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Highlights

- A new TCN-based Neoglacial glacier record for the western European Alps.
- At least five LIA-like advances occurred during the last 4200 years.
- Direct dating of a prominent glacier advance coeval with the 4.2 ka event
- Other LIA-like advances occurred 3.7 ka, 2.1 ka, 1.3 ka and 0.9 ka ago
- We review NH evidence for glacier advance related to the 4.2 ka event

¹⁰Be exposure dating of the timing of Neoglacial glacier advances in the Ecrins-Pelvoux massif, southern French Alps

Melaine Le Roy ^a, Philip Deline ^a, Julien Carcaillet ^b, Irene Schimmelpfennig ^c, Magali Ermini ^c and ASTER Team ^c *

^a EDYTEM, Université Savoie Mont Blanc, CNRS, Le Bourget du Lac, France

^b ISTerre, Université Grenoble Alpes, CNRS, Grenoble, France

^c CEREGE, Aix-Marseille Université, CNRS, IRD, Coll. France, Aix-en-Provence, France

* Georges Aumaître, Didier L. Boulès and Karim Keddadouche

Abstract

Alpine glacier variations are known to be reliable proxies of Holocene climate. Here, we present a terrestrial cosmogenic nuclide (TCN)-based glacier chronology relying on 24 new ¹⁰Be exposure ages, which constrain maximum Neoglacial positions of four small to mid-sized glaciers (Rateau, Lautaret, Bonnepierre and Etages) in the Ecrins-Pelvoux Massif, southern French Alps. Glacier advances, marked by (mainly lateral) moraine ridges that are located slightly outboard of the Little Ice Age (LIA, c. 1250-1860 AD) maximum positions, were dated to 4.25 ± 0.44 ka, 3.66 ± 0.09 ka, 2.09 ± 0.10 ka, c. 1.31 ± 0.17 ka and to 0.92 ± 0.02 ka. The ‘4.2 ka advance’, albeit constrained by rather scattered dates, is to our knowledge exposure-dated here for the first time in the Alps. It is considered as one of the first major Neoglacial advance in the western Alps, in agreement with other regional paleoclimatological proxies. We further review Alpine and Northern Hemisphere mid-to-high latitude evidence for climate change and glacier activity concomitant with the ‘4.2 ka event’. The ‘2.1 ka advance’ was not extensively dated in the Alps and is thought to represent a prominent advance in early Roman times. Other Neoglacial advances dated here match the timing of previously described Alpine Neoglacial events. Our results also suggest that a Neoglacial maximum occurred at Etages Glacier 0.9 ka ago, i.e. during the Medieval Climate Anomaly (MCA, c. 850-1250 AD). At Rateau Glacier, discordant results are thought to reflect exhumation and snow cover of the shortest moraine boulders. Overall, this study highlights the need to combine several sites to develop robust Neoglacial glacier chronologies in order to

33 take into account the variability in moraine deposition pattern and landform obliteration and
34 conservation.

35

36 Keywords: Holocene, Neoglacial, Glacier chronology, 4.2 ka event, TCN dating, ^{10}Be ,
37 southern French Alps

38

39

40 1. Introduction

41

42 The considerable quasi-global glacial wastage that has been taking place for more than a
43 century is currently accelerating in the European Alps (**Vincent et al., 2017**) in response to a
44 warming trend that is higher than the hemispheric average (**Auer et al., 2007**). Glacier mass
45 changes measured over the last three decades cannot be explained without accounting for
46 anthropogenic forcing (**Marzeion et al., 2014**). Conversely, glacier mass losses immediately
47 following the end of the Little Ice Age appear to have been primarily driven by natural
48 forcings (**Vincent et al., 2005; Lüthi, 2014; Marzeion et al., 2014; Sigl et al., 2016**). The
49 knowledge of the timing and amplitude of Holocene LIA-type glacial events and the
50 underlying natural climate forcings could improve our understanding of the glacier evolution
51 and responsible climate drivers since the LIA. Though, the temporal and spatial patterns of
52 such glacial events remain poorly constrained, making causes highly debated (**Wanner et al.,**
53 **2011; Solomina et al., 2015; 2016**). A better spatio-temporal knowledge of Holocene climate
54 characteristics is necessary to decipher forcing factors and improve climate model simulations
55 (**Schmidt et al., 2014; McCarroll, 2015**).

56 Small to medium-sized alpine glaciers react sensitively to short-term climate variations
57 (**Jóhannesson et al., 1989; Oerlemans, 2005; Six and Vincent, 2014; Roe et al., 2017**). This
58 has last been evidenced in the Alps by the decadal-scale period of minor climate deterioration
59 in the second half of the 20th century that led to significant glacier advances and moraine
60 deposition in the early 1980s (**Patzelt, 1985**). The glacial-geologic record (i.e. moraine ridge
61 stratigraphy) is hence often considered as one of the most straightforward climate proxies
62 beyond the instrumental period (e.g. **Putnam et al., 2012; Young et al., 2015**). However, full
63 use of this archive as paleoclimate proxy is restricted by its discontinuous nature, by the
64 selective preservation of glacial deposits and by possible non-climatic controls on moraine

65 deposition, such as glacier stagnation or advance due to coverage of the glacier tongue by
66 rockfall debris.

67 The glacial geologic record could differ between nearby sites due to different glacier
68 hypsometry and ice-flow dynamics (Winkler et al., 2010; Barr and Lowell, 2014). For this
69 reason, dating of moraine complexes provides not only insights into the timing of past glacier-
70 friendly periods, but also allows interpreting the glacial record in terms of landform
71 deposition, conservation and obliteration (Gibbons et al., 1984; Kirkbride and Winkler,
72 2012; Barr and Lowell, 2014). Well-distributed regionally-significant chronologies based on
73 a homogeneous set of glaciers (size, hypsometry) are therefore requisite to obtain a more
74 complete picture of glacier advances in response to local climate changes. This is particularly
75 true with regard to the Neoglacial, a period during which the glacier advances were of nearly
76 the same magnitude throughout the Northern Hemisphere (Solomina et al., 2015), favouring
77 self-censoring of the moraine record. These chronologies will also permit to assess if climate
78 was the main driver of the glacier variations, because non-climatic controls should not affect
79 several glaciers simultaneously.

80 There is still a dramatic lack of direct constraints on Holocene glacier variability in the
81 westernmost Alps. Investigations on the French side of the Alps have mainly focused on the
82 Mont Blanc massif (see Le Roy et al., 2015 and references therein). The Ecrins-Pelvoux
83 massif (hereafter EPM) is the south-westernmost currently glaciated area in the European
84 Alps – with the exception of isolated and soon-disappeared glacierets located in the
85 Belledonne massif to the west and in the Ubaye and Maritime Alps massifs to the south (Fig.
86 1a). To deal with the problem of representativeness and selective preservation of moraine
87 deposits we sampled the outermost Holocene ridges at four different glacier forefields located
88 throughout the EPM (Fig. 1b). The aim of this study was to obtain a local-scale overview of
89 the timing of Holocene/Neoglacial maxima. We present ^{10}Be exposure ages constraining
90 glacier advances that were almost as extended as the late-LIA stages, marked by the so-called
91 ‘AD 1850 moraines’. Given the location of the dated moraine segments, at most a few tens of
92 meters outboard of the LIA extent, we consider in the following text that glacier size during
93 these advances was virtually the same as during the ensuing LIA maxima.

94

95 **Fig. 1**

96

97 2. Study area and previous work

99 The EPM (centred on 44°55'N 6°17'E) belongs to the external crystalline massifs of the
100 western Alps. It consists of blocks of European basement, which was intruded by granites and
101 metamorphosed during the Hercynian orogeny, then exhumed along crustal-scale faults since
102 Oligocene–Early Miocene times. Remnants of inverted Jurassic sedimentary basins are
103 interspersed between these blocks (**Dumont et al., 2008**). Hence, bedrock lithologies of the
104 glaciated catchments are mostly quartz-rich granites and gneisses (**Barbier et al., 1976**;
105 **Barféty and Pêcher, 1984**).

106 The massif presents a high alpine topography with several summits reaching c. 4000 m
107 a.s.l. (**Fig. 1b**) and valley bottoms around 1000-1500 m a.s.l. Imprint of Quaternary
108 glaciations on present-day landscape is strong as shown by the U-shaped valley profiles,
109 valley rock steps and overdeepenings, lateral hanging valleys, and widespread glacial
110 trimlines (**Montjuvent, 1974**; **Delunel et al., 2010b**; **Valla et al., 2010**). Last Glacial
111 Maximum (LGM) ice cover has been reconstructed from mapping and interpolation of glacial
112 features (**van der Beek and Bourbon, 2008**). Geomorphic evidence also permitted to
113 reconstruct the LIA glacier extent, which amounted to c. 171 km² (**Fig. 1b**; **Edouard, 1978**;
114 **Gardent et al., 2014**). Present day glacierization is characterized by highly fractionated small
115 cirque and slope ice bodies (n = 282; mean surface: 0.24 ± 0.6 km²), which covered 68.6 km²
116 in 2009 (**Fig. 1b**; **Gardent et al., 2014**). Few valley glaciers are still present today, and most
117 of them are debris-covered. The two largest glaciers are the debris-free Girose Glacier (5.1
118 km², that forms an ice surface of 8.1 km² with the contiguous Mont-de-Lans Glacier) and
119 Glacier Blanc (4.8 km²) (2009 data; **Fig. 1b**).

120 The deglacial chronology from the LGM maximum was improved recently in the EPM
121 with TCN ages constraining ice thinning and tongue retreat (**Delunel et al., 2010b**). The
122 Younger Dryas stadial has led to readvances, which were mapped in several catchments
123 (**Edouard, 1978**; **Franco, 1981**; **Coûteaux, 1983a**; **1983b**; **Colas, 2000**; **Di Costanzo and**
124 **Hofmann, 2016**) and numerically-dated only at one location (**Chenet et al., 2016**). It appears
125 that glacier tongues were restricted to the highest cirques at the beginning of the Holocene
126 (e.g. **Coûteaux, 1983a**). State-of-the-art knowledge on the deglaciation was summarized by
127 **Cossart et al. (2011)** for the eastern part of the EPM. However, no studies have focused on
128 the dating of Holocene/Neoglacial stadials in the EPM up to now. Only LIA and post-LIA
129 advances were dated by palynology (**Tessier et al., 1986**) and lichenometry (**Cossart et al.,**
130 **2006**; **Le Roy and Deline, 2009**; **Le Roy, 2012**). In addition, it should be noted that subfossil
131 woods were found at two glacier forefields in the EPM: at Chardon Glacier (**Fig. 1b**; **Vivian,**

132 **1979**; Le Roy, unpublished data) where one of the samples yielded a radiocarbon age of 1035-
133 1410 cal. AD (med. prob.: 1245 cal. AD; **Vivian, 1979**), and at Etages Glacier (this work).
134 However, in these studies, the circumstances of sampling are either not accurately known or
135 indicate detrital woods of unknown original location. A firm link with glacier variations has
136 therefore not yet been established at these sites. In any case, these wood remains indicate
137 periods when the treeline was at higher elevation – as the sites of recovery are nowadays
138 devoid of woody vegetation – and mostly composed of *Pinus cembra*, a species which is rare
139 today in the study area (**Coûteaux, 1984**).

140 Pioneering glaciological studies in the EPM date back to the late 19th century with the first
141 punctual glacier length measurements (**Bonaparte, 1891; 1892; Kilian, 1900**) and were
142 intensified in the 20th century (**Jacob and Flusin, 1905; Allix et al., 1927; Letréguilly and**
143 **Reynaud, 1989; Reynaud, 1998; Reynaud and Vincent, 2000; Bonet et al., 2016**). Mass
144 balance is surveyed at Glacier Blanc since AD 2000 with the glaciological method, and was
145 reconstructed for earlier dates based on remote sensing data (**Thibert et al., 2005; Rabatel et**
146 **al., 2008; 2016**). Based on remote sensing of the end-of-summer snowline, average ELA was
147 determined to be 3100 ± 80 m for 11 of the main EPM glaciers during the AD 1984-2010
148 period, with the lowest values occurring during the year 1984 (2995 ± 60 m) and the highest
149 during the year 2003 (3330 ± 75 m) (**Rabatel et al., 2013**). This can be compared to the
150 average value of 2990 ± 110 m computed over the same time interval for 14 glaciers in the
151 Mont Blanc massif, located 120 km north (**Rabatel et al., 2013; Fig. 1a**). Additionally,
152 **Cossart (2011)** derived an ELA rise of c. 250 m between the LIA (2763 ± 174 m) and present
153 day (AD 2004; 3024 ± 175 m) for 60 of the main EPM glaciers applying an Accumulation
154 Area Ratio (AAR) of 0.67.

155

156 **Fig. 2.**

157

158 3. Sample sites

159

160 The investigated glacier forefields belong to the Romanche (Rateau and Lautaret Glaciers)
161 and Vénéon (Bonniepierre and Etages Glaciers) catchments (**Fig. 1b**). Characteristics of the
162 study sites are given below and in **Tab. 1**

163

164 **Tab. 1.**

165

166 *3.1. Rateau Glacier*

167 Rateau Glacier is located in the north face of Rateau (3809 m a.s.l.; **Fig. 2a**). It is
168 becoming a mountain-type glacier as the steep slope body will be soon disconnected from the
169 collapsing debris-covered tongue (**Fig. 2a**). During the LIA maxima, Rateau Glacier was
170 connected to the nearby Meije Glacier, and they had together the lowest frontal position in the
171 EPM (**Fig. 2a; Tab. 1**). At least two generations of post-LIA frontal moraines can be
172 identified at c. 2210 and c. 2260 m (**Fig. 2a**), likely deposited during the 1920s and the 1980s.
173 During periods of large Holocene extent, Rateau Glacier fed a small left-hand side marginal
174 lobe, which has built a set of nested lateral ridges (**Fig. 2a**). Eight ridges were mapped on the
175 western part of the lobe, among which the four outer were sampled here (MA-MD; **Fig. 3a**).
176 MD is the most individualized and sharp-crested ridge in the sampled set. The three outermost
177 ridges (MA-MC) features more subdued crests, due to erosion, but also because accretion of
178 material has partially covered pre-existing ridges during advances of same magnitude. Today,
179 the marginal lobe is no longer active and has evolved into rock glacier morphology. Historical
180 length surveys provide information on Rateau Glacier recurrences during the post-LIA retreat.
181 A first advance peaked around AD 1891 (> 13 m), followed by the tongue stagnation until
182 1894; a second advance began between 1907 and 1911 and culminated around 1921-24 (> 35
183 m), with the highest frontal progression in 1918-19 (**Allix et al., 1927**).

