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1 Late Holocene initiation of a deep rock slope failure in an Alpine 2 valley revealed by ^{10}Be surface exposure dating (Chamonix, France)

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11 Abstract:

12 We studied a newly identified, multiple-kilometer-long rock slope failure in the Aiguilles Rouges
13 massif (Chamonix valley, France). Owing to a high-resolution light detection and ranging (LiDAR)
14 digital elevation model (DEM) and field work, we mapped morphostructures, including scarps, open
15 fractures, and counterscarps. In some places, vertical offsets can reach tens of meters and crevasses
16 can be meters wide. The evidence of gravitational activity (boulder displacements from analyses of
17 archival satellite images) and the sharpness of the scarp outcrops together suggest very recent
18 movements. These observations agree with ground displacement rates of a few millimeters per year
19 estimated by interferometric synthetic aperture radar (InSAR) time series between 2014 and 2018.
20 We sampled two vertical profiles along the top scarps to define the chronology of the slope failure
21 using beryllium-10 (^{10}Be) surface exposure dating. Glacially polished surfaces cut by these
22 gravitational scarps were also sampled to determine glacial retreat timing as well as to constrain the
23 pre-exposure ^{10}Be inheritance. In total, 11 samples were studied.
24 Our results highlight a significant time lag (approximately 15 ka) between the first evidence of
25 nonglacial activity and the initiation of the slope failure that happened 1.3–2.5 ka ago, depending on

26 the inheritance schemes. This suggests that the delayed opening of the crevasse is only one stage of
27 a process that began when the valley was deepened and the glacial debuitressing is not the unique
28 driving factor. This process of progressive failure of an excessively steep slope may continue, and the
29 evolution of this slope failure may constitute a hazard for the upper part of the Chamonix valley.

30 Keywords: Alpine glacial valley; Rock slope failure; Cosmogenic radionuclide dating; Natural hazard;
31 Chamonix valley

32 **1. Introduction**

33 Shapes of alpine valleys have been deeply influenced by glacial cycles involving glacier retreats and
34 advances over the Quaternary period (Harbor et al., 1988; Herman et al., 2011). These processes
35 have formed U-shaped valleys with steepening slopes that are rejuvenated by gravitational processes
36 (Le Roux et al., 2009; Crosta et al., 2013; Riva et al., 2018). Glacial debuitressing and mechanical
37 stress release have often been mentioned as causes that favor slope instabilities that can lead to
38 large scale landslides or rock slope failures (Holm et al., 2004; Bigot-Cormier et al., 2005; Cossart et
39 al., 2008; Ballantyne et al., 2014). Predisposing factors, such as lithology, rock mass characteristics, or
40 geologic structures, play an important role in rock slope initiation and mountain slope evolution.
41 There are many potential triggering factors for gravitational structures in mountainous regions,
42 sometimes working together, such as exceptional rainfall episodes (Guzzetti et al., 2004; Pothérat
43 and Effendiantz, 2009), seismic activity (Chen et al., 2005; Jibson et al., 2006), changes in
44 groundwater regime (Pisani et al., 2010), or permafrost degradation (Raveland and Deline, 2011). All
45 these factors control the spatial and temporal evolution of the instabilities, the landscape
46 morphologies (Crosta et al., 2013), and the risks these instabilities represent for human communities
47 and infrastructures.

48 Rock slope failures (RSFs) such as rockslides, rock avalanches, or rockfalls, are common hazards in
49 mountainous areas. Due to climate changes and thawing permafrost, the frequency and severity
50 increase at high altitude and catastrophic events can occur (e.g. Raveland and Deline, 2011).

51 Conversely, deep-seated gravitational slope deformations (DSGSD) or slow-moving rockslides
52 generally do not seem to directly threaten humans. They evolve at slow displacement rates (several
53 mm yr⁻¹) over long periods and are characterized by specific geomorphological features such as
54 counterscarps, double ridges, trenches, and tension cracks (El Bedoui et al., 2009; Le Roux et al.,
55 2009). DSGSDs may occur with a time lag relative to glacier disappearance (Hormes et al., 2008; Ivy-
56 Ochs et al., 2009; Prager et al., 2009b), supporting the hypothesis that even when the glaciers retreat
57 they undergo progressive failure in subcritical stress conditions (Eberhardt et al., 2004; Brideau et al.,
58 2009; Riva et al., 2018). Although they can appear stable or metastable, they can be reactivated and
59 record sudden accelerations (see a review in Pánek and Klimeš, 2016). Understanding the long-term
60 evolution and identifying the triggers of such structures are thus keys to estimating potential hazards
61 in the near future.

62 Geochronological investigations allow the understanding of the temporal evolution of large RSFs (e.g.
63 Pánek, 2015; Pánek and Klimeš, 2016). In the last decade, cosmic ray exposure (CRE) dating method,
64 such as with *in-situ* produced beryllium-10 (¹⁰Be), has been widely applied either to boulders in
65 landslide debris deposits (e.g. Zeng et al., 2020) or along headscarp walls (e.g. Böhme et al., 2019). For
66 instance, these applications revealed successive failure phases of RSFs on single site but with time lag
67 between each event reaching several thousand years (e.g. Delgado et al., 2020), synchronicity of
68 slope failures with past climate changes (e.g. Sepulveda et al., 2022), long-term progressive failure
69 leading to landslide slope rate increase (e.g. Schwartz et al., 2017), among others.

70 In this study, we investigated a large RSF recently identified in the Aiguilles Rouges massif (Chamonix
71 valley, western European Alps, France). We used the ¹⁰Be dating method to establish its long-term
72 evolution. We also performed field mapping supported by aerial photograph analyses and the
73 interpretation of the high-resolution light detection and ranging (LiDAR)-derived digital elevation
74 model (DEM) to determine the geomorphology of the area and the relationship with RSF landforms.
75 This paper has the following four aims: (i) to constrain the timing of the initiation of the RSF, (ii) to

76 explore the link with local glacier extension, (iii) to identify the driving and triggering factors of the
77 RSF, and (iv) to evaluate the potential threat it represents for the downslope population.

78 **2. Settings**

79 **2.1 Geographic context**

80 The Aiguilles Rouges massif (ARM) is located on the northwestern side of the Chamonix valley
81 (western European Alps, France), west of the Mont Blanc massif (4808 m above sea level, a.s.l.) and
82 is part of the alpine external crystalline massifs (Fig. 1). The ARM stretches over 15 km from
83 northeast to southwest and goes from 1000 m a.s.l. near Chamonix to 2965 m a.s.l. at the Aiguille du
84 Belvédère. It includes the 3279-ha Aiguilles Rouges Nature Reserve created in 1974. The Chamonix
85 valley is highly populated and plays host to numerous touristic and economic activities including an
86 important railway line and an international road linking France to Italy and Switzerland.

87 Our study site, called the Argentière crevasse (Fig. 1), is located in the Chéserys area, upslope of the
88 Argentière village.

89 The massif is divided into three parts. The upper part of the Reserve is characterized by vertical walls
90 and numerous remnant glaciers and alpine lakes. Then, the rhododendron moor marks the transition
91 between the forest and alpine meadows. A dense spruce forest covers the steep slopes of the lower
92 part of the Reserve. In terms of geomorphology, the landscapes of the upper slopes have mainly
93 been shaped by glacial erosion and paraglacial processes, whereas screes and rockfall deposits from
94 periglacial processes are major components of the lower slopes. Fallen rocks can sometimes reach
95 more than 50 m across. One of these gravitational structures has been known since the 19th century
96 as the iconic Aiguillette d'Argentière (1893 m a.s.l.), a rock slice of approximately 25 m high
97 separated from the rockwall (Fig. 2)

98 On the landscape, trimlines (Fig. 1) indicate that the area was covered by ice during the Last Glacial
99 Maximum (approximately 21 ka ago; Clark et al., 2009) at approximately 2600 m a.s.l. (Coutterand

100 and Buoncristiani, 2006). Glacier thinning could have occurred very quickly after 16.6 ± 0.6 ka ago
101 (Lehmann et al., 2020). Regional synthesis, mainly based on relative geomorphological analysis and
102 relative chronology, suggest that the Col des Montets (1461 m a.s.l.) was free of ice before the
103 Younger Dryas (Jaillet and Ballandras, 1999). Using ^{10}Be cosmogenic dating on the frontal moraine of
104 the Argentière glacier confirms that the glacier was restricted to the bottom of Chamonix valley at ca.
105 12 ka ago (Protin et al., 2019) and lost its downslope continuity.

106 On the southern slopes of the ARM, small cirque glaciers were nonetheless present until the Younger
107 Dryas (Nicoud et al., 2005) and even to approximately 8 ka ago as shown by ^{10}Be cosmogenic dating
108 of glacial polish (Cara et al., 2017). Numerous small lakes and bogs have set up in the glacial
109 overdeepening. Furthermore, rock glaciers and local moraines suggest that small glaciers developed
110 during the Little Ice Age in the upper part of the two sides of the ARM. It is still glacierized above
111 2400 m a.s.l. with five small glaciers ($< 0.1 \text{ km}^2$), and locally permafrost-affected rockwalls (Magnin et
112 al., 2015). Figure 3 shows the entire study area seen from the Glacier de Lognan, on the other side of
113 the valley.

