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► **To cite this version:**

Stéphane Molliex, Gwenael Jouet, N. Freslon, D.L. Bourles, Christine Authemayou, et al.. Controls on Holocene denudation rates in mountainous environments under Mediterranean climate. *Earth Surface Processes and Landforms*, 2016, 42 (2), pp.272-289. 10.1002/esp.3987 . hal-01458854

HAL Id: hal-01458854

<https://amu.hal.science/hal-01458854v1>

Submitted on 12 Apr 2021

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Controls on Holocene denudation rates in mountainous environments under Mediterranean climate

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Abstract :

The Mediterranean domain is characterized by a specific climate resulting from the close interplay between atmospheric and marine processes and strongly differentiated regional topographies. Corsica Island, a mountainous area located in the western part of the Mediterranean Sea is particularly suitable to quantify regional denudation rates in the framework of a source-to-sink approach. Indeed, fluvial sedimentation in East-Corsica margin is almost exclusively limited to its alluvial plain and offshore domain and its basement is mainly constituted of quartz-rich crystalline rocks allowing cosmogenic nuclide Be-10 measurements. In this paper, Holocene denudation rates of catchments from the eastern part of the island of Corsica are quantified relying on in situ produced Be-10 concentrations in stream sediments and interpreted in an approach including quantitative geomorphology, rock strength measurement (with a Schmidt Hammer) and vegetation cover distribution. Calculated denudation rates range from 15 to 95 mm ka⁻¹. When compared with rates from similar geomorphic domains experiencing a different climate setting, such as the foreland of the northern European Alps, they appear quite low and temporally stable. At the first order, they better correlate with rock strength and vegetation cover than with morphometric indexes. Spatial distribution of the vegetation is controlled by morpho-climatic parameters including sun exposure and the direction of the main wet wind, so-called Libecciu'. This distribution, as well as the basement rock strength seems to play a significant role in the denudation distribution. We thus suggest that the landscape reached a geomorphic steady-state due to the specific Mediterranean climate and that Holocene denudation rates are mainly sustained by weathering processes, through the amount of regolith formation, rather than being transport-limited. Al/K measurements used as a proxy to infer present-day catchment-wide chemical weathering patterns might support this assumption.

Keywords : Be-10-derived denudation rates, Schmidt Hammer, Quaternary, quantitative geomorphology, weathering

45 Relief is the result of the interactions between denudation and sedimentation that are
46 mainly controlled by tectonic forces and climate changes (e.g., Davies *et al.*, 1977; Molnar
47 and England, 1990; Allen, 1997; Willett *et al.*, 2006; Whipple, 2009). The buildup of
48 topography by tectonics causes river channels and hill slopes to steepen, increasing relief
49 and, as a result, enhancing denudation rates and sediment flux (e.g., Willett, 1999;
50 Montgomery and Brandon, 2002). Moreover, rock fracturing induced by deformation
51 increases the likelihood of erosional transport (Molnar *et al.*, 2007). On the other hand, a
52 change in climatic parameters will directly impact denudation efficiency and rates (e.g.,
53 Whipple *et al.*, 1999; Bonnet and Crave, 2003). Recent studies, both onshore and offshore,
54 also highlighted the role played by glaciation in the destabilization and denudation of
55 mountainous environments during the Quaternary (Hinderer, 2001; Norton *et al.*, 2010; Jorry
56 *et al.*, 2011; Glotzbach *et al.*, 2013).

57 Millennial to present-day denudation rate estimates in recent orogens correlate with
58 environmental and/or geomorphic metrics such as local relief, slope gradient, channel
59 steepness, mean annual precipitations, and percentage of glacier cover (e.g., Summerfield
60 and Hulton, 1994; Brandon *et al.*, 1998; Montgomery and Brandon, 2002; DiBiase *et al.*,
61 2010; Bermudez *et al.*, 2013; Glotzbach *et al.*, 2013). For long-term scales, relief evolves
62 independently of denudation processes toward steady-state equilibrium between denudation
63 rates and isostatic rock uplift (e.g., Willett and Brandon, 2002). On continental margins, long-
64 term sediment fluxes directly depend on the water discharge and hydrology of the

65 contributing rivers (Litty *et al.*, 2016), the latter being controlled by climate and drainage
66 basin characteristics (catchment morphology, soil and vegetation developments, bedrock
67 type) (Syvitski and Morehead; 1999).

68 Recently, numerous studies used ^{10}Be -derived catchment-wide denudation rates to
69 quantify feedbacks between tectonics, climate, denudation, and the resulting catchment
70 morphometry in various different settings (*e.g.*, Riebe *et al.*, 2000; Matmon *et al.*, 2003;
71 Wittmann *et al.*, 2007; Ouimet *et al.*, 2009; Palumbo *et al.*, 2009; DiBiase *et al.*, 2010, 2012;
72 Delunel *et al.*, 2010; Henck *et al.*, 2011; Roller *et al.*, 2012; Savi *et al.*, 2014). Denudation
73 rates in mountainous settings have been shown to often be partly correlated to geomorphic
74 metrics, relief and slope in particular (*e.g.* compilation in Montgomery and Brandon, 2002;
75 Portenga and Bierman, 2011; Willenbring *et al.*, 2013), but some studies highlight the
76 importance of rock strength (Clapp *et al.*, 2001; Morel *et al.*, 2003; Palumbo *et al.*, 2009)
77 and/or bedding orientation (Chittenden *et al.*, 2014; Cruz Nunes *et al.*, 2015), vegetation
78 cover (Roller *et al.*, 2012; Torres Acosta *et al.*, 2015), precipitation and discharge
79 (Bookhagen and Strecker, 2012), and uplift rates (Wittmann *et al.*, 2007; Baran *et al.*, 2014;
80 Godard *et al.*, 2014).

81 The Mediterranean climate is determined by a close interplay between atmospheric
82 and marine processes and strongly differentiated regional topographies (Xoplaki *et al.*, 2004 ;
83 Kulhemann *et al.*, 2008). This leads to a specific subtropical climate with strong seasonal
84 variability, relatively warm temperatures and highly fluctuating precipitations. Very few
85 studies focused on catchment-wide denudation in such Mediterranean climate environment,
86 despite that the encountered specific climatic conditions are able to efficiently model
87 landscapes and create some impressive erosional features within granites such as "tafoni"
88 (Brandmeier *et al.*, 2011).

89 In this study, we investigate the denudation rates of the mountainous island of
90 Corsica in the Mediterranean Sea, particularly focusing on catchments draining the
91 northeastern part of the continental margin (the Golo River and surrounding watersheds).
92 Holocene denudation rates are deduced from in-situ produced cosmogenic ^{10}Be
93 concentration measurements from stream sediments. They are then compared to rock
94 strength and environmental and geomorphic metrics derived from digital elevation models
95 (DEM). Then, forcing parameters such as morphology, climate, uplift, rock strength,
96 vegetation patterns, and catchment area are discussed.

97 **1- Geological setting:**

98 Northeastern Corsica is characterized by a steep mountainous morphology, with
99 elevations reaching more than 2,700 m (Fig. 1). It is composed of two main structural units
100 (Fig. 1): (i) the Hercynian unit, an underlying magmatic basement related to the late
101 Paleozoic Hercynian Orogeny, constitutes the highest part of the island (up to 2704 m in
102 Monte Cinto); (ii) the Alpine unit, an overthrust nappe-unit built from Mesozoic oceanic
103 crust and sedimentary series emplaced and metamorphosed during the so-called “Pyrenean”
104 compression lasting from the Late Cretaceous to the Eocene (Mattauer *et al.*, 1981; Gibbons
105 and Horak, 1984; Harris, 1985; Fellin *et al.*, 2006; Danisik *et al.*, 2007). This Alpine unit was
106 exhumed during an early to middle Miocene (Cavazza *et al.*, 2001; Zarki-Jakni *et al.*, 2004)
107 regional extension, which occurred due to the Ligurian sea opening which itself was related
108 to the extension in the back-arc area of the Apennines orogen (*e.g.*, Jolivet *et al.*, 1998).
109 This caused the drifting of the Corsica-Sardinia block from the European continent of
110 southern France to its current location (*e.g.* Réhault *et al.*, 1984; Mauffret and Gorini, 1996;
111 Roca *et al.*, 1999; Gattacceca *et al.*, 2007) and reactivated the contact between the
112 Hercynian and Alpine units as a normal shear zone (Jolivet *et al.*, 1991; Daniel *et al.*, 1996;
113 Brunet *et al.*, 2000). Shallow water sediments were then deposited above the Alpine units in
114 small Miocene basins (Orszag-Sperber and Pilot, 1976) during the early-middle Miocene
115 (Ferrandini *et al.*, 1998; Cubells *et al.*, 1994). In the late Miocene–Pliocene, crustal extension
116 in the proto-Tyrrhenian Sea resulted in the subsidence and deposition of continental and
117 shallow marine deposits up to 2 km-thick in numerous half-grabens (onshore and offshore),
118 as in the Aleria basin (Fournier *et al.*, 1991; Loÿe-Pilot *et al.*, 2004). Since the late Miocene,
119 no major tectonic events have affected Corsica. However, transpressive faulting within the
120 late Miocene to Quaternary units suggest that far-field compressional stresses still affect
121 northeastern Corsica to the present-day (Fellin *et al.*, 2005a; Serrano *et al.*, 2013).

122 The evolution of paleo-surfaces exposed at high altitudes in Corsica has been
123 investigated using apatite fission-track and cosmogenic nuclide methods (Kuhlemann *et al.*,
124 2007; 2009). These studies suggest that the central and northernmost parts of Corsica have
125 experienced continuous exhumation since the Miocene, with high denudation rates in the
126 order of 1000–1500 mm.ka⁻¹ in the early-middle Miocene decreasing to less than 200 mm.ka⁻¹
127 from the late Miocene to the present-day (Fellin *et al.* 2005a; Kuhlemann *et al.*, 2007;
128 2009). This leads to the local preservation of old (Paleocene-Eocene) remnant planation
129 surfaces (Kuhlemann *et al.*, 2005a; Danisik *et al.*, 2012). Quaternary uplift rates were
130 investigated using elevated marine terraces and stair-cased alluvial terrace geometry
131 (Conchon, 1978; 1999; Collina-Girard, 1999; Fellin *et al.*, 2005b; Kuhlemann *et al.*, 2005b).
132 These rates range between 50 mm.ka⁻¹ along the southwestern coast, 150-300 mm.ka⁻¹ on
133 the northwestern coast and 200-400 mm.ka⁻¹ in the northeastern part. However, large

134 uncertainties are associated to these rates due to the fact that some terraces used in these
135 studies are not well dated and accurate corresponding sea-levels are difficult to estimate.
136 Moreover, recent uplift rates are also expected to vary significantly between different tectonic
137 regimes across the island (Lenôtre *et al.*, 1996). Nevertheless, while recent deformation is
138 well-documented near coastal domains and eastern alluvial plains (e.g. Serrano *et al.*, 2013),
139 the lack of significant seismicity, as well as the lack of recent deformation evidence in
140 upstream areas, leads us to consider Corsica to be a tectonically stable domain.

