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25 **Key words**

26 Late Pleistocene; glacial geochronology; cosmogenic exposure dating; moraine; plateau icefield; Massif
27 Central; Aubrac

28 **1 Introduction**

29 Recent advances in glacial chronology highlight that glaciated areas in Europe did not experience their
30 Local Last Glacial Maximums (LLGMs) synchronously during the late Pleistocene (129-11.7 ka)
31 (Gillespie and Molnar, 1995; Hughes et al., 2013). In the western Alps (Gribenski et al., 2021) and in
32 the eastern Pyrenees (Calvet et al., 2011) for example, the LLGM was probably not coeval with the
33 global Last Glacial Maximum (LGM) between 27.5 and 23.3 ka, *sensu* Hughes and Gibbard (2015).
34 LLGMs in Europe were controlled by both the North Atlantic and Mediterranean atmospheric
35 circulation patterns, which evolve during the late Pleistocene (Domínguez-Villar et al., 2013; Delmas,
36 2015; Hughes and Woodward, 2017; Monegato et al., 2017; Gribenski et al., 2021; Reixach et al., 2021).
37 This evolution is controlled by the southward displacement of the Polar Jet Front that induced
38 reorganization of atmospheric circulations (Florineth and Schlüchter, 2000; Kuhlemann et al., 2008;
39 Luetscher et al., 2015). In the Massif Central, France, the timing of LLGMs differed from one area to
40 another, according to available relative and indirect chronologies, and thus it provides a key area to
41 identify changes in atmospheric circulation patterns.

42 Veyret (1978) first proposed a regional and comprehensive synthesis of late Pleistocene glacier
43 fluctuations in the Massif Central based on geomorphological criteria, comprising delimitation of glacier
44 extensions, the prominence of glacial features, the degree of the alteration of moraines and
45 morphostratigraphy. The author thus identified three glaciated areas (Fig. 1), each controlled by
46 topographic and climatic influences. The western glacier group is composed of the Cantal, Monts Dore
47 and Cézallier coalescent glacier systems (Fig. 1). Its LLGM was in MIS 2 based on the limited
48 weathering of glacial deposits (Goër de Hervé, 1972; Veyret, 1978). A large glacier advance following
49 the LLGM, locally named the Recurrence Event (Veyret, 1978), deposited end moraines within the main
50 valleys of the Massif Central, followed by a phase of cirque glaciation that is only represented within
51 the Cantal (Valadas, 1984). The timing of full deglaciation is not well known, but it is associated with

52 the Last Glacial-to-Interglacial Transition (LGIT) (Veyret, 1978; Valadas, 1984; Vergne, 1991). The
53 southern glacier group was composed of the Margeride and Lozère Mount plateau icefields and by the
54 Velay cirque glacier, that reached its LLGM extent during the MIS 2 (Veyret, 1978). The Aubrac
55 Mountains are located at the margins of the western and southern glacier groups (Fig. 1) (Veyret, 1978).
56 The eastern glacier group corresponds to the small icecap of the Forez. It reached its LLGM extent
57 during the MIS 4 (Etlicher, 1986).

58 Inferences of climate change from deglacial sequences rely on direct dating tools applied to ice-marginal
59 landforms that provide robust glacial chronology (Kleman and Borgström, 1996; Stokes et al., 2015). In
60 particular, terrestrial cosmogenic nuclides (TCNs) have been widely applied to glacial landforms, such
61 as end moraines, glacial deposits or subglacially eroded bedrock. Resultant exposure ages constrain the
62 timing of the advance and decay of late Pleistocene glaciers (Balco, 2011; Stokes et al., 2015; Allard et
63 al., 2021). Previous reconstruction of the late Pleistocene glaciation of the Massif Central was based on
64 morphostratigraphic evidence and relative correlations with local and regional paleoenvironmental
65 proxies (Goër de Hervé, 1972; Poizat and Rousset, 1975; Veyret, 1978; Valadas, 1984; Etlicher, 1986;
66 Ancrenaz et al., 2020). Indirect paleoenvironmental proxies, such as pollen sequences (Etlicher et al.,
67 1987; Reille and Beaulieu, 1988; Ponel and Russell Coope, 1990; Vergne, 1991; Gandouin et al., 2016;
68 Ponel et al., 2016) or sedimentological stratigraphies (Valadas, 1984; Laville et al., 1986; Van Vliet-
69 Lanoë et al., 1991), are available and provide indirect chronological boundaries for glacier fluctuations.
70 However, resulting chronologies differ for the various glaciated areas of the Massif Central (cf. Etlicher
71 and Goër de Hervé, 1988; Defive et al., 2019). Hence, without direct dating of glacial deposits, important
72 uncertainties and inconsistencies remain, especially concerning the timing of both the LLGM and full
73 deglaciation. During the last glacial cycle, two periods, locally named the Early and Late Pleniglacial
74 (*sensu* Laville et al., 1986; Van Vliet-Lanoë et al., 1991), that correlate with Marine Isotopic Stage 4
75 (MIS 4; 71–57 ka; Lisiecki and Raymo, 2005) and MIS 2 (29–11.7 ka; Lisiecki and Raymo, 2005),
76 respectively, were recognised as two major phases of glacier growth in the Massif Central (Etlicher and
77 Goër de Hervé, 1988; Defive et al., 2019). In general, in the literature and according to the considered
78 glaciated area, the LLGM is attributed to one of these two periods.

79 The objectives of the present study were to assess the existing relative glacial chronologies (Poizat and
80 Rousset, 1975; Veyret, 1978; De Goër et al., 1994; Ancrenaz et al., 2020) and to provide the first direct
81 dating chronology for fluctuations of the Aubrac plateau icefield. In that respect, key glacial landforms
82 identified by the morphostratigraphy are targeted. In combination with an extensive literature review on
83 glacier fluctuations within the Massif Central, a revised and comprehensive chronostratigraphic
84 framework is proposed that focuses on the southwestern Massif Central. Furthermore, inferences of
85 glacier fluctuations from regional paleoclimates, especially favourable glacier-growth phases, are
86 proposed.

87 **2 Study area**

88 The Aubrac Mountains, in the south western Massif Central, extend over 1300 km² and culminate at the
89 Signal de Mailhebiau at 1467 m above sea level (a.s.l.) (Fig.1). Three geomorphological subareas divide
90 the Aubrac Mountains (Fig. 2). The Aubrac Highlands are composed of a SSE to NNW alignment of
91 rounded summits composed of Miocene volcanic rocks (Goër de Hervé et al., 1991; Leibrandt, 2011).
92 This topographic ridge extends along 25 km with an elevation decrease from 1400 m in the SSE to 1200
93 m in the NNW. The Aubrac Highlands separate the two following subareas: the Aubrac Plateau to the
94 northeast and Aubrac Valleys to the southwest (Fig. 2). The Aubrac Plateau, with elevations between
95 1000 and 1200 m, formed within granitic Carboniferous rocks (Couturié, 1977). The Bès valley oriented
96 from SSW to NNE represents the major hydrographic basin that drains this subarea. In addition, the
97 Aubrac Valleys, incised into the margins of the highlands, are deep and steep valleys that are 10 to 20
98 km long and cut in metamorphic rocks.

99 The Aubrac Mountains constitute the first orogenic obstacle that is encountered by atmospheric low-
100 pressure systems pushed by the westerlies from the North Atlantic into the European continent. Current
101 precipitation is characterised by a maximum during winter and a minimum during summer (Jubertie,
102 2006). Precipitation of more than 1500 mm.y⁻¹ was recorded (1980-2010) for the Aubrac Highlands at
103 Nasbinals, 1284 m a.s.l. (ID station: 48104002; <https://donneespubliques.meteofrance.fr>). Snowfall and
104 snow cover during winter highlights the role of the Aubrac Highlands as a preferential area for snow
105 accumulation (Valadas and Veyret, 1981).

106 **3 The late Pleistocene Aubrac glaciation**

107 **3.1 Evidence for a plateau icefield glaciation**

108 The topographic setting of the Aubrac Mountains exert a strong control on the spatial extension and
109 dynamics of the former plateau icefield. The plateau icefield accumulation zone was centred on the
110 Aubrac Highlands (Fig. 2), with an estimated thickness between 200 and 300 m (Poizat and Rousset,
111 1975; Ancrenaz et al., 2020). The ice divide of the plateau icefield followed the SSE–NNW orientated
112 Aubrac Highlands and separated two distinct zones of glacial dynamics. The Aubrac Plateau that was
113 situated to the east of the ice divide and was predisposed to the establishment of a plateau glacier because
114 of its relatively low-lying topography (Figs. 2-3). This plateau glacier is recorded by ice-scoured knock
115 and lochan topography, with bedrock forming *roches moutonnées* and rock basins partly infilled by
116 glacial or associated sediments (Poizat, 1973; Veyret, 1978; Goër de Hervé et al., 1994; Ancrenaz et al.,
117 2020). According to the orientation of *roches moutonnées* (Ancrenaz et al., 2020) and the dispersion of
118 erratics (Poizat and Rousset, 1975), the Bès valley was the preferential drainage route for the Aubrac
119 plateau glacier. In the Aubrac Valleys, isolated glacial deposits were associated to phases of glacier
120 retreats and overdeepened morphologies were recognised and associated with subglacial erosion by
121 outlet valley glaciers (Rousset, 1963, 1970; Poizat and Rousset, 1975).

122 Local plateau icefields are sensitive to climatic changes when the regional Equilibrium Line Altitude
123 (ELA) is close to the plateau elevation. Indeed, climatic changes inducing variations in ELAs lead to
124 substantial changes in the distribution of glacier ablation/accumulation zones, i.e. small increase or
125 decrease of the ELA induce significant glacier advance or retreat in response to strong modification of
126 the accumulation/ablation zones distribution (Manley, 1955; Rea et al., 1999; Rea and Evans, 2003;
127 Boston et al., 2015). This observation is in accordance with the landform assemblage identified in the
128 Aubrac Plateau, indicating rapid and *in situ* glacier downwasting during deglaciation (Section 2.3)
129 driven by small changes in the ELA. The accumulation zone of the Aubrac plateau icefield was located
130 along the topographic ridge of the Aubrac Highlands (Poizat and Rousset, 1975; Ancrenaz et al., 2020).
131 The mass balance adjustment of this zone is argued to control the former glacier extent in the Aubrac
132 Plateau and in Aubrac Valleys (Poizat and Rousset, 1975; Ancrenaz et al., 2020). A glacier advance (or

133 retreat) in the Aubrac Plateau was coeval with a glacier advance (or retreat) in the Aubrac Valleys in
134 response to the thickening (or thinning) of the glacier accumulation zone in the Aubrac Highlands.

