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1 **Quantifying geomorphological evolution from ^{10}Be denudation rates: insights from high-**
2 **resolution depth profiles, topsoils and stream sediments (Strengbach CZO, France)**

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14 **Keywords** : ^{10}Be isotopes, geomorphological evolution, denudation rates, watershed.

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16 **Keywords :** ^{10}Be isotopes, geomorphological evolution, denudation rates, watershed.

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20 **Abstract**

21 A rare dataset of in-situ ^{10}Be from high-resolution depth profiles, soils, rock outcrops and
22 stream sediments is combined with geochemical analysis and modeling of regolith evolution
23 to understand the variability of denudation rates in a mountain watershed (Strengbach
24 critical zone observatory). High-resolution depth profiles are key to detect the presence of
25 mobile regolith and to highlight how it affects the critical zone evolution. The modeling of
26 regolith evolution and ^{10}Be concentrations along depth profiles allow to estimate both the
27 cosmic ray exposure age (19 kyr) and the mean denudation rate (22 mm.kyr^{-1}) of the regolith
28 without any steady-state assumption on ^{10}Be concentrations. Comparison with maximum
29 denudation rates inferred from topsoil samples collected from the surface of the depth
30 profiles and calculated using the temporal steady-state assumption of ^{10}Be concentrations
31 highlight an overestimation of denudation by a factor of two. Maximum spatially-averaged
32 denudation rates determined from stream sediment samples also likely overestimate
33 denudation rates by a factor of two. These biases are significant for investigating the
34 geomorphological evolution and we propose a method to correct denudation rates using the
35 inherited ^{10}Be concentrations and the cosmic ray exposure age deduced from the high-
36 resolution depth profiles. A key result is also that a steady-state of ^{10}Be concentrations and a
37 steady-state of regolith thickness are two different equilibrium states that do not necessarily
38 coincide. The comparison between locally corrected and spatially-averaged denudation rates
39 indicates that the watershed geomorphology is not in a topographic steady-state but is
40 modulated by regressive fluvial erosion. Nonetheless, our study demonstrates that even in a
41 watershed where the steady-state assumption of ^{10}Be concentrations is not verified, the
42 spatial variations of in-situ ^{10}Be concentrations in sediments still carry qualitatively relevant
43 information on the geomorphological evolution of landscapes.

44 **1. Introduction**

45 Denudation, which is defined as the sum of chemical weathering and physical erosion, is a
46 key process for understanding the Earth's surface evolution. Denudation exhumes fresh
47 bedrock, exerts a strong control on landscape structure, and carries regolith particles from
48 watershed slopes to rivers (Vanacker et al., 2007; Lupker et al., 2012; Schoonejans et al.,
49 2016). Quantifying denudation rates is important for the understanding of regolith residence
50 time (Foster et al., 2015), the sediment transport in rivers (Dongen et al., 2019), the
51 evolution of mountain ranges (Meyer et al., 2010), or the coupling between chemical and
52 physical processes in the critical zone (CZ; West, 2012; Larsen et al., 2014).

53 In-situ terrestrial cosmogenic ^{10}Be is a powerful tool to study the Earth's surface evolution
54 and to gather quantitative estimates on denudation rates (e.g. Dixon and Riebe, 2014). Many
55 studies used in-situ ^{10}Be to estimate watershed-scale averaged denudation rates from
56 stream or river sediments (e.g. Brown et al., 1995; Granger et al., 1996). This widely used
57 approach is based on several important assumptions: (1) the denudation is uniform over
58 time and ^{10}Be concentrations have reached a steady-state, reflecting a balance between ^{10}Be
59 cosmogenic production and ^{10}Be loss by denudation and radioactive decay. (2) each eroding
60 area contributes quartz to the mixed sediment sample in proportion to its erosion rate (valid
61 with a single lithology). (3) if several lithologies are present, the contributing rock types
62 contain similar grain size distributions. (4) quartz is not enriched during weathering or
63 transport. (5) sediment storage is minimal in the watershed. And (6) the denudational time
64 scale is smaller than the timescale for radioactive decay (Von Blanckenburg, 2005; Dunai,
65 2010).

66 In parallel to the watershed-scale averaged denudation rates, soil and regolith denudation
67 rates were also locally determined from in-situ ^{10}Be in topsoil samples (e.g. Riebe et al.,
68 2003; Meyer et al., 2010). However, calculations based on topsoil samples only provide
69 estimates of the minimum exposure age under the assumption of a negligible denudation or
70 estimates of the maximum denudation rate under the assumption of steady-state ^{10}Be
71 concentrations (Braucher et al., 2009). Soil denudation rates are frequently converted into
72 soil or regolith production rates assuming a steady-state of regolith thickness (regolith
73 denudation at the soil level is balanced by the regolith production at depth; e.g. Heimsath et
74 al., 1997).

75 One of the major issues is that the steady-state of denudation and in-situ ^{10}Be
76 concentrations may not be verified for watersheds with a complex evolution including the
77 transport of mobile regolith along slopes or complex exposure histories (assumption 1, Von
78 Blanckenburg, 2005). Many studies taking place in mountain regions with potentially young
79 surfaces and/or mobile regolith do not quantify the impact of this assumption (e.g. Riebe et
80 al., 2003; Meyer et al., 2010), primarily because the analysis of in-situ ^{10}Be in superficial
81 samples cannot independently determine both regolith exposure age and denudation rate
82 (Norton et al., 2010). Another problem is the lack of a clear distinction between three
83 different equilibrium states that do not necessary coincide in time: the steady-state of
84 denudation and ^{10}Be concentrations, the steady-state of regolith thickness, and the steady-
85 state of geomorphology and topography.

86 Sampling of in-situ ^{10}Be depth profiles raises a new research perspective (e.g. Schaller et al.,
87 2004; 2018; Ackerer et al., 2016; Hidy et al., 2018). By gathering regolith and bedrock
88 samples and by combining the contributions from neutrons and muons in modeling

89 approaches, depth profiles opened the possibility of determining both exposure ages and
90 denudation rates (Braucher et al., 2009). Depth profiles also provided new opportunities to
91 investigate the thickness of mobile regolith, the mixing processes near the surface, or the
92 polyphase evolution of weathering profiles (Ackerer et al., 2016; Schaller et al., 2018).

93 We supplement the data obtained by Ackerer et al. (2016) for a high-resolution depth profile
94 located on the summit of a mountain watershed (the Strengbach watershed). Two new high-
95 resolution depth profiles in addition with topsoils, rock outcrops and stream sediments were
96 sampled across the Strengbach watershed to gather a wide database of mineralogical,
97 geochemical and in-situ ^{10}Be analyses. This rare combination allows us to investigate: (a)
98 how the mobile regolith is generated and how it affects the critical zone (CZ) evolution, (b)
99 how the assumption of steady denudation and ^{10}Be concentrations impacts the denudation
100 estimates, (c) how denudation rates vary temporally and spatially in a mountain watershed
101 and (d) what are the timescales to reach steady-state for ^{10}Be concentrations, regolith
102 thickness and geomorphology.

103 The Strengbach watershed is a particularly well suited location to investigate these points as
104 it is a small mountain watershed (0.8 km²) with a complex evolution and a first order stream
105 (Ackerer et al., 2016). The bedrock is relatively homogeneous (Fichter et al., 1998) and
106 alluvial terraces or major zones of sediment storage were not observed in the watershed
107 (Cotel et al., 2016). Quartz enrichment during weathering is not an issue and the
108 denudational time scale is clearly smaller than the timescale for radioactive decay (Ackerer
109 et al., 2016). By consequence, most of the assumptions classically used to interpret ^{10}Be data
110 are validated in this mountain watershed (assumptions 2 to 6; Von Blanckenburg, 2005),
111 with the exception of the critical assumption of balance between the productions and losses

112 of ^{10}Be within the watershed (assumption 1). Our study will thus allow to focus on the
113 specific impact of this major assumption on denudation estimates in a complex watershed.

114 **2. Study site**

115 This study was conducted in the Strengbach watershed (figure 1), one of the reference sites
116 of the French critical zone network (OZCAR), where multidisciplinary studies have been
117 conducted since 1986 (“Observatoire Hydrogéo chimique de l’Environnement”, OHGE;
118 Fichter et al., 1998; Viville et al., 2012; Ackerer et al., 2016; 2020a; Chabaux et al., 2017;
119 2019; Pierret et al., 2018; Wild et al., 2019; Ranchoux et al., 2021). The Strengbach
120 watershed is a headwater watershed (0.8 km²) located in the Vosges Mountains of
121 northeastern France at elevations between 883 and 1147 m (figures 1a and 1b).

122 The current climate is mountainous oceanic, with an annual mean temperature of 7 °C and
123 an annual mean rainfall of approximately 1400 mm. An updated summary of the Strengbach
124 hydrologic and climatic features can be found in Pierret et al. (2018). The bedrock is a
125 Hercynian granite, fractured, hydrothermally overprinted, and mainly composed of quartz
126 (35%), albite (31%), K-feldspar (22%) and biotite (6%). The granite was affected by
127 weathering processes during the Quaternary (Ackerer et al., 2016) and the watershed is
128 currently covered by a 50 to 100 cm-thick soil overlying a 0.5 to 8 m-thick regolith (Weill et
129 al., 2019; Ackerer et al., 2020a). The regolith is defined here by the layer of weathered
130 material that has lost a physical coherence with the bedrock in place. The granitic bedrock is
131 outcropping at few locations, with tors located on the summit of the northern slopes and at
132 mid-elevation on the southern slopes (figure 1).

133 The watershed geomorphology was likely affected by periglacial processes during the
134 Quaternary as the Vosges massif experienced Pleistocene glaciations (Mercier et al., 1999;

135 Heyman et al., 2013). Cold conditions likely persisted at the elevation of the Strengbach
136 watershed after the Late Glacial Maximum (LGM), and the forest cover probably developed
137 only at the beginning of the Holocene (Leroy et al., 2000). The Strengbach watershed was
138 affected by anthropogenic deforestation associated with pastoralism before the 20th century
139 (Etienne et al., 2013), and the return to a dense forest is due to the planting of spruce stands
140 at the beginning of the 20th century (Fichter et al., 1998).

141 **3. Methods**

142 **3.1 Sampling methods**

143 Samples were collected at different locations to explore the variability of denudation rates
144 within the watershed (figure 1b). Two new high-resolution depth profiles extending from the
145 topsoil (0 m) to the granitic bedrock (2-3 m) were realized by digging pits in the slope below
146 the depth profile located on the summit and analyzed by Ackerer et al. (2016). These new
147 depth profiles are located on a gentle slope of the southern part of the watershed where no
148 modern perturbations are visible (figure 1b). Our high-resolution sampling includes: (1) a
149 high number of samples (10 to 15), (2) a detailed density of sampling (one measure every 15-
150 25 cm maximum) to identify specific zones along the vertical direction and (3) a sampling not
151 neglecting the lower part of the depth profiles (the lower regolith and the bedrock), as this
152 part contains important information for the long term muonic exposure and the inherited
153 ¹⁰Be. The three depth profiles are named P1, P2 and P3 (figures 1b and 1c). Observations
154 were performed to study the CZ structure and to estimate soil and regolith thicknesses in
155 the three profiles.

156 In addition, topsoil samples were collected at different places in the watershed, with
157 independent topsoil samples taken on the top of depth profiles, and other samples taken at

158 lower elevations and on the summit of the northern part of the watershed (figure 1b). Two
159 rock samples were also collected on granitic tors outcropping nearby the summit of northern
160 slopes and at mid elevation location on southern slopes (figure 1b). Stream sediment
161 samples were finally collected along the Strengbach stream network before and at the
162 watershed outlet (figure 1b).

163 **3.2 Analytical methods**

164 For samples collected in depth profiles, depth, major element concentrations, mineralogy,
165 density and in-situ ^{10}Be concentrations were characterized (table 1, supplementary tables S1
166 and S2). For topsoil, rock outcrop and stream sediment samples, only density and in-situ ^{10}Be
167 concentrations were characterized (table 2).

168 Bulk densities of regolith samples collected from depth profiles were estimated by
169 combining measurements of fine fraction and block densities (Ackerer et al., 2016). For each
170 sample in the upper part of profiles (0-100 cm), the fine fraction density was obtained by
171 weighing a known volume of regolith sampled with a steel cylinder. The proportion of blocks
172 was estimated via macroscopic description and photos and a constant density of 2.7 g.cm^{-3}
173 was used for the granitic bedrock. The bulk density of each sample was then calculated by
174 combining the fine fraction density and the density of blocks with respect to the proportion
175 of blocks in the horizon.

176 Quartering and sequential crushing provided representative subsamples of regolith bulk
177 samples. These subsamples were used for mineralogical, major elements, and in-situ ^{10}Be
178 analysis following the methodology detailed in Ackerer et al. (2016). Mineralogical analysis
179 was performed by X-ray diffraction, and major element concentrations were determined by
180 tetraborate alkaline fusion followed by ICP-MS and ICP-AES analysis at ITES laboratory

181 (Institut Terre et Environnement de Strasbourg, France; Gangloff et al., 2014; Prunier et al.,
182 2015) and BRGM (Bureau des Recherches Géologiques et Minières, Orléans, France). The
183 analytical accuracy was checked by measuring the San Joaquin soil standard (NIST SRM
184 2709a), and the overall uncertainty was <5 % for major element concentrations.

185 For in-situ ^{10}Be , subsamples consist of approximately 150 g of 250-500 μm sieved grains.
186 Purified quartz was obtained from regolith samples by following the approach involving
187 sequential HCl and HF leaching to remove oxides, organic matter and meteoric beryllium
188 (e.g. Kohl and Nishiizumi, 1992). After addition of a ^9Be carrier, beryllium separation,
189 purification, hydroxylation, and targeting, the $^{10}\text{Be}/^9\text{Be}$ ratios were measured by accelerator
190 mass spectrometry at ASTER (CEREGE laboratory, France). The $^{10}\text{Be}/^9\text{Be}$ ratios of the blanks
191 and samples were approximately 1×10^{-15} and 1×10^{-13} , respectively, and the overall analytical
192 uncertainty for ^{10}Be concentrations was <5 % (table 1).

193 **3.3 Quantification of chemical mobility and volumetric variations in depth profiles**

194 To quantify the mobility of major elements within each depth profile, mass transfer
195 coefficients (τ_i) were calculated as (Brimhall et al., 1991; Riebe et al., 2003):

$$196 \quad \tau_i = \left(\frac{C(i)_{sample} C(j)_{bedrock}}{C(i)_{bedrock} C(j)_{sample}} - 1 \right) \quad (1)$$

197 with $C(i)_{sample}$ and $C(i)_{bedrock}$ the concentrations of the element i for a given sample and
198 for the deepest sample of the depth profile, and $C(j)_{sample}$ and $C(j)_{bedrock}$ the
199 concentrations of the immobile element j for a given sample and for the deepest sample of
200 the depth profile. Positive or negative τ_i imply the gain or loss of the element i relative to
201 the deepest bedrock sample of a profile, taken as the reference rock. Titanium (Ti) was used
202 as an immobile element for the τ_i calculation due to evidence of immobile behavior during

203 weathering in the profile P1 (Ackerer et al., 2016). In addition, volumetric strain index (ε) was
204 calculated to quantify the volumetric variations within each depth profile as (Brimhall et al.,
205 1991):

$$206 \quad \varepsilon = \left(\frac{C(j)_{bedrock} \rho_{bedrock}}{C(j)_{sample} \rho_{sample}} \right) - 1 \quad (2)$$

207

208 with ρ_{sample} and $\rho_{bedrock}$ the bulk density values for a given sample and for the deepest
209 sample of each depth profile, respectively. ε values close to zero indicate isovolumetric
210 weathering, while positive or negative values indicate expansion or compaction.