141 With a catchment area of 1,214 km² and a length of 89 km, the Golo River is the largest river
142 in Corsica (Fig. 1). It passes through the Alpine and Hercynian domains. Its average gradient
143 is 30 m/km. Preservation of fluvial terraces is low due to the very steep and confined nature
144 of the valleys. Despite the relatively steep slopes, there is no evidence of recent landslides,
145 as the colluvial fans are covered by dense vegetation (Sømme *et al.*, 2011). Its main
146 tributaries are the Asco (34.1 km; 165 km²), Tartagine (30.2 km; 136 km²), Casaluna (25.3
147 km; 100 km²), and Lagani rivers (22.1 km; 47 km²) (Fig.1). Downstream, the Golo alluvial
148 plain (Marana plain; Fig.1) is only 11 km wide. Deposits on this plain are mainly constituted
149 by pebbles in a sandy matrix and are not exclusively provided by the Golo catchment, but
150 also by two coastal catchments draining only the Alpine domain, the Bevinco to the north and
151 Fium Alto to the south (Fig.1). Corsica is experiencing a subtropical Mediterranean climate
152 with strong seasonal variability. The mean annual temperatures are relatively warm (15.9°C
153 on the Marana plain), and the precipitations are highly spatially variable, ranging from 700
154 mm.yr⁻¹ on the Marana plain to 1300-1400 mm.yr⁻¹ in the highest parts of the Golo catchment
155 (Kuhlemann *et al.*, 2007; Sømme *et al.*, 2011). Strong seasonal variations in precipitation
156 result in high flood discharge from the Golo River, with an average discharge of 20.4 m³.s⁻¹
157 and a flood peak reaching up to 734 m³.s⁻¹ (HYDRO French Database,
158 www.hydro.eaufrance.fr). Vegetation is dominated by evergreen bushes and trees in the
159 lowlands with increasing amounts of deciduous and pine forest at higher elevations (Reille *et*
160 *al.*, 1997; 1999). There is no recent glacier permanent ice in the studied area (Wilhelm,
161 1975), but the uppermost part of the Golo watershed was glaciated during the Last Glacial
162 Maximum (LGM) and subsequent glacial phases (Klaer, 1956; Kuhlemann *et al.*, 2008;
163 Krumrei, 2009).

164 **2- Material and methods**

165 **2-1) Morphometric parameters calculations**

166 For each catchment, morphometric parameters including mean elevation, mean
167 slope, shape factor, sinuosity, hypsometric integral, local relief, and the normalized

168 steepness index (Ksn) (Tab. 1), were computed from a 25 m-resolution DEM provided by the
169 French Geographical Institut (IGN).

170 Mean elevation and mean slope were computed by calculating the average value of
171 each pixel of the DEM and slope map (derivative DEM), respectively. The shape factor
172 corresponds to the ratio between maximum width and perpendicular maximum length of the
173 catchment. The sinuosity is the ratio between the straight distance from the source to the
174 outlet and the same distance along the stream.

175 The hypsometric integral (Hi) is a non-dimensional parameter representing, after
176 normalization, the repartition of the drainage area as a function of the elevation of the main
177 catchment stream (Strahler, 1952). The shape of the hypsometric curve as well as the value
178 of Hi express the volume of a basin that has not yet been eroded and thus aim to
179 quantitatively express the evolution of a catchment (Strahler, 1952). A convex hypsometric
180 curve with a high integral value (> 0.6) characterizes a weakly eroded catchment, while a
181 concave hypsometric curve with a low integral value (< 0.4) characterizes a strongly eroded
182 catchment. This parameter is highly dependent on the erodibility of the rocks (Hurtrez and
183 Lucazeau, 1999) and on the size of the catchment (Hurtrez *et al.*, 1999; Walcott and
184 Summerfield, 2008).

185 The local relief is defined as the difference in elevation between the interfluve and the
186 lowest point in the erosive channel (Ahnert, 1984). It therefore characterizes the incision, the
187 relief is high if incision is strong (e.g. ; Champagnac *et al.*, 2014). The geophysical relief is
188 another more convenient parameter, calculated as the difference between a smooth surface
189 connecting the highest points in the current landscape and the current topography (Small
190 and Anderson, 1998; Brocklehurst and Whipple, 2002). This smooth surface was created by
191 computed a flow accumulation grid from an inverse DEM. Indeed, the high values of flow
192 accumulation computed from an inverse DEM correspond to the main interfluves and
193 highpoints. The value of geophysical relief obtained by this method highly depends on the
194 “tension” of the smooth surface, i.e. the number of the high points chosen (Ahnert, 1984;
195 Lucazeau and Hurtrez, 1997; Champagnac *et al.*, 2012). For our calculation, we created two
196 smooth surfaces with different “tension” using two border values for flow accumulation (over
197 1000 cells and over 2000 cells) that correspond to the range of representative values.

198 A hydrographic network has a particular relationship between its slope and its
199 drainage area (Hack, 1957; Flint, 1974; Howard and Kerby, 1983), which can be written as
200 the slope-drainage area relationship:

201
$$S = ks.A^{\theta}$$

202 where S is the slope, A is the drainage area, and k_s and θ are the steepness and the
203 concavity of the studied stream, respectively. The concavity θ depends on erosional
204 processes and can be related to orographic rain (Schlunegger et al., 2011) and denudation
205 rates (Vanacker et al., 2015), whereas the steepness index k_s depends on erodability and
206 rock uplift. (Howard et al., 1994; Willgoose, 1994; Whipple and Tucker, 1999). Since these two
207 parameters are independent, it is possible to normalize k_s using the same reference
208 concavity for all watersheds (Snyder et al., 2000; Whipple, 2004; Duvall et al., 2004). This
209 new parameter (k_{sn}) is often used to account for rock uplift in geomorphological studies (e.g.
210 Wobus et al., 2006; Kirby and Whipple, 2012). In this study, using a reference concavity of
211 0.5, we calculated k_{sn} for each 1 km-long stream section in the studied area as well as a
212 weighted average k_{sn} for each catchment.

213 To compute the vegetation cover pattern, we used the “Corine Land Cover” European
214 program database (French data downloadable on [http://www.statistiques.developpement-](http://www.statistiques.developpement-durable.gouv.fr/donnees-ligne/li/1825.html)
215 [durable.gouv.fr/donnees-ligne/li/1825.html](http://www.statistiques.developpement-durable.gouv.fr/donnees-ligne/li/1825.html)).

216 **2-2) In-situ produced ^{10}Be concentrations and estimations of millennial-scale** 217 **denudation rates**

218 Rocks exposed to cosmic rays accumulate cosmogenic nuclides whose
219 concentrations depend on their production rates (P), their half-lives, and the denudation rates
220 (e.g. Gosse and Philips, 2001; Dunai, 2010). With long enough exposure to cosmic ray
221 derived particles, cosmogenic nuclide concentrations reach a steady-state equilibrium at
222 which cosmogenic nuclide gain due to production equals losses due to denudation and
223 radioactive decay (e.g. von Blanckenburg, 2005). At the catchment area scale (A), gain due
224 to production is ($P \times A$), while losses can be estimated by the mean concentration within river
225 sediment grains at the outlet (C) multiplied by the sediment flux (F). The concentration of a
226 particular cosmogenic nuclide (^{10}Be) measured in sediments sampled from river outlets
227 enables us to estimate the sediment flux (and therefore the denudation rate) at the
228 catchment scale, assuming constant erosion in the catchment (Brown et al., 1995; Bierman
229 and Steig, 1996; Granger et al., 1996). The integration time scale is equal to the absorption
230 depth scale (~60 cm in silicate rock; Lal, 1991) divided by the denudation rate. It often
231 corresponds to the last hundreds or thousands of years, depending on the denudation rate.

232 Sixteen samples from active riverbed sands were collected during fieldwork in 2012 at
233 the outlet of sub-catchments from the Golo watershed and surrounding areas in order to
234 determine their mean denudation rate from their in-situ produced ^{10}Be concentrations (Fig. 3;
235 Tab. 2). The physico-chemical preparation of the samples and the AMS measurements of
236 their ^{10}Be concentrations were carried out at the Laboratoire National des Nucléides

237 Cosmogéniques (LN2C) in the Centre Européen de Recherche et d'Enseignement des
238 Géosciences de l'Environnement (CEREGE), in Aix-en-Provence. Samples were prepared
239 for ^{10}Be isotopic measurements following chemical procedures modified from Brown *et al.*
240 (1991), Kohl and Nishiizumi (1992), and Merchel and Herpers (1999). For all samples, sands
241 were sieved to isolate the 250-1000 μm fraction. The magnetic fraction was removed with a
242 magnetic separator (Frantz-type), and pure quartz was obtained by at least 3 repeated \sim
243 10% $\text{H}_2\text{SiF}_6\text{--HCl}$ etchings and 3 repeated \sim 3% $\text{HNO}_3\text{--HF}$ etchings. Atmospheric ^{10}Be was
244 subsequently eliminated by 3 sequential dissolutions with diluted HF. Approximately 300 μg
245 of an in-house ^9Be carrier solution, prepared from deep-mined phenakite (Merchel *et al.*,
246 2008), was added to each sample, and residual grains were dissolved in a strong HF
247 solution. After the obtained solutions were evaporated to dryness and the residues were
248 dissolved in hydrochloric acid, Be was separated by anion and cation exchange columns.
249 After the solution volumes were reduced by heating, the Be hydroxides precipitated using
250 $\text{NH}_{3\text{aq}}$ were dried and finally ignited at 900°C to BeO. BeO targets were prepared for
251 measurement at the French National Accelerator Mass Spectrometry facility (ASTER), in
252 CEREGE, Aix-en-Provence. The obtained $^{10}\text{Be}/^9\text{Be}$ ratios were corrected for procedural
253 blanks and calibrated against the National Institute of Standards and Technology standard
254 reference material 4325 by using an assigned value of $2.79\pm 0.03\times 10^{-11}$ and a ^{10}Be half-life of
255 $1.387\pm 0.012\times 10^6$ years (Korschinek *et al.*, 2010; Chmeleff *et al.*, 2010). Analytical
256 uncertainties (reported as 1σ) include uncertainties associated with AMS counting statistics,
257 chemical blank measurements, and AMS internal error (0.5%). Long-term AMS
258 measurements of procedural blanks yield a background ratio of $3.0\pm 1.5\times 10^{-15}$ for $^{10}\text{Be}/^9\text{Be}$
259 (Arnold *et al.*, 2010). A sea-level high-latitude (SLHL) spallation production of 4.03 ± 0.18 at
260 $\text{g}^{-1} \text{yr}^{-1}$ (weighted mean of the most recently calibrated production rates in the Northern
261 Hemisphere; see Molliex *et al.*, (2013) for references) was used and scaled for latitude
262 (Stone, 2000) and elevation. The production rate is calculated for each cell of the DEM
263 constituting the catchment, and the mean catchment production rate value is calculated by
264 averaging the values of quartz-producing rock areas following the method described by
265 Delunel *et al.*, (2010) and using Balco's (2001) script to calculate the shielding factor. The
266 contribution of muons to the production rate was calculated using the physical parameters
267 evaluated by Braucher *et al.* (2011).

268 The presence of glaciers in some catchments during the integration time violates the
269 steady state assumption inherent in this method, which aims to determine a catchment-wide
270 denudation rate using in-situ produced ^{10}Be (Brown *et al.*, 1995; Bierman and Steig, 1996;
271 Granger *et al.*, 1996), and may lead to an overestimation of the denudation rates (Gosse and
272 Philips, 2001; Vance *et al.*, 2003; Portenga *et al.*, 2015). Regarding the samples we

273 analyzed, the integration times may cover different glacial periods: LGM from 23 to 19 ka
274 (Mix *et al.*, 2001); Older Dryas (OD) from 17 to 14.7 ka (Severinghaus *et al.*, 1998; Stanford
275 *et al.*, 2006), and Younger Dryas (YD) from 12.9 to 12.1 ka (Cacho *et al.*, 2002). We
276 therefore used a map of glacier coverage, produced by Krumrei (2009), for each climatic
277 period to determine the part of the studied catchments' surface affected by the presence of
278 glaciers.