135 **3.2 Deglaciation sequence in the Aubrac Plateau**

136 The majority of glacial landforms in the Aubrac Mountains are located on the Aubrac Plateau (Figs. 2-
137 3). Three distinct glacier advances were reconstructed in previous studies from investigations of
138 landform assemblages (Table 1, Fig. 3). First, the extent of the LLGM was defined by the maximum
139 extent of volcanic erratic boulders on the Aubrac Plateau (Figs. 2, 3A) that matched the maximal extent
140 of glacial sediments (Veyret, 1978; Goër de Hervé et al., 1994). On this basis, the LLGM outer limit
141 matched the occurrence of tors or saprolite associated with periglacial environments (Goër de Hervé et
142 al., 1994; Ancrenaz et al., 2020). Moreover, the Allatieux end moraine (1199 m), which corresponds to
143 a group of three discontinuous ridges that are one hundred metres long, demarcates the LLGM in the
144 Aubrac Plateau (Table 1; Fig. 4A.) (Ancrenaz et al., 2020). This landform matches (i) the maximal extent
145 of erratic boulder dispersion and (ii) the occurrence of periglacial features such as bloc fields or tors.
146 The fresh aspect with no boulder weathering rind and the well preserved glacial sediments support an
147 interpretation of that the LLGM correlates with the global LGM (Veyret, 1978; Ancrenaz et al., 2020).

148 The end of the LLGM is characterised by a pronounced retreat of the Aubrac plateau icefield (Fig. 3C).
149 An early still-stand is recorded in the Bès valley at the Longevialle end moraine (1040 m a.s.l.),
150 composed of a group of parallel and subdued ridges with pronounced crest attenuation (Table 1; Fig.
151 4B). It is located only 2 km behind the maximum erratic boulder dispersion that demarcates the LLGM
152 limit and also behind fluvio-glacial sediments deposited during LLGM retreat. Subsequent glacier
153 recession is recorded by numerous, well-preserved glacial, glacio-fluvial and glacio-lacustrine sediment
154 landform assemblages that were identified within the Aubrac Plateau (Poizat, 1973; Poizat and Rousset,
155 1975; Veyret, 1978; Goër de Hervé et al., 1994; Ancrenaz et al., 2020). These sediments were associated
156 with the formation of ice-dammed lakes which infilled the major depressions of the plateau and comprise
157 kame hills and proglacial outwash. These landforms suggest *in situ* glacier downwasting (Livingstone
158 et al., 2010; Lovell et al., 2019). Indirect chronological control on this phase of deglaciation was
159 obtained from ¹⁴C dating of sediment sequences infilling depressions along deglaciated areas of the

160 Aubrac Mountains. At the Roustières site, located in the Aubrac Plateau inside the LLGM limit (see Fig.
161 2 for the location), proglacial lake sediments are dated to the Oldest Dryas between 17.7 and 16.6 ka
162 (Gandouin et al., 2016; Ponel et al., 2016). However, the base of the sequence was not reached and thus
163 these ages provide only a minimum age constraint.

164 A glacier re-advance interrupted this retreat sequence and truncating these sediments and landforms and
165 depositing till. This glacial event is referred to as Grandvals Stade (Fig. 3C) (Poizat, 1973; Veyret, 1978;
166 Ancrenaz et al., 2020). It is demarcated in the Bès valley by the Grandvals end moraine (1059 m a.s.l.)
167 (Table 1; Fig. 5A). This landform is divided into two features: one on each valley flank. Fluvioglacial
168 deposits associated with the Grandvals end moraine formed proglacial outwash that aggraded
169 overLLGM deposits. After the glacier retreats, the Bès River deeply incised the Grandvals end moraine
170 and constructed fluvial terraces (Ancrenaz et al., 2020). Other end moraines in the Aubrac Plateau
171 encompassed truncated LLGM deposits and were correlated to the Grandvals stade (Ancrenaz et al.,
172 2020). These include the Rateylou end moraine. It is located in a flat depression between the Bès and
173 the Rimeize catchments (Table 1; Fig. 5B). This landform is composed of two subdued ridges and is
174 correlated with the Grandvals stade.

175 A final glacier advance in the Bès valley is recorded by the Bouquincan end moraine that is located up-
176 valley from the Grandvals end moraine (Table 1; Fig. 3B) (Ancrenaz et al., 2020). This end moraine
177 contains fluvio-glacial sediments deformed by glacio-tectonic processes, interpreted as the result of a
178 glacier advance in the Bès valley. All glacial deposits that were identified in the Aubrac Highlands were
179 associated with the final plateau icefield decay. For example, the Bonnecombe till demarcated the very
180 last deglaciation phase as it is located at the Bonnecombe pass (1340 m a.s.l.), one of the most elevated
181 area of the Aubrac Mountains.

182 The Aubrac plateau icefield was a climatically-sensitive system located between the western and
183 southern sectors of the Massif Central. Three distinct groups of end moraines record three glacier
184 advances in the region, especially along the Bès valley (see section 3.2), but the absolute ages of these
185 moraines (and thus the associated advances) is poorly constrained. This paper addresses this through

186 TCN dating of end moraines and erratic boulders and in doing so establishes, for the first time, a direct
187 chronological framework for the late Pleistocene glacial history of the Aubrac Mountains.

188 **4 Be-10 and Al-26 exposure dating**

189 **4.1 Sampling strategy, collection and preparation**

190 This study aims to determine the timing of three main glacier advances, as well as the timing of
191 deglaciation in the Aubrac Mountains (section 3.2). Glacial landforms and deposits in the identified from
192 the literature were checked in the field and candidates for TCN dating selected according to two
193 requirements. First, glacial landforms should be clearly identified by morphostratigraphic evidence In
194 that respect, end moraines were especially targeted (Kleman and Borgström, 1996). Secondly, the glacial
195 landform or deposit should match prerequisites for sampling, especially the availability of moraine
196 boulders and low post-depositional disturbance from natural processes or anthropogenic activities (Figs.
197 4, 5, and 6).

198 Sample collection was guided by recommendations established elsewhere (Putkonen and Swanson,
199 2003; Dunai, 2010; Heyman et al., 2016). The availability of dateable moraine boulders limited the
200 number of samples per landform. For subdued end moraines, three samples were collected when
201 available (Putkonen and Swanson, 2003). Tall boulders with their bases broadly embedded in the
202 landform were targeted (Heyman et al., 2016). Rock samples with a thickness of between one and three
203 centimetres were obtained from top and flat surfaces of moraine boulders, using a hammer and a chisel.

204 Samples were prepared for the extraction of Be-10 and Al-26 (Brown et al., 1991; Merchel and Herpers,
205 1999). Rock samples were crushed and sieved. A grain size between 0.25 and 1 mm was chosen for
206 further analysis. After magnetic separation using a Frantz Barrier separator, the nonmagnetic grains
207 (including quartz) were subjected to leaching with H_2SiF_6 and HCl to eliminate non-quartz minerals.

208 Three partial dissolutions were performed with HF to eliminate atmospheric ^{10}Be . According to Merchel
209 et al. (2008), ~ 0.15 g of a homemade (3025 ± 9 ppm) ^9Be solution was then added to the solution to fix
210 the $^{10}\text{Be}/^9\text{Be}$ ratio. To determine if an ^{27}Al spike is needed, natural concentrations of ^{27}Al were measured
211 in liquid aliquots by inductively coupled plasma–optical emission spectrometry (ICP–OES, Thermo
212 Icap6500). Quartz was totally dissolved in 40% HF solution. Afterwards, beryllium and aluminium were

213 extracted by successive anion and cation resins. Finally, the hydroxides $\text{Be}(\text{OH})_2$ and $\text{Al}(\text{OH})_3$ were
214 heated at 800°C for one hour. The resulting powders were mixed with niobium (BeO) or silver (Al_2O_3).
215 All ^{10}Be and ^{26}Al measurements were performed at the French Acceleratory Mass Spectrometry (AMS)
216 National Facility “ASTER” located at CEREGE in Aix-en-Provence, France (Arnold et al., 2010). Be-
217 10 and Al-26 were determined for all samples except for AU31, AU32 and AU33 for which only Be-10
218 measurements are available. Be-10 data were adjusted to the ASTER in-house standard, “STD11”, with
219 a Be-10/Be-9 ratio of $(1.191\pm 0.013)\times 10^{-11}$ (Braucher et al., 2015) and the Be-10 half-life of
220 $(1.387\pm 0.0012)\times 10^6$ years (Chmeleff et al., 2010; Korschinek et al., 2010). Al-26/Al-27 ratios were
221 calibrated with an ASTER in-house standard, “SM-AL-11”, with a ratio of $7.401\pm 0.064\times 10^{-12}$ (Merchel
222 and Bremser, 2004) assuming an Al-26 half-life of $7.05\pm 0.17\times 10^5$ years (Samworth et al., 1972).
223 The parameters above imply an Al-26/Be-10 spallation production ratio of $\sim 6.61\pm 0.52$. Reported
224 analytical uncertainties include counting statistics, machine stability ($\sim 0.5\%$ for ^{10}Be ; Arnold et al.,
225 2010), and blank correction. The Be-10/Be-9 and Al-26/Al-27 blank ratios were 2.45×10^{-15} ($\pm 29\%$) and
226 9.00×10^{-16} ($\pm 100\%$), for AU01 to AU17 samples respectively. The Be-10/Be-9 blank ratio was 2.60×10^{-15}
227 ($\pm 15\%$) for the AU31, AU32 and AU33 samples.

228 **4.2 Minimum Exposure Age (MEA) calculation**

229 Be-10 and Al-26 exposure ages were calculated using the CHRONUS-Earth online calculator v.3 (Balco
230 et al., 2008) (http://stoneage.ice-d.org/math/v3/v3_age_in.html). Cosmogenic Be-10 and Al-26
231 production rates at Sea Level and High Latitudes (SLHL) from spallation were computed from default
232 calibration datasets from Borchers et al. (2016), with: 4.132 ± 0.218 $\text{at}\cdot\text{g}^{-1}\cdot\text{y}^{-1}$ for Be-10 and 29.950 ± 2.955
233 $\text{at}\cdot\text{g}^{-1}\cdot\text{y}^{-1}$ for Al-26. Production rates at SLHL are scaled using the Lal-Stone time corrected scaling
234 scheme (Lal et al., 1991; Stone, 2000), the ERA40 atmosphere model and the magnetic field
235 reconstruction from Lifton (2016). Corrections for topographic shielding were applied following Dunai
236 (2010) and using the inclination of the visible horizon from each sample point extracted from a Digital
237 Elevation Model (DEM) in GIS software (ArcGis 10.3v) and compiled in Table 2. The rock density was
238 set to 2.65 $\text{g}\cdot\text{cm}^{-2}$ for all samples (Table 2). Contribution from muons were calibrated according to Balco
239 (2017).