184

185 *3.2. Lautaret Glacier*

186 Lautaret Glacier is the largest remnant of formerly connected ice bodies that occupy the
187 ENE-faced cirque located below Grand Doigt (3973 m a.s.l.) and Gaspard Peak (3881 m
188 a.s.l.; **Fig. 2b**). All these glaciers are of mountain-type after the disconnection of the presently
189 debris-covered dead-ice tongue. At least five generations of frontal moraines have been
190 mapped (**Fig. 2b**). Two latero-frontal ridges can be traced down to c. 1940 m and c. 1970 m
191 on the right bank of the stream. Around 2150-2170 m, two sets of ridges were likely deposited
192 between the 1890s and the 1920s. Finally, the youngest moraine at the margin of the stagnant
193 ice tongue dates from the early 1980s (**Fig. 3b**). Additionally, four main vegetated ridges
194 were identified atop of the large left lateral composite moraine, around 2340 m (MA-MD;
195 **Fig. 3b**). The TCN sampling site is located on the distal face of the outermost moraine ridge
196 (MA) where only one suitable boulder was found for analysis (**Fig. 3b**). This outer ridge is
197 cross-cutted by the immediately located inboard ridge MB (**Fig. 3b**). Lautaret Glacier was

198 rarely surveyed: it was found advancing by 23 m between 1890 and 1891 (**Bonaparte, 1892**),
199 and by c. 4 m between 1919 and 1924 (**Allix et al., 1927**).

200

201 *3.3. Bonnepierre Glacier*

202 Bonnepierre Glacier is a simple, debris-covered, mainly avalanche-fed, valley glacier
203 occupying a cirque located below the 1000 m-high Dôme des Ecrins (4015 m a.s.l.) NW face
204 (**Fig. 2c**). The right lateral moraine extends over 3.1 km between c. 2880 and 2120 m. This
205 large, composite, single crested moraine subdivides into four well-defined nested ridges
206 around 2550 m. However, erosion of the proximal face of the moraine makes difficult to
207 establish a firm correspondence between the innermost segments (**Fig. 3c**). All the segments
208 (MA-ME) were sampled for TCN (**Fig. 3c**). Another set of lateral ridges – not sampled – has
209 been identified at the same elevation atop of the left lateral moraine (**Fig. 2c**). In frontal
210 position, several ridges are present between 2120 m and 2360 m. Lichenometric surveys
211 carried out on these ridges revealed that the maximum frontal extent was achieved at the end
212 of the LIA (c. AD 1854) and that the youngest frontal moraine was deposited during the early
213 1940s (**Le Roy and Deline, 2009; Le Roy, 2012**). Hummocky terrain in between denotes a
214 rapid demise of a likely already debris-covered tongue after the end of the LIA. Lichen
215 measurements were also performed on the lateral ridges MA, MB and ME (60 to 100
216 measurements per surface; **Le Roy and Deline, 2009; Le Roy, 2012**). Results obtained are
217 105 mm, 72 mm and 48 mm (mean of the five largest thalli per surface), respectively (**Le Roy**
218 **and Deline, 2009; Le Roy, 2012**). Post-LIA behaviour of Bonnepierre Glacier shows a rapid
219 retreat between AD 1860 and 1890, which then slowed until 1895 and was followed by a
220 nearly stable front over the 20th century (total retreat since 1895 only amounts to 210 m).
221 Periods of still-stand (1895-1899 and 1967-1977) and small advance (+15 m between 1913
222 and 1922) show that this glacier nevertheless reacted to the 20th century climate downturns
223 (**Allix et al., 1927; Edouard, 1978**). Rock avalanche history of this site includes several
224 undated events evidenced by large deposits on the right bank of the valley (**Fig. 2c**), of which
225 one is interstratified in the upper part of the lateral moraine (**Fig. 4c**), pointing to an
226 occurrence during the last millennium. More recently, the collapse of the Clocher des Ecrins
227 (3801 m a.s.l.) summit zone ($< 1 \times 10^4 \text{ m}^3$) occurred during the winter 1931-1932 (**Bourgeat,**
228 **1990**), and a large rock avalanche ($\sim 3.6 \times 10^5 \text{ m}^3$) occurred in the Flambeau des Ecrins north
229 face (3000 m a.s.l.) on 19th July 2016.

230

231 *3.4. Etages Glacier*

232 Etages Glacier is a north-facing valley glacier located in a cirque below Pointe du Vallon
233 des Etages (3564 m a.s.l.) and Tête de l'Étret (3559 m a.s.l.). The tongue is disconnected from
234 the small remaining slope ice bodies (**Fig. 2d**). Six frontal moraine ridges were mapped
235 between 2090 and 2355 m (M1-M6; **Fig. 3d**). Sampling for TCN dating was carried out only
236 on the outermost ridge M1, a c. 2-m-high, clast-supported and boulder-rich, frontal moraine
237 remnant dissected by proglacial streams. All moraines (except M4 and M6) were extensively
238 surveyed to locate and measure the largest lichen thalli per boulder (50 to 150 measurements
239 per surface; **Le Roy and Deline, 2009; Le Roy, 2012**). On moraine M1, the mean of the five
240 largest thalli amounts to 88.6 mm, which prevents an estimate of the age of this surface based
241 on available growth curve. Moraine M2 is only preserved on the right side of the valley and
242 lacks suitable boulders whether for lichen measurement or TCN sampling. Lichen data
243 collected on this surface (mean five largest: 52.4 mm) are thus considered minimum values
244 for dating purpose. Frontal moraine M3 (c. 4 m high) is quasi-continuous over the entire
245 width of the valley. Lichen measurements (mean of the five largest: 50.2 mm) indicate a
246 stabilisation between AD 1822 and 1853 depending on the method used (see **Le Roy, 2012**).
247 This indicates a deposition of M3 during the last major advance of the LIA. Moraine M5 is
248 dated between 1919 and 1922 depending on the method (mean of the five largest: 29 mm).
249 This is consistent with the first accurate map of Etages Glacier, established in 1929-1930,
250 which shows the tongue already retreating (c. 40 m upstream) from this moraine. Moraine
251 M6, located 180 m downstream from the 2009 front, has formed in the mid-1980s. Finally, a
252 1.4-m-long detrital subfossil log was found in the proglacial stream at 2160 m, in fall 2008.
253 The sampling location is 800 m downstream from the current glacier front and immediately
254 downstream from a perennial snow patch located at the foot of the avalanche path originating
255 in the Encoula Pass area (**Fig. 2d and 3d**).

256

257 *3.5. Glacier/climate relationships*

258 Although glacier tongues are currently debris-covered at our study sites, the climate
259 sensitivity of investigated glaciers is evidenced by historical moraine deposition and available
260 punctual length variation observations presented in Section 3. They indicate that all glaciers
261 have reacted to the three post-LIA climate downturns. By analogy with alpine glaciers for
262 which continuous length observations exist (**Müller, 1988; Letréguilly and Reynaud, 1989;**
263 **Reynaud and Vincent, 2000; Hoelzle et al., 2007**) a short reaction time characterizes glacier
264 dynamics at our study sites, likely 10-20 years, Bonnepierre Glacier being the ice mass
265 slowest to react.

266

267 4. Methods

268

269 4.1. Mapping and Equilibrium Line Altitude (ELA) determination

270 Geomorphic units were mapped based on recent aerial photographs (2009) and field
271 surveys (**Fig. 2**). ELAs were determined for the four glaciers at two different time steps, the
272 Neoglacial/LIA maximum and the early 1980s, during which glaciers were at equilibrium and
273 deposited stadal moraines. For both stadials, a digital elevation model was generated by
274 drawing contour lines of the paleoglaciers (with 50 m step) according to standard procedure
275 (e.g. **Sissons, 1977**). ELAs computation was made with the toolbox developed for ArcGIS by
276 **Pellitero et al. (2015)** with a contour interval set to 10 m. Two methods were implemented.
277 First, an Accumulation Area Ratio (AAR) of 0.67 – which has been widely used for ELA
278 reconstruction in the Alps (e.g. **Gross et al., 1977; Maisch et al., 1999**) – was applied.
279 However, this ratio can be significantly lower for glaciers experiencing strong topographical
280 control and/or extensive debris-cover (e.g. Bonnepierre Glacier), resulting in an
281 underestimation of the ELA altitude. At this latter site, a lower AAR value of 0.40 was used
282 (e.g. **Kulkarni, 1992; Scherler et al., 2011**). Second, an Area-Altitude Balance Ratio
283 (AABR) of 1.6 – representative for the Alps (**Rea 2009; Pellitero et al., 2015**) – was also
284 used, lowered to 0.9 for the debris-covered Bonnepierre Glacier. Resulting ELAs are reported
285 in **Tab. 1**. The maximum elevation of the lateral moraines (MELM) is also shown for each
286 site.

287

288 4.2. ¹⁰Be sampling

289 Sampling took place in 2010 and 2012. All samples were collected from boulders that
290 were embedded in the crest or distal face of moraine ridges. In both locations, boulder
291 exposure age is thought to reflect the age of moraine stabilisation. Indeed, landform
292 degradation observations and modelling have shown that a large boulder embedded on distal
293 face mid-slope is unlikely to have experienced significant exhumation or burial on Holocene-
294 aged moraines (**O'Neal, 2006; Putkonen and O'Neal, 2006; Putkonen et al., 2008**).
295 Consequently, ¹⁰Be ages are interpreted as close minimum ages for the culmination of the
296 glacier advances (**Gosse, 2005**). Ridges were chosen for sampling based on their distal
297 location with regard to the fresh appearing late-LIA deposits. Features indicating substantial
298 exposure time (mature lichen cover, well-developed soil) helped to determine sampling sites.

299 A *Rhizocarpon* sp. local growth curve (Pech et al., 2003; Le Roy and Deline, 2009) and
300 available lichen dates (Le Roy, 2012) were used to confirm that sampled ridges were
301 deposited prior to the second half of the LIA. Rock samples were removed from the upper cm
302 of meter-scale flat boulders with hammer and chisel (Tab. 2). Location and elevation of the
303 samples were recorded with a handheld GPS, assuming a vertical and horizontal positional
304 uncertainty of ± 10 m and ± 5 m, respectively (Tab. 2). Topographic shielding was estimated
305 in the field through skyline survey using a clinometer (Tab. 2).

306

307 **Tab. 2**

308

309 4.3. ^{10}Be geochemistry and AMS measurement

310 Samples were crushed and sieved to retain the 200-500 μm size fraction. The subsequent
311 beryllium extraction protocol is adapted from Brown et al. (1991) and Merchel and Herpels
312 (1999) chemical procedures. It was carried out at the ISTERre cosmogenic laboratory in
313 Grenoble. Quartz was isolated through magnetic separation and repeated leaching in a
314 $\text{H}_2\text{SiF}_6\text{-HCl}$ mixture. Meteoric Be purification was achieved with three sequential baths in
315 diluted HF (Kohl and Nishiizumi, 1992). The purified quartz samples (between 14 g and 52
316 g) were spiked with ~ 300 μl of a 1.10^{-3} g g^{-1} Be carrier solution (Scharlab ICP Standard)
317 before being totally dissolved in concentrated HF (Tab. 3). After the evaporation of the HF,
318 perchloric and nitric acids were added and evaporated to remove organic compounds and
319 fluorides. Anion and cation exchange columns allowed the separation of Fe and Ti and the
320 isolation of the Be fraction. At last, Be hydroxide was extracted by alkaline precipitation (von
321 Blanckenburg et al., 1996). The final BeO target were oxidized and mixed with Nb powder
322 prior to loading it on the cathode for AMS measurements, which were carried out at ASTER,
323 the French National AMS facility (CEREGE, Aix-en-Provence; Arnold et al., 2010). The
324 measured $^{10}\text{Be}/^9\text{Be}$ ratios were calibrated against the NIST_27900 beryllium standard
325 Reference Material 4325 by using an assigned value of $2.79 \pm 0.03 \times 10^{-11}$ (Nishiizumi et al.,
326 2007) and a ^{10}Be half-life of $1.387 \pm 0.012 \times 10^6$ years (Chmeleff et al., 2010; Korschinek et
327 al., 2010). The ^{10}Be concentrations inferred from the measured $^{10}\text{Be}/^9\text{Be}$ ratios are corrected
328 for the corresponding full process blank ratios which range between $(1.4 \pm 0.8) \times 10^{-15}$ and
329 $(9.7 \pm 1.3) \times 10^{-15}$ ($n = 7$; Tab. 3). AMS analytical uncertainties (reported as 1σ) include the
330 uncertainties associated with the AMS counting statistics, with the chemical blank corrections
331 and with the ASTER AMS external error (0.5%; Arnold et al., 2010; Tab. 3).

332

333 **Tab. 3**

334

335 *4.4. ¹⁰Be age computation*

336 Exposure ages were computed with the online CREp program (Martin et al., 2017;
337 <http://crep.cirp.cnr-nancy.fr/#/init>; last accessed 06/09/2017) and are presented in Tab. 3.
338 Scaling to the sample locations was made according to the recent, physically-based, LSD
339 model (Lifton et al., 2014) which performs similarly to older empirical models (Borchers et
340 al., 2016). Chosen parameters include the ERA-40 atmosphere model (Uppala et al., 2005)
341 and the Lifton VDM 2016 geomagnetic database (Lifton, 2016). We retained the production
342 rate (PR) derived by Claude et al. (2014) at the Chironico landslide site (Ticino, Switzerland,
343 260 km NE from the EPM; Fig. 1a) which is the only regional PR available so far. This PR
344 has a value of 4.16 ± 0.10 at. $\text{g}^{-1} \text{yr}^{-1}$ for LSD scaling and is consistent with other recently
345 derived Northern Hemisphere PR (Balco et al., 2009; Fenton et al., 2011; Ballantyne and
346 Stone, 2012; Briner et al., 2012; Goehring et al., 2012; Young et al., 2013; Small and
347 Fabel, 2015; Stroeven et al., 2015). In order to make our results easily comparable to
348 previous studies carried in the Alps (e.g. Schimmelfennig et al., 2012; 2014) we also
349 present in Tab. 3 exposure ages computed with the CRONUS-Earth online calculator (version
350 2.2; Balco et al., 2008; <http://hess.ess.washington.edu/>) using the time-dependent ‘Lm’
351 scaling model (Lal, 1991; Stone, 2000), standard atmosphere (N.O.A.A., 1976) and the
352 ‘Arctic’ PR (Young et al., 2013) which has a value of 3.93 ± 0.15 at. $\text{g}^{-1} \text{yr}^{-1}$ for ‘Lm’ scaling
353 (Balco et al., 2008). ‘Lm/Arctic’ ages are in average only 2.7 % older (80 ± 20 years) than the
354 ‘LSD/Chironico’ ages used in the discussion hereafter. We therefore stress that the choice of a
355 different PR/scaling model couple would not affect our conclusions. We did not apply
356 erosion- nor snow-corrections given that erosion is negligible for late-Holocene surfaces and
357 that most sampled boulders lie in windswept locations (but see further discussion in Section
358 5.1). Multiple boulder exposure ages from a single surface (i.e. moraine ridge) were averaged
359 (arithmetic mean) after identification of possible outliers. The uncertainties in the mean ages
360 are represented by the standard deviations of the retained boulders ages plus the production
361 rate uncertainty (2.4 %), added in quadrature. To ensure homogeneity between the different
362 time scale notations used in the discussion and to ease comparisons with dendro-dates and
363 calibrated radiocarbon dates, all exposure ages are hereafter expressed either in ka b2k (before
364 AD 2000) or in calendar years (BC/AD). Radiocarbon dates quoted in the discussion were

365 recalibrated from original publication using Calib 7.1 (Stuiver et al., 2017) and IntCal13
366 (Reimer et al., 2013), and are expressed both by the 2σ interval of calibration and the median
367 of the probability distribution function (Telford et al., 2004).