279 Among the selected catchments, there are sub-catchments nested in "parent" basins.
280 In order to avoid biases in statistical analysis because of possible dependency between
281 parent and nested catchments, effective denudation rates for the non-nested portion outside
282 of the nested sub-catchment were computed following the methods of Granger *et al.*, (1996)
283 and Portenga *et al.*, (2015):

$$\text{Effective denudation rate} = \frac{\text{Parent basin sediment flux} - \text{Sub basin sediment flux}}{\text{Parent basin area} - \text{Sub basin area}}$$

284 The sediment flux being defined as:

$$\text{Sediment flux} = {}^{10}\text{Be denudation rate} \times \text{Catchment area}$$

285

286 **2-3) Quantification of rock strength by Schmidt Hammer in-situ measurements**

287 Variations in rock competence play an important role in determining bedrock incision
288 and denudation rates (e.g. Hack, 1957; Stock and Montgomery, 1999; Whipple *et al.*, 2000;
289 Sklar and Dietrich, 2001; Duvall *et al.*, 2004). The rebound values (Q) given by Schmidt
290 Hammer measurements are more widely used to quantify rock competence in
291 geomorphologic studies (e.g. Duvall *et al.*, 2004; Goudie, 2006; Engel *et al.*, 2014). The Q
292 value appears at a first order to be a good indicator of the resistance of rocks to erosion and
293 is now extensively used in geomorphic studies, even if some other factors, however difficult
294 to infer at this scale, such as fracturing density or chemical composition, enhance erodability.
295 The details of the operational aspects of the Schmidt Hammer method can be found in Basu
296 and Aydin (2004) and Aydin and Basu, (2005). In order to assess variations in bedrock
297 competence, 143 fresh outcrops in 31 different lithologic units in the Golo catchment were
298 used to measure rock strength and define a mean and mode rebound value (Q) for each type
299 of rock. For each site, 15 individual measurements on non-fractured and non-weathered rock
300 surfaces were collected, omitting any test that yielded a hollow sounding impact, that
301 fractured the rock, or that yielded a rebound value (Q) < 30 since the minimum instrument
302 reading is 30. Unconsolidated rocks such as marls or alluvial deposits were thus not tested,

303 but these rocks only represent less than 5% of the total Golo drainage area. A mean value
304 was then applied to each lithologic unit in order to create a map of the rock strength of the
305 Golo catchment. This map was used to estimate the catchment-averaged rebound value by
306 computing the weighted mean of each lithologic unit relative to its extension in the
307 catchment.

308 **2-4) Inferring weathering intensity from Al/K concentration**

309 The chemical erosion of silicates (i.e. weathering) is defined by the alteration of K-
310 feldspath to kaolinite. The clay mineralogy of a sediment can be inferred from the analysis of
311 two major elements (Al and K) which characterize the weathering degree of clays. A high
312 Al/K ratio reflects a high abundance of kaolinite and thus a high weathering degree (e.g.
313 Schneider *et al.*, 1997). Measurements of the Al/K ratio were performed on the <45 μ m
314 fraction of 26 river sediment samples. Corsican riverbed sediments are typically
315 characterized by coarse grains and contain very little clay material due to the rapid transport
316 of fine particles. Thus, the clay samples collected for Al/K measurements are from small
317 natural dams where fine particle sedimentation can occur. At each sampling site, a large
318 amount of sediment was sieved using one liter of river water and then only the 0-125 μ m
319 fraction was transferred to a pre-cleaned polypropylene bottle. Water was then discarded by
320 centrifugation at the laboratory and the sediment was sieved at 45 μ m. The 125-45 μ m
321 fraction was discarded and the fine fraction (< 45 μ m) was dried for 12 hours at 90°C and
322 crushed before analysis. The inorganic geochemical composition (Si, Al, Fe, Mn, Ca, Mg, K,
323 Na, Ti, P, S) of the fine sediment was determined on fusion beads by WD-XRF (Wave
324 Dispersion X-Ray Fluorescence, S8 Tiger, Bruker®) analysis according to the procedure
325 derived from Maghraoui *et al.* (1999). Calibration curves were established using a set of
326 internal and international reference materials. The relative uncertainty obtained is lower than
327 1%.

328 **3- Results**

329 **3-1) Geomorphologic study**

330 All computed parameters are compiled in Fig. 2 and Tab. 1. The Hercynian domain, in
331 the upstream part of the studied catchments, shows a higher mean slope (~ 25 %) than the
332 Alpine one (~ 20 %). Moreover, the morphology is clearly much steeper in the Hercynian
333 domain (mean residual relief: ~ 280 m) than in the Alpine domain (mean residual relief: ~
334 175 m). The limit between the Hercynian and the Alpine domains regarding slopes and the
335 normalized steepness index values (Ksn) for each 1 km-long stream portion is well marked
336 (Fig. 2A). Upstream, the Golo and Tavignano are both glacial U-shaped valleys with steep

337 flanks and flat bottoms (Fig. 2A). K_{sn} values vary from a factor of 1 to 10 in the Golo, Asco,
338 and Tavignano rivers, indicating that some parts (with higher values) of the hydrographic
339 network present anomalous slopes. These anomalies are located just downstream of the
340 lowest glacial-shaped features suggesting a glacial imprint on the present-day morphology. A
341 change in rock competence between the Alpine and Hercynian domains can also partly
342 account for such K_{sn} differences. Only small anomalies can be seen in the Alpine domain.
343 They may be explained by local lithologic contacts, for instance near the transition between
344 ophilites and phyllites as seen by comparing Fig. 1B and Fig. 2A. The same kind of
345 structurally induced dichotomy between the Alpine and Hercynian domains is visible on the
346 residual relief map (Fig. 2B). Such a dichotomy is not easily noticeable on the main rivers'
347 longitudinal profiles (Fig. 2C1; Fig. 2C2). Indeed, longitudinal profiles generally show
348 classical concave-shaped profiles. Nevertheless, some knickpoints can be distinguished on
349 many profiles, corresponding either to lithologic contacts (e.g. Bevinco; Fium Alto) or to
350 transitions from glaciated to non-glaciated areas (e.g. Golo, Asco, Tavignano), in good
351 agreement with K_{sn} values. These kind of observations have also been made in the
352 Himalayas where knickzones have been mapped downstream of formerly glaciated basins
353 (Korup and Montgomery, 2008), or in the Bolivian Andes where downstream increases of
354 knickzones and concavities correspond to orographic reaches (Schlunegger *et al.*, 2011).
355 Some authors also suggest that these kinds of anomalies may be related to a local change in
356 the uplift rate (Fellin *et al.*, 2005b). The values of the hypsometric integrals range between
357 0.33 and 0.47 indicating an intermediate state between mature and non-mature drainage
358 networks (Fig. 2C3; Tab. 1).

359 **3-2) Quantification of millennial-scale denudation rates**

360 The ^{10}Be -derived denudation rates range from 15 to 95 $\text{mm}\cdot\text{ka}^{-1}$ (Fig. 3), with most of
361 them ranging between 40 and 60 $\text{mm}\cdot\text{ka}^{-1}$. Except for catchment 5, integration times range
362 between 7 to 18 ka (Tab. 2). Catchments with the lowest values ($< 40 \text{ mm}\cdot\text{ka}^{-1}$) are mainly
363 located north of the studied area and drain either the Hercynian (Asco (# 5), the Tartagine (#
364 3)), or the Alpine (Lagani (# 2), Bevinco (# 1)) domains. The highest values ($> 70 \text{ mm}\cdot\text{ka}^{-1}$)
365 mainly concern catchments in the Alpine domain (Casaluna (# 10), Fium Alto (#12)), except
366 for Restonica (# 15), which is located in the Hercynian domain (Fig.3).

367 With the exception of two samples (5 and 14), the studied catchments present a
368 maximum ice cover surface lower than 15% during less than 20% of the maximum
369 integration time. This implies a measured ^{10}Be concentration of, at the most, 5% lower than
370 would be measured for the same permanently non-glaciated catchment. Given the
371 associated uncertainties, this does not significantly affect the apparent denudation rate. This

372 is also the case for catchment 14, with a maximum of 25 % of the drainage surface having
373 been covered by glaciers, but over ~8 % of the maximum integration period in the most
374 adverse conditions. Sample 5 integrates a long time period (~40 ka) that includes several
375 climatic changes and glacier development on more than 50% of its surface (Tab. 2). In this
376 case, the ^{10}Be -derived denudation rate of $15.4 \pm 1.8 \text{ mm.ka}^{-1}$ represents an overestimation.
377 Nevertheless, this denudation rate is already the lowest of the studied area, so an
378 overestimation of this value do not change interpretations about processes responsible for
379 such a low denudation. Considering this discussion about the significance of ^{10}Be
380 concentrations in past glaciated areas, we decided to not exclude any data because of the
381 presence of past glaciers in the following statistical analyses, even if the denudation rate
382 determined from sample 5 has to be considered as overestimated.

383 The effective denudation rate was computed for catchments containing a nested part,
384 i.e. 4, 7, 10, and 16. The rate is similar to that deduced from direct measurements for
385 catchments 7 and 16 and is roughly 20-30% higher for catchments 4 and 10. Taking into
386 account the associated uncertainties, the sediment flux at the outlet of catchment 8 is
387 consistent with the sum of the sediment fluxes of the Golo's upstream tributaries (# 2, 3, 4, 7,
388 and 10), indicating an efficient mixing of sediment up to this point which gives confidence to
389 our results. This portion of the catchment (# 8) is nonetheless too small to allow us to
390 compute an accurate effective denudation rate. An effective denudation rate along the
391 downstream part of the Golo (# 9) cannot be computed because the sediment supply
392 deduced from measured ^{10}Be concentrations do not show significant variations between the
393 upstream (# 8) and downstream (# 9) parts of the catchment portion (Fig. 3; Tab. 2). This
394 indicates that the sediments sampled from catchment 9 are mostly supplied by the upstream
395 part of the catchment. The downstream portion of the Golo River provides a larger part of the
396 finest material, so during sampling we probably miss the sediment resulting from denudation
397 of this sector. Since it is not possible to compute an accurate effective denudation rate, we
398 will exclude catchments 8 and 9 from our statistical analysis.

399 **3-3) Rock strength mapping**

400 Rebound values (Q) range from 40 ± 8 for Miocene clayed sandstone to 73 ± 5.5 for
401 rhyolite (Fig. 4A). Most of the rocks in the Alpine domain yield values lower than 60, whereas
402 the values are higher than 60 for most of the rocks in the Hercynian domain (Fig. 4B). The
403 Alpine and Hercynian domains are made up of three main structural units: phyllites,
404 Mesozoic sedimentary rocks (flysch, sandstone, and limestone), and ophiolitic complex in the
405 Alpine domain, and a crystalline metamorphic basement, with granites and acid volcanic
406 rocks in the Hercynian domain. Rebound values were averaged for each structural unit (Fig.

407 4C). The measurements show that the three harder units belong to the Hercynian domain (~
408 65 ± 9 ; Fig. 4C) while the three softer are in the Alpine domain (~ 53 ± 12 ; Fig. 4C). In order
409 to compare the ^{10}Be -derived denudation rates and mean catchment Q values, we have to
410 assume that the quartz content is almost similar in each unit, even if it difficult to accurately
411 estimate. Some specific rocks do not contain quartz (i.e. limestones) but they only represent
412 a small part of the total drainage area. Quartz-bearing rocks are contained in all units (even
413 ophiolites which are frequently cutted by acid crystalline dykes and quartz veins) and we
414 estimate that the most extended rocks contain a similar ~ 30-50% quartz-content.

415 **3-4) Inferring vegetal cover and relation with exposition**

416 Nowadays, human activities are mainly concentrated in the alluvial plain. The
417 repartition of vegetation is thus almost natural in most parts of the studied area. Vegetation in
418 the Hercynian domain is often sparse or less developed, except for some coniferous forests
419 located in high-elevation valleys. In many places, rocks are directly exposed. The Alpine
420 domain is mainly covered by deciduous or mixed forests and sclerophyllous vegetation (Fig.
421 5C). Field expositions were calculated for each type of vegetation (Fig. 5A; 5B; 5C). Results
422 show that the forests are mainly on NW-exposed slopes (Fig. 5B). This is especially true for
423 coniferous and mixed forests, whereas transitional woodland-shrub domains do not develop
424 on any preferential slope orientation. Areas covered by short vegetal species such as
425 grassland or sclerophyllous are mainly on S- to SE-exposed slopes (Fig. 5A). Areas where
426 rocks directly outcrop are also mainly orientated to the southeast (Fig. 5A). To summarize,
427 vegetation was classified into two main groups: the short vegetation domain and the forest
428 domain, respectively (Fig. 5A; 5B). More than 61% of the forests are on N- to W-exposed
429 slopes, whereas only 26% are on E- to S-exposed slopes. For the short vegetation, 50% is
430 found on E- to S-exposed slopes, whereas only 27% is on N- to W-exposed slopes.

