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1 **Tectonic and climatic control on terrace formation: coupling in situ produced ^{10}Be depth**
2 **profiles and luminescence approach, Danube River, Hungary, Central Europe**

3
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24
25
26 **Abstract**

27 The terrace sequence of the Hungarian part of the Danube valley preserves a record of
28 varying tectonic uplift rates along the river course and throughout several climate stages. To
29 establish the chronology of formation of these terraces, two different dating methods were
30 used on alluvial terraces: exposure age dating using in situ produced cosmogenic ^{10}Be and
31 luminescence dating. Using Monte Carlo approach to model the denudation rate-corrected
32 exposure ages, in situ produced cosmogenic ^{10}Be samples originated from vertical depth
33 profiles enabled the determination of both the exposure time and the denudation rate. Post-IR

34 IRSL measurements were carried out on K-feldspar samples to obtain the ages of
35 sedimentation.

36 The highest terrace horizon remnants of the study area provided a best estimate erosion-
37 corrected minimum ^{10}Be exposure age of >700 ka. We propose that the abandonment of the
38 highest terrace of the Hungarian Danube valley was triggered by the combined effect of the
39 beginning tectonic uplift and the onset of major continental glaciations of Quaternary age
40 (around MIS 22). For the lower terraces it was possible to reveal close correlation with MIS
41 stages using IRSL ages. The new chronology enabled the distinction of tIIb (~90 ka; MIS 5b-
42 c) and tIIIa (~140 ka; MIS 6) in the study area. Surface denudation rates were well
43 constrained by the cosmogenic ^{10}Be depth profiles between 5.8 m/Ma and 10.0 m/Ma for all
44 terraces. The calculated maximum incision rates of the Danube relevant for the above
45 determined >700 ka time span were increasing from west (<0.06 mm/a) to east (<0.13 mm/a),
46 toward the more elevated Transdanubian Range. Late Pleistocene incision rates derived from
47 the age of the low terraces (~0.13-0.15 mm/a) may suggest a slight acceleration of uplift
48 towards present.

49

50 Keywords: cosmogenic ^{10}Be exposure age, depth profiles, denudation rate, post-IR IRSL,
51 river terrace, incision rate, uplift rate, Quaternary

52

53 **1. Introduction**

54

55 The age and position of a fluvial terrace with respect to a reference level provides a good
56 approximation of the river incision rate. In certain occasions this is a valid time-averaged
57 proxy for the surface uplift rate. On the other hand, climatically induced changes in river style
58 can also lead to terrace formation (Bridgland, 2000; Peters and Van Balen, 2007; Bridgland
59 and Westaway, 2008; Gibbard and Lewin, 2002, 2009; Warner, 2012). Incision/uplift rates
60 calculated on the basis of age determination of river terraces in Europe mostly suggest values
61 up to ~1 mm/a for middle and late Pleistocene times (Brocard et al., 2003; Peters and van
62 Balen, 2007; Viveen et al., 2012; Necea et al., 2013; Rixhon et al, 2011, 2014) and
63 exceptionally, in tectonically active regions, like the SE Carpathians (Necea et al., 2013) or
64 for shorter periods (Antón et al., 2012) present higher rates. Burial age determination of cave
65 sediments using cosmogenic ^{10}Be and ^{26}Al in Switzerland suggested an incision rate of 0.12
66 mm/a for the early Pleistocene, and a tenfold increase was suggested for middle to late
67 Pleistocene times Häuselmann et al. (2007a). Applying the same method Wagner et al. (2010)

68 calculated rather low average bedrock incision rates of 0.1 mm/a for the last 4 Ma with a
69 decreasing trend towards present (Table 1). However, the decrease towards younger times
70 was attributed to the rise of the base level due to sediment aggradation within the valley.

71 The uplift rate of 1.75 mm/a provided by geodetic measurements in the western and
72 northern Alpine foreland (Ziegler and Dézes, 2007) represents the higher end of the uplift
73 rates based on geochronological and geomorphological constraints (Table 1)

74 The combined application of cosmogenic ^{10}Be exposure dating and luminescence dating
75 allows a more robust age determination than using either method separately, as it was
76 demonstrated by some previous studies (Anders et al, 2005; Delong and Arnold, 2007;
77 Guralnik, 2011; Viveen et al., 2012). Main objective of the present study is to provide age
78 constraints to fluvial terraces of the Danube River in order to determine its incision rate. New
79 constraints on vertical neotectonic deformation of the western Pannonian Basin and on the
80 role of climate change in terrace evolution are presented. Denudation rates provided by ^{10}Be
81 depth profiles will contribute to a better understanding of the pace of surface processes in the
82 region.

83

84 **2. Geological and geomorphological setting**

85

86 *2.1. Quaternary tectonics and drainage pattern evolution*

87

88 The Pannonian Basin (Fig. 1) represents a back-arc basin formed by Miocene crustal
89 thinning and subsequent “post-rift” thermal subsidence (Horváth and Royden, 1981, Horváth
90 et al. 2015), which led to the formation of the several 100 m deep Lake Pannon. This lake was
91 filled up during the late Miocene due to large sediment input (Magyar et al., 2007, 2013). The
92 Western Pannonian Basin was subsequently occupied by a fluvio-lacustrine system,
93 established in the latest Miocene (Uhrin et al., 2011), with paleo-rivers drained towards the
94 south (Szádeczky-Kardoss 1938, 1941; Pécsi, 1959; Gábris and Nádor, 2007). Neotectonic
95 shortening and related uplift progressively shifted from SW (Slovenia) toward NE (Tari,
96 1994; Fodor et al. 2005). The deflection of the Alpine and Carpathian rivers (including the
97 paleo-Danube) from their southerly flow towards the east, their current runoff direction,
98 followed the propagation of the neotectonic deformation. As a result, the formation of a large
99 alluvial fan in the Danube Basin was proposed during the early Pleistocene (Szádeczky-
100 Kardoss, 1938, 1941) (Fig. 2), and the Danube was forced to find its way towards the east

101 across the TR. The age of abandonment of the highest terrace of the Danube provides a good
102 approximation of the age of the onset of fold-related uplift in the northeast part of the TR.

103 Neotectonic structures developed due to changing boundary conditions of the deformation:
104 one of the major driving forces, the eastward pull (subduction rollback) beneath the Eastern
105 Carpathians largely decreased while northward push of the Adriatic plate continued (Horváth,
106 1995, Bada et al., 2006). In the new neotectonic regime transpression resulted in the
107 reactivation of strike slip faults and large-scale folding of the Pannonian lithosphere
108 manifested by the coexistence of several uplifting regions and areas of subsidence (Horváth
109 and Cloetingh 1996, Bada et al., 2006, Ruzkiczay-Rüdiger et al., 2007; Dombrádi et al.,
110 2010). One of these uplifting areas is the SW-NE trending, 300-700 m above sea level (asl)
111 high Transdanubian Range (TR) emerging between the Danube Basin and the Great
112 Hungarian Plain, two lowlands of continuous sediment accumulation since middle Miocene
113 times (Fig.1).

114 The Danube River is the only one cutting through the uplifting basement unit of the TR
115 (Fig. 1). The differential uplift across the TR, i.e. gradually decreasing uplift rates from its
116 axis towards the neighbouring lowlands, led terrace remnants in the Danube valley to perform
117 an upwarped pattern relative to the modern river profile (Fig. 2; Pécsi, 1959; Gábris, 1994;
118 Ruzkiczay-Rüdiger et al., 2005a; Gábris and Nádor, 2007; Gábris et al. 2012 and references
119 therein). As a consequence, number and position of terraces alter along the Danube and
120 geomorphic horizons of the same age may occur at different elevations across the uplifting
121 range. Moreover, this setting enabled the simultaneous development of fill terraces in the
122 margins, and of strath terraces in the central part of the TR; rising additional difficulty in
123 correlating terrace horizons along the river.

124 The uplift rates inferred from the published terrace chronological data (Ruzkiczay-
125 Rüdiger et al., 2005a) along the Hungarian segment of the Danube valley varied between 0.05
126 and 0.36 mm/a at the margins and at the axis of the TR, respectively. ³He surface exposure
127 dating of strath terraces at the Danube Bend (Fig. 1) suggested a maximum incision rate of 1.6
128 mm/a at the valley section of most intensive uplift, at the NE-SW trending axis of the
129 Transdanubian Range (TR) (Ruzkiczay-Rüdiger et al., 2005b). However, according to GPS
130 velocity measurements the maximum uplift rate in the TR does not exceed 0.5 mm/a
131 (Grenerczy et al., 2005 and pers. comm. 2014).

132 Our study area, the Győr-Tata terrace region (GTT) is located in the south-eastern margin
133 of the Danube Basin, in the transitional zone of the subsiding area towards the uplifting TR
134 (Figs 1, 2A, 3). Although the regional structural and geomorphological context of the

135 transitional zone between the Danube Basin and the TR is relatively clear, the local structural
136 setting of the GTT has not been clearly documented. The axis of uplift of the TR is NE-SW
137 (Fig. 1) which is oblique to the present (and postulated Quaternary) direction of compression
138 derived from stress data (Horváth and Cloetingh, 1996; Bada et al. 2006). This gentle folding
139 would lead to westward tilt of the western TR foothills, including the study area. A
140 pronounced Late Miocene fault is present at the western margin of the TR (Fodor et al., 2005;
141 Bada et al., 2007; Dombrádi et al., 2010 Figs. 1, 2A), which could be reactivated in the
142 Quaternary. The abundant travertine occurrences (Pécsi, 1959, Ruszkiczay-Rüdiger et al.,
143 2005a; Kele, 2009; Sierralta et al., 2009) along proposed fault may be connected to this
144 reactivation. Our study area, the GTT is situated to the west from this possible Quaternary
145 fault (Figs. 1, 2A).

146 In the south-western part of the TR ^{10}Be exposure age of wind-eroded landforms was
147 determined up to 1.56 ± 0.09 Ma (Ruszkiczay-Rüdiger et al., 2011) (Fig.1). This age provides a
148 minimum time constraint on the onset of the inversion-related uplift in the southwest part of
149 the TR.

150 At the same time, during the early Pleistocene, the Danube was already deflected from its
151 southerly flow and formed a large alluvial fan in the subsiding Danube Basin (Fig. 2A;
152 Szádeczky-Kardoss, 1938, 1941). According to Pécsi (1959), in the slightly uplifted TR at
153 least 3 terraces (tVI–tIV) were formed during this period of time (Fig. 3). In the traditional
154 terrace chronology, the tVI horizon is considered to be the first terrace of Danubian origin in
155 the central TR. Recently, the origin of the horizons above tV has been questioned: they might
156 be remnants of surfaces pre-dating the appearance of the Danube River in the area
157 (Szeberényi, 2014). We suggest that the number and age of the terraces might be different
158 along the Danube valley because of the possibility of non-developed and/or completely
159 eroded terrace remnants at diverse valley sections.

160 When the inversion-related vertical movements propagated towards the north and
161 northeast, the Danube incised into the marginal areas of its alluvial fan and into the
162 underlying late Miocene fluvial and lacustrine sediments, formation of a terrace staircase
163 started at the western and eastern side of the Danube Basin (Figs. 1, 2A,B), while continuous
164 aggradation occurred in the basin interior of ongoing subsidence (Fig. 3). The highest terraces
165 at the western (Parndorf plateau) and eastern sides (GTT) of the Danube Basin (tIV; Figs. 2A.
166 and 3) are supposed to be remnants of this ancient alluvial fan (Szádeczky-Kardoss, 1938,
167 1941; Pécsi 1959).

168 Age determination of the Danube terraces at the south-eastern margin of the Danube Basin,
169 the GTT will provide time constraints on the propagation of fold-related uplift of the TR from
170 southwest towards the northeast

171

172 *2.2. Position and age of the terraces in the Győr-Tata terrace region (GTT)*

173

174 The ages of Hungarian river terraces have traditionally been correlated to different Alpine
175 glacial stages (Günz, Mindel, Riss Würm; Penck and Brückner, 1909) and are numbered from
176 the youngest/lowest horizon (tI, high floodplain) to the higher/older levels (Pécsi, 1959;
177 Kretzoi and Pécsi, 1982). These ages were based mostly on geomorphological and
178 sedimentological correlations of terraces combined by scattered large mammal findings and
179 rare numerical ages. These “traditional” terrace ages were quantified by Ruzkiczay-Rüdiger
180 et al. (2005a), and ages have been further refined by this study considering the studies of
181 (Lisiecki and Raymo, 2005; Gibbard and Van Kolfshoten, 2005; Nitychoruk et al., 2006;
182 Ehlers and Gibbard, 2007) providing new age constraints on the MIS stages of the Alpine
183 glacial terminology (Fig. 4).

184 Mean level of the Danube in the GTT is at 109-107 m asl, from west to east; based on
185 EOVI (Uniform National Projection of Hungary) 1:10 000 topographic maps. Relative terrace
186 heights (current elevation of the surface of the fluvial material) were calculated respectively
187 (Table 2).

188 Terraces in the GTT occur only on the southern side of the river. To the north, the river is
189 escorted by a wide alluvial plain extending to the southern flanks of the Carpathians (Figs.1,
190 2A). The floodplain on the Southern riverside is narrow, flood-free terraces are close to the
191 current river. According to the traditional terrace system of the Danube in Hungary (Pécsi,
192 1959; Kretzoi and Pécsi, 1982; Gábris, 1994; Ruzkiczay-Rüdiger et al., 2005a, Gábris and
193 Nádor, 2007) five terraces occur in the GTT.

194 The lowest terrace above the floodplain is tIIa is ~9 m above mean river level (amrl). In the
195 neighbouring Gerecse Hills U/Th dated travertines provided an age of 10-40 ka (Kele, 2009)
196 for this terrace, which is the most continuous horizon along the Danube. There are some Th/U
197 age data available on tIIb (increasing height from 12 m to 22 m amrl. from west to east),
198 clustering around 100-130 ka (summary in Ruzkiczay-Rüdiger et al 2005b, and Kele 2009;
199 Table 1A), however it is still not clear whether this horizon has been deposited by the end of
200 the penultimate glacial or at the beginning of the last glacial phase. Above this level terrace

201 chronology is even more uncertain. The tIII horizon is apparently missing in the study area, its
202 existence in this valley section has not been verified yet (Figs. 2A, B, 3, 4).

203 In the western part of the GTT elevation of the tIV horizon is at 35-40 m amrl, and it is an
204 extended flat surface covered by 8-10 m thick gravel. At the easternmost part of the region
205 only small remnants of this terrace horizon, covered by thinned terrace material (0-3m), have
206 remained at an elevation of 80-87 m amrl. According to the traditional terrace chronology
207 (Pécsi, 1959) the terrace level was formed during the Mindel glaciation, which ended after
208 MIS 12 (Gibbard and Kolfshoten, 2005; Lisiecki and Raymo, 2005), providing a minimum
209 terrace age of 420 ka.

210 Krolopp (1995) described an early to middle Pleistocene malacofauna from the eastern and
211 from the western parts of the tIV (Győr and Grébics, respectively; Fig.2A). The described
212 fauna belongs to the *Viviparus boeckhi biozone* and is indicative of fluvial environment under
213 mild, even warm climate. According to Krolopp (1995) this biozone started around the
214 beginning of the Quaternary (ca. 2.6 Ma) and ended with the Günz/Mindel interglacial, which
215 suggests a minimum age of 540 ka for this terrace (Fig. 4).

216 Gábris (2008) and Gábris et al. (2012) made an attempt to find a direct link between
217 terrace formation and global climate change and made a compilation of terrace incision and
218 marine oxygen isotope data (MIS), assuming that each incision event could be linked to a
219 MIS Termination. On this basis Gábris (2008) suggested younger ages to each terrace than it
220 has been inferred from the revision of the a traditional terrace chronology based on the Alpine
221 glaciations, and divided the tIII terrace into tIIIa and tIIIb horizons (Fig.4).

222 The malacofauna of similar age (Krolopp 1995) and sedimentological investigations
223 (heavy mineral composition, pebble lithologies and roundedness Pécsi, 1959) proposed that
224 the different segments of the tIV terrace of gradually increasing elevation belong to a single
225 terrace level and thus they share the same age (Fig. 2A, 3). This is supported also by the DEM
226 derived gentle slope of this horizon ($<0.1^\circ$) between Győr and Grébics, and is in agreement
227 with the tectonic setting of the GTT between the uplifting TR and the subsiding Danube Basin
228 (Figs. 1, 2A, 3).