114 **2.3 Geology and tectonics:**

115 The ARM is mostly composed of metamorphic rocks such as gneiss, mica schist, and granite (von
116 Raumer, 1987) and is crossed by the active Vallorcine fault (Fig. 1). This fault is one of the major
117 tectonic structures of the area. It extends between the Rhône River and Chamonix and regularly
118 reactivates. The Vallorcine fault is the geological contact between the Aiguilles Rouges gneiss and the
119 Vallorcine granite. It was presumably the seismic location of the M_w 5.3 earthquake that affected the
120 area in 1905 (Cara et al., 2017) and the Vallorcine seismic swarm is running along the fault (Ritz et al.,
121 2021).

122 The Remuaz fault (Fig. 4a) extends from the southern end of the Vallorcine fault for approximately 15
123 km between the Col des Montets and the Chéserys area. It is an active fault with a dominant dextral
124 strike-slip component compatible with a tectonic origin for the postglacial deformation as suggested

125 by shifted glacial landforms (Ritz et al., 2021). The southern part of the Remuaz fault is oblique to the
126 Arve valley. The southwestern end disappears under a scree on the steep slopes and possibly under
127 sedimentary rocks in the Chamonix valley.

128 Several gravitational scarps have been mapped by the Bureau de Recherches Géologiques et
129 Minières (BRGM; Pairis et al., 1992) and some have been interpreted as deep-seated RSFs by
130 Blondeau (2018) and Blondeau et al., 2021. Furthermore, displacement rates between 2014 and 2018
131 were estimated over the western Alps using interferometric synthetic aperture radar (InSAR) time
132 series from 130 Sentinel-1 acquisitions and 800 interferograms (André et al., 2021). These satellite
133 data suggest a present-day downhill creeping in some portions of the steep slope of the ARM, but
134 more work is needed to precisely estimate a motion value. Nonetheless, there has never been a
135 quantitative study about the geometry of the geological objects and rates of the gravitational
136 processes, which is the purpose of this work. To avoid the complexity related to the spatial
137 superposition of active tectonic and gravitational features, we focus on the gravitational features
138 located downslope of the clearly expressed part of the Remuaz fault (Fig. 4a). Our study site is
139 located within the gneiss between the slightly flat area and the valley slope.

140 **3. Methods**

141 **3.1 Geomorphologic and structural methods**

142 The general strategy is based on three methods as follows: 1) A cartographic approach using aerial
143 images acquired and orthorectified by Institut national de l'information géographique et forestière
144 (IGN) with a resolution of 0.2 m and a high-resolution DEM from the Institut National de la Recherche
145 Agronomique (INRAe) made by airborne LiDAR in 2012 with meter-scale resolution. 2) A visual
146 reconnaissance from a helicopter that also allowed the acquisition of high-resolution photos that
147 were used to make a local photogrammetric three-dimensional (3D) model using Agisoft LLC
148 Metashape software (professional edition, version 1.8) and reached a resolution < 0.5 m (see the
149 result here: [Buoncristiani, 2020](#)). Cross sections derived from this DEM allowed the precise

150 determination of gravitational structures' geometries. 3) Field work to determine structural
151 measurements (orientation and dip of the fractures as well as amplitude of displacement along the
152 fractures). For open fractures (called crevasses hereafter), we chose to measure the total
153 displacement vector from characteristic irregularities located both at the footwall and at the hanging
154 wall and to define this vector by its norm (total displacement measured with a laser range finder), its
155 orientation with respect to north in the horizontal plane, and its dip (Fig. 5). The total displacement is
156 composed of two distinct vertical and horizontal components. If the dip is greater than 45°, the
157 magnitude of the vertical component is greater than that of the horizontal component. Furthermore,
158 the upslope wall of the crevasse forms a clear morphological scarp if the dip of the displacement
159 vector is greater than the dip of the topographic slope (α). The crevasse's strike and dip (θ) are mean
160 values estimated from aerial photographs and field measurements. The opening is estimated in a
161 horizontal plane perpendicular to the crevasse orientation.

162 **3.2 Cosmogenic dating**

163 To constrain the timing of both glacier retreat and initiation of the RSF, we performed CRE dating
164 using *in-situ* beryllium-10 (^{10}Be). In the southern part of the ARM, the dominant lithology favored the
165 use of ^{10}Be in quartz. Sampling locations are given in Figure 4a–d and the characteristics of the
166 sampled surfaces are given in Table 1. In total, 11 rock samples of approximately 0.5 kg to 1 kg were
167 extracted. Two types of morphologies were sampled: a near-vertical scarp of gravitational origin (9
168 samples) and glacially polished surface (2 samples). The two glacially polished surfaces are located at
169 the hanging wall (AIG20-1, top of the scarp) and the footwall of the sampled gravitational scarp
170 (AIG20-7).

171 Samples were prepared at the geo-thermo-chronology (GTC) platform [Institut des Sciences de la
172 Terre (ISTerre) lab, Grenoble, France] following a procedure adapted from Brown et al. (1991) and
173 Merchel and Herpers (1999); see Appendix A. Then, $^{10}\text{Be}/^9\text{Be}$ ratios were measured at the

174 Accelerator Mass Spectrometry facility (ASTER), CEREGE, Aix-en-Provence, France (see Appendix A
175 and references therein).

176 Exposure ages were calculated using the online CREp calculator (Martin et al., 2017). We used a
177 global averaged time-dependent sea level high-latitude (SLHL) ^{10}Be production rate of 4.09 ± 0.19
178 atoms $\text{g}^{-1} \text{yr}^{-1}$ (Martin et al., 2017). As most of the samples are from a south-exposed slope steeper
179 than 45° , preventing the persistence of snow, ages were computed without considering snow
180 shielding and assuming a negligible denudation. Results are presented in Table 3.

181 Exposure ages of samples extracted along the gravitational scarp should be considered as the
182 maximum apparent ages because a significant inheritance may have been accumulated at depth
183 before direct scarp exposure (e.g., Zerathe et al., 2013, 2014). Indeed, an inherited component of
184 cosmogenic radionuclide concentration acquired before direct exposure should be considered for
185 surface samples as well as at depth (e.g., Benavente et al., 2017; Schwartz et al., 2017; Hilger et al.,
186 2019; Břežný et al., 2021). In the present case, the sampled gravitational scarp affects a glacially
187 polished surface (Fig. 5) leading to a basic set of three hypotheses for inheritance calculation. 1) At
188 the time of the glacier retreat, the polished surface is free of cosmogenic nuclides due to the
189 previous strong ice abrasion. 2) Cosmogenic production starts on polished surface through direct
190 exposure (t_0), as well as at depth due to the penetration of high-energy particles (neutrons and then
191 muons; Braucher et al., 2013). A large quantity of inherited nuclides can form before scarp exposure
192 if the slip event is significantly delayed with respect to glacier retreat. 3) Finally, when the slip starts
193 (t_1), the scarp is directly exposed to cosmic rays and accumulates an additional quantity of
194 cosmogenic nuclides. Deciphering the different steps and the ^{10}Be accumulation component is based
195 on glacier retreat timing (t_0) deduced from surface samples. An inheritance can be calculated using
196 the classic equation (see Appendix B) used for calculation of *in-situ* cosmogenic nuclide accumulation
197 (Lal, 1991) as described by Břežný et al. (2021) and considering an integration time $t_{\text{int}} = t_0 - t_1$ and
198 an estimated depth, determined as follows:

199
$$Z_E = \frac{\sin \alpha \times L}{\sin \theta}, \quad (1)$$

200

201 where α is the dip of the topographic slope, L the distance between the sample and the top of the
202 scarp, and θ is the dip of the crevasse. Finally, this inherited ^{10}Be concentration is subtracted from
203 the measured concentration (Table 3). Corrected ages are calculated as previously described using
204 the same parameters on the online CREp calculator (Martin et al., 2017).

205 **4. Results**

206 **4.1. Geomorphology and structural geology**

207 The tectonic structures are characterized by a rather planar pattern (Fig. 4a) and the Remuaz
208 fault is a subvertical fault dipping strongly to the southeast. A second fault, which we call the
209 Aiguillette fault, is trending N52° and dipping 79° to the northwest. Downslope of the Remuaz fault,
210 we have identified a rock crevasse that is several kilometers long (Fig. 4a), extending between 1700
211 and 2000 m a.s.l. The walls of the crevasse are very irregular and formed by an amalgam of different
212 fractures. In addition, some portions of these crevasses are covered by scree associated with snow
213 avalanches and boulder/rock falls from the upper cliffs (Fig. 4a–c).

214 In the northeastern part of the Argentière crevasse (Fig. 4c), downslope of the *Tête aux*
215 *Vents*, the rupture occurred upslope of a cliff where the slope changes from 78° to 35° (Fig. 5a). The
216 vertical component that moved down the downslope wall reaches 7 m (Fig. 6a) and is greater than
217 the horizontal component, as the dip of the displacement is always greater than 45° (Table 2). The
218 total displacement between the two walls reaches 13 m and is nearly parallel to the approximately
219 N100° main dip of the hillslope. As the crevasses are composed of several segments with different
220 orientations, the distances measured perpendicular to the walls and strike–slip components are
221 variable. The opening of the rock crevasse reaches 10 m where the crevasse is oriented N42°E
222 perpendicular to the displacement, whereas the opening is only 1.5 m where the crevasse is oriented

223 N164° and associated with a large strike–slip component. The cliff present in the prolongation of the
224 northeastern termination of the Argentière crevasse (Fig. 5a) probably corresponds with the footwall
225 of a now collapsed gravitational structure. Indeed, a large scree including a chaotic array of blocks—
226 some of which reach sizes greater than 10^3 m^3 (Fig. 5f)—has been fed by collapses from this area.

