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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  $^{10}\text{Be}$  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<sup>-1</sup>. However, during moderately glaciated times**  
34 **(74–29 ka), denudation rates were similar to today ( $0.24 \pm 0.04$  mm a<sup>-1</sup>). 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<sup>1,2</sup>. Relying on sedimentation records<sup>3</sup> and thermochronology<sup>4,5</sup>  
42 several authors have reported increased denudation rates at the onset of Pleistocene  
43 glaciations around 2.6 Ma<sup>6</sup> either due to increased variability of Quaternary climatic cycles<sup>3</sup> or

44 growth of glaciers in mountain ranges, an efficient agent of erosion<sup>4,7</sup>. However, marine  
45 <sup>10</sup>Be/<sup>9</sup>Be records indicate stable weathering rates over the same period<sup>1</sup>. Other studies have  
46 highlighted the potential biases inherent in sedimentary records<sup>8</sup> and thermochronology<sup>9</sup>,  
47 and the link between weathering and erosion has also been questioned<sup>10</sup>. Moreover, the  
48 presence of glaciers does not necessarily lead to an increase in denudation rates<sup>11</sup>.

49 *In-situ* Terrestrial Cosmogenic Nuclides (TCNs) such as <sup>10</sup>Be record denudation rates in  
50 modern river systems<sup>12-14</sup> as well as over geological (10 Ma) timescales<sup>15,16</sup>. 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., denudation<sup>12-14</sup>, 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<sup>17,18</sup>,  
55 terrace deposits<sup>19,20</sup>, lake sediments<sup>21,22</sup> and molasse deposits in foreland basins<sup>15,16</sup>.  
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 forcing<sup>15</sup>. 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* <sup>10</sup>Be 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 motions<sup>25</sup>. The Var river catchment (2,800  
70 km<sup>2</sup>, maximum altitude of 3,143 m above sea level, mean altitude of 1,250 m) is a reactive  
71 watershed of this massif<sup>26</sup>, 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 km<sup>2</sup>) that  
73 limits on-land sediment storage<sup>24</sup>. 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 (<sup>10</sup>Be 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 <sup>10</sup>Be-derived denudation rate of the entire Var catchment is 0.24  
78 ± 0.04 mm a<sup>-1</sup> <sup>27</sup>. The mouth of the Var River is connected to a submarine canyon that incises  
79 both a virtually absent continental shelf (width >200 m) <sup>28</sup> and a steep (11°) continental  
80 slope<sup>29</sup>. 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)<sup>28,29</sup>.

83 Due to a colder climate than today, the European Alps were extensively glaciated  
84 during Pleistocene (2.6 Ma to 11.7 ka)<sup>23</sup>. In the Var catchment, glaciers occupied the upper  
85 valleys of the Var, Tinée, and Vésubie Rivers (Fig. 1) during the Last Glacial Maximum<sup>23</sup> (LGM,  
86 26.5–19 ka<sup>30</sup>). When the glaciers receded and completely melted most recently (19–11 ka),  
87 they left glacially polished surfaces<sup>31</sup>, and river incision began in the newly deglaciated  
88 valleys<sup>32</sup>. This shift from glacial to fluvial environments is thought to have driven strong  
89 changes in denudation processes<sup>24,33</sup>. 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<sup>26,28</sup>. This correlation is interpreted to have been driven by glacial  
93 fluctuations, as the highest sediment fluxes occurred during glacial maxima<sup>24</sup>.

94

95 A total of 26 samples of silts to fine sands (50–250  $\mu\text{m}$ ) were collected from two cores  
96 drilled in the VSR (Fig. 2).  $^{10}\text{Be}$  concentrations in the samples range from  $(0.91 \pm 0.12) \times 10^4$  at  
97  $\text{g}^{-1}$  (19.8 ka, LGM) to  $(5.62 \pm 0.44) \times 10^4$  at  $\text{g}^{-1}$  (73.9 ka, Marine Isotopic Stage (MIS) 5a<sup>34</sup>) (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  $^{10}\text{Be}$  concentrations (see Methods, Supplementary  
101 Section 2). We analyzed 19 new samples (see Methods) to extend the coverage of pre-existing  
102  $\epsilon\text{Nd}$  data from 50–20 ka<sup>24</sup> to 74–50 ka (Fig. 2c). Our new  $\epsilon\text{Nd}$  values range between –10.3 and  
103 –9.6 (average –10.1).

