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Authors response to reviewer.

Paper title: *Monitoring glacier albedo as a proxy to derive summer and annual surface mass balances from optical remote-sensing data*
Authors: L. Davaze, A. Rabatel, Y. Arnaud, P. Sirguey, D. Six, A. Letréguilly, M. Dumont
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Dear Handling Editor: Dr. Valentina Radic, dear reviewer: Dr Dozier,

We first thank you for taking the time to carefully read and comment the revised version of our paper.

According to the referee comments, we also would like to apologize for our short response in the previous review on the radiation balance computation, which was, retrospectively, lacking of argumentation.

Hereafter, we hope that you will find an exhaustive response to your comments (the text of the paper appears in blue and the changes we propose are underlined).

Best regards,

Lucas Davaze on behalf of all the authors.

Response to review #2

General comments by V. Radic:

Your revised manuscript had been sent for another round of reviews since the reviewers asked for substantial revisions. By now I have received a review from Jeff Dozier and would like to give you a chance to respond to his comments and provide any further revisions if you consider necessary.

Jeff's opinion is that the revised manuscript is publishable, but he was disappointed that you refused his suggestion to compute a radiation balance. In your response letter please provide a convincing rationale for declining this suggestion.

As a reminder, below is the comment from the first review by Prof. J. Dozier :

"A lot of work has gone into this paper, but the presentation of the results is not satisfactory. The measurement of albedo is rigorously and correctly done, following Dumont et al. (2012). However, the statistical correlation analysis of the minimum summer albedo sweeps a lot of interesting processes into it, and raises more questions. Specifically, how do the spatial and temporal variability of the albedo affect the SMB? And how does the uncertainty in the albedo estimate relate to the uncertain in the absorption, i.e. "1- α ", which is what we really want. Therefore, a much stronger paper would result if instead you compute the absorbed solar radiation by date, and integrate it over periods of interest. This calculation, perhaps called the radiative forcing in $W\ m^{-2}$, takes into account the variability on the glacier, along with the different glaciers' exposure to the incoming solar radiation. Painter has used this measure. Painter and his colleagues have used this measure effectively to characterize the effects of dust on snow. You've already had to do much of the work in order to compute the albedo values, so take the next step."

Authors reply:

We agree on the fact that snow albedo variation, especially through the darkening effect from dust deposition (Miller et al., 2016), and the intrinsic snow albedo feedback from snow grain growth due to warmer meteorological conditions (Warren, 1982), are important drivers of the snowpack melt. As stated in Prof Dozier comment, changes in glacier surface albedo have therefore a direct impact on the absorption of solar radiation by the snow/ice layers. Computing the absorbed solar radiation by date and for each glacier would then be an appropriated protocol to estimate the impact, in terms of snow or ice loss, of a changing surface albedo. However, our study is meant to propose a new method to derive summer and annual surface mass balance (SMB) of glaciers by analyzing the albedo signal from optical satellite images. It is beyond the scope of this study to describe and quantify the impact of surface albedo changes on snow and ice melt; once again our goal is to present a methodological framework to compute annual and summer glacier SMB at regional scale with the aim of substantially increase the number of monitored glaciers.

We agree that computing the albedo at pixel scale of 250m resolution, for 30 glaciers offers the opportunity to study at a regional scale the impact of a changing surface albedo onto glacier

SMB, by studying the potential effect of increasing dust content, glacier orientation, or grain growth enhancement... and this will most probably be a next step of the work we can do with the data set of albedo maps presented in this study and data are available upon request from the corresponding author.

According to your remark, we added in the conclusion of the new manuscript an assessment of our discussion on the potential of computing a radiation balance:

"For future works, the MODIS archive together with albedo maps, cloud and shadow masks processed with MODImLab, together with validation data from AWS offer a unique dataset to monitor the temporal and spatial evolution of the surface albedo of glaciers at a regional scale. Computing for instance the absorbed solar radiation (Bair et al., 2016) by date and for each glacier would then be an appropriated protocol to estimate the impact of a changing glacier surface albedo in terms of snow or ice melt. Quantifying albedo changes and resulting mass losses with such an approach would be of major interest to better understand the potential effects of possibly increasing dust content, glacier orientation or snow grain growth on glacier surface melt processes. "

Bair, E. H., Rittger, K., Davis, R. E., Painter, T. H. and Dozier, J.: Validating reconstruction of snow water equivalent in California's Sierra Nevada using measurements from the NASA Airborne Snow Observatory, *Water Resour. Res.*, 52(11), 8437–8460, doi:10.1002/2016WR018704, 2016.

Miller, S. D., Wang, F., Burgess, A. B., McKenzie Skiles, S., Rogers, M. and Painter, T. H.: Satellite-Based Estimation of Temporally Resolved Dust Radiative Forcing in Snow Cover, *J. Hydrometeorol.*, 17(7), 1999–2011, doi:10.1175/JHM-D-15-0150.1, 2016.

Warren, S. G.: Optical properties of snow, *Rev. Geophys.*, 20(1), 67–89, doi:10.1029/RG020i001p00067, 1982.

1 **Monitoring glacier albedo as a proxy to derive summer and**
2 **annual surface mass balances from optical remote-sensing data**

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16

17

18 **Abstract.**

19 Less than 0.25% of the 250,000 glaciers inventoried in the Randolph Glacier Inventory (RGI
20 V.5) are currently monitored with in situ measurements of surface mass balance. Increasing
21 this archive is very challenging, especially using time-consuming methods based on in situ
22 measurements, and complementary methods are required to quantify the surface mass balance
23 of unmonitored glaciers. The current study relies on the so-called albedo method, based on the
24 analysis of albedo maps retrieved from optical satellite imagery acquired since 2000 by the
25 MODIS sensor, onboard of TERRA satellite. Recent studies revealed substantial relationships
26 between summer minimum glacier-wide surface albedo and annual surface mass balance,
27 because this minimum surface albedo is directly related to the accumulation-area ratio and the
28 equilibrium-line altitude.

29 On the basis of 30 glaciers located in the French Alps where annual surface mass balance are
30 available, our study conducted on the period 2000-2015 confirms the robustness and reliability
31 of the relationship between the summer minimum surface albedo and the annual surface mass
32 balance. For the ablation season, the integrated summer surface albedo is significantly
33 correlated with the summer surface mass balance of the six glaciers seasonally monitored.
34 These results are promising to monitor both annual and summer glacier-wide surface mass
35 balances of individual glaciers at a regional scale using optical satellite images. A sensitivity
36 study on the computed cloud masks revealed a high confidence in the retrieved albedo maps,
37 restricting the number of omission errors. Albedo retrieval artifacts have been detected for
38 topographically incised glaciers, highlighting limitations in the shadows correction algorithm,
39 although inter-annual comparisons are not affected by systematic errors.

40

41 **1 Introduction**

42 Mountain glaciers represent only 3% of the ice volume on the Earth but contribute
43 significantly to sea level rise (e.g. Church et al., 2013; Gardner et al., 2013; Jacob et al., 2012).
44 In addition, millions of people partly rely on glaciers, either for drinking water, agriculture or
45 related glacier hazards (Baraer et al., 2012; Chen and Ohmura, 1990; Immerzeel et al., 2010;
46 Kaser et al., 2010; Sorg et al., 2012; Soruco et al., 2015). The surface mass balance (SMB) of
47 glaciers is directly driven by the climate conditions; consequently, glaciers are among the most
48 visible proxies of climate change (Dyurgerov and Meier, 2000; Haeberli and Beniston, 1998;
49 Oerlemans, 2001; Stocker et al., 2013). Measuring and reconstructing glacier SMB therefore
50 provides critical insights on climate change both at global and regional scales (Oerlemans,
51 1994).

52 Systematic SMB monitoring programs began in the late 1940s - early 1950s in most of the
53 European countries (e.g., France, Norway, Sweden, Switzerland). Gradually, more glaciers
54 have become monitored, reaching the present worldwide figure of 440. However, this
55 represents only a small sample of the nearly 250,000 inventoried glaciers worldwide (Pfeffer
56 et al., 2014). Among the existing methods to quantify changes in glacier SMB, the well-
57 established glaciological method has become a standard widely used worldwide yielding most
58 of the reference datasets (World Glacier Monitoring Service, WGMS, Zemp et al., 2015).
59 Based on repeated in situ measurements, this method requires intensive fieldwork. This
60 method is however unable to reconstruct SMB of unmonitored glaciers. The Global Terrestrial
61 Network for Glaciers (GTN-G) aims at increasing substantially the number of monitored
62 glaciers to study regional climate signal through changes in SMB. To reach this objective, the
63 development of methods complementary to the ground-based glaciological method is therefore
64 required. Since the 1970s, several methods have taken advantage of satellite imaging to
65 compute changes in glacier volume (Kääb et al., 2005; Rabatel et al., 2017; Racoviteanu et al.,
66 2008). Several glacier surface properties have thus been used as proxies for volume
67 fluctuations: changes in surface elevation from differencing digital elevation models (DEM)

68 (e.g. Belart et al., 2017; Berthier et al., 2016; Gardelle et al., 2013; Ragettli et al., 2016; Shean
69 et al., 2016); end-of-summer snow line elevation from high spatial resolution optical images
70 (Braithwaite, 1984; Chinn et al., 2005; Meier and Post, 1962; Mernild et al., 2013; Rabatel et
71 al., 2005, 2008, 2016; Shea et al., 2013); mean regional altitude of snow from low spatial
72 resolution optical images (Chaponniere et al., 2005; Drolon et al., 2016); or changes in the
73 glacier surface albedo from high temporal resolution images (Brun et al., 2015; Dumont et al.,
74 2012; Greuell, W. et al., 2007; Greuell and Knap, 2000; Shea et al., 2013; Sirguey et al.,
75 2016). Widely used over icecaps or large ice masses, satellite derived DEM could not yet be
76 confidently used to compute annual or seasonal SMB of mountain glaciers, even if recent
77 studies have revealed promising results for determining SMB changes of large mountainous
78 glacierized areas (Belart et al., 2017; Ragettli et al., 2016). The method based on the
79 correlation between the regional snow cover and glacier SMB have shown satisfying results to
80 retrieve seasonal SMB, especially for the winter period. This method allowed the
81 quantification of 55 glaciers SMB in the European Alps over the period 1998-2014 (Drolon et
82 al., 2016). The method based on the identification on high spatial resolution optical images of
83 the end-of-summer snow line altitude has shown encouraging results in the French Alps,
84 multiplying by six the available long-term annual SMB time series (Rabatel et al., 2016), but
85 need to be automatized to compute glacier SMB at regional scales. In addition, monitoring
86 glacier surface properties on the daily or weekly basis and over large glacierized regions is still
87 challenging with high spatial resolution images. The current study is based on the albedo
88 method used in Dumont et al. (2012), Brun et al. (2015) and Sirguey et al. (2016). Images
89 from the MODerate resolution Imaging Spectroradiometer (MODIS) are processed to compute
90 daily albedo map of 30 glaciers in the French Alps over the period 2000-2015. Then, we rely
91 on the methodological framework proposed by Sirguey et al. (2016) on Brewster Glacier
92 (New-Zealand), looking at the relationships between annual and seasonal SMB and the
93 glacier-wide averaged surface albedo $\bar{\alpha}$. Our overall objective is to study the relationships

