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► **To cite this version:**

Apolline Mariotti, Pierre-henri Blard, Julien Charreau, Samuel Toucanne, Stephan J Jorry, et al.. Nonlinear forcing of climate on mountain denudation during glaciations. *Nature Geoscience*, 2021, 14 (1), pp.16-22. 10.1038/s41561-020-00672-2. hal-03103081

HAL Id: hal-03103081

<https://cnrs.hal.science/hal-03103081v1>

Submitted on 8 Jan 2021

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1 **Nonlinear forcing of climate on mountain**

2 **denudation during glaciations**

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19 **ABSTRACT**

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21 **Denudation is one of the main processes that shapes landscapes. Since temperature,**
22 **precipitation and glacial extents are key factors involved in denudation, climatic**
23 **fluctuations are thought to exert a strong control on this parameter over geological**
24 **timescales. However, the direct impacts of climatic variations on denudation remain**
25 **controversial, particularly those involving the Quaternary glacial cycles in mountain**
26 **environments. Here we measure *in-situ* cosmogenic ^{10}Be concentration in quartz in**
27 **marine turbidites of two high-resolution cores collected in the Mediterranean Sea,**
28 **providing a near-continuous (~1 to 2 ka temporal resolution) reconstruction of**
29 **denudation in the Southern Alps since 75 ka. This high-resolution paleo-denudation**
30 **record can be compared with well-constrained climatic variations over the last glacial**
31 **cycle. Our results indicate that total denudation rates were ~2 times higher than**
32 **present during the Last Glacial Maximum (26.5–19 ka), the glacial component of the**
33 **denudation rates being $1.5^{+0.9}_{-1.0}$ mm a⁻¹. However, during moderately glaciated times**
34 **(74–29 ka), denudation rates were similar to today (0.24 ± 0.04 mm a⁻¹). This suggests**
35 **a nonlinear forcing of climate on denudation, mainly controlled by the interplay**
36 **between glaciers velocity and basin topography. Hence, the onset of Quaternary**
37 **glaciations, 2.6 million years ago, did not necessarily induce a synchronous global**
38 **denudation pulse.**

39

40 How climate and climatic fluctuations affect denudation rates over geological
41 timescales remains debated^{1,2}. Relying on sedimentation records³ and thermochronology^{4,5}
42 several authors have reported increased denudation rates at the onset of Pleistocene
43 glaciations around 2.6 Ma⁶ either due to increased variability of Quaternary climatic cycles³ or

44 growth of glaciers in mountain ranges, an efficient agent of erosion^{4,7}. However, marine
45 ¹⁰Be/⁹Be records indicate stable weathering rates over the same period¹. Other studies have
46 highlighted the potential biases inherent in sedimentary records⁸ and thermochronology⁹,
47 and the link between weathering and erosion has also been questioned¹⁰. Moreover, the
48 presence of glaciers does not necessarily lead to an increase in denudation rates¹¹.

49 *In-situ* Terrestrial Cosmogenic Nuclides (TCNs) such as ¹⁰Be record denudation rates in
50 modern river systems¹²⁻¹⁴ as well as over geological (10 Ma) timescales^{15,16}. TCNs are not
51 affected by the same biases as sedimentation rate estimates or thermochronology, and they
52 track both weathering and erosion rates, i.e., denudation¹²⁻¹⁴, at sufficient resolution (<1 ka in
53 better cases) to capture the impact of millennial climatic variations. Previous paleo-
54 denudation studies have used continental sedimentary archives such as cave sediments^{17,18},
55 terrace deposits^{19,20}, lake sediments^{21,22} and molasse deposits in foreland basins^{15,16}.
56 However, these records are often too discontinuous to accurately reflect the impact of short-
57 term climatic variations on denudation. Additionally, the most complete records were
58 obtained in active tectonic settings over long time scales (>5 Ma), making it difficult to
59 decipher any climatic forcing¹⁵. Continuous, well-dated, and high-resolution marine sediment
60 records in well-constrained sediment routing systems have great potential to overcome these
61 limitations; they may represent the gold standard of TCN archives of paleo-denudation rates.

62

63 In this study, we measured *in-situ* ¹⁰Be concentrations in 75-0 ka marine sediments
64 exported by the Var River, which drains the Southern French Alps into the Mediterranean Sea
65 (Fig. 1).

66

67 **Southern Alps, the reactive Var basin and samples**

68 Southern French Alps are a mountainous region that is considered tectonically stable
69 since 75 ka, with near null horizontal and vertical motions²⁵. The Var river catchment (2,800
70 km², maximum altitude of 3,143 m above sea level, mean altitude of 1,250 m) is a reactive
71 watershed of this massif²⁶, i.e., one that responds quickly to any change in denudation or
72 sediment transport, with steep slopes (mean 23°) and a small alluvial plain (<50 km²) that
73 limits on-land sediment storage²⁴. The geology of the Var basin includes two main regions: the
74 crystalline Mercantour-Argentera massif in the northern and highest part of the catchment,
75 and sedimentary terrains in the southern part. Quartz-bearing rocks (¹⁰Be being measured in
76 quartz) cover 26% of the Var watershed and are mainly present in the upper part of the
77 catchment. The present-day ¹⁰Be-derived denudation rate of the entire Var catchment is 0.24
78 ± 0.04 mm a⁻¹ ²⁷. The mouth of the Var River is connected to a submarine canyon that incises
79 both a virtually absent continental shelf (width >200 m) ²⁸ and a steep (11°) continental
80 slope²⁹. Hence, during the Quaternary, sediments draining the catchment have been
81 continuously transferred to the deep depositional system of the Var in the form of turbidites,
82 building the Var Sedimentary Ridge (VSR, <2,000 m below sea level)^{28,29}.