240 Calculated exposure ages for a given nuclide concentration is named the Minimum Exposure Age
241 (MEA) because: (i) no correction for post-glacial denudation nor for snow cover were applied, and (ii)
242 simple exposure histories, without pre-exposure to cosmic rays, were assumed (see section 5.2 for more
243 details). Hereafter, MEAs were expressed with their external uncertainty that includes analytical
244 uncertainty and production rate uncertainty. In order to compare Be-10 and Al-26 MEAs obtain from a
245 glacial landform, for MEA groups with $n \leq 3$, the reduced chi square statistic (X_R^2) was used. Only Be-
246 10 MEAs were compared for the Bonnetcombe site. For a MEA group with a $X_R^2 \leq 1$, an uncertainty-
247 weighted mean was calculated with the standard deviation of the uncertainty-weighted mean. For MEA
248 groups with a $X_R^2 > 1$, the sample with the farthest MEA from the uncertainty-weighted mean was
249 manually rejected and interpreted as an outlier. Outliers are repeatedly rejected until we obtained a $X_R^2 \leq$
250 1 or we obtained a MEA group with $n < 3$ for which no uncertainty-weighted mean was calculated.
251 Manually rejected MEAs were interpreted as the effect of geomorphologic processes and are discussed
252 in section 5.2 below.

253 **5 Results and interpretations**

254 **5.1 Minimum Exposure Ages**

255 Twenty moraine boulders were sampled from six different glacial landforms. Resultant cosmogenic
256 nuclide measurements and MEAs, with their external uncertainty, were compiled in Tables 2 and 3. Two
257 boulders (AU16 and AU17) were taken from the Allatieux end moraine crests (Figs. 4A and 6). A
258 clustered group of MEAs is obtained, between 25.41 ± 2.14 to 21.34 ± 3.11 ka, with an uncertainty-
259 weighted mean of 24.29 ± 1.28 ka. Three samples (AU10, AU11 and AU13) were collected from boulders
260 embedded in the Longevialle end moraine flattened crests (Figs. 4 and 5B). MEAs range from
261 18.12 ± 2.72 to 53.86 ± 4.72 ka and provided the most scattered results in our dataset. No uncertainty-
262 weighted mean was calculated. Three samples were taken from the crest of the Grandvals end moraine
263 (AU13, AU14 and AU15) at the western end of the landform (Figs. 5A and 6). MEAs were clustered,
264 between 21.55 ± 2.05 to 12.19 ± 1.57 ka, with an uncertainty-weighted mean of 20.43 ± 0.93 ka (Table 3).
265 Only the Al-26 MEA from the AU14 sample was interpreted as an outlier. Four moraine boulders
266 embedded in two features composing the Rateylou end moraine (AU03, AU04, AU05 and AU06) gave

267 scattered MEAs from 2.17 ± 0.41 to 20.55 ± 1.84 ka (Table 3; Figs. 5B and 6). No uncertainty-weighted
268 mean was calculated. In addition, two isolated erratic boulders at the Rateylou site (AU01 and AU02)
269 were sampled near the end moraine (Fig. 5B) and MEAs ranged from 6.55 ± 0.65 to 21.48 ± 1.86 ka. At
270 the Bouquincan end moraine, three samples (AU07, AU08 and AU09) were taken from boulders with
271 low heights: < 0.5 m above the ground (Fig. 3B). Results ranged from 6.84 ± 1.41 to 18.27 ± 1.66 ka. An
272 uncertainty-weighted mean of 16.59 ± 0.9 ka was calculated with rejection of the AU09 sample. Finally,
273 samples AU31, AU32 and AU33 are collected from moraine boulders embedded in the Bonnecombe
274 glacial deposit (Figs. 5C and 6). Only ^{10}Be MEAs are available, and calculated MEAs are 18.66 ± 1.60
275 (AU31), 16.60 ± 1.43 (AU32) and 16.68 ± 1.48 ka (AU33), with a calculated ^{10}Be uncertainty-weighted
276 mean of 16.64 ± 1.03 ka.

277 **5.2 Glacial landform age**

278 The distribution of exposure ages on end moraines often displays scattered results for the same landform
279 (Heyman et al., 2011; Applegate et al., 2012). Two common explanations are identified (see Heyman et
280 al., 2011): incomplete exposure and the pre-exposure. Boulders with incomplete exposure tend to
281 underestimate ages for end moraine formation. This generally reflects post-depositional processes, such
282 as erosion (e.g. moraine matrix erosion), leading to the landform lowering and smoothing and to boulder
283 exhumation. Exhumed boulders by post-depositional processes, were exposed during a shorter period
284 than the true landform age (Zreda et al., 1994; Putkonen and Swanson, 2003; Putkonen and O'Neal,
285 2006; Chevalier et al., 2011; Heyman et al., 2011; Palacios et al., 2019; Allard et al., 2020). Then, boulder
286 with incomplete exposure the end moraine deposition This is the case for outliers with younger MEAs
287 than the uncertainty-weighted mean: AU09 (Table 3; Fig. 7). AU01, AU03 and AU04, from the two
288 Rateylou sites, provided Holocene MEAs (11.7 ka to present) that were younger than expected and
289 interpreted as moraine boulders with incomplete exposure histories (Table 3; Fig. 7). The proximity with
290 quarries (> 100 m; Fig. 5C) could lead to potential modifications of the glacial landform morphology,
291 especially exhumation of buried boulders. For the AU14 moraine boulder, only the Al-26 MEA was
292 interpreted as an outlier since the Be-10 measurement is coherent for the landform age (Table 3; Fig. 7).

293 Conversely, moraine boulders with pre-exposure provide overestimated ages for moraine deposition due
294 to inherited cosmogenic nuclides from a previous episode(s) of exposure (Heyman et al., 2011). Pre-
295 exposed boulders were estimated to account to 3% of exposure ages from datasets analysed in Putkonen
296 and Swanson (2003). In our dataset, only one moraine boulder with a much older MEA than others was
297 identified: AU11 (Table 3; Fig. 7). As this moraine boulder was associated with a glacial landform inside
298 the LLGM extent, it was interpreted as an overestimated MEA due to cosmogenic nuclide inheritance.

299 Geomorphological context provides a strong control on processes affecting moraine boulder exposure
300 history and MEA scattering (Palacios et al., 2019; Tompkins et al., 2021). In the Aubrac Mountains,
301 post-depositional erosion of end moraines is argued to be low because: (i) subdued ridges were poorly
302 affected by post-depositional processes such as run-off or slope destabilization (Putkonen and Swanson,
303 2003) and (ii) the sedimentological composition of these glacial landforms, dominated by sand, gravel
304 and boulders (Veyret, 1978; Ancrenaz et al., 2020), favoured end moraine stabilization (Tompkins et al.,
305 2021). These local geomorphological settings were concordant with a rapid stabilization of glacial
306 landforms after their deposition. As a consequence, a MEA cluster and its uncertainty-weighted mean is
307 argued to best represent the age of the moraine stabilisation (Allard et al., 2020; Tompkins et al., 2021).
308 The age of the end moraine deposition is best represented by the oldest exposure age obtained on the
309 glacial landform (Zreda et al., 1994; Putkonen and Swanson, 2003; Briner et al., 2005; Chevalier et al.,
310 2011; Heyman et al., 2011; Chevalier and Replumaz, 2019). Considering this interpretation, glacial
311 landform stabilization (best approximated by the corresponding uncertainty-weighted mean) occurred
312 rapidly after the landform deposition (best approximated by the corresponding oldest MEA) in the
313 Aubrac Mountains (Table 3; Fig. 7).

314 **5.3 Three distinct glacial stades in the Aubrac Mountains**

315 The LLGM is dated at the Allatieux end moraine to 25.41 ± 2.14 ka (Table 3). The Aubrac plateau
316 icefield covered approximately 500 km², with outlet valley glaciers in the Aubrac Valleys of 6 to 10 km
317 in length (Figs. 2 and 8) and an ELA estimated at 1331 ± 10 m a.s.l. (Ancrenaz et al., 2020). Eastward,
318 glacier margins terminate against topographic constraints (i.e., reverse slopes), which restricted plateau
319 icefield extension (Figs. 2, 3A and 8). After the LLGM, the Aubrac plateau icefield retreated first back

320 onto the plateau. A first glacial stillstand is documented by the Longevialle end moraine dated to
321 24.31 ± 2.7 ka (Table 3). According to stratigraphic observations, the extension of glacier-free areas
322 related to this glacier retreat can be estimated (Poizat and Rousset, 1975; Veyret, 1978; Ancrenaz et al.,
323 2020), and the deglaciated area after the LLGM is at least half that of the LLGM area (Figs. 8-9). The
324 Grandvals stade is dated at three sites: the Grandvals end moraine (21.55 ± 2.05 ka), the Rateylou end
325 moraine (20.55 ± 1.84) and the Rateylou isolated erratics (21.48 ± 1.86 ka) (Tables 1 and 3, Figs. 7-9).
326 These ages overlap, suggesting that these landforms are probably related to the same glacier advance
327 (Kirkbride and Winkler, 2012; Blomdin et al., 2016), as suggested by the morphostratigraphic
328 framework established for the Aubrac Mountains (Section 2.3). The last glacial stade is documented by
329 the Bouquincan end moraine and dated to 18.27 ± 1.66 ka. During this stade, the plateau icefield was
330 restricted to the Aubrac Highlands with outlet glaciers confined to the main head valleys (Fig. 8). Steps
331 of the following glacier retreat are not recognised in the field. Subsequent stillstands during glacier
332 retreat were not recognized but tills were identified over the Aubrac Highlands (Goër de Hervé et al.,
333 1994). The Bonnecombe till is one of these deposits associated with the Aubrac plateau icefield decay
334 (Table 1; Fig. 8). The ages of both the Bouquincan stade and the Bonnecombe till (18.66 ± 1.60 ka)
335 suggest a rapid deglaciation. This trend is supported by paleoenvironmental reconstruction from the
336 Roustières site (see Figs. 2 and 8) (Gandouin et al., 2016; Ponel et al., 2016), where periglacial
337 conditions in the Aubrac Plateau were active between 17.7 and 16.6 ka, with persistent winter snow
338 cover in the Aubrac Highlands.

339 **6 Chronological update of the southwestern Massif Central** 340 **chronostratigraphy: correlations and climatic inferences**

341 The direct glacial chronology presented in this work is in good agreement with the existing
342 morphostratigraphic framework established in the Aubrac Mountains (Section 2.3) and provides robust
343 chronological control of past fluctuations of the Aubrac plateau icefield. Considering that the Aubrac
344 plateau icefield is a valuable climatic proxy (Section 2.2), this new chronology highlights the main
345 climatic conditions for the southwestern Massif Central and allows us to update and revise the regional
346 chronostratigraphy.