94 between glacier SMB and albedo by: (i) reconstructing the annual albedo cycle for 30 glaciers
95 in the French Alps for the period 2000-2015; (ii) linking the albedo signal to the summer
96 components of the SMB as well as to its annual values for 6 and 30 glaciers, respectively; (iii)
97 assessing the sensitivity of the retrieved albedo towards tuning parameters (cloud coverage
98 threshold for images processing, reliability of detected shadows). Section 2 presents the
99 available SMB datasets used for the comparison and describes briefly the in situ automatic
100 weather stations (AWS) used to assess the quality of MODIS retrieved albedo. The method to
101 retrieve albedo maps is described in Sect. 3. Results are presented and discussed in Sect. 4 and
102 5. The conclusion gathers the main results of the study and provides perspectives for future
103 works.

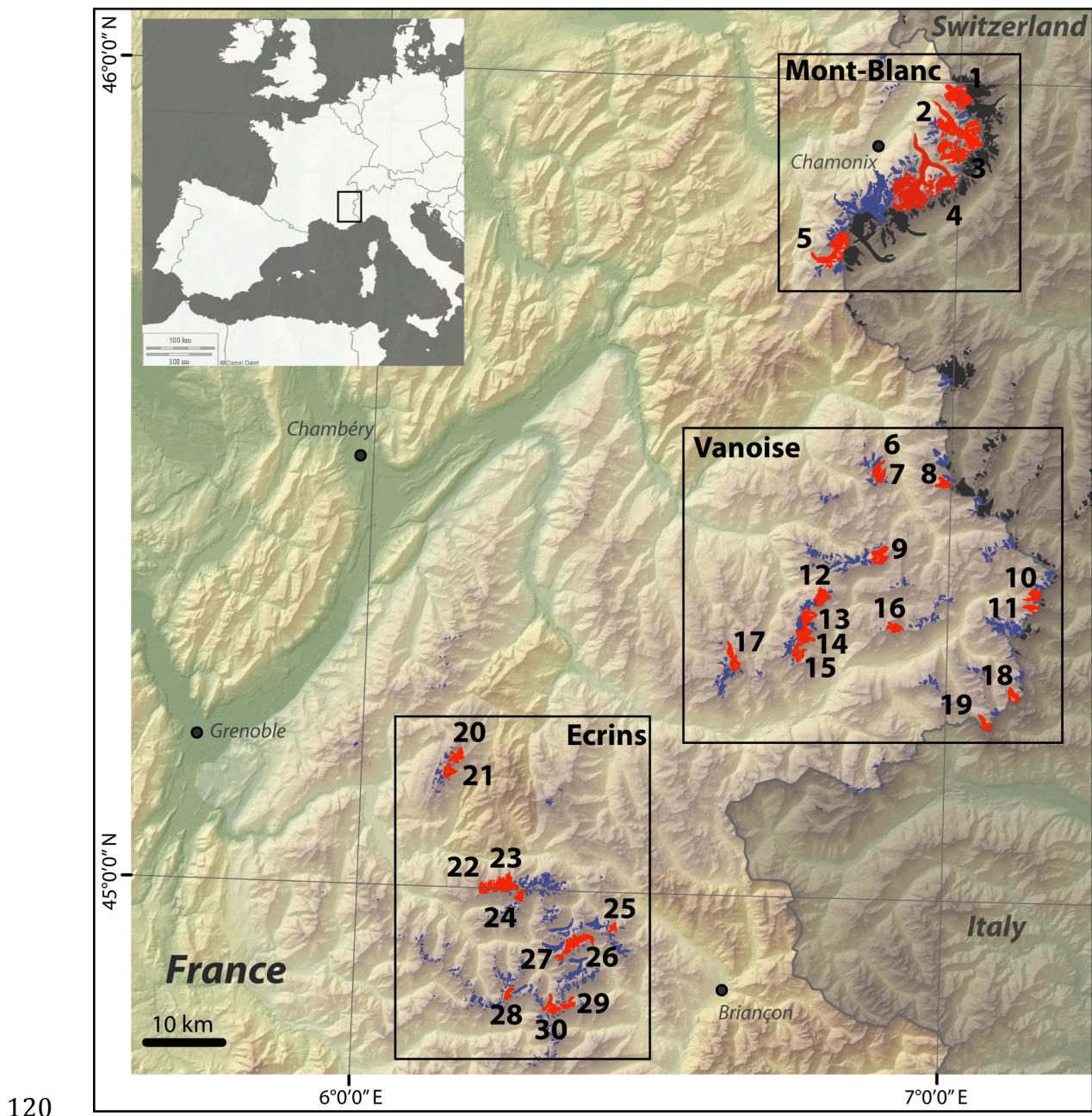
104 **2 Study area and data**

105 **2.1 Site description**

106 The study focuses on 30 glaciers located in the French Alps (Fig. 1). Each glacier can be
107 classified as mountain glacier, extending over an altitudinal range from around 1600 m a.s.l.
108 (Argentière and Mer de Glace glaciers) to 4028 m a.s.l. (Blanc Glacier), and located between
109 the coordinates: 44°51" N to 46° N and 6°09" E to 7°08" E. The cumulative glacial coverage
110 considered in the present study is 136 km², i.e. half of the glacier surface area covered by 593
111 inventoried glaciers over the French Alps for the period 2006-2009 (Gardent et al., 2014).

112 Studied glaciers have been selected following four criteria related to the availability of field
113 data and remote sensing constraints, namely: (i) the annual glacier-wide SMB for the study
114 period had to be available; (ii) the glacier surface area had to be wide enough to allow robust
115 multi-pixel analysis; (iii) the glacier had to be predominantly free of debris to allow remotely-
116 sensed observations of the albedo of snow and ice surfaces; and (iv) summer SMB records had
117 to be available to consider summer variability. Finally, 11 glaciers have been selected in the
118 Ecrins range, 14 in Vanoise and 5 in Mont-Blanc (Fig. 1, and listed Table 1).

119



121 **Figure 1** Map of the region of interest with the studied glaciers shown in red (numbers refer to
 122 Table 1). The four AWS used in the present study were set up on Saint-Sorlin Glacier (n°20).
 123 Adapted from Rabatel et al. (2016).

124
 125 **2.2 MODIS satellite images**
 126 The MODIS sensor, onboard the TERRA - EOS/AM-1 satellite is acquiring near-daily images
 127 of the Earth since February 25th, 2000. With 36 spectral bands ranging from 0.459 to 14.385
 128 μm , and spatial resolution ranging from 0.25 to 1 km depending on the spectral band, MODIS

129 is nowadays one of the most used optical sensors for land surface observations. Because of its
 130 short temporal revisit time, its long acquisition period and its moderate resolution, images
 131 from MODIS are the most suitable for the present work. We therefore rely on about 15,000
 132 MODIS calibrated Level 1B (L1B) swath images.

#	Name	Mask size [Pixel]	$b_a = P_1^a \bar{\alpha}_a^{min} + P_2^a$				$b_s = P_1^s \bar{\alpha}_s^{int} + P_2^b$			
			r^2	RMSE	P_1^a	P_2^a	r^2	RMSE	P_1^s	P_2^s
1	Tour	71	0.78	0.61	14.9	-7.8				
2	Argentière	111	0.74	0.39	16.8	-8.4	0.76	0.27	12.3	-10.1
3	Talèfre	40	0.46	0.73	17.0	-8.0	0.46	0.69	15.9	-12.1
4	Mer de Glace	246	0.16	0.89	8.7	-5.8	0.69	0.31	15.3	-12.1
5	Tré la Tête	38	0.43	1.25	22.8	-10.0				
6	Savinaz	7	0.23	1.27	12.3	-7.4				
7	Gurraz	17	0.29	0.77	9.8	-5.8				
8	Sassière	19	0.52	0.67	8.2	-4.9				
9	Grande Motte	30	0.83	0.53	13.6	-6.5				
10	Mulinet	18	0.33	0.62	7.7	-4.5				
11	Grand Méan	11	0.44	0.64	7.8	-4.2				
12	Arcelin	37	0.64	0.52	6.6	-3.7				
13	Pelvē	44	0.41	0.75	8.7	-5.7				
14	Arpont	41	0.28	1.0	9.8	-5.8				
15	Mahure	20	0.55	0.66	10.1	-5.1				
16	Vallonnet	19	0.36	0.66	3.4	-2.0				
17	Gebroulaz	23	0.62	0.45	9.1	-4.6	0.76	0.28	9.8	-7.9
18	Baounet	11	0.16	0.64	2.8	-2.5				
19	Rochemelon	11	0.31	0.67	4.3	-2.8				
20	Saint-Sorlin	31	0.86	0.37	13.8	-6.3	0.94	0.21	14.7	-11.0
21	Quirlies	15	0.60	0.54	11.4	-5.2				
22	Mont de Lans	35	0.69	0.64	11.4	-5.4				
23	Girose	60	0.70	0.43	9.1	-4.7				
24	Selle	13	0.79	0.41	9.0	-4.4				
25	Casset	7	0.73	0.47	8.9	-4.6				
26	Blanc	44	0.82	0.29	7.9	-3.9	0.72	0.26	9.2	-7.3
27	Vallon Pilatte	7	0.68	0.56	16.0	-7.2				
28	Rouies	14	0.72	0.68	18.0	-7.8				
29	Sélé	12	0.63	0.61	10.9	-5.1				
30	Pilate	18	0.68	0.83	28.1	-13.1				

133
 134 **Table 1:** List of studied glaciers, characteristics and albedo/mass balance correlations over
 135 2000-2015, except for summer coefficients (over 2000-2010). For localization, refer to Fig. 1.
 136 Highlighted rows exhibit glaciers where annual and summer in situ glacier-wide SMB data are
 137 available. The mask size is expressed in number of pixels. To obtain the glacier mask area in
 138 km^2 , one should multiply the mask size by $0.0625\ km^2$. Determination coefficients are
 139 expressed for each glacier (full plotted results are shown in supplementary material). Note the
 140 units of r^2 (%), RMSE, P_1 and P_2 (m w. e.).