431 **3-5) Al/K measurements reflecting the role of vegetation and/or lithology in** 432 **weathering intensity**

433 The Al/K ratio ranges from 1.6 to 4.4 (Table 3 and Fig. 6). It has been compared to
434 bedrock strength and vegetation cover. Indeed, the combined presence of high vegetation
435 (presence of forests) and soft bedrock promotes the efficiency of the weathering intensity
436 and regolith formation. In both cases, despite the scattered data, a correlation is noticed with
437 the Al/K ratio ($R^2 = 0.64$, linear for bedrock strength and $R^2 = 0.49$, linear for vegetal cover)
438 (Fig. 6C and Fig. 6D). The weaker the bedrock, the more important the forest cover in the
439 catchment, and the higher the Al/K ratio. Note that three samples (46, 48, 79; Tab. 3) are
440 excluded from the general regression lines. These samples result from the drainage of a
441 large amount of ophiolites whose initial K_2O concentration was significantly lower than that of

442 other Corsican rocks (Grelou-Orsini, 1977; Bailly, 2004) (Fig. 6). The initial composition of
443 the drained rocks indeed influences the Al/K ratio (Bayon *et al.*, 2012) when considering the
444 < 45 μm size fraction. This may also explain the scattering of the data (Fig. 6C and Fig. 6D).
445 Nevertheless, the overall correlation between Al/K ratio, rock strength and vegetal cover
446 leads us to believe that the Al/K ratio can adequately reflect the intensity of weathering in the
447 studied area.

448 **3-6) Comparisons between denudation rates and morphometric indexes,** 449 **climate, lithology, and vegetal cover.**

450 ^{10}Be -derived denudation rates are mostly integrated over a ~15 ka long time period. The
451 climate in Corsica was colder in the past (Kuhlemann *et al.*, 2008) and the vegetation was
452 mainly constituted by steppes during late glacial and the first thousands years of the
453 postglacial period (Reille *et al.*, 1999). The present-day pattern of forests in Corsica has been
454 nearly the same for the last 7-8 ka (Reille *et al.*, 1999). This pattern is mainly controlled by
455 constant parameters (exposition, altitude, relative humidity, type of substratum...). Without
456 strong human activities, shifts in climate conditions would have only resulted in overall
457 denser or sparser vegetation without significant changes in the spatial pattern of vegetation
458 cover. In these conditions, the use of modern vegetation indices is reliable enough to assess
459 relative distribution over similar integration periods (e.g. Portenga *et al.*, 2015; Torres Acosta
460 *et al.*, 2015). Fig. 7 shows the relationship between the ^{10}Be -derived denudation rates in the
461 studied catchments and different morphometric and climatic parameters such as mean
462 elevation (Fig. 7A), slope (Fig. 7B), local relief (Fig. 7C), normalized steepness indexes (Fig.
463 7D), and annual rainfall (Fig. 7G) (from the WorldClim database; Hijmans *et al.*, 2005). We
464 also compare ^{10}Be -derived denudation rates with the mean rebound value (Q), used as a
465 proxy of catchment rock strength (Fig. 7E), and with modern vegetation cover (Fig. 7F)
466 represented as the percentage of forests in the catchment.

467 At a first order, none of the morphometric parameters correlate with the deduced
468 denudation rates ($R^2 < 0.05$ for slope, Ksn, local relief, mean elevation), indicating that
469 morphology is not the main factor controlling the denudation. Mean annual rainfall values do
470 not correlate either ($R^2 = 0.17$). Denudation rates present a better fit with rock strength (Fig.
471 7E; $R^2 = 0.61$; linear) and percentage of forests ($R^2 = 0.42$; linear); the softer the rock and the
472 more developed the forest, the higher the denudation rate. Note that rock strength also partly
473 correlates with vegetal cover, the forests being preferentially developed on softer rocks (Fig.
474 7H). In detail, correlations between denudation, rock strength, and percentage of forest are
475 mainly driven by the three same samples: 4, 10, and 12. These samples correspond to the
476 extreme values of both rock strength and denudation rates (Tab. 2). If we compute the same

477 relationships without these three samples, correlations with rock strength and vegetation
478 weaken ($R^2 = 0.05$ for both parameters), whereas correlations with morphometric indexes
479 significantly increase ($R^2 = 0.33$ for mean elevation; $R^2 = 0.54$ for mean slope; $R^2 = 0.37$ for
480 local relief; $R^2 = 0.43$ for Ksn; Fig. 7). It appears that rock strength and/or vegetation controls
481 denudation rates at a first order but when values of these two parameters are weakly
482 distributed, a correlation between denudation rates and morphology can be noticed.

483 **4- Discussion**

484 **4-1) Comparison with previously determined denudation rates**

485 Denudation rates deduced for Corsica catchments are quite low in comparison with
486 rates determined in the foreland of the northern European Alps, which present a similar
487 geomorphic context (Wittmann et al., 2007; Norton et al., 2011; Molliex et al., 2016).

488 The contemporary averaged sediment flux (bedload and suspended load) estimated
489 from measurements performed at the Golo River outlet during the August 27th 1975 flood
490 event is $64,500 \pm 7,500 \text{ t.yr}^{-1}$ (Quenelec, 1982). Considering a mean density of 2.7 t.m^{-3} for
491 the eroded rocks, the corresponding deduced denudation rate is $23 \pm 2.7 \text{ mm.ka}^{-1}$. This value
492 is about half that deduced from ^{10}Be concentration measurements, that is $46.5 \pm 9.4 \text{ mm.ka}^{-1}$
493 integrated over the last 13.5 ka (catch. 9, Fig. 3, Tab. 2). Due to the rather short time period
494 integrated when performing the direct sediment flux measurement method, extreme flood
495 events such as millennial or centennial floods are not necessarily recorded, which may lead to
496 significant underestimations of fluxes in torrential or particular hydrological domains
497 (watersheds highly sensitive to sudden floods, for instance) (Kirchner et al., 2001; Serrat et
498 al., 2001; Molliex et al., 2016).

499 Conversely, a Holocene denudation rate of $48 \pm 25.3 \text{ mm.ka}^{-1}$ was deduced from the
500 estimation of sediment volumes deposited offshore (Calvès *et al.*, 2013; unit 3; last 16 ka;
501 61.5 to $198 \text{ t.km}^2.\text{yr}^{-1}$), similar to that deduced from in situ-produced ^{10}Be concentration
502 measurements. Nonetheless, denudation rates deduced onshore could be inconsistent with
503 sediment flux deduced offshore because of the increase of sediment storage in the
504 catchment during glacial periods (Harbor and Warburton, 1993; Hinderer, 2012). Indeed,
505 glacial denudation leads to the formation of moraines, which are preserved during the glacial
506 period because of the lower precipitation rate and the lower transport capacity due to water
507 discharge. Moraines could thus constitute a significant volume of stored sediment that can
508 easily be mobilized during deglacial and interglacial periods when fluvial processes become
509 more efficient, yielding then to a higher sediment flux during interglacial periods. In the case
510 of eastern Corsica the relatively small extension of several preserved moraines (Krumrei,

511 2009; Kuhlemann *et al.*, 2009) should not influence denudation rates inferred from sediment
512 fluxes.

513 **4-2) A peri-glacial environment at steady-state equilibrium during Holocene ?**

514 Long-term (Quaternary) denudation rates inferred from in situ-produced ^{10}Be concentration
515 measurements on granitic crests ranged from 8 to 20 mm.k^{-1} in the Golo catchment
516 (Kuhlemann *et al.*, 2007; 2009). The late Quaternary incision rate of bedrock by the main
517 Golo stream, inferred from OSL-dated terraces geometry, ranges from 160 to 475 mm.k^{-1}
518 (Fellin *et al.*, 2005a; Sømme *et al.*, 2011). The ^{10}Be derived catchment-scale denudation rate
519 determined for the Golo catchment in this study thus ranges between those estimated for the
520 crests and the main stream. In catchments where stream incision is more efficient than
521 denudation of the crests, the main stream location corresponds to the most effective
522 denudation zone and the crests are significantly less eroded, inducing an increase in relief
523 (e.g. Whipple *et al.*, 1999; Meyer *et al.*, 2010; Yanites and Ehlers, 2012; Champagnac *et al.*,
524 2014). This relief-increase model for the Golo is supported by the stair-cased geometry of
525 terraces in the catchment, indicating the efficient incision of the stream at the Quaternary
526 scale.

527 In glacial and peri-glacial environments such as Corsica, three main climatic phases can be
528 distinguished through changes in sediment fluxes and sedimentary patterns (e.g. Church and
529 Ryder, 1972; Hinderer, 2012). The glacial phase leads to an onshore incision, a strong frost-
530 cracking, and eolian alteration. The deglacial phase leads to a strong phase of aggradation
531 due to sea level rise and a high sediment supply controlled by glacial melting and moraine
532 degradation. This phase leads to the deposit of braided terraces. After the sediment supply
533 peak due to the deglaciation, sediment fluxes decrease exponentially and approach an
534 interglacial equilibrium (Ballantyne, 2002).

535 In a fluvial environment, channel steepening is linked to uplift and denudation rates. Channel
536 steepness is thus a reliable proxy that allows us to investigate these parameters (e.g. Anherst,
537 1970; Schmidt and Montgomery, 1995; Roering *et al.*, 1999, 2001; Montgomery and
538 Brandon, 2002; DiBiase *et al.*, 2012). The presented quantitative geomorphology study in
539 Corsica shows that glacial processes in the upstream part of the main catchments have
540 impacted the hydrographic network. On the other hand, global studies indicate that
541 morphology often controls denudation rates in such steep environments (e.g. Montgomery
542 and Brandon, 2002; Portenga and Bierman, 2011; Willenbring *et al.*, 2013). In the northern
543 European Alps, ^{10}Be -derived denudation rates appear to be partly correlated with
544 geomorphic metrics such as slope or relief (Wittmann *et al.*, 2007; Delunel *et al.*, 2010;
545 Norton *et al.*, 2011; Glotzbach *et al.*, 2013; Molliex *et al.*, 2016).

546 ¹⁰Be deduced denudation rates for Corsica can be considered as representative for the
547 Holocene, an interglacial period characterized by a warm and relatively wet climate (Reille et
548 al., 1997; 1999). The fact that denudation rates which integrate nearly all the Holocene
549 period still almost stable despite strong differences in morphologic indexes suggest that
550 Corsica reached the interglacial equilibrium described by Ballantyne (2002).

551 **4-3) Importance of regolith production on Holocene denudation rates in** 552 **Corsica.**

553 When the differences in rock strength and/or vegetation cover are significant,
554 denudation rates in Corsica, even if they still low, seem more evidently linked to rock strength
555 and vegetation than to morphology. Because forests are more developed in the Alpine
556 domain, where rocks are softer, than in the Hercynian domain, it is difficult to decipher
557 whether vegetation or rock strength mainly controls the denudation rates. We showed that
558 land exposure plays an important role in vegetal cover distribution, forests being well-
559 developed on N- to W-exposed land, whereas S- to E-exposed land is preferentially covered
560 by short vegetation. This repartition could be explained by climatic parameters. Indeed, the
561 “Libecciu”, the prevailing wet wind in Corsica, blows from the west (Benevent, 1914;
562 Brandmeier *et al.*, 2011). Brandmeier *et al.* (2011) previously pointed out the importance of
563 westward winds such as the “Libecciu” in the particular weathering of Corsican granites
564 characterized by the development of westward-oriented tafoni (erosional features in
565 granites). North-exposed lands have less sun exposure at a daily scale and are thus more
566 prone to keep humidity, which favors vegetation growing. The vegetation development
567 increases the efficiency of pedogenic processes: the more the vegetation is developed, the
568 more the soil and regolith are developed. Correlations between the Al/K ratio, percentage of
569 forests, and bedrock strength could suggest that the chemical weathering of the bedrock and
570 thus the development of the regolith is controlled by the vegetation and type of rocks. Norton
571 *et al.* (2014) demonstrated that at a low denudation rate, regolith production is poorly related
572 to changes in precipitation rate and temperature. This may explain why precipitation rates,
573 even if the resolution of the climatic data is poor in the studied area (WorldClim database;
574 Hijmans *et al.*, 2005), do not correlate with ¹⁰Be-derived denudation rates (Fig. 7G). The
575 denudation rates in Corsica might thus to be directly related to the regolith development, the
576 chemical weathering being more efficient than the physical processes in the denudation
577 process. Some studies also demonstrate a better correlation of denudation rates with
578 lithology and regolith development than with morphometric parameters, even in mountainous
579 environments (Clapp *et al.*, 2001; Morel *et al.*, 2003; Palumbo *et al.*, 2009). Indeed, in those
580 cases, rock strength is directly presented as being responsible for a more efficient regolith
581 production due to rock structures (joints, dipping, schistosity) (Densmore *et al.*, 2007).