229 Another interesting feature of the highest terrace of the GTT (tIV) is its appearance as
230 isolated terrace remnants bordered by steep slopes on both sides (Fig. 1, 2A,B). Towards the
231 Danube, the terrace riser is facing a lower (tIIb) terrace horizon. However, to the south,
232 similarly steep slope leads into a topographic depression with an elevation of 120-130 m
233 carved into the late-Miocene fluvial-lacustrine sedimentary sequence. We suggest that the
234 coarse alluvial sediments of the Danube have protected the surface of the tIV terrace from

235 denudation, while to the south of the former alluvial fan of the Danube the loose late Miocene
236 sediments were exposed. After the lowering of the erosion base by the incision of the Danube
237 these uncovered areas were object of faster denudation, and thus the terrace remnants became
238 topographically isolated.

239 Most important consequence of this topography for surface exposure dating is that the tIV
240 level has been disconnected from potential sediment sources, hence post abandonment
241 sediment accumulation could be disclosed. On the other hand, considerable amount of
242 denudation must have occurred, evidenced by the fact that the fine overbank deposits have
243 been stripped from above the coarse channel or point-bar facies (gravelly sand, sandy gravel)
244 exposed on the present terrace surface.

245

246

247 **3. Material and methods**

248

249 *3.1. Sampling strategy*

250

251 Terrace deposits of the study area consist of cross bedded sandy gravels, gravelly sands.
252 No silt or clay lenses were observed. Low terraces (tIIa and tIIb) were frequently covered by
253 1-1.5 m thick fine grained alluvial cover, aeolian sand or loess. The most frequent gravel
254 material was quartzite, which was the main target for the sampling in gravel layers. In most
255 cases, imbrication of the pebbles and bedding of the sand indicated original position of the
256 sediment. Obscure bedding or evidences of sediment mixing are detailed in the site
257 descriptions. Large cobbles and boulders were missing from all terraces and terrace material
258 was loose, not cemented.

259 The accumulation history the radioactive cosmogenic ^{10}Be in terrace sediments may be
260 highly complex. Its accumulation prior to deposition (inheritance), surface denudation, post-
261 depositional sediment mixing and aeolian sediment cover are processes that have to be
262 considered (Brocard et al., 2003; Braucher et al., 2009; Antón et al., 2012; Rixhon et al.,
263 2011).

264 Because of the high probability of surface denudation since terrace formation and unknown
265 amount of inherited cosmogenic nuclides, we use the cosmogenic ^{10}Be depth profile-approach
266 coupled with luminescence dating. The ^{10}Be depth profiles enable the determination of both
267 denudation rate and exposure age of the terraces by considering all particles, neutrons and
268 muons, involved in the production of ^{10}Be (Siame et al., 2004). In addition, the amount of

269 inherited ^{10}Be and the bulk density can also be estimated (Anderson et al., 1996; Brocard et
270 al., 2003; Siame et al., 2004; Braucher et al., 2003, 2009). An accurate age determination of
271 depositional surfaces can be achieved (Hedrick et al., 2013; Stange et al. 2013; 2014) as long
272 as ^{10}Be concentrations have not yet reached the steady state, with nuclide production and loss
273 (decay + denudation) in equilibrium. If the profile has reached secular equilibrium, long-term
274 denudation rate and minimum exposure age (integration time) of the profile can be
275 constrained (Lal, 1991; Matsushi et al., 2006). In case of post-depositional sediment mixing
276 no exponential decrease of ^{10}Be concentrations is observed with depth (Ward et al., 2005); in
277 that case, it is problematic to determine a terrace abandonment age using cosmogenic ^{10}Be .

278 Post IR-IRSL dating is suitable to date the time of sediment burial and provides
279 independent age control on the accumulation of terrace material. It is less sensitive of surface
280 processes than exposure age determination, but the useful time range is shorter and is not
281 applicable on gravels. Accordingly, the combined methodology is suitable for providing
282 robust time constraints on terrace formation.

283 Elevations and positions of the samples were recorded on the field using hand-held GPS
284 (WGS 84 reference datum) and cross referenced with 1:10 000 topographic maps (Table 2).
285 For sampling along depth profiles active or abandoned gravel pits with terrace material
286 exposed in at least 2 m depth in original position are suitable, which prerequisites were
287 fulfilled by only a limited number of locations. These were taken as sample locations of this
288 study (Fig.5).

289 The highest, tIV terrace horizon was sampled at three locations at increasing altitudes from
290 west to east (147m Győr; 153m Bana; 194m Grébics; Fig. 2A, 5, Table 2). The tIIIb level was
291 sampled at two locations, with somewhat different altitudes (121 m – Ács and 127 m –
292 Mocsá). 20-40 pebbles of 1-5 cm size were collected from each sampled depth. From sand
293 layers whole sand samples were collected (Figs. 5, 6). For site descriptions refer to
294 Supplementary Section 3.1.

295 One sample was collected from the recent gravel of the Danube at Ács (Dan08-27). This
296 sample consisted of quartzite pebbles, which were processed to measure ^{10}Be concentration of
297 the current load of the river. At this location the terrace IIIb lies next to the actual riverbed
298 (Fig. 2A,B).

299

300

301 *3.2. Cosmogenic ^{10}Be exposure age dating: theory and laboratory analyses*

302

303 3.2.1. Theoretical considerations

304

305 During the last decade Terrestrial in situ produced Cosmogenic Nuclides (TCN) became a
 306 widely used technique for dating geomorphic surfaces and determining denudation rates
 307 (Bierman; 1994; Cerling and Craig, 1994; Rixhon et al, 2011; Portenga and Bierman, 2011).
 308 Terrestrial in situ produced cosmogenic ^{10}Be is produced within quartz mineral lattice near the
 309 Earth's surface (Lal, 1991). Quartz is a mineral abundant in most terraces and datable time
 310 span using in situ cosmogenic ^{10}Be covers the entire Quaternary (Lal, 1991; Gosse and Philips
 311 2001, Dunai, 2010; Hancock et al., 1999.; Brocard et al., 2003; Wolkowinsky and Granger,
 312 2004; Häuselmann et al. 2007b).

313 Three main types of secondary particles are involved in the in situ production of
 314 cosmogenic nuclides: fast neutrons (L_n), stopping muons (slow or negative; $L_{\mu\text{slow}}$) and fast
 315 muons ($L_{\mu\text{fast}}$). Assuming that the production and erosion rates remained constant through
 316 time, the in situ-production of ^{10}Be is given by the following equation:

317 eq(1):

$$N_{(x,\varepsilon,t)} = \frac{P_{sp} \cdot \exp\left(-\frac{x}{L_n}\right) \left(1 - \exp\left(-t\left(\frac{\varepsilon}{L_n} + \lambda\right)\right)\right)}{\frac{\varepsilon}{L_n} + \lambda} + \frac{P_{\mu\text{slow}} \cdot \exp\left(-\frac{x}{L_{\mu\text{slow}}}\right) \left(1 - \exp\left(-t\left(\frac{\varepsilon}{L_{\mu\text{slow}}} + \lambda\right)\right)\right)}{\frac{\varepsilon}{L_{\mu\text{slow}}} + \lambda} + \frac{P_{\mu\text{fast}} \cdot \exp\left(-\frac{x}{L_{\mu\text{fast}}}\right) \left(1 - \exp\left(-t\left(\frac{\varepsilon}{L_{\mu\text{fast}}} + \lambda\right)\right)\right)}{\frac{\varepsilon}{L_{\mu\text{fast}}} + \lambda} + N_0 \cdot \exp(-\lambda \cdot t)$$

318

319 where $N(x,\varepsilon,t)$ is the nuclide concentration function of depth x (g/cm^2), denudation rate ε
 320 ($\text{g}/\text{cm}^2/\text{y}$) and exposure time t (y). Depths were defined at the centre of the sample. P_{sp} , $P_{\mu\text{slow}}$,
 321 $P_{\mu\text{fast}}$ and L_n , $L_{\mu\text{slow}}$, $L_{\mu\text{fast}}$ are the production rates and attenuation lengths of neutrons, slow
 322 muons and fast muons, respectively. L_n , $L_{\mu\text{slow}}$, $L_{\mu\text{fast}}$ values used in this paper are 160, 1500
 323 and $4320 \text{ g}/\text{cm}^2$, respectively (Braucher et al., 2003). λ is the radioactive decay constant and
 324 N_0 is the inherited nuclide concentration. $P_{\mu\text{slow}}$, $P_{\mu\text{fast}}$ are based on Braucher et al. (2011).
 325 Being aware of the importance of material density on the depth profile modelling results
 326 (Braucher et al, 2009, Rodés et al., 2011), bulk densities were determined in the laboratory by
 327 weighing mass of 100 cm^3 terrace material in wet and dry state. Resulting values for sand,
 328 gravelly-sand and sandy-gravel varied between 1.7 and $2.1 \text{ g}/\text{cm}^3$. Measured density of the
 329 fine aeolian cover (Ács) was $1.6 \text{ g}/\text{cm}^3$. Evidently, it was not possible to fit gravelly material
 330 in the 100 cm^3 cylinder perfectly and cobbles could not be represented. Hence, density was set

331 at slightly higher values as a free parameter, between 1.8 and 2.2 g/cm³, and considered as
332 constant over depth (Guralnik et al., 2011; Rixhon et al., 2011; Stange et al., 2013).

333 CosmoCalc add-in for Excel (Vermeesch, 2007) has been used to calculate sample
334 thickness scaling (with an attenuation coefficient of 160 g/cm²) and atmospheric pressures.
335 Production rates were scaled following Stone (2000) with a sea level high latitude production
336 rate of 4.02±0.36 atoms/g SiO₂/yr. This production rate is the weighted mean of recently
337 calibrated production rates in the Northern Hemisphere (Balco et al., 2009; Fenton et al.,
338 2011; Goehring et al., 2012; Briner et al., 2012).

339 For all samples the topographic shielding was negligible and therefore not corrected for.
340 Based on Dunai (2001), corrections for geomagnetic variations accounted for less than 5% for
341 the considered time span. Ages and denudation were not corrected for this variation.

342 During glacials, periglacial semi-arid loess steppe climate prevailed in the study area (Van
343 Vliet-Lanoë et al., 2004; Ruszkiczay-Rüdiger et al., 2015), with significant wind erosion
344 (Sebe et al., 2011), thus no correction of production rates for snow cover was necessary.

345 According to Cerling and Craig (1994) the effect of an old-growth fir forest on the
346 production rate of cosmogenic ³He is less than 4%. Plug et al. (2007) also concluded that the
347 shielding effect on cosmic irradiation of an old-growth boreal forest is less than 3%.
348 Differences in tree species and moisture content, and the lack of forest vegetation may result
349 in an even smaller correction of the site specific production rate, therefore ¹⁰Be production
350 rates were not corrected for the vegetation cover effect.

351 Concentrations of ¹⁰Be can be interpreted either as minimum exposure ages assuming
352 negligible denudation, or as maximum denudation rates. In these simplistic assumptions,
353 inherited TCN concentrations from pre-depositional exposure are neglected. When using
354 depth profiles denudation rate, exposure time and inheritance can be quantified if the TCN
355 concentration decreased exponentially with depth (Siame et al., 2004; Braucher et al., 2009)

356

357 3.2.2. Laboratory analysis

358

359 Chemical treatments of most samples were carried out at the CEREGE laboratory in Aix
360 en Provence (France) except for the alluvial samples from Ács (Dan13-11 to -18). For these
361 samples crushing and quartz purification were performed in the new TCN Sample Preparation
362 Laboratory of the Institute for Geological and Geochemical Research of the Hungarian
363 Academy of Sciences at Budapest.

364 Amalgamated pebble samples were crushed and sieved. 90-120 g of the 0.25-1 mm grain
365 size fraction was chemically etched. Sand samples were only sieved (0.25-1 mm) and for fine
366 sand samples (Dan 13-07 to-10) the 125-250 μm fraction was used. ~40 g pure quartz was
367 dissolved in HF in the presence of ^9Be carrier (100 μg of 3.025×10^{-3} g/g ^9Be in-house
368 solution). After substitution of HF by nitric- then hydrochloric acids, ion exchange columns
369 (Dowex 1x8 and 50Wx8) were used to extract ^{10}Be (Merchel and Herpers, 1999). Targets of
370 purified BeO were prepared for AMS (Accelerator Mass Spectrometry) measurement of the
371 $^{10}\text{Be}/^9\text{Be}$ ratios on ASTER, the French national facility, CEREGE, Aix en Provence (Arnold
372 et al., 2010). These measurements were calibrated against the NIST SRM4325 standard, using
373 an assigned value of $(2.79 \pm 0.3) \times 10^{-11}$ for the $^{10}\text{Be}/^9\text{Be}$ ratio. Analytical uncertainties (reported
374 as 1σ) include uncertainties on AMS counting statistics, uncertainty on the NIST standard
375 $^{10}\text{Be}/^9\text{Be}$ ratio, an external AMS error of 0.5% (Arnold et al., 2010) and chemical blank
376 measurement. A ^{10}Be half-life of $(1.387 \pm 0.01) \times 10^6$ years (Korschinek et al., 2010; Chmeleff
377 et al., 2010) was used.

378

379 3.3. Luminescence dating

380

381 3.3.1. Theoretical considerations

382

383 Luminescence is a radiometric dating method which provides the burial time of the
384 sediment as its most frequent components – quartz and feldspars – are acting as natural
385 radiation dosimeters (Lian and Roberts, 2006; Wintle, 2008). The previous luminescence
386 signal of the minerals can be completely or partially bleached by sunlight during
387 transportation. The rate of bleaching depends on the type of transportation and transport
388 distance. During aeolian transportation it is very likely that the minerals are completely
389 bleached by sunlight, however fluvial, glacial and marine transport may result in partial
390 bleaching.

391 The luminescence age of the sample is calculated from the equivalent dose (D_e : the natural
392 absorbed radiation dose [Gy] of the sample) divided by the annual natural ionising radiation
393 dose (dose rate [Gy/ka]) of the sample. Quartz is stimulated by blue light during the
394 measurements and generally known as Optically Stimulated Luminescence (OSL). Feldspars
395 are usually stimulated by infrared radiation, which is termed Infrared Stimulated
396 Luminescence (IRSL) to measure the equivalent dose of the samples (Lian and Roberts 2006;
397 Wintle 2008).

398

399 3.3.2. Experimental setup

400

401 The OSL of quartz has a lower dating limit, therefore it can be applied only on younger
402 samples (< 60-100 ka in Hungary). Therefore, in this study post-Infrared Infrared Stimulated
403 Luminescence (post-IR IRSL) measurements were carried out on K-feldspar samples,
404 comparing the post-IR IRSL 290 and post-IR IRSL 225 signals on the samples. The post-IR
405 IRSL signal of feldspar (Thomsen et al., 2008, Buylaert et al., 2009, 2012, Thiel et al., 2011)
406 usually has a dating limit up to 300 ka (Murray et al., 2014) and it shows negligible fading.
407 The IRSL signal of feldspars suffers from anomalous fading – an unwanted, athermal loss of
408 signal, resulting in age underestimation (Wintle, 1973). However, the post-IR IRSL signals
409 are less affected by fading (Thomsen et al., 2008; Buylaert et al., 2009; Thiel et al., 2011),
410 especially the post-IR IRSL-290 signal (Thiel et al., 2011), which does not need fading
411 correction, although a small fading rate (<1-1.5 %/decade) has been measured (Thiel et al.,
412 2011, 2014; Buylaert et al., 2012; Schatz et al., 2012). The applied post-IR IRSL-290 and -
413 225 (pIRIR-290, pIRIR-225) measurement protocols (Table S1) are described in Thiel et al.
414 (2011) and in Buylaert et al. (2009), respectively.

415 Luminescence samples were taken by pushing metal tubes into the previously cleaned
416 wall. The preparation of luminescence samples was conducted under subdued red light. The
417 coarse-grained fraction of 100-150 μm , or 150-200 μm , or 200-250 μm , or 250-300 μm was
418 extracted from the samples. All samples were treated using 0.1 N hydrochloric acid, 0.01 N
419 sodium-oxalate and 30 % hydrogen peroxide to remove carbonate, clay coatings and organic
420 matter from the samples, respectively. Feldspar and quartz grains were extracted by heavy
421 liquid separation using sodium polytungstate.

