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Miocene cave sediments record topographic, erosional and drainage development in the Western European Alps

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Cave sediment record, Cosmogenic nuclides burial dating, Detrital thermochronology*

Abstract

The dynamic Neogene evolution of the Western European Alps included exhumation of the external crystalline massifs, thrust propagation to the foreland, drainage network reorganization, and major climatic variations. To constrain possible interactions between those factors, accurate geomorphological and sedimentological archives are required. However, intra-orogenic areas are subject to erosion, and extensive glacial cover during the Quaternary erased most of the geomorphic markers in the Alps. For these reasons, the genesis of the main features of the modern landscape, such as the major valleys and the drainage network, remains poorly understood. This study highlights how recently discovered karstic archives from the perched paleo-karst of the Obiou peak (Dévoluy massif, SE France) record the tectonic and drainage-network evolution of this part of the Alps during the Neogene. The Obiou caves are located at 2250-2380 m elevation, ~1600 m above the Drac valley; they contain fluvial deposits including sand-clay units and rounded crystalline cobbles derived from the adjacent Ecrins-Pelvoux massif. As the Dévoluy and Ecrins-Pelvoux massifs are currently separated by the axial Drac valley (a major tributary of the Isère River), these cave sediments must have been deposited by a radial drainage system before incision of the modern Drac. We report new

33 multi-method results from these sediments, including cosmogenic-nuclide burial dating (^{21}Ne , ^{10}Be ,
34 ^{26}Al in quartz), provenance analysis (clast petrography and heavy-mineral analysis), and detrital
35 thermochronology (apatite fission-track and (U-Th)/He) combined with a paleo-environmental
36 reconstruction from palynology. $^{21}\text{Ne}/^{10}\text{Be}$ dating of cobbles and sand constrains the burial age to
37 11.5 ± 1.5 Ma, providing a maximum age for the modern axial drainage system and a minimum long-
38 term incision rate of ~ 140 m/Myr for the Drac valley. Comparison of the combined data to both
39 modern rivers and nearby Oligocene foreland-basin deposits provides evidence for two successive
40 drainage reorganizations. Early Miocene exhumation and development of high topography in the
41 Ecrins-Pelvoux massif, linked to localized thrusting on a crustal-scale ramp, led to initial deflection of
42 the antecedent radial drainage network, beheading its headwaters by establishment of the axial
43 upper Durance valley. Subsequent propagation of thrusting into the subalpine Dévoluy massif and
44 associated uplift during the mid to late Miocene led to establishment of the modern drainage
45 system.

46 1. Introduction

47 The fluvial drainage network sets the base level and ensures the redistribution and
48 evacuation of water and sediment fluxes generated upstream in mountain belts. The efficiency of
49 transport processes along the drainage network therefore limits erosional fluxes, controlling
50 topographic development and *in fine* the balance between rock uplift and denudation (*Whipple et*
51 *al., 1999*). Changes in drainage patterns influence sediment deposition in basins (e.g., *Kuhlemann et*
52 *al., 2006; Allen, 2008*) and can affect evolution and biodiversity (e.g., *Craw et al., 2016*). The current
53 drainage pattern of the Western European Alps appears suboptimal in its function because several
54 trunk rivers, including the Rhône, Rhine, and Isère, flow parallel to the axis of the orogen, and
55 therefore perpendicular to its overall gradient for an extended part of their course (Fig. 1A). The
56 origin of such axial drainage, which contrasts to the more common radial drainage pattern, has been
57 a major question in geomorphology for decades (e.g., *Oberlander and Morisawa, 1985*). In the
58 French Western Alps, the initiation and timing of incision of the near-continuous axial Isère and Drac
59 valleys, locally known as the “*Sillon Subalpin*”, have been discussed since the middle of the last
60 century (e.g., *Blanchard, 1947; Debeltmas, 1995*). However, none of the evolutionary scenarios
61 proposed (involving river capture, glacial incision, or gravitational collapse) satisfactorily explain the
62 current drainage pattern.

63 Fluvial drainage reconstructions based on the sedimentary record imply that the early
64 Miocene drainage system of the European Alps was mostly radial (*Schlunegger et al., 1998;*
65 *Kuhlemann et al., 2006; Jourdan et al., 2012*), suggesting major drainage reorganization during the

66 Neogene. The transition time between these two drainage configurations remains to be elucidated,
67 as do the potential links to the topographic evolution and the driving mechanisms for these changes.
68 The main Alpine axial valleys are systematically located adjacent to the External Crystalline Massifs
69 (ECM), where the highest topographies and present-day surface uplift rates in the mountain range
70 are found (*Sternai et al., 2019* and references therein). The ECM started uplifting and exhuming
71 during the late Oligocene Miocene (*van der Beek et al., 2010; Glotzbach et al., 2011a; Herwegh et al.,*
72 *2019; Valla et al., 2021*). This spatial and temporal coincidence raises the prospect of an integrated
73 geodynamic model to link tectonics, topographic development, and drainage reorganization in the
74 Western European Alps.

75 To this aim, reliable geomorphic and sedimentological archives are required, but extensive
76 Quaternary glaciations and valley incision (e.g., *Valla et al., 2011*) have erased most of these markers
77 within the mountain belt, whereas foreland-basin sediments (e.g., *Kuhlemann and Kempf, 2002;*
78 *Glotzbach et al., 2011b*) only provide range-scale integrated records. Karstic systems may, however,
79 host remarkable archives and are particularly well suited to studying the relief history of mountain
80 ranges. In particular, several studies have related cave elevations to paleo-base levels and used
81 terrestrial cosmogenic nuclide (TCN) dating to recover past fluvial incision rates in various contexts
82 (e.g., *Granger et al., 1997; Haeuselmann et al., 2007; McPhillips et al., 2016; Sartégou et al., 2018*).
83 Complementary data from karstic archives, such as thermochronology or petrography, can provide
84 key information about sediment provenance and/or paleo-exhumation rates (*Sauro et al., 2021*).
85 However, the quantitative analysis of karstic archives is challenging due to limitations in the TCN
86 dating methods (*Ma et al., 2018*), commonly limited cave preservation and access, and the potential
87 complexity of the karst system (*Malcles et al., 2020*).

88 Here, we use a multi-method approach to characterize Alpine-basement derived sediments
89 that were recently discovered in caves of the Dévoluy massif, SE France (*Jagercikova et al., 2021; Fig.*
90 *1B, C*). We present new ^{21}Ne - ^{10}Be - ^{26}Al burial dating, detrital apatite fission-track (AFT) and apatite (U-
91 Th-Sm)/He (AHe) thermochronology, as well as petrographic and heavy-mineral data. Combined with
92 previously published pollen analyses (*Jagercikova et al., 2022*), these data are used to propose a
93 consistent model for the Miocene topographic, erosional and river-drainage development of the
94 French Western Alps.

95 2. Geological and morphological setting

96 The European Alps result from the collision between Europe and Adria since the early
97 Paleogene, and are composed of a number of tectonic units corresponding to different

98 paleogeographic domains and metamorphic facies. The Western European Alps contain two major
99 domains: the external zones, corresponding to the former European rifted margin, and the internal
100 zones, corresponding to the former Piedmont-Ligurian Ocean and Briançonnais continental sliver
101 (e.g., *Masini et al.*, 2013). The internal zones have undergone subduction-related metamorphism
102 during the Eocene (*Rosenbaum and Lister*, 2005), whereas the external zones exhibit only very low-
103 grade Alpine metamorphism (*Bellanger et al.*, 2015). The external zones comprise basement units of
104 the External Crystalline Massifs (ECM), overlain by Mesozoic marly and calcareous sediments that
105 have been folded and faulted to form the Subalpine massifs, and Cenozoic synorogenic molasse
106 deposits (Fig. 1).

107 Several observations allow reconstructing the Alpine tectonic evolution of the ECM and,
108 more specifically in the scope of this study, the Ecrins-Pelvoux Massif (Fig. 1). The deposition of
109 Upper-Eocene (Priabonian; 38-34 Ma) flysch, known as the Champsaur sandstone, unconformably
110 overlying a paleo-topography (*Gupta and Allen*, 2000) implies that basement rocks were locally at the
111 surface during that period. Subsequent tectonic burial under internal-zone units and thickening
112 related to shortening led to low-grade metamorphic conditions, as recorded by phengite formation
113 in shear zones dated by $^{40}\text{Ar}/^{39}\text{Ar}$ to between 34 and 25 Ma (*Simon-Labric et al.*, 2009; *Bellanger et*
114 *al.*, 2015). Paleomagnetism and zircon fission-track (ZFT) thermochronometry record an onset of
115 exhumation for the Ecrins-Pelvoux massif between 30 and 24 Ma (*Crouzet et al.*, 2001; *van der Beek*
116 *et al.*, 2010), with basement rocks that are currently at the surface being exhumed through the AFT
117 closure depth of around 4 km between 10 and 3 Ma (*Seward et al.*, 1999; *van der Beek et al.*, 2010;
118 *Beucher et al.*, 2012). During mid-late Miocene times, thrusting and shortening propagated to the
119 Subalpine massifs (*Philippe et al.*, 1998; *Girault et al.*, 2022; *Bilau et al.*, 2023). In the Vercors and
120 Chartreuse massifs (Fig. 1), syn-tectonic deposition of molasse sediments in front of the active
121 thrusts records southeast-to-northwest propagation of deformation between the Burdigalian (20-16
122 Ma) and the Pliocene (*Kalifi et al.*, 2022). The major thrusts of the Subalpine massifs disappear
123 southwest of the Pelvoux massif, where another structure, the Digne thrust sheet, accommodated
124 shortening during Miocene times (*Schwartz et al.*, 2017; Fig. 1B).