582 The size of the studied catchments may be an important parameter since correlation
583 between morphologies and denudation rates is often highlighted for large catchments or
584 small catchments in tectonically active settings (e.g. Riebe et al., 2000) while studies
585 concluding to a non-correlation between morphometric indexes and denudation rates often
586 concern small catchments in non-tectonically active settings (< 500 km²; Palumbo *et al.*,
587 2009; Roller *et al.*, 2012). Indeed, it has been suggested that in large catchments, spatial
588 variations of denudation rates linked to local parameters such as lithology or vegetation are
589 averaged (Summerfield and Hulton, 1994; Palumbo *et al.*, 2009).

590 **Conclusion**

591 Millennial-scale ¹⁰Be-derived denudation rates obtained from the Golo sub-basins and
592 surrounding catchments are quite low, almost spatially stable and do not correlate at first
593 order with geomorphic metrics (such as slope, relief, steepness, mean elevation...), as it is
594 usually the case in similar geodynamic and geomorphologic settings. This suggests that
595 Corsica's landscape reached an interglacial steady-state equilibrium that might be caused by
596 the specific characteristics of the Mediterranean climate. Based on catchment mean rock
597 strength (Schmidt Hammer measurements) and vegetation cover distribution, we highlighted
598 that the rock strength and spatial distribution of the vegetation probably play a significant role
599 in the denudation distribution. The pattern of vegetation is partly controlled by its exposure to
600 the sun and the direction of the main regional wet-wind (so-called "Libecciu"). Even if it is
601 difficult to decipher the role of each parameter, we suggest that regolith formation by
602 chemical weathering is the main parameter controlling denudation in mountainous
603 environment experiencing such climatic context. Indeed, higher denudation rates are located
604 in areas where conditions are the most favorable for the development of regolith (weak
605 bedrock or/and developed vegetation). Al/K ratio measurements also support a weathering
606 efficiency linked to the development of the vegetation and/or to the strength of the
607 substratum. The small size of the studied catchments might play a role in the fact that short-
608 term denudation rates better reflect local processes such as regolith formation rather than
609 the catchment's general morphology.

610 **Acknowledgements**

611 S. Molliex benefited from a post-doc fellowship granted by LabexMER and IFREMER.
612 This work further benefited from a State Grant from the French "Agence Nationale de la
613 Recherche (ANR)" in the Program "Investissements d'Avenir" (ANR-10-LABX-19-01, Labex
614 Mer). The French CNRS INSU SYSTER partly financed the cosmogenic nuclide
615 concentration measurements. We thank L. Léanni, M. Arnold, G. Aumaître, and K.
616 Keddadouche for their valuable assistance during ¹⁰Be measurements at the ASTER AMS

617 national facility (CEREGE, Aix en Provence), which is supported by the INSU/CNRS, the
618 ANR through the "Projets thématiques d'excellence" program for the "Equipements
619 d'excellence" ASTER-CEREGE action, IRD and CEA. We thank B. Drussel and E. Gautier
620 for their assistance in the field. We are also grateful to P. Bierman, F. Schlunegger and 2
621 anonymous reviewers for their fruitful comments which contributed to improve the
622 manuscript. K. Kovacs post-edited the English style.

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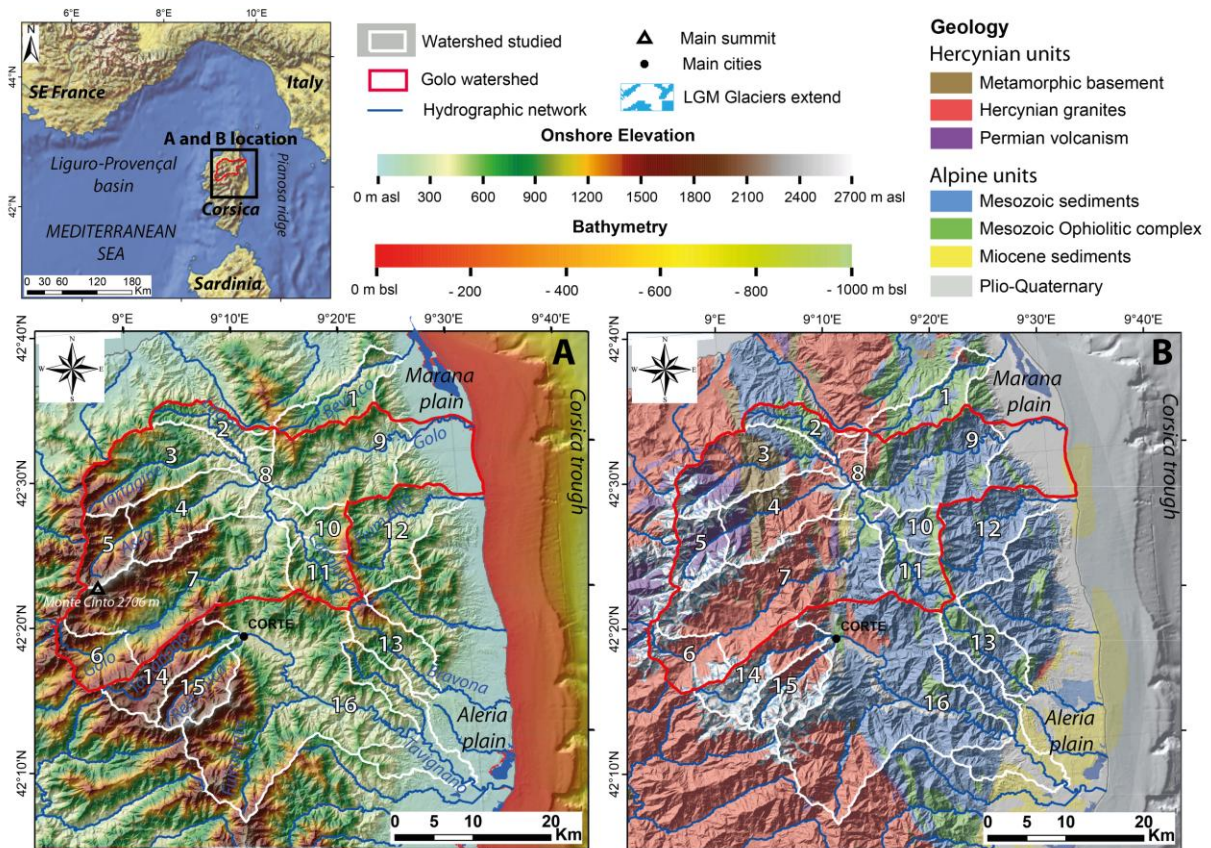
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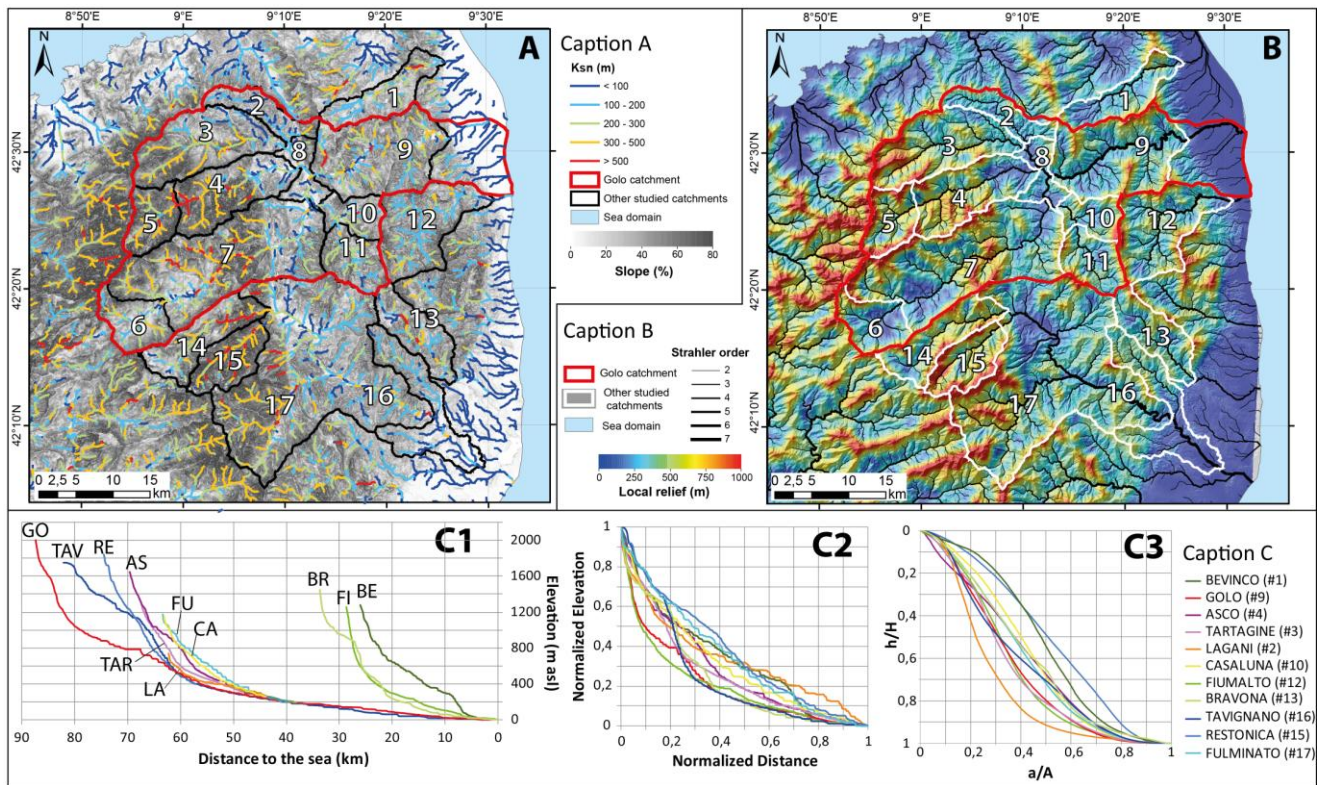
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 1012