142 **2.3 Surface mass balance data**

143 In the French Alps, six glaciers allow both the summer and annual analyses to be conducted,
144 due to the availability of summer SMB data (b_s) obtained from in situ measurements with the
145 glaciological method (unpublished data, LGGE internal report, listed in Table 1). In addition,
146 glacier-wide annual SMB of the 30 studied glaciers were computed by Rabatel et al. (2016)
147 using the end-of-summer snow line measured on optical remote-sensing images and the
148 glacier-wide mass change quantified from DEMs differencing.

149 For the six glaciers where glacier-wide annual SMB are available from the two methods, i.e.,
150 in situ and satellite measurements, the average of the two estimates was used to calibrate and
151 evaluate the albedo method, in order to derive for each glacier a single relationship SMB vs.
152 computed albedo. We did not discuss here the difference in between the considered datasets
153 because these differences have been investigated by Rabatel et al. (2016).

154 **2.4 In situ albedo measurements**

155 Albedo measurements acquired punctually using an AWS on Saint-Sorlin Glacier have been
156 used to evaluate the MODIS retrieved albedo. In situ albedo measurements were available for
157 three periods in the ablation zone (July-August 2006; June-August 2008; June-September
158 2009) and for one period in the accumulation zone (June-September 2008). Albedo data from
159 these AWS have been calculated as the ratio of the reflected to incident shortwave radiation
160 (0.3 to 2.8 μm) using two Kipp and Zonen pyranometers. With a potential tilt of the instrument
161 with respect to surface melting and the intrinsic sensor accuracy ($\pm 3\%$, Six et al., 2009), the
162 calculated albedo at the AWS shows a $\pm 10\%$ accuracy (Kipp and Zonen, 2009; Dumont et al.,
163 2012).

164 **3 Methods**

165 **3.1 MODImLab products**

166 MODIS L1B images were processed using the MODImLab toolbox (Sirguey, 2009). Image
167 fusion between MOD02QKM bands 1 and 2 at 250 m resolution and MOD02HKM bands 3 to
168 7 at 500 m resolution allows 7 spectral bands at 250 m resolution to be produced (Sirguey et

169 al., 2008). Then, atmospheric and topographic corrections are applied that include multiple
170 reflections due to steep surrounding topography (Sirguey, 2009). Various products are derived
171 from the corrected ground reflectance including snow and ice surface albedo (Dumont et al.,
172 2012). As recommended by Dumont et al. (2012) the WhiteSky (WS) albedo (estimated value
173 of the surface albedo under only diffuse illumination) is considered. The use of an anisotropic
174 reflection model for snow and ice has been preferred to the isotropic case, due to its closer
175 agreement with in situ measurements (Dumont et al., 2012). The MODImLab toolbox also
176 produces sensor geometrical characteristics at the acquisition time such as the solar zenith
177 angle (SZA) and the observation zenith angle (OZA) used for post-processing the images
178 (Sect. 3.4). The MODImLab cloud detection algorithm is more conservative than the original
179 MODIS product (MOD35), and has been preferred as recommended in (Brun et al., 2015).
180 According to Dumont et al. (2012) and further assessed by (Sirguey et al., 2016) the overall
181 accuracy of MODImLab albedo product under clear-sky conditions is estimated at $\pm 10\%$.
182 To mitigate the impact of shadows over the glaciers, MODImLab uses a DEM from the
183 Shuttle Radar Topography Mission (SRTM – 90 m resolution – acquired in 2000) to estimate
184 the sky obstruction by the surrounding topography and to correct the impact of shadows (see
185 Sirguey et al., 2009). The algorithm implemented in MODImLab is fully described in (Sirguey
186 et al., 2009) and inspired from (Dozier et al., 1981; Dozier and Frew, 1990) for the sky
187 obstruction factor processing (Horizon and Vsky in Sirguey et al., 2016), and from (Richter,
188 1998) for correction of shadows. It is first computed at 125 m resolution, providing Boolean-
189 type products of self and cast shadows per pixel. Results are then averaged and aggregated to
190 250 m resolution, producing sub-pixel fraction of shadow (further detailed in Sirguey et al.,
191 2009). Finally, MODIS data processed with MODImLab provides, among others, near-daily
192 maps of white-sky albedo at 250 m resolution together with cloud masks and cast and
193 projected shadows.

194 Albedo maps have been processed for 5,068 images for the Ecrins range, 4,973 for Mont-
195 Blanc and 5,082 for Vanoise over the period 2000-2015. Only images acquired between 09h50

196 and 11h10 AM UTC (+2h in summer for local time conversion) were selected to get minimum
197 SZA and limit projected shadows of surrounding reliefs.

198 **3.2 Glacier masks**

199 Following Dumont et al. (2012) and Brun et al. (2015), we manually created raster masks of
200 the 30 glaciers, based on the glaciers' outlines from 1985-87 (Rabatel et al., 2013) and high
201 spatial resolution (6 m) SPOT-6 images from 2014. All debris-covered areas, together with
202 mixed pixel (rock-snow/ice) have been removed to capture only the snow/ice albedo signal.

203 The resulting number of pixels per glacier is listed in Table 1.

204 **3.3 Surface albedo and glacier-wide mass balance relationship**

205 **3.3.1 Basis of the method**

206 For one glacier in the Alps (Dumont et al., 2012), two in the Himalayas (Brun et al., 2015) and
207 one in the Southern Alps of New Zealand (Sirguey et al., 2016), the summer minimum glacier-
208 wide averaged albedo ($\bar{\alpha}_a^{\min}$) has been significantly correlated to the glacier-wide annual SMB.

209 The relationship between $\bar{\alpha}_a^{\min}$ and glacier-wide SMB results from the fact that solar radiation
210 is the main source of energy for melting snow and ice, both at the surface and within the first
211 centimeters below the surface (Van As, 2011). But this is not sufficient to explain why
212 averaged surface albedo is suitable for monitoring glacier SMB.

213 If we consider a temperate glacier in the mid-latitudes, its surface is fully covered by snow in
214 winter, leading to high and uniform surface albedo ($\bar{\alpha} \approx 0.8$ in Cuffey and Paterson, 2010).
215 During the ablation season, the accumulation area is still covered with snow conversely to the
216 ablation area where the ice is exposed and sometimes covered by debris. The overall albedo of
217 the glacier surface is therefore decreasing over the course of the ablation season, providing
218 information on the ratio of these two areas. The ratio between the size of the accumulation
219 zone and the entire glacier, called the accumulation-area ratio (AAR) has often been used as a
220 predictor of SMB both qualitatively (LaChapelle, 1962; Meier and Post, 1962; Mercer, 1961)
221 or quantitatively (Dyurgerov et al., 2009). Therefore, assessing $\bar{\alpha}_a^{\min}$ provides insights of the

222 relative share between the exposed ice and the snow-covered areas at the end of the ablation
223 season, also quantified by the AAR.

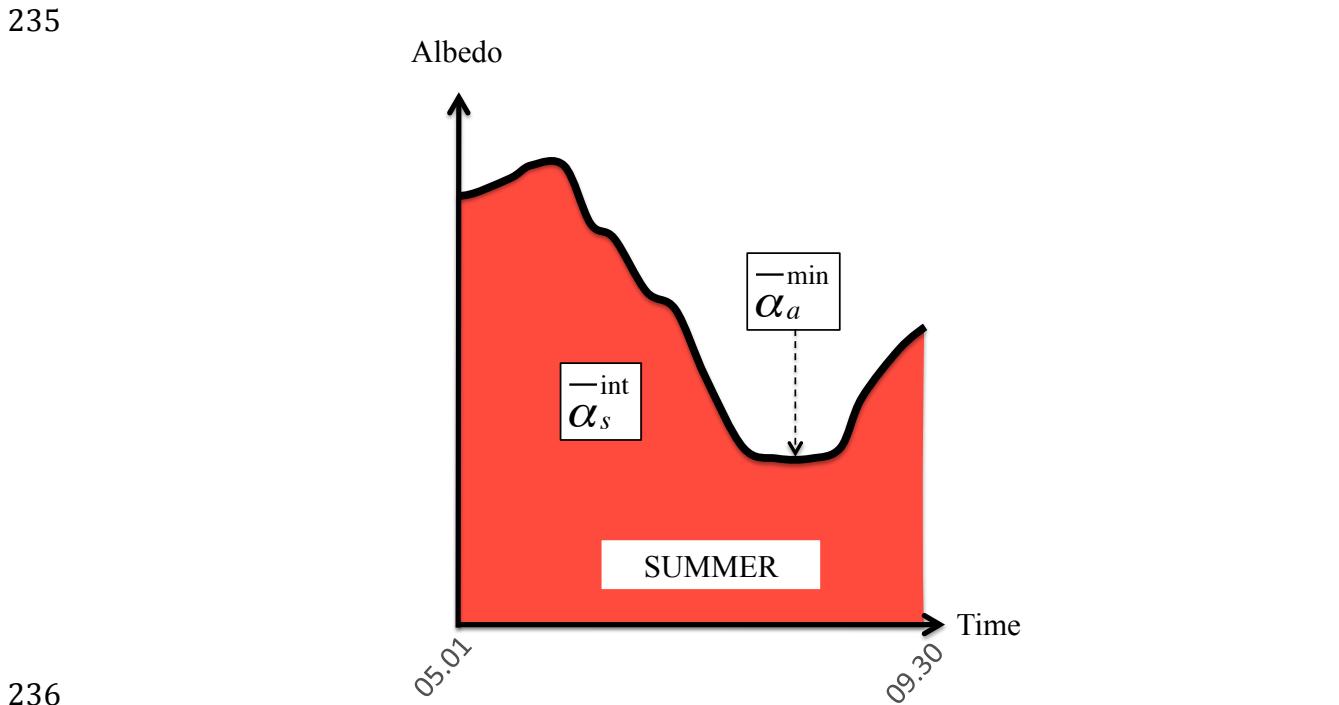
224 **3.3.2 From annual to summer surface mass balances**

225 In this study, $\bar{\alpha}_a^{\min}$ has been computed for the 30 glaciers in order to validate the method at a
226 regional scale. Only the $\bar{\alpha}_a^{\min}$ occurring in summer have been considered because minimum
227 values out of the summer period are artifacts. Then, $\bar{\alpha}_a^{\min}$ has been directly correlated to
228 available annual SMB data (listed in Table 1).