125 In contrast to their geodynamic and tectonic evolution, the morphologic and topographic
126 development of the Western European Alps remains much less understood. The Oligocene to lower
127 Miocene molasse deposited in the Subalpine massifs contains detritus derived from the internal
128 zones of the Alps (*Bocquet*, 1966; *Meckel*, 1997; *Jourdan et al.*, 2012), attesting to the existence of a
129 radial drainage network extending into the internal zones. At that time, high topography appears to
130 have been restricted to the internal zones (*Fauquette et al.*, 2015). This pattern strongly contrasts
131 with the current situation where the external zones, in particular the ECM, host the highest

132 topography with major peaks (e.g., Mont Blanc: 4808 m; Barre des Ecrins: 4102 m), and mark a
133 secondary drainage divide that is only crossed by a few large rivers (e.g., the Isère River northeast of
134 the Belledonne massif).

135 **Figure 1:** Location and setting of the studied Obiou caves. **A:** Satellite image (source:
136 <https://www.geoportail.gouv.fr/>) showing the location of the study area in the Western European Alps (black
137 square indicates extent of panel B). The major axial valleys of the Western Alps are highlighted with blue lines.
138 **B:** Regional geological map with the main tectonic units discussed in the text (*ECM*: external crystalline massif;
139 *ICM*: internal crystalline massif; *aut.*: autochthonous) and the major Alpine thrusts (thick black lines; *DT*: Digne
140 thrust; *PFT*: Penninic frontal thrust; *SAT*: Sub-Alpine thrust). The names of the main massifs are indicated in
141 bold white font. The location of the Obiou caves is indicated with a yellow star, and other sand samples used in
142 this study are shown by colored circles (yellow: modern rivers – this study; orange: modern rivers – published
143 data; pink: Dévoluy molasse) with the corresponding river catchments outlined by white dashed lines. **C:** North
144 face of the Obiou peak with the entrances of the three studied caves (yellow stars).

145 The Subalpine Dévoluy massif is located southwest of the Ecrins-Pelvoux massif (Fig. 1) and is
146 characterized by thick Upper Cretaceous limestones with a high karstification potential. We studied
147 three fossil caves (Jalabres, Petit Odieux, and Calvaire) located on the northeast flank of the Obiou
148 peak between 2250 and 2380 m elevation, ~1600 m above the modern Drac River (Fig. 1). These
149 caves are disconnected from the present-day drainage network and were not connected to other
150 massifs during Quaternary glaciations (*Monjuvent, 1978*). This high-mountain paleo-karst system
151 preserves alluvial sediments that include abundant crystalline cobbles, indicating non-local sediment
152 sources and fluvial transport (*Jagercikova et al., 2021*).

153 **Figure 2:** Morphology of the Jalabres cave (see location in Fig. 1B, C). **A:** Sub-horizontal paleo-meander incising
154 into limestone units. **B:** 60-m high dome with massive speleothem attesting of past warm conditions. **C-D-E:**
155 Alluvial deposits with sandy material and crystalline cobbles preserved as an infilled sediment pocket within the
156 paleo-meander (panel A).

157 The Jalabres cave, situated at 2380 m elevation, is a 200-m long horizontal cave preserving a
158 30-m high paleo-meander (Fig. 2A). The canyon shape and the large volume of the main karstic
159 gallery indicate that karstification occurred in the vadose zone with high water discharge (*Audra and
160 Palmer, 2011*). The cave ends abruptly on a fault, which borders a voluminous 60-m high collapsed
161 dome. A >10-m high speleothem, located at the inner extremity of the cave (Fig. 2B), testifies of past
162 warmer conditions, as the modern mean annual temperature of 1°C in the cave prevents such
163 massive calcite precipitation. Crystalline cobbles set in a sand and clay matrix have been found *in situ*
164 at different sites that are elevated above the cave floor (Fig. 2C). They indicate possible complete
165 sediment filling of the cave, which may have protected it from collapse after disconnection from the

166 local base level (i.e., the Drac River, Fig. 1). The alluvial material analyzed in this study was collected
167 from two sites. The first consists of collapsed conglomerates including crystalline pebbles and
168 cobbles interbedded with mica-rich sands and clays; the second site is a pocket infilled with sand,
169 pebbles, and cobbles (Fig. 2C-D-E). Locally derived components such as chert or limestone are rare in
170 these outcrops, implying the absence of mixing between these *in-situ* alluvial sediments and
171 autochthonous material.

172 The Petit Odieux cave, located at 2250 m elevation, has developed horizontally along a
173 stratification level in the limestone units. Its total volume is much smaller than that of the Jalabres
174 cave (the cave is roughly 1 m high at the outcrop of allochthonous sediment), and its morphology is
175 more anastomosing. The floor of the Petit Odieux cave is covered with conglomerate, including
176 products of cave ceiling collapse, allochthonous elements (sand and pebbles), and reworked salmon-
177 colored calcified clays. A section of laminated clays was found *in-situ* in a small infill and was
178 collected for pollen analysis (*Jagercikova et al., 2022*). Because of the sediment filling and posterior
179 cave collapses, the original section of the gallery is unknown. Finally, the Calvaire cave, also located
180 at 2250 m elevation, displays even more degraded outcrops and sediment infills, including some
181 angular blocks of sandstone that are composed of reworked allochthonous sand in a calcitic cement.
182 The collected sediment infill is located near the cave entrance.

183 3. Methods

184 Our analysis focused on determining the depositional age of the cave sediments, identifying
185 their provenance, and using them to infer paleo-exhumation rates of the source area(s). Cosmogenic
186 ^{21}Ne - ^{10}Be - ^{26}Al burial dating (e.g., *Granger and Muzikar, 2001; Niedermann, 2002*) was used to
187 constrain the timing of deposition. For provenance analysis, we used pebble counting combined with
188 sand petrography and heavy-mineral analysis (e.g., *Garzanti and Andò, 2019*). Detrital AFT and AHe
189 thermochronology of cave sediments was used to infer the exhumation history of the source area
190 (e.g., *Glotzbach et al., 2011b*).

191 3.1 TCN burial dating

192 To ensure reproducibility, six samples were collected in the Jalabres cave for TCN burial
193 dating, including two individual cobbles (JAL20-00, JAL20-01), two pebble mixtures (JAL20-06, JAL20-
194 09), and two sand samples (JAL20-07, JAL20-08). The first three samples were collected from the infill
195 shown in Fig. 2C-E, whereas the three last samples derive from a site closer to the entrance of the
196 Jalabres cave (site 1 of *Jagercikova et al., 2021*). We included two indurated samples from the other

197 caves: a calcite-cemented sandstone rich in crystalline rock lithic fragments collected near the
198 entrance of the Calvaire cave (CAL20-01), and crystalline cobbles from conglomerates in the Petit
199 Odieux cave (YEUX20-01).

200 Samples were prepared at the ISTERre GTC platform (University Grenoble Alpes, France).
201 After preliminary crushing of one cobble and pebbles, all samples were sieved to extract the 100-800
202 μm fraction. Quartz was separated using a Frantz[®] isodynamic magnetic separator and purified
203 through chemical dissolution in successive baths of H_2SiF_6 and HCl. Purified quartz was etched three
204 times in HF to remove possible contaminations of atmospheric ^{10}Be .

205 For ^{10}Be and ^{26}Al analyses, quartz fractions were spiked with 500 μl of a 998.0 ± 3.5 ppm Be
206 carrier (Scharlau Batch 16107901) for ^{10}Be concentration determination and dissolved with
207 concentrated HF solution; no Al spike was added during chemical preparation. The natural ^{27}Al
208 concentrations in the samples were measured on a Varian 720 ES ICP-AES at the ISTERre Mineralogy-
209 Geochemistry platform using the 396.152 nm emission line. ^{10}Be and ^{26}Al extraction followed the
210 conventional procedure of *Merchel and Herpers (1999)*. $^{10}\text{Be}/^9\text{Be}$ and $^{26}\text{Al}/^{27}\text{Al}$ ratios were measured
211 at the French National AMS facility ASTER (CEREGE, Aix-en-Provence, France; *Arnold et al., 2010*)
212 using for calibration the in-house Be standard ($^{10}\text{Be}/^9\text{Be}$ ratio of $1.19 \pm 0.01 \times 10^{-11}$; *Braucher et al.,*
213 *2015*) and the ASTER in-house standard SM-AL-11 ($^{26}\text{Al}/^{27}\text{Al}$ ratio of $7.40 \pm 0.06 \times 10^{-12}$; *Merchel &*
214 *Bremser, 2004*). Correction for a full-process blank ratio of $^{10}\text{Be}/^9\text{Be} = 7.62 \pm 0.78 \times 10^{-15}$ was applied.
215 For $^{26}\text{Al}/^{27}\text{Al}$, we measured an analytical blank ratio of $1.05 \pm 0.74 \times 10^{-15}$.

216 ^{21}Ne measurements were performed on 0.7-1.1 g of HF-etched quartz aliquots of samples
217 JAL20-01, JAL20-06, JAL20-07, and JAL20-08, at the GFZ Potsdam (Germany), following methods
218 described in *Niedermann (2002)* and *Hetzel et al. (2002)*. The abundance of cosmogenic ^{21}Ne was
219 determined by stepwise heating at 400, 600, 800, and 1200°C, and computed assuming a two-
220 component mixture of cosmogenic and trapped Ne in the 400-800°C steps, with trapped Ne having
221 either atmospheric composition or the composition determined by in-vacuo crushing of aliquots of
222 the quartz samples (Table S2). This method allows to detect nucleogenic contributions or
223 compositions dominated by isotopic mass fractionation (*Niedermann, 2002*). Aliquots of the quartz
224 standard CREU-1 that were measured during the sample batch yielded a ^{21}Ne excess of $340 \pm$
225 14×10^6 at g^{-1} and $345.8 \pm 7.4 \times 10^6$ at g^{-1} , respectively, consistent with the reference value
226 $348 \pm 10 \times 10^6$ at g^{-1} (*Vermeesch et al., 2015*).

227 The calculation of burial ages relies on an assumed production rate, which is strongly
228 dependent on the paleo-elevation, especially for slowly-eroding settings (e.g., *Granger and Muzikar,*
229 *2001; Blard et al., 2019*). Production rates were estimated with the scaling model of *Lal (1991)* and

230 the global ^{10}Be sea-level-high-latitude production rate of 4.1 ± 0.2 at $\text{g}^{-1} \text{y}^{-1}$ (Martin *et al.*, 2017). The
231 effect of latitude change is negligible at this timescale compared to other uncertainties, particularly
232 for the estimated production rate. We considered only neutron spallation, which produces ^{21}Ne and
233 ^{10}Be with a ratio of 4.2 ± 0.2 (Kober *et al.* 2011).

