horizontal. 4) Snow phase function and single scattering albedo are implicitly described by the asymmetry factor, absorption enhancement parameter, and SSA
20 (Kokhanovsky and Zege, 2004). The values of asymmetry factor and absorption enhancement parameter are assumed constant and are taken from Libois et al. (2014b). 5) Snow is clean or contains sufficiently small quantities of impurity not to impact the albedo in the considered wavelength range. This is usually the case in Antarctica (Grenfell et al., 1994).

    In these conditions, the analytical asymptotic radiative transfer (ART, Kokhanovsky and Zege, 2004; Negi et al., 2011) is valid and gives the directional and diffuse hemispheric reflectances α^{diff} and α^{dir} as a function of SSA:

25
$$\alpha^{\mathrm{diff}}(\lambda) = \exp\left(-4\sqrt{\frac{2B\gamma(\lambda)}{3\rho_{\mathrm{ice}}\mathrm{SSA}(1-g)}}\right) \tag{1}$$

$$\alpha^{\mathrm{dir}}(\lambda,\theta) = \exp\left(-\frac{12}{7}(1+2\cos\theta)\sqrt{\frac{2B\gamma(\lambda)}{3\rho_{\mathrm{ice}}\mathrm{SSA}(1-g)}}\right), \tag{2}$$

where θ is the solar zenith angle, $\rho_{\mathrm{ice}} = 917\,\mathrm{kg\,m^{-3}}$ is the ice density at $0\,°\mathrm{C}$, and $\gamma(\lambda)$ is the absorption coefficient of ice, taken from Warren and Brandt (2008). $B = 1.6$ and $g = 0.85$ are respectively the absorption enhancement parameter and the asymmetry factor values suggested by Libois et al. (2014b).

Considering that observed albedo is composed of a direct and diffuse component, the model reads:

$$\alpha^{\text{1-param}}(\lambda,\theta) = \left[r^{\text{diff}}(\lambda,\theta)\alpha^{\text{diff}}(\lambda) + \left(1 - r^{\text{diff}}(\lambda,\theta)\right)\alpha^{\text{dir}}(\lambda) \right], \tag{3}$$

where $r^{\text{diff}}(\lambda,\theta)$ is the ratio of diffuse over direct irradiance and is supposed to be known. This simple model is called hereinafter 1-parameter model because SSA is the only unknown. However, it is not the most suitable because albedo measurements are subject to various artifacts such as: variations of the illumination or poor cross-calibration between the downwelling and upwelling measurements, slopes of the surface, etc, sometimes even resulting in values larger than 1 (Pirazzini et al., 2015). To cope with artifacts that do not depend on wavelength, a free wavelength-independent scaling parameter A is introduced:

$$\alpha^{\text{2-param}}(\lambda,\theta) = A\left[r^{\text{diff}}(\lambda,\theta)\alpha^{\text{diff}}(\lambda) + \left(1 - r^{\text{diff}}(\lambda,\theta)\right)\alpha^{\text{dir}}(\lambda) \right], \tag{4}$$

This 2-parameter model is more flexible but to avoid over-fitting, we discard spectra that would result in A out of the range 0.9–1.1.

The fit is performed with a non-linear least square method (provided by the Python scipy.optimize.leastsq function) to minimize the squared difference between the model and observed spectrum.

The choice of wavelength range to perform the fit depends on the available observation. In our case, spectrometers with Silicium sensor are used so only data at wavelengths shorter than 1050 nm present a sufficient signal/noise ratio. In addition, we restrict the fit to wavelengths longer than 700 nm because the sensitivity of the albedo to SSA is stronger in the near-infrared than in the visible, and impurities are known to affect visible wavelengths (France et al., 2011).

2.2 Vertical representativeness of retrieved SSA values

Assuming the snowpack is vertically homogeneous and retrieving a single bulk SSA stabilizes the fit. However, this implies to interpret the retrieved value with caution. There are indeed some evidences of significant vertical gradient in the topmost centimeter in Antarctica. Grenfell et al. (1994) found at South Pole that adding a thin layer of very small snow grains (SSA larger than $100\,\mathrm{m^2\,kg^{-1}}$) on the surface in their radiative transfer calculations was required to match their albedo measurements. They attributed this to the deposition of snow particles having been shrunk by sublimation during blowing snow events. Carmagnola et al. (2013) at Summit in Greenland use spectral data in the visible, near and short wave infrared and found that tuning the surface layer SSA helped to improve the agreement with the observations but did not explore values as high as Grenfell et al. (1994) did. In contrast, Gallet et al. (2011) measured SSA vertical profiles at Dome C ~~(75S, 123E)with a home-made instrument presented in the first section. The second section details the construction and characterization of the light collectors (or fore optics) that are critical to achieve high quality measurements. The third section presents the data processing steps to derive spectral albedo from raw data~~ and obtained radiative transfer calculations in agreement with Hudson et al. (2006) observations without adding such a layer. Using snowpack numerical modeling, Libois et al. (2015) obtained SSA profiles at Dome C with such a significant gradient near the surface. This is particularly remarkable as their simulations do not consider some processes susceptible to bring small grains on the surface such as shrinking of air-borne particles by sublimation or clear-sky precipitation (a.k.a diamond dust).

Because of this gradient, it is important to know over which thickness the retrieved SSA is representative. Unfortunately, there is no simple answer because the penetration depth of radiation in snow depends not only on the wavelength but also on the vertical profile of snow properties, including SSA itself and density, which are both unknown. Furthermore, the estimated SSA value is not a simple average over a given thickness but is a weighted average with a kernel function decreasing with depth. This function has an exponential form for homogeneous snowpack but is more complex otherwise.

Only multi-layer radiative transfer calculations can give a precise estimation of the sensitivity-depth relationship when SSA and density profiles are known. To evaluate this relationship in the simple case of an homogeneous snowpack, we proceed as follows: the two-stream radiative transfer model TARTES (Libois et al., 2013) is run considering a semi-infinite medium with fixed SSA and a density of $270 \, \mathrm{kg \, m^{-3}}$ (typical mean surface conditions at Dome C, Libois et al. (2014a)). A layer with variable thickness h is added on top of it, with the same properties, excepted that a small perturbation δSSA is added to the SSA. The relative contribution of the uppermost layer to the albedo is then defined as the quantity $(\alpha(h) - \alpha(0))/(\alpha(\infty) - \alpha(0))$ where $\alpha(h)$ is the albedo calculated by TARTES for the thickness h and averaged over the wavelength range used by Autosolexs (700-1050 nm). The actual value of δSSA does not change the final result owing to the normalization, as long as it is small (e.g. $+1 \, \mathrm{m^2 \, kg^{-1}}$). The contribution is shown in Figure **??** (plain line) as a function of h for different SSA values. Results show that the uppermost 10 mm snow layer contributes to nearly 60% of the albedo for SSA of $20 \, \mathrm{m^2 \, kg^{-1}}$ and ~~the last section addresses the retrieval of the surface specific area of superficial snow~~ 85% for higher SSA of $100 \, \mathrm{m^2 \, kg^{-1}}$. Conversely, the layer contributing to 80% of the signal is 18, 12 and 8mm thick for SSA of 20, 50 and 100 respectively. We can conclude that 1 cm is a reasonable approximation for Dome C conditions which gives a more quantitative meaning to the term "near-surface" used throughout this article.

For comparison, we also run similar simulations at 1310 nm, a wavelength commonly used by device to measure SSA (Gallet et al., 2009; Arnaud et al., 2011) (dashed line in Figure **??**). In this case, most of the signal comes from the uppermost 5 mm in any common conditions. Note that the penetration depth is inversely proportional to density so that the thickness values presented here should be multiplied by about two in case of fresh snow or surface hoar.

2.3 SSA uncertainty due to instrumental artifacts

Even when the assumptions listed in Section 2.1 are met, the SSA retrieval can be impacted by observation imperfections, such as dark current of the radiometer, leveling and angular response of the light collector, calibration, etc. Here, such impacts are investigated using a simple theoretical framework. Most of these effects can be represented in a first approximation as either a bias in the measured irradiances (hereinafter called offset) or a small spurious wavelength-dependent trend in the spectra (hereinafter called chromatic aberration). We investigate theoretically these two cases successively in the next subsections and apply the result to real observations in Section 3. In both cases, to quantify the impact on the estimation of the SSA, we perform simple and idealized numerical experiments as follows: we first compute the perfect albedo spectrum with the model described in Equation 18 for a given SSA (called true SSA) and perturb it to mimic the considered artifact. Then, we retrieve the SSA using the algorithm described in Section 2.1 and deduce the error as the difference between the estimated and true SSA.