229 Following the work by Sirguey et al. (2016) on Brewster Glacier, a similar approach has been
230 used in order to validate the method at a summer scale but only on six glaciers (within our
231 sample of 30) for which the summer SMB are available. Conversely to Sirguey et al. (2016),
232 the summer SMB b_s has been compared to the integrated albedo signal $\bar{\alpha}_s^{\text{int}}$ during the entire
233 ablation season (1st May to 30th September) computed as follow and illustrated in Fig. 2.

234

$$\bar{\alpha}_s^{\text{int}} = \int_{05.01}^{09.30} \bar{\alpha}(t) dt \quad \text{Eq. (1)}$$



237

238 **Figure 2:** Schematic of a typical albedo cycle over one summer, displaying parameters which
239 have been linked to annual and summer (between 1st May and 30th September in the northern

240 hemisphere) SMB. $\bar{\alpha}_s^{\text{int}}$ is retrieved using Eq. (1). The summer minimum value of albedo is
241 represented by $\bar{\alpha}_a^{\text{min}}$.

242 Integrated summer albedos allow to account for snowfall events that can occur during the
243 ablation period (punctual high albedos). As an example, a strong summer snowfall event
244 leading to a rather persistent snow coverage of the glacier will 'feed' the integrated albedo, and
245 physically reduces the glacier melting, which has an impact on the SMB (Oerlemans and
246 Klok, 2004). The method therefore accounts for snowfall events to retrieve the glacier summer
247 SMB. To compare each year together and remove the impact of the variable integration time
248 period for each glacier, $\bar{\alpha}_s^{\text{int}}$ has been divided by the number of integrated days.

249 **3.4 Data filtering**

250 MODIS offers the opportunity to get daily images, but retrieving daily maps of Earth surface
251 albedo remains challenging. Indeed, various sources of error require filtering the available
252 images in order to only capture physical changes of the observed surface and not artifacts.
253 Clouds are known to be a major problem in optical remote sensing of the Earth surface
254 especially in the case of ice and snow covered surfaces. Even if some algorithms exist to
255 differentiate clouds and snow-covered areas (e.g., Ackerman et al., 1998; Sirguey et al., 2009),
256 omission errors are difficult to avoid, leading to erroneous albedo of the surface.

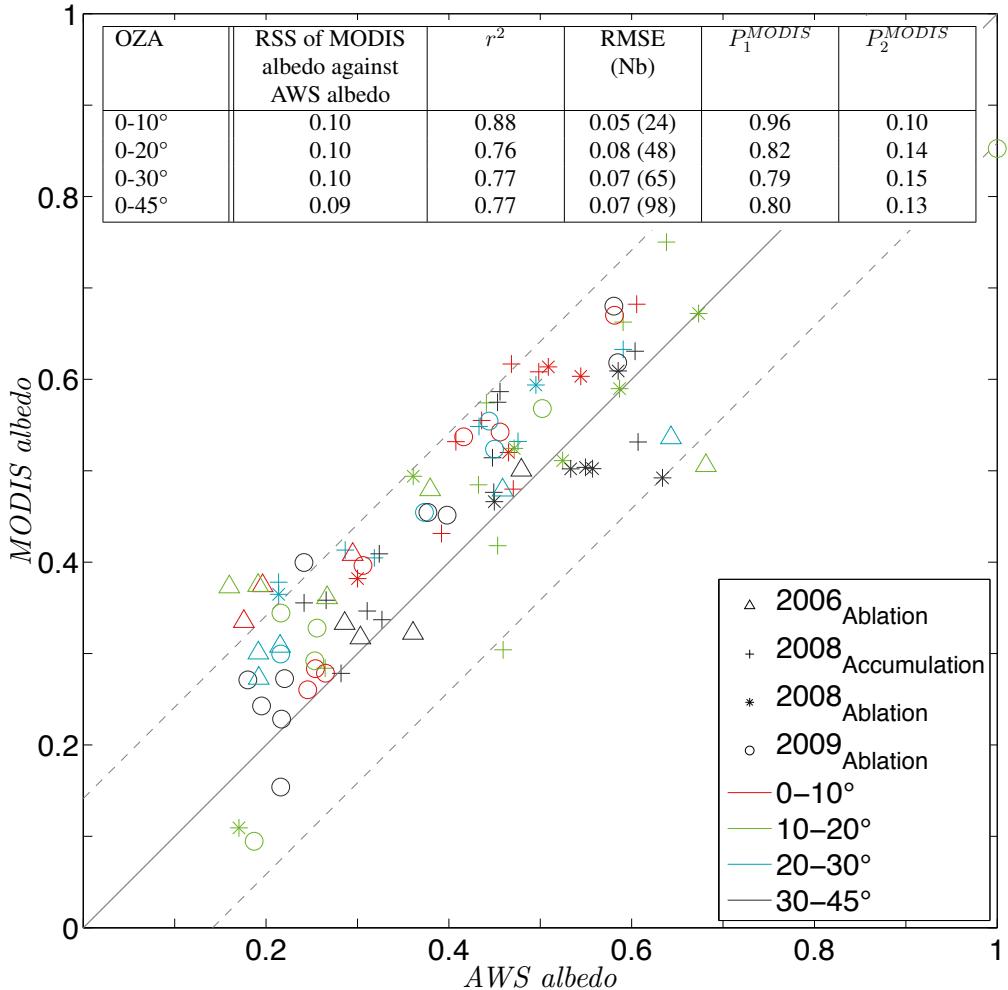
257 In this study, all images with a presence of cloud greater than 30% of the total glacier surface
258 area have been discarded. This threshold is higher than that chosen in Brun et al. (2015) on the
259 Chhota Shigri Glacier (20%), and we thus discuss Sect. 5.1 the impact of the computed cloud
260 threshold on the derived albedo results. When determining $\bar{\alpha}_a^{\text{min}}$, 0% of cloud cover has been
261 imposed as a condition and visual check for each year and each glacier has been performed.
262 Snapshots from the fusion of MODIS bands 1 to 3 and from bands 4 to 6 (Sirguey et al., 2009)
263 have been used to visually check the images, together with images from other satellites
264 (mostly from the Landsat archive) and pictures and comments from mountaineering forums.

265 This last step, although laborious meticulous when studying 30 glaciers allowed the
266 identification of the summer minimum to be improved. Visual check of the images also
267 confirms that projected shadows of clouds are not affecting the albedo map. Another source of
268 error is the impact of the OZA. As mentioned in Sirguey et al. (2016), accuracy of the MODIS
269 retrieved albedo strongly decreases for viewing angles above 45° as pixel size increases from
270 2 to 5-folds from OZA = 45° to 66° (Wolfe et al., 1998). This phenomenon is accentuated
271 when observing steep-sided snow/ice surfaces, surrounded by contrasted surfaces (rocks,
272 forests, lakes...). This distortion could lead to capture the mean albedo of a glacier plus its
273 surroundings. Following this, we decided to filter the images according to their OZA angle, as
274 further described Sect. 4.1.

275 **4 Results**

276 **4.1 Retrieved albedo assessment**

277 A quantitative evaluation of the retrieved albedo has been performed with AWS deployed on
278 Saint-Sorlin Glacier. Measurements have been synchronized between punctual albedo for
279 MODIS and a 2-hour averaged albedo around MODIS acquisition time for the AWS. It is
280 worth reminding some differences between the in situ measured albedo data and the one
281 retrieved using MODIS. The downward facing pyranometer stands at around 1 m above the
282 surface, corresponding to a monitored footprint of *ca.* 300 m² (theoretical value for a flat
283 terrain) while the pixel area of MODImLab products matches 62,500 m². Quantified albedos
284 from each method are therefore not representative of the same area. On the other hand,
285 incoming radiation data are extremely sensitive to a tilt of the sensor located on the AWS and
286 maintaining a constant angle throughout the monitoring period remains challenging, especially
287 during the ablation season. For instance, a tilt of 5° of the pyranometer at the summer solstice
288 can increase by 5% the error on the irradiance measurement (Bogren et al., 2016). No sensor
289 tilt was deployed on the AWS, thus preventing the application of tilt-correction methods (e.g.,
290 Wang et al., 2016). Nonetheless, regular visit allowed to maintain the sensor horizontal and to
291 limit errors in the irradiance measurements.



294 **Figure 3:** MODIS albedo and AWS albedo data for different OZA classes on Saint-Sorlin
 295 Glacier. Years indicated in the caption correspond to the year of acquisition while subscripts
 296 express the AWS location in the accumulation or ablation areas. The mean discrepancy
 297 between MODIS and AWS albedo per OZA is quantified by the RSS (residual sum of square).
 298 Correlation coefficient per OZA classes are also provided, with r^2 , RMSE together with the
 299 number of compared measurements (Nb), and coefficients of the equation:
 300 $MODIS_{albedo} = P_1^{MODIS} AWS_{albedo} + P_2^{MODIS}$. The continuous grey line illustrates the 1:1
 301 relationship between AWS and MODIS retrieved albedo. Thin and dotted lines represent the
 302 combined uncertainties on both AWS and MODIS retrieved albedo (absolute value of 10% for
 303 each), only accounting for intrinsic sensor accuracy and not for errors related to the acquisition
 304 context, e.g. size of the footprint.

305

306 Figure 3 illustrates the comparison between the retrieved and measured albedos at the AWS
307 locations for various OZA classes. One can note minor differences between the data plotted in
308 Fig. 3 and those presented in Dumont et al. (2012, Fig. 2). These differences are related to
309 changes in the MODImLab algorithm and different computation of the in situ albedo,
310 integrated over a two-hour period in the current study.