347 6.1.1 The GS-3: establishment of full glacial conditions

348 In the western Massif Central, a successive and gradual climatic deterioration with increasingly colder
349 temperatures started with the onset of MIS 2 (Laville et al., 1986) at ~29 ka (Lisiecki and Raymo, 2005).
350 A cold culmination is recorded and primarily associated to the beginning of the MIS 2 (Laville et al.,
351 1986), and then correlated to approximately 25 ka (Van Vliet-Lanoë et al., 1991). Dating of the infilling
352 of thermal contraction cracks in discontinuous permafrost in the Aquitaine basin resulted in ages
353 between 27 and 25.5 ka BP and suggests a climatic downturn at the beginning of MIS 2 (Bertran et al.,
354 2014). This event is synchronous with the LLGM in the Aubrac Mountains, dated to 25.41±2.14 ka. This
355 is also coeval with the global LGM between 27.5 and 23.3 ka, *sensu* Hughes and Gibbard (2015),
356 Greenland Glacial Stadial 3 (GS-3) between 27.5 and 23.3 ka (Rasmussen et al., 2014), Heinrich Stadial
357 2 (HS2) between 26.5 and 24.3 ka (Sanchez Goñi and Harrison, 2010) and finally the $\delta^{18}\text{O}$ minimum of
358 the 7H record from the Sieben Hengste cave (northern Alps) centred on 25.3 ka (Luetscher et al., 2015).
359 Moisture advection from the Mediterranean Sea, enhanced by cyclogenesis, led to favourable conditions
360 for glacier growth in the circum-Mediterranean mountains (Florineth and Schlüchter, 2000; Kuhlemann
361 et al., 2008; Luetscher et al., 2015; Monegato et al., 2017), such as the Pyrenees (Delmas et al., 2011;
362 Reixach et al., 2021), the Alps (Monegato et al., 2017; Wirsig et al., 2016) and the Iberian Peninsula
363 (Domínguez-Villar et al., 2013). These cold and wet climatic conditions related to enhanced
364 Mediterranean influences probably triggered the LLGM in the Aubrac Mountains, as well a
365 simultaneous LLGM in other glaciated mountains of the southwestern Massif Central (Cantal,
366 Margeride and Lozère Mountain). In this area, gelifluction and gelifraction deposits on non-glaciated
367 slopes should be correlated to this period (Fig. 9) because effective gelifluction or gelifraction processes
368 are supported by cold and wet climatic conditions (Valadas, 1984) that reached their optimum during
369 this period. The question of an asynchronous LLGM between the southwestern (Aubrac, Cantal,
370 Margeride and Lozère Mountain) and the eastern Massif Central remains.

371 6.1.2 The GS-2.1: aridity, a limiting factor for glacier growth

372 After the global LGM, full glacial conditions characterised by a cold and arid climate (Laville et al.,
373 1986; Reille and Beaulieu, 1988; Van Vliet-Lanoë et al., 1991) prevailed across the whole Massif Central

374 during GS-2.1 (22.9 to 14.7 ka; Rasmussen et al., 2014). For example, at the Haute-Laagerie site
375 (44°57'11" N; 1°00'12"E), reconstructed anomalies for mean annual precipitation are -800 mmy⁻¹ and -
376 7.4 to -4.7 °C for mean air temperatures at ~21 ka (Lécuyer et al., 2021). These data are comparable to
377 those reconstructed from the Lake Bouchet sequence (see Fig. 1 for location), with -6 °C for the annual
378 temperature anomaly and -772 mm for the annual precipitation anomaly (Peyron et al. 1998). Between
379 ~24 to 17 ka, the Aubrac plateau icefield was in recession reflecting cold and dry climatic conditions in
380 the Massif Central. It is suggested here that the aridity recorded in the Aquitaine Basin and in the Massif
381 Central was the limiting factor for glacier growth during this period. In the Alps, the progressive
382 starvation of Mediterranean moisture triggered glacier recession (Luestcher et al., 2015; Wirsig et al.,
383 2016 and references therein). On non-glaciated slopes of the southwestern Massif Central, reduced
384 gelifluction and gelifraction processes were associated with this period and highlight the climatic aridity
385 (Valadas, 1984; Van Vliet-Lanoë et al., 1991; Degeai and Pastre, 2009) (Fig. 9). However, two glacier
386 advances at this time, the Grandvals stade (21.55±2.05 ka) and the Bouquincan stade (18.27±1.84 ka),
387 indicated two phases of glacier favourable climatic conditions. Important glacier advances in the Alpine
388 foreland at 21–20 ka were reported (Wirsig et al., 2016 and references therein). For example, the
389 Tagliamento lobe (Italy) advanced at 23–21 ka (Monegato et al., 2007) and the Durance glacier (France)
390 advanced at approximately 20 ka (Jorda et al., 2000). The Eurasian ice sheet reached its maximal extent
391 at ~21 ka (Hughes et al., 2015).

392 This period is also characterised by climatic instabilities recorded in the western Massif Central as
393 indicated by archaeological stratigraphic investigations (Laville et al., 1986). In the eastern Massif
394 Central (Velay Mountains), slope instability (dated to ~20 ka) is interpreted as possible permafrost
395 disappearance (Poiraud, 2012). This period is also coeval with the deposition of aeolian coversands in
396 the Aquitaine Basin, dated to 24 - 14 ka, in association with cold and dry climatic conditions (Sitzia et
397 al., 2015). During this phase, periods of reduced coversand accumulation, linked to reduced aeolian
398 activity, permafrost degradation and/or milder climatic conditions, were reported between 20.8 and 17.0
399 ka (Sitzia et al., 2015). The HS1 started with the progressive cooling of the North Atlantic between 19
400 and 17.4 ka (Pascual et al., 2020). Effects of this cooling in the southwestern European climate were

401 reported between 18 to 15.6 ka (Sanchez Goñi and Harrison, 2010) and were characterized by cold and
402 arid conditions (Kageyama et al., 2005; Ludwig et al., 2018). These conditions were associated with
403 discontinuous permafrost in the Aquitaine basin (Lenoble et al., 2012) and cold and dry climates in the
404 Massif Central despite longer and warmer summers (Reille and Beaulieu, 1988). The Grandvals and
405 Bouquincan stades were potentially triggered by short-lived or slight climatic changes during the GS-
406 2.1, especially a wetter period in overall dry climatic conditions. At the scale of the entire Massif Central,
407 the Recurrence Event, i.e., glacier advance following the LLGM in the western Massif Central (Sections
408 1 and 2.3), was correlated to the Grandvals stade in the Aubrac Mountains, despite uncertainties in
409 moraine correlations (Kirkbride and Winkler, 2012).

410 Finally, the decay of the Aubrac plateau icefield (18.66 ± 1.60 ka) was coeval with the onset of the HS-1
411 (18 to 15.6 ka; Sanchez Goñi and Harrison, 2010). This event was recognised as a receding glacier phase
412 in the circum-Mediterranean mountains related to enhanced aridity (Allard et al., 2021). In the
413 southwestern Massif Central, HS-1 was a period of glacier recession. In the Roustières section (Fig. 2),
414 cold but arid conditions were reconstructed between 17.7 and 15 ka (Gandouin et al., 2016), while the
415 Aubrac Mountains were probably deglaciated. Similar climatic conditions prevailed in the Cantal, the
416 Cézallier and the Monts Dore deglaciated slopes (Ponel et al., 1991; Vergne, 1991; Miras et al., 2006).
417 This is consistent with a radiocarbon age obtained in the northern Cantal on a kame terrace deposit
418 (17.1–15.7 ka cal BP; calibrated with CLAM software v.2.3.2 and the “IntCal20” calibration curve;
419 Blaauw, 2010; Reimer et al., 2020). This glacial deposit was deposited during the Cantal glacier
420 recession (Veyret-Mekdjian et al., 1978) and supported glacier retreats in the Cantal during the HS1, as
421 the Aubrac plateau icefield. At the beginning of Greenland Interstadial 1 (GI-1; at 14.7 ka; Rasmussen
422 et al., 2014), a slight summer warming ($+4^{\circ}\text{C}$, compared to Oldest Dryas temperature) was recorded at
423 the Roustières site (Gandouin et al., 2016). This climatic amelioration was coeval with a general trend
424 of slope stabilisation caused by the development of vegetation cover across the whole of the Massif
425 Central (Fig. 9) (Reille and Beaulieu, 1988; Vergne, 1991; Degeai and Pastre, 2009; Gandouin et al.,
426 2016; Ponel et al., 2016). These milder climatic conditions potentially led to glacier recession into higher
427 elevation cirques of the Cantal and Monts Dore massifs (Veyret, 1978; Valadas, 1984; Vergne, 1991).

428 **7 Conclusion**

429 Five end moraines and one till in the Aubrac Mountains were selected, in accordance with field
430 investigations, for surface exposure dating using ^{10}Be and ^{26}Al ($n = 20$). The new glacial chronology
431 produced for the Aubrac plateau icefield southwestern Massif Central is in good agreement with (i) the
432 existing morphostratigraphy previously established in the Aubrac Plateau and (ii) other
433 paleoenvironmental records from the Massif Central and more widely in western Europe. The new data
434 indicate that the LLGM occurred during the early part of MIS 2 (28–23 ka). It was followed by two
435 glacier advances, referred to as Grandvals and Bouquincan stades, during GS-2.1 at 24–19 ka and 20–
436 16 ka, respectively. The Aubrac plateau icefield decay is dated to GS-2.1a at 18–14 ka, in accordance
437 with local paleoenvironmental data. The glacial chronology presented here is the first comprehensive
438 chronological framework for the late Pleistocene glaciation in the French Massif Central. First, the
439 LLGM in the Aubrac Mountains is synchronous with favourable conditions for glacier expansion in
440 Central Europe (Alps) related to reduced North Atlantic circulation and enhanced moisture advection
441 from the Mediterranean Sea that reached at least the southern Massif Central. Second, GS-2.1 in the
442 Massif Central is characterised by a cold and arid climate with possible wetter climate phases that
443 potentially led to two glacier advances: the Grandvals stade in the Aubrac Mountains (named the
444 Recurrence Event in the Cantal, Cézallier and Monts Dore) and the Bouquincan stade in the Aubrac
445 Mountains. These phases are associated with the potential northward shift of the westerlies. This may
446 have triggered glacier advances in the Aubrac Mountains. Aridity during this period was the main limiting
447 factor for glacier growth, especially during HS1, which is coeval with Aubrac plateau icefield decay and
448 glacier retreat recorded in the north Cantal.

449 **8 Acknowledgements**

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457 for their AMS expertise. B. Chandler and an anonymous reviewer are warmly thanked for their
458 constructive comments and their help to improve this paper.