234 3.2. Petrography

235 The petrography of sediment samples was systematically compared with sediments of
236 nearby rivers that drain regions of known lithologies (Fig. 1B). The lithology of n randomly picked
237 pebbles (diameter 1 to 10 cm) was analyzed for the Jalabres cave (JAL20-04, $n=130$) and two modern
238 sands of rivers draining the southwest part of the Pelvoux massif: the Drac River ($n=113$) and the
239 Bonne River ($n=93$; sampling locations indicated in Fig. 1B).

240 Petrographic analyses were carried out on a sandstone sample from the Calvaire cave
241 (CAL20-01), a conglomerate matrix and a sand sample from the Petit Odioux (YEUX20-05, YEUX20-13)
242 and Jalabres (JAL20-04) caves. Sands were classified according to the proportions among the three
243 main framework components (Q: quartz; F: feldspar; L: lithic fragments), considered where
244 exceeding 10% QFL and listed in order of abundance (e.g., in a feldspatho-quartzo-lithic sand
245 $L>Q>F>10\%$ QFL; Garzanti, 2019). Rock fragments were classified by protolith composition and
246 metamorphic rank.

247 From a split aliquot of the 15-500 μm size window obtained by wet sieving, heavy minerals
248 were separated by centrifuging in Na-polytungstate (density 2.90 g/cm^3) and recovered by partial
249 freezing with liquid nitrogen (Andò, 2020). For each sample, ≥ 200 transparent heavy minerals (tHM)
250 were point-counted at an appropriate regular spacing (200 or 300 μm ; Garzanti and Andò, 2019).
251 Opaque minerals, rock fragments, iron oxides, soil clasts, phyllosilicates and carbonates were not
252 considered as integral parts of the tHM suite. The same analysis was performed on two samples
253 representative of the two main local formations making up the Dévoluy Molasse: the uppermost
254 Eocene-lowermost Oligocene Souloise and lower Oligocene Saint-Disdier sandstones (Meckel, 1997).
255 New data from modern sands from the Bonne and Drac rivers and literature data for rivers draining
256 the internal Alps (Garzanti *et al.*, 2010) were used for comparison. All sample sites are indicated on
257 Figure 1B.

258 3.3. AFT and AHe thermochronology

259 In addition to samples JAL20-01, JAL20-06, JAL20-07, JAL20-08, CAL20-01, and YEUX20-01,
260 which were also analyzed for TCN burial dating, we included a pebble mix from the Jalabres cave

261 (JAL20-09) for thermochronological analysis. Apatite grains were separated from the 100-200 μm
262 fraction using standard heavy-liquid and magnetic separation techniques. Apatite aliquots were
263 hand-picked and mounted in epoxy resin, polished, and etched with 5.5 M HNO_3 for 20 s at 21°C.
264 Apatite mounts were covered with muscovite sheets as external detectors and irradiated together
265 with dosimeter glass IRMM 540R at the FRM II Research Reactor in Munich (Germany) along with
266 Durango and Fish Canyon Tuff age standards. After irradiation, mica detectors were etched in 48% HF
267 at 21°C for 18 min. Fission tracks were counted at the ISTERre GTC platform by M. Balvay using the ζ -
268 calibration approach (Hurford and Green, 1983). Track lengths were measured whenever possible, as
269 well as etch-pit widths (D_{par} values) that can be used as kinetic indicators. Apatite grains for AHe
270 analysis were hand-picked from Jalabres cave samples and measured at the ISTERre GTC platform.
271 AHe analysis was performed at the University of Potsdam (He degassing and measurement) and GFZ
272 Potsdam (U, Th, and Sm measurements) following the methodology described in Zhou et al. (2017).

273 4. Results

274 4.1. TCN burial dating

275 Concentrations of cosmogenic ^{10}Be and ^{21}Ne in quartz are reported for all cave sediments in
276 Table 1. Detailed information and full analytical data are available in the Supplementary material
277 (Tables S1-S2). All measured ^{26}Al concentrations were below the analytical blank, or close to (and
278 associated with uncertainties over 100%, Table S1), whereas ^{10}Be concentrations were all
279 significantly above the measured blank except for sample YEUX20-01. As a consequence, the
280 $^{26}\text{Al}/^{10}\text{Be}$ system only provided minimum estimates for sample burial ages (Table S1), that we did not
281 further consider in the discussion. We noticed a strong granulometric effect among JAL samples
282 (Table 1): sand samples (JAL20-07 and -08) have higher ^{10}Be concentrations than pebbles (JAL20-06),
283 and pebbles have a higher ^{10}Be concentration than the cobble (JAL20-01). These differences may be
284 due to different erosion processes before burial, as cobbles and pebbles may be derived from
285 landslides and thus brought abruptly to the surface (e.g., Mariotti et al., 2019). The cosmogenic ^{21}Ne
286 abundance was computed both as the excess relative to the fluid-inclusion composition (as
287 determined by crushing extractions; Table S2) and relative to air for samples JAL20-06, JAL20-07 and
288 JAL20-08; those samples plot along the spallation line in a Ne three-isotope diagram (Fig. 3A), and
289 they provide similar concentrations of cosmogenic ^{21}Ne . Sample JAL20-01 shows very little ^{21}Ne
290 excess and its data can essentially be explained by mass fractionation; thus, the nominal cosmogenic
291 ^{21}Ne concentration reported in Table 1 should be taken as a maximum estimate.

292 $^{10}\text{Be}/^{21}\text{Ne}$ burial plots (Fig. 3B) provide a convenient way to represent this relationship (e.g.,
293 Ma et al., 2018; Blard et al., 2019). To assess the burial-age dependence on production rates,

294 isochrons and iso-erosion lines are plotted for two end-member scenarios: a scaling factor of 1
 295 corresponding to a sea-level paleo-production rate (pink lines), and of 7 considering a paleo-
 296 elevation of approximately 2.5 km (black lines). Between those two end-member scenarios, paleo-
 297 erosion rates are directly affected by the 7-fold difference in paleo-production rates, but the burial-
 298 age estimates are robust, with <1 Ma variation between the two considered scenarios. Considering
 299 all analytical uncertainties, the different estimates of cosmogenic ^{21}Ne concentration, and the paleo-
 300 elevation effect on production rates, our best-estimate $^{10}\text{Be}/^{21}\text{Ne}$ burial ages range from 10.5 to 13
 301 Ma. This is younger than, but overlaps within 1σ error, a previous estimate based on a single sample
 302 from the Jalabres cave (15.6 ± 3.8 Ma; *Jagercikova et al., 2021*). The difference is ascribed to a much
 303 higher ^{21}Ne concentration measured in the previous study, in which only a single-step extraction was
 304 performed at 1500°C , potentially linked to some contribution of nucleogenic neon trapped in fluid
 305 inclusions.

Sample	Grain size (cm)	Quartz mass for ^{10}Be extraction (g)	^{10}Be ($\times 10^3$ at/g)	^{21}Ne excess relative to air ($\times 10^6$ at/g)	^{21}Ne excess relative to trapped component ($\times 10^6$ at/g)
JAL20-01	13	74.6	1.3 ± 0.6	$1.5^{+0.7}_{-0.4}$	Not measured
JAL20-06	2 – 10	30.0	4.2 ± 1.7	16.4 ± 0.4	$7.4^{+0.6}_{-0.5}$
JAL20-07	0.01 – 0.08	49.6	8.4 ± 1.1	22.0 ± 0.6	9.0 ± 0.6
JAL20-08	0.01 – 0.08	14.7	6.2 ± 1.6	22.7 ± 0.6	8.3 ± 0.6
CAL20-01	0.01 – 0.08	24.6	7.4 ± 1.6	Not measured	Not measured

306 **Table 1:** Measured cosmogenic ^{10}Be and ^{21}Ne concentrations (with 1σ uncertainties) in quartz from samples
 307 from the Jalabres cave. See Tables S1 and S2 (Supplementary material) for full analytical details.

308 **Figure 3:** ^{10}Be and ^{21}Ne cosmogenic nuclide measurements and estimated burial ages for JAL samples. **A:** Ne
 309 isotopic ratios (with 2σ errors). A pure mixture of trapped atmospheric and cosmogenic Ne should lie on the
 310 spallation line. Shaded areas correspond to a mix of a trapped component, as determined by crushing
 311 extractions, and cosmogenic neon. The neon released during the 1200°C heating step was not considered as its
 312 composition suggests a nucleogenic origin (see inset for details). Data from sample JAL20-01 plot close to the
 313 mass fractionation line. **B:** Cosmogenic $^{10}\text{Be}/^{21}\text{Ne}$ burial plot with 1σ error ellipses. Cosmogenic ^{21}Ne is taken as
 314 the excess relative to the result of the crushing extraction (blue) and to air (green). A single previous analysis
 315 (JAL1) from *Jagercikova et al. (2021)*, for which the gas was extracted using a single degassing step at 1500°C
 316 and the ^{21}Ne excess was computed relative to air, is shown in red. Two end-member scenarios have been
 317 considered for production rates, affecting burial ages (curves) and paleo-erosion rates (lines): a scaling factor of
 318 1 corresponding to a sea-level paleo-production rate (pink), and of 7 considering a 2.5 km paleo-elevation
 319 (black).