227 The southwestern part of the rock crevasses (Fig. 4b), located close to the Chéserys chalet, shows a
228 maximum displacement of 7.2 m (Fig. 5e) and the magnitude of the vertical component is less than
229 that of the horizontal component, with the dip of the displacement vector less than 45° (Table 2),
230 and the dip of the displacement is close to the topographic slope. At the southwestern end, the
231 motion of the crevasse system is transferred toward the north by a fracture that is nearly parallel to
232 the displacement vector and crosses the Remuaz fault (Fig. 4a). This transfer fault connects to a
233 system of several fractures on the western side of the Remuaz fault that are nearly parallel to it.

234 These fractures are less planar than the Remuaz fault and their downslope compartments are
235 lowered. Numerous fractures, parallel or perpendicular to the major crevasse, are observed. Many of
236 them are slightly open ($< 1 \text{ m}$). The vertical motion on these open fractures is very limited ($< \text{few}$
237 centimeters). The use of a sounding line (thin rope with a weight) has shown depths greater than 15
238 m. The total rock volume involved is thus estimated at $5 \pm 3 \times 10^6 \text{ m}^3$. The northeastern termination
239 of the Argentière crevasse is currently unstable and a diachronic analysis, based on IGN aerial
240 images, shows that an approximately $2.5 \times 10^3 \text{ m}^3$ block collapsed sometime between 1949 and
241 1960.

242 **4.2. Chronological results**

243 We sampled the crevasse in the portion where the opening width is the smallest and is similar to a
244 fault scarp (site 1, Table 2). This is situated where the angle between the motion vector and the
245 fracture location is the smallest, 12° (Fig. 5c, d). ^{10}Be concentrations and related exposure ages are
246 reported in Table 3. In the following section, ages are considered with external uncertainties
247 including the uncertainty on the production rate. Exposure ages obtained from the offset surface are

248 17.5 ± 1.0 ka (AIG20-1) and 18.9 ± 1.0 ka (AIG20-7), for the hanging wall scarp and its footwall,
249 respectively. These two values are indistinguishable within uncertainties and have a weighted mean
250 of 18.2 ± 0.7 ka. This confirms the geomorphological observation that the gravitational scarp cut a
251 glacially polished surface that is of the same age and that was initially continuous along the slope.
252 The raw exposure ages (i.e., calculated without inheritance correction) obtained along the scarp,
253 from the scarp top to the scarp toe, range from 3.0 ± 0.6 ka (AIG20-2) to 0.7 ± 0.3 ka ago (AIG20-6)
254 for profile P1 and from 2.3 ± 0.4 ka (AIG20-17) to 1.0 ± 0.3 ka (AIG20-19) for profile P2 (Fig. 6). These
255 ages should be considered as maximum exposure ages. Figure 7 shows that the age distributions of
256 the two profiles are almost identical and that the exposure ages decrease as a function of the
257 distance to the top scarp, except for the first sample of profile P2 (AIG20-16: 1.6 ± 0.3 ka) which is
258 considered as outlier due to scarp rejuvenation and that is thus discarded for the following
259 discussion. With the inheritance correction (see section 3.2), the exposure ages become much
260 younger, ranging from 1.3 ± 0.3 to 0.6 ± 0.3 ka and from 2.0 ± 0.3 to 0.9 ± 0.3 ka, for profiles P1 and
261 P2, respectively (Table 3). The corrected ages should be considered as minimum exposure ages.

262 **5. DISCUSSION**

263 **5.1 A complex fracture pattern above the village of Argentière**

264 This new structural work has clarified the geometry of the deep-seated gravitational slope
265 deformation system located above the village of Argentière. The crevasse system does not have the
266 same direction as the plane of the Remuaz fault (Fig. 4) and the similarly oriented Aiguillette fault
267 dips to the northwest while the crevasse system dips to the southeast (Fig. 6). At the southwestern
268 end, the crevasse system intersects the Remuaz fault. The amount and the dip of the displacement
269 increase from southwest to northeast (Table 2). Currently, at the northeastern end of the Argentière
270 crevasse, the cliff corresponds to the extension of the footwall of the crevasse that has probably
271 been exposed by the collapse of a former block (Fig. 4c, 5a). This portion of the cliff suggests that the
272 surface rupture does not propagate far below the surface but tends to emerge at the base of the

273 cliff. These observations indicate a more mature structure in the NE, and suggest a propagation of
274 the structure to the southwest. This would suggest that the structure initiated later toward the
275 southwest than in the portion where the chronological constraints were obtained. All these
276 geometric characteristics suggest that the geometry of the gravity system is largely independent of
277 the geometry of the Remuaz fault and no evidence exists to relate the crevasse development with
278 the active fault movements.

279 **5.2 Northeastern Argentière crevasse initiation and kinematics**

280 The exposure ages corrected with inheritance data provide estimates for the timing of the
281 crevasse initiation. Slip along the scarp could have occurred progressively, as corrected ages (Fig. 7b)
282 are generally younger with depth. The standard deviations of several linear regression fits are in the
283 order of measurement uncertainty. The least-squares regression method suggests an average slip
284 rate of 7.6 m/ka, with standard deviations between data and linear fits that reach 0.38 ka. On the
285 other hand, as most of the corrected exposure ages are compatible within uncertainties (Fig. 7c), an
286 instantaneous (or very fast) event could have also occurred at a time equal to the weighted mean of
287 1.3 ± 0.6 ka. An instantaneous slip at 1.3 ka provides a slightly greater standard deviation of 0.45 ka.
288 The instantaneous slip model at 1.3 ka provides a lower bound for the initiation of the crevasse.

289 The end of the motion would have occurred more than 0.6 ka ago based on the lowest slip
290 rate model. This agrees with the samples located at the base of the scarp (AIG20-6 and AIG20-19)
291 that indicate corrected exposure ages of 0.6 ± 0.3 ka and 0.9 ± 0.3 ka ago. In summary, the kinematic
292 activity of the main crevasse would have started less than 1.3–2.5 ka ago. Furthermore, it could have
293 been progressive and it cannot be excluded that movement is still ongoing on the main crevasse and
294 the other crevasses located downslope. This possible present-day activity is consistent with the
295 2014–2018 InSAR displacement fields preliminarily estimated at a few millimeters per year below the
296 northeastern end of the Argentière crevasse (André et al., 2021).

297

298 **5.3 Time lag between gravitational failure initiation and glacier retreat**

299 A time lag between gravitational collapse and glacier retreat has frequently been observed
300 elsewhere, and the fundamental role of postglacial debuitressing in the initiation of gravitational
301 structures on alpine valley flanks has been identified (Cossart et al., 2008; McColl, 2012). However,
302 information to support this relation based on dating both the initiation of the gravity structure and
303 the time of debuitressing of the slopes remains scarce. In the western Alps, the Tinée valley was
304 completely deglaciaded approximately ca. 13 ka ago (Bigot-Cormier et al., 2005) while the first
305 gravitational movement of the La Clapière landslide occurred at approximately 10 ka, implying a time
306 lag of approximately 3 ka (Bigot-Cormier et al., 2005; El Bedoui et al., 2009; Jomard et al., 2014). In
307 the Romanche Valley (France), Le Roux et al. (2009) and Schwartz et al. (2017) found that glacier
308 retreat at the summit scarp was achieved between ca. 17 and 13 ka ago while the landslide initiated
309 between 8 and 6 ka ago, implying that the rupture occurred with a delay of minimum 5 ka. At Flims
310 (Switzerland), a rockslide occurred more than 2.5 ka after deglaciation (Ivy-Ochs, 2009). In the Val
311 Viola (Italy), the rockslide occurred more than 4 ka after deglaciation (Hormes et al., 2008). The
312 Fernpass (Austria) rockslide occurred 1 ka after the retreat of the glacier (Prager et al., 2009a). These
313 studies indicate that deglaciation occurred earlier than the rockslide initiation but in most cases the
314 delay is less than 5 ka. In the present study, the time lag between the initiation of the dated scarp
315 and the disappearance of the glacier at the level of the scarp is significantly higher, with a value
316 reaching approximately 15 ka.

317 **5.4 Delayed deformation models involving both debuitressing and fluid pressure**

318 Glacial processes can be a cause of RSF initiation, although they precede the initiation by
319 several thousand years. Mechanical modeling (Riva et al., 2018) indicates that large alpine rock
320 slopes undergo long-term evolution in paraglacial to postglacial environments. Progressive RSFs of
321 deglaciaded slopes occur due to rock mass weakening, increased permeability, and damage linked to
322 fluid-to-solid hydromechanical coupling. These progressive failures promote the development of

323 potentially catastrophic rockslides that would occur several thousands to ten thousand years after
324 the debuitressing (Riva et al., 2018). The large time lag between initiation and gravitational activity
325 raises the possibility of rainfall, snowmelt, or extreme meteorological events as main triggers of the
326 gravitational activities. Ivy-Ochs et al. (2009) found that some of the largest landslides in the Alps did
327 not occur directly after deglaciation but during the Early to Middle Holocene. They initiated close to
328 the Preboreal/Boreal transition (approximately 10 ka ago). That is the case for Köfels (9.8 ka ago, Ivy-
329 Ochs et al., 1998), Kandertal (9.6 ka ago, Tinner et al., 2005) and Flims (8.9 ka ago, Ivy-Ochs et al.,
330 2009), for example. This transition is marked in Western Europe by a shift to significantly warmer and
331 wetter conditions, with an increase of the mean temperature of 1°C to 2°C (Rasmussen et al., 2006)
332 and heavy annual precipitations, as evidenced by lake level (Magny et al., 2011). Therefore, external
333 hydrological parameters may have an amplifying effect on the postdebutressing gravitational
334 instabilities. A synthesis of all large ($>10^6$ m³) rock slope instability events has been carried out across
335 the Alps, including the nonglaciated parts (Zerathe et al., 2014, and references therein). Two
336 temporal clusters have been identified regardless of process (including initiation/reactivation and
337 catastrophic collapse of Sackungen, deep-seated landslides, earthflows, and rockfalls). The first
338 cluster, between 11 and 8 ka ago, is related to paraglacial readjustment by progressive failure at the
339 transition to the Holocene (Prager et al., 2009a,b). The second, 5 -3 ka ago (Zerathe et al., 2014), is
340 related to seismicity (Prager et al., 2008) or paleohydrological cycles during the Middle Holocene
341 climatic transition (Prager et al., 2008; Zerathe et al., 2014).