459

460 **9 Tables**

461

462 **Table 1.** Summary of key glacial landforms identified in the field fulfil criteria for cosmogenic exposure dating, with their significance in the existing
 463 morphostratigraphic framework for the Aubrac plateau icefield.

Local glacier stade		Glacial landform targeted for dating		
Local name	Field evidence	Geomorphological field survey	<i>Name</i> (type of landform)	Location
Local Last Glacial Maximum	- in the Aubrac Plateau: maximum dispersion of erratic boulders, terminal moraines, - in Aubrac Valleys: maximum extent of glacially moulded morphologies and terminal moraine remnants.	<i>Veyret, 1978</i> <i>De Goër et al., 1994</i> <i>Poizat and Rousset, 1974</i> <i>Ancrenaz et al., 2020</i>	<i>Allatieux</i> (end moraine)	Aubrac Plateau Rimeize watershed 1199 m
			<i>Longevialle</i> (end moraine)	Aubrac Plateau Bès valley 1040 m
Grandvals stade	- in the Aubrac Plateau: LLGM glacier melt-out sediments truncated by till and recessional moraines.	<i>Veyret, 1978</i> <i>Poizat, 1973</i> <i>Ancrenaz et al., 2020</i>	<i>Grandvals</i> (end moraine)	Aubrac Plateau Bès valley 1059 m
			<i>Rateylou</i> (end moraine)	Aubrac Plateau Rimeize watershed 1185 m
Bouquincan stade	- in the Aubrac Plateau: Bouquincan recessional moraine.	<i>Ancrenaz et al., 2020</i>	<i>Bouquincan</i> (end moraine)	Aubrac Plateau Bès valley 1157 m
Deglaciation	- in the Aubrac Highland: glacial deposits in the uppermost reaches of the Aubrac Valleys.	<i>Veyret, 1978</i> <i>De Goër et al., 1994</i> <i>Ancrenaz et al., 2020</i>	<i>Bonnecombe</i> (glacial deposit)	Aubrac Highlands Doulounet valley 1345 m

464

Table 2. Location of the sampling sites and analytical data for Be-10 and Al-26 samples from the Aubrac Mountains.

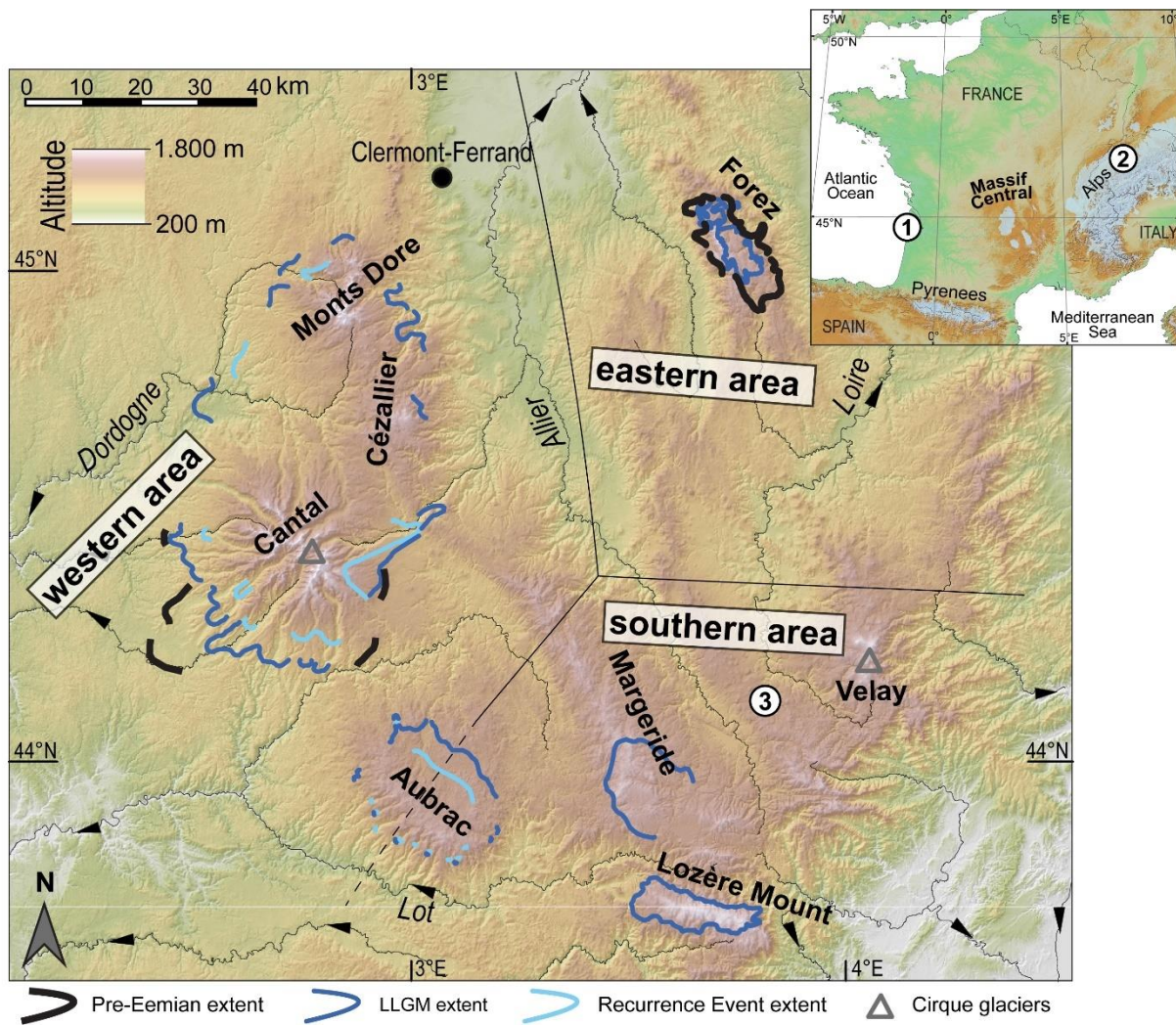
<i>Local glacier stade</i>	<i>Landform & sample ID</i>	<i>Longitude</i>	<i>Latitude</i>	<i>Altitude (m a.s.l.)</i>	<i>Topographic shielding</i>	<i>Rock density (g.cm⁻²)</i>	<i>Sample thickness (cm)</i>	<i>[¹⁰Be] (×10⁻³ at.g⁻¹)</i>	<i>[²⁶Al] (×10⁻³ at.g⁻¹)</i>
	<i>Allatieux moraine</i>								
	AU16	3°13'52.9"	44°72'37.6"	1196	0.99	2.65	3.0	270.29 ± 7.6	1686.28 ± 192.2
	AU17	3°13'88.4"	44°72'12.9"	1192	0.99	2.65	2.0	266.75 ± 8.8	1559.80 ± 155.3
LLGM	<i>Longevialle moraine</i>								
	AU10	3°03'05.9"	44°76'50.7"	1039	0.99	2.65	2.0	230.70 ± 17.7	1343.50 ± 114.6
	AU11	3°02'93.8"	44°76'51.8"	1041	0.99	2.65	4.0	500.17 ± 17.5	2538.08 ± 125.0
	AU12	3°03'31.3"	44°76'50.8"	1036	0.99	2.65	4.0	176.16 ± 9.9	1156.15 ± 123.2
	<i>Grandvals moraine</i>								
	AU13	3°02'85.7"	44°74'24.9"	1055	0.99	2.65	3.0	202.19 ± 7.7	1257.79 ± 84.1
	AU14	3°02'88.4"	44°74'25.5"	1057	0.99	2.65	2.0	207.31 ± 10.8	804.75 ± 60.5
	AU15	3°02'86.9"	44°74'25.8"	1055	0.99	2.65	2.5	192.84 ± 8.7	1250.80 ± 95.9
	<i>Rateylou isolated erratic boulder</i>								
Grandvals stade	AU01	3°10'14.0"	44°69'77.4"	1195	0.99	2.65	3.5	107.28 ± 4.6	450.02 ± 47.5
	AU02	3°11'01.8"	44°69'74.7"	1192	0.99	2.65	3.0	228.22 ± 7.8	968.18 ± 75.0
	<i>Rateylou moraine</i>								
	AU03	3°10'93.0"	44°69'75.1"	1190	0.99	2.65	2.5	34.23 ± 2.7	159.10 ± 24.8
	AU04	3°10'93.1"	44°69'74.7"	1190	0.99	2.65	2.5	70.06 ± 4.2	393.36 ± 39.2
	AU05	3°11'17.3"	44°69'66.5"	1187	0.99	2.65	2.5	218.47 ± 9.1	1344.09 ± 72.1
	AU06	3°11'34.8"	44°69'61.6"	1183	0.99	2.65	3.0	207.74 ± 9.7	1087.26 ± 70.5
	<i>Bouquincan moraine</i>								
Bouquin- can stade	AU07	3°09'32.3"	44°66'15.5"	1158	0.99	2.65	4.0	172.94 ± 11.6	1051.46 ± 67.5
	AU08	3°09'25.3"	44°66'22.0"	1156	0.99	2.65	2.5	189.55 ± 8.4	1123.39 ± 76.2
	AU09	3°09'17.0"	44°66'25.4"	1153	0.99	2.65	3.5	75.28 ± 4.3	483.40 ± 85.2
	<i>Bonnecombe till</i>								
Deglacia- tion	AU31	3°12'28.1"	44°55'97.6"	1341	0.99	2.65	3.5	221.31 ± 7.2	n.a.
	AU32	3°12'65.9"	44°56'16.2"	1342	0.99	2.65	2.5	198.66 ± 6.7	n.a.
	AU33	3°12'66.5"	44°55'98.8"	1339	0.99	2.65	2.0	199.92 ± 7.9	n.a.