320 4.2. Petrography and provenance

321 Among the 130 pebbles counted from the sediment infill of the Jalabres cave (Fig. 4), 49%
 322 are granitoids, followed by gneiss (19%), vein quartz, amphibolite, volcanic rocks, limestone,

323 micaschist, and epidote-rich metamorphic rocks. In contrast, the modern Bonne River carries only
324 29% granitoid pebbles but 41% metamorphic rocks, including numerous meta-volcanic rocks from
325 the Taillefer/Pelvoux cortex zone. Liassic dark marls and limestones are also more abundant (16 %)
326 than in the cave sediments. Modern Drac River deposits are even richer in sedimentary pebbles, with
327 39% limestone and 10% flysch (Champsaur sandstone, not found in the Jalabres cave but observed in
328 marmot burrows near the entrance; *Jagercikova et al., 2021*); granitoids represent only 19% of the
329 counted pebbles. Therefore, cave sediments appear representative of the Pelvoux basement, as
330 currently sampled by the Bonne and Drac rivers, although granite is over-represented with respect to
331 gneiss, and crystalline rocks are over-represented with respect to sedimentary rocks, relative to
332 modern river deposits. This difference may represent a bias due to transport length, as softer
333 lithologies are preferentially removed downstream (e.g., *Dingle et al., 2017*). In contrast to the
334 Dévoluy molasse (*Meckel, 1997*), no component derived from the internal Alps (such as serpentinite
335 or radiolarite) was encountered in the Jalabres cave pebbles.

336 **Figure 4:** Comparison of pebble lithologies between the sedimentary infill of the Jalabres cave and modern
337 sediments from rivers draining the Ecrins-Pelvoux massif. Number of counted pebbles - Jalabres: 130; Drac:
338 113; Bonne: 90.

339 The sand fraction of the Jalabres cave sample is litho-feldspatho-quartzose with meta-
340 granitoid (diorite, tonalite) and high-grade metamorphic grains including garnet amphibolite.
341 Volcanic lathwork and metapsammite grains also occur. The tHM suite, dominated by green-brown
342 hornblende with garnet and oxy-hornblende (Table S3; Fig. S1) is notably richer (tHM concentration
343 3.3% of bulk sample) than in modern sands of the Bonne (tHM 0.4%) and Drac (tHM 0.6%) rivers.
344 Modern river sands have similar tHM suites dominated by blue-green hornblende with epidote;
345 green augite and garnet are common in the Drac sand, whereas apatite is common in the Bonne
346 sand.

347 The Petit Odieux cave sediments consist of a matrix-supported conglomerate with granite,
348 gneiss, micaschist, dark sandstone, and volcanic clasts. Pebbles and clays, potentially reworked in the
349 cave, are mixed with coarse sand with a petrographic composition similar to both adjacent sand
350 deposits within the Petit Odieux cave and the sandstone sample from the Calvaire cave. This
351 feldspatho-litho-quartzose sand contains low- to medium/high-rank metapelite, metapsammite, and
352 micritic or sparitic carbonate rock fragments. Plagioclase and orthoclase occur in subequal
353 proportions; biotite dominates over muscovite. Samples from the Petit Odieux and Calvaire caves
354 have poor tHM suites (tHMC 0.6 and 0.1%, respectively) with apatite, epidote, garnet, and zircon;
355 most of the dense fraction consists of Ti-oxides and Fe-hydroxides, with rare hematite.

356 The Dévoluy molasse samples have a different mineralogy. The very fine-grained and
357 extensively calcified litho-quartzose Souloise sandstone contains plagioclase, slate lithics, muscovite,
358 and glauconite. The coarse-grained feldspatho-quartzo-lithic Saint Disdier sandstone contains
359 abundant serpentine schist and cellular serpentinite grains diagnostic of provenance from the
360 Internal Alps ophiolites (*Garzanti et al., 2010*), together with plagioclase, microlithic volcanic,
361 chloritoschist, slate, and metasandstone rock fragments. The tHM suite of both molasse samples is
362 extremely poor (tHMC 0.04-0.05%), with durable apatite, garnet, rutile, zircon, tourmaline, and
363 minor Cr-spinel, but no unstable ferromagnesian minerals. Garnet commonly shows corrosion facets
364 (*Andò et al., 2012*).

365 A principal-component analysis (PCA; Fig. 5) was performed on the analyzed samples to highlight the
366 different controls on the tHM suites. Dimension 1, explaining around 43% of the compositional
367 variance, discriminates based on the different chemical durability of detrital minerals (*Garzanti et al.,*
368 *2018*) and thus chiefly reflects diagenesis of foreland-basin sediments. Dimension 2, explaining 18%
369 of the variance, discriminates between greenschist/blueschist minerals derived from the Alpine
370 Internal zone and minerals declining with deep burial such as apatite and garnet (*Garzanti et al.,*
371 *2018*) preserved only in the External zone, where Alpine metamorphic overprint was minor, and thus
372 mainly reflects sediment provenance. To the latest are also correlated amphibolite-facies minerals
373 derived from higher-temperature Variscan metamorphic rocks of the ECM.

374 In the PCA plot (Fig. 5), cave sands plot in the field of relatively unweathered sediments, near
375 modern sediments in streams draining the ECM. The Jalabres cave sand plots closer to river sands
376 derived from the Ecrins-Pelvoux massif than to the other caves sediments, which plot in a more
377 intermediate position between the Ecrins-Pelvoux massif and the Molasse deposits. This difference
378 may point either to a larger contribution from the sedimentary cover of the external zone, or to
379 more intense weathering conditions in the Calvaire and Petit Odieux caves with respect to the
380 Jalabres cave.

381 **Figure 5:** Principal component analysis (PCA) of transparent-heavy-mineral suites in cave sediments and
382 modern river sands (including data from *Garzanti et al., 2010*). Ap = apatite + monazite, bl-g H = blue-green
383 hornblende, g-br H = green-brown hornblende, &A = other amphiboles; Ep = epidote group, Grt = garnet, HgM
384 = higher-grade metamorphic minerals (andalusite + kyanite + sillimanite + staurolite), LgM = lower-grade
385 metamorphic minerals (prehnite + pumpellyite + chloritoid + lawsonite + carpholite), Px = pyroxenes, Ti& =
386 anatase and titanite, ZTR = zircon + tourmaline + rutile (durable minerals).

387 4.3. Detrital thermochronology

388 Detailed AFT data (Table S4) and radial plots for each sample (generated using *RadialPlotter*;
389 *Vermeesch, 2009*) (Fig. S2) are available in the Supplementary material. Between 20-100 grains were
390 analyzed per sample and all samples pass the χ^2 test. All samples except JAL20-01 have central AFT
391 ages between $15.9^{+3.5}_{-2.9}$ and $19.5^{+2.1}_{-1.9}$ Ma (Fig. 6A, Table S4); age distributions are unimodal and show
392 very little dispersion, even for sand samples (JAL20-07 and -08). The cobble sample JAL20-01 stands
393 out from the other JAL samples as it records a central AFT age of 32.3 ± 2.5 Ma. We could measure
394 97 confined tracks on JAL samples (JAL20-06, -07, -08 and -09); combined, they show a mean track
395 length of 13.1 ± 0.2 μm with a standard deviation of 2.2 μm (Fig. S3). The measured track lengths are
396 not correlated with D_{par} values, suggesting limited compositional control on track annealing in these
397 samples. The D_{par} value of sample JAL20-01 is slightly higher than for other JAL samples (1.90 ± 0.16
398 μm versus 1.55 ± 0.22 μm).

399 Single-grain AHe data are plotted in Figure 6B, with detailed analytical results given in the
400 Supplementary material (Table S5). The median AHe age for Jalabres cave samples is 14.4 Ma,
401 excepting sample JAL20-01, which yielded single-grain AHe ages between 20.5 and 30.6 Ma. No
402 correlation between single-grain AHe ages and effective uranium (eU) is observed in the dataset (Fig.
403 S4).

404
405 **Figure 6:** Detrital thermochronology results from the Obiou caves. **A:** Kernel density distributions of AFT single-
406 grain ages; vertical bars represent central AFT ages. **B:** Distributions of single-grain AHe ages for Jalabres cave
407 samples. See Supplementary material for full analytical data.

408 5. Data Interpretation and Discussion

409 5.1. Age and depositional environment

410 We obtained Serravalian – early Tortonian (13-10 Ma) ^{21}Ne - ^{10}Be burial ages for alluvial
411 sediments in the Jalabres cave (Fig. 3). A Miocene burial age also explains the absence (or very close
412 to the analytical blank) of measurable amounts of cosmogenic ^{26}Al in all the cave samples (Table S1),
413 which is in agreement with previous attempts by *Jagecikova et al. (2021)* and only allows for
414 estimates of minimum burial time for cave samples. These ages are consistent with the
415 thermochronological data from the same samples, which provide pre-depositional cooling ages from
416 around 18 Ma (AFT) to 14 Ma (AHe).

417 *Jagercikova et al. (2022)* presented palynological results from a clay sample extracted from
418 the Petit Odieux cave, which, by comparison with results from the Western Alpine foreland
419 (*Fauquette et al., 2015*), suggest a Serravalian to earliest Zanclean age (i.e., ~5-14 Ma). Although
420 broad, this age range is consistent with our more precise TCN burial age of 11.5 ± 1.5 Ma. The pollen
421 assemblage includes both subtropical species, suggesting a low-elevation setting in a warmer-than-
422 present mid to late Miocene climate, and boreal species indicating that elevations >1900 m must
423 have existed in the region. *Jagercikova et al. (2022)* also report the occurrence of a small number of
424 non-reworked lagoonal and coastal dinoflagellates, suggesting that the caves formed in a coastal
425 environment close to sea-level. Some cobbles show surface marks suggesting lithophagus
426 perforations, which would confirm deposition in a coastal environment. If confirmed, this
427 observation would be crucial for constraining the coastline position during mid to late Miocene
428 times. Marine deposits of Serravillan-Tortonian age have been found to the southwest of the
429 Vercors massif (*Kalifi et al., 2021*), but constraints farther east remain scarce. Assuming that the
430 caves were formed close to sea level, they must have been uplifted to their present elevation of 2300
431 m since 10-13 Ma. In addition, the Obiou summit is currently situated ~500 m above the caves,
432 providing an estimate for the minimum amount of topographic relief at the time of cave formation.