2.3.1 Offset

We consider the true reflected and incident radiance spectra $S^{\mathrm{ref}}(\lambda)$ and $S^{\mathrm{inc}}(\lambda)$ are affected by a constant bias d. The measured albedo writes:

$$\alpha(\lambda) = \frac{S^{\mathrm{ref}}(\lambda) + d}{S^{\mathrm{inc}}(\lambda) + d}. \tag{5}$$

5 Dividing by the incident radiance makes explicit the true albedo α^{true}:

$$\alpha(\lambda) = \frac{\alpha^{\mathrm{true}} + \frac{d}{S^{\mathrm{inc}}(\lambda)}}{1 + \frac{d}{S^{\mathrm{inc}}(\lambda)}}. \tag{6}$$

For the incident spectrum $S^{\mathrm{inc}}(\lambda)$, we choose a Gaussian shape looking like the observations (Section 3.3):

$$S^{\mathrm{inc}}(\lambda) = S_{\mathrm{mode}} \exp\left[-\left(\frac{\lambda - 680\mathrm{nm}}{270\mathrm{nm}} \right)^2 \right]. \tag{7}$$

The amplitude of the spectrum S_{mode} is usually of the order of the resolution of the Digital Analog Converter of the radiometer
10 (2^{16} in our case) if the integration time is optimal. We define $d' = d/S_{\mathrm{mode}}$, the relative contribution of the offset with respect to the spectrum amplitude. We then estimate the SSA from the perturbed $\alpha(\lambda)$ spectrum and calculate the difference with respect to the true SSA.

Figure **??** shows the SSA error as a function of the relative offset d' ranging from 0 (perfect instrument) to 2% for several SSA values. The result is nearly independent of the true SSA. To meet the 15% accuracy criteria, the offset needs to remain
15 under d'=1%. For a spectrometer, this offset can generally be estimated using dark pixels (pixels receiving any light so that $S(\lambda) = 0$). Since their signal-to-noise ratio is usually much better than 1:100, the offset correction should not in most case be an obstacle to achieve the targeted accuracy.

2.3.2 Chromatic aberration

To investigate the impact of chromatic aberration, we consider the true albedo α^{true} is multiplied by a linear function of the
20 wavelength:

$$\alpha(\lambda, \theta) = \left(1 - b\frac{\lambda - \lambda_0}{\lambda_1 - \lambda_0} \right) \alpha^{\mathrm{true}}(\lambda, \theta) \tag{8}$$

where b controls the slope of the wavelength dependence. The limits λ_0=400 nm and λ_1=1100 nm are chosen so that the instrument is assumed perfect at 400 nm in the blue and b represents the relative response difference between the upward and downward looking channels at 1100 nm. Figure **??** illustrates the results with true SSA of 50 $\mathrm{m^2\,kg^{-1}}$ and $b = 0.05$ (actual
25 values of b are given in this section and in Section 3). We remind that the SSA estimation only uses data between 700–1050 nm. The retrieved SSA is 38.2 $\mathrm{m^2\,kg^{-1}}$ with the 2-parameter model and 27.6 $\mathrm{m^2\,kg^{-1}}$ with the 1-parameter one. This corresponds

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to relative errors of 24% and 45% respectively. The 2-parameter model is clearly better which can be explained by the fact that it is insensitive to the scaling of the albedo and depends mostly on the spectal variations of albedo within the range used for the optimization. However, this error is larger than the 15% target accuracy in SSA.

An interesting feature of the 2-parameter model is that the albedo predicted with the optimal SSA and A values (orange curve) exhibits an offset in the visible not only with respect to the true albedo (black curve) but also to the observed one (green curve). For instance, in Figure **??**, the difference is 0.024 on average over the range 400–550 nm. This difference (hereinafter referred to as residuals) is a consequence of the wavelength-dependent perturbation and can be exploited to assess the amplitude of this perturbation in real data because it does not require knowledge of the true albedo. It means that acquisitions subject to residual chromatic aberration feature a large difference in the visible range with respect to the spectrum predicted using the 2-parameter model. To explore how this difference and the relative error on retrieved SSA depend on actual SSA, Figure **??**a shows the relative error and the mean difference over the range 400–550 nm as a function of the chromaticity parameter b for several SSA. The error varies almost linearly with b and the slope increases with SSA. Hence, to attain the target accuracy of 15%, the chromatic parameter should not exceed about 0.04 for SSA=20 $\mathrm{m^2\,kg^{-1}}$ and remain under 0.02 for SSA=100 $\mathrm{m^2\,kg^{-1}}$. Figure **??**b shows the nearly univocal relationship between b and the mean residual. This suggests that mean residual values under 0.01 indicate a chromaticity b better than 0.02 and thus an SSA accuracy better than 15%, thereby providing a fast and simple spectra quality check criteria.

We implemented this criteria in our algorithm as follows: 1) predict the albedo spectrum with the 2-parameter model using the optimal SSA and A parameters obtained by fitting the model to the observed spectrum in the range 700–1050 nm. 2) calculate the mean difference in the range 400–550 nm and 3) reject the spectrum if this difference is larger than 0.01. This filter works at Dome C where snow is clean, but would not work in the presence of light absorbing impurities because the latter cause a decrease of the albedo in the blue-green domain (Warren, 1982) which would result in systematic rejection by the filter.

2.4 ~~Automatic measurements of spectral albedo~~

Several artifacts results in chromatric aberration. Leveling of the collector is one of them. Tilt angle δ of the collector with respect to the horizontal indeed tends to affect albedo by a scaling $\eta(\lambda, \theta)$ that can be written according Bogren et al. (2016) and Grenfell et al. (1994):

$$\eta(\lambda, \theta) \approx \left(1 - r^{\mathrm{diff}}(\lambda, \theta)\right) \left(\frac{\cos(\theta - \delta)}{\cos(\theta)} - 1\right) \tag{9}$$

where the diffuse term error has been neglected as suggested by Bogren et al. (2016) findings and tilt in the direction of the sun (worse case) is considered. Assuming the errors are small enough to be treated as perturbations, b can be deduced as follows:

$$b \approx \eta(\lambda_1, \theta) - \eta(\lambda_0, \theta) \tag{10}$$

$$\approx \left(r^{\mathrm{diff}}(\lambda_1, \theta) - r^{\mathrm{diff}}(\lambda_0, \theta)\right) \left(\frac{\cos(\theta - \delta)}{\cos(\theta)} - 1\right) \tag{11}$$

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For a tilt angle of 1° and under typical conditions for Dome C, we obtain $b = 0.01$ for SZA=53° (as the noon acquisition studied in Section 3.2) and $b = 0.05$ for SZA=77° (as the evening acquisition) which is weak in the former case but is too high in the latter one to reach the 15% target accuracy. These results highlight the crucial role of the leveling of the instrument. Other defects resulting in chromatic aberration are addressed in Section 3.2 on practical examples.

3 Measurements of spectral albedo

A specific instrument has been developed to measure the spectral albedo of the surface at Dome C. The following subsection describes the instrument, details on the construction and characterization of the light collectors (or fore optics) and data processing steps.

3.1 Autosolexs instrument

We designed and assembled an instrument composed of several commercial and home-made components to automatically measure spectral albedo. The main specifications driving the design ~~was~~ were the robustness to work in ~~the~~ Dome C harsh environment for several years and the ability to acquire not only albedo (incident and reflected radiation) but also the radiation within the snow at several levels (not used in this study), thus requiring a spectrometer with many inputs. The instrument is named Autosolexs (Automatic SOLar Extinction in Snow) because of the latter application.

Figure **??** shows pictures of the visible above-ground part of the system which comprises two heads for albedo measurements at about 2 m above the surface. ~~This~~ Due to snow accumulation, this height has decreased at an averaged rate of about 10 cm per year since the installation. Each head is equipped with two light collectors looking upward and downward. The collected ~~radiations are~~ radiation is transmitted by 6-m long fiber optics (core diameter of 400 μm) to the rest of the device that is buried under the snow. Neither electronic nor fragile parts are above-ground.

The instrument scheme is depicted in Fig. **??**. The fibers coming from the heads are directly connected to the inputs of a 16-to-1 optical switch (FiberSwitch® mol 1×16 19" 2 HU by Leoni). Using a mechanic-optical module driven by a digital interface, it links in a few milliseconds one of the 16 inputs to the output with a reproducibility better than 0.03 dB according to the manufacturer. This corresponds to about 0.7% uncertainty for the irradiance which in the worse case translates to 1.4% for the albedo. The output of the switch is connected to a splitter 1-to-2 that transmits radiation to two spectrometers (~~MAYA~~ MAYA2000 PRO, Ocean Optics). One covers the visible ~~near infrared and~~ and near-infrared and the second one is dedicated to the ~~ultraviolet~~near-infrared. Only the former is used in this study. It gives signal above the noise level in the range 350 - 1050 nm with an effective spectral resolution of 3 nm. One of the inputs of the optical switch is obstructed so it measures dark conditions that are used to correct the offset present in the spectrometer acquisitions. The instrument container is thermalized at +10±2°C which is necessary for ~~the good~~ nominal operation and stability of the spectrometers.

An embedded computer controls the optical switch, the spectrometers and schedules the acquisition. Every 12 min, a complete sequence of measurements is performed in the following order: head 1 (incident and then reflected channels), head 2 (incident and then reflected channels), buried fibers, heads 1 and 2 again, and at last dark. Each measurement takes 10-20

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seconds. An automatic camera is connected to the computer to take one picture per sequence like those shown in Fig. **??**. This provides qualitative but very useful information on the state of the snow surface and the presence of frost on the light collectors. The system is connected to the power supply of the Concordia station – the power consumption is typically around 50 W – and to the network to send data in near-real time.