368

369 **Fig. 3.**

370

371 5. Results and interpretation

372

373 5.1. ELAs determination

374

375 Mean ELAs computed for the Neoglacial/LIA maximum are 2737 and 2779 m,
376 depending on the method (Tab. 1), which is in good agreement with previous work in the
377 EPM (Cossart, 2011). We found a c. 90 m ELA rise (average of the results from both
378 methods) between this stadial and the early 1980s stadial (Tab. 1). Without taking into
379 account the nearly stagnant Bonnepierre Glacier, this value amounts to c. 115 m. According
380 to available estimates of ELA climate sensitivity to summer temperature (e.g. Rabatel et al.,
381 2013), this value is equivalent to a 0.8 to 1°C warming over the considered period.

382

383 5.2. Moraine exposure ages

384

385 5.2.1. Rateau Glacier

386 The nine ages from the four outer moraine ridges (MA, MB, MC and MD) range from c.
387 2.1 to c. 3.7 ka, a relatively short time span of ~1.6 ka falling within the Late Holocene
388 period. However, the order of these ages clearly disagrees with the moraine ridge stratigraphy,
389 as the four oldest ages are obtained on the inner ridges MC and MD (Fig. 3a and 4a). The
390 entire spread in ages is similar to the one observed on the single innermost sampled ridge
391 MD. Explanations for discordant results in moraine age populations generally include nuclide
392 inheritance from previous exposure durations, exhumation or toppling of some boulders, or
393 seasonal shielding by snow cover (Schildgen et al., 2005; Delunel et al., 2014b). Here, we
394 consider nuclide inheritance and boulder toppling as unlikely. In the case of nuclide
395 inheritance, a wider and more dispersed spread would be expected in the age population. We
396 retained boulder exhumation, as well as a possible shielding by snow cover of the shortest
397 boulders, as the most probable explanation. This is confirmed by the correlation existing

398 between the four exposure ages and the heights of the corresponding samples on moraine MD
399 (**Tab. 2 and 3**). Averaging ages from a given ridge is therefore not possible here. Only the
400 tallest boulders are likely to yield a robust estimate for the timing of glacier advance. We
401 therefore tentatively interpret the oldest age on MD (RAT09, sample height above surface:
402 115 cm) – which is surely the less affected by exhumation and snow cover of all sampled
403 boulders given its geometry – as the best estimate for the deposition of this ridge at $3.66 \text{ ka} \pm$
404 0.45 ka . This implies that the three outermost ridges (MA, MB, and MC) must have been
405 deposited before that date. This interpretation is in accordance with the conclusions of
406 **Heyman et al. (2016)** who have demonstrated that the fraction of well-clustered ages on
407 moraine surfaces increases with boulder height cut-off values up to 160 cm. In comparison,
408 mean height of the samples from Rateau Glacier is $0.80 \pm 0.24 \text{ m}$ (**Tab. 2**). According to this
409 explanation, ages from the ridges MA, MB and MC displayed on **Fig. 3a** must be considered
410 as distant minimum ages and are not taken into account hereafter. As an indication, only
411 apparent ages based on the single oldest age per surface are displayed on **Fig. 4a** for these
412 three outer ridges.

413

414 5.2.2. Lautaret Glacier

415 The unique sample collected at this site from a 110 cm-tall, well-embedded boulder yields
416 an age of $1.31 \text{ ka} \pm 0.17 \text{ ka}$ for the outermost lateral ridge MA. Given the size and location of
417 this boulder we consider this is a reliable estimate for the deposition of MA, although it rests
418 on a single age. This gives a maximum age for the three inboard lateral ridges MB-MD that
419 were mapped atop of the left lateral composite moraine (**Fig. 3b and 4b**).

420

421 **Fig. 4**

422

423 5.2.3. Bonnepierre Glacier

424 The four ages acquired on the outermost ridge MA range from 4.7 to 3.7 ka (**Fig. 3c**).
425 Although there is 1 ka spread within this age population and the oldest (BON03) and the
426 youngest (BON00) ages might be slightly over- and underestimated, respectively, no outlier
427 can be identified based on the Chauvenet's criterion. The mean age of the four boulders is
428 $4.25 \pm 0.44 \text{ ka}$, which is in good agreement with the consistent ages of boulders BON02 and
429 BON04 (mean: $4.35 \pm 0.11 \text{ ka}$). We therefore consider the average of all four ages a reliable
430 estimate for the timing of deposition of MA (**Fig. 4c**). Given the volume of MA (**Fig. 4c**)

431 several glacier advances have likely been necessary to build the core of this landform. Only
432 the last advance has been dated here. Ages of these older advances cannot be constrained
433 easily and could date either from the early Holocene or from the Neoglacial before c. 4.3 ka.
434 Finally, the ‘4.2 ka advance’ could possibly have been multi-phased as shown by the presence
435 of a subdued ridge, mid-slope of the distal face of MA (**Fig. 3c**). Anyway, if this has been the
436 case these distinct glacier pulses would have occurred in a very short interval as demonstrated
437 by the similar ages obtained on both the moraine crest (BON02) and at the bottom of the
438 distal face (BON04) (**Fig. 3c**).

439 The age of the next prominent ridge MB is constrained by two consistent ages (BON05
440 and BON06; **Fig. 3c**), which give a mean of 3.66 ± 0.10 ka. The small segment MC, which is
441 adjoined to the proximal face of MB (no counter-slope), yields an age of 3.65 ± 0.35 ka based
442 on a single boulder (BON07). This age is indistinguishable from that of MB. We therefore
443 interpret this ridge as having been deposited during the same glacial phase as MB.

444 Due to erosion, the two innermost segments (MD and ME) don’t feature crests. Only part
445 of the distal face is available for sampling (**Fig. 4c**). This resemblance might suggest that both
446 segments would have originally been part of the same ridge. The two boulders sampled on
447 segment MD yield ‘too old’ ages with respect to the stratigraphy (4.22 ± 0.40 ka for BON12
448 and 5.63 ± 0.68 ka for BON11; **Fig. 3c**) and were therefore considered as outliers. This
449 inherited component could originate from the deposition of avalanched material pre-exposed
450 in the cirque rock-walls or from the reworking of previously deposited moraine boulders.
451 Since this is a low-level inheritance, leading to Neoglacial ages, we favour here the latter
452 interpretation. On segment ME, boulder BON10 (3.53 ± 0.22 ka) was also considered an
453 outlier as internal uncertainty does not overlap with the other two. Given the similarity of the
454 age of this boulder with the mean age of the immediately outboard located blocky ridges
455 MB/MC, it is likely that this boulder was reworked from the proximal face of MB/MC, before
456 its deposition with the ridge ME (**Fig. 4c**). The two remaining boulders on ME (BON08 and
457 BON09) yield consistent ages and allow proposing a deposition of this ridge at 2.09 ± 0.10
458 ka, in agreement with the stratigraphy. We have privileged the youngest ages on ME as best
459 estimates of its true deposition age due to morphostratigraphical considerations and to the
460 significantly different lichen data obtained on MB and ME (see Section 3.3), suggesting that
461 these two ridges were not deposited during the same glacial phase. Moreover, based on field
462 observations, it is highly unlikely that these two boulders were exhumed or even toppled after
463 deposition.

464 Exposure ages acquired at this site show that lichen data (see Section 3.3) are not relevant
465 to estimate the timing of deposition of mid-to-late Holocene moraines in the area. Despite
466 their internal consistency (i.e. conforms to the stratigraphy), lichen measurements available
467 here greatly underestimate moraines ages (see **Le Roy, 2012**). This is likely because of a
468 rejuvenation of the population as the first colonizers are no longer present and saturation of
469 the surfaces occurs. ‘Too young’ lichen ages were also reported in other settings (e.g. **Wiles
470 et al., 2010**) even if a fair matching between lichenometric and cosmogenic ages could also be
471 noted elsewhere (**Young et al., 2009**) in regions where saturation likely occurs later than in
472 the Alps, due to lower growth rates.

473

474 5.2.4. *Etages Glacier*

475 Frontal moraine M1 is dated to 0.92 ± 0.02 ka (AD 1085 \pm 20) based on the mean of two
476 consistent ages (ETA01 and ETA02; **Fig. 3d**). This date marks the Neoglacial maximum of
477 Etages Glacier.

478 Independent evidence reinforces the TCN dating of moraine M1 during the MCA. A
479 detrital log (eta01) was found in the proglacial stream below the snow patch, c. 400 m
480 upstream from M1 (**Fig. 3d**). The tree-ring series from this sample was crossdated with two
481 different *Pinus cembra* reference chronology: the Eastern Alpine Conifer Chronology from
482 Tirol (EACC; **Nicolussi et al., 2009**) and a reference from Mont Blanc massif (**Le Roy et al.,
483 2015**). Consistent results indicate this tree lived more than 385 years between AD 632 and
484 1016+, an unknown number of outer rings being missing due to abrasion (**Fig. 5**). It is not
485 possible to state if the tree death is directly related to glacial activity. This seems however
486 unlikely given the marked eccentric pattern of the tree-rings (**Fig. 5**) – probably resulting
487 from a growth on a steep slope just above the glacier forefield – and the sampling site located
488 directly downstream of a major avalanche path (**Fig. 3d** and **4d**). Nevertheless, its
489 preservation required a fairly rapid burial by glacial sediments. This surely occurred by the
490 end of the 11th century AD (c. 0.9 ka ago), when Etages Glacier likely advanced beyond the
491 location of the avalanche deposit during the course of its Medieval advance.

492

493 **Fig 5.**

494

495 Interestingly, a ¹⁰Be inventory of glaciogenic material performed by **Delunel et al.
496 (2014a)** in the Etages valley (see **Fig. 3d** for location of their samples) shows good agreement

497 with our data. They report apparent ages (recalculated here with a ~15 % lower PR) for sand
498 and gravel samples taken at the glacier portal (c. 400 yrs; Rd 59), on the main frontal outwash
499 near moraine M1 (c. 830 and 1410 yrs; Rd 61 and 62), as well as at two locations on the large
500 latero-medial moraine that the authors call ‘LIA moraine’ (c. 530 and 680 yrs; Rd 60 and 57).
501 Moreover, a zero-concentration was determined for a sample taken from the glacier surface
502 (Rd 58). According to the authors’ interpretation, this dataset indicates a significant inherited
503 ^{10}Be component at some locations, because some ages are considered ‘too old’ for LIA
504 moraines. However, given our results, we consider the latero-medial moraine to be built
505 during part of the Neoglacial, with the two most important advances occurring at c. AD 1100
506 (deposition of M1) and somewhere between 1100 and the 19th century (deposition of M2).
507 Therefore, ^{10}Be concentrations corresponding to a c. 1 ka-long exposure for samples from the
508 outwash located near moraine M1 are in good agreement with our dating of this ridge at 0.9
509 ka, as is the early-LIA apparent age determined on the latero-medial moraine (**Fig. 3d**).
510 Finally, the low to null ^{10}Be concentrations measured by **Delunel et al. (2014a)** for the
511 supraglacial load reinforce confidence in ages we obtained on moraine M1, as inheritance
512 seems absent at this site.

513

514 5.3. Summary

- 515 - The moraines that we were able to date all fall into the Late Holocene period, between
516 c. 4.3 and 0.9 ka (**Fig. 6**).
- 517 - The oldest moraine in the data set is dated to 4.3 ± 0.4 , which can most likely be
518 correlated with the classical ‘4.2 ka event’ (**Walker et al., 2012; Fig. 6**).
- 519 - Around 3.7 ka, several moraines were deposited in a short interval at Bonnepierre
520 Glacier and also probably at Rateau Glacier. Other than that, none of the dated
521 moraines is contemporaneous with any one from the other sites. Anyway, because of
522 the close spatial proximity, we consider that the moraines ages are complementary and
523 reflect a common climate signal modulated by respective glacier dynamics.
- 524 - The other advances occurred at 2.1 ± 0.1 and likely at 1.3 ± 0.2 ka. The youngest one
525 occurred during the classical MCA at 0.9 ka.
- 526 - Exhumation seems to be the main process explaining boulder age discrepancy at
527 Rateau Glacier site, while three boulders showing low-level inheritance were
528 identified on the innermost moraine segments at Bonnepierre Glacier site.

529 - Mean Neoglacial/LIA ELAs at our studied glaciers were 2737 and 2779 m, depending
530 on the method. We found an ELA rise of c. 90 m (up to c. 115 m excluding
531 Bonnepierre Glacier) between the Neoglacial maxima and the early 1980s.

532

533 6. Discussion

534

535 6.1. The ‘4.2 ka advance’, one of the first major Neoglacial advance in the Alps

536

537 We interpret the ‘4.2 ka advance’ – to our knowledge directly (i.e. exposure-) dated here
538 for the first time in the Alps – as one of the first major Neoglacial advance in this area.
539 Together with data from the literature (Sections 6.1.1, 6.1.2 and 6.2), we propose this advance
540 as possibly marking the onset of this period (**Fig. 6**), defined here as a late-Holocene interval
541 during which prominent advances reached Holocene maxima and no multi-centennial (> 500
542 yrs) climate optima occurred. This differs from the definition sometimes found, i.e. the
543 renewal of glacial activity (without reaching Holocene maxima on large valley glaciers)
544 following the Holocene Hypsithermal – which starts between c. 6.6 and 5.2 ka, both in the
545 Alps (**Fig. 6c-f**; **Baroni and Orombelli, 1996**; **Orombelli, 1998**; **Nicolussi and Patzelt,**
546 **2001**; **Wipf, 2001**; **Deline and Orombelli, 2005**; **Joerin et al., 2008**; **Holzhauser, 2010**;
547 **Nicolussi et al., 2013**; **Gabrielli et al., 2016**) and in northwestern North America (**Menounos**
548 **et al., 2009**; **Harvey et al., 2012**; **Osborn et al., 2012**; **Luckman et al., 2017**).

549

550 **Fig 6.**

551

552 6.1.1. Paleoclimate evidence

553 There is a wealth of paleoclimatic evidence indicating that the period 4.4–4.0 ka has seen
554 a cooling and an increase in humidity around the north Atlantic. Likewise, this period marks a
555 multi-centennial-scale ‘tipping point’ in several proxy time-series (**Fig. 6**). North Atlantic
556 lake-sediments records, Alpine treeline variations, and high-elevation speleothems $\delta^{18}\text{O}$
557 records show a marked cooling (**Fig. 6a/b**; **Andresen et al., 2006**; **Fohlmeister et al., 2013**;
558 **Nicolussi et al., 2005**). In central north-western Europe, lake-levels, fluvial geomorphic
559 activity and erosion in the montane zone peaked at that time (**Fig. 6h-j**; **Hoffmann et al.,**
560 **2008**; **Magny et al., 2011**; **2012a**; **2012b**; **Bajard et al., 2016**), as well as peat humification
561 proxies in oceanic north-western Europe (**Barber et al., 2003**; **Eckstein et al., 2010**). In the

562 southern French Alps, a major detrital pulse centred on 4.2 ka was recorded in a high-
563 elevation catchment (Lake Petit; **Brisset et al., 2013**; see **Fig. 1a** for location). In the same
564 area, a cluster of TCN ages constraining rock avalanches – probably triggered by high rainfall
565 events – appears coeval with this period (**Zerathe et al., 2014**). Close to our study sites, the
566 rock avalanche deposit damming Lauvitel Lake (see **Fig. 1b** for location) was also TCN-dated
567 to this interval, at 4.65 ± 1.3 ka (2650 ± 1300 BC; mean age recalculated here from **Delunel**
568 **et al., 2010a** with four additional ages and without taking into account possible outliers) –
569 which is consistent with the dating of the lake sedimentary fill (**Fouinat et al., 2013**).