422 Luminescence measurements were performed using an automated Risø TL/OSL-DA-20
423 reader at the Department of Physical Geography in the Eötvös Loránd University, Institute of
424 Geography and Geology. The reader is equipped with a bialkali EMI 9235QB photomultiplier
425 tube, IR diodes ($\lambda=875$ nm) and a $^{90}\text{Sr}/^{90}\text{Y}$ β -source. Schott BG-39 and BG-3 filters were
426 placed in front of the photomultiplier, transmitting wavelengths between 350 and 420 nm.

427 Dose rates were obtained from the potassium, uranium and thorium content (Table S2), as
428 measured by gamma spectrometry fitted with a HPGe (High-Purity Germanium) N-type
429 coaxial detector in the laboratory at the Leibniz Institute for Applied Geophysics, in
430 Hannover. Polypropylene Marinelli beakers were filled with 700 g of the sediment and stored
431 for a minimum period of 4 weeks to allow the ^{226}Rn – ^{222}Ra equilibrium to be re-established.

432 A potassium content of $12.5\pm 1\%$ (Huntley and Baril, 1997) was applied to the K-rich feldspar
433 fraction to account for the internal dose rate. An average a -value of 0.08 ± 0.02 (Rees-Jones,
434 1995) was used for the feldspar IRSL age calculation. The cosmic radiation was corrected for
435 altitude and sediment thickness (Prescott and Hutton, 1994), assuming a water content of
436 $10\pm 5\%$ for all samples. Dose rate conversion is based upon the factors of Adamiec and Aitken
437 (1998).

438

439 *3.4. Multielectrode resistivity profiles*

440

441 Ten 2D multielectrode resistivity profiles (MUEL) were measured on the tIV terrace
442 surface aiming at an insight on the thickness of coarse-grained alluvial material deposited by
443 the paleo-Danube (Geomega, 2004). MUEL profiles enable the distinction among the
444 sediments of different textures down to depths of 40-60 m. The profiles were inverted using
445 RES2DINV software. Topographical data were incorporated to the measured data before
446 inversion and elevation changes were accounted for during the inversion process. The
447 inverted profiles were plotted using a horizontal scale of 1:2000, and vertical scale of 1:1000,
448 thus a vertical exaggeration of 2 was applied. The colour scale has been specifically chosen to
449 help discriminating the coarse grained terrace sediments and the underlying finer grained late
450 Miocene strata. Red and purple colours (ie. resistivities above 160 Ohmm) indicate coarse
451 grain sediments. When these sediments are located near the surface of the tIV terrace, they
452 most probably correspond to the deposits of the ancient Danube.

453

454

455 **4. Results**

456

457 *4.1. The cosmogenic ^{10}Be depth profile approach*

458

459 The time needed to reach the steady state concentration considering muon particles at
460 depth is much longer than at surface, where neutrons are the predominant particles in the total
461 production of cosmogenic nuclides (Braucher et al, 2003). Thus, as a first step, to determine
462 whether a depth profile has reached the secular equilibrium or not, one may assume an infinite
463 exposure age for all samples along the depth profile, then calculate the correlated maximum
464 denudation rates using eq(1) (Table 2). If the denudation rates are increasing from the top to
465 the bottom, then the steady state has not been reached, and an exposure age can be

466 determined. If the denudation rates remain constant implies that steady state has been reached,
467 which allows the determination of a minimum age of terrace abandonment only. When
468 denudation rates are constant until a certain depth and increase below that depth, then the
469 upper part of the profile will fix the denudation rate and the exposure time can be assessed
470 using the bottom part of the profile.

471 Constant or slightly decreasing maximum denudation rates for the Danube profiles at the
472 upper 1.5-2 m (save the Ács profile) suggest steady state conditions at the top (Table 2). The
473 ^{10}Be concentrations, however, become lower for the deepest samples, which may indicate a
474 difference of inherited ^{10}Be inventories. However, this shift of the ^{10}Be concentrations is not
475 reflected by a visible change of the sedimentary succession, as it could be expected in case of
476 a major change in source area, or at a larger time gap within the terrace deposit. At Grébics,
477 maximum denudation rates are constant with depth within uncertainties suggesting steady
478 state for all samples of the profile.

479 The calculation of the integration time (Lal, 1991), a most conservative minimum age
480 estimate of terraces is described in Supplementary Section 4.1.1.

481

482 Aiming at an age estimate of the studied terraces, χ^2 minimizations were performed using a
483 modified version of the *^{10}Be profile simulator 1.2* of Hidy et al. (2010) for profiles where the
484 ^{10}Be concentrations showed the expected exponential decrease with depth. After preliminary
485 quick tests, user defined χ^2 cut-off values were defined for each dataset to provide 100 000
486 model solutions for each simulation, and fulfill the Bayesian statistics. The applied Monte
487 Carlo simulation parameters are presented in Table 3 and are described in the Supplementary
488 Section 4.1.2.

489 Results presented in this paper are based on Bayesian probability density functions (PDF)
490 of the *depth profile simulator*. The Bayesian 1σ and 2σ upper and lower values (Table 4)
491 represent the uncertainty of the simulation results, thus were used to calculate the relevant
492 error bars. Bayesian 2σ lower ages are taken as minimum ages when ^{10}Be concentrations are
493 in secular equilibrium. This approach leaves only 2.5% probability of a lower age estimate.
494 This was the case for Grébics, Bana and Győr locations. For the Mocsá site, where ^{10}Be
495 concentrations have not reached a secular equilibrium yet, the reported uncertainties are
496 Bayesian 1σ lower and upper values. For denudation rate and inheritance the Bayesian most
497 probable values with a 1σ confidence level are discussed.

498 At the Ács site the alluvial sediments had a cover of aeolian origin with no exponential
499 decrease of the TCN concentrations. Here a different approach has been applied, which is
500 described later.

501 Complete data tables of the simulator results and statistics are found in table S4.

502

503 *4.2. Exposure age – denudation rate determination of the terrace IV*

504

505 Three sample sites, Győr, Bana and Grébics belong to this horizon. According to field
506 evidences, considerable thickness of fluvial material has been eroded from the tIV terrace.
507 Fine overbank deposits, which are usually deposited on top of glacial terraces (Bridgland,
508 2000) are usually missing from the terrace surface. Borehole data from both the actual
509 floodplain and from terraces in the neighbouring Gerecse Hills, where freshwater limestone
510 locally protected the complete alluvial sequence, suggest that several meters of overbank
511 sediments on top of the coarse material was typical along the Danube. Based on these
512 evidences, a minimum total denudation of 5 m for the lower, more extended terrace remnants
513 (Győr, Bana, Figs. 2A, 3) and 6 m for the highest, most eroded location (Grébics, Figs. 2A, 3)
514 was constrained during the simulations to get the most probable minimum ages. The upper
515 threshold of total denudation was set as high as 15 m and 16 m, respectively. The upper and
516 lower age limits for the simulations were determined using stratigraphical and
517 paleontological evidences (described in Section 2.2) (Fig 4, Table 3).

518

519 The **Győr profile** (Dan08-02 to -12, Table 2) is the only one where sand and amalgamated
520 pebble sub-samples could be separated by sieving and were processed separately, as there was
521 enough material from both grain sizes to have a sample set of at least five samples, the
522 minimum sample number for depth profile modelling (Hidy et al., 2010).

523 The ^{10}Be concentrations resulted to be consistently slightly higher in sand than in pebble
524 sub-samples, with both showing an exponential decrease towards depth (Table 2, Fig. 7A).
525 The variable inherited TCN concentrations of individual pebble samples was described by
526 several authors (e.g. Schmidt et al., 2011; Codilean et al., 2014), however this effect can be
527 overcome by sampling along depth profiles and using amalgamated pebble and/or sand
528 samples (Anderson et al., 1996); the approach followed by this study. Aguilar et al. (2014)
529 suggested that different denudation processes on the catchment and the effect of size
530 reduction of larger boulders may be responsible for the smaller ^{10}Be concentrations in alluvial
531 gravel than sand in a semi-arid catchment of the central Andes. On the other hand, several

532 authors described a positive correlation between grain size and ^{10}Be concentrations (Schmidt
533 et al, 2011), while others reported variable or similar TCN concentrations in different grain
534 sizes (Schaller et al., 2001; for a revision refer to Carretier et al., 2015). The discussion of
535 catchment-scale denudation processes of the Danube River is out of the scope of this study
536 due to the size and complexity of the drainage area.

537 The observed exponential decrease of ^{10}Be concentrations with depth for both pebble and
538 sand suggest that the inherited ^{10}Be inventory is constant along the profile, although at
539 different concentrations for pebble and sand samples. Therefore, the post-depositional
540 exposure history can be modelled using TCN concentrations of each grain size fraction (Hidy
541 et al., 2010). Both models are expected to provide similar exposure age – denudation rate
542 pairs, but with a different amount inherited ^{10}Be concentration.

543 Profile simulations (*depth profile simulator* of Hidy et al., 2010) were performed
544 separately for the sand and for the pebble profiles. Simulation settings were similar (Table 3)
545 as they share the same post-depositional exposure history. A wide user defined range of
546 inheritance was chosen to cover the expected values for both the sand and the pebble profiles.
547 The user defined χ^2 cut-off value was much lower for the pebble profile than for the sand (2
548 and 15, respectively), suggesting a better model fit of the pebble profile (n=5) than for the
549 sand profile (n=8).

550 Considering all sand samples the simulator was not able to find a solution because of the
551 distinctly low ^{10}Be concentration of the bottom samples (Dan08-11 and-12; Table 2, Fig. 7A).
552 These samples represent a cross-bedded sand layer underlying the coarser, graded channel
553 deposits of the upper 4.3 meters of the profile and showed sharply different maximum
554 denudation rate values (Figs. 5A, 6A, Table 2). The measured ^{10}Be concentrations suggest
555 that this sand unit originates from a previous sedimentation phase. Therefore, samples #11
556 and #12 were excluded from further simulations.

557 The simulation results approved that the ^{10}Be concentrations have reached secular
558 equilibrium (Figs. S1A, S2A, Table S4). Bayesian 2σ lower ^{10}Be exposure age of the Győr
559 profile is >693 ka for the pebble and >708 ka for the sand profile, suggesting a minimum
560 exposure age of the terrace around >700 ka (Table 4) for this site.

561 Probability density functions (PDF) are similar for pebble and sand (Fig. S1A). They show
562 well determined peaks for denudation rate ($6.5_{-0.7}^{+1.6}$ m/Ma and $7.0_{-1.2}^{+0.5}$ m/Ma for pebble and
563 sand) with values in agreement within error with each other and with the previously
564 calculated maximum denudation rate (Table 4). The inherited ^{10}Be component resulted to be

565 lower for the pebbles than for the sand (70_{-5}^{+14} kat/g and 95_{-12}^{+27} kat/g, respectively), as it was
566 expected (Table 4, Fig. 7A).

567

568 From the sample set at **Bana** (Dan08-20 to -26) the Dan 08-22 sample had considerably
569 higher ^{10}Be concentration than suggested by its subsurface depth (Table 2, Fig 7B), therefore
570 it was excluded from depth profile modelling as an outlier. The previously suggested steady
571 state ^{10}Be concentrations were confirmed by the depth profile modelling (Figs. S1B, S2B,
572 Table S4). Monte Carlo simulations yielded a Bayesian 2σ lower exposure age of >477 ka
573 with a well constrained most probable denudation rate of $10.0_{-0.1}^{+0.2}$ m/Ma and inheritance of
574 106_{-14}^{+35} kat/g (Table 4).

575

576 The ^{10}Be concentrations of the top samples of the **Grébics** profile (Dan08-40 and -41)
577 suggested that the upper 50 cm could be mixed during quarrying (Table 2, Figs. 5C, 6C).
578 Accordingly, the arithmetic mean of the ^{10}Be concentrations of the two upper samples was
579 placed in their average depth, in order to provide a rough estimate of ^{10}Be concentration
580 relevant for the uppermost part of the profile (Fig. 7C). This *Dan08-40-41avg* sample was
581 considered during the depth profile simulations.

582 The simulation results suggested also secular equilibrium of ^{10}Be concentrations (Figs.
583 S1C, S2C, Table S4). The Bayesian 2σ lower exposure age of the Grébics profile resulted to
584 be >659 ka; with a well determined denudation rate of $8.7_{-0.8}^{+1.2}$ m/Ma and an inheritance of
585 85_{-16}^{+24} kat/g (Table 4).

586

587

588 *4.3. Exposure age – denudation rate determination of the terrace IIb*

589

590 At **Mocsa** (Dan08-31 to -35) the sandy gravel channel facies of the terrace was covered by
591 fine overbank sediments in 1.3 m thickness. Therefore, ^{10}Be samples were taken from the 1.3
592 to 4 meter depth range (Figs. 5D/1, 6D/1, 7).

593 For the depth profile modelling, user defined lower and upper age thresholds were
594 determined as 14 and 360 ka, respectively (Table 3), on the basis of geological, stratigraphical
595 and paleontological data integrated by the revision of the terrace chronology (Fig. 4). Total
596 amount of eroded material was maximised at 500 cm, due to the presence of the overbank fine
597 sediments and low position of the terrace.

598 Simulation results are presented by Figs S1D, S2D, Table S4. The Bayesian most probable
599 age of the terrace at Mocsa was 149_{-13}^{+160} ka, with an associated denudation rate of $5.8_{-2.6}^{-2.1}$
600 m/Ma and an inherited ^{10}Be concentration of 91_{-4}^{+4} kat/g (Table 4). As a result of the lack of
601 uppermost part of the depth profile, the exposure age could be constrained only with large
602 uncertainty towards older ages.

603

604 At **Ács** the measured ^{10}Be concentrations were significantly higher in the sandy gravel of
605 fluvial origin than in the aeolian cover sediments. The measured ^{10}Be concentrations did not
606 display the expected exponential decrease with depth (Table 2, Fig. 7E). The mean ^{10}Be
607 concentration of the cover (84 ± 8 kat/g) is similar (within error) to that of the actual Danube
608 gravel (Dan08-27; Table 2) and to the mean inheritance signal of the other profiles.

609 The lack of exponential decrease in the cover may suggest (1) vertical sediment mixing in
610 the upper 130 cm, (2) dominance of the inherited component due to the very young age of the
611 sediment or (3) recent truncation of the upper part of the profile (at least 3 m material stripped
612 off).

613 The possibility of mixing (1) could not be excluded due to the proximity of the present day
614 soil, and the presence of recent biogalleries (Fig. 5E). The recent deposition of the loess with
615 a constant inherited ^{10}Be concentration (2) is improbable as the youngest published OSL age
616 of fine-sandy loess was 14 ± 2 ka (Thamó-Bozsó et al., 2010) and in the Pannonian Basin all
617 data suggest that loess formation ceased during the Holocene. The likelihood of recent
618 truncation of several meters of material (3) from the top of the terrace surface seems to be
619 limited due to the low position and flat topography of the terrace and also due to the
620 conservative denudation rates calculated for all terraces, including those at higher positions
621 (Table 4). The surface of this terrace segment is hummocky, with variable depth of wind-
622 blown material. Therefore, wind erosion and formation of deflation hollows may have
623 occurred.

624 In the lower, fluvial part of the profile, ^{10}Be concentrations are roughly constant near 130
625 kat/g. A slight decreasing trend with depth can be observed only in the upper 60 cm (samples
626 Dan13-11 – 13; Table 2). Field observations (Supplementary Section 3.1.5 and Figs. 5E, 6E)
627 suggest that considerable cryoturbation affected the fluvial sediment at this site. The lack of
628 exponential decrease and the high ^{10}Be concentrations (significantly above the inheritance of
629 the other terraces and the actual Danube gravel; Tables 2, 6) suggest that periglacial mixing
630 might have affected the entire profile.