83 Due to a colder climate than today, the European Alps were extensively glaciated
84 during Pleistocene (2.6 Ma to 11.7 ka)²³. In the Var catchment, glaciers occupied the upper
85 valleys of the Var, Tinée, and Vésubie Rivers (Fig. 1) during the Last Glacial Maximum²³ (LGM,
86 26.5–19 ka³⁰). When the glaciers receded and completely melted most recently (19–11 ka),
87 they left glacially polished surfaces³¹, and river incision began in the newly deglaciated
88 valleys³². This shift from glacial to fluvial environments is thought to have driven strong
89 changes in denudation processes^{24,33}. The high frequency of turbidites in the VSR is attributed
90 to flood events, and their geochemical signatures (εNd) have been correlated to climatic
91 changes in the Var basin, both over millennial Dansgaard-Oeschger cycles and glacial-

92 interglacial periods^{26,28}. This correlation is interpreted to have been driven by glacial
93 fluctuations, as the highest sediment fluxes occurred during glacial maxima²⁴.

94

95 A total of 26 samples of silts to fine sands (50–250 μm) were collected from two cores
96 drilled in the VSR (Fig. 2). ^{10}Be concentrations in the samples range from $(0.91 \pm 0.12) \times 10^4$ at
97 g^{-1} (19.8 ka, LGM) to $(5.62 \pm 0.44) \times 10^4$ at g^{-1} (73.9 ka, Marine Isotopic Stage (MIS) 5a³⁴) (see
98 Supp. Table 1 and 2, Methods, Supp. Section 1, Extended Data Fig. 1). Both Dansgaard-
99 Oeschger stadials and interstadials are documented by our sampling (Fig. 2a/b). Our data
100 show that grain size has no impact on the ^{10}Be concentrations (see Methods, Supplementary
101 Section 2). We analyzed 19 new samples (see Methods) to extend the coverage of pre-existing
102 ϵNd data from 50–20 ka²⁴ to 74–50 ka (Fig. 2c). Our new ϵNd values range between –10.3 and
103 –9.6 (average –10.1).

104

105 **From ^{10}Be concentrations to paleo-denudation rates**

106 Deriving basin-wide denudation rates from ^{10}Be concentrations requires computing
107 the ^{10}Be production rate at the time of sediment production. This parameter is mainly a
108 function of basin altitude, which has likely remained steady over the last 75 ka. ^{10}Be
109 production is also controlled by glacial extent because glaciers shield underlying rocks from
110 cosmic rays, effectively eliminating ^{10}Be production beneath several decameters of ice. To test
111 the sensitivity of the glacial shielding on the denudation rates, we considered two scenarii to
112 compute paleo-production rates³⁶. The first assumes that no glaciers were present between 0
113 and 75 ka (Scenario 1: ‘Ice-free’). Although this approach contradicts geomorphological
114 evidence of glaciers in the Var catchment^{31,37}, this endmember scenario provides maximum
115 denudation rate estimates. In the second scenario (Scenario 2: ‘Glaciated’), we computed
116 paleo-production rates by reconstructing glacial extents. We used $\delta^{18}\text{O}$ data from planktonic

117 foraminifera^{26,28} sampled in both cores to compute sea surface paleo-temperatures, which we
118 then used to reconstruct the glacial extents in the Var watershed using a simple 2D glacier
119 flow model including a positive degree-day mass-balance³⁸ (details in Supp. Section 3). The
120 basin-averaged paleo-production rates were then calculated with the BASINGA code,
121 assuming no cosmogenic production under ice³⁶. Our modeling yields average glacial extents
122 covering 21%, 15%, and 30% of quartz-bearing rocks during MISs 4, 3, and LGM (19-26.5 ka),
123 respectively. This glacial cover lowered production rates by 10% to 30% compared to the
124 respective ice-free conditions (Supp. Table 4 and Extended Data Fig. 2 and 3).

125

126 The paleo-denudation rates (both scenarii) range between 0.15 ± 0.01 and 1.26 ± 0.16
127 mm a^{-1} (all uncertainties are 1σ), implying ^{10}Be integration times of 0.5–3.9 ka (see Methods),
128 sufficiently short to observe the potential influence of major climatic changes both throughout
129 the last glacial cycle and during the deglaciation. Moreover, with such integration times,
130 potential inter-sample timescale biases are lower than 25%³⁹. Between 75 and 29 ka (MIS 5a–
131 3), both cosmogenic production models yield comparable and rather stable denudation rates
132 between 0.15 ± 0.01 (Scenario 2, 73.9 ka) and $0.41 \pm 0.04 \text{ mm a}^{-1}$ (Scenario 1, 31.3 ka), with
133 mean values of 0.32 ± 0.03 (Scenario 1) and $0.25 \pm 0.02 \text{ mm a}^{-1}$ (Scenario 2) (Fig. 2). These
134 rates are similar to the present-day value of $0.24 \pm 0.04 \text{ mm a}^{-1}$ ²⁷.

135

136 **Potential causes of denudation variations**

137 The coldest event of MIS 4, corresponding to a ~1 ka long episode according to the local
138 $\delta^{18}\text{O}$ (66 - 67 ka) (Fig. 2b), did not decrease the ^{10}Be concentration. Additionally, neither
139 climatic and glacial fluctuations reported in the Alps^{40,41} during MIS 4–3, nor Atlantic
140 Dansgaard-Oeschger oscillations (well identified from $\delta^{18}\text{O}$ records in the studied cores) had a
141 noticeable impact on the ^{10}Be signal, suggesting that denudation rates are not significantly

142 impacted by these millennial climatic variability^{1,33}. The observed fluctuations in the
143 turbidites frequency during MIS 4–3 (Fig. 2d) could thus reflect changing transport conditions
144 in the sedimentary system rather than variations in denudation rates²⁴. Nevertheless, ϵNd
145 data indicate that partially glaciated valleys (Tinée and Vésubie) represented a significant
146 contribution to the sedimentary supply between 75 and 20 ka (MIS 5a–2) (Fig. 2c).