570 The most accurate temporal constraint on the ‘4.2 ka event’ is given by a bog-pine dying-
571 off phase which coincides with a major disruption in bog-oak growth indices from 2168 BC
572 onwards in northwest Germany – indicating wet and cool conditions with regional
573 significance (**Eckstein et al., 2010**). Since some other northern Europe bog-tree low-growth
574 and dying-off events – e.g. during the AD 4th and 6th centuries (**Sass-Klaassen and Haenrets,**
575 **2006**) – correspond to known glacier advances in the western Alps (**Holzhauser et al., 2005;**
576 **Le Roy et al., 2015**), this wet/cool c. 4.2 ka period could have triggered positive mass balance
577 in the Alps as well. Cool conditions are confirmed by the $\delta^{18}\text{O}$ record of high-elevation
578 speleothems, a proxy for annual temperature, which rose markedly at c. 2165 BC, reaching
579 for the first time since the Preboreal high values that characterize late Holocene cold periods
580 (**Fig. 6a; Fohlmeister et al., 2013**). Moreover, dendro-densitometry analyses carried out on
581 timberline subfossil larch trees in the Swiss Alps show an extended drop, reflecting low
582 summer temperature, between 2150 and 2030 BC, with a minimum around 2110 BC
583 (**Renner, 1982**). Our estimate of the age of moraine MA at Bonnepierre Glacier at 4.25 ± 0.44
584 ka (2250 ± 440 BC) thus overlaps with independent evidence for cooling and wetter
585 conditions presented above.

586

587 *6.1.2. Glacier-related evidence*

588 In the Alps, some glaciers were advancing slightly beyond modern positions as early as
589 4.5 ka (**Fig. 6d/f**). In Austrian Alps, an advance of the Schwarzenstein Glacier is recognized
590 after 2870–2500 cal. BC (med. prob.: 2685 cal. BC; **Pindur and Heuberger, 2010**; all cited
591 locations from previous studies are shown in **Fig. 1a**) and the Gepatsch and Pasterze Glaciers
592 reached their AD 1960 and AD 2000 levels, respectively, around 2500 BC (**Nicolussi and**
593 **Patzelt, 2001**). However, in these cases, the maximum extent achieved at that time is not
594 known.

595 Data relating to the 4.2 ka glacier advance itself are spatially better constrained, as some
596 scarce moraines could tentatively be assigned to this period. At Brenay Glacier (Swiss Alps),
597 a soil buried by the outermost ridge of the Holocene moraine complex has yielded an age of
598 3485–1965 cal. BC (med. prob.: 2685 cal. BC; **Schneebeli, 1976**). Despite large dating
599 uncertainties, this ridge may have been deposited at the end of the third millennium BC.
600 Similarly, Schmadri Glacier has reached or surpassed its LIA maximum extent after 2460–
601 2040 cal. BC (med. prob.: 2240 cal. BC) as shown by a soil buried below a lateral ridge
602 located 50 m outboard of the LIA maximum moraine (**Wipf, 2001**). Other evidence for
603 contemporary glacier advance without preserved moraine ridges come from Morteratsch
604 Glacier, which slightly exceeded its AD 1862 extent after 2570–2205 cal. BC (med. prob.:
605 2390 cal. BC) as demonstrated by glacially-sheared *in situ* stumps (**Holzhauser, 2010**). Alike,
606 Rhône Glacier experienced a large advance, slightly exceeding its LIA maximum extent, just
607 after 2570–2135 cal. BC (med. prob.: 2340 cal. BC) as indicated by a reworked buried soil
608 (**Zumbühl and Holzhauser, 1988; Holzhauser, 2007**).

609 At some other sites, the maximum extent of the c. 4.2 ka advance is not precisely known
610 but evidence was found that this cool/wet event represented the end of a long lasting climate
611 optimum. Unambiguous, glacier-related, evidence for this warm period comes as continuous
612 peat growth under the location of the current tongue of Pasterze Glacier (Austrian Alps)
613 between 3360–3100 cal. BC (med. prob.: 3210 cal. BC) and 2460–2210 (med. prob.: 2330
614 cal. BC). To allow peat colonization, the glacier had to be significantly smaller than its AD
615 2007 extent during this c. 1 ka interval (**Kellerer-Pirklbauer and Drescher-Schneider,**
616 **2009**). In the Italian Alps, radiocarbon-dating of the last rings of a subfossil log found in the
617 Forni Glacier forefield indicates its tongue likely exceeded its AD 1960 level after 2251–2082
618 cal. BC (med. prob.: 2186 cal. BC), ending a retreat upstream of this position that lasted at
619 least 300 years (**Pelfini et al., 2014**). On the Tisenjoch Pass – the site where Ötzi was
620 discovered – the youngest buried soil was dated to 2560–2150 cal. BC (med. prob.: 2360 cal.
621 BC). It testifies a thinning of the Niederjoch Glacier accumulation area leading to a
622 significant ice-free period at this high-elevation crest site (3215 m a.s.l.) during the third
623 millennium BC (**Baroni and Orombelli, 1996**). In the Swiss Alps, Gault Glacier exceeded its
624 AD 1980 level for the first time in 400 years after 2835–2155 cal. BC (med. prob.: 2445 cal.
625 BC; **Wäspi, 1993**). Similarly, the slow reacting Gorner Glacier exceeded its AD 1963 level
626 (or a more upstream location) slightly after 2231 BC, as shown by dendro-dating of detrital
627 logs whose outermost rings are lacking. This advance ended a glacier extent similar or shorter
628 than today that lasted at least 650 years at this site (**Holzhauser, 2010**). The Subboreal

629 Climate Optimum (c. 3200–2200 BC) is also particularly visible in the dataset of radiocarbon-
630 dated archeological artefacts from Schnidejoch Pass (**Fig. 7b**; **Hafner and Schwörer, 2017**).
631 The subsequent marked drop of the summed probability density functions (PDFs) of the
632 calibrated radiocarbon ages centred on 2300-2200 BC (**Fig. 7b**) likely indicates an
633 enlargement of the adjoining Chilchli Glacier leading to unfavourable conditions to cross the
634 pass during this interval.

635 South of the Alps, the renewed growth of Calderone Glacier just after 2565–2155 cal. BC
636 (med. prob.: 2370 cal. BC), as shown by a soil reworked by the glacier tongue and
637 incorporated into a moraine, is considered as marking the onset of the Neoglacial in the
638 Apennines (**Giraudi, 2005**).

639 Finally, between the ‘4.2 ka advance’ and the culmination of the next period of glacier
640 advance (Löbben Advance Period, see Section 6.3.1), a marked and possibly multi-phased
641 retreat occurred. Morteratsch Glacier reached a position at least as retracted as c. AD 2000
642 before to advance over this extent after 2195–1775 cal. BC (med prob.: 1995 cal. BC;
643 **Holzhauser, 2010**). Renewed peat growth at Pasterze Glacier between 1935–1745 cal. BC
644 (med. prob.: 1835 cal. BC) and 1625–1445 cal. BC (med prob.: 1540 cal. BC) indicates multi-
645 secular ice free conditions (extent < AD 2007) at the beginning of the Löbben Advance
646 Period in the forefield of this slow reacting glacier (**Kellerer-Pirklbauer and Drescher-
647 Schneider, 2009**). Mostly-glacier-hostile conditions characterizing this period are also clearly
648 reflected by the temperature-sensitive $\delta^{18}\text{O}$ record from high-elevation speleothems (**Fig. 6a**)
649 as well as by the Mer de Glace and Schnidejoch records (**Fig. 7a/b**)

650

651 *6.2. The ‘4.2 ka advance’ in the Northern Hemisphere*

652

653 Glacier advances coeval with our data and marking the onset of the local Neoglacial were
654 also reported from Greenland (**Balascio et al., 2015**), Svalbard (**van der Bilt et al., 2015**)
655 Iceland (**Striberger et al., 2012; Larsen et al., 2012**) and Scandinavia (**Bakke et al., 2010;**
656 **Vasskog et al., 2012; Jansen et al., 2016**), based on lake-sediments. Outside of the North
657 Atlantic realm, numerous glacially-sheared *in situ* stumps and some wood mats were
658 radiocarbon-dated around 4.2 ka in the Canadian coastal cordilleras (see review in **Menounos
659 et al., 2008**; see also **Menounos et al., 2009; Harvey et al., 2012; Maurer et al., 2012;**
660 **Osborn et al., 2013; Mood and Smith, 2015; Luckman et al., 2017**). In this area, the
661 advance probably did not reach Holocene maxima but was quite prominent, at least similar to
662 the AD 1940 extent (e.g. **Maurer et al., 2012**). Glacier chronologies from Alaska show a

663 prominent advance dated between 4.6 and 4.0 ka marking the onset of the local Neoglacial
664 (**Krivicich, 2009; McKay and Kaufman, 2009; Badding et al., 2013**), with an ELA
665 depression around 4.2 ka comparable to the one of the Holocene maxima (**Boldt et al., 2012**)
666 and a frontal position exceeding the LIA moraines (**Badding et al., 2013**). A significant
667 increase in snow cover, persisting almost continuously from that time up to the 20th century
668 warming, has also been evidenced from wood remnants buried by perennial snow patches in
669 the Colorado Front Range (**Benedict et al., 2008**). At high latitudes, current ice-cap melt rates
670 were found to be without analog since 4.2 ka in the Canadian Arctic (**Fisher et al., 2012;**
671 **Miller et al., 2013; Margreth et al., 2014**). This whole set of data point to a hemispheric
672 climatic event leading to prominent glacier advances in most Northern Hemisphere mountain
673 ranges.

674

675 **Fig 7.**

676

677 *6.3. Other EPM Neoglacial advances in the Alpine context*

678

679 *6.3.1. The '3.6 ka advance'*

680 At Bonnepierre Glacier and likely also at Rateau Glacier, evidence is provided for a
681 prominent and multi-phased glacier advance which culminated at 3.66 ± 0.09 ka (1660 ± 90
682 BC; mean age of the three moraines at the two sites). This advance is coeval with the so-
683 called 'Löbben' period in the Alps (Löbben Advance Period, LAP; c. 1900–1450 BC; **Patzelt**
684 **and Bortenschlager, 1973; Nicolussi et al., 2006**). This result is in line with the extensive
685 body of work that reports dating of LAP moraines in the Alps (**Fig. 6d-f**; see review in
686 **Holzhauser 2010** and **Moran et al., 2017**; see also **Le Roy et al., 2015**). At several sites,
687 these moraines are the first lateral ones preserved inboard of the Lateglacial/early Holocene
688 positions (e.g. **Wipf, 2001; Schimmelpfennig et al., 2012**) and more rarely the most down-
689 valley preserved Neoglacial frontal ridges (**Wipf, 2001; Moran et al., 2017**). Our tentative
690 interpretation of the Rateau Glacier dating results implies that the three outermost ridges MA
691 to MC (on which the resulting boulder ages are most likely underestimated; **Fig. 3a** and **4a**)
692 either pre-date the LAP or were deposited during the LAP shortly before moraine MD. In the
693 Swiss Alps, the Tschingel Glacier forefield gives evidence for a multi-phased LAP, with four
694 ridges attributed to this period (**Wipf, 2001**). The higher climate sensitivity of Rateau Glacier

695 with respect to Bonnepierre Glacier (see Section 3) could explain why four ridges would have
696 been deposited during the LAP by the first, and only two by the latter.

697 Our TCN ages agree remarkably with dendro-dating evidence for the time span of the
698 LAP culmination in the Alps, placed between 1660 and 1545 BC from analyses carried out on
699 tens of glacially-buried subfossil logs (**Fig. 7a**; **Nicolussi and Patzelt, 2001**; **Le Roy et al.,**
700 **2015**). Similarly, the glacio-lacustrine record from Lake Bramant, located 20 km north of the
701 EPM, shows a rise in clastic sedimentation between c. 1700 and 1400 BC indicating a
702 contemporaneous advance of Saint Sorlin Glacier (**Fig. 7e**; **Guyard et al., 2013**). Further
703 evidence comes from the radiocarbon ages of archeological finds on Schnidejoch Pass, which
704 drop drastically at 3.6 ka, likely reflecting a glacier-induced cessation of the pass traverse
705 (**Fig. 7b**; **Hafner and Schwörer, 2017**). At Tsidjiore Nouve Glacier in the western Swiss
706 Alps, recalculation of the age of ARO-12 – the tallest and most ice-distal boulder from the
707 LAP-aged moraine dated by **Schimmelpfennig et al. (2012)** – with the PR used in this study
708 yields an age of 3.60 ± 0.11 ka (1600 ± 110 BC). Alike, recalculation of the data from frontal
709 moraine M2A at Längental Glacier, western Austria (**Moran et al., 2017**), yields a mean age
710 of 3.54 ± 0.18 ka (1540 ± 180 BC). This means that deposition of these ridges was
711 synchronous with that of our dated ridges in the EPM and consistent with independent
712 (dendro-based) evidence constraining the culmination of the LAP.

713 Finally, Alpine paleoclimatic proxies show unambiguously a sharp drop in temperature
714 between 3.6 and 3.5 ka (**Fig. 6a**; **Fohlmeister et al., 2013**; see also **Le Roy et al., 2015**, their
715 Fig. 8f), also reflected in the treeline record (**Fig. 6b**; **Nicolussi et al., 2005**).

716 On a global scale, firm evidence now indicate that glacier advances contemporary to the
717 Lössen Advance Period occurred at other Northern Hemisphere locations (**Osborn et al.,**
718 **2013**; **Luckman et al., 2017**; **Schweinsberg et al., 2017**)

719
720 *6.3.2. The '2.1 ka advance'*

721 At Bonnepierre Glacier, the next prominent advance for which evidence is preserved took
722 place in the late Iron Age/early Roman times, at 2.09 ± 0.1 ka (90 ± 100 BC). Little
723 geomorphic evidence yet exist in the Alps for an advance occurring at that time. To our
724 knowledge, only two sites provided evidence for a large advance in the first century BC. In
725 the Austrian Alps, Gepatsch Glacier advanced beyond its AD 1930 extent around 200 cal.
726 BC–130 cal. AD (median prob.: 35 cal. BC; **Patzelt, 1995**; **Nicolussi and Patzelt, 2001**). In
727 the French Alps, Argentière Glacier likely achieved advanced positions after 174+ BC as
728 shown by a log embedded in till 15 m below the moraine crest (dendro-date for the tree

729 ARG04; **Le Roy, 2012**). At the nearby glacier Mer de Glace, indirect evidence based on two
730 detrital logs could indicate that an advance occurred after 87+ BC (**Fig. 7a**). Moreover, this
731 period stands out in the glacio-lacustrine record of Lake Bramant, reflecting enhanced glacial
732 activity centred on 70 cal. BC (**Fig. 7e; Guyard et al., 2013**). Minimum-limiting temporal
733 constraints for the ‘2.1 ka advance’ include the first germination dates of subfossil trees
734 retrieved from the Mer de Glace lateral moraine, recorded from 23 BC (**Fig. 7a**). These dates
735 probably reflect the shift to warmer conditions at the onset of the Roman Era. Likewise,
736 radiocarbon age PDFs of the archeological artefacts found at Schnidejoch Pass start around
737 2.15 ka, supporting a change from glacier-friendly to glacier-hostile conditions at that time
738 (**Fig. 7b**).