631 To simulate this profile, two models can be proposed:

632 1) In a simplistic two-step approach in step one, the lower part of the profile is deposited
633 with an important inheritance signature; after short exposition, this deposit was covered by
634 fine material. This scenario yields to an age of ~4ka for the fluvial terrace abandonment with
635 an inherited concentration of 124 kat/g and a present fine deposition event with inherited ^{10}Be
636 concentrations of ~83 kat/g. This scenario can be excluded due to the position of the terrace
637 12 m above the river, to the time needed for the development of the reddish paleosoil on top
638 of the terrace material and also due the cryoturbation and loess formation on top of the terrace
639 material, both requiring periglacial and/or arid climate conditions (in contrast with the present
640 day temperate climate of the area).

641 2) After sedimentation, the alluvial material was exposed and a reddish paleosoil
642 developed on its top. Soil formation was stopped by climate deterioration, and the sequence
643 was mixed under periglacial conditions. This event erased the former exponential decrease of
644 ^{10}Be concentrations with depth. Cryoturbation was followed by loess deposition.

645 Accepting this latter scenario, it was possible to assess the duration of buried exposure
646 time of the alluvium using the slight exponential decrease of ^{10}Be concentrations in the upper
647 part of the fluvial sediments. Simulation settings appear in Table 3, results are presented in
648 Table 4, Fig. S1E, S2E. The user defined denudation rate was between 0 and 10 m/Ma, based
649 on denudation rate data obtained from the other sample locations (Table 4). Our tentative age
650 estimate for the age of loess deposition (buried exposure of the terrace) is 12_{-6}^{+26} ka.

651

652

653 *4.4. Luminescence characteristics of the samples, post-IR IRSL ages*

654

655 *4.4.1 Dose-recovery, residuals, fading*

656

657 The D_e values for the pIRIR signals are obtained by integrating the first 2.5 s of the IRSL
658 decay curve and the final 100 s of the stimulation is subtracted in order to remove the
659 background. Dose response curves were fitted using single or double exponential function
660 (Fig. 8A,B). For all experiments only aliquots with a recycling ratio consistent with unity ± 10
661 % and with a recuperation $< 10\%$ were accepted.

662 Prior to the equivalent dose measurements dose-recovery tests were carried out on one
663 sample from each profile to test the reliability of the applied measurement protocol. The
664 aliquots were bleached in a Hönle solar simulator for three hours. Dose-recovery ratios range
665 between 0.94 ± 0.04 and 1.15 ± 0.05 . These values – except the sample from Mocsa, which has

666 the highest dose-recovery ratio (1.15 ± 0.05) – are surprisingly low. Other authors reported
667 higher dose-recovery values (often >1.1) (Stevens et al., 2011; Thiel et al., 2011, 2014; Schatz
668 et al., 2012). Murray et al (2014) compared the dose-recovery values after bleaching in solar
669 simulator or window (sunlight bleaching) and concluded, that the bleaching in solar simulator
670 results in better dose-recovery ratios. In our case the reason for the unexpectedly good dose-
671 recovery ratios is probably the same.

672 Residual signals were also measured after sunlight bleaching in window and a small
673 residual signal ($<2\text{Gy}$) was only observed for the samples. This residual dose was not
674 subtracted either from the recovered dose, or from the natural dose estimates because this can
675 be considered negligible.

676 Fading tests were carried out on six samples (4-8 aliquots), one from each profile at Bana
677 and Mocsa and on all samples from the profile at Ács. The fading rates (g-values) are shown
678 in Table S5. Large inter- and intra-sample variation was observed. The fading measurements
679 resulted in very large scatter in g-values and sometimes negative g-values, probably as
680 laboratory measurement artefacts. However, our observation is not unique, many authors have
681 reported about negative or even anomalously high g-values in some cases (Murray et al.,
682 2014; Trauerstein et al., 2014; Lowick et al., 2012). Our results range between -4.35 ± 2.54 and
683 4.56 ± 2.36 %/decade (Table S5), and some of them are much higher than those reported in
684 other studies ($<1-1.5\%$ /decade) (Thiel et al., 2011, 2014; Murray et al., 2014; Schatz et al.,
685 2012). This difference might be explained by mineral and grain-size differences (polymineral
686 fine grained versus coarse-grained K-feldspar samples). The detected large scatter of the
687 fading rates indicates that the measured values are probably unreliable for our samples.
688 (However, averaging our measured fading rates resulted in g-value of $1.5 \pm 1.2\%$ /decade
689 (Table S5), which can be acceptable for the pIRIR-290 signal.) Nevertheless, it has already
690 been demonstrated by previous studies (Thiel et al., 2011; Buylaert et al., 2012) that the
691 pIRIR-290 signal is stable, not influenced by anomalous fading (Wintle, 1973) (e.g.: natural
692 pIRIR-290 signal is consistent with saturation), therefore we did not correct the ages for
693 fading at all.

694

695 *4.4.2 Equivalent dose, incomplete bleaching, saturation, age estimates*

696

697 D_e values range from 46.2 ± 3.2 Gy to 386.8 ± 9.5 Gy (Table 5). The pIRIR-290 signal has
698 not reached the saturation level for five samples (Mocsa-1, -2, Ács-A1, -A2, -A4) (Figs. 5D,E,
699 8A,B). These D_e values correspond to age estimates from 15.2 ± 1.3 ka to 146 ± 10 ka.

700 Representative shine-down and dose-response curves from Mocsa-2 and Ács-A2 are shown in
701 Fig.8 A,B.

702 Considering the fluvial origin of the samples, we can expect partially bleached minerals,
703 which would result in age overestimation (Rodnight et al., 2006). Therefore, we applied small
704 aliquots (10-30 grains/aliquot) for the measurements, which are more or less able to mimic the
705 single grain measurements. We found that all of our samples (which were not in saturation)
706 resulted in D_e distributions almost within ± 2 sigma. Radial plots of the small aliquots (Fig. 9)
707 do not show wide D_e distributions therefore severe incomplete bleaching can be excluded for
708 the samples. Although, as Reimann et al. (2012) and Trauerstein et al. (2014) reported, signal
709 averaging can still be detected in case of small aliquots compared to single grain dating, but
710 this might be more emphatic for young samples. In this study the fading corrected results of
711 the pIRIR-225 protocol (Table S6) show good agreement with the pIRIR-290 ages (Table 5),
712 which indicates that our samples are not influenced by severe incomplete bleaching as the
713 pIRIR-225 signal is easier bleachable than the pIRIR-290 signal. The results of the pIRIR-225
714 measurement protocol are described in the Supplementary Section 4.4.3. Nevertheless, in case
715 we failed to detect the effect of incomplete bleaching, the residual dose measured on a
716 modern fluvial sample (Trauerstein et al., 2014) would correspond to a potential pIRIR-290
717 age overestimation of about 13.5 ka, which is about 10% of the age of our fluvial samples.
718 The residual signal would be significant for younger samples when the magnitude of the
719 residual dose is similar to the natural dose. Therefore the mean age was calculated to estimate
720 the pIRIR ages.

721 The pIRIR-290 signal was observed in saturation for four samples out of nine (Bana-A1, -
722 A2, -B3, Ács-B3) (Fig.8C). The ratio of the sensitivity-corrected natural signal to the
723 laboratory saturation level was calculated for these samples and they ranged from 0.87 to 0.95
724 (Table 5). Therefore only minimum equivalent doses and age estimates were calculated for
725 these samples (Wintle and Murray, 2006; Murray et al., 2014) (a single exponential growth
726 curve was fitted to the data points and the $2 \cdot D_0$ value was calculated to assess minimum age
727 estimates). These minimum dose estimates scatter around 1000 Gy, range from 904 Gy to
728 1120 Gy similarly as it is reported by Murray et al. (2014). They suggest that finite dose
729 estimates cannot be used beyond ~ 1000 Gy, which results in age estimate around 300 ka
730 considering a dose-rate of 3-4 Gy/ka for fine-grained sediment. In this study the total dose
731 rates are lower, ranging from 2.26 ± 0.21 Gy/ka to 3.51 ± 0.17 Gy/ka (Table 5) therefore age
732 estimates up to 400 ka are possible.

733 All three samples from the Bana profile (tIV, Figs. 5B, 6B) were beyond the datable time
734 range because the pIRIR-290 signals were already in saturation (Fig. 8C). Therefore only
735 minimum ages were estimated from this profile, >410 ka for sample Bana-A1, >386 ka for
736 sample Bana-A2 and >320 ka for sample Bana- B3 (from the underlying late Miocene sand)
737 (Table 5). As the uppermost sample yielded the oldest minimum age, the entire profile can be
738 considered as >410 ka, which is in accordance the ^{10}Be exposure age estimate of this terrace
739 (Table 6).

740 Two sand samples from the tIIb terrace at Mocsá (Figs. 5D/2, 6D/2) provided age
741 estimates of 139 ± 9 and 146 ± 10 ka for the sediment burial (Table 5), which are in agreement
742 with our ^{10}Be exposure age estimate (Table 6).

743 Four samples were collected from the terrace profile at Ács (Figs. 5E, 6E). The sample
744 Ács-B3 was taken from the material underlying the terrace and it approaches saturation,
745 therefore a minimum age of >258 ka was determined for this sample (Table 5). The fluvial
746 terrace material, samples Ács-A2 and Ács-A4 provided age estimates of 95 ± 6 ka and 86 ± 6 ka,
747 respectively. Accumulation of the aeolian cover of the terrace was dated to 15 ± 1 ka, which is
748 similar to the buried exposure time estimated for the fluvial sediments (Table 4).

749
750

751 **5. Discussion**

752

753 Both post-IR IRSL and TCN dating methods were successfully applied to determine the
754 age of deposition of Danube terraces. For the older terraces only minimum ages could be
755 assessed by both methods. For the higher terrace remnants (tIV) the depth profile modelling
756 of ^{10}Be data using the *^{10}Be depth profile simulator* (Hidy et al., 2010) allowed determining
757 older minimum terrace ages compared to luminescence dating at one location (Bana; Table 6).
758 Consequently, no further post-IR IRSL samples were processed from this horizon. On the
759 other hand, for the lower/younger horizons the post-IR IRSL method provided better
760 constrained terrace ages. The combination of the two methods enabled a more robust age
761 determination than using either method separately (Guralnik et al., 2011).

762

763 *5.1. Early Pleistocene age of the terrace IV of the study area*

764

765 Accurate dating of the tIV terraces was difficult due to secular equilibrium of ^{10}Be
766 concentrations and saturation of the luminescence signal. Nevertheless, ^{10}Be minimum ages

767 from >477 to >708 ka and a post-IR IRSL minimum age of >410 ka could be determined for
768 this horizon (Table 6). ¹⁰Be depth profile modelling enabled the determination of the oldest
769 minimum terrace age at Győr due to the lowest denudation rate at this location (Tables 4, 6).
770 Accordingly, we consider the >700 ka ¹⁰Be exposure age as the best estimate for the entire
771 tIV terrace range, which is concordant with the other minimum ages younger than >700 ka.
772 This age is considerably older than the ~420 ka age of terrace abandonment suggested by the
773 revised traditional terrace chronology (Fig. 4) and is well in agreement with the 540-2600 ka
774 age range provided by malacological investigations of Krolopp (1995). The >700 ka minimum
775 age of the highest terrace flight of the GTT is comparable with the estimated age of the tV or
776 tVI horizon of the TR (Figs. 3, 4). As the tVI horizon is considered to be the first Danubian
777 terrace in the TR marking the beginning of river incision in the Hungarian Danube valley, we
778 propose that the highest terrace of our study area (tIV) may share the same age as tVI in the
779 TR. Consequently, in the following part of this study we use the name tIV-VI for this terrace.

780 On the other hand, our results do not support the proposed link between terrace formation
781 and MIS terminations (Gábris 2008, Gábris et al., 2012; Fig. 4) during the middle Pleistocene.
782 Interestingly, we did not find any evidence on the existence of terrace horizons between
783 143±10 ka and >700 ka in the study area. Middle and late Pleistocene times are usually
784 represented by at least 4-5 terraces in Europe (Gibbard and Lewin, 2009 and references
785 therein) and in the TR (Pécsi, 1959; Ruzsáczay-Rüdiger et al, 2005a, Gábris and Nádor,
786 2007; Gábris et al, 2012), with well-expressed horizons usually linked to the Mindel glacial
787 (MIS 12). We propose that these horizons may have also developed in the study area, but
788 were most probably destroyed by the subsequent lateral erosion of the Danube. The uplifting
789 barrier of the TR composed of more resistant lithology could lead to an increased sinuosity
790 (and thus severe lateral erosion) of the Danube in the study area (Schumm et al., 2002),
791 situated just upstream of the TR (Figs. 1, 3)

792

793

794 *5.2. Separation of the late Riss/Saale (tIIIa) and Early Würm/Weichsel (tIIb) terraces*

795

796 It was in question whether all the terrace remnants considered formerly as tIIb (with
797 elevations 12-16 m and 17-22 m above the Danube) belonged to the same level sharing the
798 same age, as suggested by Pécsi (1959). Mapping of the base of the Quaternary sediments
799 (using shallow borehole data; Kaiser, 2005) revealed a 6-8 m high riser between the base of

800 the lower and higher parts of this terrace, suggesting the existence of two separate terrace
801 horizons, whose surface scarp was usually masked by erosional processes.

802 Post-IR IRSL dating provided well constrained burial ages of 139 ± 9 and 146 ± 10 ka
803 (averaging at 143 ± 10) for the Mocsa site (20m amrl; Tables 5, 6). Cosmogenic in situ ^{10}Be
804 dating provided 149^{+160}_{-13} ka exposure age, coinciding within error with IRSL results both
805 suggesting terrace deposition during MIS 6 (130-190 ka; Lisiecki and Raymo, 2005). The
806 luminescence age of 95 ± 6 ka and 86 ± 6 ka (averaging at 91 ± 6) of the terrace at Ács (12m
807 amrl) proposes an early Würmian (MIS 5b-c) age of this terrace.

808 Considering the elevation and the age differences of the Mocsa and Ács locations (Table
809 6), we propose the separation of these terraces. We suggest that a higher level has developed
810 during the late Riss glacial (MIS 6; Mocsa), which we index as tIIIa (after Gábris and Nádor,
811 2007) and that the lower, tIIIb level has formed during the early Würm glaciation (MIS 4-5;
812 Ács) (Fig. 4).

813

814 *5.3. Climate control on the onset of terrace formation*

815

816 Our new chronological data suggests that MIS based terrace chronology proposed by
817 Gábris (2008) is reasonable for the low terraces up to tIIIa, but it is not applicable for the
818 higher terraces (Fig. 4). Apparently, during the middle Pleistocene not all the climate
819 fluctuations led to terrace formation or these terraces were destroyed later by the river.

820 The first major worldwide Pleistocene glacial events with substantial ice volumes in
821 continental areas outside the polar regions occurred around 0.8-0.9 Ma (MIS 22) (Ehlers and
822 Gibbard, 2007; Clark et al., 2006; Gibbard and Lewin, 2009). After the “mid-Pleistocene
823 climate transition” (Clark et al., 2006; ca. 1.2-0.7 Ma) dominantly cold climate conditions
824 with short episodes of glaciations and brief interglacials prevailed. The >700 ka minimum age
825 of the uppermost, tIV-VI horizon in our study area is reconciliable with a river style change
826 triggered by the “mid-Pleistocene climate-transition” when mechanism of larger rivers in
827 Western Europe shifted from valley widening to incision (Gibbard and Lewin, 2009). Burial
828 age determination of cave sediments in the northern Swiss Alps revealed an abrupt increase of
829 valley incision rates (from 0.12 mm/a to 1.2 mm/a) at 0.8-1.0 Ma (Häuselmann et al., 2007a),
830 in good agreement with the time of the mid-Pleistocene climate transition.

831 Hence, we suggest that abandonment of the former wide valley and onset of incision may
832 have started in connection with this climate event as a complex response of the river to the
833 propagation of the compression related vertical movements from the SW towards the basin

834 interior and of climate change. Accordingly, the minimum age of >700 ka of the highest
835 terraces may be valid for the Hungarian Danube valley as the onset valley incision, instead of
836 the Plio-Pleistocene boundary, suggested by Pécsi (1959).

837 Resolution of our data does not enable the distinction between climatically or tectonically
838 triggered terrace formation. Fold related uplift of the TR and subsidence of the neighbouring
839 lowlands (Figs. 1, 3) necessarily led to river incision in and around the uplifting TR, where
840 terraces were formed as a consequence of the interplay (positive and negative feedbacks)
841 between the ongoing uplift and climate change.