3.2 Design and characterization of the light collectors

The light collectors are an important component. Albedo – or more rigorously bi-hemispherical reflectance (Schaepman-Strub et al., 2006) – is the ratio of the upwelling and downwelling ~~fluxes~~spectral irradiances, and the flux is the energy (per unit of surface, of time and of wavelength) crossing an horizontal surface. To measure flux, the collector must have a so-called cosine response so that the energy coming with an oblique direction is weighted by the cosine of the zenith angle (Grenfell et al., 1994; Morrow et al., 2000).

In principle, a flat surface which transmits a constant proportion of the incident beam whatever its direction acts as a perfect collector. For this reason, collectors are often made of a flat disk of highly diffusing material like sintered teflon grains. Nevertheless, teflon grains, like snow grains, have a strong forward scattering behavior so that flat collectors are more reflective at grazing angles than for the normal incidence. As a result, they transmit less at grazing angles than ~~what is~~ required for a perfect cosine response. ~~When~~ In clear sky conditions, when the sun is low on the horizon~~and in clear sky conditions, the downwelling irradiance has a strong and grazing direct component and the~~, the upward-looking collector ~~thus tends~~ tends thereby to under-estimate the incoming flux. In contrast, the irradiance reflected ~~back~~ by the snow is ~~always~~ more diffuse and the down-looking collector is ~~therefore almost not affected by this issue~~almost unaffected by the collector quality. This differential behavior usually results in an over-estimation of the albedo which increases with the solar zenith angle and depends on the direct/diffuse ratio of the sky.

This problem is particularly serious because it strongly depends on the wavelength. This arises because ~~both~~ scattering by the light collector materials and the ~~sky depend on scattering intensity which usually~~ atmosphere strongly decreases with longer wavelengths. Hence, the collector angular response changes with the wavelength from a near-ideal cosine response in the blue part of the solar spectrum to a nadir peaked response at longer wavelengths (Grenfell et al., 1994). This effect is enhanced by the decrease of the direct/diffuse component ratio of the downwelling irradiance under clear-sky conditions, with the wavelength. ~~The albedo is therefore subject to chromatic aberration (wavelength dependent response)~~ This results in chromatic aberration that depends on the sun elevation, cloudiness, and other factors affecting the spectral and angular characteristics of the illumination.

To address this difficulty, we ~~took~~ pursued a two-fold strategy consisting of building and optimizing our own collectors ~~,~~and applying an *ad hoc* correction to remove the residual imperfections~~using the actual~~. The latter requires the angular response of the collectors to be measured.

Our collectors are made of a bumped surface cap (Figure **??**) following the design proposed by Bernhard and Seckmeyer (1997) to compensate for the reduced transmission at grazing angles. However, a side-effect of this design is that the light transmitted through the cap is not emerging symmetrically with respect to the collector axis (shown in Figure **??**) but is

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more intense in the azimuthal direction of the sun. This effect increases with the solar zenith angle. This has a very negative side-effect because the spectrometer does not collect all the light transmitted by the collector but only a part coming from a rectangular area that corresponds to the optical image of the slit used in input of the spectrometer to limit the aperture before the grating (the fiber does not blur enough this image to weaken this effect). The combination of the eccentered illumination on the cap in the azimuthal direction of the sun and the collection of the light from another unknown azimuthal direction results in a general azimuthal response of the system. This response is about 5-10% in amplitude, which is unacceptable for measuring albedo with 1% accuracy. To solve this critical problem, we inserted a second flat disc between the fiber and the bumped cap whose role is to homogenize the light coming from the cap. Moreover, by adding more scattering material, this improves the overall diffusion of the collector and ~~thus~~ thereby the quality of the angular response. We could not have used thicker caps or disks to further improve the scattering because the overall transmissivity of the collector is already of the order of 0.001, which imposes ~~relatively long integration time ,~~ integration time of the order of 1 second. Longer integration time would result in increasingly ~~uncorrectable dark current and noise~~ reduced signal-to-noise ratios. The optimization of the shape of the collector cap was done empirically by a series of trials and errors.

To measure the collector response, we set up an optical bench with a collimated light beam illuminating the collector mounted on a motorized rotation stage. The collector is connected to the spectrometer through a flexible fiber optic. Both the spectrometer and the stage are controlled by computer allowing rapid and reproducible measurements of the irradiance $C(\lambda,\theta)$ as a function of the incident angle θ. ~~The angular response of the collector $C(\lambda,\theta)$ is calculated as the ratio between the measured irradiance and the ideal cosine response (Section 3.3) so that a value of 1 throughout the spectrum indicates an ideal collector.~~ A typical response is depicted in Figures **??** and **??** as a function of the angle and wavelength respectively. It has been normalized by the ideal cosine and considering exact response at 0° following (Grenfell et al., 1994). For angles lower than 70°, the deviation remains within ±6% of the ideal response which is comparable to Grenfell et al. (1994) but twice better than Carmagnola et al. (2013) who used a flat collector. The deviation between 70° and 80° degrades significantly which suggests ~~difficulties to get accurate~~ uncorrectable measurements in this range of SZA, however it remains within ±15% while Grenfell et al. (1994) show twice larger errors. The values under 500 nm are extrapolated (Fig. **??**) because our light source is too weak in this wavelength range to get signal above noise level. The response is relatively independent of the wavelength for SZA=50° and becomes more and more ~~dependent~~ wavelength-dependent at larger angles. The measured responses are used for the *ad hoc* correction described in the next section.

3.3 Processing of raw measurements into albedo

The calibration of the raw spectra acquired by the spectrometer requires several processing steps: 1) dark and stray light corrections (abbreviated 'dc' and 'sl' hereinafter), 2) integration time scaling ('it'), 3) cross and absolute calibrations ('cal'), 4) cosine correction and cloud detection ('cc'), 5) albedo calculation. These steps are described in details in the following:

Dark and stray light corrections. The raw spectra acquired by the spectrometer are in numerical counts ranging from 0 to about 65000 for each pixel of the CMOS sensor (16-bit Analog-to-Digital converter). As any photo-sensitive sensor, the spectrometer is subject to dark current and other effects that result in non-zero outputs even when no light enters in the device.

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It is important to remove this offset ~~as highlighted in Section 2.3.1~~ especially in weak illumination conditions or near the borders of the bandwidth where the sensor sensitivity is ~~weak~~the weakest. The dark current of our spectrometers depends on ~~the~~ temperature and integration ~~times and is most likely~~ time and may be subject to aging of the spectrometer and other environmental parameters. Because it is not possible to control all these parameters, dark spectra are measured ~~for~~in every sequence (every 12 min) using the dedicated channel. ~~Acquisitions~~ Since we found a linear relationship between the offset and the integration time, acquisitions at two extreme integration times (T1=13 ms the minimum of the spectrometer and T2=1000 ms the typical time for our measurements) are necessary and sufficient to correct all the spectra acquired during the same sequence at any integration time. ~~We indeed found that the offset is a linear function of the integration time.~~ Hence, to correct a spectrum $S^T(\lambda)$ acquired with a given integration time T, we first estimate the offset at the actual acquisition time by linear interpolation of the two extreme dark spectra ($D^{T1}(\lambda)$ and $D^{T2}(\lambda)$) and then subtract this estimate from the measured spectrum as follows:

$$S_{dc}^T(\lambda) = S^T(\lambda) - \left[\frac{D^{T2}(\lambda) - D^{T1}(\lambda, T_1)}{T_2 - T_1}(T - T_1) + D^{T1}(\lambda, T_1) \right] \tag{12}$$

where $S_{dc}^T(\lambda)$ is the dark current corrected spectrum.

Stray light is another typical artifact of spectrometers (Kantzas et al., 2009). It comes from the small fraction of the beam that instead of passing through the grating to be dispersed is deviated and after internal ~~reflexions~~ reflections ends up on the CMOS sensor adding signal to the pixels. Second order dispersion from the grating is another cause of stray light but our spectrometer is mounted with a second-order filter that prevents this effect. For spectrometers of the same brand as ours, Kantzas et al. (2009) found stray light amounting to between 2 and 12% of the maximum intensity. This effect results in an offset which depends on the total intensity entering the spectrometer. To estimate the stray light, we assume it affects all the pixels equally (i.e. independent of the wavelengths) and estimate it using the average of the counts in the range ~~200 – 260~~200–260 nm where no radiation is supposed to come from the grating, because the fiber optics cut-off is around 290 nm and because the weak sensitivity of CMOS sensor in this range. The spectrum corrected from stray light writes:

$$S_{sl+dc}^T(\lambda) = S_{dc}^T(\lambda) - \left\langle S_{dc}^T(\lambda) \right\rangle_{\lambda=200-260nm} \tag{13}$$

Integration time scaling. Considering that the CMOS sensor accumulates charge with the same efficiency whatever the integration time, the spectra in counts are divided by the integration time T to obtain a signal proportional to the incoming radiance:

$$S_{it+sl+dc}(\lambda) = S_{sl+dc}^T(\lambda)/T \tag{14}$$

Calibrations. The calibration of each channel to obtain absolute irradiance is performed in two steps, first the cross-calibration of the channels to each others and second the absolute calibration of the set. We came to this two-step strategy because it was difficult to design an experiment to perform the absolute calibration directly with a reproducibility better than 1% as required for our target accuracy. Since the albedo only depends on the relative calibration between the channels, we put more effort on the cross-calibration than on the absolute one.