211 **4. Analytical results**

212 **4.1 Subdivision of depth profiles**

213 Soil thickness is variable and reaches 50 cm, 70 cm and 80 cm in depth profiles P1, P2 and
214 P3, respectively (table 1). Soil type changes from alocrisol in P1 to ochric podzol in P2 and P3,
215 and few weakly weathered blocks of granite are observed within the upper soil in P2 and P3
216 (figure 1c). Lower regolith extends between 50 and 100 cm, 70 and 190 cm, and 80 and 190
217 cm of depth for P1, P2 and P3, respectively (table 1, figure 1). In P1, the amount of blocks
218 gradually increases from top to bottom of the lower regolith, while the proportion of granitic
219 grus decreases (figure 1c). In P2 and P3, the thicker lower regolith has a different texture,
220 with higher amount of clay minerals and some large blocks showing angular and sharp faces
221 with signs of transport. The amount of blocks is also much more variable along the vertical
222 direction in P2 and P3. Granitic bedrock extends below 100 cm for P1 and below 190 cm for
223 P2 and P3, respectively (figure 1c). Bedrock is in place for the three depth profiles and shows
224 weaker traces of weathering. As also detailed in Ackerer et al. (2016), profile P1 has a simple

225 and continuous vertical structure indicating a soil and a regolith developed from in-situ
226 bedrock weathering. P2 and P3 show more complex vertical structures indicative of regolith
227 mobility along the slope.

228 **4.2 Mineralogy, chemical mobility and density in depth profiles**

229 A clear enrichment of interstratified clay minerals is observed in the soil of all depth profiles
230 (supplementary table S1). Minerals resistant to weathering tend to accumulate in the soil
231 (quartz, K-feldspar) while illites, biotite and albite are depleted. Variations from bedrock to
232 regolith are relatively smooth in profile P1 with enrichment of quartz and depletion of most
233 other primary minerals (table S1). In P2 and P3, abrupt depletion of albite and enrichment of
234 illites and micas occur at the transition between bedrock and mobile regolith. Albite is less
235 depleted in the upper part of the mobile regolith (table S1), indicating a non-continuous
236 weathering along the vertical direction in profiles P2 and P3.

237 τ_i values close to zero indicate limited mobility of major elements in the bedrock of the
238 three profiles (figure 2, concentrations in table S2). In P1, τ_i values point to a progressive
239 depletion of major elements from the bottom of the regolith ($\tau_{CaO} = -0.1$) to the topsoil
240 ($\tau_{CaO} = -0.5$, figure 2a). In P2 and P3, the mobile regolith shows stronger variations of τ_i
241 coefficients with marked depletion of sodium and calcium ($\tau_{Na2O} = -0.8$, 130 cm depth, P2,
242 figure 2b; $\tau_{Na2O} = -0.7$, 180 cm depth, P3, figure 2c). τ_i values are less negative in the
243 upper part of the mobile regolith and more negative in the soil of P2 and P3, supporting a
244 non-continuous weathering along the vertical direction in these two profiles (figures 2b and
245 2c). These complex variations are less visible for silica, aluminum and potassium as these
246 elements are controlled by less weatherable minerals (quartz, K-feldspar, ...). Overall, the
247 mobility of major elements follows the structure of depth profiles, with a progressive

248 depletion of major elements in P1 and more complex variations in the mobile regolith of P2
249 and P3.

250 Bulk density is assumed constant in the bedrock of the three profiles ($d=2.7 \text{ g.cm}^{-3}$) before
251 decreasing in the lower regolith ($d=1.8\text{-}2.6 \text{ g.cm}^{-3}$) and in the soil ($d=0.9\text{-}1.8 \text{ g.cm}^{-3}$, table 1
252 and figure 3). Bulk density primarily follows the soil and regolith thicknesses. Volumetric
253 variations are limited in the bedrock and the lower regolith, but expansion occurs in the soil
254 of the three depth profiles ($\epsilon=0.5\text{-}1.4$, figure 3b).

255 **4.3 In-situ ^{10}Be variations in depth profiles**

256 At first order, in-situ ^{10}Be concentrations increase towards the soil surface within the three
257 depth profiles (table 1, figure 3a). ^{10}Be concentrations are similar in the granitic bedrock for
258 all profiles with concentrations around 25000-30000 atoms. g qtz^{-1} . The exponential increase
259 of ^{10}Be concentrations is continuous within the bedrock and the lower regolith of P1, while
260 ^{10}Be concentrations show a linear trend in the soil of P1 (figure 3a). For P2 and P3 profiles,
261 ^{10}Be concentrations show an exponential trend within the mobile regolith, but
262 concentrations are systematically higher than the concentrations measured in the lower
263 regolith of P1. ^{10}Be concentrations are also higher in the soil of P2 and P3 compared to P1,
264 with scattered concentrations in the soil of P3 (figure 3a). Similarly to the profile subdivision
265 and major elements, ^{10}Be concentrations point to a relatively continuous evolution in P1 and
266 more complex variations in P2 and P3. This complexity tends to increase from the summit to
267 the lower part of the hillslope.

268 **4.4 In-situ ^{10}Be variations in topsoils, rock outcrops and stream sediments**

269 In-situ ^{10}Be concentrations measured in topsoil samples tend to decrease with increasing
270 elevation, with concentrations around 180000-190000 atoms. g qtz^{-1} near the main crests

271 and concentrations around 200000-220000 atoms. g qtz⁻¹ at mid-slope (table 2). ¹⁰Be
272 concentrations in rock outcrops are the highest measured in the Strengbach watershed with
273 values of approximately 220000-230000 atoms. g qtz⁻¹ (table 2). For the stream sediment
274 samples, ¹⁰Be concentrations are comparable in the upper part of the stream (around
275 190000 atoms. g qtz⁻¹, table 2). The stream sediment sample collected at the watershed
276 outlet stands out with the lowest ¹⁰Be concentration measured in surface samples across the
277 watershed (table 2).

278 **5. Modeling methods**

279 Mean denudation rate and cosmogenic exposure age of profile P1 were determined from a
280 numerical inverse procedure in Ackerer et al. (2016). Due to the complex structure of the
281 two new depth profiles P2 and P3, an accurate estimation of denudation rates and exposure
282 ages is not possible from a numerical inverse procedure for these profiles. We therefore
283 propose a direct modeling of the evolution of the depth profiles P2 and P3 integrating
284 regolith production, denudation, and the effect of the regolith mobility on ¹⁰Be
285 concentrations. This direct modeling of regolith evolution at the depth profile scale is
286 supplemented by denudation rates derived from topsoil, rock outcrop and stream sediment
287 samples.

288 Our modeling approaches are based on the same general differential equation describing
289 the evolution of ¹⁰Be concentration (C) as a function of depth and time (e.g. Lal, 1991):

$$290 \quad \frac{\partial C}{\partial t} = P \exp\left(\frac{-\rho z}{K}\right) + d \frac{\partial C}{\partial z} - \lambda C \quad (3)$$

291 with P the total surface production rate (in atoms. g qtz⁻¹. yr⁻¹), t the cosmic ray exposure
292 age of the profile (in yr), z the depth (in cm), ρ the integrated density (in g. cm⁻³), K the

293 attenuation length of cosmic ray particles (g. cm^{-2}), d the mean denudation rate of the
 294 surface (in cm. yr^{-1}) and λ the radioactive decay constant of the ^{10}Be (in yr^{-1}). This general
 295 differential equation 3 was adapted for the interpretation of ^{10}Be concentrations from depth
 296 profile, topsoil, rock outcrop and stream sediment samples.

297 **5.1 Modeling of ^{10}Be concentrations in depth profiles**

298 By considering constant values over time of inherited concentration from previous exposure
 299 C_0 and of respective proportions of secondary neutrons, negative muons and fast muons in
 300 ^{10}Be production, the evolution of ^{10}Be concentration (C) as a function of time (t), depth (z)
 301 and integrated density (ρ) for a given denudation rate (d) along a depth profile can be
 302 determined from equation 3 as (adapted from Siame et al., 2004):

$$\begin{aligned}
 303 \quad C(t, z, \rho) = & C_0 \exp(-\lambda t) + \frac{P \cdot P_n}{\frac{\rho d}{K_n} + \lambda} \exp\left(\frac{-\rho z}{K_n}\right) \left(1 - \exp\left(-t\left(\frac{\rho d}{K_n} + \lambda\right)\right)\right) \\
 304 \quad & + \frac{P \cdot P_{nm}}{\frac{\rho d}{K_{nm}} + \lambda} \exp\left(\frac{-\rho z}{K_{nm}}\right) \left(1 - \exp\left(-t\left(\frac{\rho d}{K_{nm}} + \lambda\right)\right)\right) \\
 305 \quad & + \frac{P \cdot P_{fm}}{\frac{\rho d}{K_{fm}} + \lambda} \exp\left(\frac{-\rho z}{K_{fm}}\right) \left(1 - \exp\left(-t\left(\frac{\rho d}{K_{fm}} + \lambda\right)\right)\right) \quad (4)
 \end{aligned}$$

306 with P_n , P_{nm} and P_{fm} the relative contributions of 97.85, 1.5 and 0.65 % to the total
 307 production rate P , and K_n , K_{nm} and K_{fm} the effective attenuation lengths of 150, 1500 and
 308 5300 g.cm^{-2} , for secondary neutrons, negative muons and fast muons, respectively (Braucher
 309 et al., 2009).

310 - Modeling of depth profiles

311 A direct modeling was developed to determine the variations of soil and regolith production
 312 rates, soil and regolith thicknesses, and bulk and integrated densities during depth profile

313 evolution (flow chart in supplementary figure S1). Simulations based on a finite difference
314 scheme were performed for each depth profile along a 1D vertical direction. The CZ vertical
315 structure was subdivided into three parts: the soil, the lower regolith and the bedrock.
316 During simulations, a constant mean denudation rate d was imposed at the soil surface of
317 each profile by following:

$$318 \quad dz_{soil_surface} = d * dt \quad (5)$$

319 with $z_{soil_surface}$ the elevation of the soil surface. In parallel, the soil is formed from lower
320 regolith weathering and the lower regolith is formed from bedrock weathering. Calculation
321 of soil and regolith production rates was based on a law of exponential decline with
322 increasing soil and regolith thicknesses (e.g. Heimsath et al., 2010; Ma et al., 2010):

$$323 \quad F_{sol} = F_0 \exp(-\alpha h_{sol}) \quad (6)$$

$$324 \quad F_{reg} = F_0 \exp(-\beta h_{reg}) \quad (7)$$

325 with h_{sol} and h_{reg} the soil and regolith thicknesses (in cm), F_{sol} and F_{reg} the production
326 rates of soil and regolith (in $\text{g.cm}^{-2}.\text{yr}^{-1}$), F_0 the maximum soil and regolith production rate
327 when soil and regolith thicknesses equal zero (in $\text{g.cm}^{-2}.\text{yr}^{-1}$), and α and β the depth scaling
328 factors of soil and regolith production rates (in cm^{-1}). h_{sol} and h_{reg} were actualized at each
329 new time step during simulations by considering the competition between soil surface
330 lowering due to denudation and soil and regolith production. For the parameter F_0
331 governing the maximum soil and regolith production rate, we used for all profiles a high
332 production rate determined by Egli et al. (2014) on a very thin regolith from a mountain
333 environment (table S3). Scaling factors α and β controlling the exponential decrease of soil
334 and regolith production rates with soil and regolith thicknesses were adjusted for each

335 profile to obtain the best agreement between modeled and observed thicknesses at present
336 time.

337 We assumed that bulk densities of topsoil, soil base and granitic bedrock remained constant
338 through time and we used present day measurements of density for these three interfaces
339 (values in table S3). Bulk density variations throughout the regolith were calculated by a
340 linear interpolation between topsoil, soil base and bedrock densities. The integrated density
341 at a given depth was then calculated by taking the average of bulk densities along the
342 regolith column above the given depth.

343 - Initial state and inherited ^{10}Be

344 Previous modeling of ^{10}Be concentrations in the depth profile P1 located at the summit of
345 the watershed indicated that a major erosion event removing the soil and part of the
346 regolith occurred at approximately 19 kyr BP (Ackerer et al., 2016). This major erosion event
347 removed approximately 2 m of material and was followed by a much more stable exposure
348 period during the Holocene. The inherited ^{10}Be concentrations are derived from the deeper
349 bedrock samples ($\approx 15\,000$ atoms. $\text{g}\ \text{qtz}^{-1}$), that would correspond to a long-term muonic
350 component equivalent to an exposure age of approximately 150 kyr.

351 Based on the results of Ackerer et al. (2016), a relic regolith in place and a bedrock in place
352 were taken as initial state for the direct modeling of P1 at 19 kyr BP (parameters in table S3).

353 Given the proximity of the three profiles, P2 and P3 were also probably affected by the
354 major erosion event at 19 kyr BP. In addition, a mobile regolith within profiles P2 and P3 was
355 highlighted by profile section descriptions and geochemical data (section 4). We therefore
356 used an extremely thin soil after erosion, a mobile regolith stabilized on the slopes, and a
357 bedrock in place as initial state for P2 and P3 at 19 kyr BP (table S3).

358 The inherited ^{10}Be concentration from previous exposure (C_0 in equation 4) was adapted for
359 each depth profile in agreement with their initial states. In the bedrock of all depth profiles,
360 C_0 was set to have an inherited ^{10}Be concentrations from a long-term muonic pre-exposure
361 of 150 kyr (table S3). The same C_0 was used for the lower regolith of the profile P1, as the
362 relic regolith probably experienced the same muonic exposure. By contrast, in the mobile
363 regolith of P2 and P3, C_0 values represent the ^{10}Be concentrations in the regolith transported
364 along the slope and stabilized on P2 and P3 after the important erosive event. C_0 values in
365 the mobile regolith of P2 and P3 are unknown and were adjusted to obtain the best
366 agreement between modeled and measured ^{10}Be concentrations in these profiles.

367 - Final ^{10}Be concentrations

368 The surface total ^{10}Be production rates were determined for the three depth profiles
369 following Martin et al. (2017) (scaling LAL-ERA Musch). After this step, ^{10}Be production rates
370 along depth were calculated for each time step from equation 4 and from the time evolution
371 of the integrated density. Final ^{10}Be concentrations were computed by taking the sum of
372 ^{10}Be produced during simulation time, and exposure age and denudation rate can be
373 estimated by comparing modeled and observed ^{10}Be concentrations. Best fits were
374 identified with the help of Kling-Gupta coefficients (Gupta et al., 2009). Overall, the
375 denudation rate d is inferred from equation 4 with the help of the direct modeling of depth
376 profile evolution that does not rely on a steady-state assumption for the ^{10}Be
377 concentrations.

378 **5.2 Modeling denudation rates from topsoil, rock outcrop and stream sediment samples**

379 In-situ ^{10}Be concentrations are also widely used to estimate local or catchment-wide
380 denudation rates from their analysis in topsoil, rock outcrop or stream sediment samples

381 (e.g. Granger et al., 1996; Meyer et al., 2010; Portenga and Bierman, 2011). By assuming that
 382 there is no inherited ^{10}Be concentration, that ^{10}Be concentration has reached a steady-state,
 383 and that the exposure age is much longer than the effective irradiation time ($t \gg T_{eff}$; with
 384 $T_{eff} = \frac{1}{\lambda + \frac{\rho}{K_n} d_{max}}$; Lal, 1991), the ^{10}Be concentration for a surface sample ($z=0$) can be
 385 expressed as (Charreau et al., 2019):

$$386 \quad C(z_0) = \frac{P \cdot P_n}{\frac{\rho \cdot d_{max}}{K_n} + \lambda} + \frac{P \cdot P_{nm}}{\frac{\rho \cdot d_{max}}{K_{nm}} + \lambda} + \frac{P \cdot P_{fm}}{\frac{\rho \cdot d_{max}}{K_{fm}} + \lambda} \quad (8)$$

387 with d_{max} the maximum denudation rate inferred from the surface sample. A good
 388 knowledge of spatial variations of ^{10}Be production rates is necessary for the calculation of
 389 d_{max} and for the comparison of its value inferred from samples collected at different
 390 locations. The d_{max} estimates were calculated similarly to Mariotti et al. (2019) using the GIS
 391 toolbox Basinga developed by Charreau et al. (2019), where d_{max} is approximated by:

$$392 \quad d_{max} \approx \left(\frac{P \cdot P_n}{C} \right) \frac{K_n}{\rho} + \left(\frac{P \cdot P_{nm}}{C} \right) \frac{K_{nm}}{\rho} + \left(\frac{P \cdot P_{fm}}{C} \right) \frac{K_{fm}}{\rho} \quad (9)$$

393 We used a 5 m resolution digital elevation model (DEM) of the Strengbach watershed from
 394 the French National Geographic Institute (IGN, RGE ALTI® version 2.0) combined with the
 395 QGIS software to determine the geomorphological features (elevation, slope, area, ...) across
 396 the watershed. Secondary neutrons, negative and fast muons were considered and we used
 397 the Lal/Stone time-dependent scaling model (Lal, 1991; Stone, 2000) and revised ^{10}Be
 398 production rates (Martin et al., 2017). Shielding effect is limited in the watershed and was
 399 estimated by following the method from Codilean (2006).