342 The initiation of the Argentière crevasse occurred during the Late Holocene, later than all the
343 previously mentioned gravitational instabilities, and this slope failure had likely required the
344 superposition of several factors. Topography first plays a major role in the rock slope instability.
345 In the case of the Argentière crevasse (Fig. 8), the rupture occurred upslope of a rock wall
346 with a slope of 78°. This topography is the result of the glacier erosion that shaped the
347 slopes at the valley scale before the LGM. However, the very long time lag (ca. 15 ka)
348 between the complete deglaciation, which rapidly occurred at Argentière, releasing the

349 entire slope between 18 and 15 ka ago (Fig. 8), and the RSF initiation (1.3 – 2.5 ka ago),
350 reasonably exclude a direct effect of slope debuttressing here. However, it is not excluded
351 that debuttressing could have led to previous rockfalls 15 ka ago along the valley of
352 Chamonix as several blocks were observed along the lower slopes, but this topic would
353 require a dedicated study. Eventually, wet climatic events, and particularly those beginning
354 approximately 2.7 ka ago at the transition from sub-Boreal to sub-Atlantic periods (Fig. 9),
355 may have favored the hydraulic loading of interconnected fractured areas during progressive
356 mechanical damage. Then, a drift toward a peak of wetter conditions that is evidenced
357 around 1.6 ka ago (Arnaud et al., 2012), as recorded in the sedimentary archives of the
358 Bourget lake (Western Alps) (Fig. 9), may also have accelerated the failure.

359 Regarding the permafrost, there is no observable evidence that ice is currently present in the
360 Argentière crevasse, and Boeckli et al. (2012) did not model permafrost at the same altitude (1900 m
361 a.s.l.) in their Alpine Permafrost Index Map. Furthermore, Heiri et al. (2015) who used biotic proxies
362 in lake sediments from French and Swiss Alps to reconstruct palaeotemperatures (15000-90 Cal BP),
363 shown that mid-Holocene July air temperatures generally range between 16.5 and 17.5°C (Heiri at
364 al., 2015). Records for the Anterne Lake (2000 m a.s.l), located less than 10 km to the northwest of
365 the crevasse as the crow flies, show temperatures ranging from 7.5 to 11°C for the past 2 ka. Affolter
366 et al. (2019), thanks to Milandre Cave fluid inclusion temperature record, shown that temperature
367 anomaly for central Europe was about -5°C 12000 years ago. These elements suggest that permafrost
368 can be excluded at the altitude of our study site. Further studies would be necessary to determine
369 the role played by seismicity in the initiation of the Argentière crevasse.

370 **5.5 Gravitational hazard**

371 The gravitational structure of the Argentière crevasse constitutes a possible significant
372 hazard for the downslope village of Argentière, as well as road and rail infrastructures. In the
373 documents concerning the risks to the town of Chamonix–Mont-Blanc, this structure has been

374 identified. It has been noted that the Béchar spur constitutes a natural bund protecting the valley
375 downslope of the northeastern Argentière crevasse (Plan de Prévention des Risques Naturels
376 prévisibles, 2000). Modeling using Rapid mass movement simulation (RAMMS) software (Tavernier,
377 2020) confirms that the village of Argentière is protected by the Béchar spur for blocks $< 100 \text{ m}^3$.
378 Tavernier (2020) shows that a very small percentage (less than 0.3%) of modeled rocks passes the
379 Béchar spur and reaches the first infrastructures in the valley. However, further southwest, the valley
380 is directly exposed to possible collapses as almost 40% of the modeled rocks cross the railway line
381 (Tavernier, 2020). Eventually, the Argentière crevasse raises questions about its cohesion and its
382 resistance to possible seismic shocks. Two scenarios are possible: either the different units that
383 compose it fragment and form unitary blocks, or they fall apart in a single unit. On the one hand, the
384 mobilized volume is of a hundred cubic meter, on the other hand, several tens of thousands. If
385 unitary blocks can be stopped by the Béchar, the collapse of blocks with greater volume ($5 \pm 3 \times 10^6$
386 m^3) would threaten the valley and its inhabitants.

387 **6. Conclusion**

388 This work describes initial research on a deep rock slope failure in the Aiguilles Rouges Nature
389 Reserve located in the upper Chamonix valley, French Alps. It shows that a large crevasse extends for
390 approximately 1 km in the gneiss above the village of Argentière. The total displacement, as well as
391 the ratio of vertical to horizontal displacement component, decreases from northeast to southwest,
392 suggesting a southwestward propagation. The rock mass located downslope of the Argentière
393 crevasse is itself affected by numerous open fractures. The ^{10}Be dates of the escarpment at its
394 northeastern termination suggest that the main development initiated at 1.3–2.5 ka ago and may still
395 be ongoing. It should be stressed that the main development of this structure occurred well after the
396 deglaciation of this area, which occurred about 18 ka ago. Other case studies focused on similar
397 gravitational processes in the Alps demonstrated that the time lag between the initiation of RSFs and
398 glacier retreat was in most cases less than 5 ka. Our study suggests a time lag reaching approximately
399 15 ka, which reasonably allow to discard this effect here. Predisposing factors (lithology, rock mass

400 characteristics, and geologic structures) and potential triggering factors (exceptional rainfall events,
401 seismic activity, changes in groundwater regime) have been discussed to understand the origins of
402 the initiation of the Argentière crevasse. Its evolution is likely to constitute a future major risk for the
403 upper part of the Chamonix valley. Further studies on the measurement of the current deformation,
404 the stability of this zone, and the modeling of the trajectories of the rockfalls in this area are
405 necessary.

406

407 **Author contributions**

408

409 L. Courtial-Manent performed her Master's thesis on the subject, analyzed the data and wrote the
410 paper, J.L. Mugnier led the project, worked on the tectonics aspects and wrote the paper. S. Zerathe
411 worked on the age interpretation, the geomorphology mapping and wrote the paper, J. Carcaillet
412 checked the quality of the geochemical processes and worked on the ^{10}Be concentrations. R.
413 Vassallo, J.F. Buoncristiani and L. Ravanel have helped in the mapping process and in reviewing the
414 manuscript. L. Tavernier did his first year of Master's thesis on this subject and he participated in the
415 fieldwork to measure the displacement vectors.

416 **Data availability**

417 Dataset related to this article (photography and precise description of the ^{10}Be sites, three-
418 dimensional (3D) model, etc.) can be asked to the corresponding author.

419 **Declaration of competing interest**

420 The authors declare that they have no known competing financial interests or personal relationships
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430

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692

693 **Figure captions**

694

695 Figure 1. Geological context of the studied area with shaded 75 m Digital Elevation Model (source:

696 Institut national de l'information géographique et forestière, IGN). Altitudes are in m a.s.l. Faults and

697 geology are based on the Bureau de Recherches Géologiques et Minières (BRGM) mapping (Pairis et

698 al., 1992a; Pairis et al., 1992b, Janjou, 2004). Northings and eastings are according to

699 RGF93/Lambert-93 grids. Black dashed lines represent Last Glacial Maximum glacial upper limits

700 (Coutterand & Buoncristiani, 2006; Coutterand, 2010). Glacier extents are from the Randolph Glacier

701 Inventory (RGI Consortium, 2017) 6.0.

702

703 Figure 2. Photo of the Aiguillette d'Argentière, famous rock-climbing site.

704

705 Figure 3. Photograph of the entire mountain slope taken from the Glacier de Lognan, on the opposite

706 side of the Arve valley.

707

708 Figure 4. a) Geomorphologic settings of the study area showing the main structural and gravitational
709 features. Slope map with inverse symbology (flattest slopes in white and highest in black, threshold
710 set at 55°) derived from the Digital Elevation Model of the Institut National de la Recherche
711 Agronomique (INRAe) 1 m high-resolution LiDAR . Slopes range from 0° to 87°. b) Zoom in on the
712 southwestern part of the crevasse. c) Zoom in on the northeastern part of the crevasse. Numbers
713 refer to the location of studied sites described in Table 2. Black lines correspond to the profiles
714 through the northeastern termination of the Argentière crevasse, see Fig. 6.

715

716 Figure 5. a) Global view of the northeastern Argentière rock crevasse. b) Location of the glacially
717 polished surface sample (AIG20-7). c) Sample location for the profile P2 (red dots). d) Sample location
718 for the profile P1 (black dots and blue triangle). e) Rock crevasse in the southwestern part of the
719 structure. The displacement vector is estimated by fitting the characteristic shapes of one wall to the
720 other. f) Collapsed rock from the northeastern part of the rock failure. Its estimated volume is
721 greater than 1000 m³.

722

723 Figure 6. Profiles through the eastern termination of the crevasse and location of the samples. The
724 red dotted lines indicate the potential surfaces before the scarp development. The red dashed lines
725 indicate the crevasses. *Location of sample AIG20-7 has been projected. The profiles were extracted
726 from the photogrammetric three-dimensional (3D) model.