The experiment to obtain data for the cross-calibration consists in measuring successively the upward and downward channels under the same illumination conditions. To do this, before the installation, we first acquired spectra with the head in normal position, with special care at the horizontal leveling, and performed a second acquisition immediately after flipping the head up side down. The elapsed time between these operations was about 30 s and the experiment was conducted under clear-sky conditions which ensured constant illumination. For each channel, the spectrum taken when looking downward $S_{it+sl+dc}^{\text{cross}}(\lambda)$ is used to calibrate any spectra acquired after installation for this particular channel (whatever the direction of looking once installed) as follows:

$$S_{ccal+it+sl+dc}(\lambda) = \frac{S_{it+sl+dc}(\lambda)}{S_{it+sl+dc}^{\text{cross}}(\lambda)} \tag{15}$$

where the $ccal$ subscript refers to the cross-calibration step. The downward direction is chosen instead of the upward because the illumination coming from the snow is mostly diffuse which ~~provides a reduced~~ limits the sensitivity to the head leveling. The experiment was done just before the installation independently for both heads so that it is useful to cross-calibrate both channels of the same head (which is relevant for the albedo) but is inadequate for inter-header calibration.

The absolute calibration aims at inter-calibrating the heads and provide absolute irradiance (in $\text{W m}^{-2} \text{nm}^{-1}$ for instance). The experiment to collect reference spectra was done after the installation of Autosolexs and uses a commercial spectral radiometer (HR1024, Spectra Vista Corporation) as the reference. ~~Incident spectra~~ Spectra have been acquired simultaneously with the two instruments and for the two heads which allows ~~to normalize~~ any spectra acquired later from any channel to be normalized as follows:

$$S_{cal+it+sl+dc}(\lambda) = S_{ccal+it+sl+dc}(\lambda) \frac{\text{SVC}(\lambda)}{S_{ccal+it+sl+dc}^{\text{abs}}(\lambda)} \tag{16}$$

where $\text{SVC}(\lambda)$ and $S_{ccal+it+sl+dc}^{\text{abs}}$ are the spectra taken with the reference spectral radiometer and Autosolexs respectively. We do not expect an accuracy better than 20% for this step owing to the limited intrinsic accuracy of the reference and the many possible artifacts of the experiment. However this does not ~~affect the albedo uncertainty~~ impact the albedo as explained before.

Collector angular response correction. The imperfect angular response of the collector mainly affects the incident channel under clear-sky conditions. Following Grenfell et al. (1994), we use the collector response measured in the laboratory $C_{it+sl+dc}(\lambda, \theta)$ at different angles θ and corrected for the dark current, stray light and integration time beforehand. Azimuthal symmetry was checked and appeared to be very good. The correction is applied ~~to the incident channel only:~~ only to the direct component as the diffuse component has already been calibrated at the absolute calibration step, hence only the incident channel is concerned:

$$S_{cc+cal+it+sl+dc}(\lambda) =$$
$$S_{cal+it+sl+dc}(\lambda) \left[r^{\text{diff}}(\lambda, \theta) + \left(1 - r^{\text{diff}}(\lambda, \theta)\right) \frac{2\cos\theta}{C_{it+sl+dc}(\lambda, \theta)} \int_0^{\pi/2} C_{it+sl+dc}(\lambda, \theta') \sin\theta' d\theta' \right] \tag{17}$$

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where $r^{\text{diff}}(\lambda,\theta)$ is the diffuse fraction. The first term in the brackets on the right-hand side represents the diffuse contribution which is assumed isotropic. The anisotropy of the sky diffuse component is neglected which is second order as long as the correction is small. The second term is the direct contribution that is normalized by the collector angular response ($C_{it+sl+dc}(\lambda,\theta)$) relatively to the ideal cosine ($\cos\theta$).

The $r^{\text{diff}}(\lambda,\theta)$ quantity is estimated using the SBDART atmospheric model (Ricchiazzi et al., 1998) considering clear-sky conditions. The atmosphere for summer Arctic is used with the altitude adapted to Dome C. The column water vapor totals 0.4 mm (Tremblin et al., 2011) and ozone is set to 300 DU. Aerosol optical depth at 440 nm is set to 0.02. This calculation does not account for the temporal variations of the atmospheric composition. It could be improved in the future by exploiting the measured incident spectra to infer and better represent the actual atmospheric conditions.

For the present study, we ~~simply~~ filtered out the data acquired in cloudy conditions. To detect such conditions, the incident spectrum is compared to a spectrum chosen as a reference when the sky was known to be clear (2013-11-15 03:36 UTC close to the local zenith). Each spectrum is then scaled beforehand by the cosine of the solar zenith angle to account for the geometrical difference. When the scaled spectrum is more than 10% lower than the reference spectrum in the range 405 – 550nm, we consider the sun as being partially obstructed and the data are discarded.

Eventually the spectral albedo is calculated by taking the ratio between the upwelling and downwelling spectra. The shadow of the mat and the heads on the snow surface is very small compared to when an operator takes manual measurements (Carmagnola et al., 2013) and is neglected.

3.4 ~~Specific surface area (SSA) retrieval from spectral albedo~~

~~Snow SSA is obtained by fitting the observed albedo spectra to a theoretical model with SSA as the main unknown. Even if the number of observables per spectrum is very large (up to 2000 different wavelengths), the errors and the auto-correlation between these observations require to keep the number of unknown low to obtain a stable fit. The following assumptions are made: 1) The snowpack is horizontally and vertically homogeneous which means only one SSA value is retrieved. 2) The surface is flat. Roughness which tends to smooth the solar zenith angular response of the snow Warren et al. (1998) is neglected. 3) The surface and the sensor are horizontal. 4) Snow phase function and single scattering albedo are implicitly described by asymmetry factor, absorption enhancement parameter, and SSA. The values of anisotropy factor and absorption enhancement parameter are taken from Libois et al. (2014b). 5) Snow is clean or contains sufficiently small quantities of impurity not to impact the albedo in the considered wavelength range (Warren et al., 2006).~~

~~In these conditions and in the visible and near-infrared wavelengths, the analytical asymptotic radiative transfer (ART, Kokhanovsky and~~ ~~valid and gives the directional and diffuse hemispheric reflectances α^{diff} and α^{dir} as a function of SSA:~~

$$\alpha^{\text{diff}}(\lambda) = \exp\left(-4\sqrt{\frac{2B\gamma_\lambda}{3\rho_{\text{ice}}\text{SSA}(1-g)}}\right)$$

$$\alpha^{\text{dir}}(\lambda,\theta) = \exp\left(-\frac{12}{7}(1+2\cos\theta)\sqrt{\frac{2B\gamma(\lambda)}{3\rho_{\text{ice}}\text{SSA}(1-g)}}\right),$$

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where θ is the solar zenith angle, $\rho_{\text{ice}} = 917$ is the ice density at 0, and $\gamma(\lambda)$ is the absorption coefficient of ice, taken from Warren and Brandt (2008). $B = 1.6$ and $g = 0.85$ are respectively the absorption enhancement and the asymmetry factor values suggested by Libois et al. (2014b).

Considering both direct and diffuse radiations as in Equation (17), the model reads:

$$\alpha^{\text{1-param}}(\lambda, \theta) = \left[r^{\text{diff}}(\lambda, \theta) \alpha^{\text{diff}}(\lambda) + \left(1 - r^{\text{diff}}(\lambda, \theta)\right) \alpha^{\text{dir}}(\lambda) \right],$$

where $r^{\text{diff}}(\lambda, \theta)$ is calculated as for the cosine correction. This model is called hereinafter 1-parameter model because only the SSA is unknown. However, it is not the most suitable because the measurements are often affected by errors which result in a slight wavelength-independent scaling of the albedo. Typical examples include the variations of the illumination between the two successive acquisitions of downwelling and upwelling irradiance, the effect of sloping surface and the imperfection of the cross-calibration. To cope with these errors without negatively impacting the SSA estimation, another model including a free scaling parameter A is also used:

$$\alpha^{\text{2-param}}(\lambda, \theta) = A \left[r^{\text{diff}}(\lambda, \theta) \alpha^{\text{diff}}(\lambda) + \left(1 - r^{\text{diff}}(\lambda, \theta)\right) \alpha^{\text{dir}}(\lambda) \right],$$

Nevertheless, considering that a value of A significantly different from 1 reveals a serious issue with the observations, we reject any fit that gives A outside the range $0.9 - 1.1$.

The fit is performed with a non-linear least square method (provided by the Python scipy.optimize.leastsq function) to minimize the squared difference between the model and observed spectrum. Smoothing is done beforehand to Before applying the SSA retrieval algorithm, albedo spectra are smoothed to reduce the noise using a first order Butterworth low-pass provided by the Python scipy.signal.butter function with cut-off of 0.1.

Given the stronger sensitivity of the albedo to SSA in the near-infrared with respect to the visible, only the data for wavelengths in the range $700 - 1050$ are used for the fit. This choice is a compromise. Extending the range towards the shorter wavelengths does not bring useful information for the SSA retrieval and may add artifacts, while narrowing it would reduce the number of observations and result in a greater sensitivity to noise. Extending the range in the longer wavelengths is not possible due to the weak sensitivity of the spectrometer.