147

148 Denudation rates obtained using both scenarii during MIS 2 and the LGM were more
149 variable and on average two times higher than the present-day rate and those during MIS 4–3
150 (Fig. 2e, Extended Data Fig. 4). Indeed, the average ^{10}Be concentrations of samples from the
151 LGM are 50% lower than those of MIS 4–3 and Holocene samples (see Supplementary Table
152 1). This period of lower ^{10}Be coincides with the largest glacial extent over the last 75 ka:
153 during the LGM (19–26.5 ka), glaciers covered nearly 30% of the quartz-bearing surfaces in
154 the Var catchment (Extended Data Fig. 3). This drop in concentration could be due to
155 contributions from ^{10}Be -depleted sediments previously stored in the catchment. Such a
156 mixing mechanism would lower the average ^{10}Be concentration, thus artificially enhancing
157 denudation rates. A potential source of ^{10}Be -depleted material is Pliocene sediments stored in
158 the Var Canyon (mapped as Neogene-Quaternary near the outlet) (Fig. 1), whose incision may
159 have been triggered by ~40 m of sea level drop between the end of MIS 3 and MIS 2⁴². During
160 warm ice-free periods, such as the Holocene, the ϵNd signature of the Var sediment is low (-11
161 to -10.5), indicating a lower contribution of radiogenic highland terrains (Fig. 2)²⁴. Because
162 Pliocene sediments were deposited during a warm and non-glaciated period, they are
163 probably also characterized by low ϵNd value (-11 to -10.5). Any reworking of these
164 sediments during sea level drops should thus be associated with a drop in ϵNd values. The
165 measured ϵNd values during the LGM (mean -10.1) are similar to the values measured during
166 MIS 3–4, indicating that reworking of Pliocene sediments is unlikely. Another potential source

167 of ^{10}Be -depleted materials could be paleo-moraines that accumulated on slopes during MIS 4–
168 3. Assuming a steady denudation rate of $\sim 0.24 \text{ mm a}^{-1}$ during the LGM (similar to MIS 4–3
169 and present-day values²⁷), a total of $\sim 1 \text{ km}^3$ of sediments was produced over this 7.5-ka-long
170 period. The same volume of reworked MIS 4–3 moraine sediments would be required to
171 lower the LGM ^{10}Be concentrations by 50% (see Supp. Section 4 and Extended Data Fig. 5 and
172 6). Assuming a simple triangular moraine shape, 1 km^3 MIS 4–3 moraines would have been
173 $\sim 130 \text{ km}$ long, 150 m wide, and 100 m high (see Supp. Section 4). Such large moraines are
174 unlikely, largest pre-Holocene moraines observable today in the Southern Alps being only 25 m
175 high and 80 m wide⁴³. Another mechanism is therefore required to explain the 50% drop of
176 cosmogenic ^{10}Be concentrations observed in Var sediments during the LGM.

177

178 When LGM glaciers intruded into lower valleys, they were probably able to increase
179 bedrock incision and local denudation rates. The main mechanism here could be the basal
180 velocity of the glaciers, that controls the denudation efficiency⁴⁴: by flowing into narrow
181 valleys, LGM glaciers had higher local velocities than MIS 3-4 glaciers, and, hence, induced
182 higher glacial incision^{44,45} (Fig. 3).

183

184 After a lag time of few thousand years necessary to remove the first meters of the ^{10}Be -
185 rich fluvial landscape soil, the glacial incision then delivered shielded material to the system.
186 After mixing with the fluvial sediments in the lower part of the watershed, this glacial input
187 induced a drop in ^{10}Be concentrations.

188

189 Such denudation solely due to the glacial incision can be quantified using a mass
190 balance approach, which assumes that the ^{10}Be concentration measured in each sample is a
191 mix of glacial and fluvial contributions⁴⁶ (see Supp. Section 5). The resulting average glacial

192 erosion rate during the LGM is $1.5_{-1.0}^{+0.9}$ mm a⁻¹, roughly four times above the value of $0.4_{-0.5}^{+0.4}$
193 obtained during MIS 4–3 (Figs. 2, 3, Extended Data Fig. 7 and 8). This LGM glacial denudation
194 is similar to the maximum local glacial incision estimated in the upper Tinée during MIS 5d-2
195 (1.8 mm a⁻¹)³¹, and to the value proposed for present-day small temperate glaciers in the
196 Swiss Alps (1 mm a⁻¹)⁷. At such incision rates, glacial incision requires 2 to 4 kyr to reach
197 substratum with null ¹⁰Be concentration. This is consistent with the observation that ¹⁰Be
198 dropped after 25 ka, ~4 kyr after the onset of the cold conditions of MIS 2 (Fig. 2).

199

200 During MIS 4-3, between 75 and 26.5 ka, glaciers existed but were likely restricted
201 most of the time to the highest parts of the catchment, in cirques and hanging valleys, with a
202 low erosive power⁴⁵ (Extended Data Fig. 9). Only for ~1 ka during the MIS 4 glaciers possibly
203 reached the lower valley as during the LGM (according to the local $\delta^{18}\text{O}$ curve (Fig. 2b)).
204 However, this glacial advance was likely too short to significantly carve the landscape and
205 deeply erode the bedrock to produce ¹⁰Be depleted sediments. During the LGM, the advance of
206 glaciers in deep valleys likely increased denudation during 7 ka. Along with the reworking of
207 MIS 4–3 moraines, this contributed to lowering the ¹⁰Be concentrations and increasing the
208 overall denudation rates of the Var basin (Figs. 2e, 3).

209

210 During deglaciation (19–11 ka)³⁷, the overall ¹⁰Be-denudation rates dropped,
211 becoming close to the present-day value (Fig. 2e). The denudation decrease between 19 and
212 15 ka is concomitant with the ϵNd -inferred shift of sediment sources from exposed crystalline
213 massifs to sedimentary cover (Fig. 2c), and with a decrease in riverine sediment inputs (Fig.
214 2d)²⁶. These changes suggest that denudation dynamics returned to fluvial conditions during
215 that time²⁴. At 10 ka, we observe a spike in denudation rates synchronous with a positive ϵNd
216 excursion (Fig. 2e). This short-term ¹⁰Be spike is only supported by one core sample and

217 occurred during a period of increased rainfall in the Mediterranean region⁴⁷. Together with
218 the ϵNd spike, this excursion is characteristic of a paraglacial lag in the sedimentary transport
219 of previously eroded glacial sediments⁴⁸ and should not be considered as an actual increase in
220 bedrock denudation²⁴. After this spike, Holocene denudation rates are similar to present-day
221 rates and those observed during MIS 4–3 (Figs. 2, 3).