727

728 Figure 7. Plot of the ¹⁰Be exposure ages as a function of the distance from the top of the scarp. a) All
729 the obtained ages without inheritance correction. B) Corrected ages of the scarp considering
730 inheritance before the initiation of the RSF. C) Probability density plot of the corrected ages where

731 thin black lines are individual ages (sample AIG20-16 in dashed line as outlier) and the thick blue line
732 is the summed probability curve (AIG20-16 excluded).

733 Figure 8. Section through the upper Arve valley illustrating the position of the eastern end of the
734 large crevasse in relation to the tectonic structures and the evolution of the deglaciation. (1) From
735 Coutterand and Buoncristiani (2006); (2) from this study; (3) from Jaillet and Ballandras (1999); (4)
736 from Protin et al. (2019).

737

738 Figure 9. Sedimentation rate and terrigenous contribution in the Lac du Bourget (from Arnaud et al.,
739 2012)

740

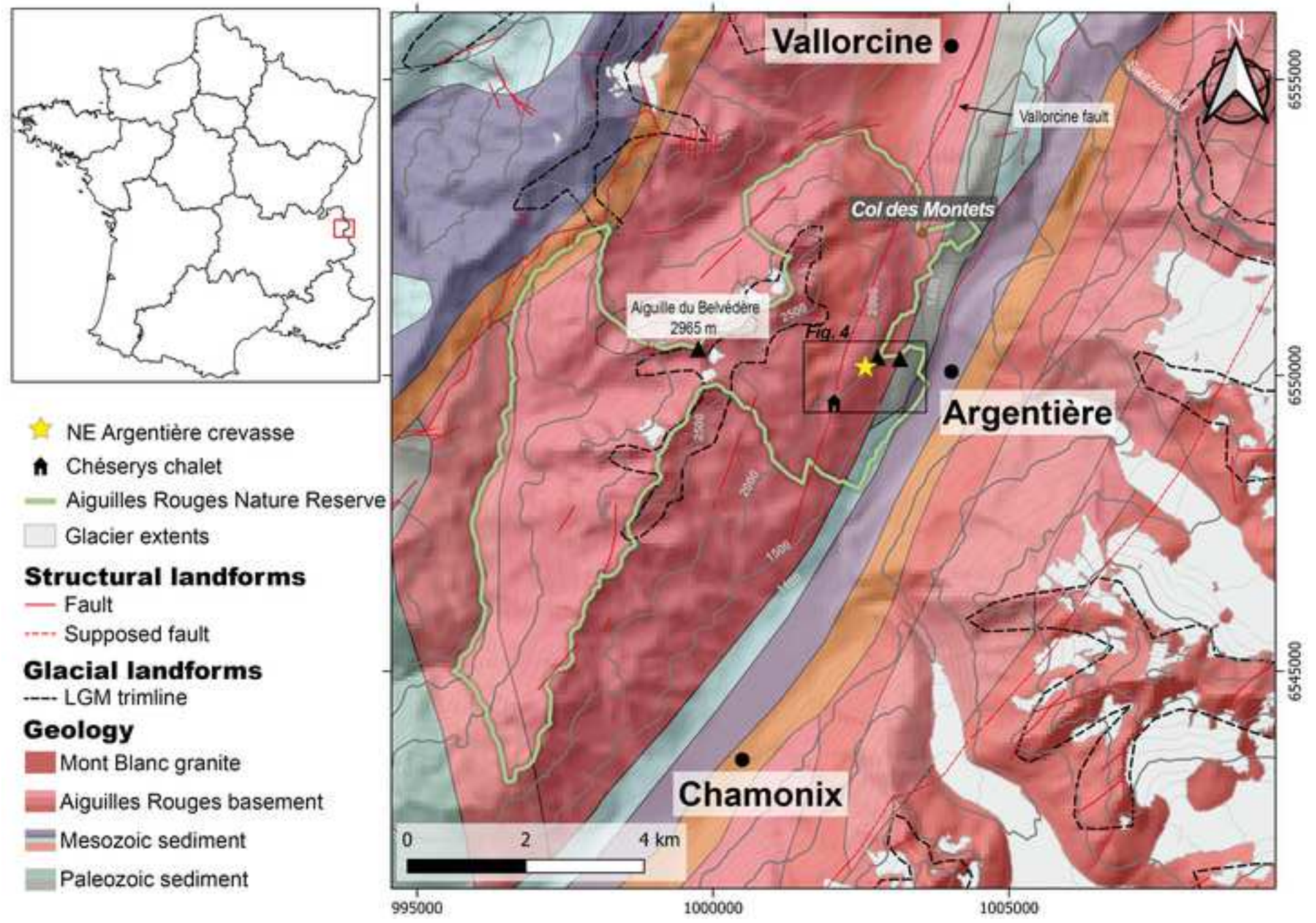
741

742 **Table captions**

743 Table 1. Sample characteristics. T is the sample thickness; Z is the distance of the sample from the
744 top of the scarp; Z_E corresponds to the estimated depth before the scarp was exhumed; S is the
745 shielding; and M_{Quartz} is the quartz mass.

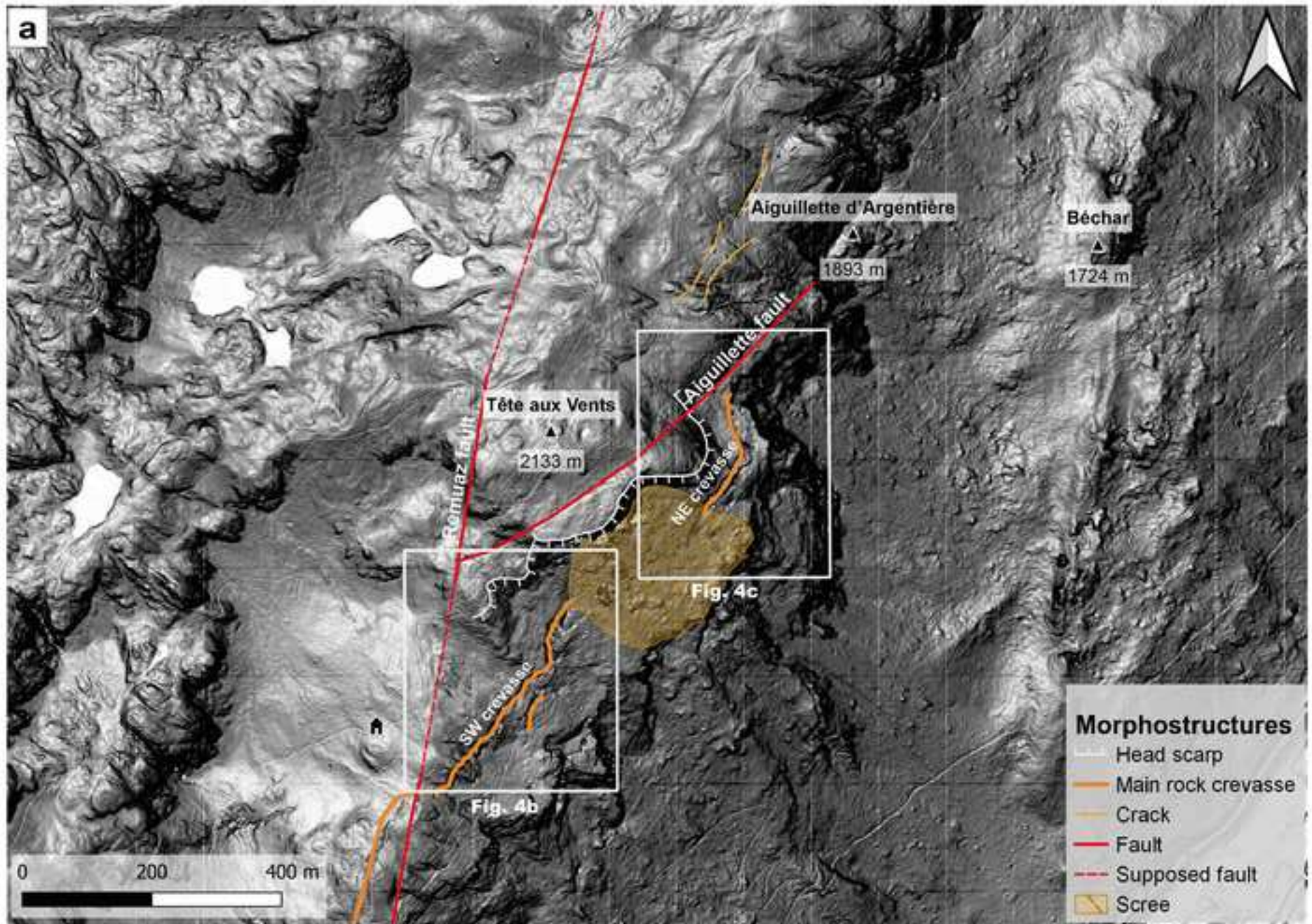
746 Table 2. Structural characteristics of the main crevasses from northeast to northwest (site locations
747 shown in Fig. 4).

748 Table 3. ^{10}Be concentrations, calculated and corrected ages considering inheritance. Internal
749 uncertainty corresponds to the 1σ uncertainty without production rate errors. The in-house standard
750 (Braucher et al., 2015) and accelerator mass spectrometer internal error (0.5%; Arnold et al., 2010)
751 have been included. A blank value with a $^{10}\text{Be}/^9\text{Be}$ ratio of 6.95×10^{-15} was used to correct the
752 samples.