400 For topsoil and rock outcrop samples, local ^{10}Be production rates were calculated at each
 401 location and equation 9 gives the maximum denudation rate at the sample location (referred

402 as local d_{max}). The local d_{max} obtained from a single topsoil sample independently
403 collected on the surface of a depth profile can then be compared to the denudation rate d
404 inferred from equation 4 to quantify the impact of the assumption of steady-state ^{10}Be
405 concentrations.

406 For stream sediment samples, average elevation was used to calculate the average ^{10}Be
407 production rate within each subcatchment. Equation 9 gives the spatially-averaged
408 maximum denudation rate of the catchment above the sampling point (referred as spatially-
409 averaged d_{max}). This calculation is based on steady-state ^{10}Be concentrations at the
410 catchment scale and on the other assumptions stated in the introduction (assumptions 1 to
411 6). Finally, spatially-averaged d_{max} obtained from stream sediments can be used to
412 determine subcatchment maximum denudation rates between two sampling points
413 (Granger et al., 1996):

$$414 \quad d_{max \text{ subcatchment}} = \frac{d_{max,2}A_2 - d_{max,1}A_1}{A_2 - A_1} \quad (10)$$

415 with A_2 and A_1 the areas of downstream and upstream catchments, respectively.

416 One of the important issue for the calculation of d_{max} from equation 9 is the choice of
417 density of the eroding material. Apart for the case of granitic tors, for which the density of
418 2.7 g.cm^{-3} is relatively obvious, several possibilities arise for topsoil and stream sediment
419 samples. One possibility is to use the current average densities of topsoil ($\approx 1 \text{ g.cm}^{-3}$, as in
420 Meyer et al., 2010) and of stream sediment samples ($\approx 2.2 \text{ g.cm}^{-3}$) for the calculation of local
421 and spatially-averaged d_{max} . Another approach is to use the current average density of
422 regolith ($\approx 1.7 \text{ g.cm}^{-3}$) for both cases, arguing that this density could be a better estimate for
423 the available erodible materials at the surface. However, it must be recalled that the

424 assumption of steady-state ^{10}Be concentrations implies a long time of cosmic ray exposure
425 ($t \gg T_{eff}$) and this point must be kept in mind for the choice of densities. If soils are young
426 and thin, and if most of the ^{10}Be accumulation occurs in the deeper regolith or in the
427 bedrock, using lower density values measured from topsoils is biasing the denudation rates
428 inferred from equation 9 toward higher values. A strategy more coherent with this
429 assumption is to use the bedrock density (2.7 g.cm^{-3}) for all calculations. This reflects that at
430 longer timescales, all the surface processes (soil development during interglacial and soil
431 scraping during glacial times) are averaged out and what is eroding is in effect the underlying
432 bedrock. We quantitatively evaluate these three approaches below.

433 **6. Modeling results**

434 **6.1 Evolution, exposure age and denudation rates of depth profiles**

435 Modeled soil production rates are variable between profiles and tend to increase with
436 decreasing elevation (from P1 to P2 and P3, adjusted parameters in table S3). By contrast,
437 regolith production rates are much lower in profiles P2 and P3 due to the thicker regolith
438 already present at the initial state (table S3). Modeled soil thicknesses rapidly increase
439 during the first 5 kyr before reaching more stable values after 10 kyr of simulation time
440 (illustration for profile P1 in figure 4). Soil thicknesses at present time reach 50 cm for P1 and
441 75 cm for P2 and P3 in accordance with measurements. Final simulated bulk and integrated
442 densities are also consistent with measurements from all depth profiles (figures 4a, 4b, 5a,
443 5b, 6a and 6b).

444 The best agreement between modeled and measured ^{10}Be concentrations are obtained with
445 the same cosmic ray exposure age of 19 kyr, the same muonic pre-exposure age of 150 kyr in
446 the bedrock, and the same denudation rate d of 22 mm.kyr^{-1} for the three depth profiles.

447 The ^{10}Be concentrations inherited from muons are presented in figure 4d. The best fit of
448 data was characterized by Kling-Gupta coefficients of 0.96, 0.95 and 0.82 for P1, P2 and P3
449 profiles, respectively (Gupta et al., 2009). In profile P1, ^{10}Be concentrations in the bedrock
450 and the lower regolith are well explained by the direct modeling (figure 5c). The ^{10}Be
451 concentrations are well explained in the whole profile P1 after considering a recent soil
452 expansion constrained by the volumetric strain index (by expanding the soil layers towards
453 the surface; as in Ackerer et al., 2016; figure 5d). In P2 and P3 bedrock, ^{10}Be concentrations
454 are well explained by the direct modeling, similarly to the P1 bedrock (figures 6c and 6d).
455 Accurate fitting of ^{10}Be concentrations in P2 and P3 regolith is more challenging due to
456 impact of mobile regolith within these profiles. Higher ^{10}Be concentrations in the mobile
457 regolith of P2 and P3 are explained by the inherited ^{10}Be from mobile regolith pre-exposure
458 (C_0 in table S3). Best agreements between modeled and measured ^{10}Be concentrations in P2
459 lower regolith are obtained with a full mixing of mobile regolith during the transport event
460 that occurred before 19 kyr BP, implying a constant C_0 value over depth in P2 lower regolith
461 (table S3, figure 6c). Best results in P3 lower regolith are obtained with a limited mixing of
462 mobile regolith and a variable C_0 value over depth (table S3, figure 6d). Scattered ^{10}Be
463 concentrations, probably due to surface mixing in P3 soil, is not described by our direct
464 modeling approach (figure 6d).

465 **6.2 Maximum denudation rates from topsoils, rock outcrops and stream sediments**

466 Maximum denudation rates inferred from equation 9 are shown in figure 7 for topsoils, rock
467 outcrops and stream sediments. For this calculation, we used a constant granite density of
468 2.7 g.cm^{-3} for all samples. Calculated spatially-averaged d_{max} are similar for all the stream
469 sediments ($36\text{-}40 \text{ mm.kyr}^{-1}$), apart from the sediments collected at the outlet that indicate

470 higher denudation rate (43 mm.kyr⁻¹, table 2, figure 7a). The highest spatially-averaged d_{max}
471 (56 mm.kyr⁻¹) is calculated for the closest subcatchment to the watershed outlet (equation
472 10, figure 7a). Local d_{max} for rock outcrops are lower (30-34 mm.kyr⁻¹), while local d_{max} for
473 soils range between 30 and 42 mm.kyr⁻¹ with higher denudation near the crests (figure 7a).
474 Local d_{max} calculated for topsoil samples independently collected at the surface of depth
475 profiles (34-42 mm.kyr⁻¹) are approximately two times higher than the denudation rate d
476 determined from the direct modeling of the high-resolution depth profiles (22 mm.kyr⁻¹).
477 Other choices of density for topsoil and stream sediment samples (i.e. section 5.2) increase
478 the biases observed between the local d_{max} determined on depth profiles and the
479 denudation rate d determined on the same depth profiles from the direct modeling (from a
480 factor three to four).

481 We find no relation between catchment mean slope orientations and denudation rates
482 (figure 7b, data in table S4), suggesting a limited role of differences in freezing-thaw cycles
483 between north facing and south facing slopes. There is also no clear relation between
484 catchment mean slopes and denudation rates (figure 7c, table S4): similar d_{max} are obtained
485 for catchments with moderate mean slopes (12-18°) and only above a threshold of around
486 20° higher denudation is observed for the outlet subcatchment (figure 7b and table S4).

487 **7. Discussion**

488 **7.1 Origin of mobile regolith**

489 Our study on the Strengbach watershed indicates that two different types of mobile regolith
490 can occur along a slope: (1) a deep mobile regolith able to transport the soil and the lower
491 regolith and (2) a near surface mobile regolith generating an expansion or a mixing of soil
492 layers.

493 For the deep mobile regolith, modeling results obtained for the three depth profiles indicate
494 that regolith mobilization occurred around the last glacial maximum (LGM, around 24-19
495 kyr). The homogenization of mobile regolith was total (in P2) or partial (in P3) during
496 transport, and stabilization occurred along the slope around 19 kyr BP. Given the moderate
497 slopes around the three depth profiles (10-12°), the probable lack of vegetation and the
498 prevalence of cold conditions at the elevation of the Strengbach watershed during the LGM
499 (Leroy et al., 2000; Heyman et al., 2013), the transport of deep mobile regolith is probably
500 explained by periglacial solifluction and cryoturbation processes. Studies of periglacial
501 processes in present day cold environments indicate that rock debris can be transported at
502 velocities up to 10-50 cm.yr⁻¹ (Kääb and Vollmer, 2000; Wangensteen et al., 2006). With a
503 similar downward velocity during the LGM, the distance of 200 m between P1, P2 and P3
504 profiles would be covered by rock debris in only 400-2000 years. These estimates indicate
505 that the transport of mobile regolith in a periglacial context can be fast compared to the ¹⁰Be
506 production timescale, explaining why ¹⁰Be concentrations within the three profiles are well
507 modeled by the same exposure age of 19 kyr.

508 Study of creeping velocity for regolith particles in a warmer and drier environment was
509 conducted by Heimsath et al. (2002). Much slower velocity was inferred for the downward
510 migration of regolith particles by processes analogous to diffusive erosion (6 et 12 m.kyr⁻¹;
511 Heimsath et al., 2002). By taking a mean velocity of 10 m.kyr⁻¹, 19 kyr would be needed for
512 regolith particles to cover the distance of 200 m between P1, P2 and P3 profiles. This time is
513 too long and incompatible with the similar ¹⁰Be concentrations observed in the upper part of
514 all profiles (table 1) and with the modeling results pointing to a similar exposure age across
515 the slope (table S3). Therefore, the deep mobile regolith observed in the watershed is most
516 likely explained by rapid peri-glacial processes occurring during the LGM (episodic erosion;

517 Norton et al., 2010), and not by continuous creeping as suggested in other mountain
518 catchments (West et al., 2013).

519 In contrast, the near surface (< 1 m of depth) regolith mobility is probably a recent
520 mechanism (Holocene to actual). A longer cosmic ray exposure of a stable profile would
521 inexorably produce a simple exponential decrease of ^{10}Be concentrations with depth, which
522 is not observed in P2 and P3. Given the moderate slopes and the dense forest cover
523 currently present around the depth profiles, recent erosion processes are probably
524 dominated by diffusive erosion (Montgomery and Brandon, 2002) and soil layers are likely
525 impacted by biological processes (Wilkinson et al., 2009). In P1, the linear trend of ^{10}Be
526 concentrations observed in the soil is explained by the volumetric expansion due to the
527 incorporation of organic matter (Ackerer et al., 2016). In P2 and P3, the soil mixing appears
528 incomplete, as ^{10}Be concentrations are not homogenized over depth as illustrated by Dixon
529 and Riebe (2014). This soil mixing is likely related to recent diffusive erosion, bioturbation by
530 living organisms, tree throw, or anthropogenic perturbations during the last centuries
531 (Etienne et al., 2013).

532 The differences between superficial and deep mobile regolith are rarely discussed in the
533 literature (e.g. Dixon and Riebe., 2014; Schaller et al., 2018). One reason is that most of
534 studies investigated shallower depth profiles or that sampling resolution in the deeper part
535 of depth profiles is too low to highlight complex ^{10}Be concentration trends and the presence
536 of deep mobile regolith. We therefore emphasize the importance of a high-resolution
537 sampling in the lower part of depth profiles to identify deep mobile regolith and to better
538 constrain the long-term evolution of the regolith along hillslopes.

539 **7.2 Role of mobile regolith in CZ evolution**

540 Our data and modeling results indicate that the mobile regolith has a key role in the CZ
541 evolution as it modifies and complexifies the CZ structure (summary figure 8). Firstly, the
542 occurrence of deep mobile regolith implies a disconnection between the bedrock in place
543 and the regolith. This induces a greater heterogeneity, explaining the sharper variations of
544 mineralogical and chemical composition at the bedrock/regolith interface within P2 and P3
545 profiles (figure 2). The deep mobile regolith also modifies the inherited ^{10}Be concentrations
546 (figures 3 and 8), increasing ^{10}Be concentrations in the regolith of profiles located downslope
547 (P2 and P3).

548 The deep mobile regolith also impacts soil and regolith production rates. Higher soil
549 formation rates from lower regolith weathering are modeled in P2 and P3 profiles (section
550 5.2), and thicker and more evolved soils (ochric podzols) are observed within profiles
551 developed on a deep mobile regolith (figures 1 and 8). These points are potentially explained
552 by a faster weathering of the transported material found in the mobile regolith compared to
553 the direct weathering of a lower regolith in place. By contrast, modeled values of regolith
554 production rates from bedrock weathering are faster at the summit (P1) than at lower
555 elevation along the hillslope (P2 and P3).

556 Ma et al. (2010) focused on a watershed affected by periglacial processes and based on U-
557 series isotopes also concluded that regolith production rates decrease with increasing
558 regolith thickness and distance from the crests. Our results support this scheme and show
559 that one possible explanation of this functioning is the deposition of mobile regolith along
560 slopes. In fact, this process is explaining both the thickening of regolith at lower elevation
561 and the protection of bedrock from weathering, which reduces regolith production rate.

562 The deep mobile regolith has also implications on the hydrological processes and water
563 storage across landscape. It induces a thickening of weathered material in the lower parts of
564 the watershed, which tends to enhance water storage capacity. Higher water storage
565 capacity at mid-slope locations compared to crests was identified as an important feature to
566 correctly model the hydrogeochemical functioning of the watershed (Weill et al., 2019;
567 Ackerer et al., 2020a).

568 Many studies investigating the CZ evolution pictured that the mobile regolith is relatively
569 superficial, affecting soils, and implying an almost complete mixing of soil layers (e.g.
570 Anderson et al., 2007; Dixon and Von Blanckenburg, 2012; Riebe et al., 2017; Schaller et al.,
571 2018). If a near surface mobile regolith is typically identified, the deep mobile regolith
572 extending below the soil is too often neglected to understand the CZ structure and its
573 evolution. As shown by the detailed modeling of depth profiles, the deep mobile regolith is
574 not always well mixed, suggesting that a depth dependent transport law as proposed by
575 Heimsath et al. (2005) is also valid in the context of shallow gradient slopes of watersheds
576 affected by periglacial processes. The mobile regolith may also change the relationship
577 between physical erosion and chemical denudation rates across landscape, by complexifying
578 the idealized soil mantled hillslope where soil is directly produced from bedrock weathering
579 (e.g. Ferrier and Kirchner; 2008).

580 **7.3 Impact of the assumption of steady-state ^{10}Be concentrations on denudation rates**

581 The results show that the denudation rate d (22 mm.kyr⁻¹) determined from the direct
582 modeling of depth profiles without any steady-state assumption on ^{10}Be concentrations is
583 approximately two times lower than the local d_{max} (34-42 mm.kyr⁻¹, figure 7a) calculated on
584 the same profiles with independent topsoil samples. This bias rises to a factor of three or

585 even four when considering current average regolith or topsoil densities ($\approx 1.7 \text{ g.cm}^{-3}$ and ≈ 1
586 g.cm^{-3} respectively) instead of the bedrock density (2.7 g.cm^{-3}) in the calculation of local
587 d_{max} .