842

843 5.4. Denudation of the terrace surfaces

844

845 Multielectrode resistivity profiles (MUEL; Fig. 10) and field experiences demonstrate that
846 the tIV-VI horizon is represented by a sheet of coarse-grained material in up to 8-10 m
847 thickness, which is underlain by fine sand (of late Miocene age; Magyar et al., 2007, 2013). It
848 is also well visible, that the base of the terrace remnant at Bana is around 150 m asl. (Fig.
849 10A,B) and goes as high as ~185 m in the eastern termination of the GTT (Fig. 10D,C).
850 Decreasing thickness of the gravel sheet towards the margins of the terrace remnant hills is
851 indicative of more intensive denudation towards the terrace scarps. It is also visible that
852 coarse terrace material is thinner at the eastern, higher section of the GTT, where their
853 maximum apparent thickness is 3-5 m (Geomega, 2004).

854 Cosmogenic ^{10}Be depth profiles provided well constrained denudation rates of the terrace
855 surfaces (Table 6). The highest denudation rates were calculated at Bana $10.0^{+1.1}_{-0.6}$ m/Ma) and
856 at Grébics ($8.7^{+1.2}_{-0.8}$ m/Ma), at the edge of the terrace remnant and on the small, isolated terrace
857 hill at the most elevated part of the terrace region, respectively. Lower denudation rate was
858 detected on larger flat surfaces: $6.5^{+1.6}_{-0.7}$ m/Ma and $7.0^{+0.5}_{-1.2}$ at Győr and $5.8^{+2.1}_{-2.6}$ m/Ma at
859 Mocsá. Considering the minimum exposure ages of the terraces these rates suggest 5-8 m of
860 material eroded from the tIV-VI terraces and ~1.0 m denudation from the tIIb-tIIIa at Mocsá.

861 The raw calculations of maximum denudation rate (Table 2) provided slightly lower values
862 than denudation rates provided by the *depth profile simulator* (Hidy et al., 2010). This is
863 because during depth profile simulations the inherited amount of ^{10}Be is considered and
864 density is a free parameter between 1.8 and 2.2 g/cm³. On the other hand, when maximum
865 denudation rate is calculated for each sample (Table 2), it is not possible to account for the
866 inheritance and density is a fixed value (at 2 g/cm³).

867 Portenga and Bierman (2011) reported that basin-averaged ^{10}Be denudation rates are
868 usually faster than outcrop erosion. In their study mean and median basin-averaged
869 denudation rates in the temperate climate zone were 277 m/Ma and 84 m/Ma, while for
870 outcrops these were 25 m/Ma and 16 m/Ma, respectively. The difference between the 30-80
871 m/Ma typical basin-averaged ^{10}Be denudation rates of middle European rivers (Schaller et al.,
872 2002) and low (up to 10 m/Ma) rates of terrace denudation in the GTT of this study suggest
873 increasing relief of the study area, as it is expected by the river incision and terrace formation
874 described in the area. Coarse grained fluvial material have protected the high terraces from
875 surface denudation, which led to surface inversion: the former riverbed has now the highest
876 topography of the area. Due to their slow denudation, these terraces serve as appropriate tools
877 to reconstruct former stages of valley evolution and thus to quantify valley incision and
878 tectonic uplift (see in section 5.6).

879

880 *5.5. Cryoturbation features*

881

882 The periglacial involutions in the outcrop at Ács were observed down to a depth of 1-1.5 m
883 from the top of the alluvial material, with no signs of mixing in the aeolian cover (Figs. 5E,
884 6E). Periglacial sedimentary features were frequently observed in the western Pannonian
885 Basin (Pécsi, 1961, 1997; Van Vliet-Lanoë et al., 2004). Their development was attributed to
886 glacial periods of the Würm and/or Riss glaciations and possibly in earlier glacial phases, at
887 locations where local climate and soil conditions were suitable for cryoturbation (e.g.
888 Matsuoka, 2011; Ruszkiczay-Rüdiger and Kern, 2015).

889 Our sampling occurred outside these well visible signs of sediment mixing (Figs. 5E, 6E).
890 However, the lack of exponential decrease of the measured ^{10}Be concentrations suggests that
891 cryoturbation affected the former terrace, even outside the involutions and at least down to a
892 depth of 3 m (Table 2, Fig. 7).

893 The post-IR IRSL age of the terrace sedimentation was 91 ± 6 ka and the age of loessy-fine-
894 sand cover proved to be 15 ± 1 ka (Table 6). The latter age is similar to the youngest age of
895 loess formation in the region (Thamó-Bozsó et al, 2010) during the latest Pleistocene (end of
896 MIS 2). The ^{10}Be exposure age estimate (covered exposure time of the alluvial sediments) of
897 12_{-6}^{+26} ka of the cover, is in agreement with its post-IR IRSL age.

898 These ages are bracketing the development of a 30-40 cm thick reddish-brown paleosoil
899 developed on top of the alluvial sediments, which was subsequently deformed by the
900 cryoturbation before the deposition of the aeolian cover. The post-IR IRSL dated loess-

901 paleosoil sequence of the Süttő quarry in neighbouring Gerecse Hills showed that several
902 paleosoil horizons were formed between ~85 and ~30 ka, with the most prominent one during
903 the MIS 3 (Novothy et al, 2011). Accordingly, the formation of the paleosoil on the alluvial
904 sediments at Ács during this period is highly probable, which could be affected by
905 cryoturbation under the periglacial climate conditions of MIS 2 (Ruszkiczay-Rüdiger and
906 Kern, 2015).

907

908 *5.6. Incision rates and tectonic uplift*

909

910 Uplift of the TR was related to the neotectonic shortening and related large-scale folding of
911 the Pannonian lithosphere (Horváth and Royden, 1981, Horváth et al. 2015), which
912 progressively shifted from SW (Slovenia) toward NE (Tari, 1994; Fodor et al. 2005). The
913 >700 ka minimum age of the highest terraces of the area put forward that the onset of uplift in
914 the northeastern part of the TR started before this time, but probably somewhat later than in
915 the southwest TR (Fig.1), where the minimum age of the onset of the uplift was 1.56 ± 0.09
916 Ma, as determined by in situ ^{10}Be exposure age of wind-abraded landforms (Ruszkiczay-
917 Rüdiger et al., 2011).

918 The maximum incision/uplift rates relevant for the minimum age of the IV-VI terrace level
919 (>700 ka) were calculated. Elevation of the terraces was corrected for the material thickness
920 eroded from their surface since their abandonment by the river (calculated using ^{10}Be
921 denudation rates, Table 6, Section 5.4). The maximum incision rates show a positive
922 correlation with the elevation of the terrace remnants with values from <0.06 mm/a to <0.13
923 mm/a (Győr: <0.06 mm/a; Bana: <0.07 mm/a; Grébics: <0.13 mm/a) (Fig. 11, Table S7). The
924 increase of incision rates from west to east along the Danube reflects well the trend towards
925 higher uplift rates closer to the uplifting TR (Figs. 1, 3). These rates are half of the maximum
926 rates calculated based on the revised traditional chronology (Fig. 11) and are in the same
927 order of magnitude as incision/uplift rates revealed by former studies in Europe (Table 1).

928 For the young terraces, incision rates of 0.15 ± 0.01 mm/a (Mocsa) and 0.13 ± 0.01 mm/a
929 (Ács) were computed using the post-IR IRSL data. Our results suggest somewhat faster uplift
930 rates for the young terraces (Mocsa, Ács) than for the tIV-VI terraces at the same position
931 along the river (Fig. 11). This might indicate slight acceleration of uplift towards present,
932 which would be in accord with the gradual built-up of compressional stress-field in the
933 Pannonian Basin resulting in an increasing trend of vertical movements (Bada et al., 2006).

934 However further data are necessary to decide whether our data indicates a true acceleration of
935 vertical movements.

936 Proposing that the abandonment of the highest terrace level of the Hungarian Danube
937 valley was triggered by the same tectonic and climatic processes, we tentatively extrapolated
938 the >700 ka minimum age to the highest terraces of the valley across TR and calculated the
939 uplift rates for each valley segment (blue dotted line on Fig. 11, Table S7). The comparison of
940 the uplift rates along the Danube calculated using our minimum age estimate of >700 ka and
941 the uplift rates calculated on the basis of the quantification of the traditional terrace
942 chronological data (orange line of Fig. 11) it is visible, that while in the GTT the traditional
943 data suggest considerably higher values, within the TR the maximum uplift rates are similar,
944 with <0.36 mm/a and <0.33 mm/a at the axis of the TR (Danube Bend; Fig. 11). By the >700
945 ka based uplift curve (blue line and blue dotted line on Fig. 11) the anomalously high uplift
946 rates suggested by the traditional terrace chronology for the eastern part of the GTT were
947 eliminated, and the curve has a smooth continuation towards the TR.

948 A slight break can be suggested in the extrapolated uplift rate curve blue dotted line on Fig.
949 11), next to the locations of the travertine occurrences. As we noted earlier, there is a fault at
950 the western margin of the Gerecse Hills, which was clearly active in the Late Miocene (Bartha
951 et al. 2014). In case of Quaternary activity along this fault, the uplift rate may have changed at
952 this fault. However, the spatial resolution of our quantitative data set is not dense enough to
953 support or reject unequivocally the break in the curve (presence of active fault).

954 Uplift rates calculated using the time span determined by Th/U ages of travertines covering
955 terrace surfaces (Kele, 2009; Kele et al., 2009, 2011; Sierralta et al., 2009) were 0.1-0.5 mm/a
956 for the Gerecse and Buda Hills. The lower values of these estimates are in agreement with
957 incision rates calculated in this study (Fig. 11, Table S7). We suggest that the higher incision
958 rates were most probably derived from travertines deposited on the valley slope, i.e. not
959 connected to the base level, thus providing excessively young ages relative to the
960 abandonment of the dated terrace. Accordingly, we propose that for the quantification of
961 valley incision and uplift rates the oldest travertine ages of a certain height are to be
962 considered.

963 Th/U series dating of cave sediments in the Buda Hills (Szanyi et al., 2012) issued 0.15—
964 0.32 mm/a uplift rate at the eastern margin of the TR; this is equally in agreement with our
965 projected uplift rate values for the SE margin of the TR (Fig. 11, Table S7).

966 The maximum uplift rate estimated by this study for the Danube Bend is in accordance
967 with novel GPS velocity measurements suggesting that uplift rates in the axial zone of the TR
968 do not exceed 0.5 mm/a (Grenerczy et al., 2005, and oral communication 2014).

969 On the other hand, Ruszkiczay-Rüdiger et al. (2005b) measured very young ^3He minimum
970 exposure ages on the strath terraces of the Danube Bend and derived a maximum incision rate
971 of 1.6 mm/a. On the andesite strath surfaces terrace ages were estimated using surface
972 samples only. We suggest that the sampled strath surfaces were most probably affected by
973 significant denudation, which could not be accounted for using surface samples only. We
974 propose that ^3He concentrations of Ruszkiczay-Rüdiger et al. (2005b) may be better
975 interpreted as denudation rates providing 3-5 m/Ma denudation for the flat or gently dipping
976 strath surfaces ($0-5^\circ$) and 16-17 m/Ma for the lowest horizon with a slope of 10° towards the
977 Danube, which values are reconcilable with denudation rates provided by the depth profiles of
978 this study.

979

980

981 **6. Conclusions**

982

983 The combined application of TCN and post-IR IRSL technique on alluvial terraces can
984 lead to robust age determination. On old terraces TCN depth profile modelling provided better
985 age estimates (or minimum age) than luminescence, even if the site was affected by
986 considerable denudation. On the other hand, luminescence is not sensitive of sediment
987 mixing, high denudation rates or anthropogenic processes like stripping of the upper layers, as
988 long as the sediment is not brought to surface after its burial. Therefore, on locations affected
989 by any of these processes post-IR IRSL was a better approach, as long as sediment burial
990 occurred within its applicability range (300-400 ka).

991 Depth profile modelling of ^{10}Be concentrations of pebble and sand samples at Győr
992 provided similar exposure age - denudation rate pairs suggesting a terrace age of >700 ky and
993 a denudation rate of ~ 6.7 mm/a (Table 6). Accordingly, we suggest that both amalgamated
994 pebble and sand samples are suitable for age determination until the depth profile method is
995 used, because it allows considering their different amount of inherited ^{10}Be . For the highest
996 terrace flight of the study area (tIV-VI) this minimum age is considered as the best estimate
997 due to the lowest denudation rate calculated at this location. Besides, it is reconcilable with
998 the lower minimum ages of the other sample locations (Fig.4).

999 The >700 ka minimum age of terrace abandonment is considered as the lower time
1000 constraint of the onset of the valley incision in the northern part of the Pannonian Basin.
1001 Nevertheless, we emphasize that terraces were formed due to the complex effect of the
1002 propagation of fold-related vertical movements towards the basin interior and of climate
1003 change. The resolution of our data does not enable the differentiation between the effects of
1004 the two processes. However, we can safely conclude that the uplift of the northeastern TR
1005 started before 700 ka, and that the shift from the former wide valley to a narrower terraced
1006 valley may have been triggered by the mid-Pleistocene climate transition (0.7-1.2 ka; Clark et
1007 al., 2006), as it was recognised at several European rivers (Gibbard and Lewin, 2009).

1008 We propose that the evolution of the Hungarian Danube valley was governed by the same
1009 tectonic and climatic processes, and that the highest terrace of the study area (tIV) shares the
1010 same age as the highest terrace horizon (tVI) of the TR. Accordingly, the >700 ka ¹⁰Be
1011 minimum exposure age of the tIV-VI level of the GTT could be extrapolated to the
1012 subsequent valley sections. Maximum incision/uplift rates were calculated accordingly, and
1013 revealed <0.06-0.13 mm/a uplift rates increasing from west to east for the study area, and
1014 enabled an extrapolation suggesting the highest rate of <0.33 mm/a in the axial zone of the
1015 TR (Fig. 11, Table S7). These maximum rates show moderate tectonic deformation of the TR
1016 from the middle Pleistocene onwards. The rate of vertical deformation is in the same order of
1017 magnitude as published in other parts of Europe (Table 1).

1018 Post-IR IRSL data at two locations on the former tIIb terrace enabled the differentiation of
1019 a higher horizon (20 m amrl) with an estimated age of 143±10 ka (Mocsa) and a lower
1020 horizon (12 m amrl) of 91±6 ka (Table 6). These results are the first geochronological
1021 evidences on the definition of the tIIb level of the Hungarian Danube valley as early Würm
1022 (MIS 5b-c), and to distinguish the higher, late Riss (MIS 6) terrace level as tIIIa. We could
1023 not find evidence on the existence of terrace remnants between <700 ka and ~143 ka in the
1024 GTT, therefore suggest that most of the terraces of middle Pleistocene times were possibly
1025 destroyed by the enhanced lateral erosion of the Danube.

1026

1027

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1029

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1047

1048

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1050

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1413
 1414

1415 **Figure captions**

1416

1417 **Fig. 1.** SRTM-based digital elevation model of the Western Pannonian Basin and schematic
 1418 pattern of neotectonic deformation structures (after Fodor et al, 2005; Bada et al, 2007;
 1419 Dombrádi et al, 2010; Ruszkiczay-Rüdiger et al., 2007). DB: Danube Bend; GTT: Győr-
 1420 Tata terrace region, W: location of ^{10}Be exposure age dated, 1.56 ± 0.09 Ma old wind
 1421 polished landforms (Ruszkiczay-Rüdiger et al, 2011).

1422

1423 **Fig. 2A.** Terraces at the marginal zones of the Danube Basin. Terraces in the western side of
 1424 the DB were tentatively mapped using the Surface geological map of the DANREG
 1425 project (Császár et al., 2000). Rectangles show the locations of the multielectrode
 1426 resistivity (MUEL) profiles of Fig. 10. Sample locations: Gy: Győr; B: Bana; Á: Ács; M:
 1427 Mocsá; Gr: Grébics. VB: Vienna Basin, EA: Eastern Alps, TR: Transdanubian Range, G:
 1428 Gerecse Hills.

1429 **Fig. 2B.** Simplified geological cross section of the central part of the Győr-Tata terraces
 1430 (modified after Pécsi, 1959). Location appears on Fig. 2A. Black dots indicate sample sites
 1431 (depth profiles at Ács and Bana).