433 *5.2. Incision and drainage development*

434 The Jalabres cave is located ~1600 m above the modern Drac River (Fig. 1), providing
435 evidence for km-scale incision of the Drac river occurring after 13-10 Ma. The minimum incision rate
436 since abandonment of the caves is around 140 m/Myr; the actual rate of incision could have been
437 significantly higher if incision started later, during late-Miocene or Pliocene-Quaternary times. This
438 rate is at least 5 times lower than late Pleistocene – Holocene incision rates of the Drac and adjacent
439 Buëch rivers, as inferred from TCN dating of alluvial terraces (*Brocard et al., 2003*). A significant part
440 of this incision may thus have taken place since the onset of Quaternary glaciations, as reported
441 elsewhere in the Western European Alps (e.g., *Haeuselmann et al., 2007; Valla et al., 2011*).

442 The studied sediments from the Jalabres cave are clearly derived from the Ecrins-Pelvoux
443 massif (Figs. 4, 5). The deposits in the two other caves probably have a similar origin, but show more
444 recycling and/or weathering. This provenance signal requires the existence of a radial drainage
445 pattern at 10-13 Ma, which was subsequently replaced by the current drainage system, separating
446 the Dévoluy massif from the Ecrins-Pelvoux massif by the axial Drac valley (Fig. 1B). The composition
447 of the cave deposits also clearly differs from that of the Oligocene molasse deposits in the Dévoluy
448 massif, which contain detritus from the Alpine internal zones lacking in the cave deposits. The south-
449 flowing Durance River currently drains the internal zones of the Western Alps in this region (Fig. 1B).

450 Thus, these deposits provide evidence for two subsequent stages of drainage reorganization: (1) an
451 Oligocene radial drainage system that deposited the Dévoluy molasse extended to the internal
452 zones; the headwaters of this drainage system were cut off during the early Miocene by the
453 establishment of the axial upper Durance valley along the Penninic Frontal Thrust; and (2) the
454 resulting shorter Late- Miocene radial drainage network, with headwaters in the Ecrins-Pelvoux
455 massif, was subsequently reorganized into the current axial drainage with the establishment of the
456 Drac valley. The first drainage reorganization stage is coeval with reorganization of the upper Rhone
457 and Rhine rivers in the Swiss Alps (*Schlunegger et al., 1998*); these rivers are also in the same
458 structural position as the upper Durance, i.e., between the Aar ECM and the internal zones (Fig. 1).

459 *5.3. Thermal and exhumation history of the Ecrins-Pelvoux massif*

460 All samples except JAL20-01 provide similar AFT ages and track-length distributions with low
461 single-grain age dispersion (<10%), particularly for sand samples. This suggests that the cave samples
462 registered a uniform source-area exhumation history, consistent with data indicating provenance
463 from the Ecrins-Pelvoux massif. We employ inverse thermal-history modeling (using the QTQt code;
464 *Gallagher, 2012*) to infer time-temperature paths from the thermochronology data. We used all
465 available AHe and AFT data from the Jalabres cave (including track-length and D_{par} data) as input, and
466 included as the sole temperature constraint that the samples must have been at surface
467 temperatures when they were deposited in the cave between 13 and 10 Ma. Our model output (Fig.
468 7A) predicts slow cooling of $\sim 3^{\circ}\text{C}/\text{Myr}$ between 24 and 15-16 Ma, increasing to $>14^{\circ}\text{C}/\text{Myr}$
469 thereafter. Assuming a uniform geothermal gradient of $\sim 30^{\circ}\text{C}/\text{km}$, this thermal history is equivalent
470 to an exhumation rate increasing from ~ 0.1 km/Myr up to ~ 0.5 km/Myr after 15-16 Ma.

471

472 **Figure 7: A.** Thermal history inferred from the Jalabres cave sediments, using available thermochronology data
473 as input and TCN burial ages as an independent constraint on depositional age (red dashed box). Inferred
474 temperature history is shown as relative probability (colors) with “best-fit” histories using different metrics and
475 their 95% credible interval shown as black lines (see *Gallagher, 2012* for details). **B.** Comparison with regional
476 Time-temperature paths and constraints from the literature. The dashed lines represent two alternative
477 scenarios discussed in the text.

478 This inferred thermal history appears to be consistent with regional exhumation patterns
479 (Fig. 7B) as recorded by bedrock thermochronology in the Ecrins-Pelvoux and adjacent Grandes
480 Rousses and Belledonne massifs (*Seward et al., 1999; Tricart et al., 2007; van der Beek et al., 2010;*
481 *Girault et al., 2022*). These massifs record an onset or acceleration of exhumation during the early-
482 middle Miocene, between ~ 20 and 15 Ma (Fig. 7B). *Girault et al. (2022)* linked this acceleration to

483 the transition from distributed shortening to localized thrusting and uplift on crustal-scale ramps
484 beneath the ECM. A similar transition was recorded slightly earlier (22-20 Ma) in the Aar massif to
485 the northeast (*Herwegh et al., 2019*). Our new data provide tighter constraints on this Miocene
486 acceleration of exhumation rates for the Ecrins-Pelvoux massif.

487 The maximum burial temperature reached by the sedimentary cover of the Ecrins-Pelvoux
488 massif, before the onset of exhumation, is constrained by Raman spectroscopy of carbonaceous
489 material (RSCM; *Bellanger et al., 2015*), with values increasing from <200°C in the southwest to
490 ~350°C in the central part of the massif. These temperatures record heating of the massif and its
491 cover by tectonic burial below the advancing Penninic Frontal Thrust (Fig. 1B) in the early Oligocene
492 (*Simon-Labric et al., 2009*). They also require rapid Oligocene cooling of the central Pelvoux massif,
493 similar to what is observed in the Grandes Rousses (Fig. 7B, path 1). In contrast, detrital zircon
494 fission-track (ZFT) ages from Priabonian flysch in the southern Pelvoux are older than the
495 depositional ages, implying that they were not, or only partially, reset (*Seward et al., 1999; Bernet,*
496 *2013*). The ZFT data thus imply limited Oligocene reheating in the southern Pelvoux massif during
497 underthrusting beneath the Penninic Front. This argues for a lower-temperature scenario (Fig. 7B,
498 path 2), taking into account that the source rocks for the sediments deposited in the cave at 10-13
499 Ma must have been situated structurally above the currently exposed bedrock, and therefore
500 recorded cooler conditions. Sample JAL20-01 is the only sample to record Oligocene cooling, with an
501 AFT central age of 32 Ma. It could have derived from a now-eroded, partially-reset zone at high
502 elevations in the Pelvoux basement. Several single-grain AHe ages of ~29 Ma in sample YEUX20-01
503 may also indicate a contribution of early-exhumed bedrock, fitting with scenario no. 2 of Fig. 7B.

504 The widespread exposure of Pelvoux basement rocks in mid-Miocene times, as suggested by
505 the petrography of the cave sediments, requires the structure and/or exhumation history of the
506 Ecrins-Pelvoux massif to be reconsidered. *Dumont et al. (2012)* proposed a geometrical
507 reconstruction of the basement-cover interface in the Ecrins-Pelvoux massif, which implied that the
508 thickness of eroded basement does not exceed 3 km in the most incised valleys. However, these
509 valleys also record young (3-5 Ma) AFT ages, implying considerable exhumation since that time,
510 which appears inconsistent with both early basement exposure and a missing basement section of
511 only 3 km.

512 *5.4. A model for regional topographic and drainage-network evolution*

513 The Obiou cave sediments provide evidence for two successive drainage reorganizations: the
514 headwaters of a long radial drainage system, reaching into the internal zones, were beheaded by the
515 establishment of the axial upper Durance valley between the Oligocene and the mid-Miocene, and

516 the shorter radial drainage sourced in the Ecrins-Pelvoux massif was replaced by the axial Drac valley
517 during the late Miocene. The timing of the first reorganization is coeval with the acceleration of
518 exhumation in the Ecrins-Pelvoux massif, which has been linked to the onset of localized thrusting on
519 a crustal-scale ramp (*Girault et al., 2022*). This basement thrust would correspond to the Alpine
520 frontal thrust (Fig. 8) identified on the nearby CIFALP seismic profile as a major crustal discontinuity
521 (*Nouibat et al., 2022*). We suggest that thrusting over this ramp induced surface uplift of the Ecrins-
522 Pelvoux massif and split the former drainage network (Fig. 8). Such a tectonically driven process of
523 drainage deflection corresponds to a classical concept documented in various contexts, including the
524 Zagros Mountains and the Appalachians (*Oberlander and Morisawa, 1985*), the Siwaliks (*van der*
525 *Beek et al., 2002*), and the Aar massif in the external zone of the Swiss Alps (*Schlunegger et al., 1998*;
526 *Kühni and Pfiffner, 2001*).

527 A similar model applies to the modern drainage system that is deflected around the Subalpine
528 Chartreuse, Vercors and Dévoluy massifs (Fig. 1). Steep thrusts in these Subalpine massifs were active
529 during Miocene times and propagated outward (*Kalifi et al., 2022; Bilau et al., 2023*). Initial thrusting
530 in the eastern Dévoluy massif during the middle Miocene would have uplifted the Jalabres cave area
531 above sea level, created the inferred ~500 m of local relief, induced erosion of most of the molasse
532 cover deposits, and started speleogenesis. Continued thrusting and uplift of the Subalpine massifs
533 probably controlled the formation of the axial drainage of the “*Sillon Subalpin*” since the constrained
534 sediment burial in the Obiou cave system. The modern orientation of the Drac valley, flowing
535 northward for around 90 km and thus opposite to the shortest route to its final base level (i.e., the
536 Mediterranean Sea), and perpendicular to the topographic gradient of the orogen, cannot be
537 explained by fluvial or glacial processes alone. We thus suggest that the modern drainage must have
538 initiated simultaneously to deformation and uplift of the subalpine massifs, which occurred during
539 the mid- to late Miocene (*Bilau et al., 2023*). The “*Sillon Subalpin*” terminates at the middle Durance
540 valley to the south of the Dévoluy massif, and we suggest that both the lack of resistant lithologies
541 and the flat décollement layer under the Digne nappe explains the preservation of the radial
542 drainage network in the southern part of the Western European Alps. These new findings contribute
543 to resolving a long-standing controversy on the origin of these axial valleys of the French Western
544 Alps and the timing of their incision (*Blanchard, 1947; Debelmas, 1995*). Glacial carving of the “*Sillon*
545 *Subalpin*” occurred preferentially in soft Upper-Jurassic marls that were, according to *Debelmas*
546 (1995), exposed after a supposed peneplanation phase. However, such a peneplanation phase
547 appears inconsistent with our new data on the Miocene paleotopography. We show that significant
548 relief existed before axial-valley development and that the modern drainage formed during the
549 Miocene in response to tectonic activity.