104

### 105 **From $^{10}\text{Be}$ concentrations to paleo-denudation rates**

106 Deriving basin-wide denudation rates from  $^{10}\text{Be}$  concentrations requires computing  
107 the  $^{10}\text{Be}$  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.  $^{10}\text{Be}$   
109 production is also controlled by glacial extent because glaciers shield underlying rocks from  
110 cosmic rays, effectively eliminating  $^{10}\text{Be}$  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 rates<sup>36</sup>. 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<sup>31,37</sup>, 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<sup>26,28</sup> 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-balance<sup>38</sup> (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 ice<sup>36</sup>. 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  $0.15 \pm 0.01$  and  $1.26 \pm 0.16$   
127  $\text{mm a}^{-1}$  (all uncertainties are  $1\sigma$ ), implying  $^{10}\text{Be}$  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%<sup>39</sup>. Between 75 and 29 ka (MIS 5a–  
131 3), both cosmogenic production models yield comparable and rather stable denudation rates  
132 between  $0.15 \pm 0.01$  (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  $0.32 \pm 0.03$  (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  $0.24 \pm 0.04 \text{ mm a}^{-1}$ <sup>27</sup>.

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  $^{10}\text{Be}$  concentration. Additionally, neither  
139 climatic and glacial fluctuations reported in the Alps<sup>40,41</sup> 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  $^{10}\text{Be}$  signal, suggesting that denudation rates are not significantly

142 impacted by these millennial climatic variability<sup>1,33</sup>. 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 rates<sup>24</sup>. Nevertheless,  $\epsilon\text{Nd}$   
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  $^{10}\text{Be}$  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  $^{10}\text{Be}$  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  $^{10}\text{Be}$ -depleted sediments previously stored in the catchment. Such a  
156 mixing mechanism would lower the average  $^{10}\text{Be}$  concentration, thus artificially enhancing  
157 denudation rates. A potential source of  $^{10}\text{Be}$ -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  $\sim 40$  m of sea level drop between the end of MIS 3 and MIS 2<sup>42</sup>. During  
160 warm ice-free periods, such as the Holocene, the  $\epsilon\text{Nd}$  signature of the Var sediment is low ( $-11$   
161 to  $-10.5$ ), indicating a lower contribution of radiogenic highland terrains (Fig. 2)<sup>24</sup>. Because  
162 Pliocene sediments were deposited during a warm and non-glaciated period, they are  
163 probably also characterized by low  $\epsilon\text{Nd}$  value ( $-11$  to  $-10.5$ ). Any reworking of these  
164 sediments during sea level drops should thus be associated with a drop in  $\epsilon\text{Nd}$  values. The  
165 measured  $\epsilon\text{Nd}$  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  $^{10}\text{Be}$ -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 values<sup>27</sup>), 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  $^{10}\text{Be}$  concentrations by 50% (see Supp. Section 4 and Extended Data Fig. 5 and  
172 6). Assuming a simple triangular moraine shape,  $1 \text{ km}^3$  MIS 4–3 moraines would have been  
173  $\sim 130 \text{ km}$  long,  $150 \text{ m}$  wide, and  $100 \text{ m}$  high (see Supp. Section 4). Such large moraines are  
174 unlikely, largest pre-Holocene moraines observable today in the Southern Alps being only  $25 \text{ m}$   
175 high and  $80 \text{ m}$  wide<sup>43</sup>. Another mechanism is therefore required to explain the 50% drop of  
176 cosmogenic  $^{10}\text{Be}$  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 efficiency<sup>44</sup>: by flowing into narrow  
181 valleys, LGM glaciers had higher local velocities than MIS 3-4 glaciers, and, hence, induced  
182 higher glacial incision<sup>44,45</sup> (Fig. 3).