739 Possible drivers for the ‘2.1 ka advance’ could be the strong rise in spring (April, May,
740 June) precipitation that occurred around 100 BC (**Büntgen et al., 2011**) associated with
741 slightly below average summer (June, July, August) temperatures (**Luterbacher et al., 2016**).
742 Shortly after, a marked drop of summer temperature occurred at 44 BC (**Büntgen et al.,**
743 **2011**). This cooling episode was caused by the third largest volcanic event of the last 2,500
744 yrs, leading to one of the coldest decade of the Late Holocene (**Sigl et al., 2015**).

745

746 6.3.3. The ‘1.3 ka advance’

747 Another Neoglacial maximum might be represented by the single boulder dated to $1.31 \pm$
748 0.17 ka (AD 690 \pm 170) at Lautaret Glacier. This dating coincides with the regionally called
749 ‘Göschenen II’ advance period (GII; c. 200-850 AD; **Zoller et al., 1966**). GII glacier
750 advances surpassed the LIA extent at a number of locations in the Alps, e.g. at Rhône Glacier
751 (**Zumbühl and Holzhauser, 1988**) and Simony Glacier (**Patzelt and Bortenschlager, 1973;**
752 **Patzelt, 1974**). At Lautaret Glacier, the deposition pattern is constrained by topography as the
753 composite moraine is backed against the rock wall (**Fig. 3b and 4b**). It is therefore possible
754 that immediately preceding Neoglacial advances were of nearly the same magnitude. Within
755 uncertainties, the Lautaret Glacier Neoglacial maximum is contemporary with the prominent
756 GII advance of Mer de Glace, dendro-dated to AD 606+, with an extent that had reached
757 approximately the AD 1870 level (**Le Roy et al., 2015**). First millennium prominent detrital
758 peaks in the glacio-lacustrine records presented on **Fig. 7d/e** could also be linked to this
759 advance, given that both age models are loosely constrained on this time period.

760 The end of the GII, also called the Late Antique Little Ice Age (**Büntgen et al., 2016**),
761 was characterized by a volcanic-induced cooling that started in AD 536 (**Toohey et al., 2016**).
762 Cold conditions with at least hemispheric relevance persisted then until AD 660 (**Büntgen et**

763 **al., 2016**), possibly responsible for the low juvenile growth of the Eta01 tree near the Etages
764 Glacier forefield (**Fig. 5**).

765

766 6.3.4. The '0.9 ka advance'

767 Moraine M1 at Etages Glacier is the only site where a pre-LIA frontal moraine has been
768 dated in the EPM. The age of 0.92 ± 0.02 ka (AD 1085 ± 20), which is constrained by two
769 consistent TCN ages and by independent evidence as well (see Section 5.2.4), agrees with a
770 deposition during a High Medieval Advance (HMA, c. 1100–1250 AD; **Nicolussi et al.,**
771 **2006**).

772 In the EPM, evidence for a HMA is also present in the sediment record of proglacial Lake
773 Muzelle (**Fouinat et al., 2017**; see **Fig. 1a** for location). In this lake, clastic sedimentation
774 increased towards the end of the MCA and peaked a first time at c. AD 1070-1080, in
775 remarkable agreement with our TCN age (**Fig. 7d**). After a drop centred on 1150, a larger
776 clastic excursion occurred between c. AD 1240 and 1305, marking the onset of the LIA (**Fig.**
777 **7d**). Overall, this pattern is very similar to other lacustrine detrital record from the French
778 Alps partly controlled by glacier size (**Fig. 7e**; **Arnaud et al., 2012**; **Guyard et al., 2013**).

779 Geomorphic evidence for a HMA has been identified at numerous sites in the Alps.
780 Although it does not represent a Neoglacial maximum, significant advances have been proven
781 for several large (> 10 km²) glaciers (see **Holzhauser, 2010** and **Le Roy et al., 2015**). When
782 accurate dendro-dates are available, they indicate advances that peaked during the early to late
783 12th century (between AD 1125 and 1186), depending on glacier response times. At Mer de
784 Glace, this advance is well-constrained between AD 1120 and 1178, an interval during which
785 the glacier grew from the AD 1890 level to the AD 1870 level (**Le Roy et al., 2015**). By AD
786 1100, the slow-reacting Aletsch Glacier was advancing over its AD 1926 extent (**Holzhauser**
787 **et al., 2005**). Because of its smaller size, Etages Glacier could have reached a maximum
788 slightly earlier than large glaciers, in response to a temperature drop spanning the second half
789 of the 11th century (**Büntgen et al., 2006**; **Millet et al., 2009**) and this position could have
790 been more advanced relative to large glaciers.

791

792 Conclusions

793

794 We presented the first numerical constraints on Neoglacial glacier advances in the French
795 Alps outside of the Mont Blanc Massif. Our results indicate that large advances, almost

796 similar to the LIA maximum, occurred repeatedly – at least five times – since c. 4.2 ka in the
797 Ecrins-Pelvoux massif.

798 Some advances dated here match the timing of widely dated Neoglacial advances in the
799 Alps, like the Löbben Advance Period, whose culmination has occurred at c. 3.6-3.5 ka, and
800 the Göschenen II advance, that peaked around 1.4 ka. The multi-phased nature of the LAP,
801 first identified in the central and eastern Alps, is confirmed here, as several moraines were
802 attributed to this period in the EPM.

803

804 Novel key findings are:

- 805 - The first direct dating of a moraine contemporary with the ‘4.2 ka event’ in the Alps,
806 an event recognized as a key marker for the subdivision of the Holocene (**Walker et**
807 **al., 2012**).
- 808 - The identification of an extensive advance occurring at the beginning of the Roman
809 Era, around 2.1 ka. To our knowledge, no other moraine ridges contemporary with this
810 advance were previously dated in the Alps, although evidence for concomitant glacial
811 activity is present in lateral moraine stratigraphy at least at two other sites, as well as
812 in some glacio-lacustrine records.
- 813 - The characterization of a High Medieval Advance that represents a Neoglacial
814 maximum during the Medieval Climate Anomaly is noteworthy as the magnitude of
815 this advance is smaller at the other sites where it was dated in the Alps.

816

817 Taking into account previous work carried out in the western Alps (**Le Roy et al., 2015**) c.
818 12 periods of glacier advance have now been identified during the Neoglacial with no more
819 than 5 centuries between two consecutive events.

820 The combination of multiple sites allowed us to identify more Neoglacial glacier advances
821 that this would have been possible at a single site. However, firm conclusions regarding the
822 unequivocal climatic origin of most of the glacier advances could not be drawn from our data
823 given that only one period of advance has been dated at two sites (LAP at Bonnepierre and
824 Rateau Glaciers). Nevertheless our dated advances are all consistent with an array of
825 paleoclimatic evidence (see Sections 6.1.1 and 6.3), which support their probable climatic
826 origin.

827 Finally, in the EPM (and most generally in the Alps), geomorphological evidence for
828 Neoglacial advances that pre-date the LIA are mostly preserved in lateral position in sectors
829 that were not overtopped during the LIA. This is due to the progressive channelization of ice

830 flow by lateral moraines in the course of the Neoglacial. Attention must be drawn on the
831 preservation of these landforms. In non-stabilized sectors, as erosion proceeds, some of the
832 ridges dated in this study will disappear in a near future due to moraine backwall retreat (e.g.
833 the innermost ridges at Bonnepierre Glacier). This will lead to a net loss of a valuable
834 paleoclimatic archive.

835

836

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852 References

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Fig. 1. (a) Location of the Ecrins-Pelvoux Massif (EPM) in the Alps. White areas: AD 2003 glacier extent (**Paul et al., 2011**). Glaciers (black font) and other locations (grey font) cited in the text are indicated ('Schwarz.' stands for Schwarzenstein Glacier). Western Alps glacierized massifs are shown with outlined font, MBM: Mont Blanc Massif, VA: Vanoise, BLD: Belledonne, GR: Grandes Rousses, UB: Ubaye, MA: Maritime Alps. Lower right inset: location of the Alps in Europe; **(b)** Shaded-relief map of the EPM with Little Ice Age (LIA) and present day (AD 2009) glacier extents (**Gardent et al., 2014**). Study sites are highlighted in red, and other EPM glaciers (5 to 7) and lakes cited in the text are indicated. The digital elevation model is an extract of the BD ALTI database from IGN (Institut national de l'information géographique et forestière). Eastings and northings are according to the IGN Lambert 93 grid in km.

Fig. 2. Geomorphological maps of the four study sites. **(a)** Rateau Glacier; **(b)** Lautaret Glacier; **(c)** Bonnepierre Glacier; **(d)** Etages Glacier. Contour interval is 50 m. Eastings and northings are according to the IGN Lambert 93 grid in m.

Fig. 3. ^{10}Be moraine boulder dating results. Exposure ages (yellow boxes) are reported as ka before AD 2000 rounded to the nearest decade and are grouped by moraine ridge. Errors are 1σ internal uncertainties (analytical errors only). **(a)** Rateau Glacier moraines MA to MD; **(b)** Lautaret Glacier moraine MA; **(c)** Bonnepierre Glacier moraines MA to ME; **(d)** Etages Glacier moraine M1. Orange box is the dendro-date of subfossil log eta01. Purple boxes show apparent ages derived from glaciogenic material (sands and gravels) collected by **Delunel et al (2014a)**. Etages Glacier map is not at the same scale. Eastings and northings are according to the IGN Lambert 93 grid in m.

Fig. 4. Field photographs summarizing moraine ages determined in this study (see **Fig. 3** for location of shooting places). Reported errors include PR uncertainty; **(a)** Rateau Glacier sampled ridges MA to MD. Ages indicated for MA to MC are 'apparent ages' based on the oldest boulder age per moraine (rather than mean ages) owing to discordant results at this site (see main text). Age for MD, also based on a single boulder, is thought to be the most reliable at this site (see main text). Non-sampled ridges MF to MH are also shown (ridge ME is obscured); **(b)** Lautaret Glacier composite left lateral moraine (left image). Sampled ridge MA is cross-cut by immediately inboard ridge MB. Right image: close-up of boulder LAU01 on moraine MA with sample location (white arrow). Notebook (30 cm-long) for scale; **(c)**

Bonnepierre Glacier sampled ridges MA to ME with mean ages. No mean age was computed for MD as only outliers were identified on this ridge. Moraine MC and MD are obscured; **(d)** Etages Glacier frontal ridges M1 to M6 and mean age for ridge M1 (left image). Red arrow: avalanche path; orange circle: sampling location of the subfossil detrital log eta01. Right image: close-up of moraine M1 with sampled boulders (white arrows).

Fig. 5. Eta01 subfossil log indexed tree-ring series (67 % spline) along with crossdating results computed with TSAP-Win software (**Rinntech, 2005**) on two distant references: the Eastern Alpine Conifer Chronology (**Nicolussi et al., 2009**) in red and the Mont Blanc massif chronology (**Le Roy et al., 2015**) in purple. Light blue intervals highlight periods of depressed growth. The close-up photograph of the wood section shows the eccentric growth pattern of this tree.

Fig. 6. Holocene paleoclimate proxies for the Alps. The period of rapid climate change between 4.4 and 4.0 ka, proposed as marking the onset of the Neoglacial period, is highlighted (light blue shading). **(a)** COMNISPA II, speleothem $\delta^{18}\text{O}$ composite record from Spannagel Cave (2340 m a.s.l.; western Austrian Alps; **Fohlmeister et al., 2013**); **(b)** Treeline variations in the Kauner Valley (western Austrian Alps; **Nicolussi et al., 2005**); **(c)** Radiocarbon and dendrochronological dating indicating smaller-than-present (AD 1990-2010) glacier extent in the Swiss and Austrian Alps (**Nicolussi and Patzelt, 2001**; **Joerin et al., 2006**; **2008**; **Drescher-Schneider and Kellerer-Pirklhauer 2008**; **Nicolussi and Schlüchter 2012**). Data falling into the 4.4-4.0 ka interval come only from the slow reacting Unteraar Glacier; **(d)** Advances of Gepatsch Glacier (western Austrian Alps) similar to – or beyond – its AD 1960 extent (**Nicolussi and Patzelt, 2001**); **(e)** Radiocarbon dating carried on subfossil soils and woods, indicating Holocene glacier maxima in the Swiss Alps (**Furrer et al., 1987**); **(f)** Upper Grindelwald Glacier (Bernese Swiss Alps) tongue elevation reconstructed from speleothem fabrics (**Luetscher et al., 2011**); **(g)** Summed Gaussian probability curves of TCN-dated Neoglacial moraines in the Ecrins-Pelvoux massif (this work). Thin white lines are individual boulder probability curves generated with external uncertainties (including PR uncertainty). Dotted thin lines represent outliers ages not included in moraine age calculation. Rateau Glacier dating are not displayed here, except moraine MD, because of discordant results (see main text). Summed curve for the LAP include mean ages of Bonnepierre Glacier moraines MB and MC and Rateau Glacier moraine MD (three moraines); **(h)** Terrigenous input into sub-alpine Lake Bourget (northern French Prealps; **Arnaud et al., 2012**); **(i)** Ledro

lake-level record (northern Italy; [Magny et al., 2012a](#)); **(j)** Cerin lake-level record (French Jura Mountains; [Magny et al., 2011](#)).

Fig. 7. Comparison of the timing of Neoglacial advances in the Ecrins-Pelvoux massif with selected proxies of glacial activity in the western Alps. **(a)** Sample depth of the dendro-dated subfossil trees retrieved from the Mer de Glace right lateral moraine (Mont Blanc massif; [Le Roy et al., 2015](#), supplemented). Replication peaks are interpreted as reflecting Neoglacial interstadial conditions and abrupt dying-off phases are interpreted as burial events caused by glacier advances; **(b)** Summed probability distribution of the calibrated radiocarbon dates from archaeological artefacts found at Schnidejoch Pass (2756 m a.s.l., Bernese Swiss Alps; [Hafner and Schwörer, 2017](#)) generated with Oxcal 4.3 ([Bronk Ramsey and Lee, 2013](#)). This dataset is thought to reflect contraction periods of the neighbouring Chilchli Glacier, allowing easier traverse of the pass; **(c)** Summed Gaussian probability curves of TCN-dated Neoglacial moraines in the EPM (this work). Thin white lines are individual boulder probability curves generated with external uncertainties (including PR uncertainty). Dotted thin lines represent outliers ages not included in moraine age calculation. Rateau Glacier dating are not displayed here, except moraine MD, because of discordant results (see main text). LAP: Lössen Advance Period; GII: Göschenen 2; HMA: High Medieval Advance. Summed curve for the LAP include mean ages of Bonnepierre Glacier moraines MB and MC and Rateau Glacier moraine MD (three moraines). Vertical dashed lines are moraine mean ages; **(d)** Proglacial Lake Muzelle clastic sedimentation ([Fouinat et al., 2017](#); [Fig. 1b](#)) based on the RABD (I-band) reflectance spectroscopy index ([Boldt et al., 2015](#)); **(e)** Proglacial Lake Bramant clastic sedimentation (2448 m a.s.l., Grandes Rousses massif; [Fig. 1a](#)). K/Ti anomalies are shown up to 800 AD (left axis); Ti anomalies are shown from 800 AD onwards (right axis). Both anomalies are considered proxies for the Saint Sorlin Glacier activity during these two time periods ([Guyard et al., 2013](#)).