4 Results

3.1 StabilityIllustration of the processing steps

Before deriving albedo and SSA over long periods of time, it is important to check the overall stability of the instrument. The seasonal variations of downwelling irradiance are plotted in Figure ?? for the two heads and for the three summer seasons. The irradiance is integrated between 400 and 1000 and averaged between $10h - 14h$ (local time). These data include all weather conditions. Overall, there is a good year to year agreement. For instance, the irradiance observed by the head 1 during the 30 days after the solstice (period in common for the three seasons) has increased by +0.5 % and +4.2% the second and third years with respect to the first one. These values become -0.5%, and +3.0% respectively for the head 2, which is relatively

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small for an unattended instrument and considering that these values include the inter-annual variability of cloudiness. Only the third season shows a significant increase (around 4-5 %) which is likely due to a degradation of the leveling. The prevailing southerlies winds may have tilted the mast northwards, resulting in the observed increase of the noon irradiances. The influence of tilt angle on the estimation of the SSA is not critical because small geometrical imperfections are partially compensated when using the 2-parameter model that includes the scaling factor A. This influence is evaluated in Section 2.3.2 but we do not know the actual tilt of the heads because the inclinometers mounted on the system appeared to be deficient at low temperature.

3.2 Measurement processing

The different steps to process the raw measurements and obtain albedo are The effect of the processing steps is illustrated in this section for two acquisitions taken in two different contrasting solar conditions: a first one at local noon the 10 January 2013 with the sun relatively high in the sky for Dome C (SZA = 53°) and a second one at 8 pm the same day (SZA = 77°). Figure Figures **??** (a) and (b) illustrate the contribution of each step (curves with different colors) for these acquisitions.

The two first steps (dark and stray light corrections) remove first step (dark correction) is shown in Figure **??** (orange). It removes the overall offset that is clearly visible in the raw data (Figure **??**) at wavelengths where the CMOS sensor has little sensitivity (under 350 nm and around 1100 nm). This correction has a positive side-effect as it also reduces the small peaks due to damaged pixels like at 862 and 1069 nm. With the aging of the spectrometer, the number of such pixels tend tended to increase (data not shown), but they remain isolated and the correction is efficient.

The albedo is affected by this correction especially near the margins of the spectrum where the sensitivity of the CMOS sensor fades. Because both the incident and reflected spectra are equally affected by the same offset value, the albedo calculated with un-corrected data tends to 1 near the margins. Since clean snow albedo is expected to tend to about $0.98 - 0.99$ in the green and blue wavelengths, un-corrected data may appear better, but it is an artifact and the offset really needs to be corrected. At the other side of the spectrum, in the near-infrared, the correction significantly changes the shape of the absorption feature around 1030 nm by lowering the local minimum which has a huge impact on the SSA estimation. Using the procedure described in Section 2.1 Applying the SSA retrieval algorithm on the albedo to estimate SSA spectra , we find that estimated SSA drops from 53 to 41 $\mathrm{m^2\,kg^{-1}}$ after the correction for the noon acquisition and from 61 to 24 $\mathrm{m^2\,kg^{-1}}$ for the evening one. These values huge differences highlight the necessity of this correction and consequently the need to accurately measure for each acquisition the dark current and stray light.

To evaluate this processing step within the modeling framework developed in Section 2.3.1, we consider typical values of the noon acquisition: the spectrum has an amplitude of nearly 50000 and the dark current (offset) is around 3800. We further assume that this offset can estimated with an accuracy of 10% (pessimistic estimate for illustration), which results in $d' = 0.8\%$. Figure **??** shows the perfect albedo (black curve) calculated for a true SSA of 50 $\mathrm{m^2\,kg^{-1}}$ and the perturbed albedo (green curve) with $d' = +0.8\%$. Despite the small difference between the spectra, the estimated SSA is 55.4 $\mathrm{m^2\,kg^{-1}}$ which corresponds to a relatively large over-estimation of 10%, but remains under our criteria of 15%. Considering that the dark current can be usually estimated by much better than 10%, the impact on the SSA is always minor.

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The next processing step (stray light correction) is not shown because it has a negligible effect. Nevertheless, it is worth noting that we assume the stray light effect to be uniform on all the pixels of the CMOS sensor. If this is not the case, the correction of the stray light would require a more complex method and the evaluation of the error proposed here would be only a lower bound of the error.

5 The cross and absolute calibration steps (from orange in ~~graph a~~ top graph to violet curve in ~~graph b~~middle graph) also radically change the shape of the spectra. The most visible change on the radiance is mainly due to the absolute calibration because it takes into account the sensitivity of the spectrometer. As expected for a silicon-based sensor, the correction increases the signal in the blue and near-infrared relatively to the yellow and red. Nevertheless, the absolute calibration is not important for the albedo as explained before. The difference on the albedo graphs ~~(c)~~ before (orange curve) and after (violet curve)

10 the calibrations is only due to the cross calibration step between the downward- and upward-looking channels ~~. The correction factor (not shown)~~ and corresponds to a scaling that linearly decreases from 0.95 in the blue to 0.85 in the near-infrared. Because of this wavelength dependence, the SSA estimation is strongly affected, values decrease from 41 $m^2\,kg^{-1}$ before correction to 36 $m^2\,kg^{-1}$ after for the noon acquisition and from 24 $m^2\,kg^{-1}$ to 20 $m^2\,kg^{-1}$ for the evening one. ~~This correction is significant but it is~~ In the theory presented in Section 2.3.2 this corresponds to approximately $b = 0.10$. Using Figure **??**, it

15 means that the cross-calibration is definitely required to reach the 15% accuracy of SSA. If we assume that the correction is precise within 10% (for instance because of the limited reproducibility of the experiment to determine the reference spectra) it leads to a residual trend of $b \approx 0.01$ which is weak enough to reach the 15% accuracy. In addition, it is worth noting that ~~it~~ this correction is constant over the whole time-series and has therefore no impact on the relative temporal variations of SSA.

 ~~In contrast, the~~ The collector angular response correction (from violet to pink curves) has very small impact on the radiance

20 (~~graphs b~~middle graph) whatever the acquisition hour. The impact is also apparently moderate on the albedo ~~graph (c~~(bottom graph). The correction factor ranges from 0.98 in the blue to 0.99 in the near infra-red at noon and 1.00 to 1.07 at 8 pm. Even if these values are weak (maximum of 7%), the correction succesfully removes the decreasing trend in the shorter wavelengths of the visible (400 to 600 nm) that affects the evening acquisition. Albedo measurements by Nicolaus et al. (2010) show a similar artifact. Such a trend is visible throughout the timeseries at large SZA and the collector response correction usually performs

25 well by recovering a nearly constant value as expected.

 Regarding SSA estimation, the correction has a little impact for the noon acquisition (36.7 and 36.2 $m^2\,kg^{-1}$ respectively before and after the correction) as expected. In contrast, for the evening acquisition the SSA estimate increases from 20 $m^2\,kg^{-1}$ to 35 $m^2\,kg^{-1}$, the latter being close to the value at noon. This clearly shows that the correction of the collector response is crucial~~and performs well~~. The theoretical spectrum that fits the fully-corrected spectrum is shown in black in

30 Figure **??**. The differences between the fit and the observation are small, we can note a slight over-estimation at 800 nm and under-estimation at 970 nm which are very likely due to a small error in the refractive index of the ice ~~, since they are not specific to our instrument (Carmagnola et al., 2013).~~already pointed out in other studies (e.g. Carmagnola et al., 2013). In contrast, the over-estimation at 1030 nm is more likely an error in the observations resulting from the low sensitivity of the spectrometer in this wavelength range.

To further investigate the impact of the albedo uncertainties due to the instrument artifacts on the retrieved SSA , we use numerical experiments.

3.2 Stability

3.3 Numerical evaluation of the uncertainty

5 Most artifacts affecting the spectral radiometer results in a first approximation in either an offset in the radiances or a chromatic aberration. To quantify the respective effects of these artifacts on the estimation of the SSA, we perform simple and idealized modeling experiments as follows: we first compute the perfect albedo spectrum with the model described in Equation 18 for a given SSA (called true SSA) and perturb it to mimic the imperfect response Before deriving albedo and SSA over long periods of time, it is important to assess the overall stability of the instrument. Then, we estimate the SSA using the same procedure as

10 used with real data and deduce the error as the difference between the estimated and true SSA.

3.2.1 Offset

To study the effect of the dark current and stray light correction, we model the measured albedo as:

$$\alpha = \frac{S^{\mathrm{ref}}(\lambda) + d}{S^{\mathrm{inc}}(\lambda) + d},$$

where the true reflected and incident radiance spectra $S^{\mathrm{ref}}(\lambda)$ and $S^{\mathrm{inc}}(\lambda)$ are affected by an offset d. Dividing by the incident

15 radiance to make explicit the true albedo α^{true} yields:

$$\alpha(\lambda) = \frac{\alpha^{\mathrm{true}} + \frac{d}{S^{\mathrm{inc}}(\lambda)}}{1 + \frac{d}{S^{\mathrm{inc}}(\lambda)}}.$$

For $S^{\mathrm{inc}}(\lambda)$ we chose a Gaussian shape looking like the raw observations in Figure **??**:

$$S^{\mathrm{inc}}(\lambda) = S_{\mathrm{mode}} \exp\left[-\left(\frac{\lambda - 680\mathrm{nm}}{270\mathrm{nm}}\right)^2\right].$$

The amplitude of the spectrum S_{mode} is usually of the order of the resolution of the Digital Analog Converter of the spectrometer

20 (2^{16} in our case) if the integration time is optimal. We define $d' = d/S_{\mathrm{mode}}$, the relative contribution of the offset with respect to the spectrum amplitude. We then estimate the SSA from the perturbed $\alpha(\lambda)$ spectrum and calculate the difference with respect to the true SSA.