222

223 Our 75-ka-long denudation record from the Southern French Alps indicates a nonlinear
224 link between climate and denudation, as shown by the relationship between $\delta^{18}\text{O}$ -derived sea
225 surface temperatures and ^{10}Be -derived denudation rates (Fig. 4).

226

227 **Implications for other regions and climatic periods**

228 Our data suggest that climatic variations may only strongly affect denudation beyond a
229 certain threshold, probably controlled by the glacier dynamics^{44,45}, the duration of glacial
230 advances and temperature driven processes (e.g. frost cracking)^{49,50}. Since the denudation is
231 enhanced during the LGM period, when glacier advanced and carved deep U-shaped narrow
232 valleys, basal ice velocity was probably the main driver of this non-linear response. This
233 mechanism was suggested for present-day⁴⁴ and at million years timescale¹⁸, and our study is
234 the first to detect such behaviour at the glacial-interglacial timescale.

235 In conclusion, our study indicates that the denudation response to Quaternary
236 glaciations is complex and non-linear in glaciated areas. This result has strong and broad
237 implications for the erosion dynamics: climatic variations affect denudations on the glacial-
238 interglacial timescales in a non-linear way, basin topography and duration of glacial advances
239 being major forcings. Since glacier dynamics are nonlinearly controlled by climate and
240 topography, climatic impacts on denudation in glaciated regions may thus be
241 spatiotemporally variable. Therefore, our study suggests that, in glaciated regions, the

242 response of the landscapes to the onset of Quaternary glaciation is not necessarily a
243 synchronous global pulse³.

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269 **REFERENCES**

- 270 1. Willenbring, J. K. & Jerolmack, D. J. The null hypothesis: globally steady rates of erosion,
271 weathering fluxes and shelf sediment accumulation during Late Cenozoic mountain uplift and
272 glaciation. *Terra Nov.* **28**, 11–18 (2016).
- 273 2. Herman, F. & Champagnac, J.-D. Plio-Pleistocene increase of erosion rates in mountain
274 belts in response to climate change. *Terra Nov.* **28**, 2–10 (2016).
- 275 3. Peizhen, Z. *et al.* Increased sedimentation rates and grain sizes 2–4 Myr ago due to the
276 influence of climate change on erosion rates. *Nature* **410**, 891–897 (2001).
- 277 4. Herman, F. *et al.* Worldwide acceleration of mountain erosion under a cooling climate.
278 *Nature* **504**, 423–426 (2013).
- 279 5. Valla, P. G., Shuster, D. L. & Van Der Beek, P. A. Significant increase in relief of the
280 European Alps during mid-Pleistocene glaciations. *Nat. Geosci.* **4**, 688–692 (2011).
- 281 6. Batchelor, C. L. *et al.* The configuration of Northern Hemisphere ice sheets through the
282 Quaternary. *Nat. Commun.* **10**, 3713 (2019).
- 283 7. Hallet, B., Hunter, L. & Bogen, J. Rates of erosion and sediment evacuation by glaciers: A
284 review of field data and their implications. *Glob. Planet. Change* **12**, 213–235 (1996).
- 285 8. Sadler, P. M. & Jerolmack, D. J. Scaling laws for aggradation, denudation and
286 progradation rates: the case for time-scale invariance at sediment sources and sinks. *Geol. Soc.*
287 *London, Spec. Publ.* **404**, 69–88 (2015).
- 288 9. Schildgen, T. F., Van Der Beek, P. A., Sinclair, H. D. & Thiede, R. C. Spatial correlation
289 bias in late-Cenozoic erosion histories derived from thermochronology. *Nature* **559**, 89–93
290 (2018).
- 291 10. Norton, K. P. & Schlunegger, F. Lack of a weathering signal with increased Cenozoic
292 erosion? *Terra Nov.* **29**, 1–8 (2017).

- 293 11. Thomson, S. N. *et al.* Glaciation as a destructive and constructive control on mountain
294 building. *Nature* **467**, 313–317 (2010).
- 295 12. Brown, E. T., Stallard, R. F., Larsen, M. C., Raisbeck, G. M. & Yiou, F. Denudation rates
296 determined from the accumulation of in situ-produced ^{10}Be in the luquillo experimental
297 forest, Puerto Rico. *Earth Planet. Sci. Lett.* **129**, 193–202 (1995).
- 298 13. Bierman, P. R. & Steig, E. J. Estimating rates of denudation using cosmogenic isotope
299 abundances in sediment. *Earth Surface Processes and Landforms* **21**, 125–139 (1996).
- 300 14. Granger, D. E., Kirchner, J. W. & Finkel, R. C. Spatially Averaged Long-Term Erosion
301 Rates Measured from in Situ-Produced Cosmogenic Nuclides in Alluvial Sediment. *J. Geol.* **104**,
302 249–257 (1996).
- 303 15. Puchol, N. *et al.* Limited impact of Quaternary glaciations on denudation rates in
304 Central Asia. *Geol. Soc. Am. Bull.* **129**, 479–499 (2017).
- 305 16. Charreau, J. *et al.* Paleo-erosion rates in Central Asia since 9Ma: A transient increase at
306 the onset of Quaternary glaciations? *Earth Planet. Sci. Lett.* **304**, 85–92 (2011).
- 307 17. Anthony, D. M. & Granger, D. E. A new chronology for the age of Appalachian erosional
308 surfaces determined by cosmogenic nuclides in cave sediments. *Earth Surf. Process. Landforms*
309 **32**, 874–887 (2007).
- 310 18. Haeuselmann, P., Granger, D. E., Jeannin, P.-Y. & Lauritzen, S.-E. Abrupt glacial valley
311 incision at 0.8 Ma dated from cave deposits in Switzerland. *Geology* **35**, 143 (2007).
- 312 19. Schaller, M. *et al.* A 30 000 yr record of erosion rates from cosmogenic ^{10}Be in Middle
313 European river terraces. *Earth Planet. Sci. Lett.* **204**, 307–320 (2002).
- 314 20. Bekaddour, T. *et al.* Paleo erosion rates and climate shifts recorded by Quaternary cut-
315 and-fill sequences in the Pisco valley, central Peru. *Earth Planet. Sci. Lett.* **390**, 103–115
316 (2014).