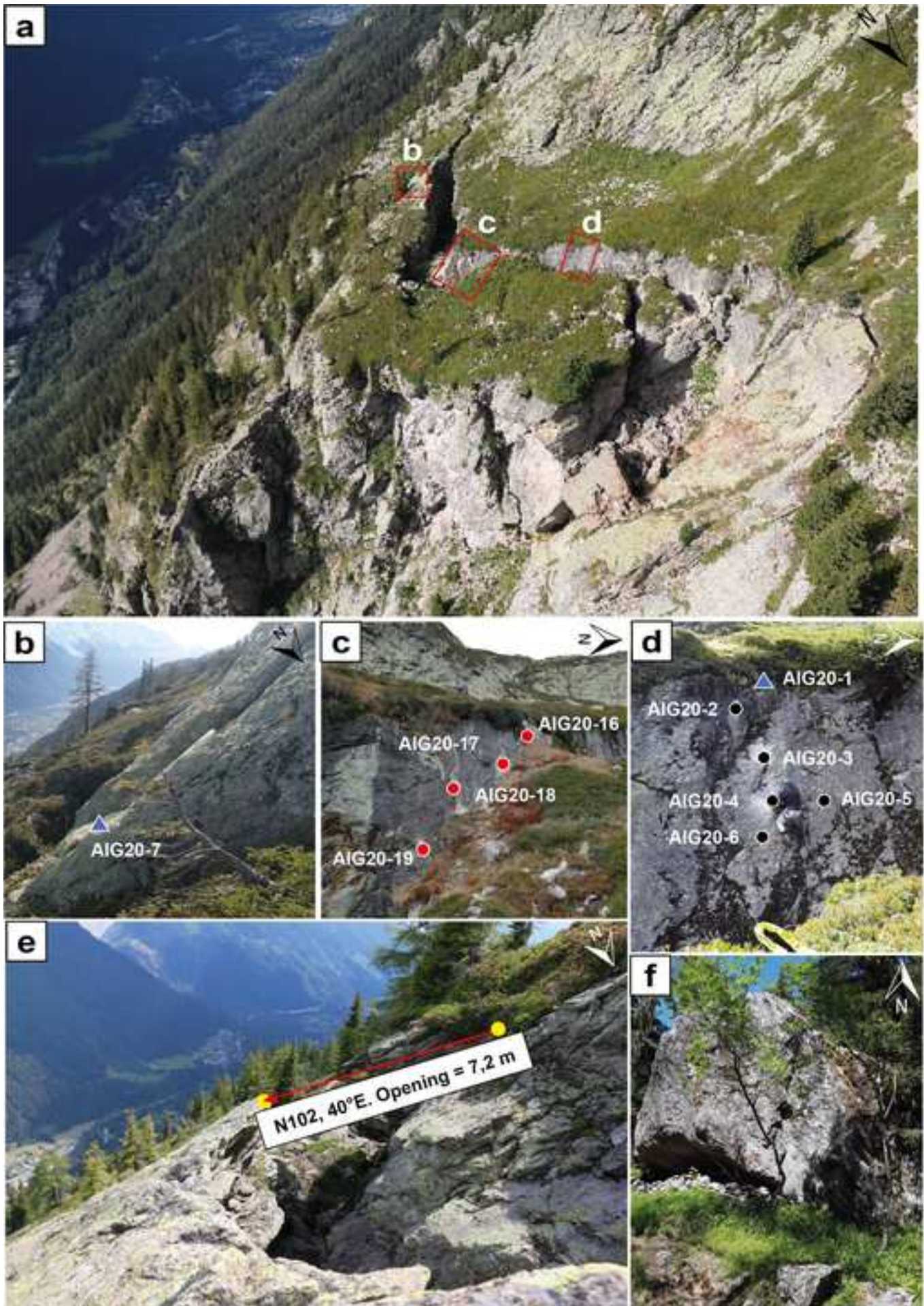


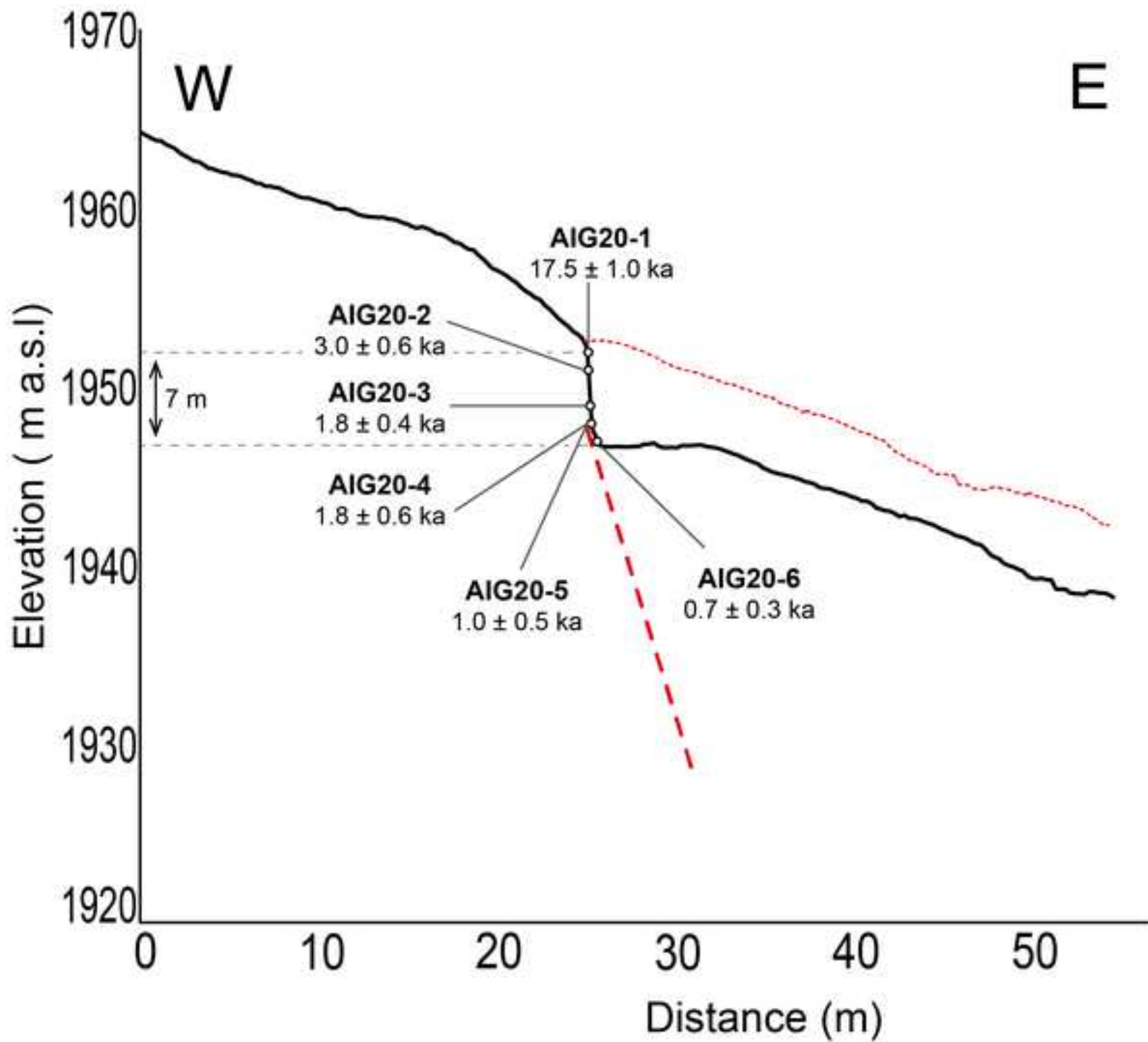