1432

1433 **Fig. 3.** Longitudinal sketch of the terrace levels along the Danube from the Danube Basin to
1434 the Great Hungarian Plain (modified after Pécsi, 1959). Variations of the elevation of the
1435 horizons indicate Quaternary vertical deformation (orange arrow: uplift, yellow arrow:
1436 subsidence). The study area, the Győr-Tata terraces (GTT, light red shadow) is in the
1437 marginal zone of the uplifting Transdanubian Range. The Danube Bend (DB) is at the axial
1438 zone of the uplifting area characterised by most intensive uplift. Dotted lines indicate
1439 subsurface horizons. Levels below tIIb do not appear in the figure. For location see Fig.1.
1440 Sample sites appear with capital letters as in Fig.2A.

1441

1442 **Fig.4.** Periods of terrace formation in the Hungarian Danube valley according to previous
1443 studies and results of this study plotted above the benthic $\delta^{18}\text{O}$ record and MIS stages
1444 (Lisiecki and Raymo, 2005).

1445

1446 **Fig. 5.** Field images of the sample sites. A: Győr; B: Bana, C: Grébics, D/1: Mocsa gravel pit
1447 (TCN samples), D/2 Mocsa, sand pit (postIR-IRSL samples) E: Ács. For sample codes
1448 refer to Table 2 and Fig. 6. The TCN samples of the Mocsa profile (D/1) were taken in a
1449 gravel pit 1 km away from the postIR-IRSL location in a sand pit of the same terrace
1450 horizon (D/2). BanaB3 and Ács B3 postIR-IRSL samples were collected from the fine sand
1451 underlying the terrace material ~20 meters away from the depth profiles. Hence these
1452 samples do not appear on the photos. Zero point of the scale-bar indicates the original
1453 surface.

1454

1455 **Fig. 6.** The sampled terrace profiles with TCN and post-IR IRSL sample locations. Post-IR
1456 IRSL samples of the Mocsa profile were taken in a sand quarry of the same terrace horizon
1457 1 km away from the TCN location (Mocsa1 and -2 profiles). BanaB3 and Ács B3 IRSL
1458 samples were collected from the underlying late Miocene fine sand ~20 meters away from
1459 the depth profiles from the underlying fine sand. These samples were not considered for
1460 the age determination of the terraces.

1461

1462 **Fig. 7.** Measured ^{10}Be concentrations plotted against sample depth. Black square of the subset
1463 C (Grébics) shows the average ^{10}Be concentration of the mixed upper horizon. Pebbles
1464 means amalgamated pebble sample. Error bars of 1σ analytical uncertainty remain
1465 invisible where they are within the size of the symbols.

1466

1467 **Fig. 8.** Representative examples for not saturated dose-response curves: samples Ács-A1 (A)
1468 and Mocsa-1 (B). Sample Bana-B3 (C) has reached saturation level.

1469

1470 **Fig.9.** Small aliquot De distributions determined with the pIRIR-290 protocol for sub-samples
1471 of Mocsa-1, Mocsa-2, Ács-A2 and Ács-4.

1472

1473 **Fig. 10.** MUEL profiles on terrace tIV-VI horizon. For locations refer to Fig. 2A. Coarse
1474 grained sediments, gravel and sandy gravel layers appear with high resistivity values (>100
1475 Ωm). Small resistivity values are indicative of fine sediments like fine sand and silt (<40
1476 Ωm). MUEL measurements were calibrated by shallow boreholes and outcrop evidences.
1477 MUEL survey was done by the Geomega Ltd. in 2004.

1478

1479 **Fig. 11.** Uplift rates along the Danube using different proxies.

1480 [1] this study; [2] Pécsi (1959), Ruzkiczay-Rüdiger et al. (2005a) and this study, for
1481 terraces up to tIV (420 ka), and the subsidence rate in the Danube Basin was calculated
1482 using 400 m thickness of Quaternary strata (Rónai, 1974); [3] Kele, 2009, Sierralta et al
1483 (2009); [4] Szanyi et al. (2012); [5] Ruzkiczay-Rüdiger et al. (2005b); [6] Grenerczy et al.
1484 (2005). Note the break in the vertical axis above 0.6 mm/a.

1485 The ^{10}Be dating-based trend of incision rates was based on the >700 ka age of the onset of
1486 valley incision suggested by the ^{10}Be exposure age of the highest terrace level (tIV-VI) of
1487 the study area (yellow shading), which was tentatively extrapolated to the entire TR section
1488 of the Danube valley (dotted line). See details in the text. Uplift rate data are presented in
1489 Table S7.

1490

1491 **Table captions appear in a separate file together with the tables**

	incision/uplift rate (mm/a)	dated landform	method	location	timespan (ka)
Antón et al. (2012)	2.0-3.0	strath terraces	cosmogenic ^{10}Be , ^{21}Ne	Duero Basin	100-0
Brocard et al. (2003)	0.8	alluvial terraces	cosmogenic ^{10}Be	French Western Alps	190-0
Giachetta et al. (2015)	0.25-0.55	-	numerical modeling	Iberian Chain	3000-0
Häuselmann et al. (2007a)	~0.12 ~1.2	cave sediments	cosmogenic ^{10}Be , ^{26}Al	Switzerland	>800 800-0
Necea et al. (2013)	0.1-2.2	alluvial terraces	IRSL	SE Carpathians	780-0
Peters and van Balen (2007)	0.01-0.16	alluvial terraces	correlation, relative chronology	Upper Rhine Graben	800-0
Rixhon et al. (2011, 2014)	0.08-0.14*	alluvial terraces	cosmogenic ^{10}Be	NE Ardennes	725-0
Viveen et al. (2012)	0.07-0.08	alluvial terraces	cosmogenic ^{10}Be , OSL, IRSL	Miño River, Iberia	650-0
Wagner et al. (2010)	0.1	cave sediments	cosmogenic ^{10}Be , ^{26}Al	Eastern Alps	4000-0
Ziegler-Dézes (2007)	1.75	-	geodetic data	Variscan Massifs in the Alpine foreland	2600-0
this study**	<0.06-0.13 (up to 0.33***)	alluvial terraces	cosmogenic ^{10}Be	Danube River	>700-0
this study	0.13-0.15	alluvial terraces	post-IR IRSL	Danube River	140-0

Table 1. Some incision/uplift rates derived from terrace studies and geodetic data in Europe.

* The incision/uplift rate was calculated by this study based on the data published by Rixhon et al. (2011, 2014).

** The exposure ages are minimum ages therefore the incision/uplift rates are maximum rates.

*** Incision/uplift rate inferred by data extrapolation along the valley.

Sample	Location	Latitude (DD)	Longitude (DD)	Alt. asl./amrl (m)	Thickness (cm)	Depth (cm)*	Mass of quartz dissolved (g)	¹⁰ Be concentration (atoms/g)		max. denudation rate (m/Ma)
Dan08-02s	Győr	47.6582	17.7251	147	7	49	34.5131	369 537	±11 710	5.8 ±0.2
Dan08-03s	Győr			/38	7	79	35.4654	292 737	±7 891	5.3 ±0.1
Dan08-04s	Győr				7	109	33.0804	230 199	±10 362	4.9 ±0.2
Dan08-04p	Győr				7	109	39.0470	199 552	±5 246	5.9 ±0.2
Dan08-05s	Győr				7	139	37.5003	200 303	±5 130	4.1 ±0.1
Dan08-05p	Győr				7	139	42.0290	158 609	±5 429	5.7 ±0.2
Dan08-06s	Győr				7	184	36.4694	118 567	±5 284	5.3 ±0.2
Dan08-06p	Győr				7	184	41.5389	110 608	±6 288	5.9 ±0.3
Dan08-07s	Győr				7	234	40.0643	133 263	±3 412	2.5 ±0.1
Dan08-08s	Győr				7	284	40.2015	88 745	±4 943	3.3 ±0.2
Dan08-09s	Győr				7	334	41.1239	87 704	±2 241	2.1 ±0.1
Dan08-09p	Győr				7	334	42.3270	73 809	±2 764	3.4 ±0.1
Dan08-10p	Győr				7	394	44.5935	71 048	±1 783	2.4 ±0.1
Dan08-11s	Győr				7	484	39.9080	36 908	±1 489	10.0 ±0.4
Dan08-12s	Győr				7	750	40.7842	16 058	±554	30.7 ±1.1
Dan08-20p	Bana	47.6749	17.88614	153	8	0	39.4791	430 686	±14 233	8.9 ±0.3
Dan08-21p	Bana			/45	8	22.5	40.5013	339 089	±9 589	8.8 ±0.3
Dan08-22p	Bana				8	67.5	43.9504	329 116	±9 450	5.3 ±0.2
Dan08-23p	Bana				8	142.5	46.0544	136 767	±4 210	6.7 ±0.2
Dan08-24p	Bana				8	255	44.0128	99 823	±4 752	3.4 ±0.2
Dan08-25p	Bana				8	375	44.6229	102 889	±2 986	0.7 ±0.0
Dan08-26s	Bana				8	425	40.8223	77 599	±6 185	1.3 ±0.1
Dan08-27p	Ács (actual)	47.7447	18.00424	108 /0	3	0	39.9286	81 254	±2 335	52.7 ±1.5
Dan08-31p	Mocsa	47.6792	18.2133	127	6	135	41.0265	197 414	±6 086	5.3 ±0.2
Dan08-32p	Mocsa			/20	6	185	44.1590	113 208	±3 299	6.6 ±0.2
Dan08-33p	Mocsa				6	245	41.3162	100 597	±3 196	4.4 ±0.1
Dan08-34p	Mocsa				6	305	46.2831	106 516	±3 276	2.1 ±0.1
Dan08-35p	Mocsa				6	405	41.9937	98 208	±3 928	0.7 ±0.0
Dan08-40p	Grébics	47.6588	18.2580	194	8	5	41.9371	340 904	±12 575	11.7 ±0.4
Dan08-41p	Grébics			/87	8	40	41.4907	472 052	±16 606	5.1 ±0.2
Dan08-40-41 avg**	Grébics				8	22	-	406 478	±16 606	7.5 ±0.3
Dan08-42p	Grébics				8	80	40.5622	223 208	±8 519	7.5 ±0.3
Dan08-43p	Grébics				8	115	40.8454	176 707	±6 216	6.7 ±0.2
Dan08-44p	Grébics				8	155	41.4021	157 497	±6 442	5.0 ±0.2
Dan08-45p	Grébics				8	195	40.5254	107 100	±3 793	5.7 ±0.2
Dan08-46p	Grébics				8	225	40.4120	103 398	±5 173	4.5 ±0.2
Dan13-07fs	Ács	47.7425	18.0064	121	6	10	24.6485	77 385	±6 704	55.9 ±4.8
Dan13-08fs	Ács			/12	6	33	22.4840	83 135	±10 860	41.8 ±5.5
Dan13-09fs	Ács				6	63	23.9188	93 930	±7 401	27.8 ±2.2
Dan13-10fs	Ács				6	113	21.9429	83 537	±6 148	20.5 ±1.5
Dan13-11s	Ács				8	135	15.3875	142 269	±8 877	8.8 ±0.6
Dan13-12s	Ács				8	165	15.7705	137 354	±4 827	6.7 ±0.2
Dan13-13ps	Ács				8	195	16.2782	132 721	±5 654	5.1 ±0.2
Dan13-14s	Ács				8	235	15.7782	132 957	±4 804	3.2 ±0.1
Dan13-15ps	Ács				8	275	14.8034	137 143	±5 768	1.9 ±0.1
Dan13-16ps	Ács				8	325	16.0647	127 170	±7 210	1.1 ±0.1
Dan13-17ps	Ács				8	360	15.2088	119 417	±5 157	0.8 ±0.0
Dan13-18ps	Ács				8	400	15.6942	139 288	±6 217	0.1 ±0.0

Table 2. Samples sites and cosmogenic ¹⁰Be data. Samples were corrected for self-shielding. Topographic shielding factor is 1 for all sites. ¹⁰Be/⁹Be ratios of process blanks were $(3.07 \pm 0.7) \times 10^{-15}$ (for samples Dan08-02 to-12); $(1.23 \pm 0.4) \times 10^{-15}$ (samples Dan08-20 to -46), and $(2.37 \pm 0.7) \times 10^{-15}$ (samples Dan13-07 to -18). Ages and denudation rates were not corrected for geomagnetic variations. Alt. asl/amrl: Altitude above sea level/above mean river level; s: sand sample, p: amalgamated pebbles; fs: fine-sand, ps: mixed sand and amalgamated pebbles.

* mean depth of the sample;

**Sample depth and ^{10}Be concentration calculated as the arithmetic mean of the depths and ^{10}Be concentrations of samples Dan08-40 and -41.

Thresholds	Győr		Bana	Grébics	Mocsa	Ács*
	(p)	(s)				
χ^2 cutoff	2	15	6.5	5	31	<2.5**
min erosion rate (cm/ka)	0.2	0.2	0.5	0.2	0.1	0.0
max erosion rate (cm/ka)	1.2	1.0	1.5	1.4	1.2	1.0
total erosion min. (cm)	500	500	500	600	1	0
total erosion max. (cm)	1500		1500	1600	500	500
min. age (ka)	150		150	150	14	0
max. age (ka)	2600		2600	2600	360	100
site specific spallogenic production rate (at/g/a)	4.6185		4.6415	4.8394	4.5398	4.5124

Table 3. Monte Carlo simulation parameters used for the depth profile simulator of Hidy et al (2010). 100 000 solutions were obtained for each simulation. Győr (p): amalgamated pebble samples; Győr (s): sand samples.

*Only the buried exposure time of the alluvial sediments could be simulated. See details in text.

** 2σ confidence level.

	<i>age (ka)</i>	<i>inheritance (10³ at/ g)</i>	<i>denudation rate (m/Ma)</i>
Győr - pebble			
Bayesian most probable	1 561	70	6.5
Bayesian 2σ upper	2 450	97	9.3
Bayesian 2σ lower	693	59	5.1
Bayesian 1σ upper	2 064	84	8.0
Bayesian 1σ lower	993	65	5.8
Győr - sand			
Bayesian most probable	1 783	95	7.0
Bayesian 2σ upper	2 448	149	8.3
Bayesian 2σ lower	708	72	5.2
Bayesian 1σ upper	2 098	122	7.5
Bayesian 1σ lower	1 024	83	5.8
Bana			
Bayesian most probable	1242	106	10.0
Bayesian 2σ upper	1541	140	12.2
Bayesian 2σ lower	477	92	8.7
Bayesian 1σ upper	1332	127	11.1
Bayesian 1σ lower	640	100	9.4
Grébics			
Bayesian most probable	1482	85	8.7
Bayesian 2σ upper	2003	133	11.0
Bayesian 2σ lower	659	53	7.1
Bayesian 1σ upper	1696	109	9.8
Bayesian 1σ lower	883	70	7.8
Mocsa			
Bayesian most probable	149	91	5.8
Bayesian 2σ upper	352	99	10.4
Bayesian 2σ lower	99	83	1.3
Bayesian 1σ upper	309	95	7.9
Bayesian 1σ lower	136	87	3.1
Ács (buried exposure)			
Bayesian most probable	12	127	9.9
Bayesian 2σ upper	65	138	9.8
Bayesian 2σ lower	0	114	0.2
Bayesian 1σ upper	38	132	8.6
Bayesian 1σ lower	6	120	1.7

Table 4. Best constrained model solutions of the ¹⁰Be depth profile simulator (Hidy et al., 2010). For simulation settings refer to Table 3. Complete data tables appear in Table S1). Bayesian 1σ and 2σ upper and lower values represent the 68 and 95 % uncertainties of the Bayesian most probable values.