To apply this theory to assess the impact of an imprecise dark and stray light correction, we take typical values from the noon acquisition studied in Section 3.2 : the spectrum has an amplitude of nearly 50000 and the offset is around 3800. We further

25 assume that this offset can estimated with an accuracy of 10%, which results in a residual bias of 380 and thereby $d' = 0.8\%$. Figure **??** shows the perfect albedo (black curve) calculated for a true SSA of 50 and the perturbed albedo (green curve) with

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$d' = +0.8\%$. Despite the small difference between the spectra, the estimated SSA is 55.4 which corresponds to a relatively large over-estimation of 10%, but remains under our criteria of 15%.

To help in designing future spectral radiometer devices, we use the same approach to compute the error in SSA as a function of the relative offset d' ranging from 0 (perfect instrument) to 2% for several SSA values (Figure **??**). The result is nearly independent of the true SSA. To meet the 15% criteria, the offset due to dark current and stray light should be corrected with an accuracy better than d'=1%. In theory, such an accuracy can be achieved using a single measurement of one dark pixel as our spectrometer feature a signal-to-noise ratio of 1:500. By averaging many dark pixels, as we do, we ensure an even better accuracy meaning that the offset correction is not an issue in our case. Nevertheless, it is worth noting that we assume the stray light effect to be uniform on all the pixels of the CMOS sensor. If this is not the case, the correction of The leveling is a first and important parameter. It was carefully done at the installation of Autosolexs in December 2012 within the uncertainty allowed by spirit level (about 0.2°) but could have degraded. Nevertheless, in December 2015, new measurements with an electronic devices yielded insignificant movement of the stray light would require a more complex method and the evaluation of the error proposed here would be only a lower bound of the error. structure.

3.2.1 ~~Chromatic aberration~~

~~To investigate the effect of chromatic aberration, we consider the true albedo is multiplied by a linear function of the wavelength:~~

$$\alpha = \left(1 - b\frac{\lambda - \lambda_0}{\lambda_1 - \lambda_0}\right)\alpha^{\text{dir}}(\lambda, \theta)$$

where b controls the slope of the wavelength dependence. The limits λ_0=The radiometric stability is another parameter. The seasonal variations of downwelling irradiance are plotted in Figure **??** for the two heads and for the three summer seasons. The irradiance is integrated between 400 and λ_1=1100 1000 nm are chosen so that the instrument is assumed perfect at 400 in the blue and b represents the differential response of the upward and downward looking channels at 1100 . Figure **??** illustrates the results with true SSA of 50 and $b = 0.05$ (see below for actual values of b). We remind that the SSA estimation only uses data between 700 and averaged between 10h – 1050 . The SSA is estimated to 38.2 by the 2-parameter model and 27.6 for 1-parameter one. This corresponds to relative error of 24% and 45% respectively. The 2-parameter model is clearly better which can be explained by the fact that it is insensitive to the scaling of the albedo but relies mostly on the relative variations of albedo over the wavelength range used for the optimization. However, this error is larger than the 15% target accuracy.

An interesting feature of the 2-parameter model is that the albedo predicted with the optimal SSA and A values (orange curve) exhibits a bias in the visible not only with respect to the true albedo (black curve) but also to the observed one (green curve)14h (local time). These data include all weather conditions. Overall, there is a good year to year agreement. For instance, in Figure **??**, the difference is 0.024 on average over the range 400 − 550 . This difference (residuals hereinafter) is a consequence of the wavelength-dependent perturbation and can be exploited to assess the amplitude of this perturbation in real data because it does not require knowledge of the true SSA. It means that the acquisitions subject to residual chromatic aberration should feature a large difference in the visible range with respect to the spectrum predicted using the 2-parameter

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~~model. The previous calculation with $b = 0.05$ and true SSA=50 gives a difference of 0.024 and was found to corresponds to relative error of 24% of SSA. To explore how these values vary in other conditions, we perform several simulations. Figure ??a shows the relative error and the mean difference over the range $400 - 550$ as a function of the chromaticity parameter b for several SSA. The error varies almost linearly with b and the slope increases with SSA. Hence, to attain the target accuracy of 15% , the chromatic parameter should not exceed about 0.04 for SSA=20 and even less, 0.02, for SSA=100 . Figure ??b shows the univocal relationship between b and the mean residual. This result suggests that mean residual values under 0.01 indicate a chromaticity better than 0.02 and thus an SSA accuracy better than 15% .~~

~~We apply this criteria on the real data by calculating the albedo spectrum predicted by 2-parameter model using the optimal SSA and A obtained by fitting the model to the observations in the range $700 - 1050$. We then calculate the mean difference in the range $400 - 550$ and reject the acquisition if this difference is larger than 0.01. The skills of this filter are illustrated in the next section. It is worth noting that it can only work if the light-absorbing impurities are negligible because dust or black carbon usually lead to a decrease of the albedo in the blue-green (Warren, 1982) which would be interpreted as an artifact by our filter that is based on pure snow albedo calculation.~~

~~It is interesting to estimate the chromatic aberration associated with the different steps of the processing or other external sources of error. For instance, the cross-calibration factor ranges from 0.95 in the blue to 0.85 in the infrared (Section 3.2) which corresponds to approximately $b = 0.10$. It means that the cross-calibration is definitely required to reach the 15%accuracy. If we assume that the correction factor is accurate within 10%, for instance because of the limited reproducibility of the experiment to determine the reference spectra, it leads to a residual trend of $b \approx 0.01$ which is weak enough to reach the 15% accuracy. A similar conclusion can be drawn for the collector angular response correction that resulted in similar scaling factors, between 1.00 to 1.07 for the blue and near infra-red wavelengths respectively for the evening acquisition studied in Section 3.2 and much lower factors for the noon acquisition.~~

~~Errors of the leveling of the measurement heads also indirectly results in chromatic aberration because it affects the direct/diffuse balance. Following ? and Grenfell et al. (1994), the albedo error $\eta(\lambda, \theta)$ due to a tilt angle δ in the direction of the sun (worse case) can be written:~~

$$\underline{\eta(\lambda, \theta) \approx \left(1 - r^{\text{diff}}(\lambda, \theta)\right) \left(\frac{\cos(\theta - \delta)}{\cos(\theta)} - 1\right)}$$

~~where the diffuse term error has been neglected as suggested by ? results. Considering small errors, b can be estimated as follows:~~

$$\underline{b} \qquad \underline{\approx \eta(\lambda_1, \theta) - \eta(\lambda_0, \theta)}$$

$$\underline{\approx \left(r^{\text{diff}}(\lambda_1, \theta) - r^{\text{diff}}(\lambda_0, \theta)\right) \left(\frac{\cos(\theta - \delta)}{\cos(\theta)} - 1\right)}$$

~~For a tilt angle of~~ the irradiance observed by the head 1 ~~and under Dome C typical conditions, we obtain $b = 0.01$ for SZA=53 (as the noon acquisition studied in Section 3.2) and $b = 0.05$ for SZA=77 (as the evening acquisition) which is weak in the former case but is too high in the latter one to reach the 15% target accuracy. These results highlight the crucial role of~~

the leveling of the instrument. When, Autosolexs was installed, the leveling was carefully done within the uncertainty allowed by spirit level (about 0.2) but it is likely that it has degraded over time. Given that measuring tilt angle with electronic devices in Dome C winter conditions is challenging and that the instrument was not serviced to preserve the surface, the confidence in our data is based on the *ad hoc* filtering based on the 2-parameter model residual in the visible. during the 30 days after the solstice (period in common for the three seasons) has increased by +0.5 % and +4.2% the second and third years with respect to the first one. These values become -0.5%, and +3.0% respectively for the head 2, which is relatively small for an unattended instrument and considering that these values include the inter-annual variability of cloudiness. Only the third season shows a significant increase (around 4-5 %) which is unexplained.

The effect of the different artifacts has been evaluated independently of each other, but in real data, they combine, sometimes compensating, sometimes not. The assessment of the interactions between the artifacts is difficult and is not addressed here, but reaching 15% accuracy in real data may require stricter tuning of the filters

4 Observations, interpretation and comparison

This section presents the 3-year long time-series of retrieved SSA and provides an analysis of its quality and information content.

4.1 Diurnal cycle of SSA

The SSA estimated over the course of a day for each head is depicted in Figure **??**a for the two heads for the 10 January 2013 as a function of the local hour and the corresponding SZA. Gray symbols represent data rejected for one of the three reasons: presence of clouds, excessive chromatic aberration or A outside the range $0.9 - 1.1$.