311 In Fig. 3, the spread between MODIS and AWS albedos is higher for low albedos (i.e. ablation
312 area). This is related to the footprint difference as described earlier, accentuating the albedo
313 differences when monitoring heterogeneous surface (snow patches, melt ponds...), even
314 more pronounced in summer. One can also note that MODIS albedo often over-estimate the
315 AWS albedo value. This over-estimation could be explained by: (1) the MODImLab albedo
316 retrieval algorithm. Indeed, under-estimation of the incoming radiation computed in the
317 MODImLab algorithm would lead to over-estimated retrieved albedo values, in addition the
318 atmospheric corrections used to compute the incident radiation could be hypothesized as
319 source of error (e.g. modeled transmittance through a simplified computed atmosphere, refer
320 to (Sirguey et al., 2009) for further description); (2) the AWS albedo measurements. Indeed,
321 view angles of AWS pyranometers (170°) could influence the retrieved albedo by monitoring
322 out-of-glacier features (e.g. moraines, rock walls, ...), resulting in under-estimated albedo
323 values. However, it is worth noting that most of the points are within the combined uncertainty
324 of both sensors and these differences in albedo retrieved from MODIS and the AWS are thus
325 hard to interpret.

326 Finally, Fig. 3 shows substantial differences between $OZA < 10^\circ$ and other OZA classes. For
327 $OZA < 10^\circ$, MODIS albedos better agree with AWS albedos than for the three other classes.
328 Integrating MODIS images with $OZA > 10^\circ$ substantially deteriorate the agreement with AWS
329 albedos (in term of r^2 , RMSE and the slope P_I^{MODIS}), especially on "narrow" targets as alpine
330 mountain glaciers. We therefore chose to prioritize images acquired with low OZA to avoid

331 detection of non-glacierized surfaces. Therefore, four classes of images have been selected
 332 following the criteria presented in Table 2.

Class	OZA (°)	Criteria
I	$OZA \leq 10$	All retained
II	$10 < OZA \leq 20$	Retained if more than 7 days between consecutive images from class I
III	$20 < OZA \leq 30$	Retained if more than 7 days between consecutive images from class I+II
IV	$OZA > 30$	Not retained

333

Table 2: Filtering the images from OZA values.

335
 336 For the rest of the computation, the absolute $\pm 10\%$ accuracy per pixel estimated in Dumont et
 337 al. (2012) has been considered. We determined the uncertainty on $\bar{\alpha}$ by accounting for the
 338 spatial variability of the albedo signal within the glacier and considering that our sets of pixels
 339 are independent from each other (Eq. (2)):

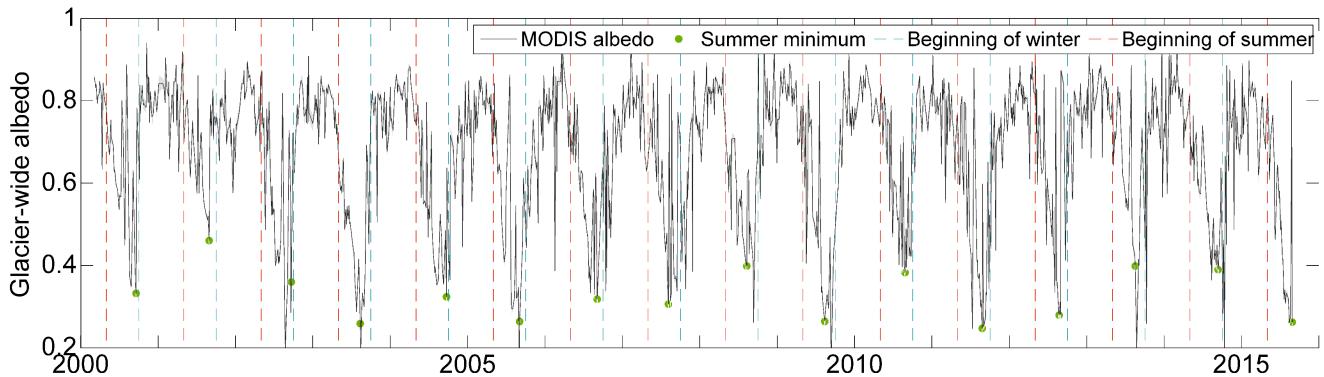
340
$$\sigma_{\bar{\alpha}} = \frac{\sigma}{\sqrt{N}} \quad \text{Eq. (2)}$$

341 where σ stands for the standard deviation of the pixels albedo with N the number of pixels.

342 **4.2 Temporal variability of the albedo signal**

343 Using the "step-by-step" filtering procedure explained in Sect. 3.4, the ~ 16 -yr albedo cycle of
 344 each of the 30 glaciers was obtained (results available in the supplementary material). Figure 4
 345 illustrates the entire albedo time series for Saint-Sorlin Glacier over the period 2000-2015. We
 346 observed that the albedo decreases from the beginning of summer (dashed red line), reaching
 347 $\bar{\alpha}_a^{\min}$ in August/September and rising again at the end of September. This cyclicity is a proxy
 348 of surface processes. The snow cover decreases at the beginning of summer until reaching its

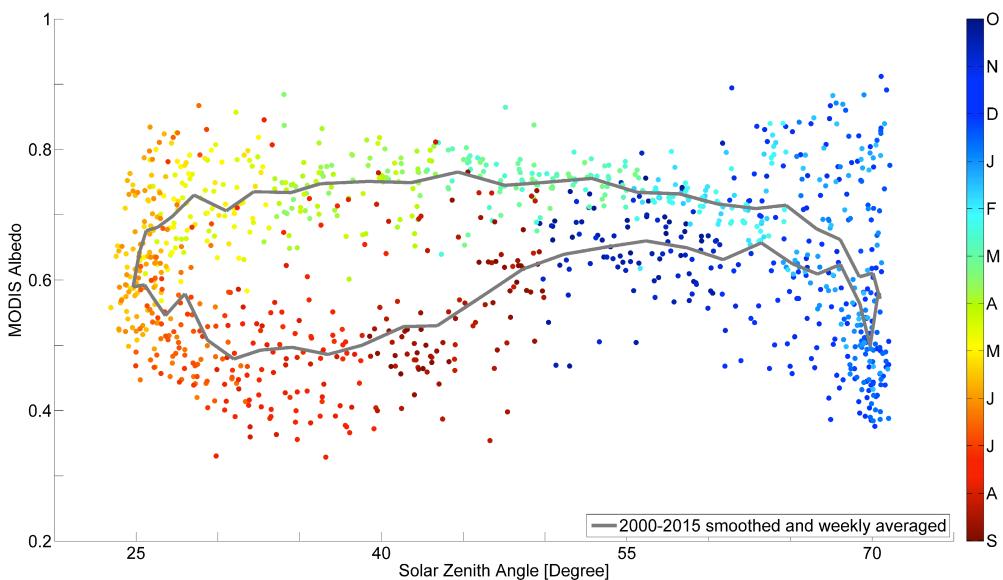
349 lowest extent, and finally increases again with the first snowfall in late summer to reach its
 350 maximum extent in winter/spring.



352 **Figure 4:** ~16-yr albedo course for Saint-Sorlin Glacier. Glacier-wide averaged albedo is
 353 represented with the continuous black line. The green dots spot for each summer the minimum
 354 average albedo, and have been manually checked for all years and glaciers. Dashed red and
 355 blue lines stand for the beginning of the defined ablation and accumulation seasons (May and
 356 October 1st respectively).

357
 358 The periodicity of the albedo signal is however not so well defined for some of the studied
 359 glaciers. For instance, Argentière Glacier exhibits a severe drop of $\bar{\alpha}$ in winter, reaching
 360 values as low as summer minimum ($\bar{\alpha} \approx 0.4$). The observed drop of albedo in winter occurs
 361 during more than one month centered on the winter solstice (December 21st) and is observed
 362 for nine glaciers (Argentière, Baounet, Casset, Blanc, Girose, Pilatte, Vallon Pilatte, Tour and
 363 Séle glaciers, refers to supplementary material for full results). These glaciers are located
 364 within the three studied mountain ranges but have the common characteristic to be very
 365 incised with steep and high surrounding faces. We studied the albedo series as a function of
 366 the SZA to reveal possible shadowing on the observed surfaces. Figure 5 displays the same
 367 cycle as Fig. 4 for Argentière Glacier but providing information about SZA. As a reminder,
 368 the MODImLab white-sky albedo is independent of the illumination geometry but the
 369 computed albedo for each pixel can be subject to shadowing from the surrounding topography.

370 Two main observations stand out from the winter part of the cycle in Fig. 5: (i) most of
 371 MODIS $\bar{\alpha}$ severely decrease under $\bar{\alpha} = 0.6$ for SZA greater than 60° corresponding to
 372 November to January images, (ii) these drops are not systematic and we rather observe a
 373 dispersion cone than a well-defined bias. As there are no physical meanings to systematic
 374 change of the surface albedo during a part of the winter period and owing to the fact that this
 375 dispersion is only observed for topographically incised glaciers, these decreases in albedo have
 376 been considered as artifacts. These observations led us to perform a sensitivity study on the
 377 validity of the shadow mask produced by MODImLab, and to study the impact of these
 378 shadows on the retrieved glacier-wide albedo (see 5.2).