Table 3. Be-10 and Al-26 Minimum Exposure Age (MEA) results of samples from the Aubrac Mountains

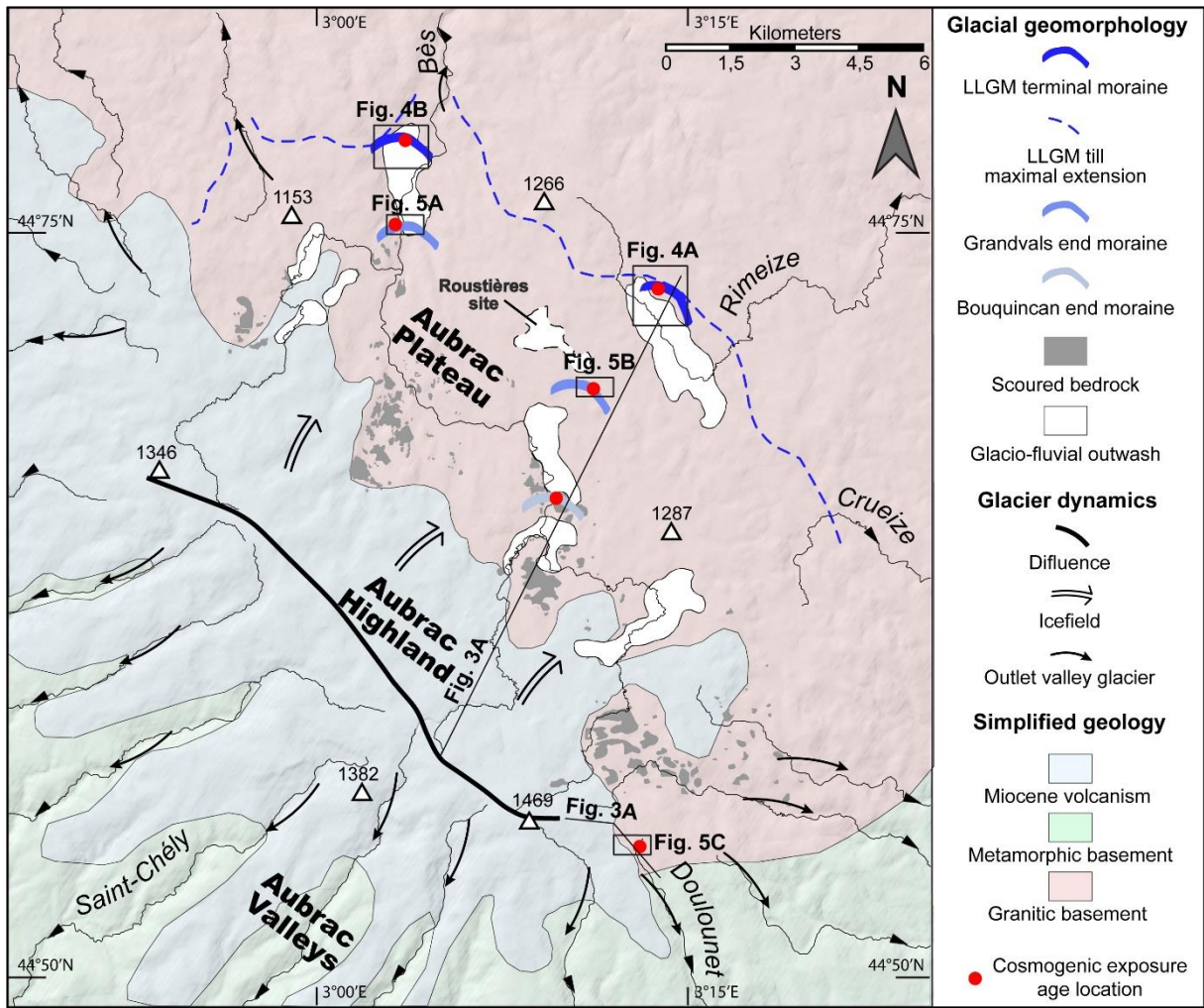
<i>Local glacier stade</i>	<i>Landform & sample ID</i>	$^{26}\text{Al}/^{10}\text{Be}$	^{10}Be MEA (ka)	^{26}Al MEA (ka)	<i>Landform stabilization (uncert.-weighted mean in ka)</i>	<i>Landform age (oldest MEA in ka)</i>		
LLGM	<i>Allatieux end moraine</i>							
		AU16	6.2 ± 0.7	25.41 ± 2.14	23.21 ± 3.62	24.29±1.28	25.41 ± 2.14	
		AU17	5.8 ± 0.6	24.95 ± 2.15	21.34 ± 3.11			
		<i>Longevialle end moraine</i>						
		AU10	5.8 ± 0.7	24.31 ± 2.7	20.7 ± 2.81	n.a.	24.31 ± 2.7	
		AU11*	5.1 ± 0.3	53.86 ± 4.72	40.05 ± 4.7			
	AU12	6.6 ± 0.8	18.88 ± 1.84	18.12 ± 2.72				
Grandvals stade	<i>Grandvals end moraine</i>							
		AU13	6.2 ± 0.5	21.22 ± 1.87	19.31 ± 2.41	20.43±0.93	21.55 ± 2.05	
		AU14**	3.9 ± 0.4	21.55 ± 2.05	12.19 ± 1.57			
		AU15	6.5 ± 0.6	20.14 ± 1.84	19.11 ± 2.49			
		<i>Rateylou isolated erratic boulders</i>						
		AU01***	4.2 ± 0.5	10.09 ± 0.91	6.17 ± 0.92	n.a.	21.48 ± 1.86	
		AU02	4.2 ± 0.4	21.48 ± 1.86	13.29 ± 1.73			
		<i>Rateylou end moraine</i>						
		AU03***	4.6 ± 0.8	3.2 ± 0.36	2.17 ± 0.41	19.68±1.11	20.55 ± 1.84	
		AU04***	5.6 ± 0.7	6.55 ± 0.65	5.36 ± 0.77			
	AU05	6.2 ± 0.4	20.55 ± 1.84	18.49 ± 2.18				
	AU06	5.2 ± 0.4	19.67 ± 1.81	15.03 ± 1.86				
Bouquincan stade	<i>Bouquincan end moraine</i>							
		AU07	6.1 ± 0.6	16.83 ± 1.75	14.95 ± 1.84	16.59±0.9	18.27 ± 1.66	
		AU08	5.9 ± 0.5	18.27 ± 1.66	15.82 ± 1.98			
	AU09***	6.4 ± 1.2	7.31 ± 0.72	6.84 ± 1.41				
Bonnecombe deposit	<i>Bonnecombe till</i>							
		AU31	n.a.	18.66 ± 1.6	n.a.	17.23±0.87	18.66 ± 1.6	
		AU32	n.a.	16.6 ± 1.43	n.a.			
	AU33	n.a.	16.68 ± 1.48	n.a.				

468 * MEA rejected for the uncertainty weighted-mean calculation and interpreted as prior-exposure; ** Al-26 MEA rejected for the uncertainty-weighted mean
469 calculation; *** MEA rejected for the uncertainty-weighted mean calculation and interpreted as incomplete exposure age

470 **10 Figure captions**

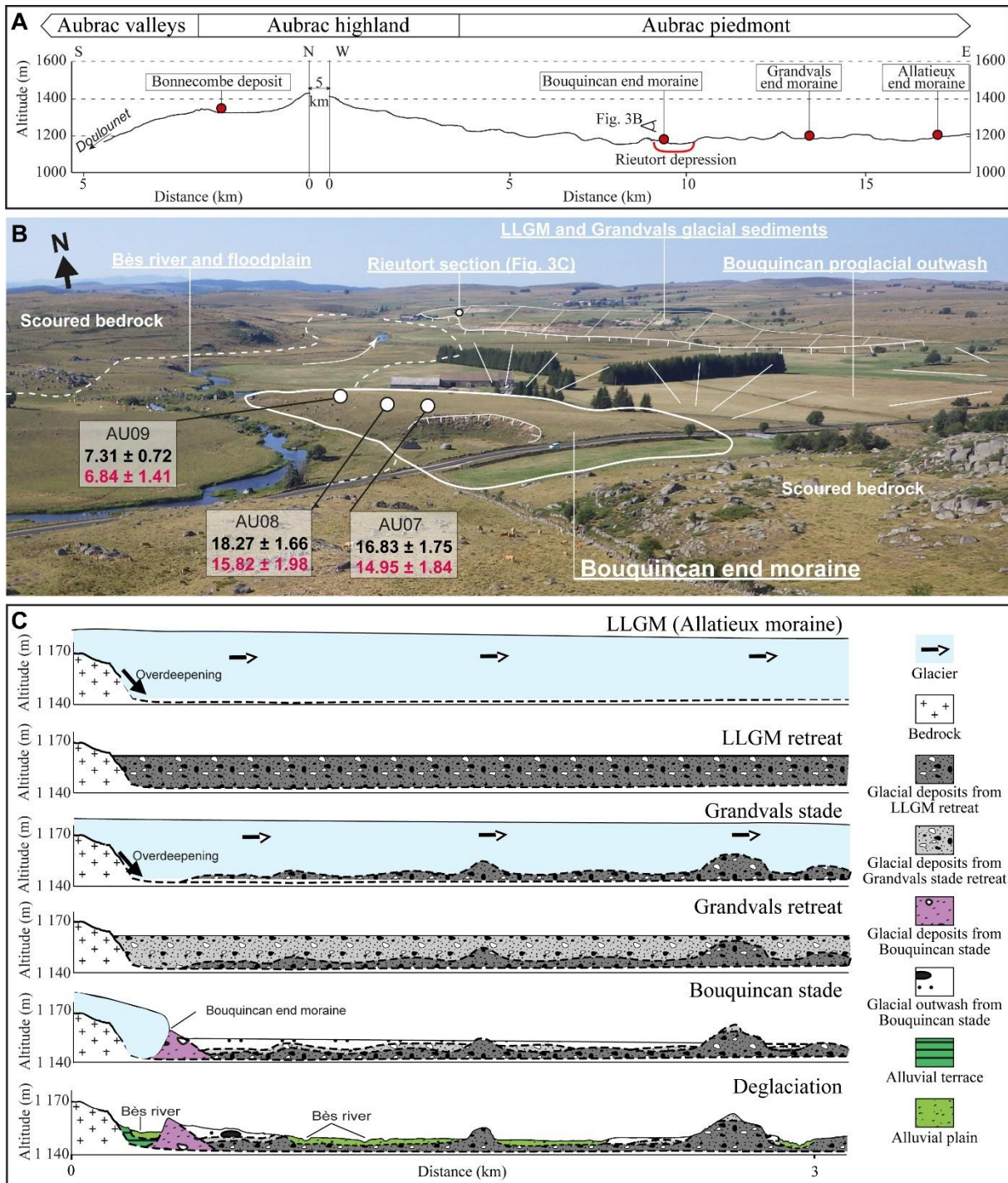


472 **Figure 1.** Extent and chronology of late Pleistocene glaciation in the Massif Central (Etlicher and De
473 Goër, 1988) and the three groups of glaciated areas according to Veyret (1978). Localities cited in the
474 main text are as follows: 1, the Aquitaine Basin; 2, the Sieben Hengste Cave 7H speleothem record
475 (Luetscher et al., 2015); and 3, the Lake Bouchet pollen sequence (Reille and Beaulieu, 1988).