Tab. 1. Basic topographic characteristics of the studied glaciers. Neoglacial and current (AD 2009) glacier surface data are taken from [Gardent et al. \(2014\)](#).

^a Neoglacial extent is considered to be similar to LIA extent (see main text)

^b Numbers in brackets represent the total area of the presently separated ice bodies that were connected during the Neoglacial maxima.

^c Current average slope is the weighted mean slope for these bodies (except for Rateau Glacier).

^d Debris-covered tongue is now separated from the main glacier body.

^e Calculated with an Accumulation Area Ratio (AAR) of 0.67 (0.4 for Bonnepierre Glacier) and 10 m contour interval.

^f Calculated with an Area Altitude Balance Ratio (AABR) of 1.6 (0.9 for Bonnepierre Glacier) and 10 m contour interval.

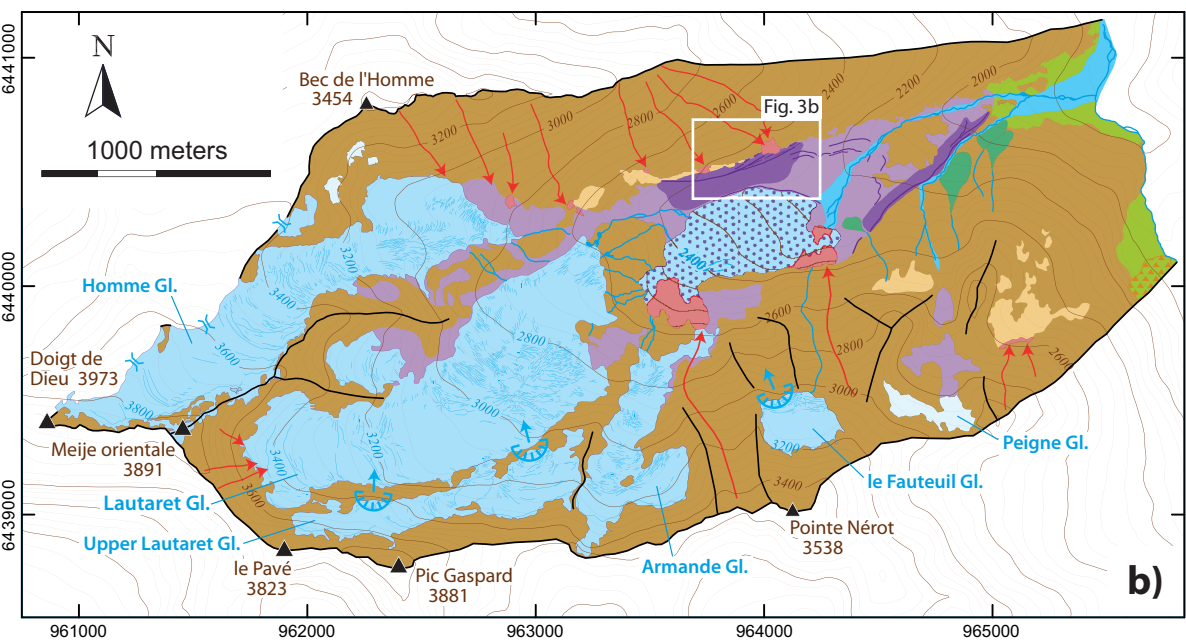
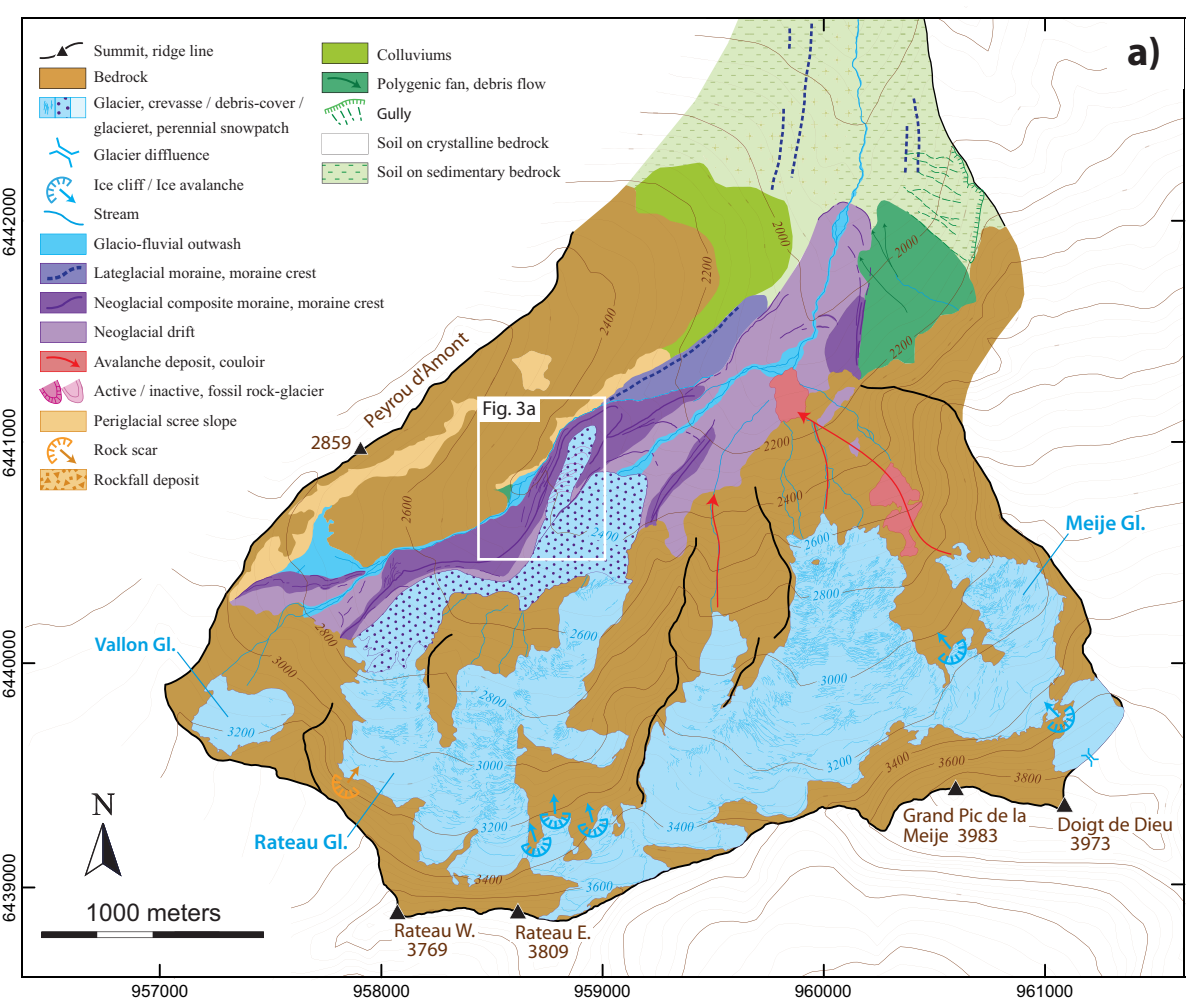
Tab. 2. ¹⁰Be sample characteristics.

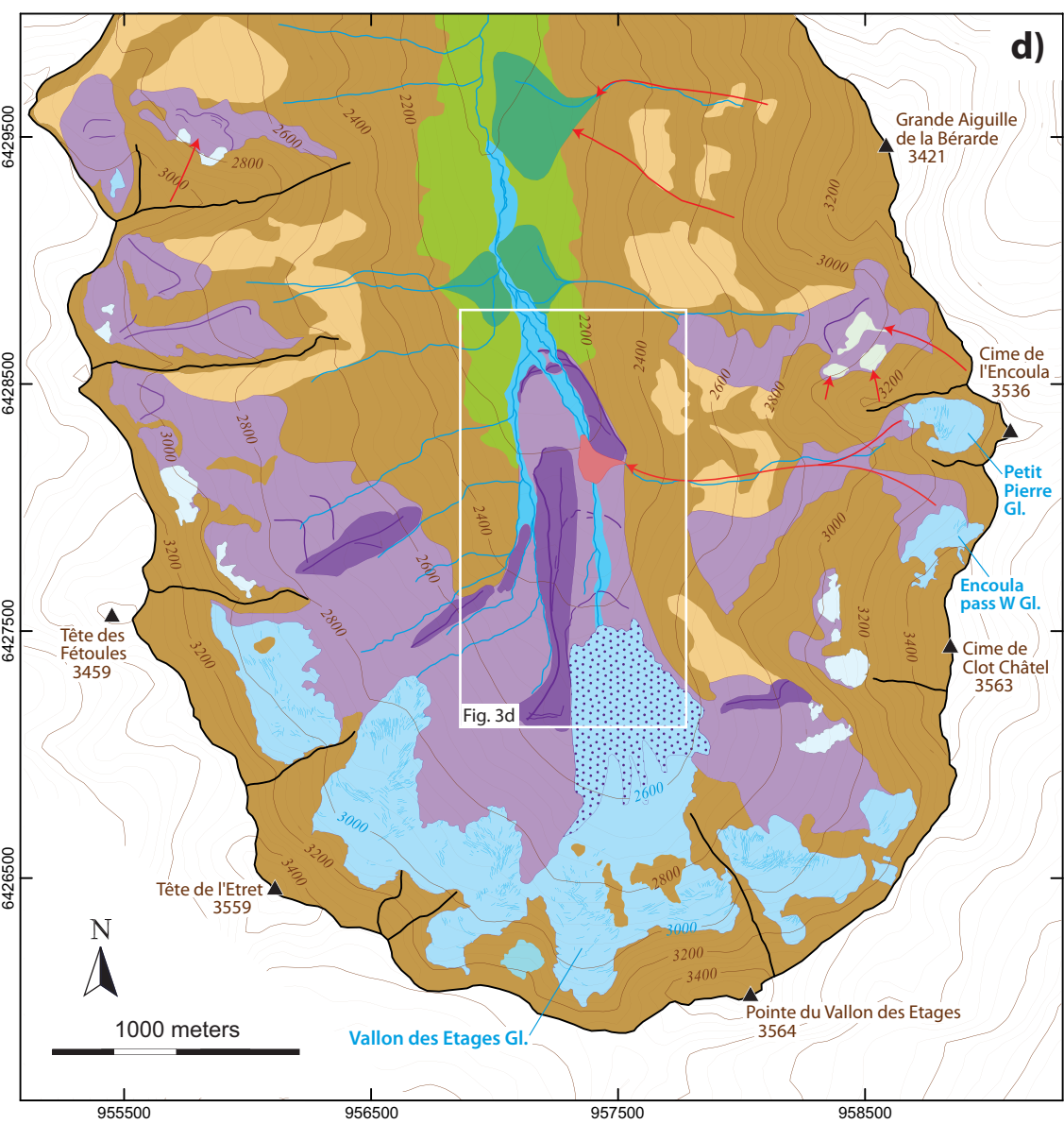
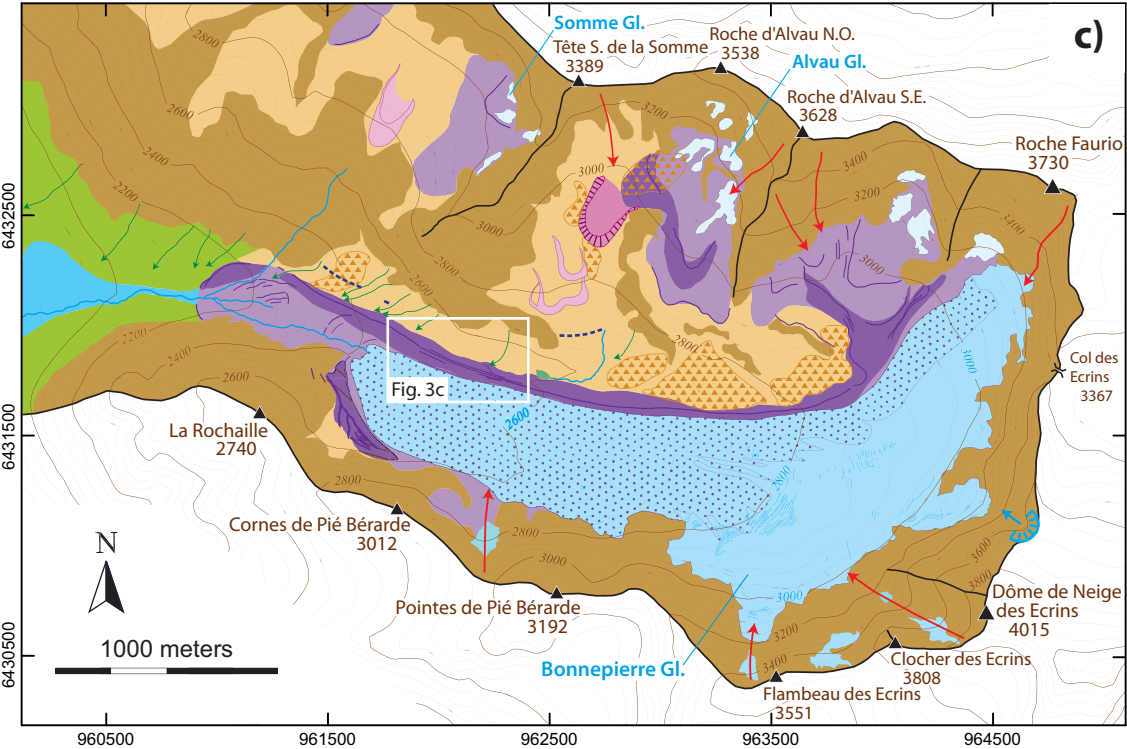
^a Rateau Glacier sample heights above the ground are the average of four point measurement per boulder (along and perpendicular to the long axis).

Tab. 3. ¹⁰Be sample exposure ages.

^a ¹⁰Be/⁹Be ratios of batch-specific analytical blanks used for the correction are $(1.4 \pm 0.8) \times 10^{-15}$ and $(2.0 \pm 1.0) \times 10^{-15}$ for samples marked with (†), $(2.0 \pm 0.9) \times 10^{-15}$ for samples marked with (‡), $(2.8 \pm 1.3) \times 10^{-15}$, $(2.5 \pm 1.1) \times 10^{-15}$ and $(2.7 \pm 1.2) \times 10^{-15}$ for samples marked with (§) and $(9.7 \pm 1.3) \times 10^{-15}$ for samples marked with (#). When multiple blanks have been processed within a single batch, mean value was used for the correction.

^b Age calculations use density values of 2.54 and 2.62 g cm³ for granite and gneiss, respectively (**Delunel, 2010**) and zero erosion. ‘LSD/Chironico’ exposure ages are calculated with the CREp program (**Martin et al., 2017**) using the LSD scaling model (**Lifton et al., 2014**) and the Chironico landslide production rate (**Claude et al., 2014**). ‘Lm/Arctic’ exposure ages are calculated with the CRONUS-Earth online calculator (version 2.2; **Balco et al., 2008**) using the time-dependent Lal/Stone scaling model (**Lal, 1991; Stone, 2000**) and the ‘Arctic’ production rate (**Young et al., 2013**).