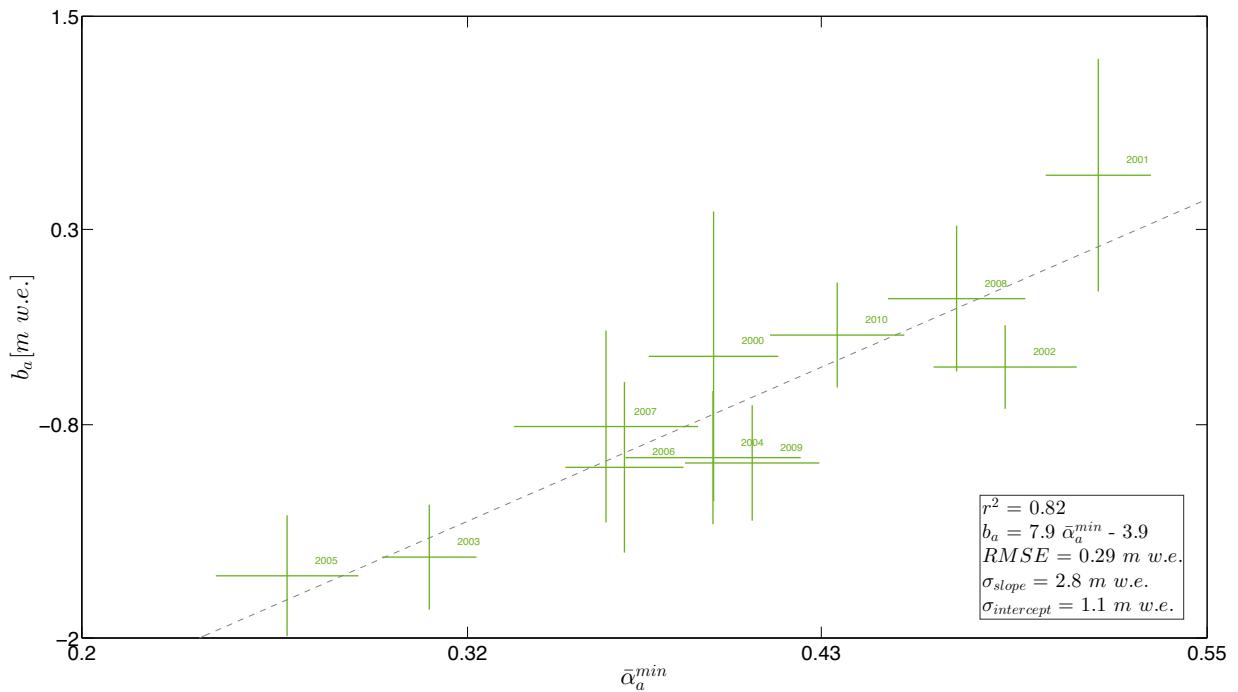
379
 380 **Figure 5:** Albedo cycle for Argentière Glacier as a function of the SZA. Each point
 381 corresponds to glacier-wide averaged albedo for each available image. The 16 years are
 382 displayed. Color scale gives indication on the date of the used image. The thick grey line
 383 describes the weekly albedo averaged over the entire study period. For readability purpose, the
 384 averaged albedo has been smoothed, using a 7 points running average.

385

386 **4.3 Albedo and glacier-wide surface mass balance**

387 **4.3.1 $\bar{\alpha}_a^{\min}$ and annual surface mass balance**

388 The summer minimum average albedo for each year and each glacier has been linearly
 389 correlated to the glacier-wide annual SMB. Figure 6 illustrates the relationship between $\bar{\alpha}_a^{\min}$
 390 and b_a for Blanc Glacier (all the other glaciers are shown in the supplementary material). Error
 391 bars result from the dispersion of the SMB dataset for each year, and from the glacier intrinsic
 392 variability of the albedo signal the day of $\bar{\alpha}_a^{\min}$ acquisition. For the glaciers where the glacier-
 393 wide annual SMB is available from the SLA method, the uncertainty is about ± 0.22 m w.e. on
 394 average (ranging from 0.19 to 0.40 m w.e. depending on the glacier, Rabatel et al., 2016).



396 **Figure 6:** Annual SMB as a function of the MODIS retrieved summer minimum glacier-wide
 397 average albedo for Blanc Glacier. Error bars result on the dispersion of the available annual
 398 SMB data and on the quadratic sum of the systematic errors made on each albedo
 399 measurement. The thin dashed grey line illustrates the line of best fit, along with regression
 400 coefficients and significance.

401
 402 Twenty-seven glaciers show significant correlations (refer to Table 1 for full results) if
 403 considering a risk of error of 5% (according to a Student's t test) and confirms the robust
 404 correlation between $\bar{\alpha}_a^{\min}$ and b_a . However, the linear correlation has no statistical significance

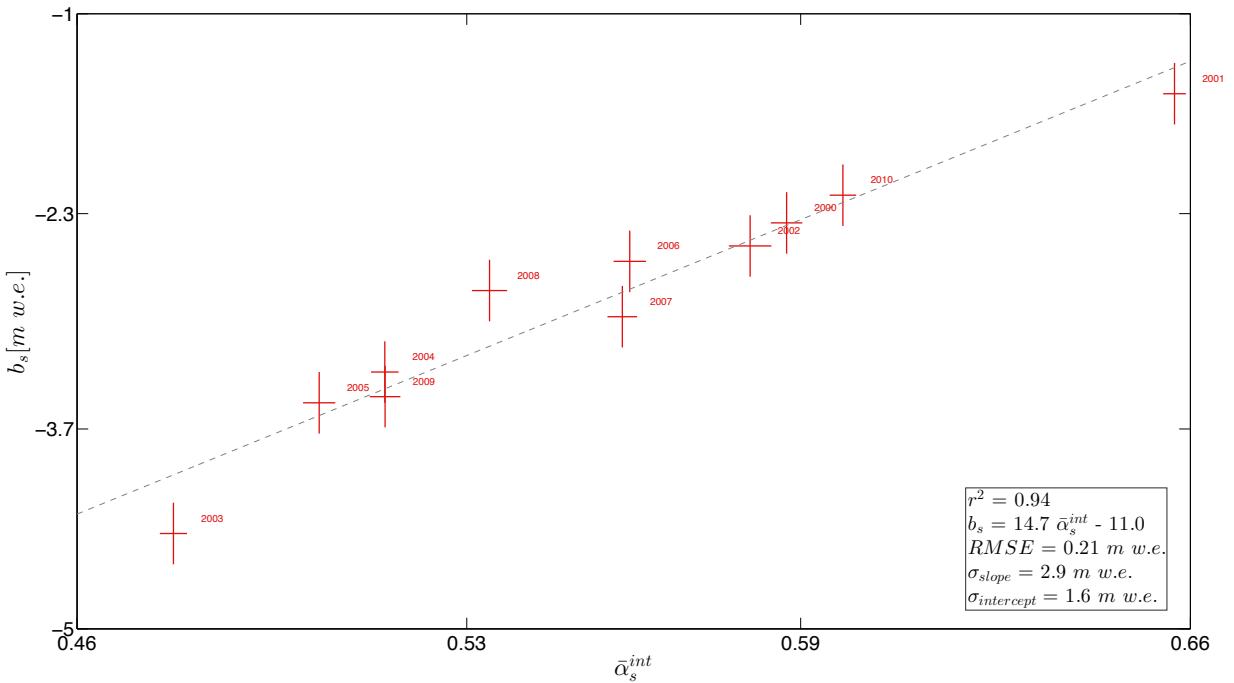
405 for three glaciers with $r^2 < 0.25$. A possible explanation is the high number of removed images
406 in summer due to manually checked thin overlying clouds not detected by the MODImLab
407 cloud algorithm.

408 Looking at the 27 glaciers for which significant relationships have been found, 2001 is
409 regularly identified as an outlier. According to existing SMB datasets, 2001 is the only year of
410 the period 2000-2015 for which the annual SMB has been positive for all the studied glaciers
411 ($+0.80 \text{ m w.e. yr}^{-1}$ in average).

412 To predict correctly the surface mass balance values for the year 2001 using the albedo
413 method, monitored minimum glacier-wide average albedo would need to be extremely high
414 (often greater than 0.7, i.e. 0.83 and 0.95 for Rochemelon and Vallonnet glaciers,
415 respectively), to match the regression line derived from other years of the time series (Table
416 1). Taking into consideration snow metamorphism during the summer period, melting at the
417 surface and possible deposition of debris or dusts, monitoring such high albedo values
418 averaged at the glacier scale is unrealistic. As removing 2001 from the time series does not
419 increase the number of glaciers for which the correlation is significant, 2001 has been
420 conserved in the time series. However, this observation reveals a limitation of the albedo
421 method by under-estimating the annual SMB value for years with very positive annual SMB.

422 **4.3.2 $\bar{\alpha}_s^{\text{int}}$ and summer surface mass balance**

423 Studying the integral of the albedo signal during the ablation season can provide insights on
424 the intensity of the ablation season and thus on the summer SMB b_s . As described in Sect.
425 3.3.2, $\bar{\alpha}_s^{\text{int}}$ has been computed and connected to the in situ b_s . Figure 7 illustrates the results
426 for Saint-Sorlin Glacier.



443 $\bar{\alpha}_s^{\text{int}}$ has been significantly correlated to b_s and is therefore a reliable proxy to record the
444 ablation season.

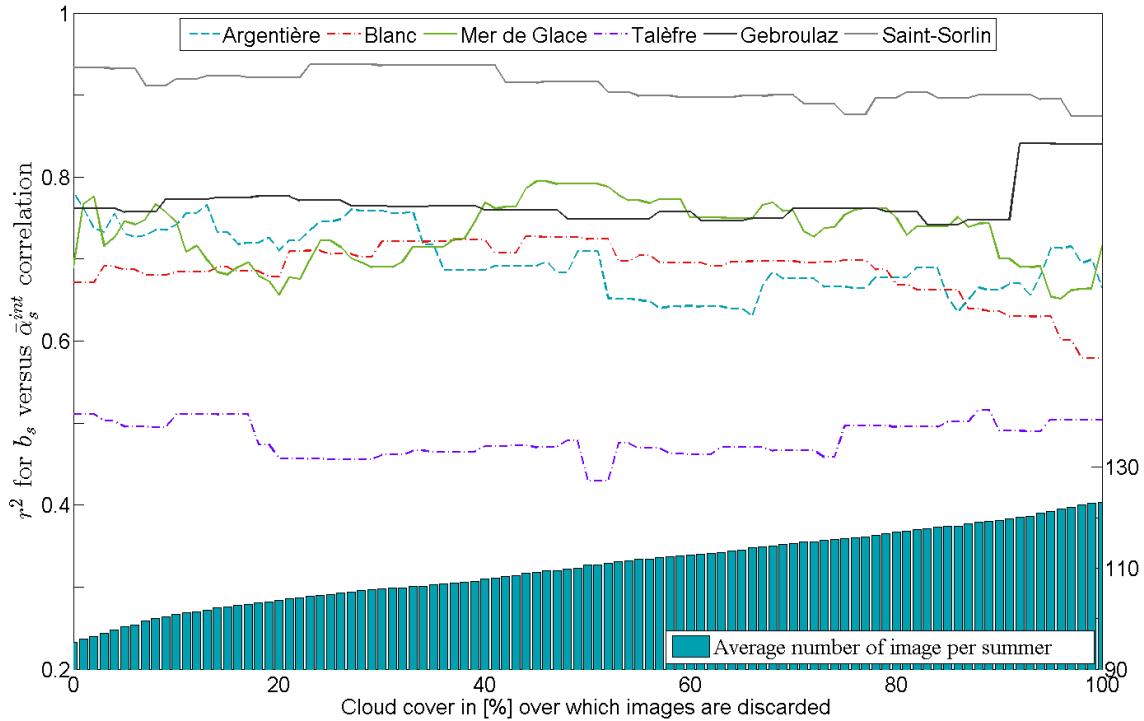
445

446 **5 Discussion**

447 In this section, we first discuss the impact of the threshold applied to the cloud cover fraction
448 on the obtained results. Then, a sensitivity study focused on the algorithm correcting the
449 shadows is presented. We finally express the main limitations and assessments of the albedo
450 method.