476

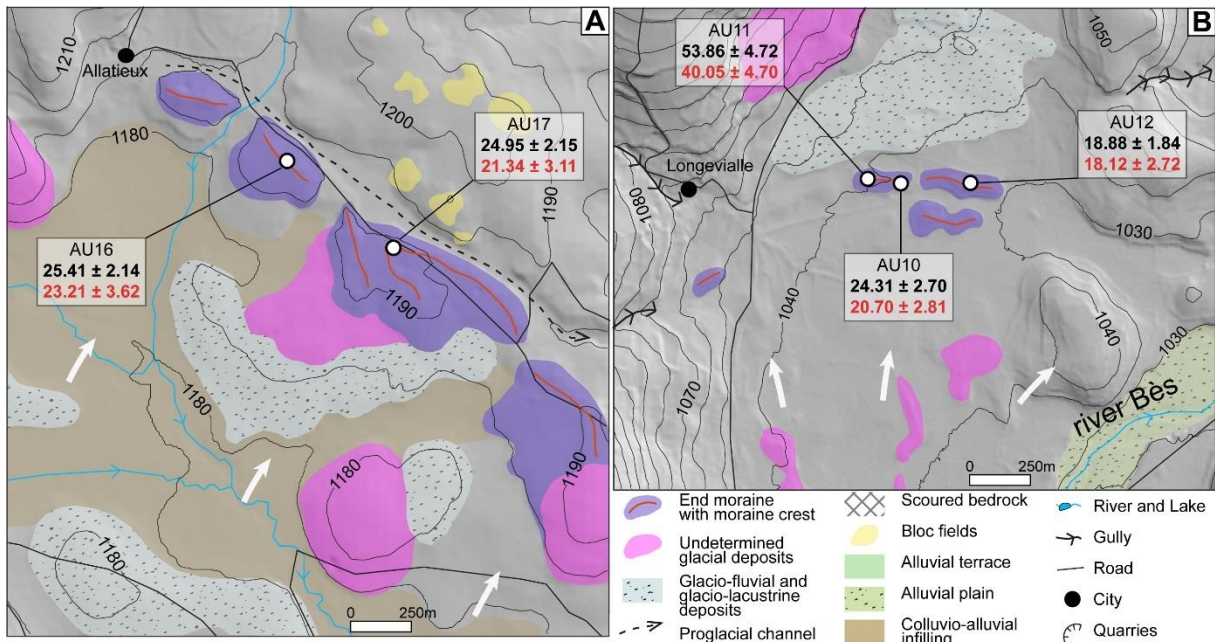
477 **Figure 2.** Aubrac Mountains lithostructural context and glacial geomorphology, with the location of
 478 sample sites for cosmogenic exposure age dating.



479

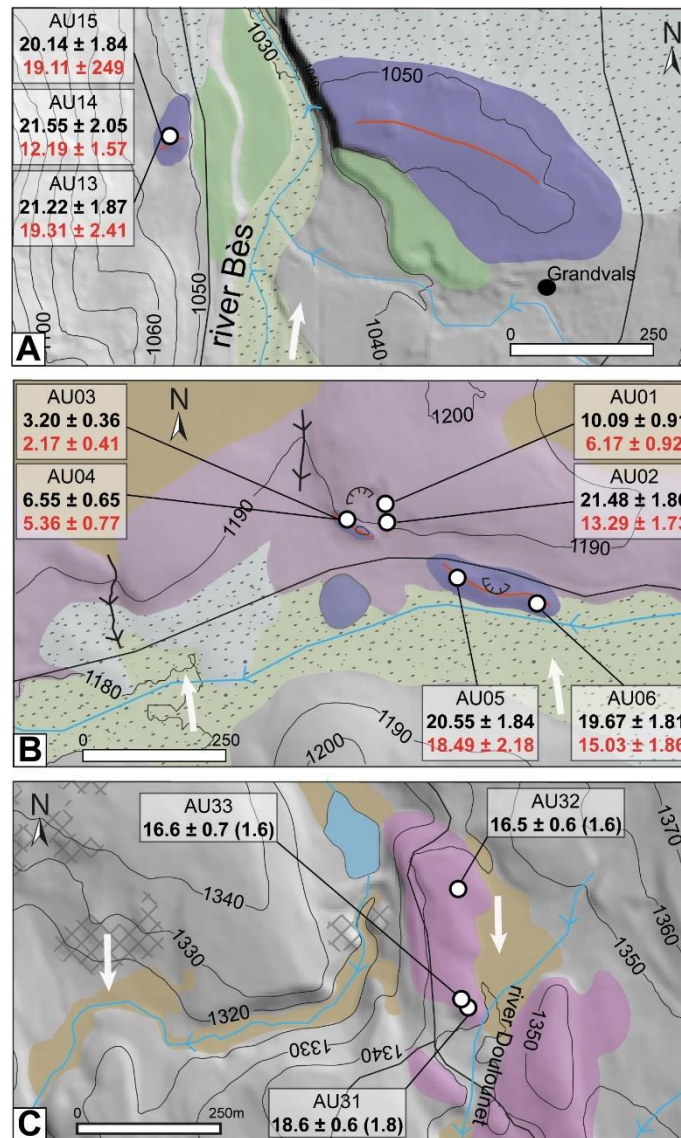
480 **Figure 3.** Locations of the main glacial landforms and sediments in the Aubrac Mountains from which
 481 the morphostratigraphic framework was built. A) Topographic profile from south to east across the
 482 Aubrac Mountains showing the locations of ice-marginal landforms and dated glacial deposits. For the
 483 location of the profiles, see Fig. 2. B) Annotated photograph of the glacial landform assemblages
 484 identified in the Rieutort depression (Poizat, 1973; Veyret, 1978; Ancrenaz et al., 2020). The upper
 485 (black) and lower (red) ages refer to ^{10}Be and ^{26}Al minimum exposure ages, respectively. C) Glacial

486 fluctuations reconstructed from the stratigraphic assemblages from the Rieutort section, located in the
487 Rieutort depression (modified from Ancrenaz et al., 2020).



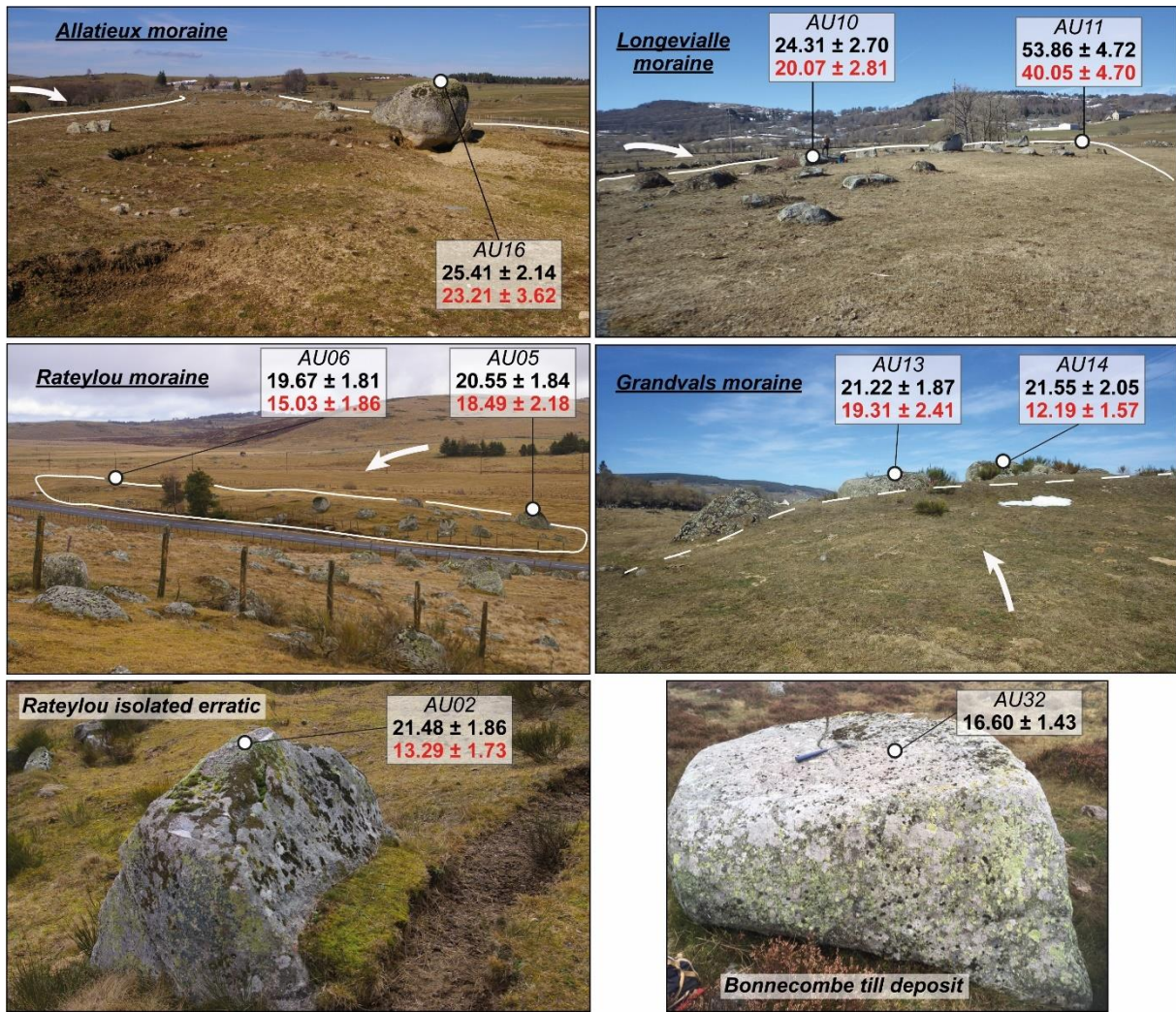
488

489 **Figure 4.** Geomorphological settings of (A) the Allatieux and (B) Longevialle end moraines, with the
 490 locations of sampled moraine boulders. Upper (black) and lower (red) ages refer to ^{10}Be and ^{26}Al
 491 MEAs, respectively, with their uncertainty (analytical and production rate). White arrows indicate the
 492 direction of the glacier flow. For the location of the captions, see Fig. 2.