Sample name	Depth [m]	Grain size [μm]	Dose rate [Gy/ka]	Equivalent dose [Gy] pIRIR- 290	Age [ka] pIRIR- 290	Saturation*
Bana-A1	4.1	150-250	2.26 \pm 0.21	> 924.6	> 410	0.87
Bana-A2	4.6	200-250	2.56 \pm 0.15	> 988.8	> 386	0.88
Bana-B3	10.4	150-200	3.40 \pm 0.19	> 1119.1	> 320	0.95 \pm 0.01
Mocsa-1	1.0	100-150	2.72 \pm 0.17	380.0 \pm 10.0	139 \pm 9	
Mocsa-2	1.6	250-300	2.65 \pm 0.17	386.8 \pm 9.5	146 \pm 10	
Ács-A1	1.0	150-200	3.03 \pm 0.16	46.2 \pm 3.2	15.2 \pm 1.3	
Ács-A2	1.7	200-250	2.68 \pm 0.15	253.1 \pm 4.7	95 \pm 6	
Ács-B3	8.4	150-200	3.51 \pm 0.17	> 903.6	> 258	0.93 \pm 0.02
Ács-A4	3.9	150-200	2.78 \pm 0.15	238.0 \pm 12.1	86 \pm 6	

Table 5. Sample characteristics for coarse-grained K-feldspar post-IR IRSL-290 dating. Sample depth, grain size, dose rate, equivalent dose and corresponding age. *Saturation levels were measured on one aliquot/sample for Bana-A1 and A2, and on 3 aliquots/sample for Bana-B3 and Ács-B3. The data table for the pIRIR-225 protocol is Table S6.

	<i>Győr-pebble</i>	<i>Győr-sand</i>	<i>Bana</i>	<i>Grébics</i>	<i>Mocsa</i>	<i>Ács</i>
Exposure age (ka)	>693	>708	>477	>659<	149 ⁺¹⁶⁰ ₋₁₃	–
Denudation rate (m/Ma)	6.5 ^{+1.6} _{-0.7}	7.0 ^{+0.5} _{-1.2}	10.0 ^{+1.1} _{-0.6}	8.7 ^{+1.2} _{-0.8}	5.8 ^{-2.1} _{+2.6}	–
Inheritance (10³ at/g)	70 ⁺¹⁴ ₋₅	95 ⁺²⁷ ₋₁₂	106 ⁺²¹ ₋₆	85 ⁺²⁴ ₋₁₅	91 ⁺⁴ ₋₄	–
Post-IR IRSL age	–	–	>410	–	143 \pm 10*	91 \pm 6*
Terrace level (traditional - this study)	tIV - tVI	tIV - tVI	tIV - tVI	tIV - tVI	tIIb - tIIIa	tIIb
height above the Danube (m)	38	38	45	87	20	12

Table 6. Exposure time, denudation rate, inheritance and post IR IRSL age of the studied terraces. Bayesian most probable values, with a 1 σ confidence window are presented. Minimum exposure ages are Bayesian 2 σ lower ages. Height above Danube refer to the present elevation of the surface of the alluvial material.

* arithmetic mean of the measured ages in Table 5.

Figure1
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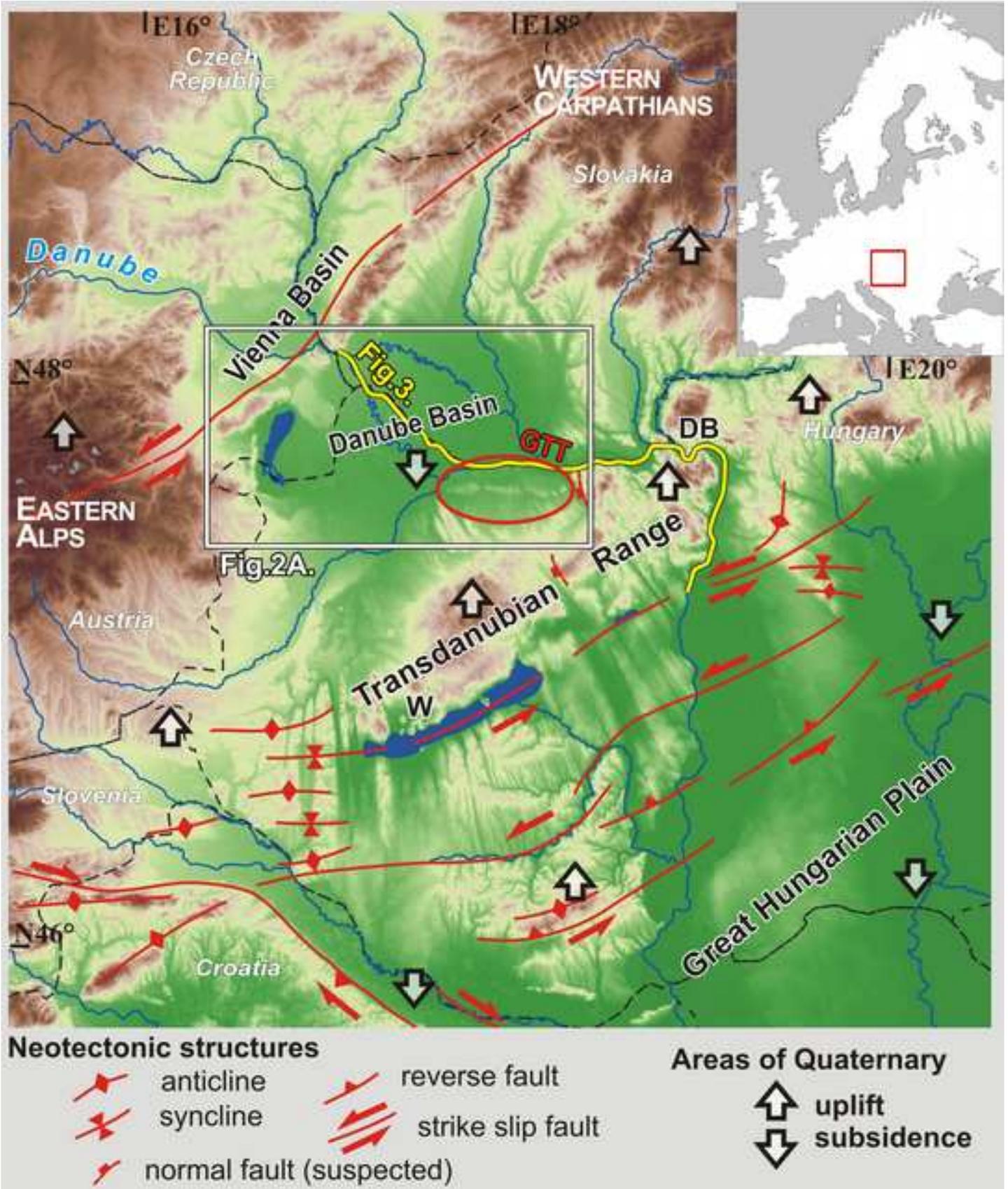


Figure2

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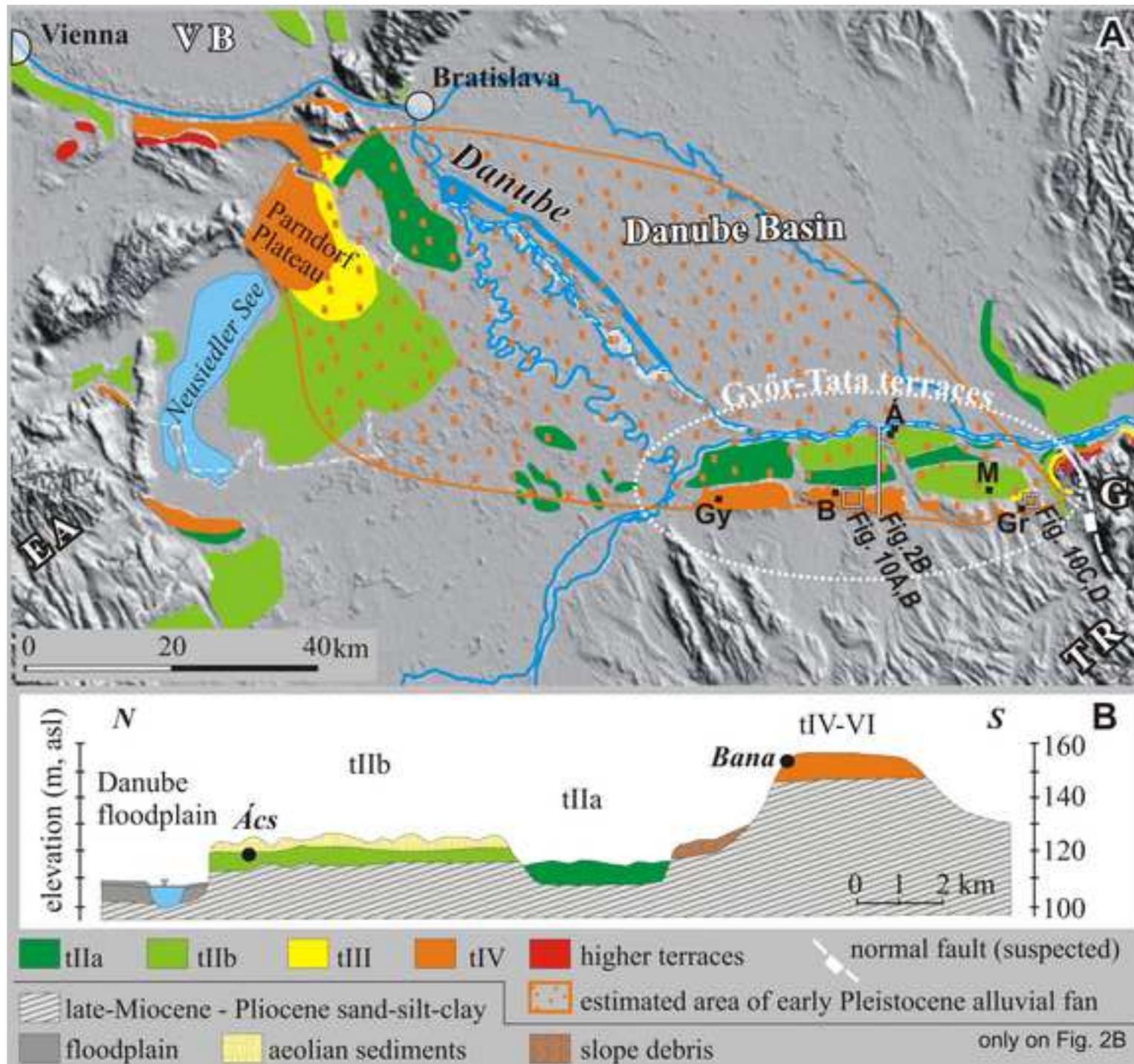


Figure3
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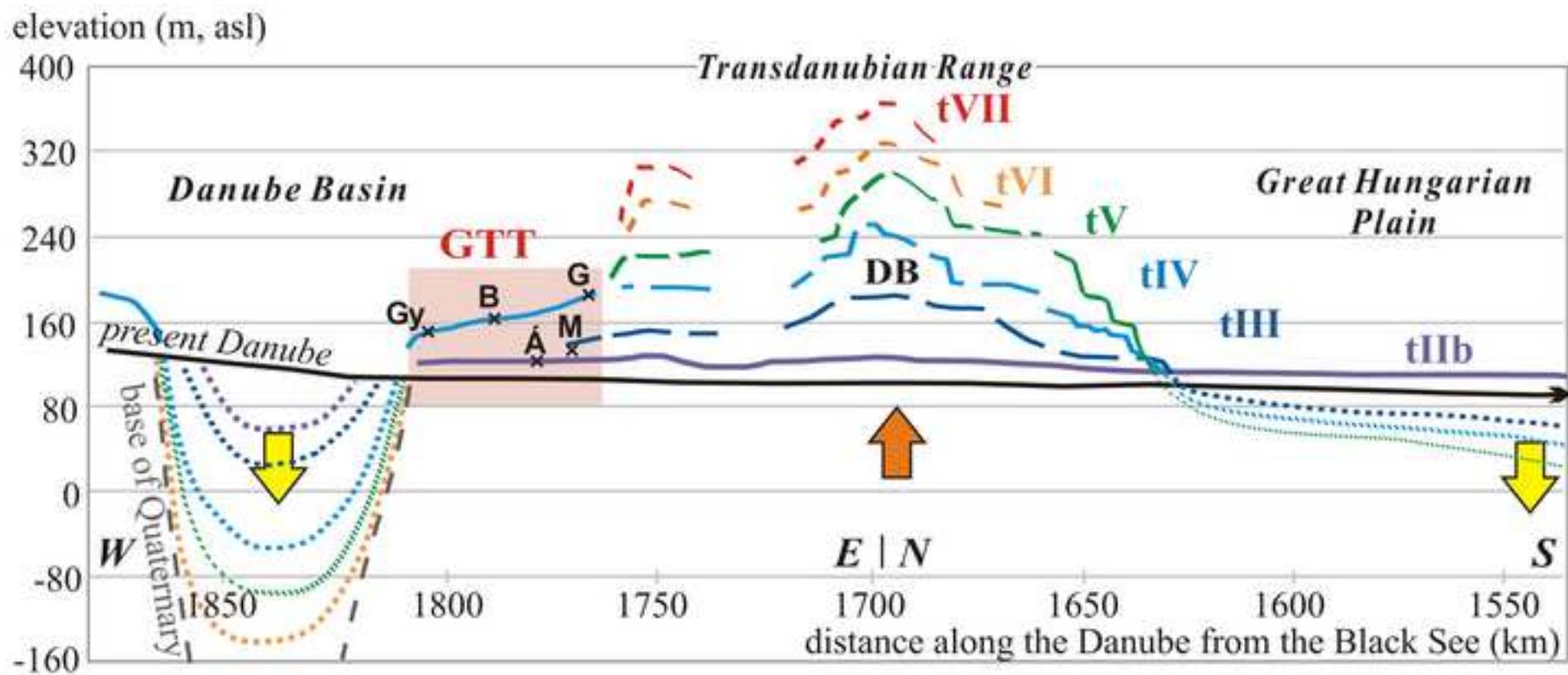


Figure4

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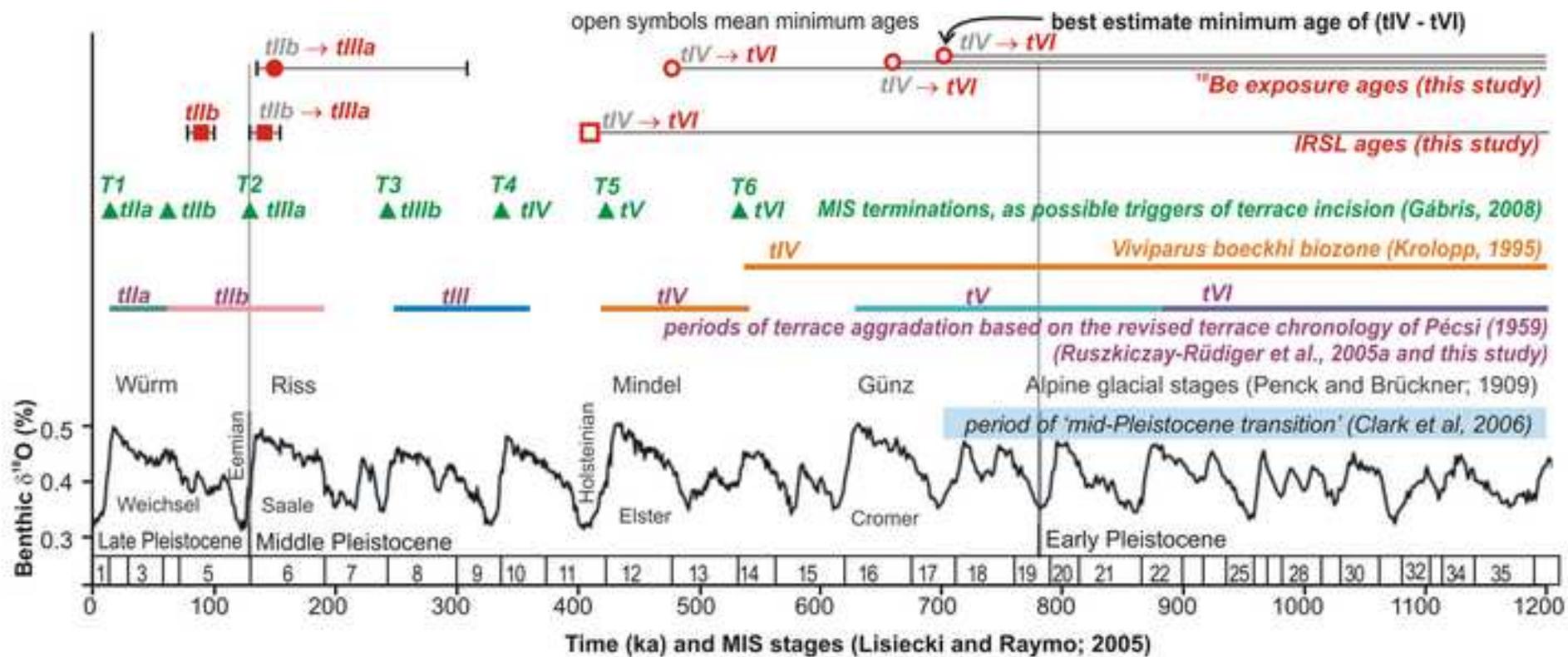