451 **5.1 Cloud coverage threshold**

452 As stated in Sect. 3.4, a value of 30% of cloud coverage over the glacier mask has been
453 defined as the acceptable maximum value for considering the albedo map of the day. We
454 computed a sensitivity study on the impact of this threshold on the value of the obtained
455 correlations between the integrated summer albedo and the in situ summer SMB. The summer
456 period has been chosen as it represents the period when the albedo of the glacier is the most
457 contrasted, between bare ice and snow/firn. The glacier-wide average albedo in this period is
458 therefore more sensitive to possible shading of a part of the glacier. Figure 8 illustrates the
459 results for the six seasonally surveyed glaciers. The used value of the allowed cloud coverage
460 appears not to have a substantial impact on the correlation. This observation first implies that
461 the MODImLab cloud product is reliable enough to only compute surface albedo and to avoid
462 too frequent misclassification between the clouds and the surface. It also suggests that
463 removing too many images because of partial cloud cover removes information about the
464 glacier-wide average albedo variability. However, allowing all images, even when the glacier-
465 wide average albedo is computed on only 10% of the glacier (90% of detected cloud
466 coverage), does not reduce significantly the correlation for most of the six glaciers.



467
468 **Figure 8:** r^2 for the six seasonally surveyed glaciers for the albedo summer integral *versus*
469 summer SMB relationship against the cloud threshold above which images have been
470 discarded during the summer season. For the computation, hundred thresholds have been
471 tested between 0 and 100%. The inner histogram illustrates the number of considered images
472 per summer and averaged on the six glaciers.

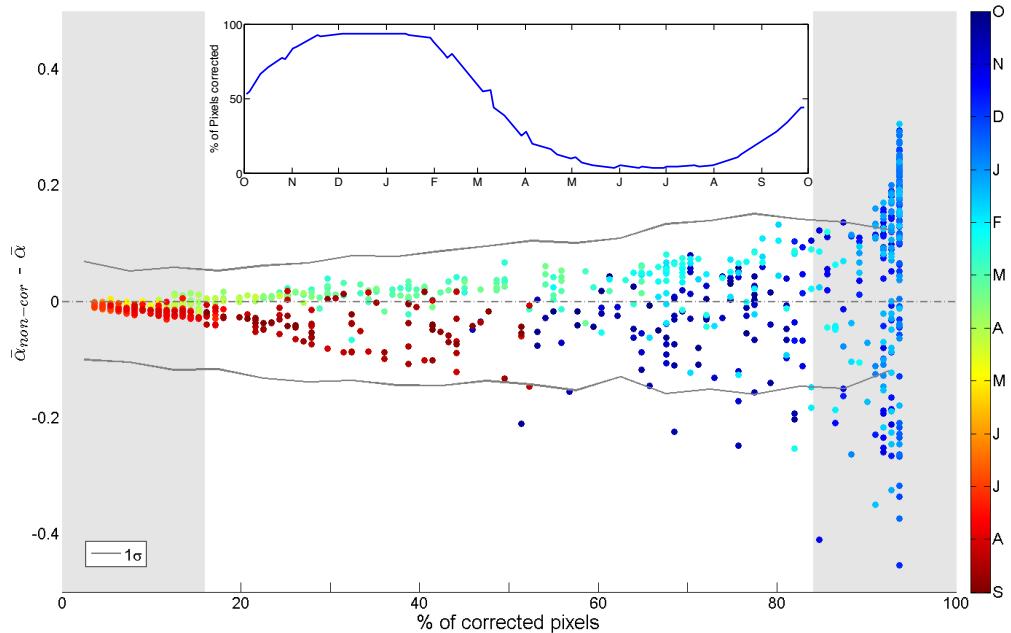
473
474 Nevertheless, hypothesizing that the glacier-wide average albedo of a small fraction of the
475 glacier (e.g., 10%) is suitable to represent the entire glacierized surface is questionable. It
476 therefore depends on the size of the observed glacier, where 10% of a glacier of 3 and 30 km²
477 have not the same meaning, but also on the delineated mask (ablation area not entirely
478 considered because of debris coverage...). The summer-integrated albedo is also highly
479 dependent on the time gap between useful images. In other words, if an image has an
480 "anomalous" glacier-wide average albedo because of high cloud coverage, the impact on the
481 integrated value will be smaller if "normal condition" albedos are monitored at nearby dates.

482 The average number of available images per year does not largely differ between the various
483 computed cloud coverage thresholds. It varies in average from 95 to 123 images per summer
484 period for respectively 0% and 100% cloud coverage threshold. Intermediate values are 106,
485 111 and 116 images per summer for 30, 50 and 75% cloud coverage threshold, respectively.
486 The difference in significance of r^2 (according to a Student's t test) between opting for 0% and
487 100% is almost negligible, and choosing the best cloud threshold value is rather a compromise
488 between the number of used images and the resulting correlation with glacier-wide SMB. We
489 finally concluded that selecting cloud coverage threshold to 30% presents the best
490 determination coefficients between the integrated summer albedo and the summer balance for
491 most of the six glaciers without losing too much temporal resolution.

492 **5.2 Assessment of the impact of shadows on retrieved albedos**

493 In light of the documented dispersion on $\bar{\alpha}$ during some of the winter months on several
494 studied glaciers (Sect. 4.2), sensitivity of the MODIS retrieved albedo against correction of
495 shadows had been assessed. This work has only been conducted on the 250 m resolution raster
496 products and specifically on the cast shadow product because self-shadow corrections can be
497 considered as reliable enough because only related to the DEM accuracy. We thus defined a
498 pixel as “corrected” when at least one of its sub-pixels was classified as shadowed. From then
499 on, two glacier-wide albedos $\bar{\alpha}$ have been defined: (i) $\bar{\alpha}_{\text{non-cor}}$ computed on non-corrected
500 pixels only, classified as non-shadowed; (ii) $\bar{\alpha}$ of both corrected and non-corrected pixels,
501 equal to the glacier-wide average albedo. Figure 9 illustrates the difference between $\bar{\alpha}_{\text{non-cor}}$
502 and $\bar{\alpha}$ as a function of the percentage of corrected pixels over the entire glacier. The study
503 was performed on Argentière Glacier (111 pixels) that exhibited large $\bar{\alpha}$ artifacts in winter
504 (Fig. 5). The inner diagram allows emphasizing the annual “cycle” of modeled shadows,
505 contrasted between nearly no cast shadows in summer and an almost fully shadowed surface
506 in winter. We represent the 1 standard deviation of $\bar{\alpha}$, averaged by classes of 5% corrected
507 pixels. In other words, it illustrates the mean variability of the glacier-wide surface albedo.

508 Therefore, for images with $\bar{\alpha}_{\text{non-cor}} - \bar{\alpha}$ within the interval defined by 1 st.dev. of $\bar{\alpha}$, errors
 509 resulting from the correction algorithm are smaller than the spatial variability of the glacier-
 510 wide albedo glacier. We also selected only significant values, following a normal distribution
 511 of the averaged $\bar{\alpha}$. Consequently, only values at $\pm 1\sigma$ (68.2%) in term of percentage of
 512 corrected pixel have been retained (i.e. when the relative share of corrected pixels ranged from
 513 15.9 to 84.1%). Between 0 and 15.9%, $\bar{\alpha}_{\text{non-cor}}$ and $\bar{\alpha}$ are not sufficiently independent
 514 because of low number of corrected pixels, and beyond 84.1%, $\bar{\alpha}_{\text{non-cor}}$ is computed over a too
 515 small number of pixels. As a consequence, even if the albedo correction in the shadowed parts
 516 of the glacier could be improved, most of the errors related to this correction do not depreciate
 517 the results. Above 80% of corrected pixels (December to early February), differences between
 518 $\bar{\alpha}_{\text{non-cor}}$ and $\bar{\alpha}$ exceed the monitored spatial variability of $\bar{\alpha}$. These anomalies are at the root
 519 of the observed artifacts Fig. 5 by the severe drops of albedos and described Sect. 4.2.



520
 521 **Figure 9:** Impact of the ratio of corrected pixels toward the difference between non-corrected
 522 and glacier-wide albedo. Each point corresponds to one acquisition and the 16 years are
 523 therefore displayed on this graph. Color scale gives indication on the date of the acquired
 524 image. Grey shaded areas correspond to ratios of corrected pixels for which $\bar{\alpha}_{\text{non-cor}} - \bar{\alpha}$ has

525 low statistical robustness (refer to the main text). Thin grey lines represent 1σ standard
526 deviation of $\bar{\alpha}$, averaged by classes of 5% corrected pixels. The inner graph illustrates the
527 amount of corrected pixel, function of the selected month.

528

529 In addition, a seasonality in the albedo signal can be observed with $\bar{\alpha}_{\text{non-cor}} - \bar{\alpha} > 0$ in early
530 spring (February to April) while $\bar{\alpha}_{\text{non-cor}} - \bar{\alpha} < 0$ in summer and autumn (June to November).
531 This could be explained by different localizations of shadowed area for a given ratio of
532 corrected pixel. As an example, a glacier could have in October a snow- and shadow-free
533 snout and a covered by fresh snow and shadowed upper section. This configuration would
534 induce a negative difference as we observe from June to November. Conversely, this glacier
535 could present in March (same ratio of corrected pixels than October) a complete snow
536 coverage, leading to a smaller difference between $\bar{\alpha}_{\text{non-cor}}$ and $\bar{\alpha}$ (<0.1) that could even result
537 on positive difference as we observe from February to April.