- 317 21. Marshall, J. A., Roering, J. J., Gavin, D. G. & Granger, D. E. Late Quaternary climatic
318 controls on erosion rates and geomorphic processes in western Oregon, USA. *Geol. Soc. Am.*
319 *Bull.* **129**, 715–731 (2017).
- 320 22. Grischott, R. *et al.* Millennial scale variability of denudation rates for the last 15 kyr
321 inferred from the detrital ^{10}Be record of Lake Stappitz in the Hohe Tauern massif, Austrian
322 Alps. *The Holocene* **27**, 1914–1927 (2017).
- 323 23. Ehlers, J., Gibbard, P. L. & Hughes, P. D. *Quaternary Glaciations - Extent and Chronology*
324 *A Closer Look : part I: Europe.* (Elsevier Ltd., 2011).
- 325 24. Bonneau, L. *et al.* Glacial erosion dynamics in a small mountainous watershed
326 (Southern French Alps): A source-to-sink approach. *Earth Planet. Sci. Lett.* **458**, 366–379
327 (2017).
- 328 25. Nocquet, J.-M. *et al.* Present-day uplift of the western Alps. *Sci. Rep.* **6**, 28404 (2016).
- 329 26. Bonneau, L., Jorry, S. J., Toucanne, S., Silva Jacinto, R. & Emmanuel, L. Millennial-Scale
330 Response of a Western Mediterranean River to Late Quaternary Climate Changes: A View
331 from the Deep Sea. *J. Geol.* **122**, 687–703 (2014).
- 332 27. Mariotti, A., Blard, P.-H., Charreau, J., Petit, C. & Molliex, S. Denudation systematics
333 inferred from in situ cosmogenic ^{10}Be concentrations in fine (50–100 μm) and medium
334 (100–250 μm) sediments of the Var River basin, southern French Alps. *Earth Surf. Dyn.* **7**,
335 1059–1074 (2019).
- 336 28. Jorry, S. J., Jégou, I., Emmanuel, L., Silva Jacinto, R. & Savoye, B. Turbiditic levee
337 deposition in response to climate changes: The Var Sedimentary Ridge (Ligurian Sea). *Mar.*
338 *Geol.* **279**, 148–161 (2011).
- 339 29. Savoye, B., Piper, D. J. W. & Droz, L. Plio-Pleistocene evolution of the Var deep-sea fan
340 off the French Riviera. *Mar. Pet. Geol.* **10**, 550–571 (1993).
- 341 30. Clark, P. U. *et al.* The Last Glacial Maximum. *Science (80-.)*. **325**, 710–714 (2009).

- 342 31. Darnault, R. *et al.* Timing of the last deglaciation revealed by receding glaciers at the
343 Alpine-scale: impact on mountain geomorphology. *Quat. Sci. Rev.* **31**, 127–142 (2012).
- 344 32. Petit, C. *et al.* Recent, climate-driven river incision rate fluctuations in the Mercantour
345 crystalline massif, southern French Alps. *Quat. Sci. Rev.* **165**, 73–87 (2017).
- 346 33. Koppes, M. & Montgomery, D. R. The relative efficacy of fluvial and glacial erosion over
347 modern to orogenic timescales. *Nat. Geosci.* **2**, 644–647 (2009).
- 348 34. Railsback, L. B., Gibbard, P. L., Head, M. J., Voarintsoa, N. R. G. & Toucanne, S. An
349 optimized scheme of lettered marine isotope substages for the last 1.0 million years, and the
350 climatostratigraphic nature of isotope stages and substages. *Quat. Sci. Rev.* **111**, 94–106
351 (2015).
- 352 35. Andersen, K. K. *et al.* High-resolution record of Northern Hemisphere climate extending
353 into the last interglacial period. *Nature* **431**, 147–151 (2004).
- 354 36. Charreau, J. *et al.* Basinga: A cell - by - cell GIS toolbox for computing basin average
355 scaling factors, cosmogenic production rates and denudation rates. *Earth Surf. Process.*
356 *Landforms* **44**, 2349–2365 (2019).
- 357 37. Rolland, Y. *et al.* Deglaciation history at the Alpine - Mediterranean transition
358 (Argentera - Mercantour, SW Alps) from 10 Be dating of moraines and glacially polished
359 bedrock. *Earth Surf. Process. Landforms* **45**, 393–410 (2020).
- 360 38. Harper, J. T. & Humphrey, N. F. High altitude Himalayan climate inferred from glacial
361 ice flux. *Geophys. Res. Lett.* **30**, 3–6 (2003).
- 362 39. Ganti, V. *et al.* Time scale bias in erosion rates of glaciated landscapes. *Sci. Adv.* **2**,
363 e1600204–e1600204 (2016).
- 364 40. Heiri, O. *et al.* Palaeoclimate records 60–8 ka in the Austrian and Swiss Alps and their
365 forelands. *Quat. Sci. Rev.* **106**, 186–205 (2014).