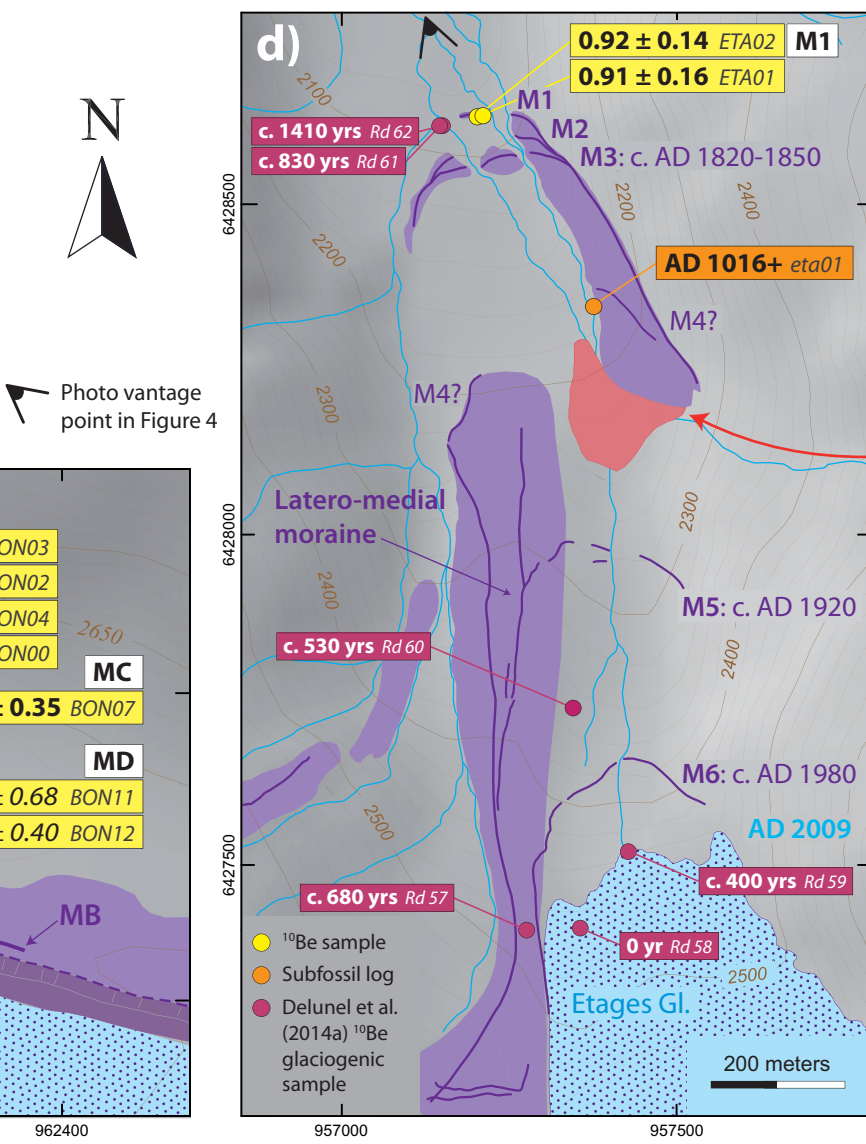
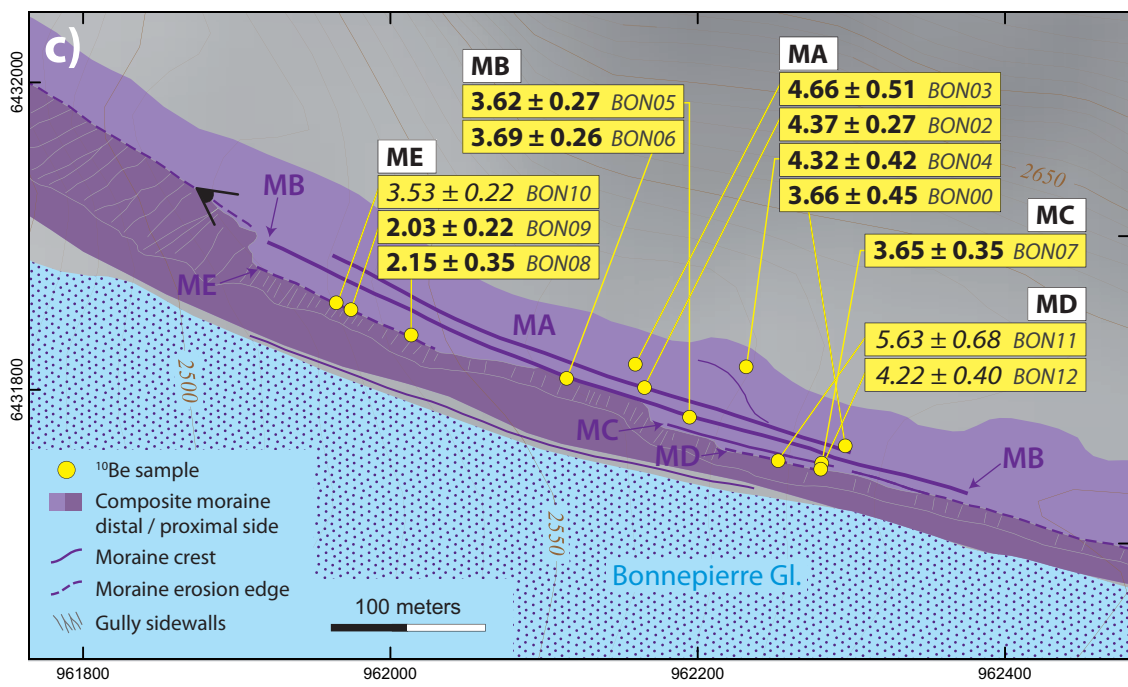
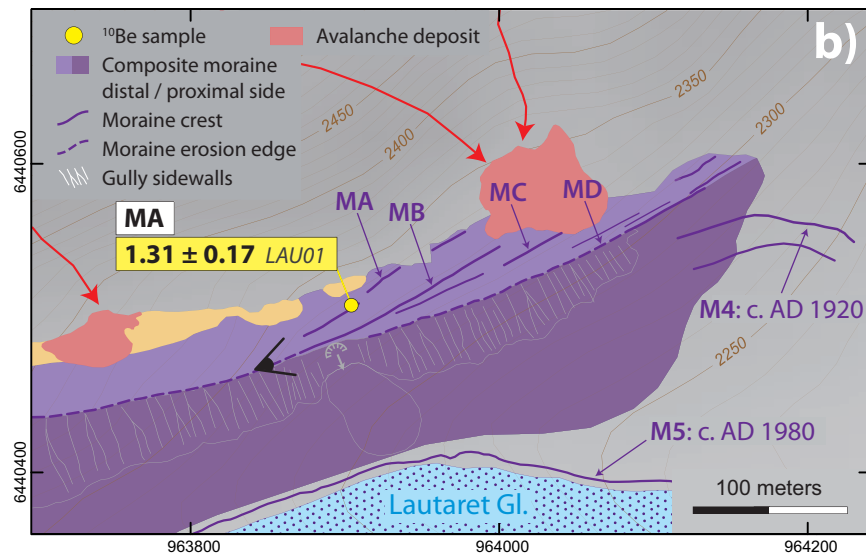
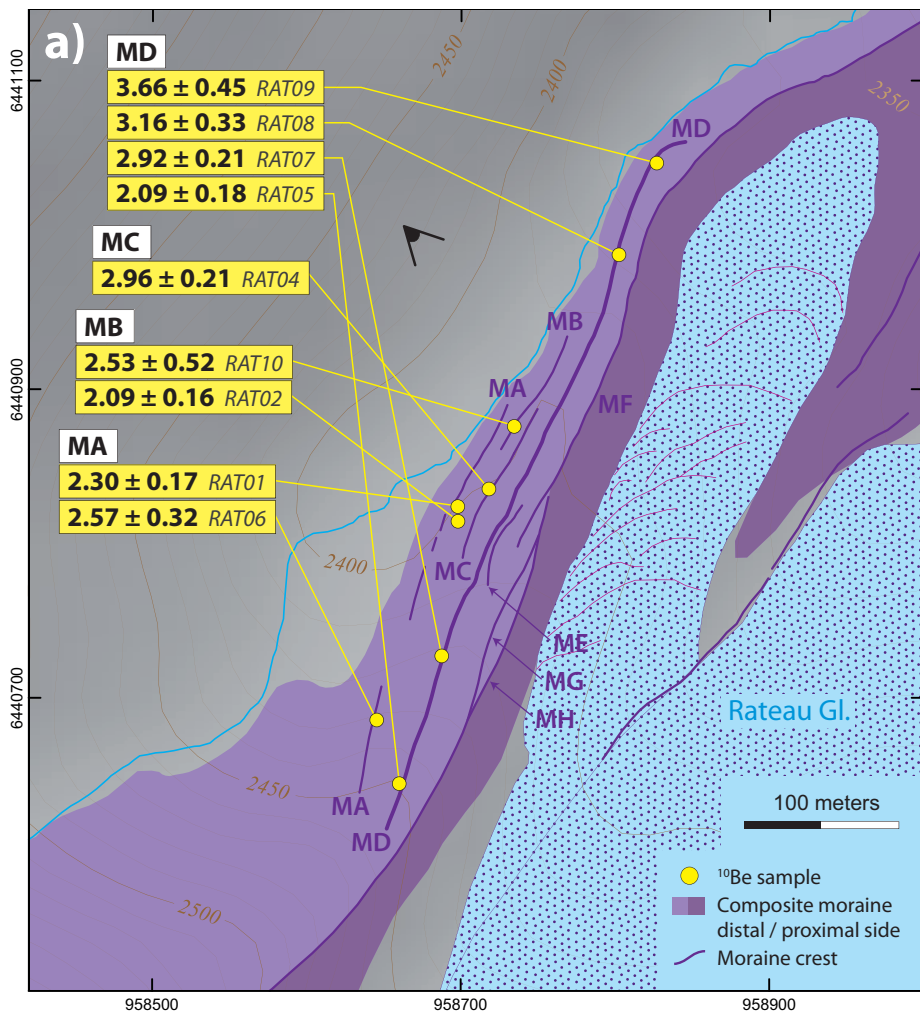
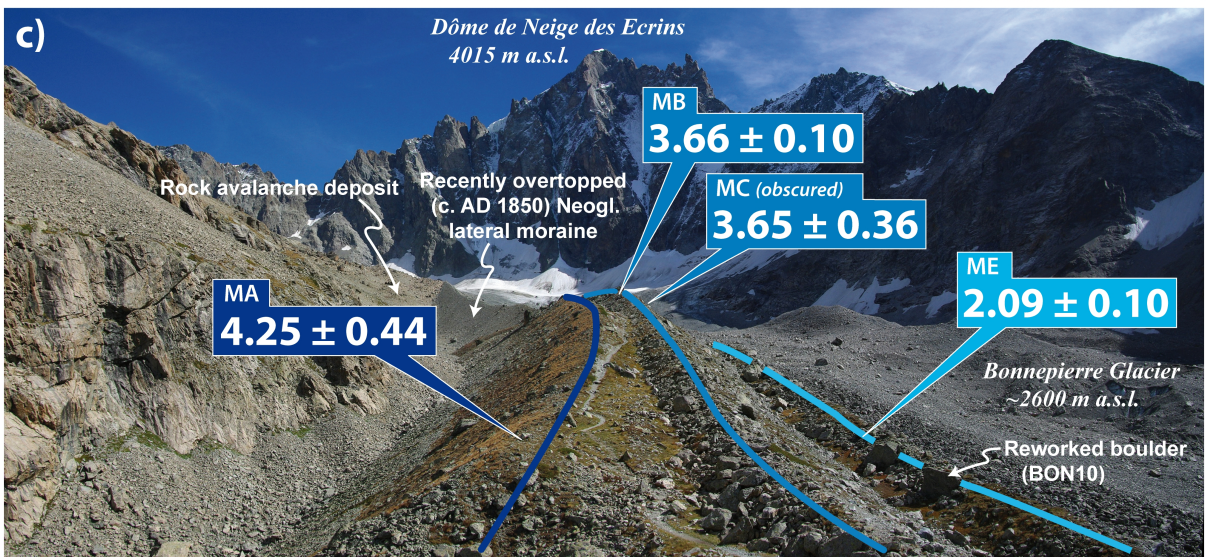
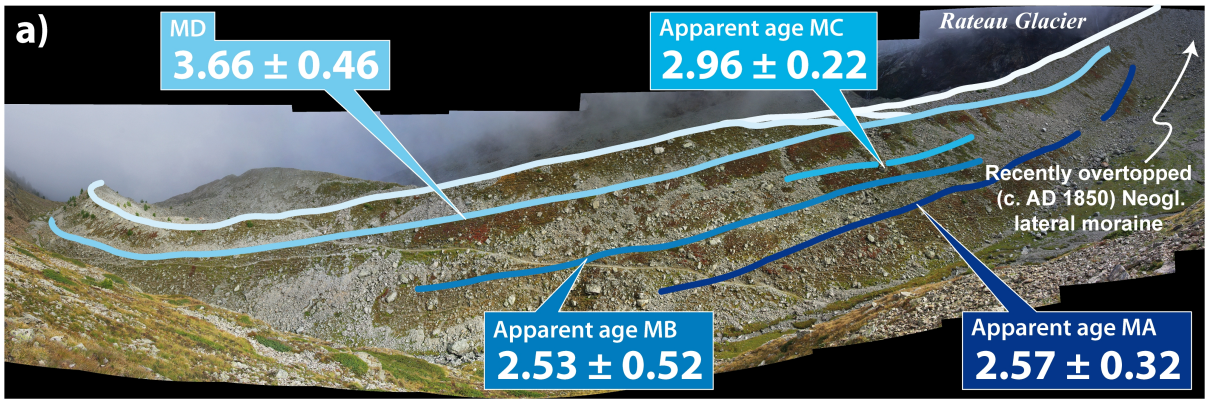
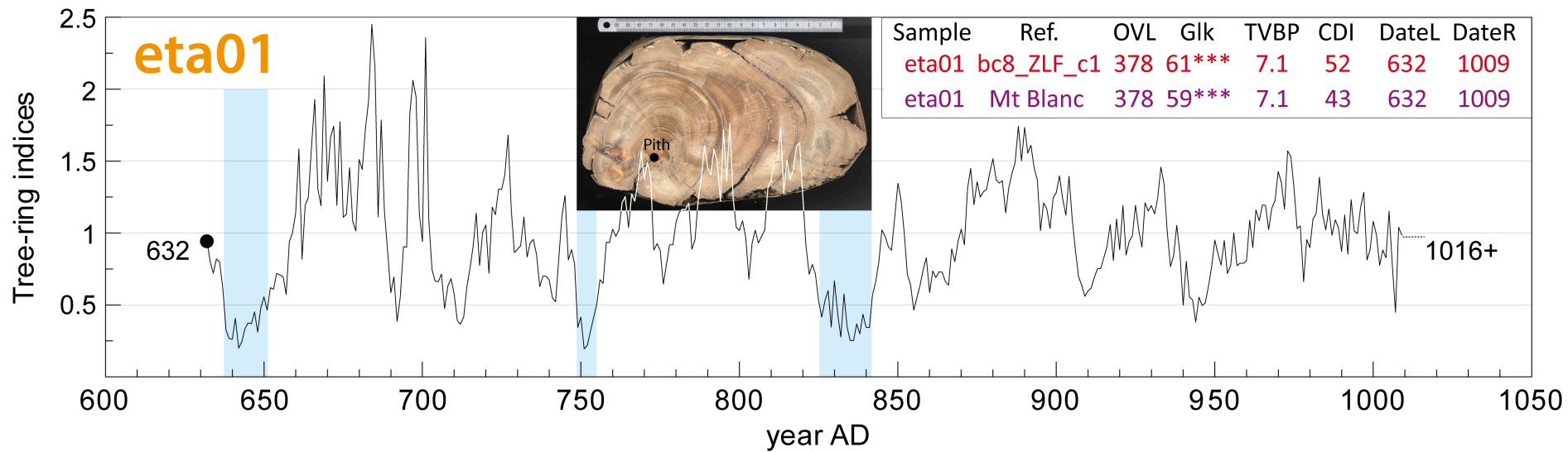
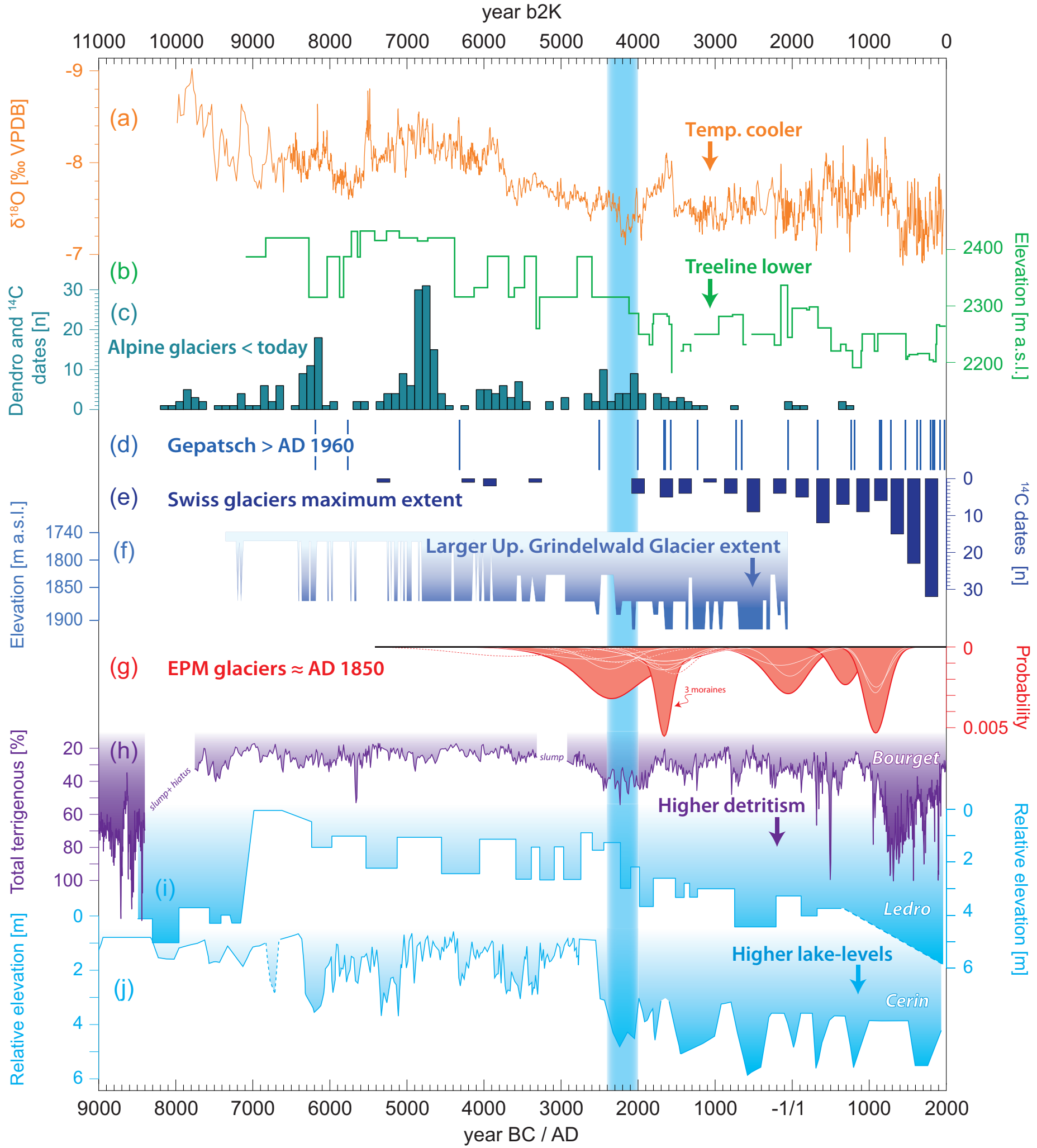
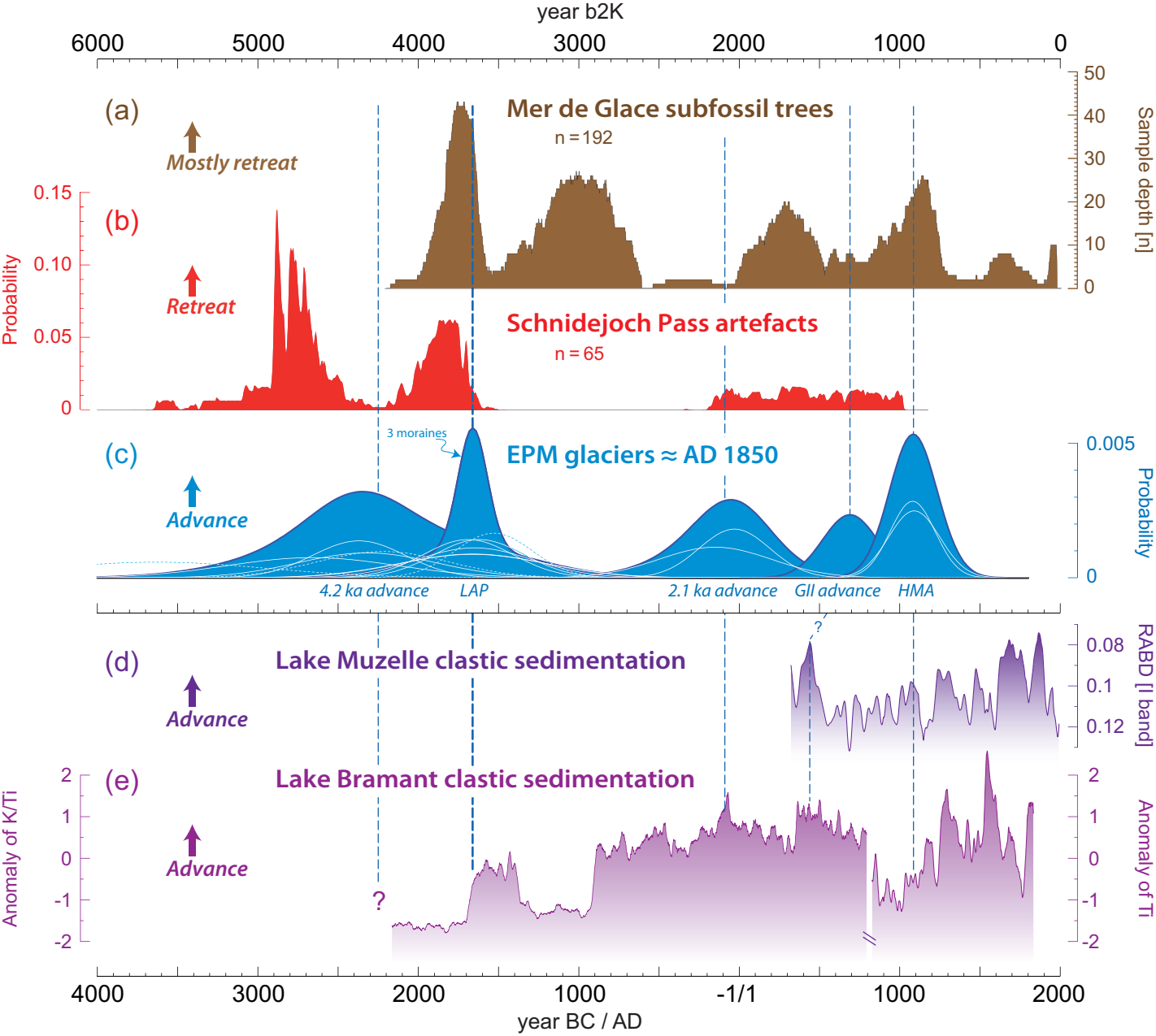