588 This bias is explained by the fact that in the Strengbach catchment, the ^{10}Be concentrations
589 are currently not at steady-state, i.e. production, denudation and radioactive decay are not
590 balanced. Additional simulations were performed on the profiles for the next 5,000 or
591 10,000 years, indicating a significant increase in ^{10}Be concentrations in the soil, regolith and
592 bedrock for all depth profiles. Such behavior is related to the short cosmic ray exposure age
593 of the regolith. Calculation of the effective irradiation time (T_{eff}) indicates that the exposure
594 age should be longer than 40-50 kyr to validate the assumption of steady-state ^{10}Be
595 concentrations, which is not the case given the important erosion event that affected the
596 profiles at around 19 kyr BP (i.e. $t < T_{eff}$).

597 A key finding is that the depth profiles, although not in steady-state for the ^{10}Be
598 concentrations, are close to a steady-state for their soil and regolith thicknesses (figure 4).
599 The simulated regolith production rates are close to the simulated denudation rates within
600 the Holocene timeframe. This conclusion is supported by the observation that our simulated
601 regolith production rate in P1 (approximately 14 mm.kyr^{-1} over the last 15 kyr, table S3) is
602 consistent with the regolith production rate independently determined on the millennial
603 timescale from U-Th-Ra isotopes (approximately 12 mm.kyr^{-1} , Ackerer et al., 2016).

604 We also highlight the importance of considering soil expansion (non-isovolumetric
605 weathering) when deciphering the evolution of depth profiles. In P1, a steady-state of
606 regolith thickness and mass balance has been reached with a regolith production rate of 12
607 mm.kyr^{-1} and a denudation rate of 22 mm.kyr^{-1} : the weathering of a given thickness of

608 bedrock produces approximately twice more thickness of soil due to the volumetric
609 expansion and the decrease in density. When considering the differences in density and
610 expansion processes, regolith production and denudation rates expressed in specific mass
611 fluxes are in fact similar in P1 (30 ± 10 and 37 ± 15 T.km⁻².yr⁻¹, respectively; Ackerer et al.,
612 2016).

613 We therefore advocate for a clear distinction between two different equilibrium states: a
614 steady state of ¹⁰Be concentrations achieved when ¹⁰Be production is balanced by
615 denudation and radioactive decay, and a steady-state of regolith thickness (or mass balance)
616 reached when regolith production is balanced by total denudation. Our results demonstrate
617 that these two steady-states are not necessarily reached over the same timescale. In the
618 case of the Strengbach watershed and nearby the crests, the steady-state of regolith
619 thickness is reached before that of the ¹⁰Be concentrations.

620 One way to correct the bias induced by the assumption of steady-state ¹⁰Be concentrations
621 on the local d_{max} is to rather estimate denudation rates d from equation 4, at $z=0$, using the
622 inherited ¹⁰Be concentrations and the exposure age both deduced from the modeling of
623 high-resolution depth profiles. The determination of these corrected denudation rates d can
624 be done with a numerical inverse procedure (Ackerer et al., 2016), by minimizing the
625 difference between ¹⁰Be concentrations measured in samples and ¹⁰Be concentrations
626 calculated from equation 4.

627 On our depth profiles, the corrected local d from topsoil samples indicate that denudation
628 rates range between 33 mm.kyr⁻¹ (in P1) and 14 mm.kyr⁻¹ (in P2), much more consistent with
629 the value of 22 mm.kyr⁻¹ estimated by the direct modeling of depth profiles (table 2). We
630 used for this calculation the average regolith density of 1.7 g.cm⁻³, as this density is

631 representative of the eroded regolith during the time-window from 19 kyr BP to present.
632 The scattering of the revised denudations is certainly affected by the mixing processes of the
633 shallow samples. Even though the corrected denudation values are only calculated for 3
634 samples, it is interesting to note that the average of these three d values is 22 mm.kyr^{-1} , i.e.
635 identical within error to the value derived from the modelling of depth profiles. This result is
636 consistent with the classical approach that recommends averaging the denudation rates of
637 several samples for a given area, rather than relying on a single sample estimate.

638 The bias induced by the classical assumption of steady-state ^{10}Be concentrations (an
639 overestimation of local denudation rates by a factor two to four) has probably a limited
640 consequence for studies dealing with first-order estimates of denudation rates and/or
641 comparing several watersheds with very different denudation rates. However, this bias is
642 significant for studies attempting to better understand the geomorphological evolution of a
643 watershed, as for example studies comparing local denudation of ridge crests with
644 watershed-scale averaged denudation (e.g. Meyer et al., 2010). Our study shows that the
645 assumption of steady-state ^{10}Be concentrations should be used very carefully in mountain
646 environments where regoliths with short cosmic ray exposure ages are present. Combining
647 high-resolution depth profiles with local regolith and topsoil samples is a way to quantify the
648 impact of this assumption on denudation estimates.

649 **7.4 Spatiotemporal variability of denudation rates and geomorphological implications**

650 The temporal variability of denudation can firstly be estimated from the direct modeling of
651 depth profiles. Denudation rates likely strongly decreased between the LGM and the last 19
652 kyr, as indicated by the cosmic ray exposure age of depth profiles pointing to important
653 erosion events before 19 kyr BP and a period of relative stability after 19 kyr BP (figure 8). A

654 second comparison can be made from the modern measurements of denudation. The
655 research conducted in the Strengbach watershed shows that the current denudation
656 determined from solute and solid fluxes at the outlet (around $15 \text{ T.km}^{-2}.\text{yr}^{-1}$, with $5 \text{ T.km}^{-2}.\text{yr}^{-1}$
657 1 of solute exports and $10 \text{ T.km}^{-2}.\text{yr}^{-1}$ of solid exports over 20 yr of record; Viville et al., 2012;
658 Cotel et al., 2016; Ackerer et al., 2020) is lower than the long-term denudation inferred from
659 ^{10}Be data (around $37 \text{ T.km}^{-2}.\text{yr}^{-1}$, determined on the last 19 kyr on P1; Ackerer et al., 2016). A
660 general trend of denudation decrease with time is then highlighted from the LGM to the last
661 19 kyr, and from this 19 kyr time-window to present. A probable explanation is the
662 progressive development of vegetation during the Holocene at this elevation (Leroy et al.,
663 2000), and the dense forest cover observed today in the watershed (Pierret et al., 2018).
664 Another explanation which cannot be excluded is that modern measurements were not
665 affected by rare but high magnitude erosion events, as suggested by Kirchner et al. (2001).

666 The spatial variability of denudation can be discussed by the comparison of denudations
667 rates determined from depth profile, topsoil, rock outcrop and stream sediment samples. As
668 discussed in section 7.3, the assumption of steady-state ^{10}Be concentrations is not applicable
669 for these sites. This suggests that the same assumption is probably too strong for the
670 determination of denudation rates from other soil, outcrop, or sediment samples in the
671 watershed and that the determined d_{max} values are biased. The approach to correct these
672 estimates developed in section 7.3 has been applied for these different samples to estimate
673 more realistic denudation rates from ^{10}Be concentrations. For all the Strengbach samples, a
674 corrected denudation rate d has been determined by numerical inverse procedure from
675 equation 4, with the assumption that the entire watershed is characterized by the same
676 exposure age and the same inherited ^{10}Be concentration as established on depth profiles.

677 Comparison of corrected denudation rates inferred from topsoil, rock outcrop and stream
678 sediment samples allows for a detailed understanding of the evolution of the watershed
679 geomorphology (figure 9, table 2). On the upper part of the watershed, lower denudation
680 rates on rock outcrops compared to soil and sediment denudation rates indicate an ongoing
681 individualization and growth of bare rocks. The observation that rock outcrops have lower
682 denudation rates than other parts of the basin is classically invoked to explain the
683 prominence of these structures in the landscape (Riebe et al., 2003; Heimsath et al., 2010).
684 With denudation rates of around 22 mm.kyr^{-1} for the regolith and 4 to 9 mm.kyr^{-1} for the
685 tors, in 19 kyr the topographic difference between the soil surface and tors would be a few
686 tens of cm, consistent with field observations for the STR 11-1 site. However, these values
687 are far too low to explain the morphologies observed at site STR11-5 where the bedrock
688 outcrop surface overhangs the surrounding soils by nearly 2 m. With the estimated
689 denudation rates presented above, such topographic differences would require nearly 150
690 kyr to be obtained, i.e. the time of the penultimate strong denudation event of the site
691 based on the interpretations of Ackerer et al. (2016). We therefore propose that bedrock
692 outcrops in these areas result from their exposure during stripping events of locally resistant
693 bedrock portions, accounting for denudation at lower rates in these areas compared to
694 other rocks. This interpretation remains to be validated by detailed studies of these tors,
695 which is beyond the scope of this study.

696 It can also be noted that in the upper part of the watershed, the highest local denudation
697 rates are observed on the crests ($33\text{-}36 \text{ mm.kyr}^{-1}$), with values significantly higher than the
698 local rate determined above the main catchment knickpoint (i.e. STR11-7; 12 mm.kyr^{-1} ;
699 figure 9). By comparing corrected spatially-averaged d from sediments from the upper part
700 of the stream with corrected local d from soils on crests (as in Meyer et al., 2010), the relief

701 decrease between the crests and above the river knickpoint would be around 15 mm.kyr⁻¹.
702 However, it must be noted that this relief variation is close to uncertainties, if we consider
703 the substantial variability of local denudation rates determined near the crest at the surface
704 of depth profiles (33-14 mm.kyr⁻¹; figure 9). Therefore, except the significant topographic
705 growth of local granitic outcrops, the upper part of the watershed is likely affected by a
706 limited relief variation.

707 This is clearly in contrast with the variability of denudation rates at the scale of the entire
708 watershed. The lower part of the watershed stands out from the rest of the watershed, as
709 the highest denudation rate in the whole watershed is undeniably determined for the outlet
710 subcatchment (99 mm.kyr⁻¹, figure 9). With the above assumptions (i.e. same exposure age
711 and inherited ¹⁰Be for the entire watershed), the high denudation rate of the outlet
712 subcatchment compared to the ones determined from the sediments collected in the upper
713 part of the stream (23-18 mm.kyr⁻¹) implies that a regressive erosion from the bottom of the
714 valley steepens the topography nearby the stream outlet. By comparing corrected spatially-
715 averaged d in the outlet subcatchment with corrected local d from soils on crests, the total
716 watershed relief increase would be around 65 mm.kyr⁻¹, following the approach given in
717 Meyer et al. (2010). From the above discussion, it can also be proposed that the Strengbach
718 watershed is subdivided into two different geomorphological systems, with an upper part
719 with a nearly constant relief and a lower part characterized by a relief increase localized near
720 the stream incision.

721 Different quantitative estimates would be determined for the geomorphological evolution of
722 the watershed from denudation rates uncorrected or differently corrected. The lowest relief
723 change between crests and outlet would be obtained with uncorrected denudation rates

724 ($\approx 20 \text{ mm.kyr}^{-1}$, figures 7). But from a qualitative point of view this would not change the
725 conclusion that the Strengbach watershed is marked by a noticeable spatial variability of its
726 denudation with an upstream subcatchment with limited relief change and a downstream
727 one with a localized regressive denudation near the stream outlet. The Strengbach case
728 therefore illustrates that even in a watershed where the steady-state assumption of ^{10}Be
729 concentrations is not verified ($t < T_{eff}$), the spatial variations of in-situ ^{10}Be concentrations
730 in sediments still carry qualitatively relevant information on the geomorphological evolution
731 of landscapes.

732 **8. Conclusion**

733 We presented in this study a rare combination of in-situ ^{10}Be data from high resolution
734 depth profiles, soils, rock outcrops and stream sediments in a small mountain watershed.
735 The high resolution of depth profile sampling was a key point to understand how the mobile
736 regolith is generated and how it affects the CZ evolution (summary figure 8). Two different
737 types of mobile regolith were identified: a deep mobile regolith explained by periglacial
738 processes exclusively occurring during the LGM and a superficial mobile regolith generating a
739 recent expansion or mixing of soil layers. Deep mobile regolith clearly modifies ^{10}Be
740 concentrations and mineralogical, geochemical and physical evolution of the CZ. The
741 modeling of regolith evolution and of ^{10}Be concentrations along depth allows to estimate the
742 cosmic ray exposure age (19 kyr) and the mean denudation rate (22 mm.kyr^{-1}) of high-
743 resolution depth profiles without any assumption of steady-state ^{10}Be concentrations.
744 Comparison with maximum denudation rates inferred from topsoil samples independently
745 collected at the surface of depth profiles and interpreted under the assumption of steady-
746 state ^{10}Be concentrations shows that this assumption leads to an overestimation of local

747 denudation rates by a factor of approximately two. One important conclusion of our study is
748 to show that if one wants to estimate denudation rates by following the classical approach
749 (i.e. single topsoil sample, assumption of steady-state ^{10}Be concentrations), the best
750 estimates are obtained when using the density of the bedrock. Results also indicate that
751 maximum spatially-averaged denudation rates determined from stream sediment samples
752 overestimate denudation rates by a factor of approximately two. These biases are significant
753 to understand the precise geomorphological evolution of a watershed, and we propose a
754 method to estimate corrected denudation rates. The comparison between corrected local
755 and spatially-averaged denudation rates indicates that the watershed geomorphology is
756 evolving and that regressive erosion increases the relief nearby the outlet. A key finding is
757 also that a steady-state of ^{10}Be concentrations and a steady-state of regolith thickness (or
758 mass balance) are two different situations that do not necessarily overlap. In the case of the
759 Strengbach watershed crests, a steady-state of ^{10}Be concentrations is not achieved while
760 regolith thickness tends to a steady-state. However, even in a watershed where the steady-
761 state assumption of ^{10}Be concentrations is not verified ($t < T_{eff}$), the spatial variations of
762 in-situ ^{10}Be concentrations in sediments still provide relevant information on the
763 geomorphological evolution of landscapes.

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Figure caption

950 Figure 1: study watershed and sampling location. (a) Strengbach watershed location in northeastern
951 France. (b) Sampling location of ^{10}Be samples including topsoil, rock outcrop, stream sediment and
952 high-resolution depth profile samples. (c) Photography of the three high-resolution depth profiles
953 realized on the southern part of the watershed. Yellow rectangles represent collected samples of soil,
954 regolith or bedrock. Blue diamonds represent samples for which mineralogical, geochemical and ^{10}Be
955 analysis were performed. The profile P1 was investigated by Ackerer et al. (2016). Mor litter is
956 defined by a thick mat of undecomposed to partially decomposed litter that is not significantly
957 incorporated into the soil, present in coniferous forests. Moder litter is defined by an undecomposed
958 and partially decomposed remains of broad-leaved deciduous forest litter that is shallowly
959 incorporated into the soil.

960 Figure 2: chemical mobility coefficients ($\tau_i = \left(\frac{C(i)_{\text{sample}} C(j)_{\text{bedrock}}}{C(i)_{\text{bedrock}} C(j)_{\text{sample}}} - 1 \right)$) for the three high-resolution
961 depth profiles on the southern slope of the watershed. (a) Subplot for the depth profile P1. (b)
962 Subplot for the depth profile P2. (c) Subplot for the depth profile P3.

963 Figure 3: ^{10}Be data, volumetric variations, and bulk density from the three high-resolution depth
964 profiles (P1, P2 and P3). (a) ^{10}Be concentrations measured in bulk samples. (b) Volumetric strain index
965 ($\varepsilon = \left(\frac{C(j)_{\text{bedrock}} \rho_{\text{bedrock}}}{C(j)_{\text{sample}} \rho_{\text{sample}}} \right) - 1$). (c) Bulk density.

966 Figure 4: direct modeling of the evolution of P1 profile. Black lines represent from top to bottom, soil
967 surface, soil base and bedrock interfaces. (a) Bulk density evolution. (b) Integrated density evolution.
968 (c) ^{10}Be production rate evolution. (d) ^{10}Be concentrations from muonic pre-exposure (150 kyr).
969 Calculations of the muonic pre-exposure consider the erosion of regolith over time, and especially
970 the important erosion event that removed approximately 2 m of regolith during the Last Glacial
971 Maximum (Ackerer et al., 2016).

972 Figure 5: modeling results for the P1 profile. (a) Measured and modeled bulk density. (b) Measured
973 and modeled integrated density. (c) Comparison between measured and simulated ^{10}Be
974 concentrations. (d) Comparison between measured and simulated ^{10}Be concentrations including a
975 correction of soil expansion.