183

184           After a lag time of few thousand years necessary to remove the first meters of the  $^{10}\text{Be}$ -  
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  $^{10}\text{Be}$  concentrations.

188

189           Such denudation solely due to the glacial incision can be quantified using a mass  
190 balance approach, which assumes that the  $^{10}\text{Be}$  concentration measured in each sample is a  
191 mix of glacial and fluvial contributions<sup>46</sup> (see Supp. Section 5). The resulting average glacial

192 erosion rate during the LGM is  $1.5_{-1.0}^{+0.9}$  mm a<sup>-1</sup>, 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 ( $1.8 \text{ mm a}^{-1}$ )<sup>31</sup>, and to the value proposed for present-day small temperate glaciers in the  
196 Swiss Alps ( $1 \text{ mm a}^{-1}$ )<sup>7</sup>. At such incision rates, glacial incision requires 2 to 4 kyr to reach  
197 substratum with null <sup>10</sup>Be concentration. This is consistent with the observation that <sup>10</sup>Be  
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 power<sup>45</sup> (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 <sup>10</sup>Be 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 <sup>10</sup>Be concentrations and increasing the  
208 overall denudation rates of the Var basin (Figs. 2e, 3).

209

210         During deglaciation (19–11 ka)<sup>37</sup>, the overall <sup>10</sup>Be-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  $\epsilon\text{Nd}$ -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)<sup>26</sup>. These changes suggest that denudation dynamics returned to fluvial conditions during  
215 that time<sup>24</sup>. At 10 ka, we observe a spike in denudation rates synchronous with a positive  $\epsilon\text{Nd}$   
216 excursion (Fig. 2e). This short-term <sup>10</sup>Be spike is only supported by one core sample and

217 occurred during a period of increased rainfall in the Mediterranean region<sup>47</sup>. Together with  
218 the  $\epsilon\text{Nd}$  spike, this excursion is characteristic of a paraglacial lag in the sedimentary transport  
219 of previously eroded glacial sediments<sup>48</sup> and should not be considered as an actual increase in  
220 bedrock denudation<sup>24</sup>. 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  $^{10}\text{Be}$ -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<sup>44,45</sup>, the duration of glacial  
230 advances and temperature driven processes (e.g. frost cracking)<sup>49,50</sup>. 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-day<sup>44</sup> and at million years timescale<sup>18</sup>, 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 pulse<sup>3</sup>.

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391

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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  $^{10}\text{Be}$  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)<sup>23</sup>, and  $\epsilon\text{Nd}$  sampling points from Bonneau et al., (2017)<sup>24</sup>.  
423 Sampling sites of cores ESSK08-CS01 and ESSK08-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)<sup>35</sup>.  
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  $^{14}\text{C}$  control points from the studied cores<sup>24,26,28</sup>. Point colours  
430 indicate the core (data in Supp. Table S2). c  $\epsilon\text{ND}$  data from the ESSK08-CS01 and ESSK0-CS13  
431 cores, reflecting sediment provenance. The most recent data (since ~50 ka) are from Bonneau  
432 et al.<sup>24</sup>, and the points labeled "this study" are new data. The orange shaded area corresponds  
433 to  $2\sigma$  uncertainties (data in Supp Table S3). d Mean turbidite fluxes over the two cores<sup>26</sup>. e  
434  $^{10}\text{Be}$ -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 ( $1\sigma$ , y-axis) for each sample. The range of  
437 present-day values is from Mariotti et al.<sup>27</sup>, and the LGM is from Clark et al.<sup>30</sup>. 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 day<sup>27</sup>. For MIS 3–4 and LGM,  $\epsilon_G$  values represent purely glacial denudation (i.e.,  
446 without any contribution from fluvial processes), and were estimated using variable <sup>10</sup>Be  
447 concentrations in glaciated areas (see Methods).  $\epsilon_{mix}$  values are weighted means of the  
448 catchment-wide (glacial and fluvial) denudation rates.  $\epsilon_F$  is the present-day catchment-wide  
449 (fluvial) denudation rate<sup>27</sup>. All uncertainties are  $1\sigma$ . 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.** <sup>10</sup>Be-  
454 derived denudation rates as a function of  $\delta^{18}O$  values measured in the cores, according to the  
455 two scenarii used to convert <sup>10</sup>Be concentrations to denudation rates (see Methods).  
456 Uncertainties on the denudation rates correspond to  $1\sigma$ . Although there is no clear  
457 relationship between the two parameters, higher denudation rates only occur for  $\delta^{18}O$  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 2008<sup>51</sup> (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 <sup>14</sup>C dates, as well as high-  
468 resolution planktonic oxygen isotope stratigraphy<sup>26,28</sup> (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<sup>-1</sup>, 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* <sup>10</sup>Be analyses in quartz<sup>27</sup>. 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 <sup>10</sup>Be concentrations thus reflect  
475 only denudation rate variations. Sample preparation and analytical methods are described in  
476 the Supplementary Information.