This particular day clouds were detected only at 3:25 and after 21:00 but the detection may not be reliable at high SZA because of the simplistic approach used to scale the reference spectrum to a given SZA (Section 3.3). The other cases of rejection are equally due to the two other criteria. It appears clearly that it usually happens when the sun is low on the horizon, with a SZA higher than 70°, which is expected. However, in many cases, the SSA of rejected acquisitions is close to valid ones which could indicate that our criteria are too conservative. On the other hand, some data at very high SZA are not rejected while we believe they should have been. For this reason, we let the criteria being conservative and even discarded acquisitions at SZA higher than 75° in the following.

Valid data show small diurnal variations between 32 and 35 $\mathrm{m^2\,kg^{-1}}$ for head 1 and slightly larger between 31 and 40 $\mathrm{m^2\,kg^{-1}}$ for head 2. Root mean squared variations is respectively 0.9 and 2.6 $\mathrm{m^2\,kg^{-1}}$ which is well under the 15% target accuracy. The mean difference between both heads is 2.6 $\mathrm{m^2\,kg^{-1}}$. The nearly perfect symmetry around noon suggests that these variations are not do not come from real SSA variations but are rather errors linked to SZA real but are caused by SZA-dependent errors either due to uncorrected artifacts in the observations or limitation of the theoretical model used for the estimation.

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The diurnal cycle is shown in Figure **??**b (1 March 2013). The SSA is much higher than in the previous case and the number of hours with SZA lower than 75° is more limited. In addition, more data are rejected, mostly because of the chromatic aberration. Altogether, the valid values range between 53 and 74 $m^2\,kg^{-1}$ representing a large amplitude of 30%. The standard deviation is about 3.5 $m^2\,kg^{-1}$ for both heads, that is 10% for a $\pm 1\sigma$ variation. It was impossible to determine whether these variations are real or result from measurement and retrieval errors. These variations must therefore be interpreted as an upper bound estimate of the uncertainty of the retrieved SSA.

4.2 Statistical characteristics of the seasonal variations

The seasonal variations of SSA estimated from the two heads over three summers (December 2012 to March 2015) are shown in Figure **??**. The daily mean of valid data is plotted in color and each individual valid data is shown in blue in the background to give an idea of the diurnal variability (or maybe residual error). In this paper, we perform a quality assessment of this series. The interpretation of the geophysical features is addressed in Libois et al. (2015).

At the beginning and end of the seasons, the SSA is usually high and exhibits large day to day variations that are suspicious. Several issues like the those due to high SZA, the weaker sensitivity of the SSA – albedo relationship at higher SSA (Domine et al., 2006; Gallet et al., 2009) and the lower number of valid data entering the daily mean during these periods with short daylight, probably combine during these periods. Because of this lower quality, the time-series in Libois et al. (2015) only uses noon data to minimize the high SZA issue and is limited to SZA lower than 67° which corresponds at Dome C to the period 18 October to 27 February. Here, because we introduce the filter based on the detection of chromatic aberration, we apply the algorithm for SZA up to 75°, relying on the filter to reject low quality data at high SZA. As a result, the period presented is slightly longer (about two weeks each sides but with little data), but the interpretation at the extremes is still subject to caution. Using the daily mean of the valid data instead of the noon value has very little impact, the difference ranges between -1.0 and +1.5 $m^2\,kg^{-1}$ depending on the season and head.

The two heads give very close values (within the daily range of variation) for some periods while significant differences are observed for others usually in the spring (October 2013 and most remarkably October-November 2104) and in the fall (March 2013 and 2014). If we consider December-January to assess the statistics, we find the difference between heads 1 and 2 to be -2.7, 1.0 and 0.0 $m^2\,kg^{-1}$ on average and 3.2, 2.4 and 2.2 $m^2\,kg^{-1}$ root mean square (rms), respectively for the three seasons. This rules out any significant bias between the two heads which could have come from issues with the cross-calibration, the only processing step that uses different data depending to the head. This also indicates that the seasonal stability of the instrument and processing is of the same order as the diurnal standard deviation (Section 4.1). Conversely, it means that the difference observed during some periods might originate from the difference of SSA between the footprints of the two heads (Section 4.3).

Altogether these results converge to a precision much better than the targeted 15% which ensures that the seasonal variations can be interpreted as a geophysical signal (Libois et al., 2015). However, it does not presume of the absolute accuracy which mostly depends on the validity of the model in general and the choice of the parameter values (shape factors, assumption of a flat surface...).

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4.3 ~~Representativeness of the estimated SSA: horizontal~~ Horizontal variability between both heads

~~Both heads depict an~~

The two heads located 2-m apart show very similar variations most of the time (Figure **??**) and agree on the overall seasonal decrease from values higher than ~~60 − 80~~ 60–80 $\mathrm{m^2\,kg^{-1}}$ in October to a minimum around 25 $\mathrm{m^2\,kg^{-1}}$ occurring in January and/or February. ~~While they show very similar values~~ However, they also significant differ during some periods~~, significant differences occur during others. The two heads are located 2-m apart.~~. Being at 2 m above the surface and accounting for the cosine weighting of the collector, we can estimate that 50% of the ~~signal they receive from the surface is reflected from~~ received signal comes from a disk of 2 m in radius (12 $\mathrm{m^2}$ in surface area), and 75% from a disf of 3.5~~m~~ m in radius (38 $\mathrm{m^2}$). Despite the proximity and a slight partial overlap of the footprints, differences between the two heads can be particularly marked as in Spring 2014. We could invoke the fact that snow accumulation is ~~known to be an irregular and heterogeneous process~~ irregular at Dome C (Petit et al., 1982; Libois et al., 2014a) leading to irregular deposition on the surface. However, this argument is not compatible with the ~~lack~~ absence of difference observed ~~in the summers between the two heads while~~ during the summers although several precipitation events ~~do occur (Libois et al., 2015) and have a clear impact on SSA~~ usually occurred with a clear effect on the observed SSA (Libois et al., 2015) but did not produces differences between both heads.

In spring 2014, the large and persistent difference could be ascribed to the presence of an erosional feature under head 1 (Figure **??**) and a wind-form feature (elongated dune) under head 2. The latter could be composed of recently deposited snow while the former feature sastrugi and older snow layers emerging on the surface. Nevertheless, this interpretation is subject to caution because visible photography is insensitive to SSA, as evidenced by the ~~weak difference between~~ resemblance of the two pictures taken 12 October and 27 November despite a three-fold decrease of the SSA.

~~The diurnal cycles of the estimated SSA during spring 2014 is small compared to the difference between the heads, which suggests the difference between the two heads is significant, but the cycle is~~ During this period, the diurnal variations of SSA are not symmetrical with respect to noon (not shown) as opposed to the cycles shown in Figure **??**. ~~This could be caused by~~ A possible reason is the inhomogeneity of the surface SSA. Since the surface is particularly rough and the sun low on the horizon during this period, the shadows are significant and change with the solar time. Hence, the "average" spectrum measured by the instrument comes from different areas depending on the time of the day and those areas may have different SSA. The direct effect of the roughness could also affect the symmetry of the diurnal cycle when the surface is rough but assessing the spectral effect of the roughness is difficult and beyond the scope of this study.

4.4 ~~Representativeness of the estimated SSA: vertical gradient~~ Comparison with ASSSAP

~~A straightforward approach to determine the validity and accuracy of the instruments and algorithm is to compare estimated SSA with manual measurements optical devices like ASSSAP (Arnaud et al., 2011; Libois et al., 2015) . Although this device uses similar principles to the algorithm presented in Section 2.1, an important difference is the use of a longer wavelength (~~ Manual measurements of surface snow SSA were taken with ASSSAP (Arnaud et al., 2011; Libois et al., 2015) a device using reflectance at 1310 nm ~~) resulting in a significant difference in the depths probed by both instruments.~~

Figure ?? shows individual manual measurements taken during the 2013-2014 seasons with ASSSAP along with Autosolexs SSA time-series (average of the two heads). Measurements of surface snow SSA with ASSSAP follow to estimate SSA using similar theory as used here. For these measurements, we followed a specific protocol that takes care to preserve the surface and prevent post-sampling transformation of the snow. More than 5 manual measurements were taken every day in a site located approximately 1 km from Autosolexs. The sampling strategy was not random but aimed at covering the variety of different surface facies identified by the operator. It implies that the average of the values may differ from the true areal average of SSA when the spread of SSA is large. Figure ??

Figure ?? shows individual manual measurements taken during the 2013-2014 seasons with ASSSAP along with Autosolexs SSA time-series (average of the two heads). Figure ?? clearly shows that most manual measurements significantly differ from the SSA estimated from Autosolexs. However the lowest values seem to agree with Autosolexs during the first part of the season, until about 8 January. After this, the spread of manual measurements is significantly reduced and despite higher values of SSA with ASSSAP, the temporal variations are correlated with those of Autosolexs.

Although this comparison does not help validate or estimate the accuracy of Autosolexs SSA, it This comparison clearly demonstrates that both SSA – abusively qualified of surface snow SSA – are representative of different thicknesses owing to the different wavelengths used by the respective instruments which combines with a sharp gradient of SSA in the first millimeters under the surface the SSA retrieved in different wavelength ranges differ, which we attribute to strong heterogeneity of SSA between the uppermost 5 mm and 1 cm according to the calculation presented in Section 2.2.