550 6. Conclusions

551 The Obiou cave system provides a rare paleo-environmental and geodynamic long record
552 from an intra-orogenic setting, formed in a karstic environment during Middle to Late Miocene
553 times. New analyses of the preserved cave sediments, sourced from the nearby Ecrins-Pelvoux ECM,
554 provide new constraints on the Neogene structural, topographic, and drainage evolution of this part
555 of the Western Alps.

556 The alluvial sediments from the Jalabres cave yield TCN burial ages of 11.5 ± 1.5 Ma. The
557 existence of sediments derived from the Ecrins-Pelvoux massif in the Dévoluy massif has implications
558 for both the timing of drainage rearrangement and the long-term incision rate of the Drac valley,
559 which is estimated around 140 m/Myr. Detrital thermochronology of the cave sediments records
560 early exhumation of the Ecrins-Pelvoux massif with an unprecedented resolution. The increase in
561 exhumation rate from 0.1 to 0.5 km/Myr just after 15-16 Ma may coincide with the activation of a
562 crustal-scale thrust and can be linked to surface uplift leading to high topography in the Ecrins-
563 Pelvoux massif. Consequently, the regional drainage network was reorganized, with a major change
564 in sediment sources during the Miocene from the Internal to the External zones of the Western
565 French Alps. Subsequent thrusting in the Subalpine massifs during the Late Miocene led to a second
566 drainage rearrangement and the development of the modern axial drainage system.

567 **Figure 8:** Tectonic, topographic, and drainage development of the western Alps in the vicinity of the Ecrins-
568 Pelvoux and Dévoluy massifs. Cross-sections and maps show three stages of evolution, from top to bottom:
569 Oligocene (*Jourdan et al., 2012; Fauquette et al. 2015*), middle-late Miocene (as inferred from cave sediments),
570 and modern setting.

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578

579 Appendix A. Supplementary data

580 Supplementary data to this article can be found online at XXXXX.

581

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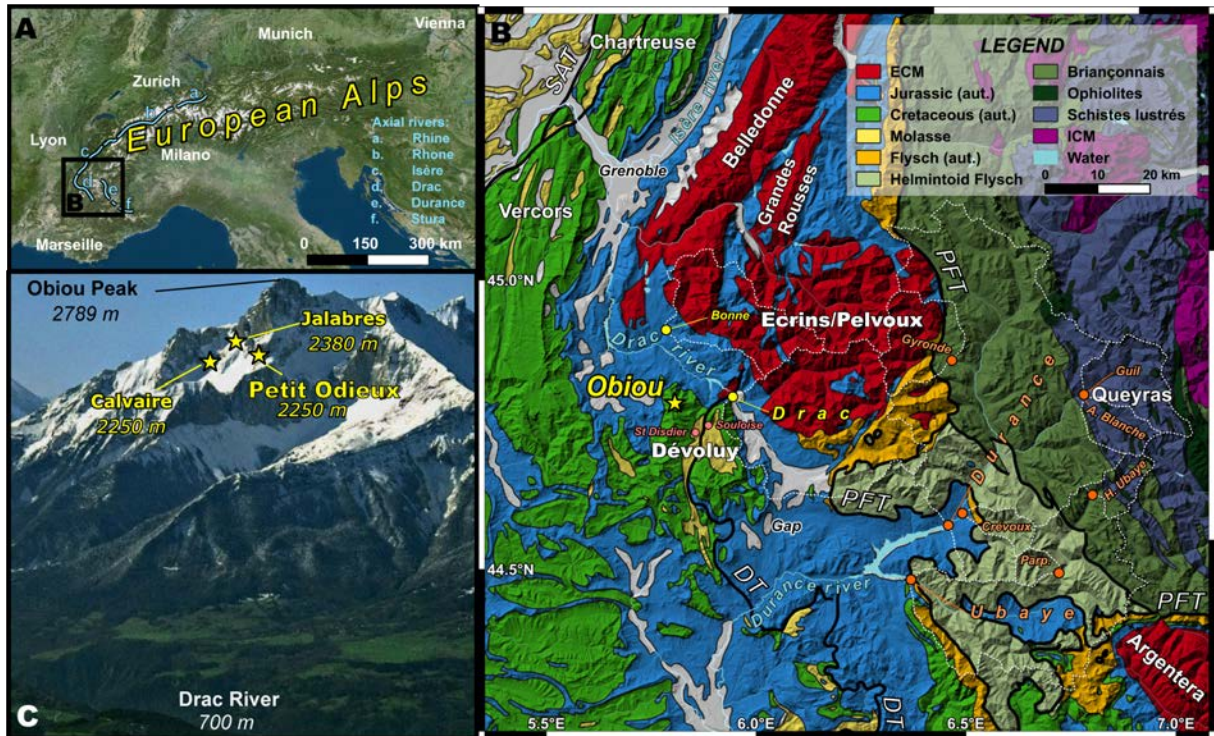
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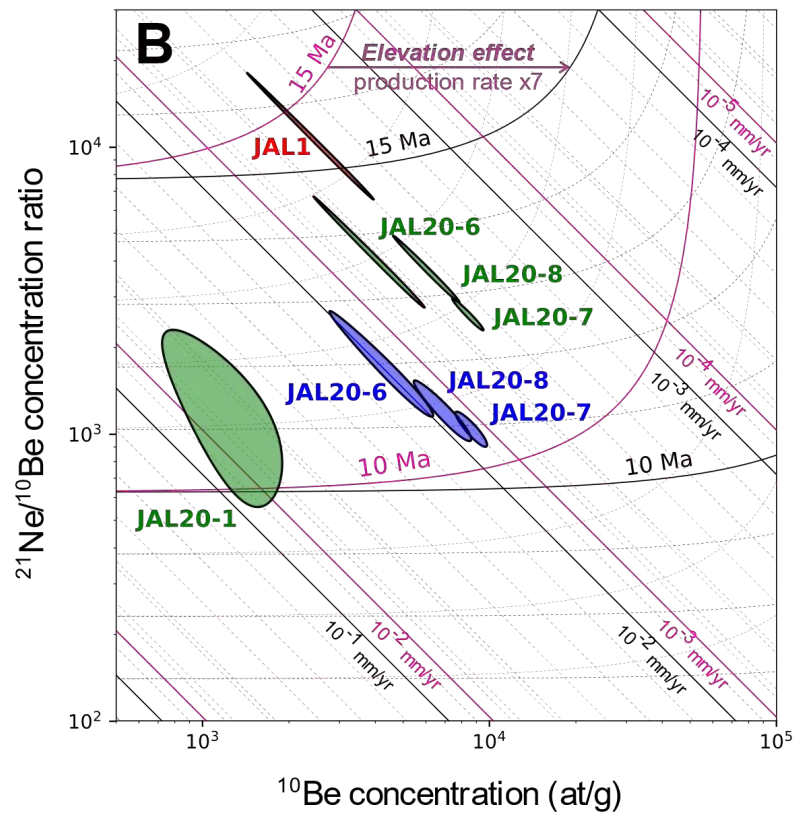
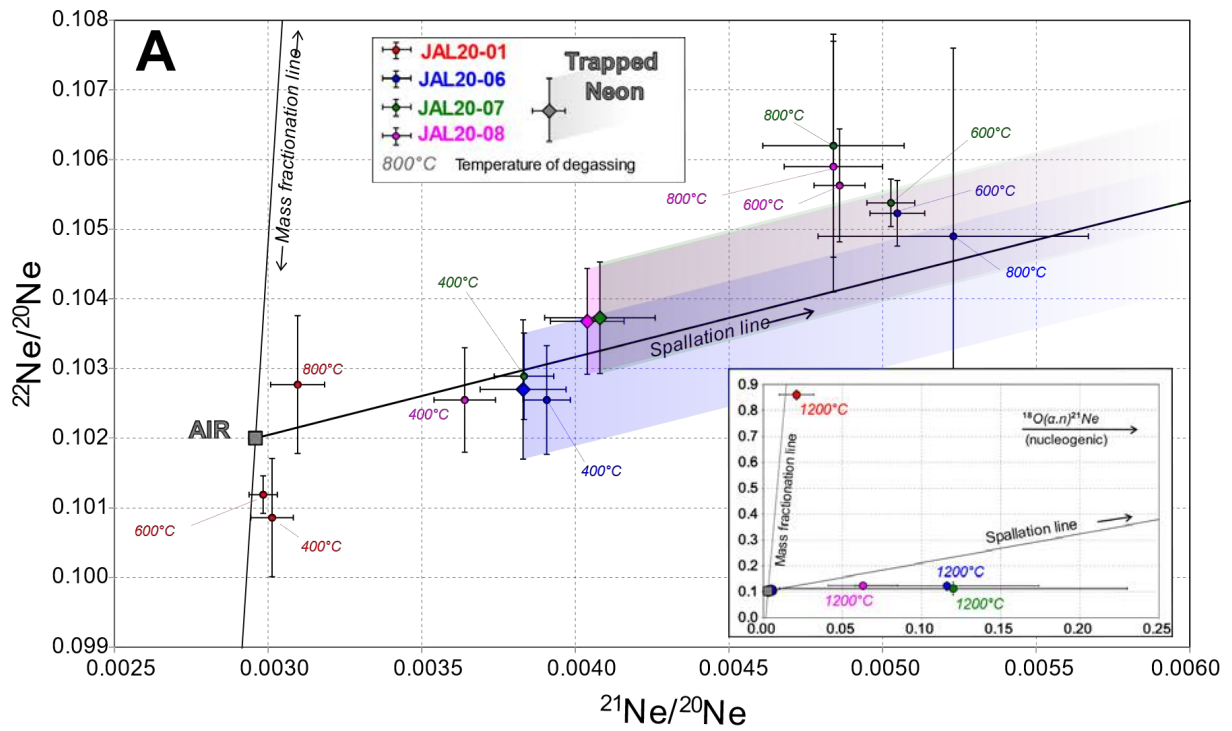


778
 779 **Figure 1:** Location and setting of the studied Obiou caves. **A:** Satellite image (source:
 780 <https://www.geoportail.gouv.fr/>) showing the location of the study area in the Western European
 781 Alps (black square indicates extent of panel B). The major axial valleys of the Western Alps are
 782 highlighted with blue lines. **B:** Regional geological map with the main tectonic units discussed in the
 783 text (*ECM*: external crystalline massif; *ICM*: internal crystalline massif; *aut.*: autochthonous) and the
 784 major Alpine thrusts (thick black lines; *DT*: Digne thrust; *PFT*: Penninic frontal thrust; *SAT*: Sub-Alpine
 785 thrust). The names of the main massifs are indicated in bold white font. The location of the Obiou
 786 caves is indicated with a yellow star, and other sand samples used in this study are shown by colored
 787 circles (yellow: modern rivers – this study; orange: modern rivers – published data; pink: Dévoluy
 788 molasse) with the corresponding river catchments outlined by white dashed lines. **C:** North face of
 789 the Obiou peak with the entrances of the three studied caves (yellow stars).