1013 **Figures captions**

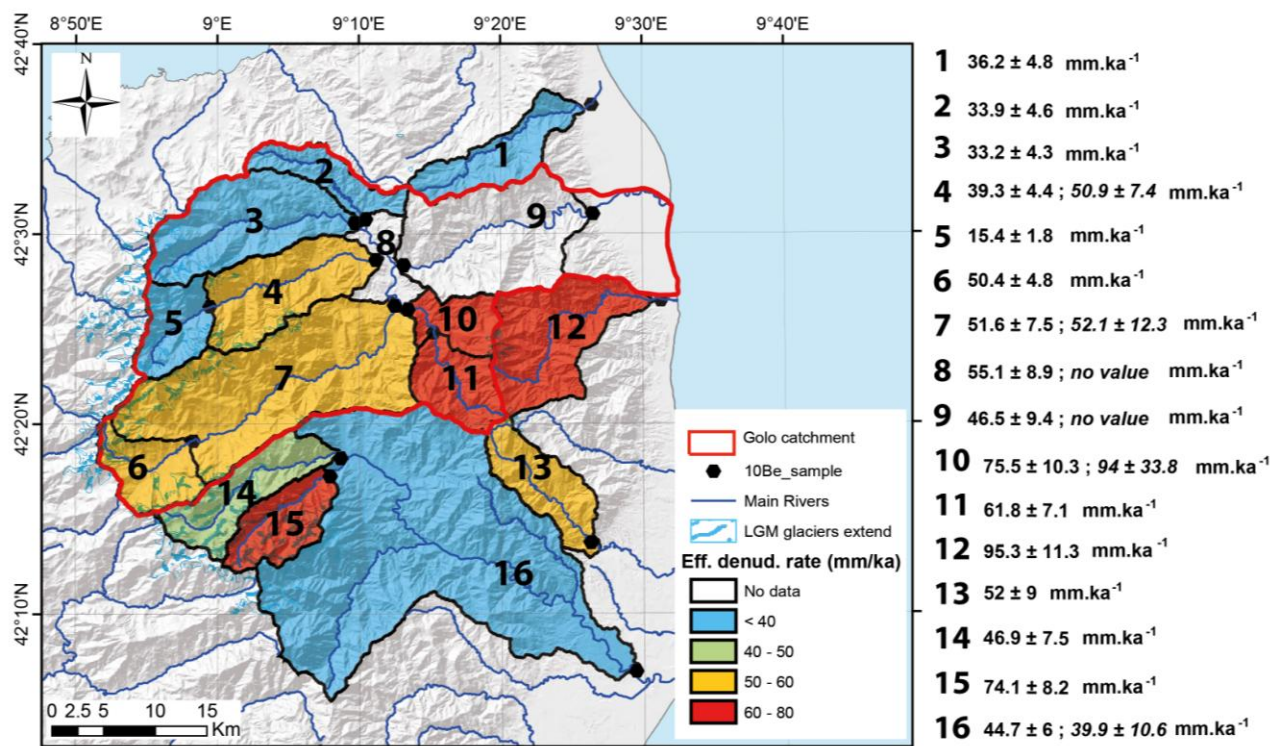


1014
 1015 Fig. 1: The geomorphological and geological setting of Corsica. A) Topographic map. B)
 1016 Geological map (from Chantraine *et al.*, 1996). LGM glaciers extend from Kuhlemann *et al.*,
 1017 (2005b). Names of studied catchments: 1) Bevinco; 2) Lagani; 3) Tartagine; 4) Asco; 5) Asco
 1018 glacial-shaped upstream; 6 to 9) Golo with 6) Golo glacial-shaped upstream; 7) Golo
 1019 upstream; 8) Golo after confluence with main tributaries; 9) Golo downstream; 10) Casaluna;
 1020 11) Casaluna upstream; 12) Fium Alto; 13) Bravona; 14) Tavignano upstream; 15)
 1021 Restonica; 16) Tavignano downstream.



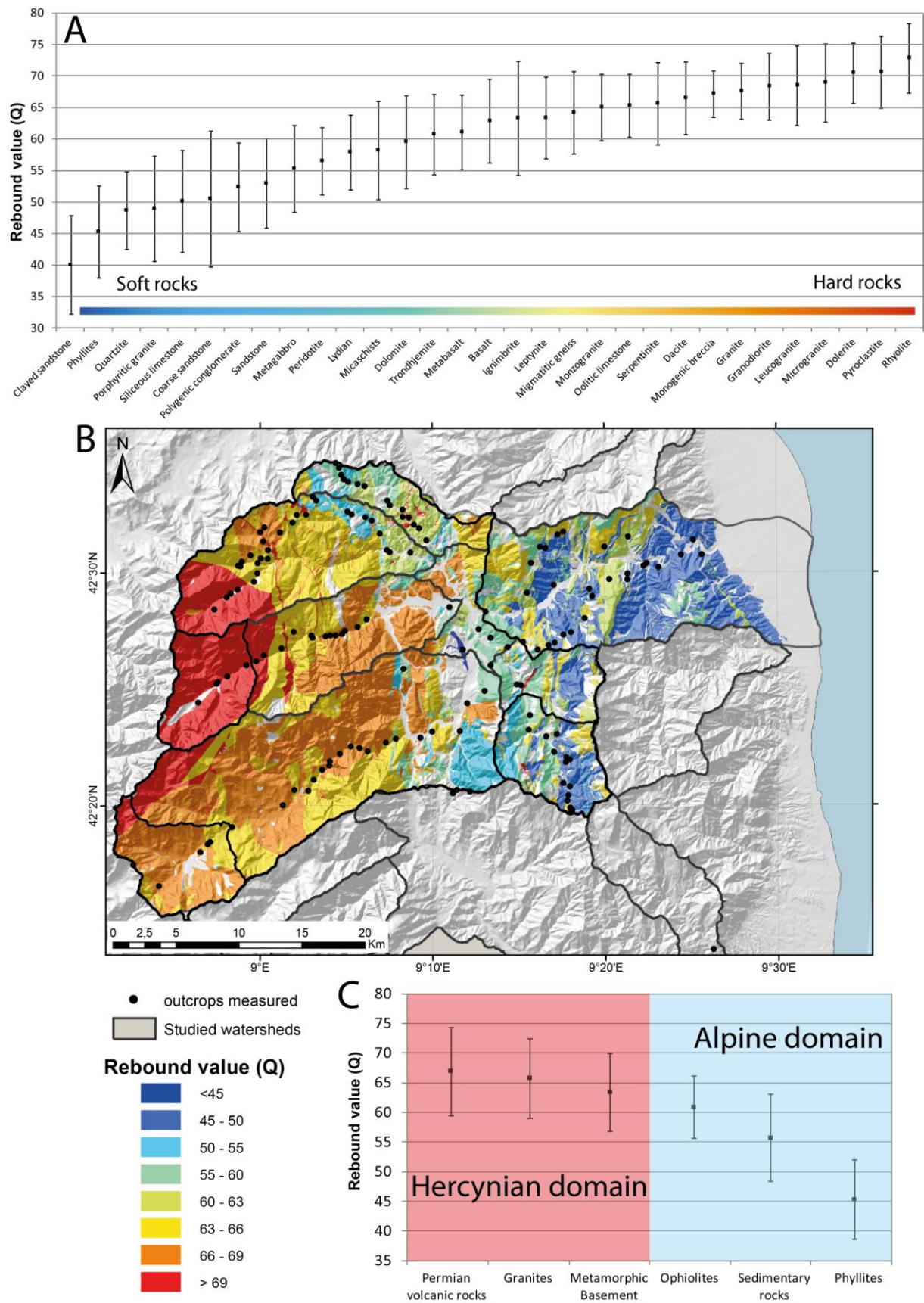
1022

1023 Fig. 2: Geomorphic parameters of the northeastern Corsica: A) Slope map and distribution of
 1024 normalized steepness index (Ksn) values of main streams. B) Map of local relief. C)
 1025 Longitudinal profiles parameters of the studied Rivers. C1) Longitudinal profiles with respect
 1026 to the distance to the coast. C2) Normalized longitudinal profiles. This representation aims to
 1027 provide a better comparison of the general shape between profiles. C3) Hypsometric curves
 1028 of each studied catchments. See Fig. 1 caption for the catchment number corresponding to
 1029 each name.



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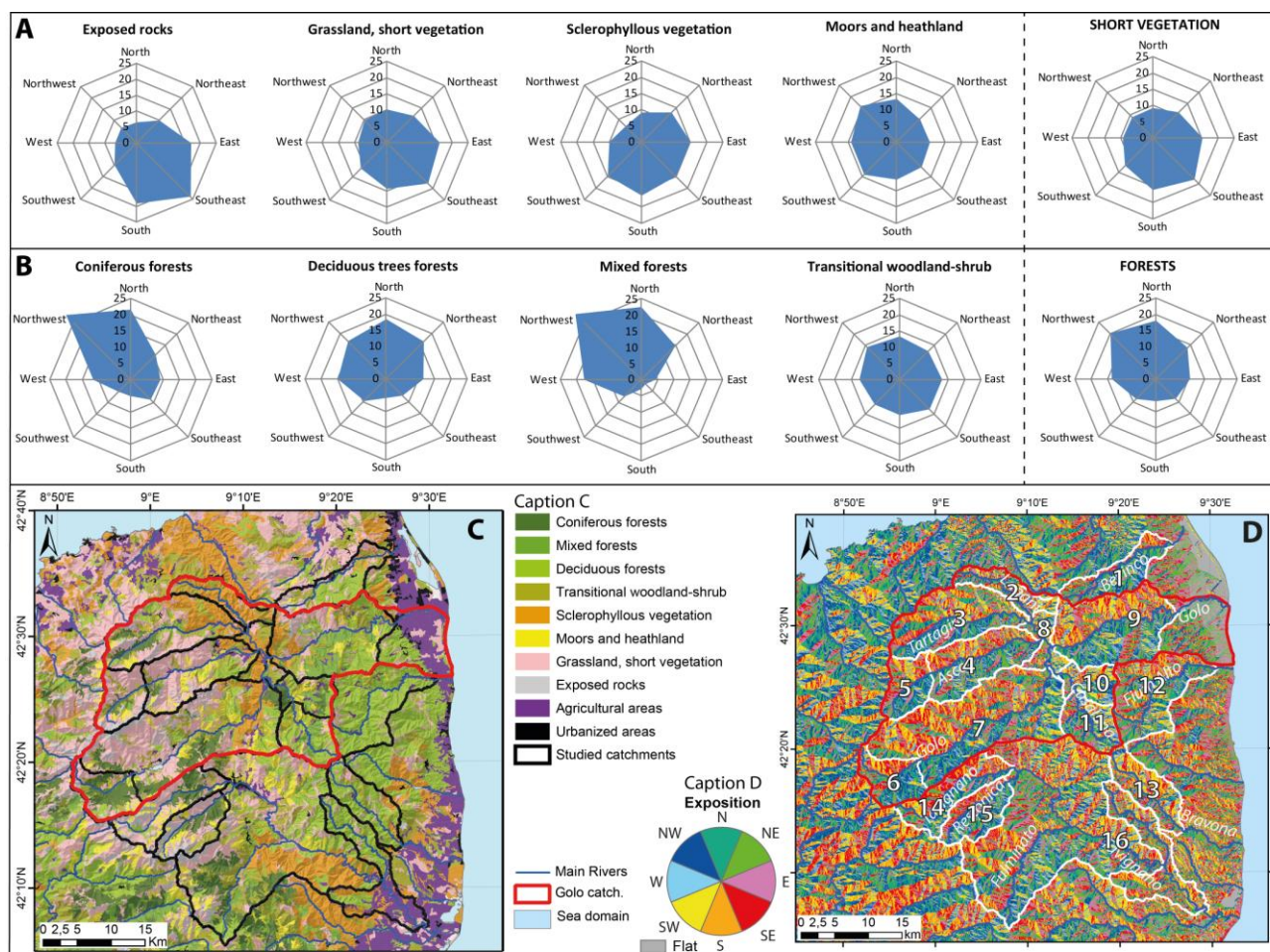
1031 Fig. 3: Repartition of catchment-wide millennial-scale denudation rates deduced from in situ-
 1032 produced ¹⁰Be cosmogenic nuclide concentration on present-day stream sediments. See Fig.
 1033 1 caption for the catchment number corresponding to each name. Denudation rates directly
 1034 deduced from ¹⁰Be concentrations are in regular form. Effective denudation rates resulting
 1035 from the subtraction of nested sub-catchments sediment fluxes are in italic form.