477

#### 478 **Computation of <sup>10</sup>Be-derived denudation rates**

479 Basin-averaged denudation rates were calculated using a simplified version of Eq. 1 of  
480 Brown<sup>12</sup>:

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

482 where *i* refers to the different <sup>10</sup>Be production pathways (n for neutron, μs for slow muons,  
483 and μf for fast muons);  $\bar{P}_i$  is the mean basin-wide <sup>10</sup>Be 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;  $\Lambda_i$  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  $\Lambda_{\mu f} = 4,320 \text{ g cm}^{-2}$ )<sup>52</sup>;  $\rho$  is the mean  
487 density of the eroded material (2.7 g cm<sup>-3</sup>); and  $\bar{C}$  is the <sup>10</sup>Be 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 tool<sup>36</sup> and the Lal-Stone scaling model<sup>53</sup>. 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/>)<sup>52,54</sup>. 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 material<sup>55</sup>.

497 Quartz-free areas (based on 1:250,000-scale French Geological Survey geological  
498 maps<sup>56,57</sup>) 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 time<sup>36</sup>. Topographic shielding factors were  
501 calculated using an ArcGIS toolbox<sup>58</sup> 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.<sup>59</sup> for  
505 samples dating to 0–60 ka and data from Valet et al.<sup>60</sup> 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  $\lambda$  is the  
512 neutronic attenuation length ( $g\ cm^{-2}$ ),  $\rho$  the rock density ( $g\ cm^{-3}$ ) and  $\varepsilon$  the denudation rate  
513 ( $cm\ a^{-1}$ )<sup>12,61</sup>. 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* <sup>10</sup>Be 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 <sup>10</sup>Be concentrations at the river outlet<sup>27</sup>. To verify that  
520 submarine processes did not induce a grain size-related bias, here we analyzed <sup>10</sup>Be  
521 concentrations in the same size fractions when the sample size was large enough (17 of 26  
522 samples). <sup>10</sup>Be 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 <sup>10</sup>Be 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 ESK08-CS01 and ESK08-CS13 provide a reliable means to distinguish sediment  
532 contributions from glaciated and non-glaciated areas<sup>24</sup> (Fig. 2c). Here we used the same  
533 methodology<sup>24</sup> to complete the εNd record to the bottom of core ESK08-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 <sup>10</sup>Be 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  $^{10}\text{Be}$  data),  
542 <https://doi.org/10.24396/ORDAR-47> (Table S2 - Core age models),  
543 <https://doi.org/10.24396/ORDAR-48> (Table S3 -  $\epsilon\text{Nd}$  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 <sup>®</sup> code used to determine glacial erosion rates is available upon request. Send an  
550 email to P.-H. Blard: [blard@crpg.cnrs-nancy.fr](mailto:blard@crpg.cnrs-nancy.fr)

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