976 Figure 6: modeling results for P2 and P3 profiles. (a) Measured and modeled integrated density for
977 the profile 1. (b) Measured and modeled integrated density for the P2 profile. (c) Comparison
978 between measured and simulated ^{10}Be concentrations for the P2 profile. (d) Comparison between
979 measured and simulated ^{10}Be concentrations for the P3 profile.

980 Figure 7: slopes, catchments of stream sediments and spatial variability of denudation rates in the
981 Strengbach watershed. (a) Map of the watershed. Maximal denudation rates (d_{max}) written in black
982 are inferred from ^{10}Be concentrations measured in topsoil, rock outcrop and stream sediment
983 samples with the GIS toolbox Basinga (Charreau et al., 2019) using equation 9 ($d_{\text{max}} \approx \left(\frac{P \cdot P_n}{C} \right) \frac{K_n}{\rho} +$

984 $\left(\frac{P.P_{nm}}{c}\right)\frac{K_{nm}}{\rho} + \left(\frac{P.P_{fm}}{c}\right)\frac{K_{fm}}{\rho}$) and under the assumption of steady state ^{10}Be concentrations. A constant
 985 bedrock density of 2.7 g.cm^{-3} was used in all calculations. The denudation rate written in blue is
 986 determined through the modeling of high-resolution depth profiles without the assumption of steady
 987 state ^{10}Be concentrations (d , equation 4). The denudation rate written in bold black for the outlet
 988 subcatchment is calculated from equation 10. Denudation rates are expressed in mm.kyr^{-1} and
 989 uncertainties include the uncertainties on the ^{10}Be concentrations, blank correction, and ^{10}Be
 990 production rates (uncertainties in table 2). (b) Spatially-averaged d_{max} calculated for stream
 991 sediment catchments as a function of mean catchment aspect ($^{\circ}$ from N). (c) Spatially-averaged d_{max}
 992 calculated for stream sediment catchments as a function of mean catchment slope ($^{\circ}$). Error bars
 993 represent one standard deviation of data.

994 Figure 8: summary diagram of the four major stages of the CZ evolution along the southern slope of
 995 the Strengbach watershed. Stage (1) around 25 000 BP: presence of paleo profiles with inherited in-
 996 situ ^{10}Be concentrations in the bedrock corresponding to the long term muonic exposure during
 997 approximately 150 kyr. Stage (2) around 24 000-20 000 BP: major erosion event removing the paleo
 998 profile soil and regolith in P2 and P3 but keeping in place part of the regolith in P1. Stage (3) around
 999 19 000 BP: stabilization of deep mobile regolith on the bedrock of P2 and P3 profiles. P1 is still
 1000 covered by part of the in-situ regolith. Stage (4) from 19 000 BP to present: period with lower
 1001 denudation rates explaining the soil development and the exposure age of 19 000 yr determined for
 1002 all profiles. In P1, a thin soil was developed from the weathering of in situ regolith and was affected
 1003 by near surface mobile regolith and volumetric expansion. In P2 and P3, thicker soils were formed
 1004 from the weathering of deep mobile regolith and were affected by near surface mobile regolith and
 1005 surface mixing. Slopes and distances are purely illustrative but height variations were represented in
 1006 accordance with modeling results.
 1007

1008 Figure 9: corrected denudation rates calculated from equation 4 by assuming a uniform cosmic ray
 1009 exposure time across the watershed equal to the exposure time determined on depth profiles ($t=19$
 1010 kyr). Corrected denudation rates written in black are inferred from ^{10}Be concentrations measured in
 1011 topsoil, rock outcrop and stream sediment samples. We used in calculations the density of 2.7 g.cm^{-3}
 1012 for rock outcrops and the average regolith density of 1.7 g.cm^{-3} for topsoil and stream sediment
 1013 samples. Denudation rate written in blue is determined through the modeling of high resolution
 1014 depth profiles without assumption of steady state ^{10}Be concentrations (d , equation 4). Denudation
 1015 rate written in bold black for the outlet subcatchment is calculated from equation 10. Denudation
 1016 rates are expressed in mm.kyr^{-1} and uncertainties include the uncertainties on the ^{10}Be
 1017 concentrations, blank correction, and ^{10}Be production rates (uncertainties in table 2).

1018

1019

Table caption

1020 Table 1: sample type, depth, bulk density, integrated density and ^{10}Be data for the three high
 1021 resolution depth profiles from the southern part of the Strengbach watershed.
 1022

1023 Table 2: sample type, location, elevation and ^{10}Be data for topsoil, rock outcrop and stream sediment
 1024 samples collected across the Strengbach watershed. Maximum denudation rates (d_{max}) were
 1025 inferred from the GIS toolbox Basinga (Charreau et al., 2019) using equation 9 and assuming a steady
 1026 state of ^{10}Be concentrations. A constant bedrock density of 2.7 g.cm^{-3} was used in d_{max} calculations.
 1027 Corrected denudation rates d were calculated without assuming a steady state of ^{10}Be

1028 concentrations (equation 4) by using inherited ^{10}Be and cosmic ray exposure age determined on
1029 depth profiles. The density of 2.7 g.cm^{-3} for rock outcrops and the average regolith density of 1.7
1030 g.cm^{-3} for topsoil and stream sediment samples were used in corrected d calculations.