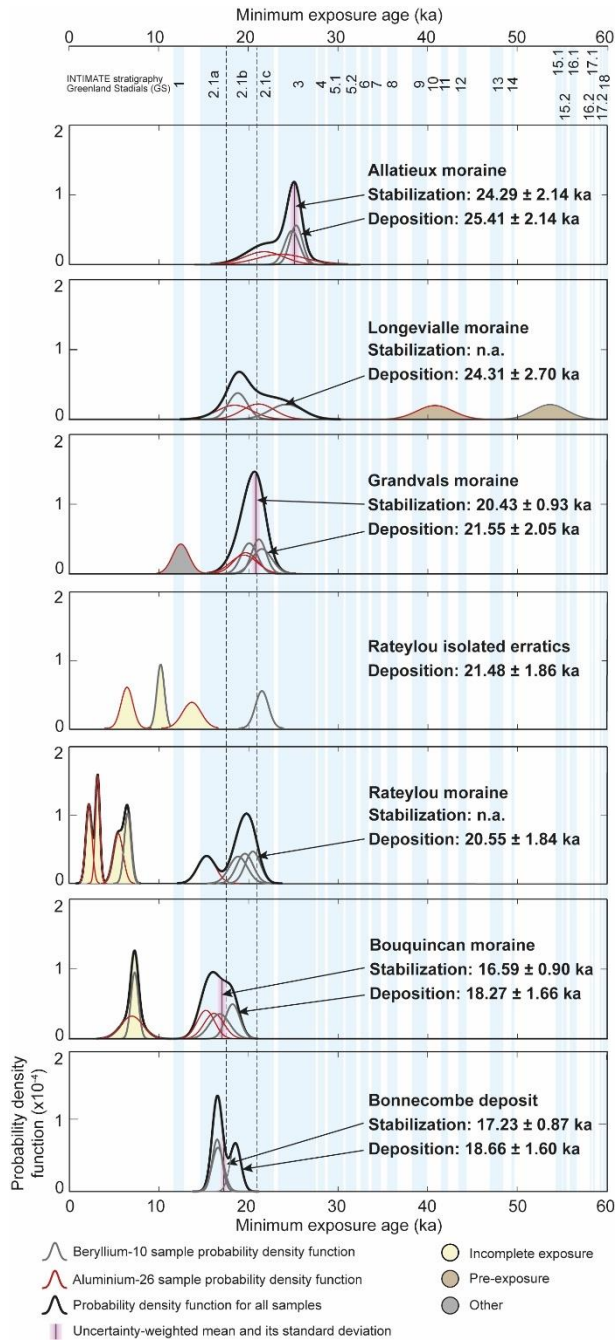
493

494 **Figure 5.** Geomorphological settings of (A) the Grandvals end moraine, (B) the Rateylou end moraine
 495 and associated isolated boulders and (C) the Bonnecombetill, with sample locations of morainic
 496 boulders. Upper (black) and lower (red) ages refer to ^{10}Be and ^{26}Al MEAs, respectively, with their
 497 uncertainty (analytical and production rate). White arrows indicate the direction of glacier flow.. For
 498 the location of the captions, see Fig. 2. Same legend as that for Fig. 4.



499

500 **Figure 6.** Photographs of sampled granitic erratic boulders embedded in glacial landforms in the
 501 Aubrac Mountains. Only sample AU02 is collected from an isolated erratic boulder. The upper (black)
 502 and lower (red) ages refer to ^{10}Be and ^{26}Al minimum exposure ages, respectively. Glacier flow is
 503 indicated by the white arrows.



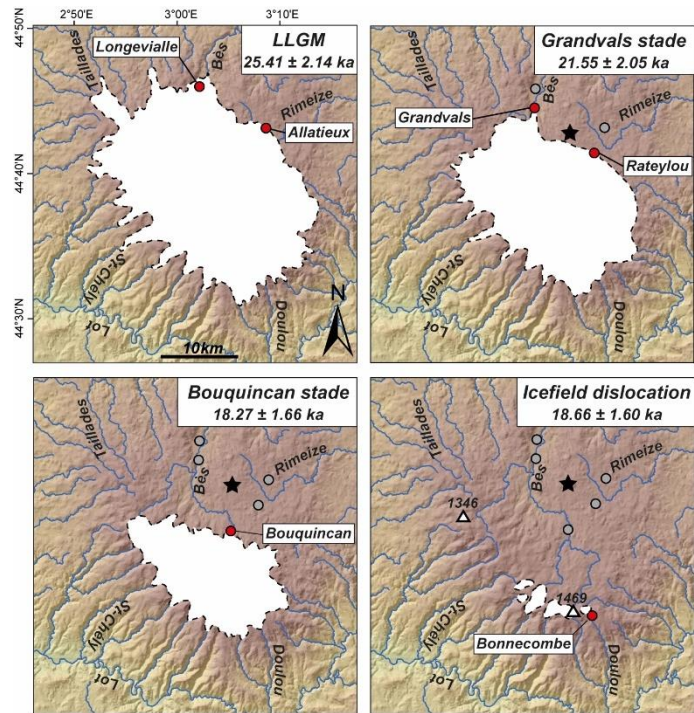
504

505

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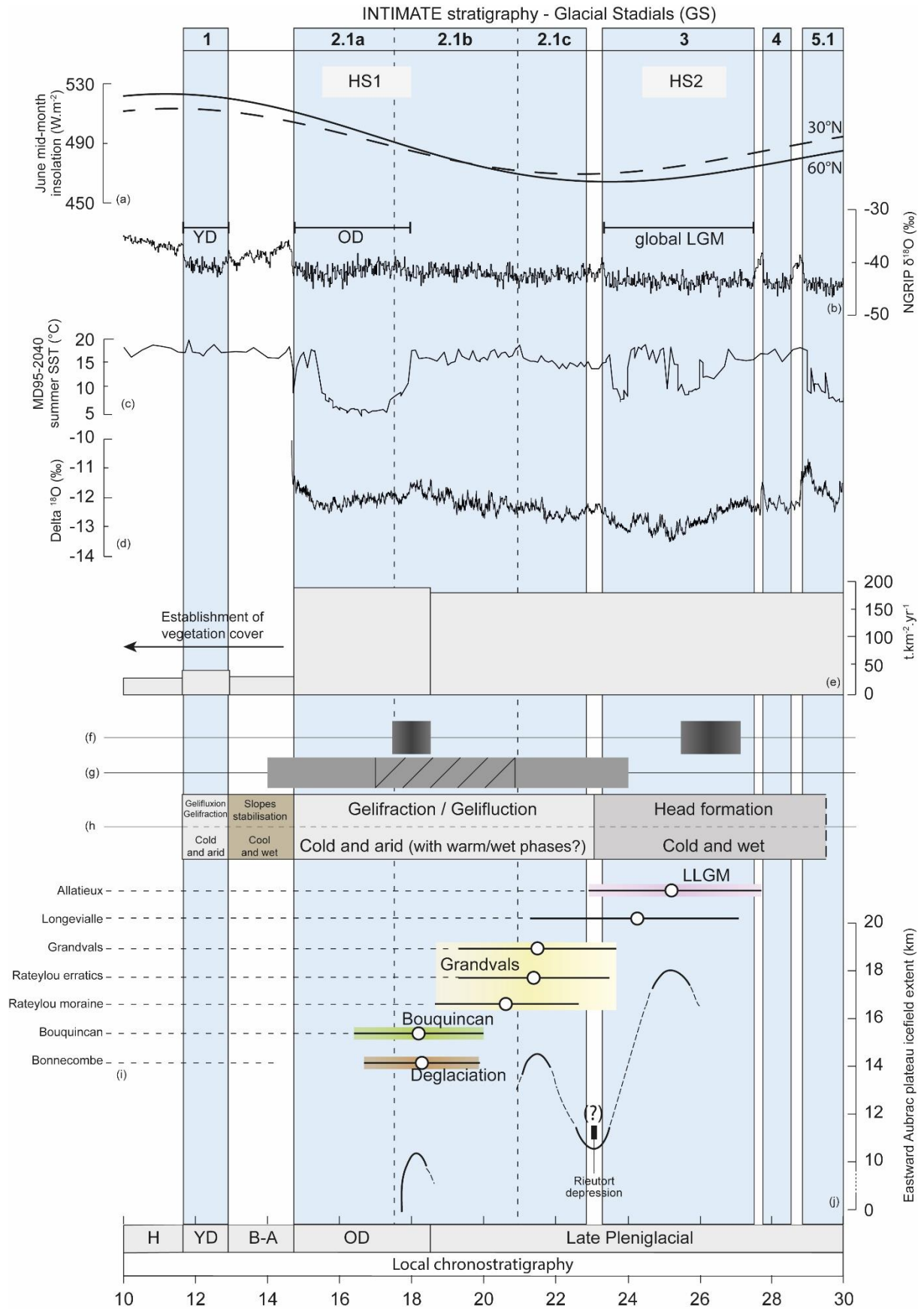
507

Figure 7. Probability density function of MEAs by location versus stadials from the INTIMATE stratigraphy (Rasmussen et al., 2014). The landform age is inferred from the oldest MEA that is not rejected by the X_R^2 test (Section 6.1). Only the internal uncertainty is shown.



508

509 **Figure 8.** Diachronic reconstructions of the Aubrac plateau icefield extent during MIS 2. Red dots are
 510 glacial landforms whose exposure ages indicate the minimum age of deglaciation. Glacial limits are
 511 uncertain and based on geomorphological (subglacially moulded morphologies and topographic
 512 constraints) and sedimentological observations (extension of till) (Poizat and Rousset, 1975; Veyret,
 513 1978; Ancrenaz et al., 2020). The black star represents the location of the Roustière site (Gandouin et
 514 al., 2016; Ponel et al., 2016); note that it is located inside the LLGM outer limit (see Section 2.3 for
 515 further details).



516

517

Figure 9. Synthetic chronological framework for the southwestern Massif Central against main

518 paleoenvironmental proxies cited in the main text. a) June mid-month insolation in $W.m^{-2}$ at $60^{\circ}N$
519 (black line) at $30^{\circ}N$ (dashed line) (Berger and Loutre, 1991), b) North Greenland Ice Core Project
520 (NGRIP) $\delta^{18}O$ from the Greenland ice core (Rasmussen et al., 2014). c) Summer Sea Surface
521 Temperature (SST) reconstructed from the MD95-2040 oceanic sediment core from the western
522 Iberian Margin (Salgueiro et al., 2010). d) Sieben Hengste $\delta^{18}O$ record (Luetscher et al., 2015). e)
523 Erosion rate in the Lake Bouchet watershed (Degeai and Pastre, 2009). f) Timing of episodes of
524 permafrost in southwestern France (Lenoble et al., 2012; Bertran et al., 2014). g) Timing of aeolian
525 sand cover accumulation in the Aquitaine Bassin (southwestern France). The dashed part corresponds
526 to period of reduced aeolian activity related to permafrost degradation (Sitzia et al., 2015). h) Resumed
527 stratigraphic chronology for the Massif Central nonglaciaded slopes (Van Vliet-Lanoë et al., 1991). i)
528 Ages of glacial landforms in the Aubrac Mountains according to exposure ages for erratic boulders
529 (this study). j) Reconstructed extension of the eastern part of the Aubrac plateau icefield. For the
530 boundaries of the LGM, the Oldest Dryas (OD), the Younger Dryas (YD) and of the Heinrich Stadials
531 1 and 2 (HS-1 and -2), see Hughes and Gibbard (2015), Shakun and Carlson (2010), Carlson (2013)
532 and Sanchez Goñi and Harrison (2010), respectively.

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