The first study to suggest such a gradient is Grenfell et al. (1994). They found at South Pole that adding a thin layer of very small snow grains (over 100) on the surface in their radiative transfer calculations was required to match their albedo measurements. Carmagnola et al. (2013) at Summit Greenland uses spectral data in the wavelength range encompassing Autosolexs and ASSSAP operating ranges and found that tuning the surface layer SSA helps to improve the agreement with the observations but did not explore values as high as Grenfell et al. (1994) did. In contrast, Gallet et al. (2011) measured SSA vertical profiles at Dome C and obtained radiative transfer calculations in agreement with Hudson et al. (2006) observations without adding such a layer. Using snowpack numerical modeling, Libois et al. (2015) obtained SSA profiles at Dome C with a sharp gradient near the surface. It is worth noting that their simulation large difference in the first part of the season can be attributed to the frequent blowing snow events occurring during this period but the difference persists over 8 January despite calmer conditions. This persistence of large difference of SSA over a small vertical distance is not well understood (Gallet et al., 2011). Clear-sky precipitation (a.k.a diamond dust) is a possible candidate but snow evolution simulation performed by Libois et al. (2015) uses ERA-Interim as input which forecasts occasional precipitation events leading . The reanalysis successfully forecasts the precipitation events which leads to marked rises of SSA (Figure ??) but does not forecast clear-sky precipitation (a.k.a diamond dust) despite its frequent observation at Dome C. This include diamond dust. Since the simulation agrees with the SSA observations, this indicates that diamond dust is not required in the model to get a sustained high surface SSA and that including the diamond dust in the simulation could increase even more the predicted gradient on the surface. We can conclude that a sharp significant SSA gradient in the first uppermost centimeters exists with SSA possibly exceeding 100

$\mathrm{m}^2\,\mathrm{kg}^{-1}$ on the very surface and that this gradient persists over long periods, even during periods without precipitation ~~and blowing snow~~.

As a consequence, the vertical representativeness of the SSA estimated in this study is an important information in order to analyze the data and compare them to models or other observations. It is therefore important to determine the thickness of the

5 layer sampled by Autosolexs. The answer is not straightforward nor univocal because the penetration depth of radiation depends on the wavelength and the vertical profile of snowproperties, including SSA itself and density. Furthermore, the estimated SSA value is not a simple average over a given thickness but is weighted by a decreasing function of the depth. This function is exponential if the snow is homogeneous, but more complex otherwise.

Only radiative transfer calculations can give a precise estimation of the weight function. To evaluate it for an homogeneous

10 snowpack, we run the two-stream radiative transfer model TARTES (Libois et al., 2013) considering a semi-infinite medium with a density of 270 corresponding to mean surface conditions (Libois et al., 2014a) and a fixed SSA. A layer with variable thickness h is added on top of it, with the same properties, excepted that a small perturbation is added to the SSA (e.g. +1 , but the exact value does not alter the final result as long as it is small compared to the actual value of SSA). The relative contribution of the uppermost layer to the albedo is then defined as the quantity $(\alpha(h) - \alpha(0))/(\alpha(\infty) - \alpha(0))$ where $\alpha(h)$ is

15 the albedo calculated by TARTES for the thickness h and averaged over the wavelength range used by Autosolexs (700-1050). The contribution is shown in Figure **??** (plain line) as a function of h for different SSA values. Results show that the uppermost 10 mm snow layer contributes to nearly 60% of the albedo for SSA of 20 and 85% for higher SSA of 100 . Conversely, the layer contributing to 80% of the signal is 18, 12 and 8thick for SSA of 20, 50 and 100 respectively. As a conclusion, 1 is a good approximation for typical Dome C conditions. Similar simulations have been run at 1310 to illustrate the difference with

20 ASSSAP (dash line in Figure **??**). In this case most of the signal comes from uppermost 5 , at most, in any conditions. Note that this depth is inversely proportional to the density so that the thickness values presented here should be multiplied by about two in case of fresh snow or surface hoar.

The question of the representativeness of the SSA estimated with the instruments is relevant for comparing the observations with snow models like Crocus or to study the snow processes. As far as the broadband albedo and the surface radiative budget

25 are concerned, SSA estimated in the range 700 – 1050 is the most relevant as most of the solar energy is absorbed in the near-infrared range (Gardner and Sharp, 2010).

5 Discussions and Conclusion

~~This study presents~~ This study introduces an algorithm to retrieve the specific surface area of the surperficial snow from spectral albedo measurements, along with Autosolexs, a multi-channel spectral radiometer used to automatically measure

30 spectral albedo in the visible and near infrared range (under 1000 nm)~~and an algorithm to retrieve the surface specific area of the surface snow~~. Despite the specificity of this home-made instrument and the operating conditions of Dome C, several general recommendations can be made regarding the design and deployment of such devices: 1) Using fiber optics to transmit the light ~~make~~ makes possible to bury the temperature-sensitive components of the instrument under the snow. This was necessary for

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Dome C considering that temperature is frequently below -40°C even during the daylight. Maintaining the device at constant above freezing temperature avoids deterioration and drift of the spectrometer. However, it costs some energy which would require adequate dimensioning of solar panels or wind turbines in the case of autonomous stations. The instrument has been working since December 2012, even in winter, with only a few short interruptions due to power outage. 2) Use of optical switch instead of two or more spectrometers is a good option. It is cost effective, avoids the need of inter-calibration of the spectrometers (Nicolaus et al., 2010) which may drift independently over time (France et al., 2011) but it is subject to switching delay and reproducibility, and involves mechanical parts. It allowed us to deploy 2 measurement heads located a few meters apart which provided evidence of spatial difference during some periods. With a 16-channel optical switch, it is technically possible to set up 7 measurement heads 3) The leveling of the head should be and remain horizontal not only for albedo which has been highlighted by numerous authors (e.g. ?)(e.g. Bogren et al., 2016), but also for SSA retrieval. Under typical polar summer conditions or in winter in the mid-latitudes, we estimate the leveling must be better than 1° to get at least the noon acquisition accurate enough to retrieve estimate SSA with an accuracy of 15%. Maintaining automatically the leveling or and measuring the inclination at low temperature is challenging. 4) Designing good light collectors and correcting from for their imperfect response is necessary for SZA above 60° and becomes more and more difficult critical above 70°. The determination of the atmospheric diffuse/direct ratio and cloud mask deserves further work compared to what has been achieved implemented here. Using the measured incident spectrum to infer the actual ratio is an avenue. 5) Filtering data based on the difference between observed and estimated albedo in the visible range allows to discard discards data subject to any artifact that produces excessive chromatic aberration. The approach developed here does not require high spectral resolution data, it can be applied to multi-band radiometers like those carried by satellites. However, it requires some assumptions like negligible impurity content and could interfere with the atmospheric correction often based on the blue channels for space-borne sensors.

Part of the difficulties addressed in this study arises from the weak sensitivity of the SSA-reflectance relationship at wavelengths under 1 μm, especially for the high SSA encountered at Dome C ($50 \, \mathrm{m^2 \, kg^{-1}}$ and higher) (Domine et al., 2006; Gallet et al., 2009). Extending the present instrument further in the infrared seems attractive because the sensitivity to SSA is greater so that larger error on the albedo measurement would be acceptable for a similar target accuracy of SSA. However it comes at a price. Building light collectors with a good angular response is increasingly challenging at longer wavelengths due to the decreasing scattering. The balance between theoretically more sensitive wavelengths and increasing chromatic aberrations needs to be weighted.

The influence of the surface roughness on the albedo and estimated SSA also deserves further work. Warren et al. (1998) explained that "introducing roughness will cause no change to the albedo of a surface whose albedo is 0.0 or 1.0. The greatest effect is for intermediate values of albedo; in the case of snow, this means near-infrared wavelength" which implies that roughness change tends to reduce the albedo in the 1030 nm absorption feature more than at shorter wavelengths where it is closer to 1. . This implies that our algorithm, by assuming a flat interface, would under-estimate the SSA over rough surfaces. Moreover, with measurements heads only 2 m above the surface, the scale of the measurements and that of the roughnesses are of the same order resulting in a more complex problem than that considered in modeling studies (Leroux and Fily, 1998). Spatial survey of albedo and SSA variations could be used to evaluate this effect. The anisotropy of the roughness and the

slope of the terrain (relevant to mountainous regions only) are expected to introduce a spurious diurnal cycle and need to be considered in the future as well.

The present study, by retrieving 3-year long time series of SSA, confirms the general coarsening of snow grains during the summer observed in several studies (Jin et al., 2008; Picard et al., 2012) and recently shown to be predicable by snow metamorphism modeling (Libois et al., 2015). However, despite its length, it features relatively little inter-annual variations compared to that observed by satellite and provides little evidence yet of modulation of the grain size by precipitation at the seasonal timescale as hypothesized by Picard et al. (2012). An important perspective of this work is to maintain Autosolexs at Dome C and to develop the installation of ground-based spectral radiometers in Antarctica.

Data Availability

The corrected and quality-checked albedo spectra and SSA time-series will be available from the Pangea database once the peer-reviewed process has validated the method.

Author contributions. G. Picard and L. Arnaud developed and built Autosolexs, L. Arnaud and Q. Libois deployed it at Dome C and performed the calibration experiments. G. Verin performed the collector characterization. M. Dumont conducted the atmospheric simulations. G. Picard and Q. Libois developed the retrieval algorithm and prepared the manuscript with contributions from the other authors.

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