1031

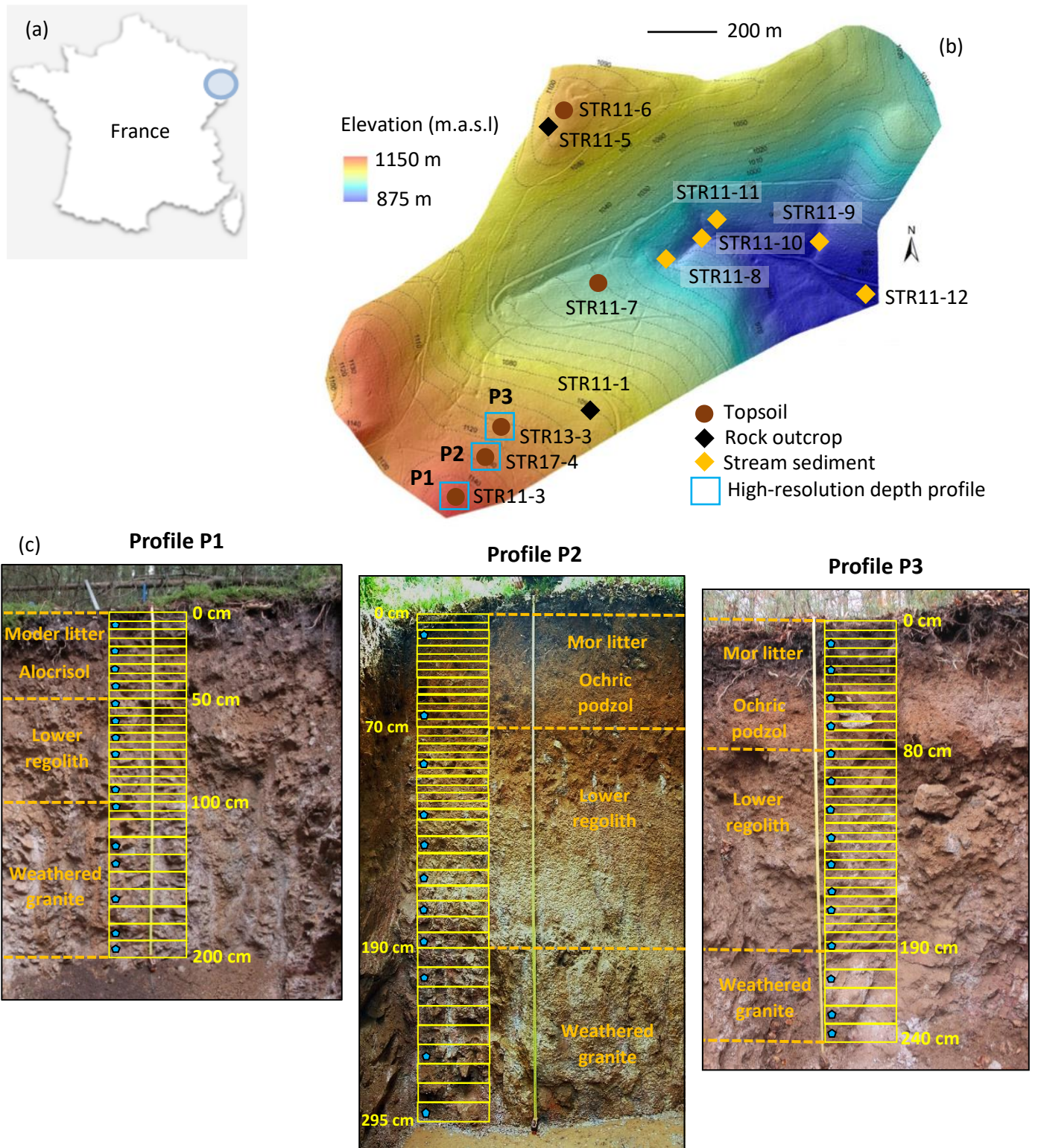


Figure 1

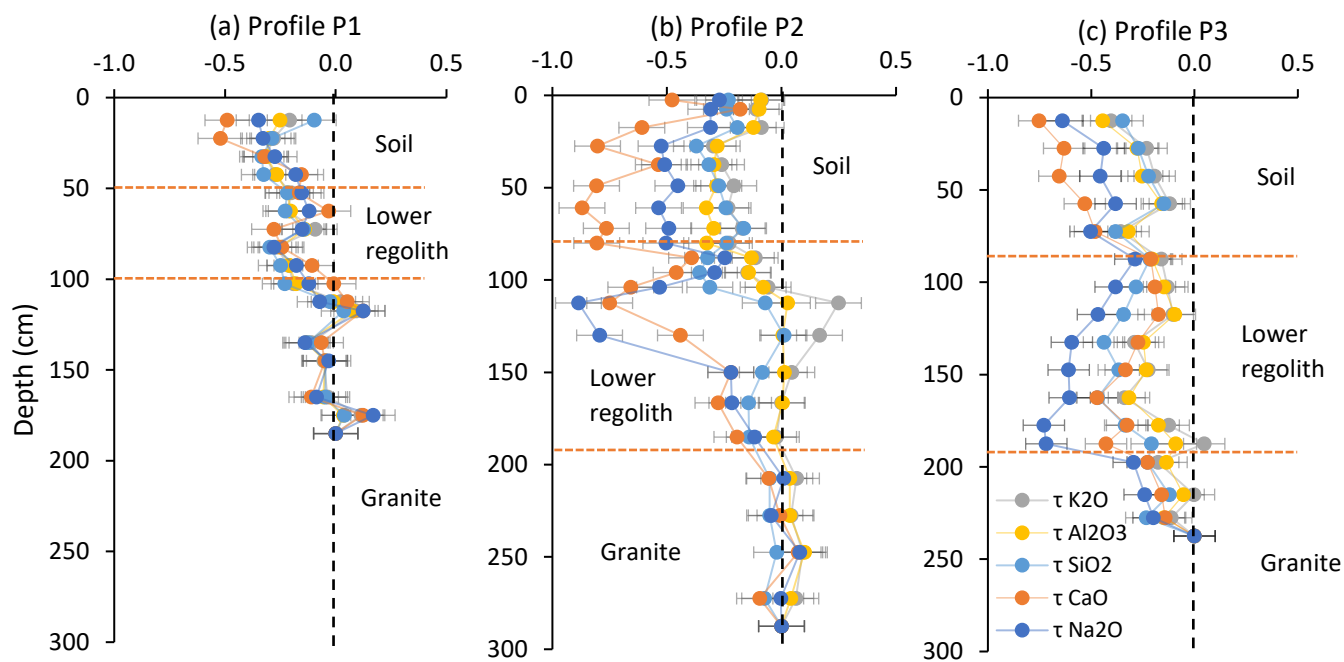


Figure 2

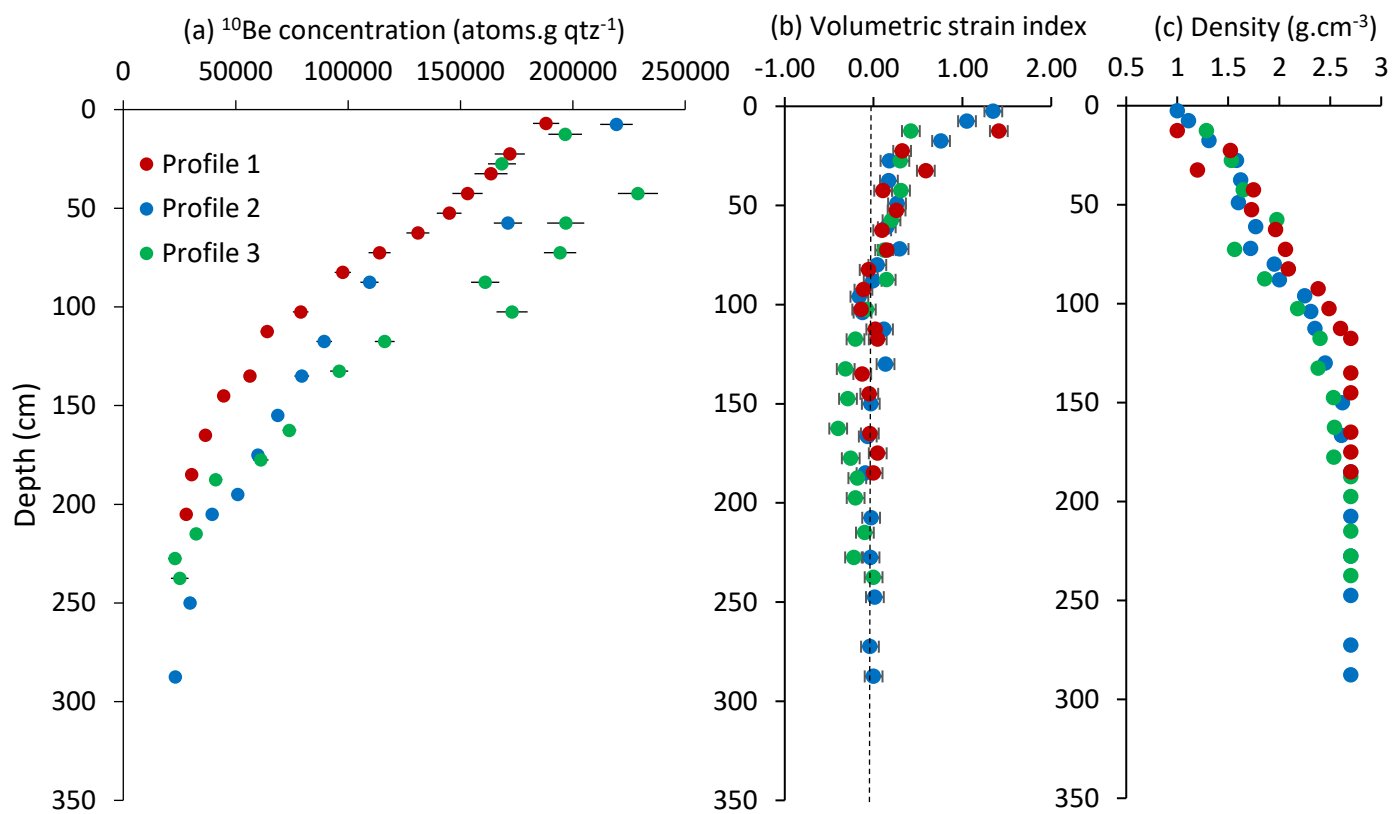


Figure 3

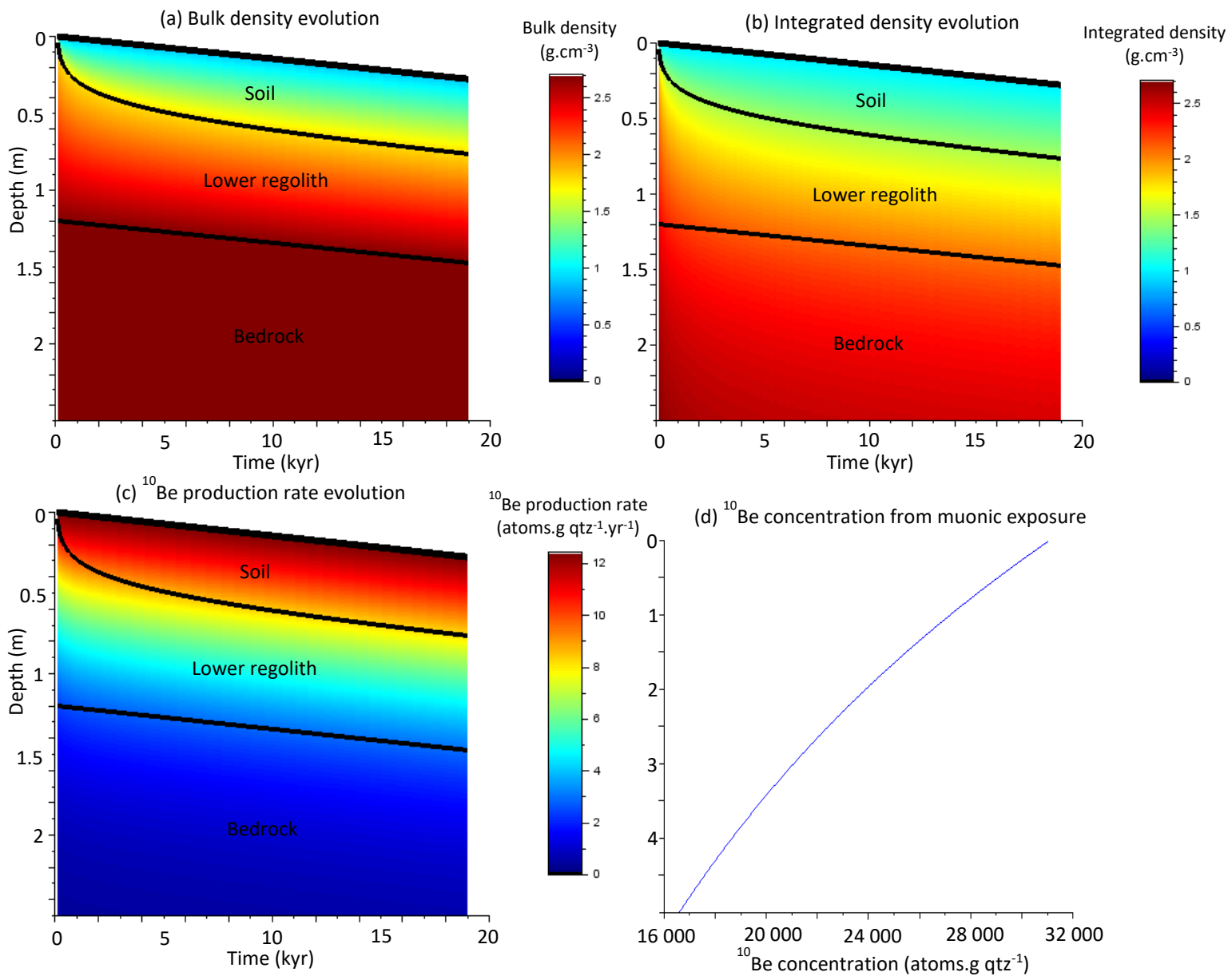


Figure 4

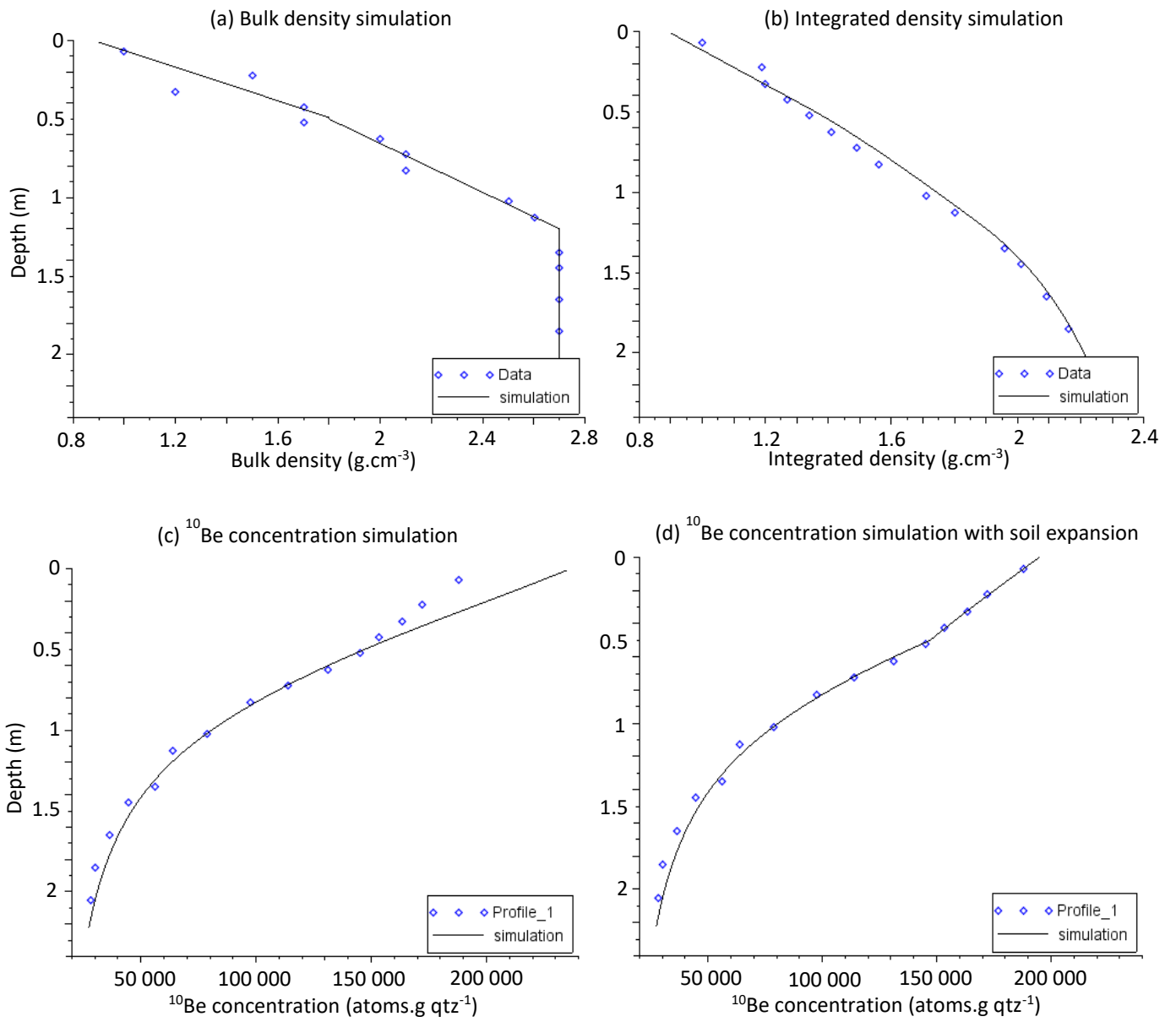


Figure 5

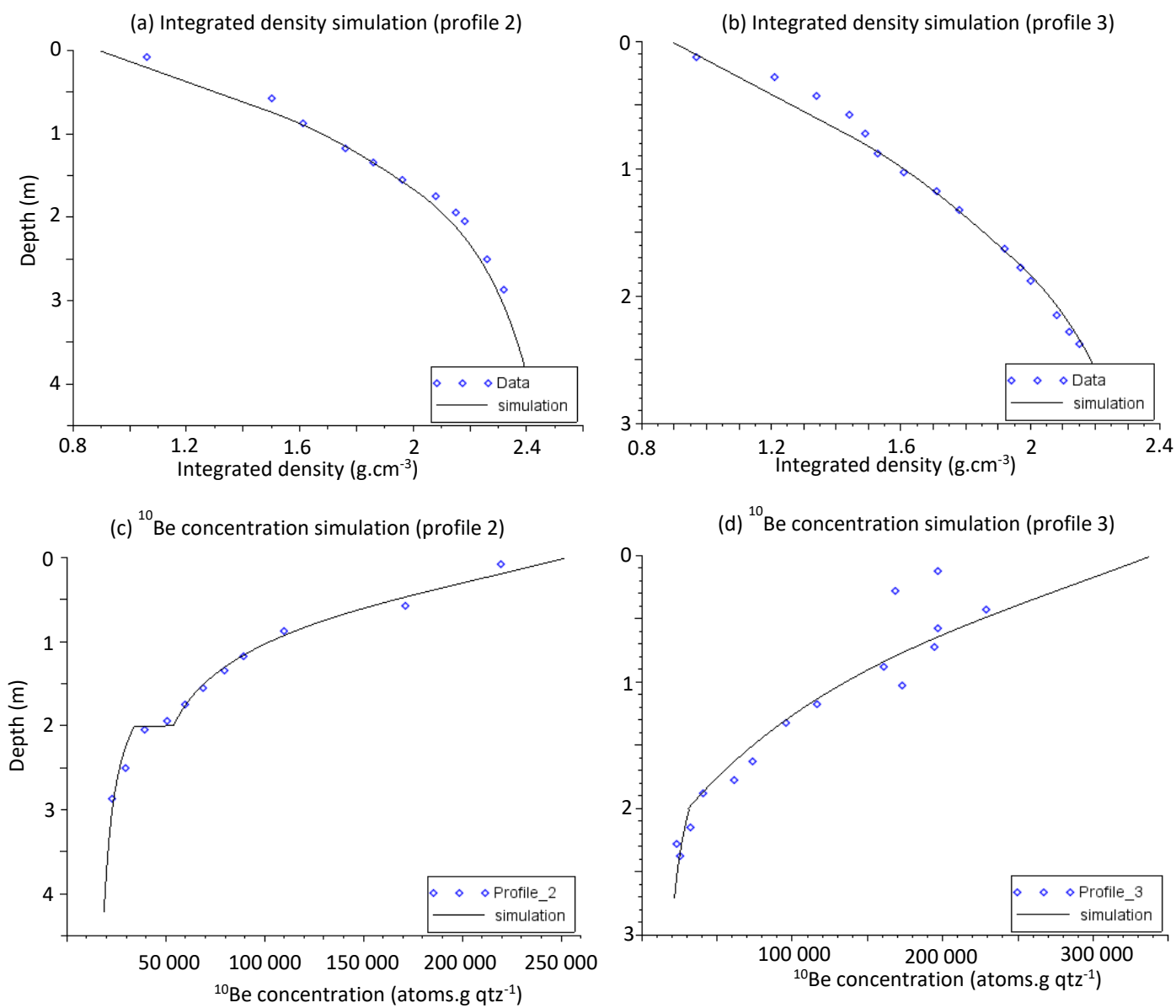


Figure 6

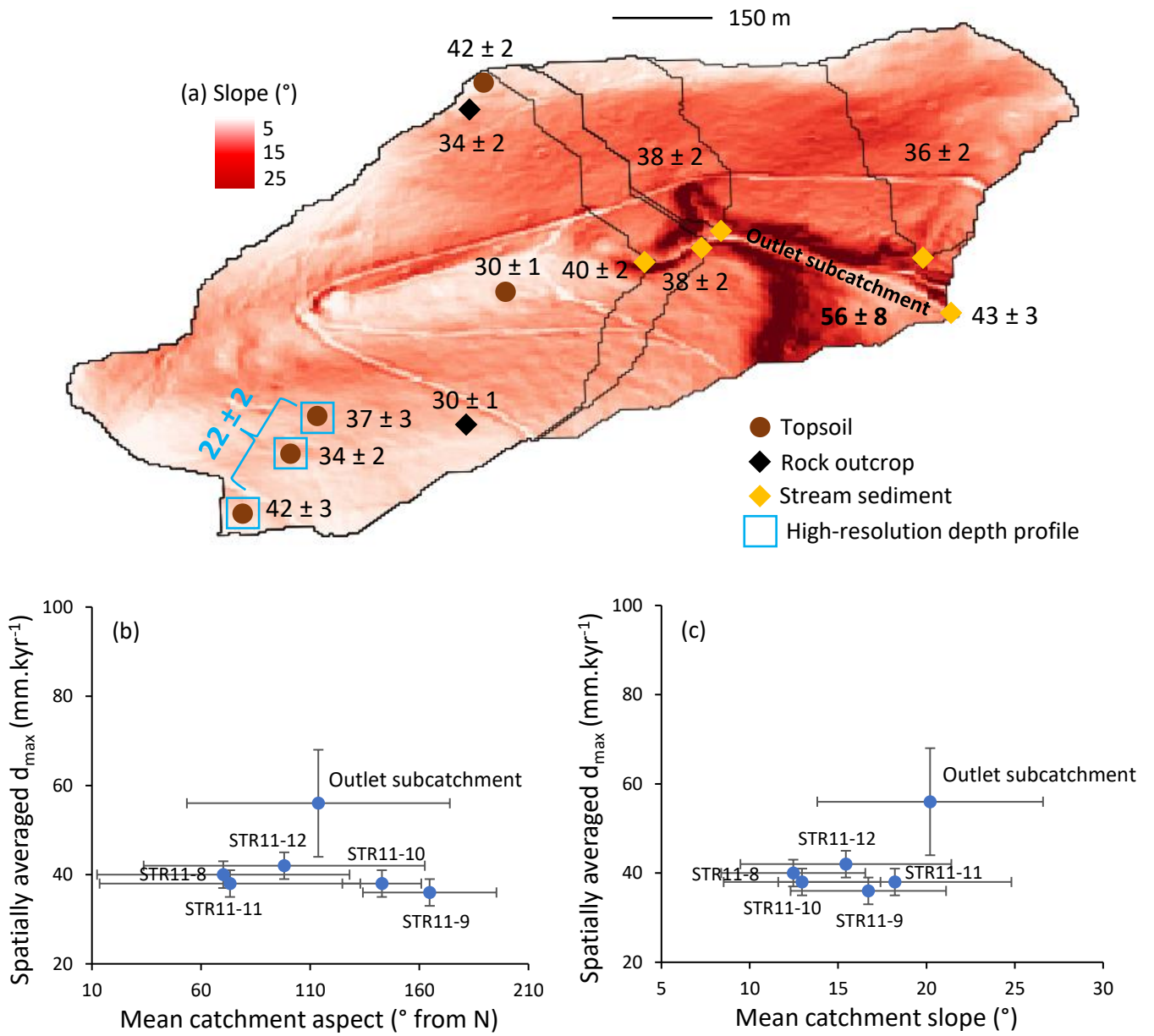


Figure 7

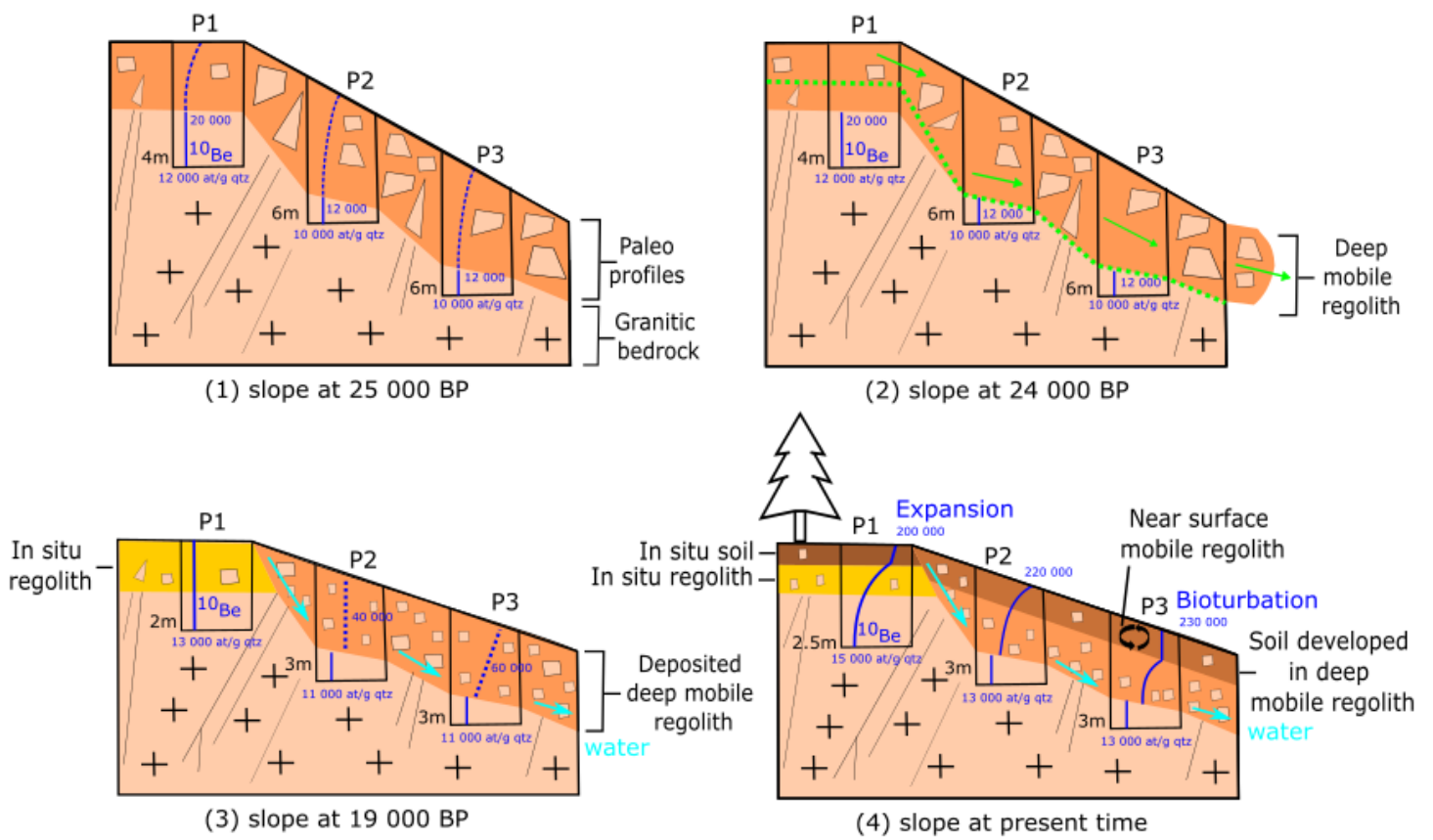


Figure 8

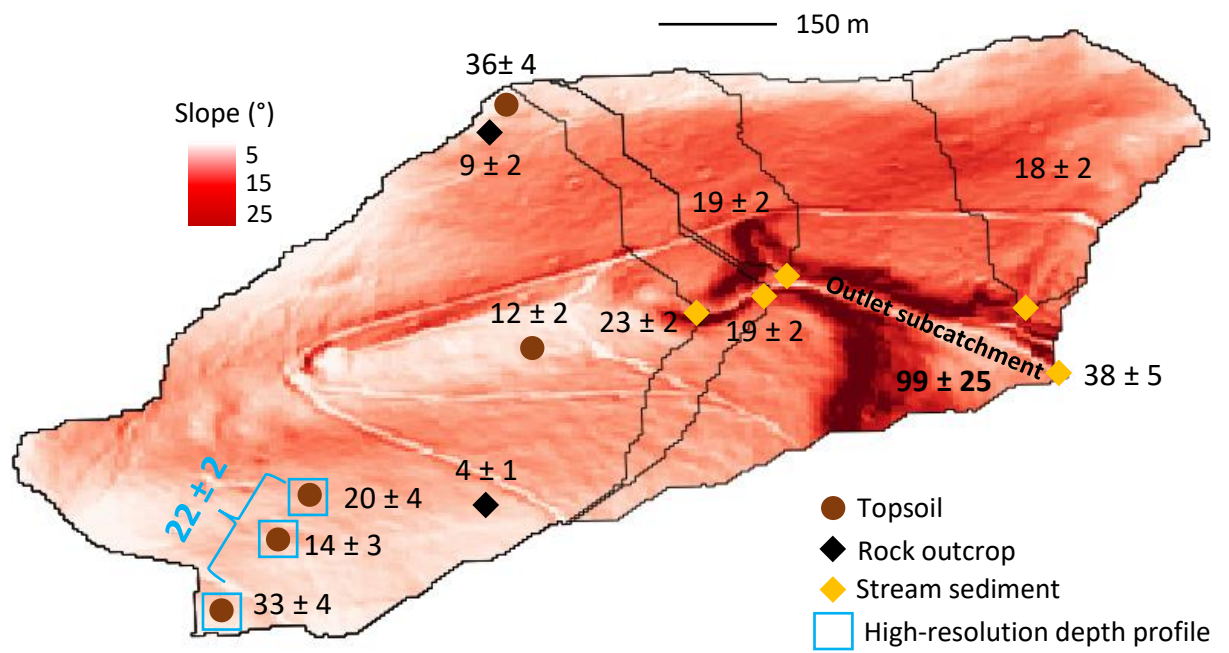


Figure 9

Sample	Type	Depth (cm)	Bulk density (g.cm ⁻³)	Integrated density (g.cm ⁻³)	Quartz mass (g)	⁹ Be (mg)	¹⁰ Be/ ⁹ Be	Err (%)	In-situ ¹⁰ Be (atoms. g qtz ⁻¹)	Err (%)
Profile P1 (48.208 °N, 7.191 °E, 1146 m)										
STR11-3	Soil	7.0	1.0	0.9	25.02	0.477	1.47E-13	3.1	188144	3.5
STR13-43	Soil	22.5	1.5	1.2	17.23	0.285	1.55E-13	3.4	171954	3.9
STR13-45	Soil	32.5	1.2	1.2	19.72	0.293	1.65E-13	4.0	163601	4.5
STR13-47	Soil	42.5	1.7	1.3	14.68	0.290	1.16E-13	3.9	153162	4.4
STR13-49	Soil	52.5	1.7	1.3	22.54	0.287	1.70E-13	3.9	144995	4.4
STR13-51	Lower regolith	62.5	2.0	1.4	18.60	0.277	1.32E-13	3.3	131079	3.9
STR13-53	Lower regolith	72.5	2.1	1.5	23.79	0.281	1.45E-13	3.8	114030	4.3
STR13-55	Lower regolith	82.5	2.1	1.6	22.03	0.252	1.23E-13	3.2	97672	3.7
STR13-59	Lower regolith	102.5	2.5	1.7	22.45	0.303	8.75E-14	3.9	78918	4.4
STR13-61	Bedrock	112.5	2.6	1.8	18.32	0.274	6.39E-14	4.0	63992	4.5
STR13-42B	Bedrock	135.0	2.7	1.9	29.31	0.291	8.49E-14	3.9	56360	4.4
STR13-43B	Bedrock	145.0	2.7	2.0	40.35	0.248	1.09E-13	5.2	44712	5.5
STR13-45B	Bedrock	165.0	2.7	2.1	38.00	0.278	7.48E-14	4.5	36532	4.9
STR13-47B	Bedrock	185.0	2.7	2.2	48.84	0.268	8.31E-14	3.9	30440	4.4
STR13-49B	Bedrock	205.0	2.7	2.2	57.09	0.280	8.55E-14	3.8	28057	4.3
Profile P2 (48.20900 °N, 7.19214 °E, 1131 m)										
STR17-4	Soil	7.5	1.1	1.1	25.49	0.519	1.61E-13	2.6	219429	3.3
STR17-14	Soil	57.5	1.9	1.5	21.67	0.518	1.08E-13	3.1	171153	3.7
STR17-20	Lower regolith	87.5	2.3	1.6	24.30	0.514	7.79E-14	3.1	109549	3.7
STR17-26	Lower regolith	117.5	2.6	1.8	30.84	0.511	8.01E-14	3.3	89399	3.9
STR17-28	Lower regolith	135.0	2.7	1.9	32.12	0.515	7.38E-14	3.6	79421	4.1
STR17-30	Lower regolith	155.0	2.7	2.0	37.16	0.516	7.41E-14	3.1	68662	3.7
STR17-32	Lower regolith	175.0	2.7	2.1	34.09	0.515	5.87E-14	3.1	59877	3.7
STR17-34	Bedrock	195.0	2.7	2.2	31.20	0.521	4.58E-14	3.4	50977	3.9
STR17-36	Bedrock	205.0	2.7	2.2	31.14	0.519	3.55E-14	4.0	39510	4.4
STR17-40	Bedrock	250.0	2.7	2.3	38.61	0.519	3.31E-14	4.9	29688	5.3
STR17-43	Bedrock	287.5	2.7	2.3	36.58	0.518	2.45E-14	6.1	23124	6.4
Profile P3 (48.209 °N, 7.193 °E, 1120 m)										
STR13-3	Soil	12.5	1.3	1.0	21.0	0.295	2.17E-13	3.2	196609	3.8
STR13-6	Soil	27.5	1.5	1.2	15.6	0.298	1.30E-13	3.1	168477	3.7
STR13-9	Soil	42.5	1.6	1.3	13.5	0.294	1.56E-13	3.3	228979	3.9
STR13-12	Soil	57.5	1.9	1.4	18.7	0.295	1.88E-13	3.7	196839	4.2
STR13-15	Soil	72.5	1.6	1.5	16.0	0.297	1.53E-13	3.1	194345	3.7
STR13-18	Lower regolith	87.5	1.9	1.6	17.6	0.297	1.39E-13	3.3	161022	3.9
STR13-21	Lower regolith	102.5	2.2	1.7	24.4	0.295	2.18E-13	3.4	172978	4.0
STR13-24	Lower regolith	117.5	2.4	1.7	14.3	0.294	8.81E-14	3.2	116320	3.8
STR13-27	Lower regolith	132.5	2.4	1.8	13.8	0.295	6.94E-14	3.7	96054	4.2
STR13-33	Lower regolith	162.5	2.5	1.9	21.6	0.297	8.00E-14	3.6	73919	4.1
STR13-36	Lower regolith	177.5	2.5	2.0	32.4	0.312	9.57E-14	5.3	61181	5.6
STR13-26B	Bedrock	187.5	2.7	2.0	47.4	0.314	9.50E-14	6.9	41096	7.2
STR13-28B	Bedrock	215.0	2.7	2.1	48.8	0.313	9.06E-14	7.1	32472	7.4
STR13-29B	Bedrock	227.5	2.7	2.1	30.1	0.295	3.74E-14	13	22949	13
STR13-30B	Bedrock	237.5	2.7	2.1	58.3	0.311	9.99E-14	15	25116	15

Sample	Type	Location	Elevation (m)	Quartz mass (g)	⁹ Be (mg)	¹⁰ Be/ ⁹ Be	Err (%)	In-situ ¹⁰ Be (atoms. g qtz ⁻¹)	Err (%)	d _{max} (mm.kyr ⁻¹)	Corrected d (mm.kyr ⁻¹)
STR11-1	Rock outcrop	48.20951 °N - 7.19587 °E	1115	25.09	0.469	1.87E-13	3.5	233458	4.0	30±1	4±1
STR11-5	Rock outcrop	48.21582 °N - 7.19528 °E	1117	24.96	0.487	1.68E-13	3.2	218546	3.7	34±2	9±2
STR11-3	Soil (top profile P1)	48.20817 °N - 7.19111 °E	1146	25.02	0.477	1.45E-13	3.1	185587	3.7	42±3	33±4