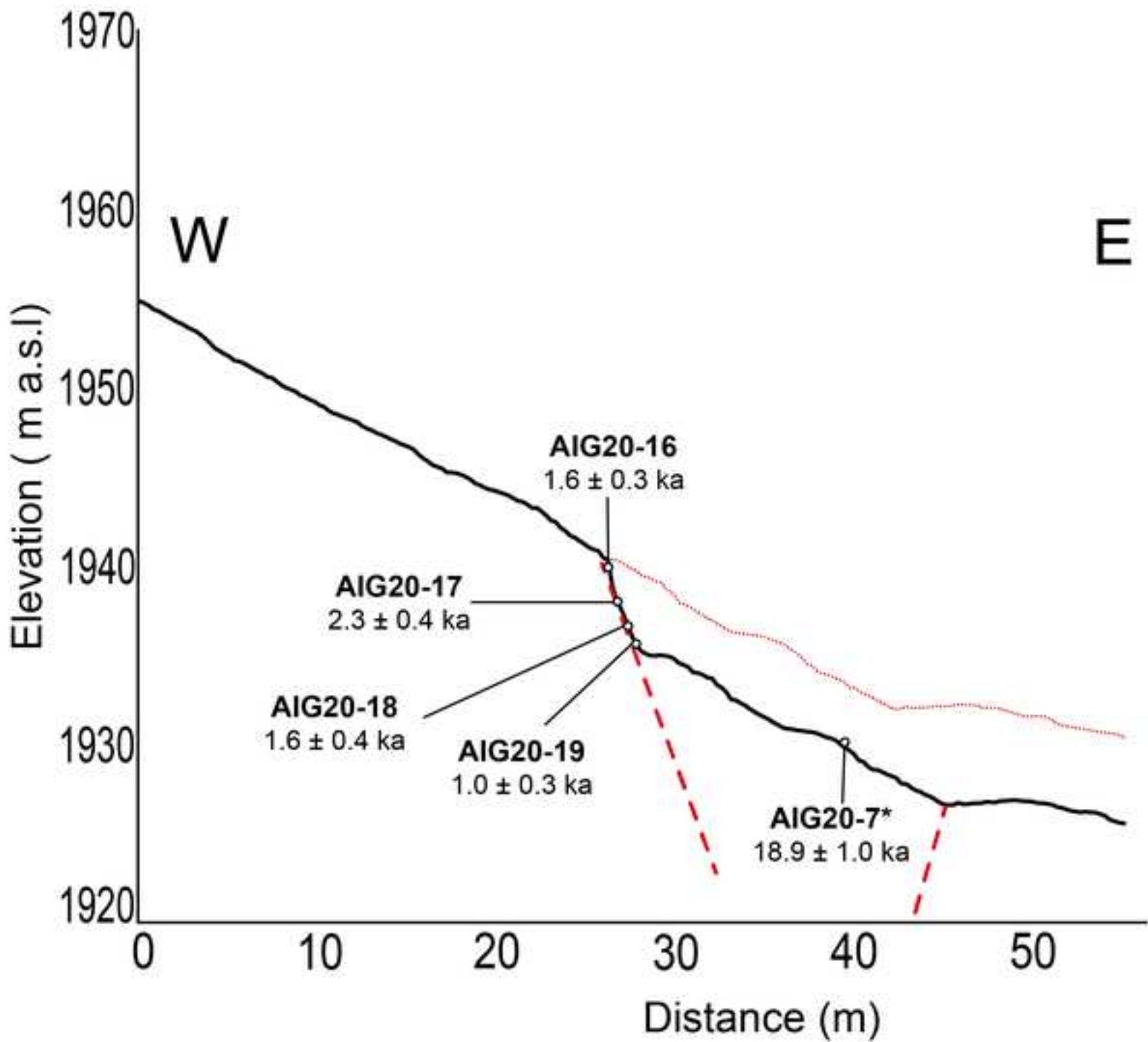


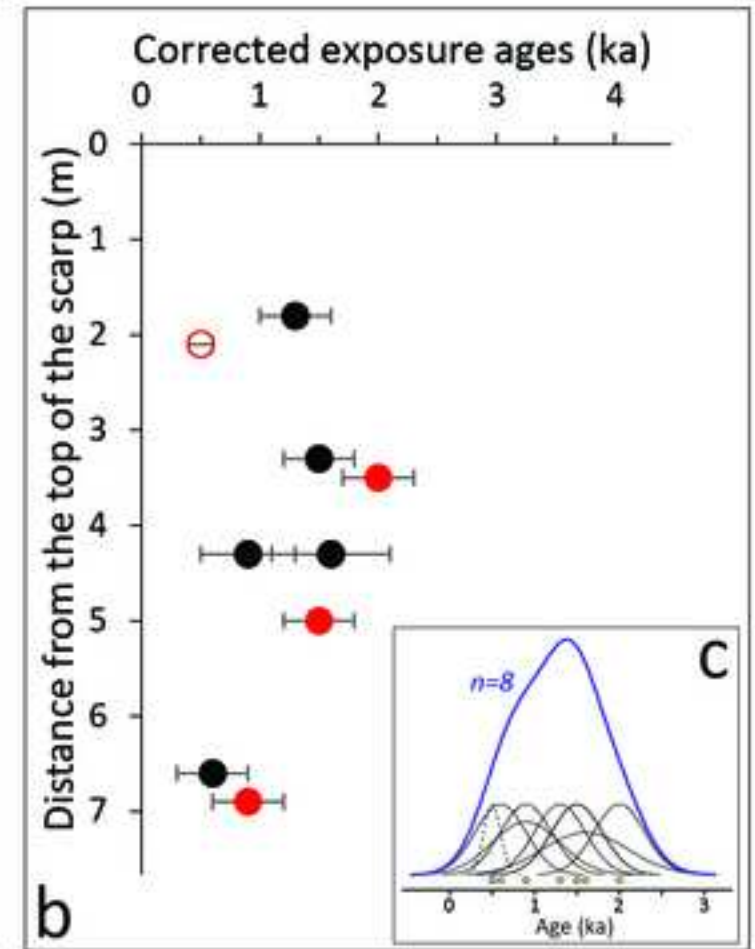
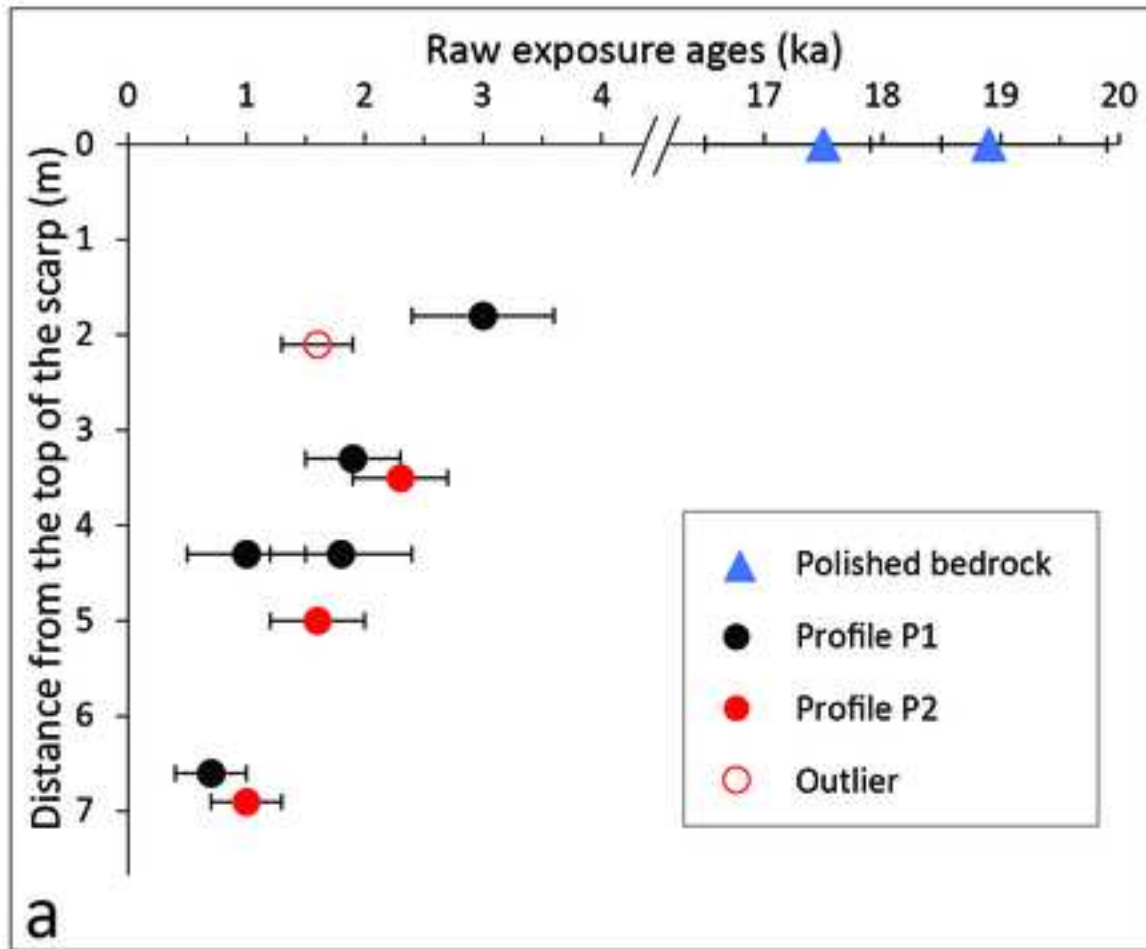