- 366 41. Ivy-Ochs, S. *et al.* Chronology of the last glacial cycle in the European Alps. *J. Quat. Sci.*
367 **23**, 559–573 (2008).
- 368 42. Waelbroeck, C. *et al.* Sea-level and deep water temperature changes derived from
369 benthic foraminifera isotopic records. *Quat. Sci. Rev.* **21**, 295–305 (2002).
- 370 43. Federici, P. R. *et al.* Last Glacial Maximum and the Gschnitz stadial in the Maritime Alps
371 according to ^{10}Be cosmogenic dating. *Boreas* **41**, 277–291 (2012).
- 372 44. Herman, F. *et al.* Erosion by an Alpine glacier. *Science (80-.)*. **350**, 193–195 (2015).
- 373 45. Brædstrup, C. F., Egholm, D. L., Ugelvig, S. V. & Pedersen, V. K. Basal shear stress under
374 alpine glaciers: insights from experiments using the iSOSIA and Elmer/Ice models. *Earth Surf.*
375 *Dyn.* **4**, 159–174 (2016).
- 376 46. Guillon, H. *et al.* Improved discrimination of subglacial and periglacial erosion
377 using ^{10}Be concentration measurements in subglacial and supraglacial sediment load of the
378 Bossons glacier (Mont Blanc massif, France). *Earth Surf. Process. Landforms* **40**, 1202–1215
379 (2015).
- 380 47. Toucanne, S. *et al.* Tracking rainfall in the northern Mediterranean borderlands during
381 sapropel deposition. *Quat. Sci. Rev.* **129**, 178–195 (2015).
- 382 48. Church, M. & Ryder, J. Paraglacial Sedimentation: A Consideration of Fluvial Processes
383 Conditioned by Glaciation. *Bull. Geol. Soc. Am.* **83**, 319–342 (1972).
- 384 49. Delunel, R., Van Der Beek, P. A., Carcaillet, J., Bourlès, D. L. & Valla, P. G. Frost-cracking
385 control on catchment denudation rates: Insights from in situ produced ^{10}Be concentrations in
386 stream sediments (Ecrins–Pelvoux massif, French Western Alps). *Earth Planet. Sci. Lett.* **293**,
387 72–83 (2010).
- 388 50. Egholm, D. L., Andersen, J. L., Knudsen, M. F., Jansen, J. D. & Nielsen, S. B. The periglacial
389 engine of mountain erosion - Part 2: Modelling large-scale landscape evolution. *Earth Surf.*
390 *Dyn.* **3**, 463–482 (2015).

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396 **ACKNOWLEDGEMENTS**

397 This research was funded by the ANR JC EroMed project (N° ANR-17-CE01-0011-01; Principal

398 Investigator P.-H. Blard). This work is part of the PhD thesis of Apolline Mariotti, whose

399 scholarship was jointly supported by the CNRS and the Région Grand Est. The ASTER AMS

400 national facility (CEREGE, Aix en Provence) is supported by the INSU/CNRS, the ANR through

401 the "Projets thématiques d'excellence" program for the "Equipements d'excellence" ASTER

402 CEREGE action and IRD. Fruitful discussions with Carole Petit, Vincent Godard and Lucille

403 Bonneau benefited the design of the study and interpretation of the results. We are grateful to

404 Laëticia Léanni and Régis Braucher for their expertise in ^{10}Be wet chemistry at LN2C/CEREGE

405 and to Anne Trinquier at IFREMER for the Nd measurements. Robert Dennen is acknowledged

406 for proofreading and improving the overall readability of the manuscript.

407

408 **AUTHOR CONTRIBUTIONS**

409 A.M., P.-H.B., S.T., J.C. and S.J.J designed the study. P.-H.B, S.T., S.M., and A.M. collected the

410 samples for analysis. A.M. prepared the samples for ^{10}Be and Nd analysis. D.B, G.A and K.K.

411 measured the $^{10}\text{Be}/^9\text{Be}$ ratios using the French Service National AMS ASTER. A.M., P.-H.B., J.C.,

412 S.T. and S.M., analyzed the data. AM wrote the initial manuscript, all authors commented and

413 contributed to the final version.

414

415 **COMPETING INTERESTS STATEMENT**

416

417 The authors declare that they have no conflict of interest.

418

419 **FIGURE CAPTIONS**

420 **Fig. 1 | The Var watershed and sedimentary system.** Geological data is from the French
421 Geological Survey (references provided in the Methods). Last Glacial Maximum (LGM) glacial
422 extent based on Ehlers et al., (2011)²³, and ϵNd sampling points from Bonneau et al., (2017)²⁴.
423 Sampling sites of cores ESK08-CS01 and ESK08-CS13 are shown in the Var Sedimentary
424 Ridge. Bathymetry from <https://www.emodnet-bathymetry.eu>

425

426 **Fig. 2 | Climate, sediment provenance, and flood proxies compared to denudation rates.**

427 a The high-resolution $\delta^{18}\text{O}$ record of Greenland ice core NGRIP (a proxy of climate)³⁵.
428 Numbers and labels indicate D-O events. b $\delta^{18}\text{O}$ values of *G. bulloides* (a local proxy of sea
429 surface temperature) and ^{14}C control points from the studied cores^{24,26,28}. Point colours
430 indicate the core (data in Supp. Table S2). c ϵND data from the ESK08-CS01 and ESK0-CS13
431 cores, reflecting sediment provenance. The most recent data (since ~50 ka) are from Bonneau
432 et al.²⁴, and the points labeled "this study" are new data. The orange shaded area corresponds
433 to 2σ uncertainties (data in Supp Table S3). d Mean turbidite fluxes over the two cores²⁶. e
434 ^{10}Be -derived denudation rates according to the ice-free and glaciated scenarii (Scenarii 1 and
435 2, respectively. Data in Supp. Table S4). Colored squares adjacent to the data points indicate
436 the integration time (x-axis) and uncertainty (1σ , y-axis) for each sample. The range of
437 present-day values is from Mariotti et al.²⁷, and the LGM is from Clark et al.³⁰. Purely glacial
438 erosion rates were calculated by mass balance, accounting only for erosion in glaciated areas
439 (i.e., excluding fluvial contributions to the denudation rate; see text and Supplementary
440 Information, data in Supp. Table S5).