Photo vantage point in Figure 4









	Neogl. area (km ²) ^a	Current area (km ²) ^b	Current altitudinal range (m a.s.l.)	Neogl. front (m a.s.l.)	Current average slope (°) ^c	Current aspect (°)	Debris cover (% 2009 area)	MELM	AAR Neogl. ELA (m a.s.l.) ^e	AAR early 1980s ELA (m a.s.l.) ^e	AABR Neogl. ELA (m a.s.l.) ^f	AABR early 1980s ELA (m a.s.l.) ^f
Rateau / Meije	4.83	1.17 (2.95)	3502-2301	1870	34.2	N (13)	28	2760	2651	2834	2761	2914
Lautaret	3.36	0.88 (2.21)	3806-2234	1910	31.0	NE (50)	27 ^d	2540	2737	2832	2847	2932
Bonnepierre	2.62	1.94 (1.97)	3264-2412	2115	20.1	W (287)	54	2880	2862	2876	2792	2816
Etages	3.56	1.12 (1.49)	3224-2431	2095	30.7	N (13)	23	2700	2697	2784	2717	2794
									2737 ± 91	2832 ± 38	2779 ± 55	2864 ± 69
									+ 95 m		+ 85 m	

Sample	Latitude (°N)	Longitude (°E)	Elevation (m a.s.l.)	Height above ground (cm) ^a	Lithology	Thickness (cm)	Shielding factor
RAT01	45.018202	6.285005	2399	96	Coarse-grained granite	1.5	0.922
RAT02	45.018125	6.285001	2401	71	Coarse-grained granite	2	0.922
RAT04	45.018282	6.285259	2403	119	Cataclased granite	2.5	0.922
RAT05	45.016606	6.284437	2432	65	Coarse-grained granite	4	0.924
RAT06	45.016990	6.284280	2414	59	Leucocratic gneiss	1.5	0.921
RAT07	45.017344	6.284831	2415	62	Coarse-grained granite	2.5	0.924
RAT08	45.019633	6.286424	2370	70	Coarse-grained granite	2	0.903
RAT09	45.020154	6.286780	2363	115	Massive amphibolite	2.5	0.903
RAT10	45.018670	6.285508	2393	62	Cataclased and chloritised coarse-grained granite	2.5	0.922
LAU01	45.013394	6.350837	2343	110	Fine-grained granite	1.75	0.848
BON00	44.935358	6.325778	2591	230	Aplitic granite	1.75	0.954
BON02	44.935748	6.324149	2581	180	Migmatised gneiss	4	0.955
BON03	44.935888	6.324073	2577	280	Anatectic granite	5	0.955
BON04	44.935833	6.324965	2573	310	Migmatised gneiss	2.5	0.932
BON05	44.935540	6.324488	2584	130	Fine-grained granite	2	0.955
BON06	44.935798	6.323495	2574	150	Fine-grained granite	2.5	0.955
BON07	44.935262	6.325607	2590	110	Migmatised gneiss	2	0.951
BON08	44.936101	6.322237	2553	180	Fine-grained granite	2.5	0.951
BON09	44.936275	6.321760	2545	190	Fine-grained granite	3	0.949
BON10	44.936307	6.321639	2543	190	Migmatised gneiss	2	0.949
BON11	44.935287	6.325250	2587	110	Oriented granite	1.3	0.955
BON12	44.935227	6.325597	2589	85	Anatexic granite	2	0.951
ETA01	44.909085	6.259764	2090	70	Deformed granite	2	0.920
ETA02	44.909090	6.259700	2090	65	Deformed granite	3.5	0.920

Sample	Quartz (g)	Be carrier (mg)	$^{10}\text{Be}/^9\text{Be}$	AMS measured error (%)	$[^{10}\text{Be}]$ (at.g $^{-1}$) ^a	Error (at.g $^{-1}$)	^{10}Be age 'LSD/Chironico' (a) ^b	External uncertainty (a)	^{10}Be age 'Lm/Arctic' (a)	External uncertainty (a)
RAT01	38.3085	0.2897	1.14043E-13	5.1	56800 [†]	3800	2310	180	2390	190
RAT02	24.3267	0.2953	6.61407E-14	4.6	52100 [‡]	3600	2100	160	2200	170
RAT04	29.6413	0.2981	1.09016E-13	5.0	71900 [‡]	4900	2970	220	3040	240
RAT05	49.7081	0.2986	1.32735E-13	6.1	52600 [†]	4100	2100	190	2210	190
RAT06	49.7575	0.2953	1.61997E-13	10.3	63600 [†]	7300	2580	320	2650	320
RAT07	27.7882	0.3003	1.00990E-13	5.1	71500 [‡]	5100	2930	230	2990	240
RAT08	37.3874	0.2893	1.44004E-13	8.9	73600 [†]	7600	3170	330	3220	350
RAT09	22.4590	0.2924	9.88726E-14	12.4	84500 [†]	10300	3670	460	3720	480
RAT10	50.5087	0.2963	1.58371E-13	19.5	61400 [†]	12000	2540	520	2610	520
LAU01	45.9494	0.3005	7.03952E-14	11.0	29600 [§]	3700	1320	170	1410	180
BON00	45.2676	0.2983	2.41154E-13	11.1	105000 [§]	12900	3670	450	3750	480
BON02	51.7196	0.2956	3.21659E-13	3.7	122200 [†]	7400	4380	290	4450	320
BON03	49.6532	0.2993	3.22216E-13	10.1	129100 [†]	14500	4670	530	4750	560
BON04	36.4583	0.2986	2.18606E-13	8.5	118700 [†]	11600	4330	430	4400	460
BON05	30.8571	0.2883	1.67223E-13	5.8	103300 [†]	7600	3630	280	3710	310
BON06	36.1911	0.3148	1.80638E-13	5.0	104000 [†]	7300	3700	280	3770	300
BON07	16.9800	0.3051	9.66976E-14	8.3	104500 [#]	9800	3660	360	3760	380
BON08	40.2259	0.2970	1.26189E-13	13.9	60900 [§]	9200	2160	350	2270	350
BON09	40.0585	0.3028	1.15729E-13	8.6	57100 [§]	5800	2040	220	2150	230
BON10	44.3182	0.2955	2.21443E-13	4.2	97500 [§]	6200	3540	240	3630	270
BON11	13.7900	0.3037	1.21111E-13	12.2	164000 [#]	21800	5640	690	5790	800
BON12	28.0600	0.3023	1.76846E-13	8.9	120300 [#]	11300	4230	410	4310	440
ETA01	46.9355	0.3017	4.46136E-14	16.7	18400 [†]	3100	920	160	980	170
ETA02	44.1835	0.3005	4.20809E-14	16.0	18400 [†]	2800	930	140	980	160