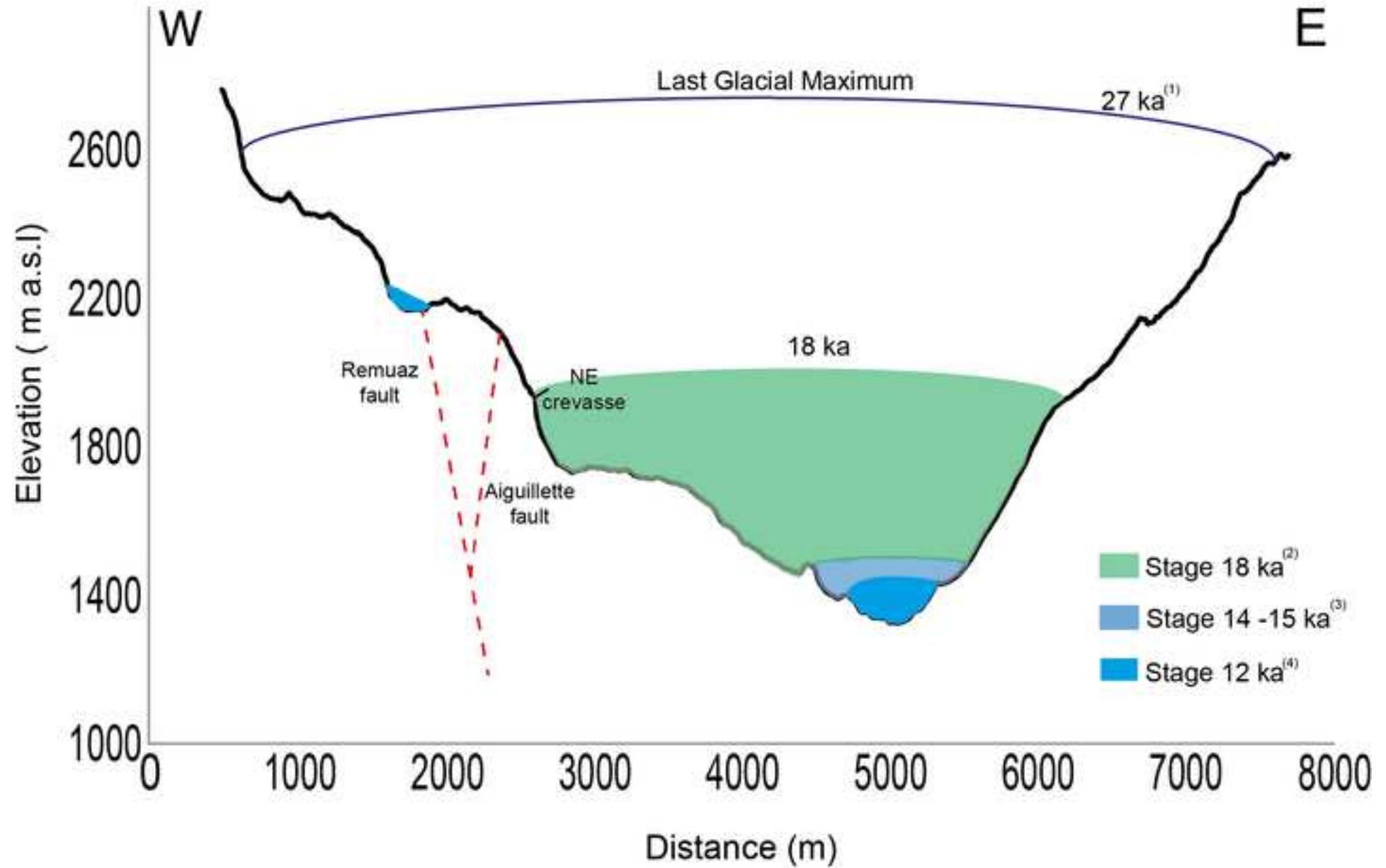


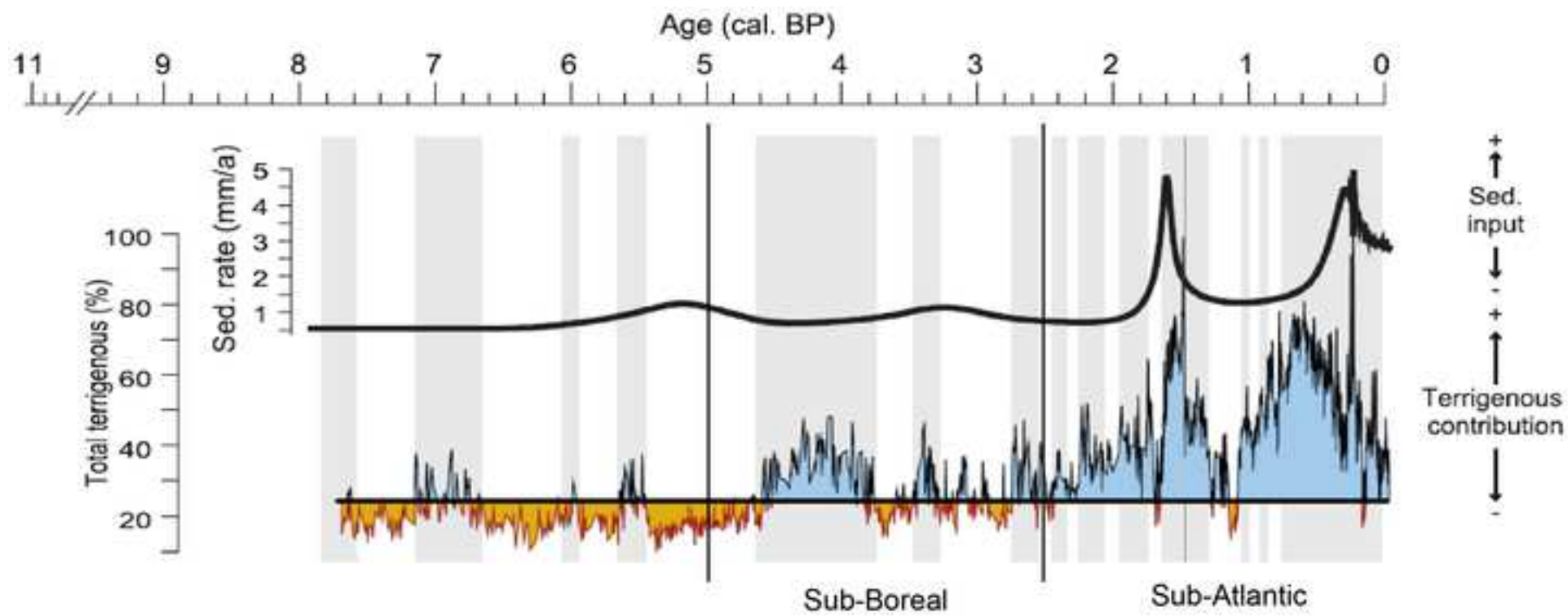












Site	Sample	Type	Latitude	Longitude	Elevation	T	Aspect	Z	Z _E	S	M _{quartz}	Carrier	¹⁰ Be at	¹⁰ Be/ ⁹ Be
			(dd)	(dd)	(m a.s.l)	(cm)	(°)	(m)	(m)		(g)	(mg ⁹ Be)	x10 ⁵	x10 ⁻¹⁴
NE crevasse (Profile P1)	AIG20-1	Scarp	45.982667	6.910054	1937	4	72	0.3	0.2	0.584	21.33	0.50	23.08	6.87 ± 0.26
	AIG20-2	Scarp			1936	4	72	1.8	1.5	0.584	16.90	0.50	3.07	0.92 ± 0.18
	AIG20-3	Scarp			1934	4	72	3.3	2.7	0.584	25.59	0.50	2.85	0.86 ± 0.17
	AIG20-4	Scarp			1933	4	72	4.3	3.5	0.584	22.17	0.51	2.09	0.61 ± 0.19
	AIG20-5	Scarp			1933	7	72	4.3	3.5	0.584	29.34	0.51	1.69	0.5 ± 0.23
	AIG20-6	Scarp			1931	6	72	6.6	5.3	0.584	29.22	0.50	1.13	0.34 ± 0.16
	AIG20-7	Polished bedrock	45.98186	6.910152	1895	4	60		0	0.682	35.86	0.50	44.57	13.4 ± 0.45
NE crevasse (Profile P2)	AIG20-16	Scarp	45.982404	6.910190	1933	3	70	2.1	1.8	0.584	28.07	0.52	2.44	0.71 ± 0.15
	AIG20-17	Scarp	45.982373	6.910219	1932	4	70	3.5	3	0.584	26.68	0.50	3.22	0.96 ± 0.15
	AIG20-18	Scarp	45.982353	6.910233	1930	4	70	5	4.3	0.584	38.37	0.50	3.30	0.99 ± 0.21
	AIG20-19	Scarp	45.982325	6.910252	1929	3	70	6.9	5.9	0.584	33.06	0.50	1.99	0.6 ± 0.18

Site	Displacement vector (norm)	Strike of displacement vector	Dip of displacement vector	Mean crevasse orientation	Mean dip of the crevasse (a)	Mean opening of the crevasse
1	12 m	N 100°	65°E	N164°E	72°E	1.5 m
2	13 m	N 130 °	50°E	N42°E	Vertical	10 m
3	7.2 m	N 102°	40°E	N8°E	65°E	3.7 m
4	3.7 m	N 115°	45°E	N22°E	67°E	2 m

Sample	¹⁰ Be x10 ⁴ at.g ⁻¹	¹⁰ Be age ± external uncertainty (ka)	Internal uncertainty (ka)	Inherited ¹⁰ Be concentration (at.g ⁻¹)	Inheritance/ measured (%)	Corrected concentration (at.g ⁻¹)	Corrected age ± external uncertainty (ka)	Internal uncertainty (1σ) (ka)
AIG20-1	19.30 ± 0.74	17.51 ± 1.01	0.65	-	-	-	-	-
AIG20-2	3.14 ± 0.60	3.01 ± 0.59	0.57	1,80E+04	57	1,35E+04	1.26 ± 0.30	0.29
AIG20-3	1.94 ± 0.39	1.84 ± 0.38	0.37	3,13E+03	16	1,62E+04	1.53 ± 0.33	0.31
AIG20-4	1.86 ± 0.58	1.76 ± 0.57	0.56	1,31E+03	7	1,73E+04	1.64 ± 0.52	0.52
AIG20-5	1.04 ± 0.48	1.01 ± 0.46	0.45	1,31E+03	13	9,07E+03	0.88 ± 0.40	0.40
AIG20-6	0.69 ± 0.32	0.66 ± 0.30	0.30	5,86E+02	6	6,32E+03	0.61 ± 0.27	0.29
AIG20-7	23.62 ± 0.81	18.87 ± 1.03	0.62	-	-	-	-	-
AIG20-16	1.65 ± 0.35	1.56 ± 0.34	0.34	1,13E+04	68	5,23E+03	0.48 ± 0.11	0.11
AIG20-17	2.37 ± 0.36	2.25 ± 0.36	0.35	2,44E+03	10	2,12E+04	2.01 ± 0.33	0.32
AIG20-18	1.65 ± 0.36	1.56 ± 0.35	0.34	8,17E+02	5	1,56E+04	1.47 ± 0.33	0.32
AIG20-19	1.00 ± 0.31	0.95 ± 0.29	0.29	5,37E+02	5	9,49E+03	0.89 ± 0.28	0.27