441

442 **Fig. 3 | Schematic representations of denudation dynamics in the Var watershed.**

443 Denudation was calculated according to the glaciated scenario (see Methods and Supp Info) in
444 the Var watershed. a MIS 4–3 (71–29 ka). b the Last Glacial Maximum (LGM, 26.5–19 ka). c
445 present day²⁷. For MIS 3–4 and LGM, ϵ_G values represent purely glacial denudation (i.e.,
446 without any contribution from fluvial processes), and were estimated using variable ^{10}Be
447 concentrations in glaciated areas (see Methods). ϵ_{mix} values are weighted means of the
448 catchment-wide (glacial and fluvial) denudation rates. ϵ_F is the present-day catchment-wide
449 (fluvial) denudation rate²⁷. All uncertainties are 1σ . In the upper part of the catchment,
450 orange arrows indicate slow glacial flow and limited glacial erosion. In distal narrow valleys
451 downstream, red arrows indicate high flow glacial velocity and high subglacial denudation
452 rates.

453 **Fig. 1 | Comparison of denudation rates and sea surface temperature proxy data.** ^{10}Be -
454 derived denudation rates as a function of $\delta^{18}\text{O}$ values measured in the cores, according to the
455 two scenarii used to convert ^{10}Be concentrations to denudation rates (see Methods).
456 Uncertainties on the denudation rates correspond to 1σ . Although there is no clear
457 relationship between the two parameters, higher denudation rates only occur for $\delta^{18}\text{O}$ values
458 higher than 3.5‰.

459

460

461 **METHODS**

462

463 **Cores and samples**

464 We sampled sediments (Supplementary Table 1) from two Calypso piston cores collected in
465 2008⁵¹ (Fig. 1): ESK08-CS01 (22 m long) is from the VSR levee crest (2,146 m water depth)
466 and ESK01-CS13 (24 m long) from the southwestern flank of the VSR (2,473 m water depth).

467 Age models for both cores were previously anchored by 25 ¹⁴C dates, as well as high-
468 resolution planktonic oxygen isotope stratigraphy^{26,28} (Fig. 2a, b, Supplementary Table 2).
469 Accordingly, core ESK08-CS01 spans 0.3–29.8 ka and ESK08-CS13 spans 0.5–74.2 ka,
470 corresponding to mean sedimentation rates of 80 and 30 cm ka⁻¹, respectively.

471 Twenty-six samples (19 in ESK01-CS13 and 4 in ESK08-CS01) were selected from
472 coarser (>50 μm) turbiditic facies suitable for *in-situ* ¹⁰Be analyses in quartz²⁷. Because of the
473 reactive nature of the watershed and the depth of the VSR, all samples were buried quickly
474 and have not been re-exposed to cosmic radiation: changes in ¹⁰Be concentrations thus reflect
475 only denudation rate variations. Sample preparation and analytical methods are described in
476 the Supplementary Information.

477

478 **Computation of ¹⁰Be-derived denudation rates**

479 Basin-averaged denudation rates were calculated using a simplified version of Eq. 1 of
480 Brown¹²:

$$481 \quad \bar{\varepsilon} = \frac{1}{\rho} \sum_{i,x} \frac{\bar{P}_i \Lambda_i}{\bar{C}} \quad (1)$$

482 where *i* refers to the different ¹⁰Be production pathways (n for neutron, μs for slow muons,
483 and μf for fast muons); \bar{P}_i is the mean basin-wide ¹⁰Be production rate, computed from the
484 arithmetic mean of spallogenic and muogenic production rates using specific scalings for each
485 production pathway as a function of elevation and latitude; Λ_i are the attenuation lengths for
486 each particle ($\Lambda_n = 160 \text{ g cm}^{-2}$, $\Lambda_{\mu s} = 1,500 \text{ g cm}^{-2}$, and $\Lambda_{\mu f} = 4,320 \text{ g cm}^{-2}$)⁵²; ρ is the mean
487 density of the eroded material (2.7 g cm⁻³); and \bar{C} is the ¹⁰Be concentration measured in the
488 sediments sampled from the cores, after correction for radioactive decay (maximum
489 correction of 4% for the oldest sample, 73.9 ka). \bar{P}_i was calculated using the BASINGA GIS
490 tool³⁶ and the Lal-Stone scaling model⁵³. We used the following values as sea-level, high-

491 latitude production rates: $P_n = 4.11 \pm 0.19$ atoms $g^{-1} a^{-1}$, $P_{\mu s} = 0.011 \pm 0.001$ atoms $g^{-1} a^{-1}$, and
492 $P_{\mu f} = 0.039 \pm 0.004$ atoms $g^{-1} a^{-1}$ (global mean derived from the CREp online calculator;
493 <https://crep.otelo.univ-lorraine.fr/>)^{52,54}. All production rates are given in Supplementary
494 Tables 3 and 4. All measured $^{10}Be/^{9}Be$ ratios were calibrated using the CEREGE STD11 in-
495 house normalization, which is similar to the KNSTD07 normalization and assumes a $^{10}Be/^{9}Be$
496 ratio of $(2.79 \pm 0.03) \times 10^{-11}$ for the SRM4325 standard material⁵⁵.

497 Quartz-free areas (based on 1:250,000-scale French Geological Survey geological
498 maps^{56,57}) were excluded from the basin-averaged production rate calculations. Production
499 rates were corrected for topographic shielding, ice cover, and changes in paleo-magnetic
500 variations occurring during the integration time³⁶. Topographic shielding factors were
501 calculated using an ArcGIS toolbox⁵⁸ that computes both self-shielding and shading. Ice cover
502 was estimated independently for each sample, and production rates were considered null
503 under the extent of the ice cover. Paleomagnetic changes were averaged over the 3 ka
504 preceding the age of the sample. We used paleomagnetic data from Muscheler et al.⁵⁹ for
505 samples dating to 0–60 ka and data from Valet et al.⁶⁰ for older samples. Paleomagnetic
506 correction factors ranged from 0.90 to 1.06.