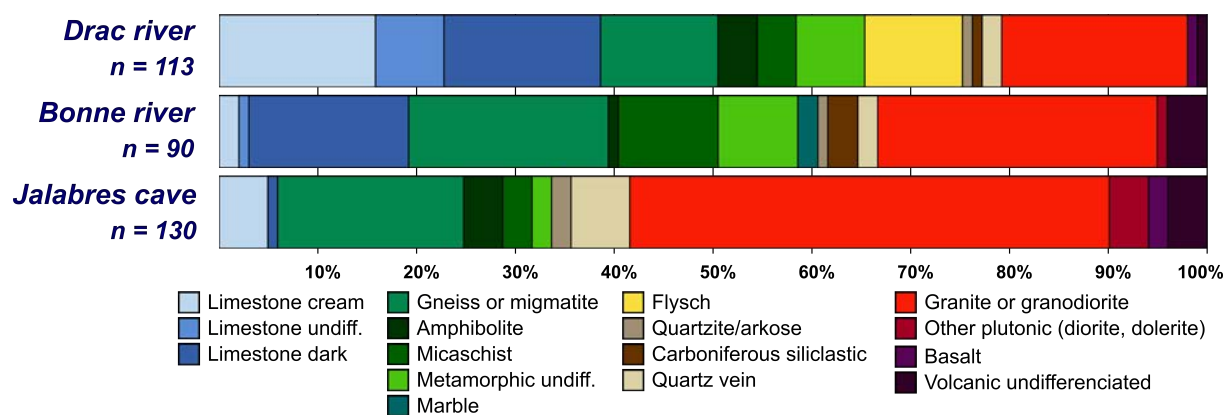
790
 791 **Figure 2:** Morphology of the Jalabres cave (see location in Fig.1B, C). **A:** Sub-horizontal
 792 paleo-meander incising into limestone units. **B:** 60-m high dome with massive speleothem

793 attesting of past warm conditions.C-D-E: Alluvial deposits with sandy material and crystalline
794 cobbles preserved as an infilled sediment pocket within the paleo-meander (panel A).
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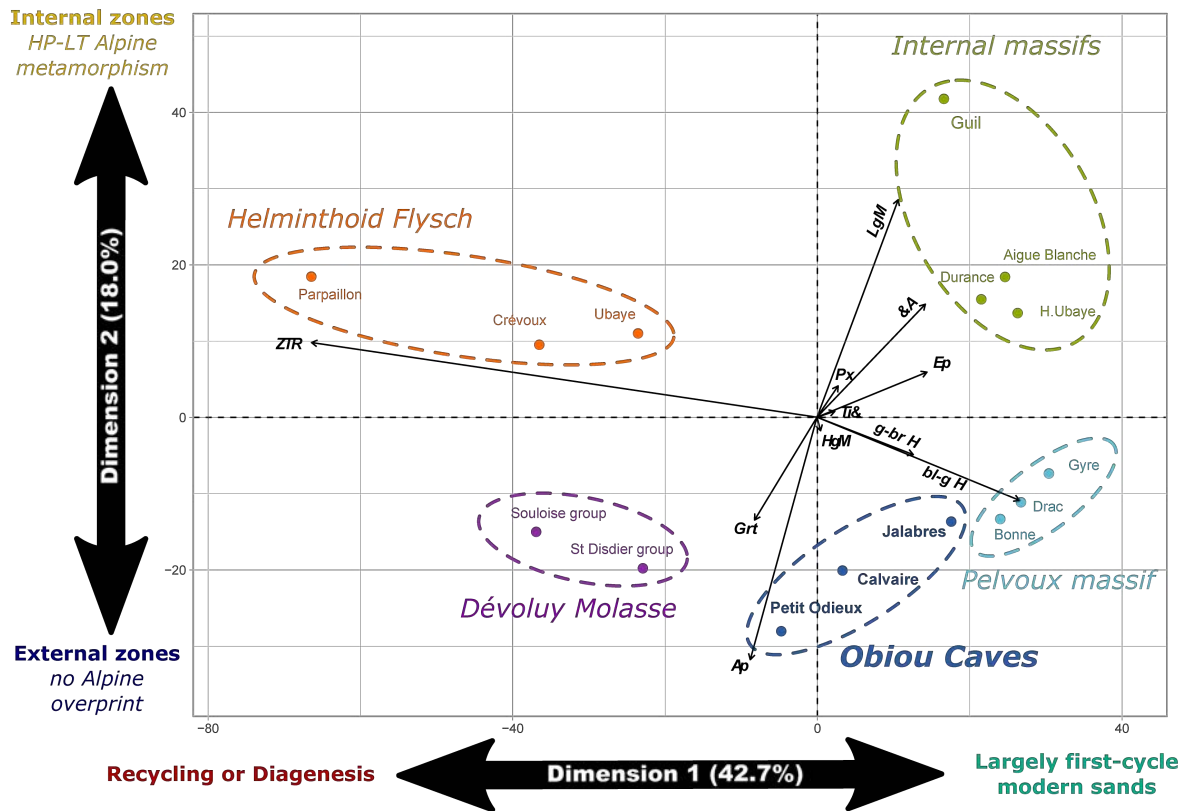


796
 797 **Figure 3:** ^{10}Be and ^{21}Ne cosmogenic nuclide measurements and estimated burial ages for JAL
 798 samples. A: Ne isotopic ratios (with 2σ errors). A pure mixture of trapped atmospheric and
 799 cosmogenic Ne should lie on the spallation line. Shaded areas correspond to a mix of a trapped

800 component, as determined by crushing extractions, and cosmogenic neon. The neon released during
 801 the 1200°C heating step was not considered as its composition suggests a nucleogenic origin (see
 802 inset for details). Data from sample JAL20-01 plot close to the mass fractionation line. B: Cosmogenic
 803 $^{10}\text{Be}/^{21}\text{Ne}$ burial plot with 1σ error ellipses. Cosmogenic ^{21}Ne is taken as the excess relative to the
 804 result of the crushing extraction (blue) and to air (green). A single previous analysis (JAL1) from
 805 Jagercikova et al. (2021), for which the gas was extracted using a single degassing step at 1500°C and
 806 the ^{21}Ne excess was computed relative to air, is shown in red. Two end-member scenarios have been
 807 considered for production rates, affecting burial ages (curves) and paleo-erosion rates (lines): a
 808 scaling factor of 1 corresponding to a sea-level paleo-production rate (pink), and of 7 considering a
 809 2.5 km paleo-elevation (black).



810
 811 **Figure 4:** Comparison of pebble lithologies between the sedimentary infill of the Jalabres cave and
 812 modern sediments from rivers draining the Ecrins-Pelvoux massif. Number of counted pebbles -
 813 Jalabres: 130; Drac: 113; Bonne: 90.



814

815 **Figure 5:** Principal component analysis (PCA) of transparent-heavy-mineral suites in cave sediments

816 and modern river sands (including data from Garzanti et al., 2010). Ap =apatite +monazite, bl-g H

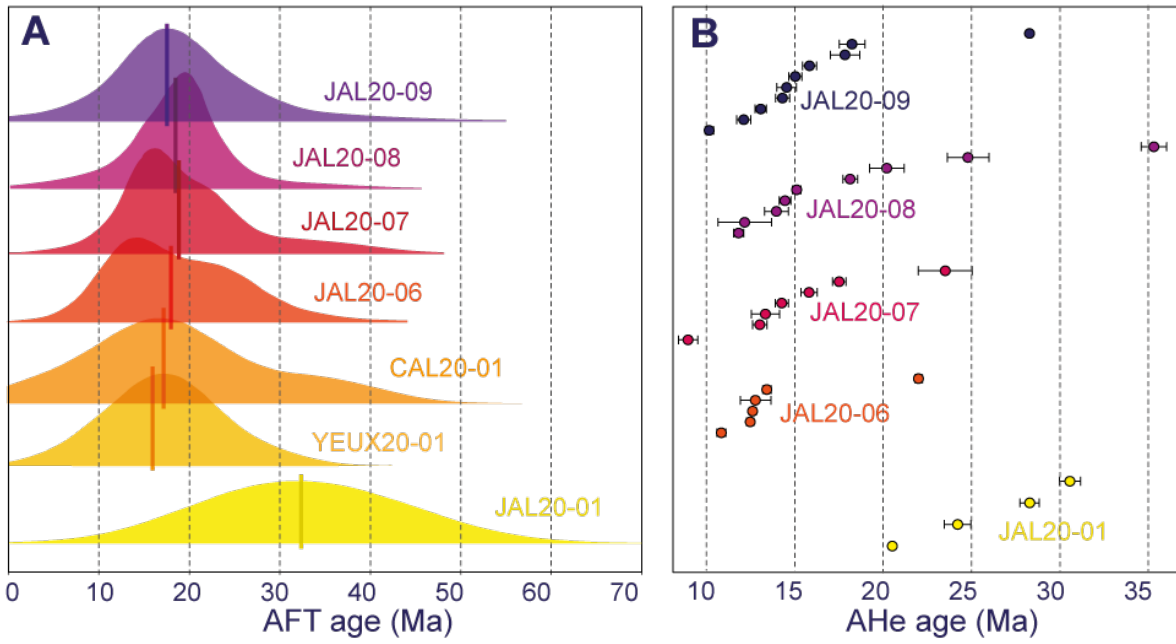
817 =blue-green hornblende, g-br H =green-brown hornblende, &A =other amphiboles; Ep =epidote

818 group, Grt =garnet, HgM =higher-grade metamorphic minerals (andalusite +kyanite +sillimanite

819 +staurolite), LgM =lower-grade metamorphic minerals (prehnite +pumpellyite +chloritoid +lawsonite

820 +carpholite), Px =pyroxenes, Ti& =anatase and titanite, ZTR =zircon +tourmaline +rutile (durable

821 minerals).