538 Finally, observed albedo artifacts in winter are most likely due to the correction of shadows.
539 On the other hand, correcting shadows accurately and consistently is extremely challenging.
540 As illustrated by Fig. 9, a way to confidently consider the albedo signal is to exclude values
541 with too large share of corrected pixels. However, because of the inter-annual approach carried
542 out in this study, such systematic artifact is not depreciating the results but would be a major
543 issue on studies focused on albedo values themselves (e.g. maps of snow extents...).

544 **5.3 Limits of the albedo method**

545 In agreement with Dumont et al. (2012) and Brun et al. (2015), retrieving the glacier annual
546 SMB from albedo summer minimums proves to be an efficient method. Low correlations often
547 result from high and persistent cloud coverage during summer, reducing the chance of spotting
548 the albedo summer minimum. For SMB reconstruction purpose, a future line of research could
549 rest upon linking morpho-topographic features of the glacier such as glacier surface area,
550 mean altitude or slope to the regression coefficients of both annual and seasonal SMB vs.

551 albedo relationships, giving the opportunity to establish analogy between monitored and
 552 unmonitored glaciers. Tests have been carried out but no significant and satisfying results have
 553 been obtained, due to a presumably too heterogeneous data set, where large glaciers ($>10 \text{ km}^2$)
 554 and/or south-facing glaciers are largely under-represented. Larger scale studies and multi-
 555 variable correlations in between morpho-topographic features could be for instance envisaged.
 556 Rabaté et al., (2017) recently proposed an alternative approach to reconstruct the annual mass
 557 balance of unmonitored glacier on the basis of the albedo method. This approach relies on the
 558 ELA-method (Rabaté et al., 2005), but using the remotely sensed monitored $\bar{\alpha}_a^{\min}$ together
 559 with the accumulation area ration (AAR), the glacier hypsometry, and the regional SMB-
 560 elevation gradient (which is the annual SMB gradient in the vicinity of the glacier ELA). For
 561 an exhaustive description of this approach, refer to Rabaté et al., (2017).

562 Using the albedo method for the summer period has shown promising results, with significant
 563 correlations found for the six seasonally monitored glaciers. There is still in this approach a
 564 step to retrieve the summer SMB of an unmonitored glacier with high confidence.

565 The winter period has also been considered in the framework of this study, but has not been
 566 presented in the main body of this publication because of underwhelming results. The albedo
 567 signal between 1st October and 30th April has been computed similarly to Sirguey et al. (2016)
 568 by integrating the winter albedo signal, only when exceeding a certain threshold $\bar{\alpha}_T$, as
 569 described by Eq. (3):

$$570 \bar{\alpha}_w^{\text{int}} = \int \bar{\alpha}(t) \begin{cases} \text{if } \bar{\alpha}(t) \text{ is found between 10.01 and 04.30} \\ \text{Only if } \bar{\alpha}(t) \geq \bar{\alpha}_T \end{cases} \quad \text{Eq. (3)}$$

571 According to Sirguey et al., 2016, the use of $\bar{\alpha}_T$ allows to detect all snowfall events on the
 572 glacier, by monitoring abrupt rises of $\bar{\alpha}$. One of the main conclusions of the latter study was
 573 the ability of the computed $\bar{\alpha}_w^{\text{int}}$ to monitor the frequency of snowfall events, themselves proxy
 574 of the accumulation of snow on the glacier, known to be one of the main component of the
 575 winter SMB.

576 $\bar{\alpha}_T$ has been chosen to maximize the correlation between the retrieved cumulative winter
 577 albedo $\bar{\alpha}_w^{\text{int}}$ and the winter SMB. Threshold values have been computed independently for each
 578 of the six seasonally monitored glaciers. To evaluate the impact of this threshold, $\bar{\alpha}_w^{\text{int}}$ has also
 579 been computed without threshold over winter months (equivalent to $\bar{\alpha}_T = 0$). Table 3 gathers
 580 all the coefficients obtained from the relationship $\bar{\alpha}_w^{\text{int}} \text{ vs. } b_w$, with and without the use of an
 581 albedo threshold $\bar{\alpha}_T$.

Glacier	$\bar{\alpha}_T$	r^2 using $\bar{\alpha}_T$	r^2 without $\bar{\alpha}_T$
Saint-Sorlin	0.76	0.75	0.21
Argentière	0.58	0.88	0.76
Talèfre	0.68	0.59	0.25
Mer de Glace	0.53	0.90	0.87
Gebroulaz	0.75	0.36	0.25
Blanc	0.70	0.33	0.21

582 **Table 3:** Coefficients of determination for the relationship between the winter SMB b_w and the
 583 integrated winter albedo, computed with and without the albedo threshold $\bar{\alpha}_T$.

585 For Argentière and Mer de Glace glaciers, a significant correlation is found whatever the value
 586 of the albedo threshold $\bar{\alpha}_T$. For the four other glaciers, using $\bar{\alpha}_T$ largely improves the
 587 correlation. However, $\bar{\alpha}_T$ is far from being uniform on the six glaciers ($0.53 \geq \bar{\alpha}_T \geq 0.76$). In
 588 addition, for most of the considered glaciers, correlation coefficients abruptly deteriorate when
 589 changing this threshold, which does not allow using a "regional" threshold for all considered
 590 glaciers. On the other hand, Argentière and Mer de Glace without the use of $\bar{\alpha}_T$ provide the
 591 best correlation coefficients compared to the other four glaciers; it is noteworthy that they are
 592 by far the largest glaciers of our monitoring set (14.59 and 23.45 km² for Argentière and Mer
 593 de Glace glaciers, respectively). With a glacier snout reaching 1600 m a.s.l., the tongue of
 594 these glaciers can experience melting events (resulting in contrasted pixels in terms of albedo
 595 value), even during the winter season. Another difference between our study and Sirguey et al.
 596 (2016) is that their work focused only on Brewster glacier, defined as a maritime glacier.

597 These types of glaciers, even during the accumulation period can experience strong varying
598 albedos in their lower reaches, which leads to similar behaviors in winter as for Argentière and
599 Mer de Glace glaciers. We therefore reconsider the idea of Sirguey et al. (2016) to use a
600 threshold as a representative value of fresh snowfall, as there is no physical reason that this
601 threshold varies, at least within the same region. However, an interesting perspective would be
602 to apply the method without threshold, on a set of other maritime or large glaciers ($> 10 \text{ km}^2$).
603 An additional approach has been carried out, aiming at retrieving b_w by deduction from the
604 reconstructed b_a and b_s from the albedo signal. This approach, not using the winter albedo
605 signal, is poorly correlated ($r^2 < 0.16$) to in situ b_w for the six seasonally monitored glaciers.
606 Indeed, the result extremely depends on the quality of the correlations between b_a , b_s and the
607 albedo signals. Saint-Sorlin Glacier is a good example, being one of the glaciers with the
608 highest correlations for the annual ($r^2 = 0.86$) and summer ($r^2 = 0.94$) SMB. Subtracting b_s
609 from b_a to computed b_w leads to an average difference between computed and measured b_w of
610 $\pm 0.41 \text{ m w.e}$ for the 10 simulated years. As a consequence, in case of low correlations between
611 SMB and albedo, errors in the computed winter SMB become exacerbated.
612

613 **6 Conclusion**

614 In this study, we used the so-called albedo method to correlate annual and summer SMB to
615 glacier-wide average albedos obtained from MODIS images. This method has been carried on
616 30 glaciers located in the French Alps, over the period 2000-2015. Images processing has been
617 performed using the MODImLab algorithm, and filters on the images have been applied,
618 removing images with more than 30% cloud coverage, and excluding images with satellite
619 observation angles greater than 30° . Quality assessment has been performed and close
620 agreement has been found between albedos from AWS installed on Saint-Sorlin Glacier and
621 MODIS retrieved albedo values. Annual SMB have been significantly correlated to the
622 summer minimum albedo for 27 of the 30 selected glaciers, confirming this variable as a good
623 proxy of the glacier-wide annual SMB. For the six seasonally monitored glaciers, summer

624 SMBs obtained from the glaciological method have been significantly linked to the integral of
625 the summer albedo. However, calculating the integral of the winter albedo to quantify the
626 winter SMB as done by Sirguey et al. (2016) has shown underwhelming results. Monitoring
627 winter glacier surface albedo may provide good insights on the frequency of snow
628 accumulation at the surface of the glacier but lacks in quantifying the amount of accumulation.
629 Glaciers that experience complete snow coverage during most of the winter season showed the
630 lowest correlation ($r^2 \leq 0.33$) while the two glaciers showing the best correlations are subject
631 to some events of surface melting in their lower reaches. Yet, this approach should not be
632 definitively forsaken but requires improvements to confidently retrieve winter SMB.

633 Sensitivity study on the impact of the considered cloud coverage has revealed a high
634 confidence in the MODImLab cloud algorithm, limiting pixel misclassifications, and a rather
635 high tolerance of the integrated signal to the number of partly cloud-covered images. This
636 confidence on cloud filters is very promising to document unmonitored glaciers. Correction of
637 shadows by the MODImLab algorithm has however revealed some limitations when a large
638 share of the glacier is shadowed by the surrounding topography (around winter solstice).
639 Despite this, severe and artificial drops of albedo in winter have not been identified as an
640 obstacle for monitoring both summer and winter SMB. Such systematic errors are not an issue
641 for inter-annual studies, but would be a serious issue on studies focused on albedo values
642 themselves. For future works, the MODIS archive together with albedo maps, cloud and
643 shadow masks processed with MODImLab, together with validation data from AWS offer a
644 unique dataset to monitor the temporal and spatial evolution of the surface albedo of glaciers
645 at a regional scale. Computing for instance the absorbed solar radiation (Bair et al., 2016) by
646 date and for each glacier would then be an appropriated protocol to estimate the impact of a
647 changing glacier surface albedo in terms of snow or ice melt. Quantifying albedo changes and
648 resulting mass losses with such an approach would be of major interest to better understand
649 the potential effects of possibly increasing dust content, glacier orientation or snow grain
650 growth on glacier surface melt processes.

651 To conclude, the use of optical satellite images to estimate glacier surface processes and
652 quantify annual and summer SMB from the albedo cycle is very promising and should be
653 expanded to further regions. Using images from different satellites, combining high spatial and
654 temporal resolution instruments, could substantially reduces uncertainties, especially for
655 spotting the albedo summer minimum with more confidence, but also to improve the temporal
656 resolution. This method could then in the short term, become reliable for retrieving SMB of
657 monitored and unmonitored glaciers.

658

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669

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