507

508 **Integration time**

509 A crucial consideration in cosmogenic dating is the method's ability to detect changes
510 in denudation rates. This mainly depends on the integration time, T_{int} , which is the amount of
511 time required to remove about 60 cm of bedrock, calculated as $T_{int} = \lambda/(\rho\varepsilon)$, where λ is the
512 neutronic attenuation length ($g\ cm^{-2}$), ρ the rock density ($g\ cm^{-3}$) and ε the denudation rate
513 ($cm\ a^{-1}$)^{12,61}. To detect past changes, this integration time must be shorter than the timescale
514 of denudation rate changes. In this dataset, the computed integration time ranges from 0.5 to
515 3.9 ka.

516

517 **Impact of grain-size on *in-situ* ¹⁰Be concentration measurements**

518 Our previous study showed that, in modern Var sediments, both the 50–100 μm and
519 100–250 μm size fractions have similar ¹⁰Be concentrations at the river outlet²⁷. To verify that
520 submarine processes did not induce a grain size-related bias, here we analyzed ¹⁰Be
521 concentrations in the same size fractions when the sample size was large enough (17 of 26
522 samples). ¹⁰Be concentrations were compatible at the 2σ level for both size fractions in all 17
523 samples, and the best-fit regression is statistically compatible with the 1:1 line (Extended
524 Data Fig. 1). These results suggest that submarine turbidite dynamics do not induce a record
525 bias. Consequently, we used the weighted mean ¹⁰Be concentrations from both grain sizes to
526 derive denudation rates (Supplementary Table 1).

527

528 **Sediment provenance**

529 We use εNd values measured in bulk sediments (finer than 63 μm) to trace sediment
530 provenance across the catchment. Indeed, it was previously shown that the εNd signatures in
531 cores ESSK08-CS01 and ESSK08-CS13 provide a reliable means to distinguish sediment
532 contributions from glaciated and non-glaciated areas²⁴ (Fig. 2c). Here we used the same
533 methodology²⁴ to complete the εNd record to the bottom of core ESSK08-CS13, i.e., from 50 to
534 74 ka (Supplementary Table 5). Prior to sample analysis, we removed carbonates, Fe-Mn
535 oxyhydroxides, and organic compounds by successive leachings. The completed εNd record
536 allows direct comparison with the ¹⁰Be data over the entire study period (Fig. 2c).

537

538 **DATA AVAILABILITY**

539 The datasets generated and analyzed during the current study are available in the ORDaR
540 Repository:

541 <https://doi.org/10.24396/ORDAR-46> (Table S1 - Raw ^{10}Be data),
542 <https://doi.org/10.24396/ORDAR-47> (Table S2 - Core age models),
543 <https://doi.org/10.24396/ORDAR-48> (Table S3 - ϵNd data),
544 <https://doi.org/10.24396/ORDAR-49> (Table S4 - Denudation rates),
545 <https://doi.org/10.24396/ORDAR-50> (Table S5 - Glacial erosion rates),
546 <https://doi.org/10.24396/ORDAR-51> (Table S6 - Monte Carlo draws parameters).

547

548 **CODE AVAILABILITY**

549 The Matlab [®] code used to determine glacial erosion rates is available upon request. Send an
550 email to P.-H. Blard: blard@crpg.cnrs-nancy.fr

551

552 **REFERENCES**

553

- 554 51. Lefort, O. ESSDIV Cruise - RV Pourquoi pas ? *IFREMER* (2008). doi:10.17600/8030010
- 555 52. Braucher, R., Merchel, S., Borgomano, J. & Bourlès, D. L. Production of cosmogenic
556 radionuclides at great depth: A multi element approach. *Earth Planet. Sci. Lett.* **309**, 1–9
557 (2011).
- 558 53. Stone, J. O. H. Air pressure and cosmogenic isotope production. *J. Geophys. Res. Solid*
559 *Earth* **105**, 23753–23759 (2000).
- 560 54. Martin, L. C. P. *et al.* The CREp program and the ICE-D production rate calibration
561 database: A fully parameterizable and updated online tool to compute cosmic-ray
562 exposure ages. *Quat. Geochronol.* **38**, 25–49 (2017).
- 563 55. Braucher, R. *et al.* Preparation of ASTER in-house $^{10}\text{Be}/^9\text{Be}$ standard solutions. *Nucl.*
564 *Instruments Methods Phys. Res. Sect. B Beam Interact. with Mater. Atoms* **361**, 335–340
565 (2015).

- 566 56. Kergkhove, C. & Montjuvent, G. *Carte géologique de la France à 1/250 000 - Gap*. (BRGM,
567 1977).
- 568 57. Roure, J., Autran, A., Prost, A., Rossi, A. & Rousset, C. *Carte géologique de la France à*
569 *1/250 000 - Nice*. (1976).
- 570 58. Codilean, A. T. Calculation of the cosmogenic nuclide production topographic shielding
571 scaling factor for large areas using DEMs. *Earth Surf. Process. Landforms* **31**, 785–794
572 (2006).
- 573 59. Muscheler, R., Beer, J., Kubik, P. W. & Synal, H.-A. Geomagnetic field intensity during the
574 last 60,000 years based on ¹⁰Be and ³⁶Cl from the Summit ice cores and ¹⁴C. *Quat. Sci.*
575 *Rev.* **24**, 1849–1860 (2005).
- 576 60. Valet, J.-P., Meynadier, L. & Guyodo, Y. Geomagnetic dipole strength and reversal rate
577 over the past two million years. *Nature* **435**, 802–805 (2005).
- 578 61. Lal, D. Cosmic ray labeling of erosion surfaces: in situ nuclide production rates and
579 erosion models. *Earth Planet. Sci. Lett. Elsevier Sci. Publ. B.V* **104**, 424–439 (1991).
- 580 62. Douguedroit, A. & de Saintignon, M.-F. Décroissance des températures en montagne de
581 latitude moyenne : exemple des Alpes françaises du Nord. *Rev. géographie Alp.* **64**, 483–
582 494 (1976).
- 583 63. Harper, J. T. & Humphrey, N. F. High altitude Himalayan climate inferred from glacial ice
584 flux. *Geophys. Res. Lett.* **30**, 3–6 (2003).

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