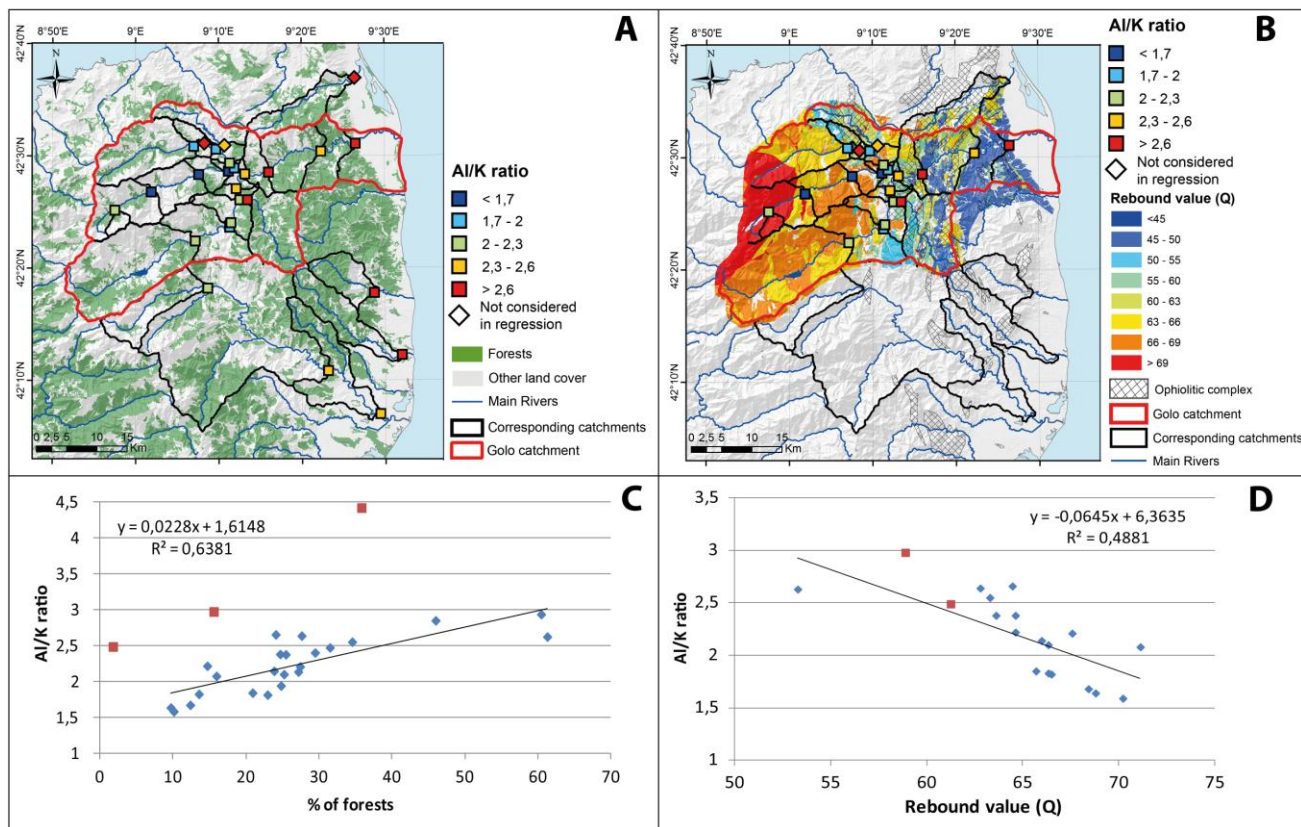
1036

1037 Fig. 4: Schmidt Hammer results and repartition of rock strength in the Golo catchment. A)
 1038 Mean rebound value (Q) interpreted from Schmidt Hammer results for each type of rock

1039 tested. Rock types are arranged from softer to the left to harder to the right. B) Map of rock
 1040 strength interpolated from mean rebound value (Q) determined for each type of rock and
 1041 location of in-situ measurements. C) Mean rebound value for each structural unit. The Alpine
 1042 domain is composed by the three softer structural units which are phyllites, sedimentary
 1043 rocks, and ophiolites. The Hercynian domain is composed by the three harder structural
 1044 units, which are metamorphic rocks from basement, granites, and Permian acid volcanism.

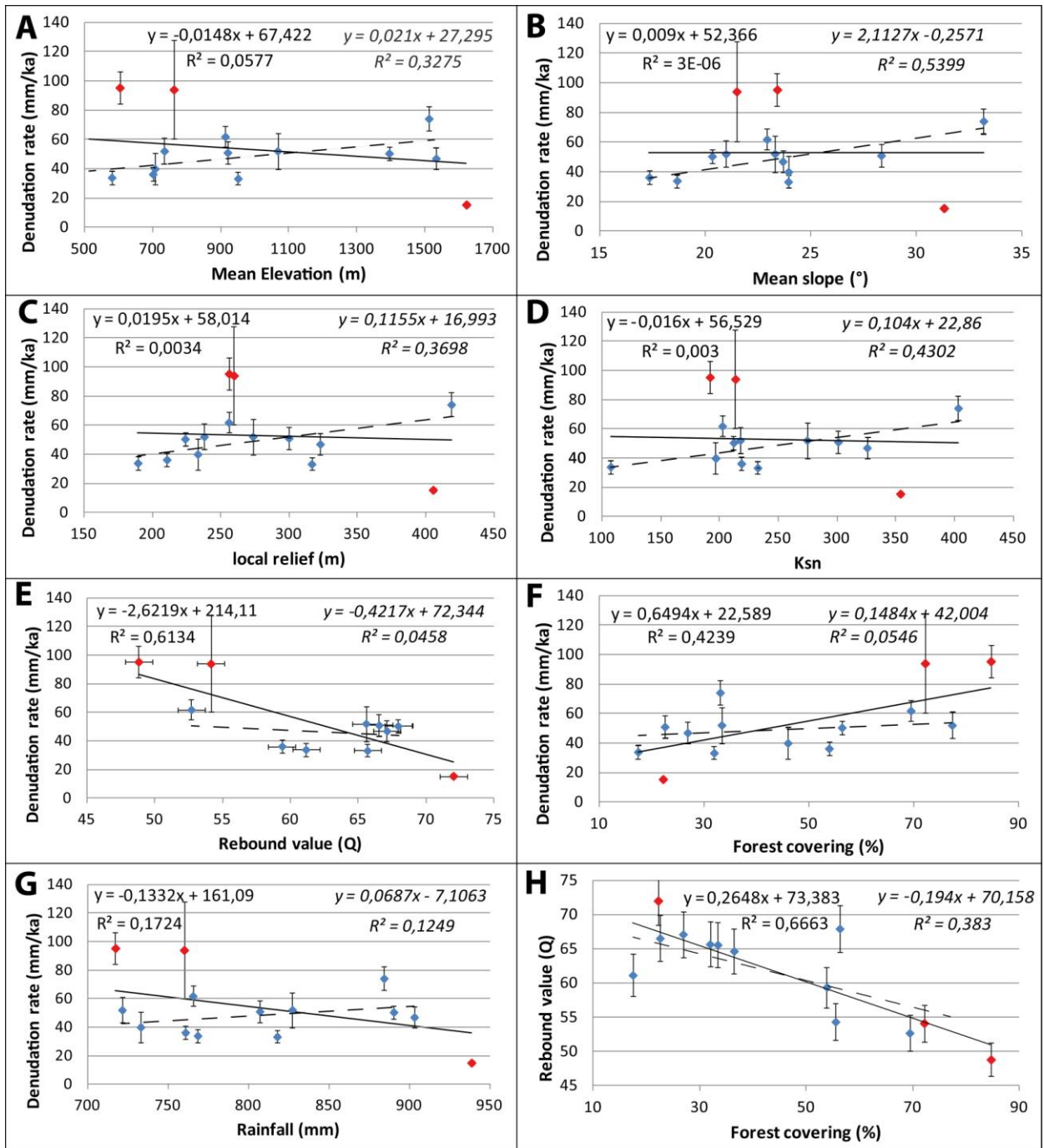


1045
 1046 Fig. 5: Landcover and relation with exposition. A) Repartition as a function of the land
 1047 exposition for each short vegetation type. B) Repartition in function of the land exposition
 1048 for each elaborated (trees) vegetation types. For A) and B) the graphs on the farther right
 1049 correspond to compilations for short vegetation and all forest types, respectively. Short
 1050 vegetation developed preferentially on S- to E-exposed lands whereas forests are
 1051 preferentially located on N- to W-exposed lands. C) Repartition of vegetation in the studied
 1052 area. Data comes from the “Corine Land Cover” European program database
 1053 (<http://www.statistiques.developpement-durable.gouv.fr/donnees-ligne/li/1825.html>). D) Map
 1054 of land exposition direction. The land is considered flat when slope is $<0.1^\circ$.



1055

1056 Fig. 6: AI/K measurements: location, results, and comparisons with parameters influencing
 1057 weathering. A) Location of river sediment samples, corresponding catchments, and AI/K
 1058 value repartition on the vegetal cover map (from “Corine Land Cover” European program
 1059 database). B) Location of river sediment samples, corresponding catchments, and AI/K value
 1060 repartition on the geological map (from Chantraine *et al.*, 1996). C) Relationship between
 1061 AI/K ratio and percentage of forests in the corresponding catchment. D) Relationship
 1062 between AI/K ratio and mean rebound value (Q) of bedrock in the corresponding catchment.
 1063 Data only concerns the Golo catchment since we do not have rebound values for the others.
 1064 For C) and D), red squares correspond to samples located in ophiolitic domain. They are not
 1065 considered in the linear regression calculation.



1066

1067 Fig. 7: Comparisons of effective ^{10}Be -derived denudation rates with morpho-climatic,
 1068 lithologic, and land-cover parameters. A) Mean catchment elevation. B) Mean catchment
 1069 slope. C) Catchment local relief. D) Mean normalized steepness index (Ksn). E) Rock
 1070 strength quantified from mean rebound value (Q). F) Percentage of forests in the catchment
 1071 area. G) Annual rainfall. H) Relationship between rock strength and forests. Samples in red
 1072 correspond to catchments with extreme values (5, 10, 12). Linear regressions considering all
 1073 samples are represented by the continuous straight line and regular-type equation and R^2
 1074 value (left of each graph). Linear regressions concerning only the blue samples are
 1075 represented by the dashed straight line and italic-type equation and R^2 value (right of each

1076 graph). Rock strength and percentage of forests present at first order a moderate correlation
1077 with denudation rates, which suggests control by weathering processes (regolith
1078 development being more efficient on lands covered by forests and constituted by softer
1079 rocks). Note that forests seem more developed on softer rocks (G). Without catchments 5,
1080 10, 12 to which extreme values of the denudation rate and rock strength are associated, a
1081 moderate correlation with morphometric parameters can be highlighted.