STR17-4	Soil (top profile P2)	48.20900 °N - 7.19214 °E	1131	25.49	0.518	4.82E-13	2.6	219429	3.3	34±2	14±3
STR13-3	Soil (top profile P3)	48.20942 °N - 7.19278 °E	1120	20.97	0.295	2.17E-13	3.2	204375	3.8	37±3	20±4
STR11-6	Soil	48.21621 °N - 7.19550 °E	1110	39.95	0.473	2.28E-13	3.1	180207	3.6	42±2	36±4
STR11-7	Soil	48.21300 °N - 7.19646 °E	1017	29.96	0.474	2.13E-13	3.3	225541	3.9	30±1	12±2
STR11-8	Stream sediments	48.21319 °N - 7.19866 °E	986	39.91	0.501	2.18E-13	3.1	182825	3.7	40±2	23±2
STR11-9	Stream sediments	48.21341 °N - 7.20348 °E	922	40.86	0.508	2.28E-13	3.0	189279	3.6	36±2	18±2
STR11-10	Stream sediments	48.21357 °N - 7.19994 °E	962	39.34	0.500	2.22E-13	3.1	188241	3.7	38±2	19±2
STR11-11	Stream sediments	48.21377 °N - 7.20012 °E	964	40.06	0.481	2.34E-13	3.0	188057	3.6	38±2	19±2
STR11-12	Stream sediments	48.21214 °N- 7.20447 °E	891	39.35	0.504	1.87E-13	3.1	159699	3.7	43±3	38±5

Supplementary materials

Supplementary tables

Sample	Type	Depth (cm)	Quartz (%)	Plagioclase Feldspars (%)	K-feldspars (%)	Micas and illites (%)	Kaolinite (%)	Interstratified clays (%)	Hematite (%)
Profile P1 (48.208 °N, 7.191 °E, 1146 m)									
STR13-41	Soil	12.5	43	17	25	8	2	5	0
STR13-43	Soil	22.5	43	18	25	8	1	5	0
STR13-45	Soil	32.5	40	20	23	9	2	6	0
STR13-47	Soil	42.5	37	22	26	8	2	5	0
STR13-49	Soil	52.5	38	18	31	9	3	1	0
STR13-51	Lower regolith	62.5	46	15	21	16	2	0	0
STR13-53	Lower regolith	72.5	48	15	18	17	2	0	0
STR13-55	Lower regolith	82.5	52	16	17	11	4	0	0
STR13-57	Lower regolith	92.5	45	16	22	15	2	0	0
STR13-59	Bedrock	102.5	35	22	22	20	1	0	0
STR13-61	Bedrock	112.5	34	20	26	19	0	0	1
STR13-41B	Bedrock	117.5	37	18	29	14	2	0	0
STR13-42B	Bedrock	135	40	22	22	13	2	0	1
STR13-43B	Bedrock	145	33	19	30	14	2	0	2
STR13-45B	Bedrock	165	37	22	28	11	2	0	0
STR13-47B	Bedrock	185	32	20	29	17	2	0	0
Profile P2 (48.20900 °N, 7.19214 °E, 1131 m)									
STR17-4	Soil	7.5	51	11	20	7	1	10	0
STR17-6	Soil	17.5	49	10	20	12	1	7	0
STR-17-8	Soil	27.5	48	13	23	7	1	8	0
STR17-12	Soil	49	48	14	22	6	1	9	0
STR17-14	Soil	61	47	16	25	10	1	0	0
STR17-18	Lower regolith	80	38	15	20	27	0	0	0
STR17-20	Lower regolith	88	38	18	20	24	0	0	0
STR17-24	Lower regolith	104	40	10	20	24	1	4	1
STR17-26	Lower regolith	112.5	51	4	21	16	0	6	1
STR17-28	Lower regolith	130	55	4	15	24	0	1	1
STR17-30	Lower regolith	150	43	15	18	24	0	0	0
STR17-32	Lower regolith	166.5	45	17	18	15	0	4	1
STR17-36	Bedrock	207.5	43	17	20	19	0	1	0
STR17-40	Bedrock	247.5	43	18	17	21	0	1	0
STR17-43	Bedrock	287.5	48	17	16	18	0	1	0
Profile P3 (48.209 °N, 7.193 °E, 1120 m)									
STR13-3	Soil	12.5	46	15	25	8	3	3	0
STR13-6	Soil	27.5	39	19	26	9	1	6	0
STR13-9	Soil	42.5	42	17	27	8	1	5	0
STR13-12	Soil	57.5	43	19	25	7	2	4	0
STR13-15	Soil	72.5	41	20	26	8	4	1	0
STR13-18	Lower regolith	87.5	38	21	26	14	1	0	1
STR13-21	Lower regolith	102.5	42	23	23	10	3	0	1
STR13-24	Lower regolith	117.5	33	18	18	29	0	0	1
STR13-27	Lower regolith	132.5	30	19	20	31	0	0	0
STR13-30	Lower regolith	147.5	35	16	21	28	0	0	0
STR13-33	Lower regolith	162.5	33	18	20	26	3	0	1
STR13-36	Lower regolith	177.5	44	6	24	24	0	2	0
STR13-26B	Bedrock	187.5	40	10	24	26	0	0	1
STR13-27B	Bedrock	197.5	37	23	24	13	3	0	0
STR13-28B	Bedrock	215	39	20	24	17	0	0	0
STR13-29B	Bedrock	227.5	36	25	24	15	0	0	1
STR13-30B	Bedrock	237.5	35	26	24	15	0	0	0

Table S1: mineralogical composition of the samples collected within the three depth profiles from the southern part of the Strengbach watershed. The mineralogical compositions of all the powdered 250- μ m bulk samples were determined by XRD with a Siemens D5000 diffractometer (BRGM) or a X BRUKER

D8 ADVANCE Eco Diffractometer (ITES). The $<2\ \mu\text{m}$ clay fractions were separated by centrifugation and analyzed after air drying and treatment with glycol ethylene and hydrazine to distinguish illites, smectites, kaolinites and inter-stratified clay minerals. Estimation of mineral proportion has been performed by the Rietveld method with the Siroquant software. (Uncertainty of $\approx 3\%$ for quartz and plagioclase mass proportion and $\approx 5\%$ for micas and clay minerals).