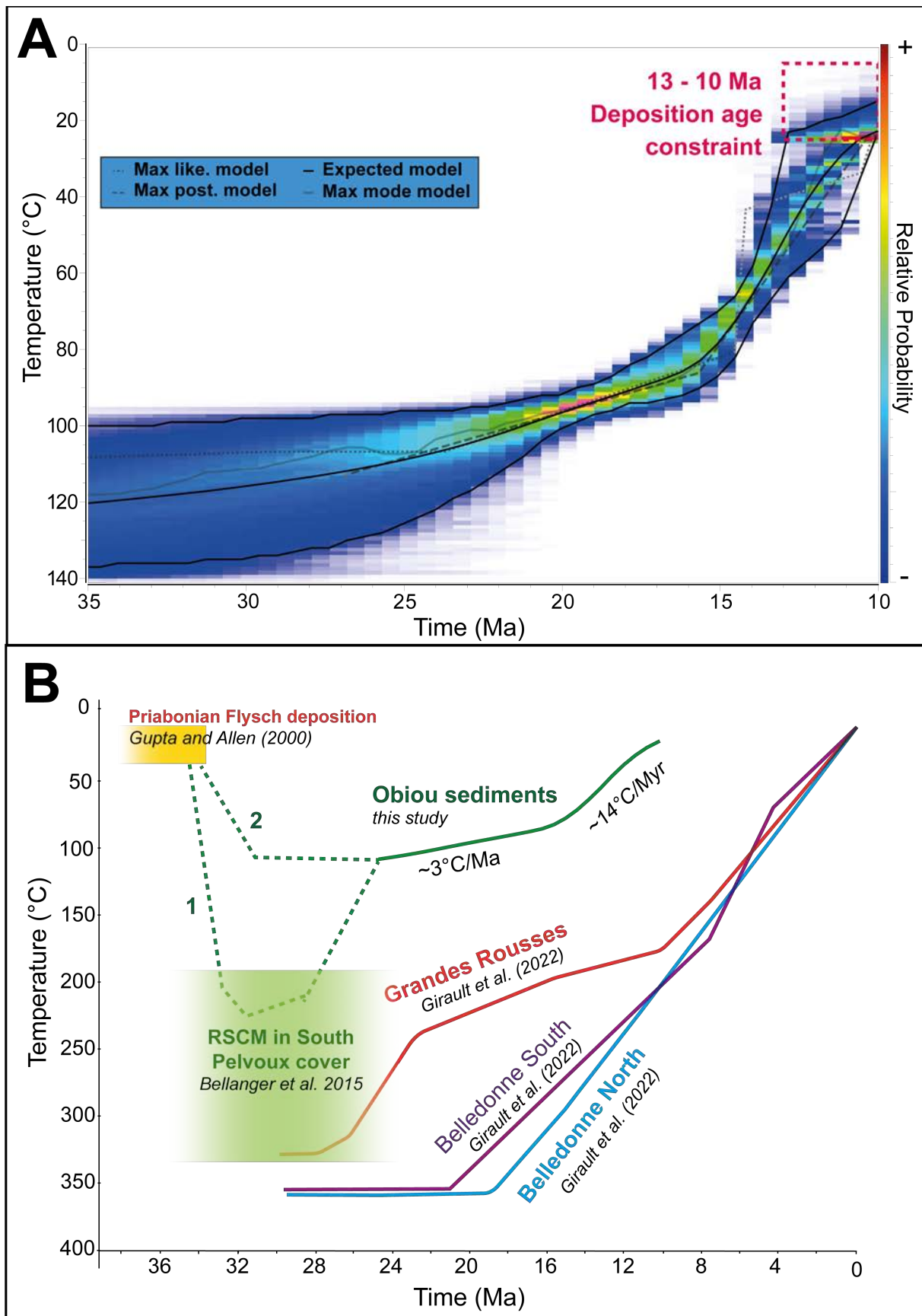
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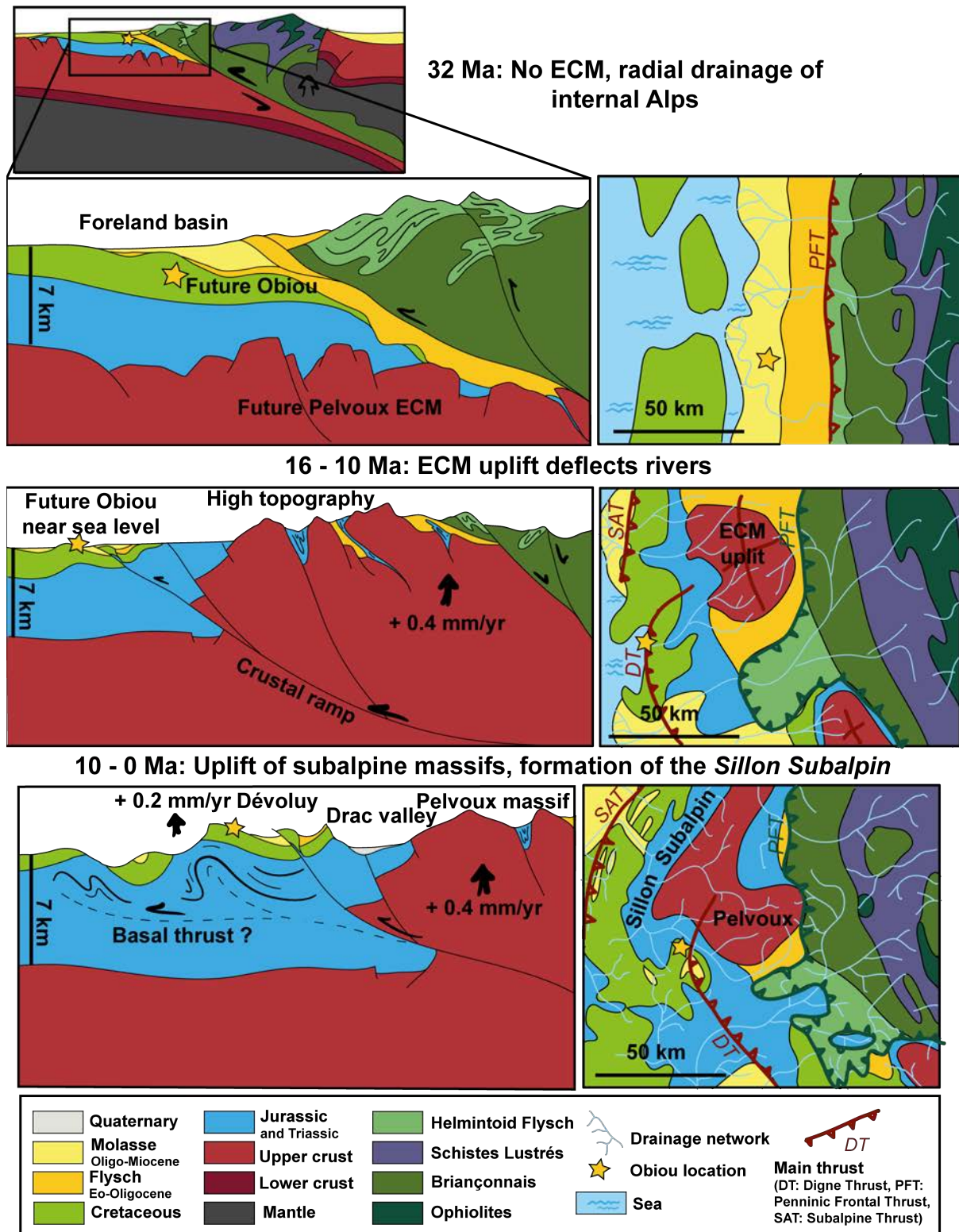
Figure 6: Detrital thermochronology results from the Obiou caves. A: Kernel density distributions of AFT single-grain ages; vertical bars represent central AFT ages. B: Distributions of single-grain AHe ages for Jalabres cave samples. See Supplementary material for full analytical data.



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827 **Figure 7:** A. Thermal history inferred from the Jalabres cave sediments, using available
 828 thermochronology data as input and TCN burial ages as an independent constraint on depositional

829 age (red dashed box). Inferred temperature history is shown as relative probability (colors) with
 830 “best-fit” histories using different metrics and their 95% credible interval shown as black lines (see
 831 Gallagher, 2012 for details). B. Comparison with regional Time-temperature paths and constraints from
 832 the literature. The dashed lines represent two alternative scenarios discussed in the text.



833 **Figure 8:** Tectonic, topographic, and drainage development of the western Alps in the vicinity of the
 834

835 Ecrins-Pelvoux and Dévoluy massifs. Cross-sections and maps show three stages of evolution, from
836 top to bottom: Oligocene (Jourdan et al., 2012; Fauquette et al., 2015), middle-late Miocene (as
837 inferred from cave sediments), and modern setting.

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