1082

Catch.	Name	Area km ²	Diam km	Sh. Fact.	Hypso. Int.	Sinuos.	Str. Slope	Str. Length km	Drop m	Drain. Dens. km.km ⁻²	X _{out} RGF93	Y _{out} RGF93	Z _{out} m	Strahler order	Z _{mean} m	Z _{max} m	Loc. Rel. m	Slope °	Sl. basin m/m	Rainfall mm.yr ⁻¹	Temp °C	Disch. m ³ .s ⁻¹	Ks	Theta	Ksn 0.5	Forests %
1	Bevinco	67.5	18.64	0.441	0.475	1.249	0.0524	21.89	1147	4.197	1227860	6189210	19	6	700.9	1458	210.6	17.3	0.3123	760.7	11.89	0.63	9.2	0.30	123.3	53.8
2	Lagani	47.2	16.32	0.421	0.267	1.481	0.0237	18.67	440	3.824	1207640	6176760	261	6	579.9	1456	189.4	18.7	0.3377	768.4	12.64		11.9	0.35	64.4	17.4
3	Tariagine	135.7	21.46	0.543	0.332	1.277	0.0435	25.54	1111	3.807	1207090	6175910	241	6	951.0	2370	316.6	23.9	0.4436	817.8	10.75		57.1	0.41	187.9	31.9
4	Asco	164.2	26.52	0.483	0.381	1.243	0.0490	30.79	1508	4.460	1209590	6173390	205	6	1150.9	2681	335.4	28.3	0.5390	851.6	9.57		183.7	0.47	256.5	22.4
5	Asco up.	53.8	11.83	0.620	0.431	1.406	0.1239	10.07	1248	4.756	1193262	6167162	812	6	1622.2	2681	405.4	31.3	0.6080	937.8	7.04		332.0	0.50	262.6	22.2
6	Golo up.	94.7	13.84	0.703	0.368	1.486	0.0774	11.47	888	4.034	1192912	6154212	837	6	1395.3	2696	223.9	20.3	0.3703	890.3	8.37		75.5	0.44	161.0	56.2
7	Golo up.	336.4	33.80	0.543	0.367	1.580	0.0411	47.28	1941	4.049	1210510	6169460	214	6	1125.6	2696	265.1	23.3	0.4305	837.7	9.73	2.94	29.8	0.36	164.0	37.3
8	GoloHercy	816.5	41.85	0.683	0.328	1.470	0.0304	47.64	1447	4.078	1211312	6172562	180	8	1004.2	2696	277.7	23.6	0.4368	821.9	10.34		202.7	0.50	185.4	36.4
9	Golo	983.0	53.61	0.585	0.338	1.531	0.0271	78.01	2134	4.079	1229340	6179060	21	8	926.0	2696	252.6	23.2	0.4290	807.0	10.75	14.4	45.8	0.39	175.1	40.0
10	Casaluna	99.8	16.51	0.605	0.402	1.382	0.0408	22.12	903	3.904	1210640	6169840	212	6	831.6	1752	253.0	21.5	0.3936	760.0	11.23		28.1	0.37	153.4	70.9
11	up.	57.3	11.36	0.666	0.426	1.478	0.0713	15.62	1114	3.909	1215037	6161587	336	5	913.3	1752	255.8	22.9	0.4227	765.4	69.39		132.5	0.47	199.6	69.4
12	Fium Alto	123.4	18.12	0.613	0.336	1.524	0.0352	24.51	863	4.142	1236060	6171160	19	6	603.5	1758	256.1	23.4	0.4328	717.4	12.54	1.34	147.2	0.48	179.3	84.6
13	Bravona	68.0	16.38	0.503	0.398	1.445	0.0660	21.59	1425	4.046	1232740	6146410	79	6	734.3	1719	237.9	21.0	0.3833	721.4	11.65	0.82	20.3	0.35	113.3	77.2
14	Tavignano	80.8	18.68	0.481	0.583	1.362	0.0543	24.18	1312	4.203	1206840	6153940	437	5	1533.0	2320	322.8	23.7	0.4382	903.1	7.62		16.4	0.32	186.8	26.8
15	Restonica	65.3	15.72	0.514	0.501	1.172	0.0783	16.66	1305	4.402	1207410	6153190	413	5	1512.0	2609	418.8	33.2	0.6537	884.2	8.01		81.1	0.40	263.8	33.0
16	Tavignano	585.3	40.69	0.595	0.364	1.619	0.0263	65.17	1712	4.067	1230090	6141110	40	7	975.1	2609	260.3	23.9	0.4438	788.8	10.57	11.7	33.3	0.37	178.9	42.5
17	Fulminato	161.9	17.69	0.719	0.383	1.567	0.0612	25.58	1564	3.994	1215210	6145540	201	7	1121.0	2597	322.6	27.4	0.5182	813.8	9.74		43.7	0.38	234.5	58.1

1086 Table1: Geomorphic and climatic parameters of the studied catchments and rivers. Row
 1087 captions from left to right are: Catchment number; catchment name; drainage area; diameter
 1088 of the catchment; shape factor; hypsometric integral; mean catchment sinuosity; mean main
 1089 stream slope; main stream length; gap between the highest and the lowest elevation in the
 1090 catchment; mean drainage density; latitude of the outlet; longitude of the outlet; elevation of
 1091 the outlet; Strahler order at the outlet; mean catchment elevation; maximum catchment
 1092 elevation; local relief; mean catchment slope in degree; mean catchment slope; mean annual
 1093 rainfall; mean annual temperature; mean annual discharge; steepness index; curvature
 1094 index; normalized steepness index (curvature is normalized); percentage of forests.

Catch.	River	Sample name	Area km ²	Y WGS 84	X WGS 84	Elev. m	Qz mass g	Measure ratios ^a ¹⁰ Be/ ⁹ Be (x 10 ⁻¹⁴)
1	Bevinco	BV79	67.5	9.441	42.613	19	8.323	4.957 ± 0.44
2	Lagani	BV48	47.2	9.174	42.513	261	20.451	13.012 ± 1.188
3	Tartagine	BV47	135.7	9.162	42.509	241	7.696	6.57 ± 0.555
4	Asco	BV43	164.2	9.186	42.478	205	10.248	8.668 ± 0.573
5	Asco	BV33	53.8	8.992	42.436	812	6.987	19.229 ± 1.366
6	Golo	BV15	94.7	8.971	42.319	837	14.899	11.41 ± 0.582
7	Golo	BV41b	336.4	9.208	42.438	214	20.917	13.389 ± 1.352
8	Golo	BV63	816.5	9.219	42.473	180	19.445	10.732 ± 1.256
9	Golo	BV65	983.0	9.441	42.518	21	7.799	5.033 ± 0.788
10	Casaluna	BV50	99.8	9.224	42.434	212	18.187	6.53 ± 0.596
11	Casaluna	BV76	57.3	9.254	42.413	336	17.955	8.084 ± 0.564
12	Fium Alto	BV34	123.4	9.471	42.445	19	16.432	3.914 ± 0.285
13	Bravona	BV84	68.0	9.438	42.230	79	15.790	7.099 ± 0.906
14	Tavignano	BV61	80.8	9.146	42.305	437	19.273	17.28 ± 1.996
15	Restonica	BV62	65.3	9.131	42.289	413	23.810	13.224 ± 0.871
16	Tavignano	BV36	585.3	9.488	42.116	40	18.452	10.959 ± 0.981

1095

Table II. Continued

Catch.	Blanks ¹⁰ Be/ ⁹ Be (x 10 ⁻¹⁴)	⁹ Be carrier x 10 ⁻⁶ g	¹⁰ Be conc. 10 ³ at.g(SiO ₂)	Prod. rate ^b at.g ⁻¹ .yr ⁻¹	Shielding	Pressure Mpa	Density kg.m ⁻³
1	0.306 ± 0.056	303.5	113.5 ± 10.8	5.92 ± 0.26	0.849	959.1	2.7
2	0.305 ± 0.064	305.0	126.2 ± 11.5	6.03 ± 0.27	0.961	945.7	2.7
3	0.222 ± 0.064	305.6	166.9 ± 14.1	8.11 ± 0.36	0.940	905.1	2.7
4	0.222 ± 0.064	305.9	167.4 ± 11.1	9.78 ± 0.44	0.945	884.1	2.7
5	0.306 ± 0.056	305.5	553.6 ± 40.0	13.18 ± 0.59	0.930	833.5	2.7
6	0.306 ± 0.056	304.5	151.9 ± 8.0	11.49 ± 0.51	0.957	856.9	2.7
7	0.305 ± 0.064	305.1	127.1 ± 12.8	9.71 ± 0.43	0.962	886.5	2.7
8	0.305 ± 0.064	304.8	108.7 ± 12.7	8.8 ± 0.39	0.956	899.6	2.7
9	0.305 ± 0.064	305.0	122.1 ± 19.1	8.31 ± 0.37	0.957	908.2	2.7
10	0.305 ± 0.064	304.8	69.1 ± 6.3	7.5 ± 0.34	0.967	917.7	2.7
11	0.306 ± 0.056	304.3	88.2 ± 6.4	7.9 ± 0.35	0.958	908.6	2.7
12	0.306 ± 0.056	304.7	44.8 ± 3.6	6.01 ± 0.27	0.931	943.4	2.7
13	0.306 ± 0.056	305.3	87.9 ± 11.7	6.54 ± 0.29	0.902	928.8	2.7
14	0.222 ± 0.064	305.3	180.2 ± 20.8	12.81 ± 0.57	0.953	842.8	2.7
15	0.222 ± 0.064	306.6	111.5 ± 7.3	12.48 ± 0.56	0.936	845.7	2.7
16	0.305 ± 0.064	304.8	117.1 ± 10.5	7.61 ± 0.34	0.936	922.2	2.7

1096

1097 Table 2: ¹⁰Be concentrations and derived denudation rates.

1098 a) The obtained ¹⁰Be/⁹Be ratios are corrected for procedural blanks and calibrated
 1099 against the National Institute of Standards and Technology standard reference
 1100 material 4325 by using an assigned value of 2.79 ± 0.03x10⁻¹¹ and a ¹⁰Be half-life

1101 of $1.387 \pm 0.012 \times 10^6$ years (Korschinek *et al.*, 2010; Chmeleff *et al.*, 2010).
1102 Analytical uncertainties (reported as 1σ) include uncertainties associated with
1103 AMS counting statistics, chemical blank measurements and AMS internal error
1104 (0.5%). Long-term AMS measurements of procedural blanks yield a background
1105 ratio of $3.0 \pm 1.5 \times 10^{-15}$ for $^{10}\text{Be}/^9\text{Be}$ (Arnold *et al.*, 2010). The contribution of muons
1106 was calculated using the physical parameters evaluated by Braucher *et al.* (2011).

1107 b) Catchment-averaged production rate is calculated by averaging the values of
1108 quartz-producing rocks with a sea-level high-latitude (SLHL) spallation production
1109 of 4.03 ± 0.18 at $\text{g}^{-1} \text{yr}^{-1}$ scaled following Stone's (2000) method. The catchment-
1110 averaged topographic shielding factor is computed from Dunne *et al.* (1999) using
1111 algorithms modified from Balco (2001).

1112 c) Integration time corresponds to the time spent in the equivalent mean attenuation
1113 length (~ 60 cm taking into account a material density of $2.7 \text{ kg}\cdot\text{m}^{-3}$) (e.g. Lal *et al.*,
1114 1991).

1115 d) Maximum ice cover in percentage considering the maximum integration period
1116 computed using the Krumrai (2009) maps of the Würmian ice extent in Corsica.

1117 e) Effective denudation rates obtained subtracting the sediment fluxes from sub-
1118 basins nested in the parent basins and lacking the sub-catchment part
1119 overlapping, as described by Granger *et al.* (1996) and Portenga *et al.* (2015).

1120 f) Environmental parameters are computed to be compared to effective denudation
1121 rates, i.e. without overlapping between catchments.

ID	Latitude	Longitude	River name	% forest	Al/K	Q Schmidt
66	42.374	9.119	Golo	27.4	2.21	67.57
68	42.401	9.191	Golo	25.2	2.10	66.33
41	42.435	9.209	Golo	27.2	2.14	65.98
42	42.451	9.202	Golo	34.5	2.55	63.29
63	42.473	9.219	Golo	24.7	2.38	64.62
64	42.476	9.267	Golo	24.1	2.66	64.46
78	42.507	9.372	Golo	25.5	2.38	63.61
65	42.518	9.442	Golo	27.6	2.64	62.79
50	42.435	9.224	Casaluna	61.3	2.63	53.29
45	42.514	9.117	Tartagine	23.0	1.82	66.49
47	42.510	9.161	Tartagine	20.9	1.85	65.69
49	42.489	9.189	Tartagine	14.7	2.22	64.62
69	42.447	9.032	Asco	10.1	1.59	70.21
70	42.473	9.126	Asco	9.7	1.64	68.79
43	42.477	9.186	Asco	12.4	1.68	68.42
44	42.482	9.198	Asco	13.6	1.83	66.34
30	42.420	8.958	Manica	16.0	2.08	71.12
67	42.394	9.189	Tiolata	24.8	1.94	54.07
40	42.297	9.479	Alesani	60.4	2.94	
39	42.204	9.534	Bravone	46.0	2.85	
61	42.304	9.145	Tavignano	23.8	2.15	
37	42.181	9.386	Tavignano	29.5	2.40	
36	42.117	9.490	Tavignano	31.5	2.48	
48	42.513	9.174	Lagani	1.8	2.49	61.25
46	42.518	9.138	San Pietro	15.6	2.98	58.89
79	42.613	9.441	Bevinco	35.8	4.42	

1122

1123 Table 3: Location of samples used for the Al/K ratios determination. ID is the name of the
1124 sample. Percentages of forest and rebound values (Q) are also reported for each
1125 corresponding catchment. The analytical error associated with the Al/K ratio is lower than
1126 0.05% (2 σ).

1127