Sample	Type	Depth (cm)	SiO ₂ (%)	Al ₂ O ₃ (%)	CaO (%)	Na ₂ O (%)	K ₂ O (%)	MgO (%)	TiO ₂ (%)
Profile P1 (48.208 °N, 7.191 °E, 1146 m)									
STR13-41	Soil	12.5	73.2	12.0	0.16	1.7	4.5	0.21	0.18
STR13-43	Soil	22.5	69.3	13.5	0.18	2.1	4.9	0.26	0.21
STR13-45	Soil	32.5	68.5	14.0	0.27	2.4	5.0	0.32	0.22
STR13-47	Soil	42.5	68.1	14.7	0.33	2.6	5.2	0.36	0.22
STR13-49	Soil	52.5	70.2	13.9	0.29	2.4	5.1	0.30	0.20
STR13-51	Lower regolith	62.5	70.1	14.4	0.34	2.6	5.1	0.34	0.20
STR13-53	Lower regolith	72.5	70.6	14.1	0.23	2.2	5.3	0.26	0.18
STR13-55	Lower regolith	82.5	69.5	14.8	0.29	2.3	5.1	0.39	0.22
STR13-57	Lower regolith	92.5	69.6	14.6	0.32	2.4	5.2	0.39	0.20
STR13-59	Bedrock	102.5	70.2	14.9	0.35	2.6	5.2	0.20	0.20
STR13-61	Bedrock	112.5	71.8	14.5	0.30	2.2	5.3	0.20	0.16
STR13-41B	Bedrock	117.5	71.6	14.8	0.30	2.5	5.3	0.20	0.15
STR13-42B	Bedrock	135	72.6	14.3	0.30	2.3	5.2	0.25	0.18
STR13-43B	Bedrock	145	72.6	14.4	0.28	2.4	5.1	0.26	0.17
STR13-45B	Bedrock	165	72.1	14.2	0.26	2.2	5.2	0.20	0.16
STR13-46B	Bedrock	175	71.8	14.2	0.30	2.6	5.5	0.26	0.15
STR13-47B	Bedrock	185	72.4	14.4	0.28	2.3	5.1	0.39	0.16
Profile P2 (48.20900 °N, 7.19214 °E, 1131 m)									
STR17-3	Soil	2.5	67.7	13.7	0.21	1.7	4.9	0.30	0.16
STR17-4	Soil	7.5	69.0	14.0	0.34	1.7	4.9	0.32	0.16
STR17-6	Soil	17.5	72.4	13.4	0.16	1.7	4.9	0.26	0.16
STR-17-8	Soil	27.5	69.8	13.6	0.10	1.4	4.7	0.29	0.20
STR17-10	Soil	37.5	74.4	13.1	0.23	1.4	4.9	0.27	0.20
STR17-12	Soil	49	74.3	12.5	0.09	1.5	4.9	0.24	0.19
STR17-14	Soil	61	77.0	11.7	0.06	1.2	4.7	0.22	0.19
STR17-16	Lower regolith	72	77.6	11.2	0.10	1.2	4.7	0.21	0.17
STR17-18	Lower regolith	80	77.3	11.7	0.09	1.4	4.8	0.23	0.19
STR17-20	Lower regolith	88	70.8	15.5	0.29	2.1	5.6	0.36	0.19
STR17-22	Lower regolith	96	70.2	16.0	0.27	2.1	5.7	0.40	0.20
STR17-24	Lower regolith	104	70.3	16.1	0.16	1.3	5.9	0.47	0.19
STR17-26	Lower regolith	112.5	73.2	13.8	0.09	0.2	6.0	0.50	0.14
STR17-28	Lower regolith	130	75.2	12.8	0.19	0.4	5.3	0.47	0.14
STR17-30	Lower regolith	150	74.7	14.1	0.29	1.7	5.2	0.35	0.15
STR17-32	Lower regolith	166.5	72.6	14.5	0.28	1.8	5.2	0.37	0.15
STR17-34	Bedrock	185	72.4	13.9	0.31	2.0	5.0	0.32	0.15
STR17-36	Bedrock	207.5	74.7	13.9	0.34	2.1	5.1	0.29	0.14
STR17-38	Bedrock	227.5	75.2	14.0	0.36	2.0	5.0	0.31	0.14
STR17-40	Bedrock	247.5	73.9	14.2	0.37	2.2	5.0	0.29	0.14
STR17-42	Bedrock	272.5	74.0	14.2	0.33	2.1	5.2	0.27	0.15
STR17-43	Bedrock	287.5	76.6	13.1	0.35	2.0	4.7	0.26	0.14
Profile P3 (48.209 °N, 7.193 °E, 1120 m)									
STR13-3	Soil	12.5	71.6	11.5	0.11	1.5	4.5	0.22	0.20
STR13-6	Soil	27.5	73.3	13.7	0.15	2.2	5.3	0.27	0.19
STR13-9	Soil	42.5	72.4	13.1	0.13	1.9	5.2	0.19	0.17
STR13-12	Soil	57.5	72.1	13.4	0.16	2.0	5.1	0.21	0.16
STR13-15	Soil	72.5	70.9	14.7	0.24	2.2	5.1	0.29	0.21
STR13-18	Lower regolith	87.5	73.8	14.1	0.30	2.6	5.4	0.26	0.18
STR13-21	Lower regolith	102.5	71.4	16.0	0.33	2.3	6.0	0.35	0.19
STR13-24	Lower regolith	117.5	69.0	17.9	0.35	2.1	6.5	0.43	0.20
STR13-27	Lower regolith	132.5	69.0	17.4	0.36	1.9	6.0	0.45	0.23
STR13-30	Lower regolith	147.5	70.6	16.1	0.30	1.7	6.0	0.42	0.21
STR13-33	Lower regolith	162.5	69.9	16.9	0.28	2.0	6.0	0.47	0.24
STR13-36	Lower regolith	177.5	70.8	16.6	0.29	1.1	6.4	0.46	0.20
STR13-26B	Bedrock	187.5	72.1	15.6	0.21	1.0	6.6	0.43	0.17
STR13-27B	Bedrock	197.5	72.0	15.2	0.29	2.5	5.3	0.34	0.17
STR13-28B	Bedrock	215	72.4	14.7	0.28	2.4	5.7	0.29	0.15
STR13-29B	Bedrock	227.5	73.0	15.3	0.33	2.9	5.8	0.29	0.18
STR13-30B	Bedrock	237.5	74.3	14.0	0.30	2.8	5.1	0.26	0.14

Table S2: sample type, sampling depth and major element concentrations measured in the three high resolution depth profiles from the southern part of the Strengbach watershed (see Gangloff et al., 2014 or Prunier et al., 2015 for a detail of the analytical procedure).

	Profile P1	Profile P2	Profile P3
General parameters			
Profile depth (cm)	250	450	300
Space step (cm)	1	1	1
Time step (yr)	100	100	100
Total simulation time (yr)	19 000	19 000	19 000
Initial thicknesses			
Initial soil thickness (cm)	4	4	4
Initial regolith thickness (cm)	120	190	200
Soil and regolith evolution			
Mean denudation rate (d , cm.yr ⁻¹)	0.0022	0.0022	0.0022
Maximum regolith production rate (F_0 , g.cm ⁻² .yr ⁻¹)	0.23	0.23	0.23
Scaling factor for soil (α ,cm ⁻¹)	0.090	0.055	0.055
Scaling factor for regolith (β ,cm ⁻¹)	0.034	0.034	0.034
Typical regolith production rate (mm.kyr ⁻¹)	14	3	2
Typical soil production rate (mm.kyr ⁻¹)	16	18	19
Density			
Average regolith density (g.cm ⁻³)	1.7	1.7	1.7
Topsoil density (g.cm ⁻³)	0.9	0.9	0.9
Soil base density (g.cm ⁻³)	1.8	2.1	2.0
Bedrock density (g.cm ⁻³)	2.7	2.7	2.7
In situ ¹⁰Be			
C_0 in lower regolith (atoms g qtz ⁻¹)	Equals muonic pre-exposure (see figure 4d)	Constant value over depth of 40000 (full mixing of mobile regolith)	Variable value over depth between 15000 and 120000 (limited mixing of mobile regolith)
C_0 in bedrock (atoms g qtz ⁻¹)	Equals muonic pre-exposure (see figure 4d)	Equals muonic pre-exposure (see figure 4d)	Equals muonic pre-exposure (see figure 4d)
Attenuation length neutrons (K_n , g.cm ⁻²)	150	150	150
Attenuation length fast muons (K_{fm} , g.cm ⁻²)	5300	5300	5300
Attenuation length negative muons (K_{nm} , g.cm ⁻²)	1500	1500	1500
Surface total production rate (P , atoms g qtz ⁻¹ .yr ⁻¹)	12.43	12.24	12.05
Relative contribution neutrons (P_n)	0.9785	0.9785	0.9785
Relative contribution fast muons (P_{fm})	0.0065	0.0065	0.0065
Relative contribution negative muons (P_{nm})	0.0150	0.0150	0.0150

Table S3: variables and numerical values used for the direct modeling of the three depth profiles from the southern part of the watershed. P is the total production rate of ¹⁰Be at the surface of the soil (determined with the Basinga GIS toolbox; Charreau et al., 2019). P_n, P_{nm} and P_{fm} are the relative contributions of 97.85, 1.5 and 0.65 % to the total production and K_n, K_{nm} and K_{fm} are the effective attenuation lengths of 150, 1500 and 5300 g.cm⁻², for secondary neutrons, negative muons and fast muons, respectively (Braucher et al., 2009). We used $\lambda=5E-7$ yr⁻¹ for the radioactive decay constant of ¹⁰Be.

Catchment	Outlet location	Outlet Elevation (m)	Area (m ²)	Mean slope (°)	Mean orientation (° from N)	Spatially-averaged d_{max} (mm/kyr)	Corrected spatially-averaged d (mm/kyr)
STR11-8 catchment	48.21319 °N - 7.19866 °E	986	387370	12.5	70.2	40±2	23±2
STR11-9 catchment	48.21341 °N - 7.20348 °E	922	90856	16.7	164.7	36±2	18±2
STR11-10 catchment	48.21357 °N - 7.19994 °E	962	432397	12.9	73.2	38±2	19±2
STR11-11 catchment	48.21377 °N - 7.20012 °E	964	173057	18.2	142.8	38±2	19±2
STR11-12 catchment	48.21214 °N - 7.20447 °E	891	741344	15.4	98.1	43±3	38±5
Outlet subcatchment	48.21214 °N - 7.20447 °E	891	175011	20.2	113.7	56±8	99±25

Table S4: catchment characteristics for stream sediment samples collected across the Strengbach watershed. Maximum spatially-averaged denudation rates (spatially-averaged d_{max}) were inferred from the GIS toolbox Basinga (Charreau et al., 2019) using equation 9 and assuming a steady state of ¹⁰Be concentrations. A constant bedrock density of 2.7 g.cm⁻³ was used in spatially-averaged d_{max} calculations. Corrected spatially-averaged denudation rates d were calculated without assuming a steady state of ¹⁰Be concentrations (equation 4) by using inherited ¹⁰Be and cosmic ray exposure age determined on depth profiles. The density of 1.7 g.cm⁻³ for stream sediment samples were used in corrected d calculations.

Supplementary figures

Model flow chart

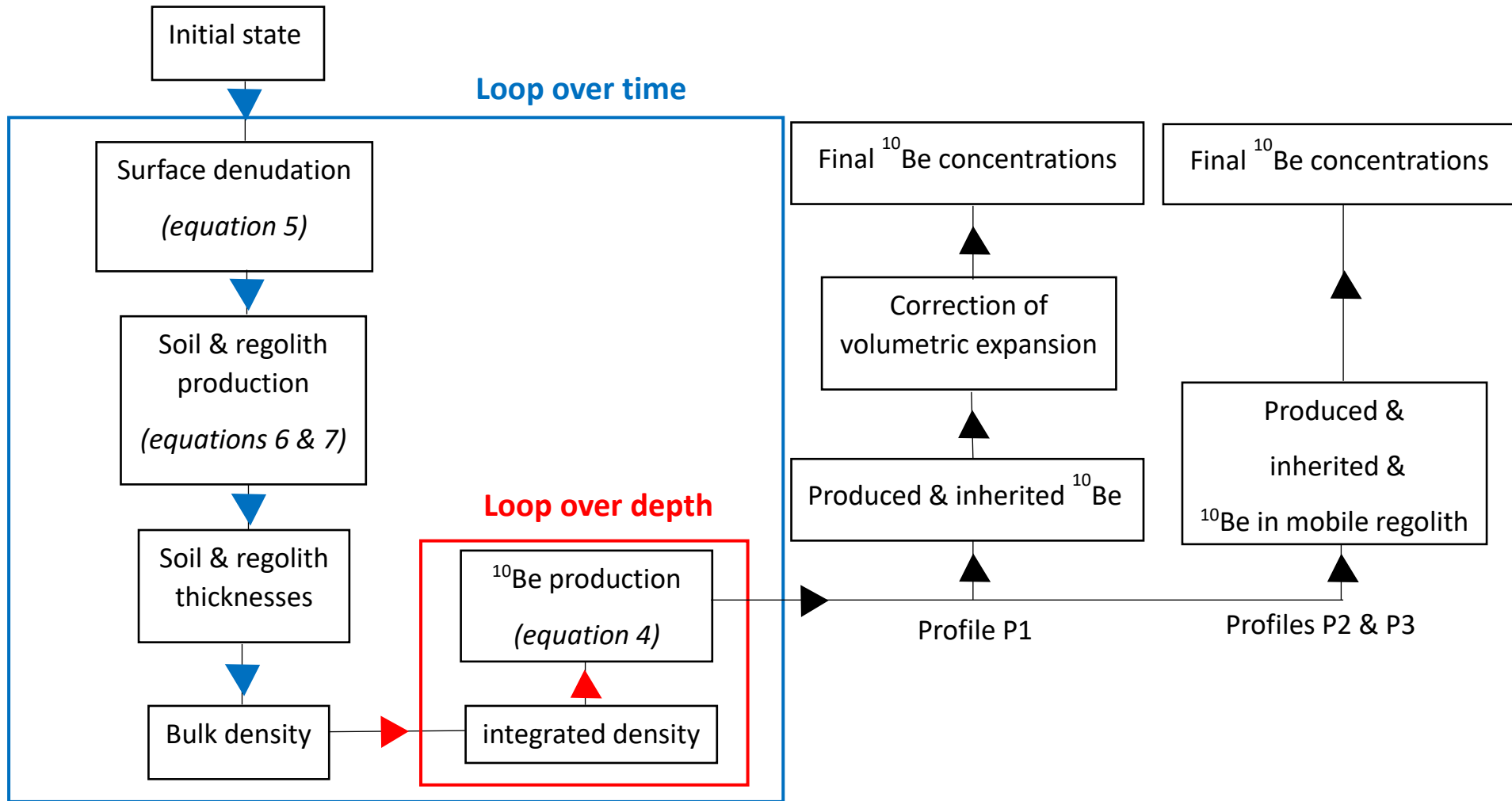


Figure S1: flow chart of the modeling of depth profile evolution and of ¹⁰Be concentration.