Figure 5

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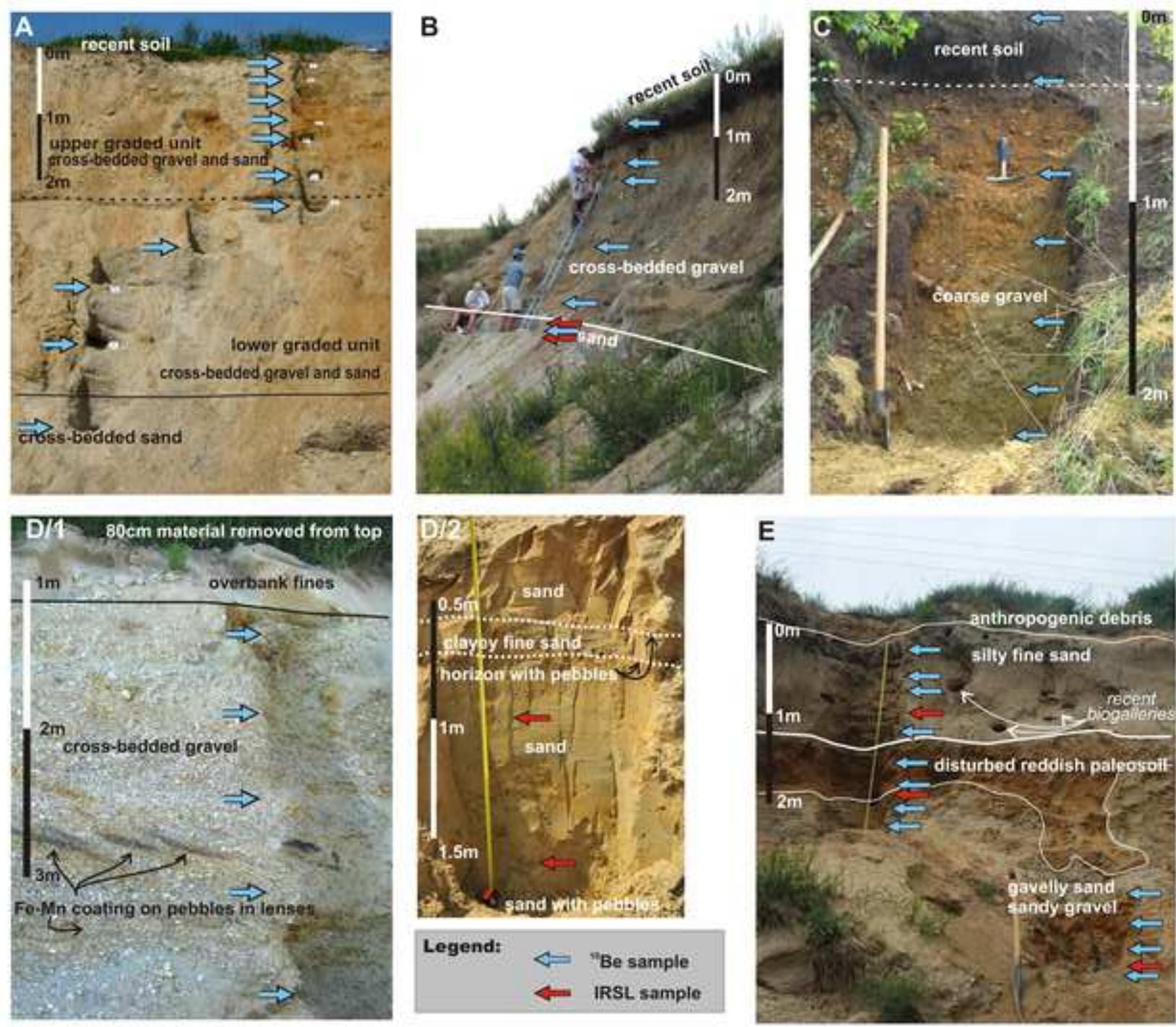


Figure6

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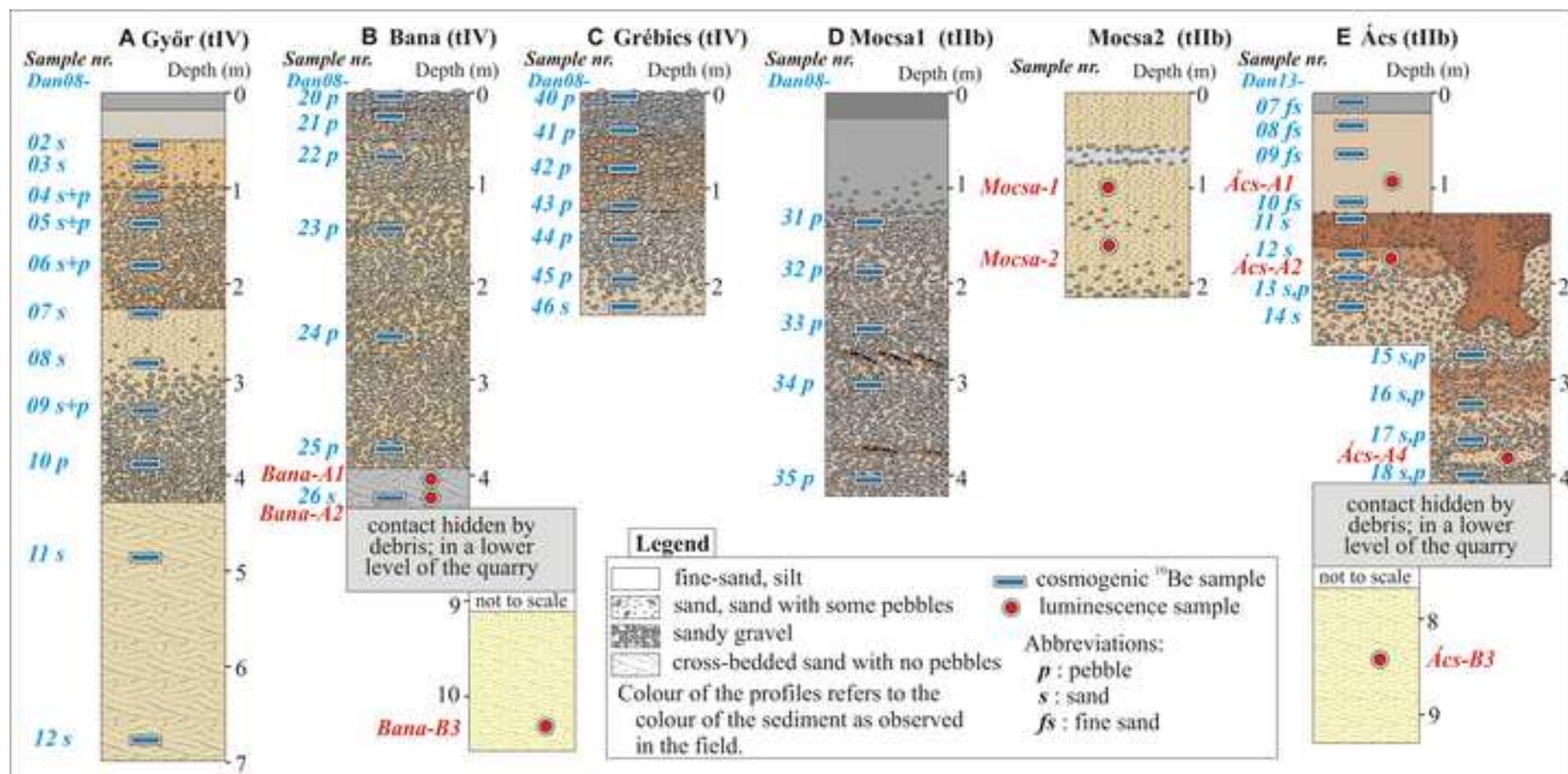


Figure 7

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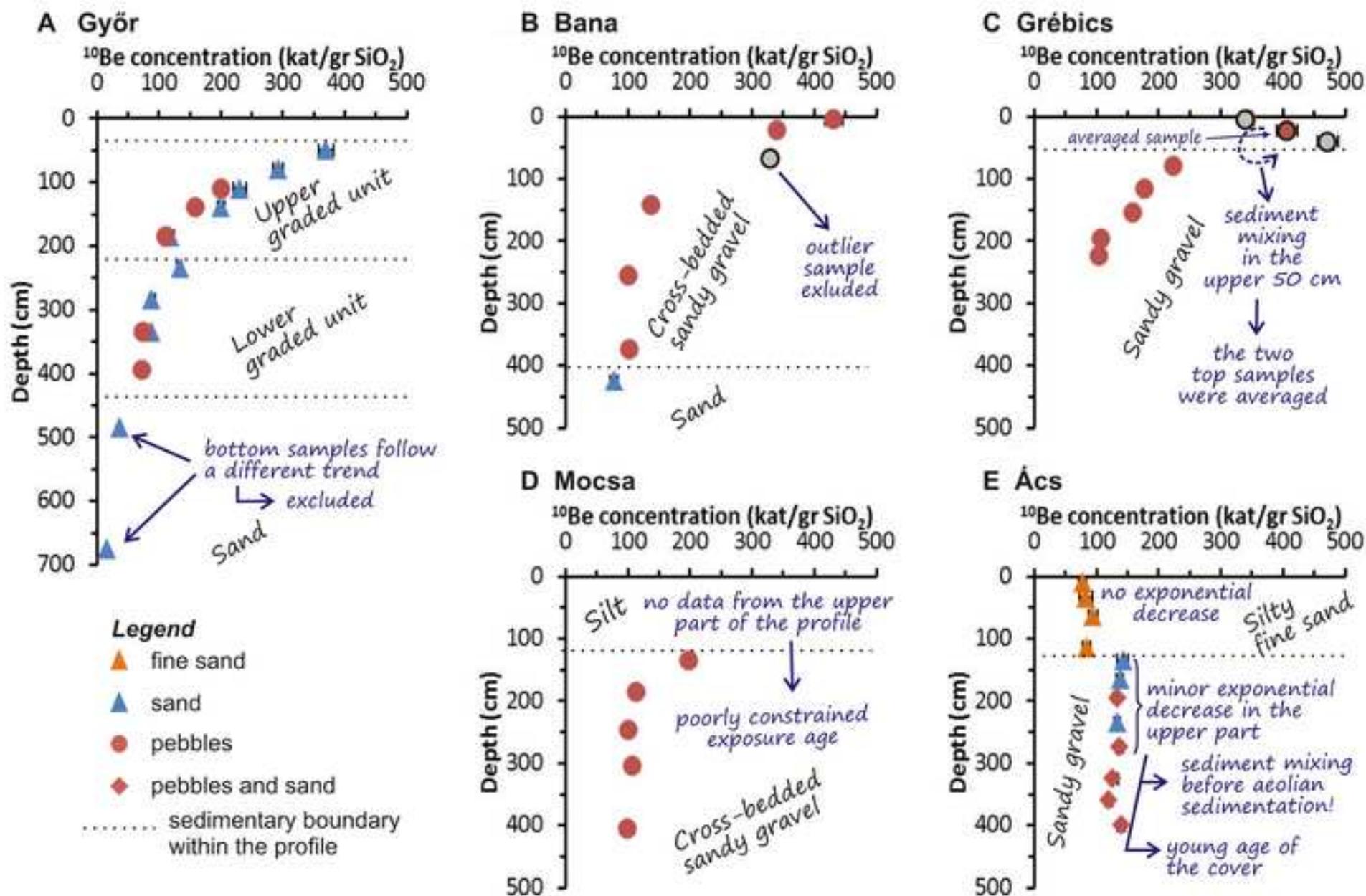


Figure8

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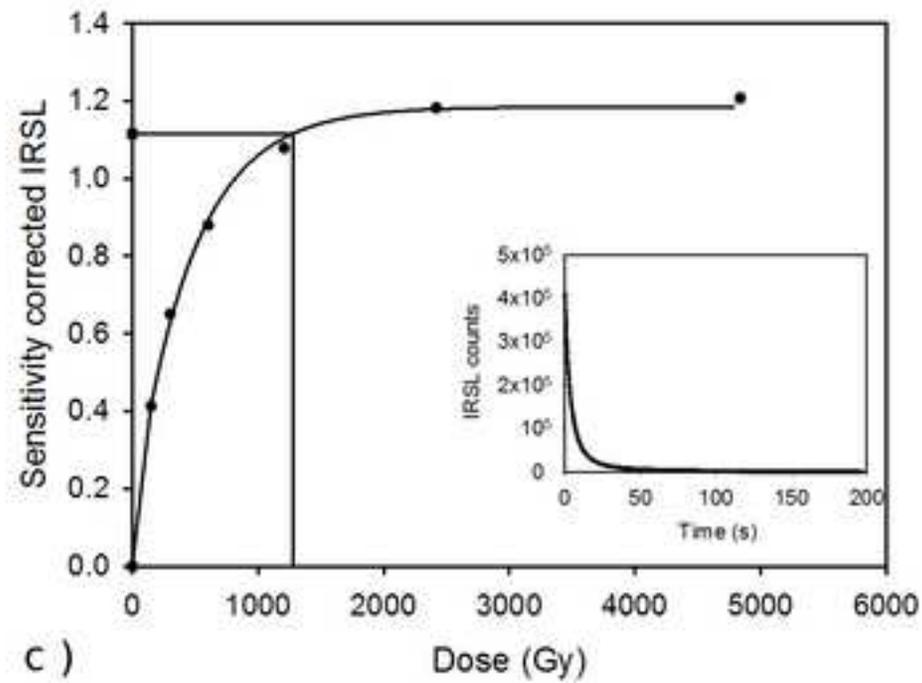
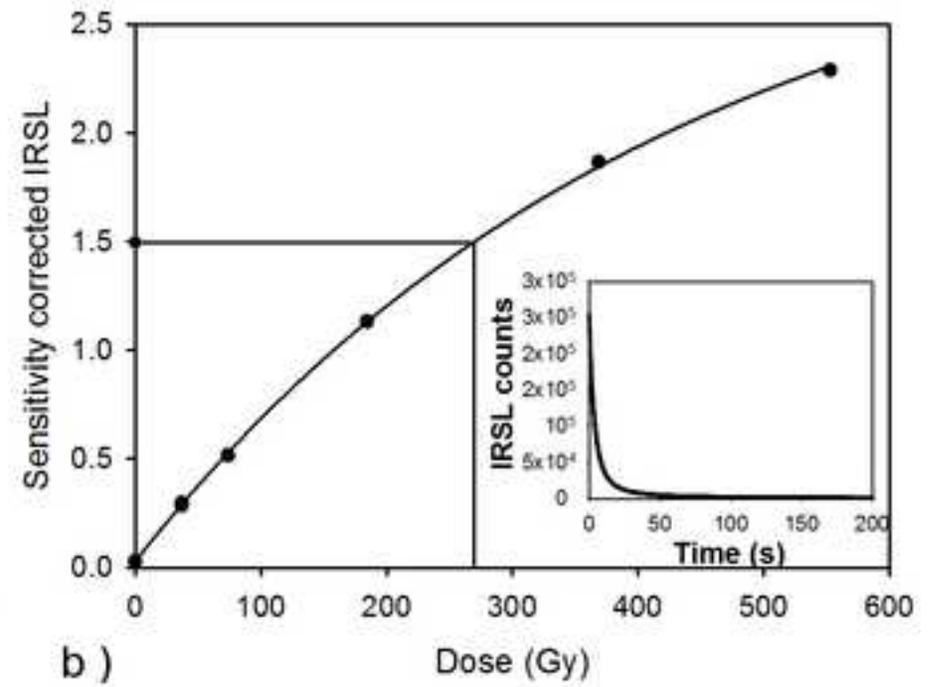
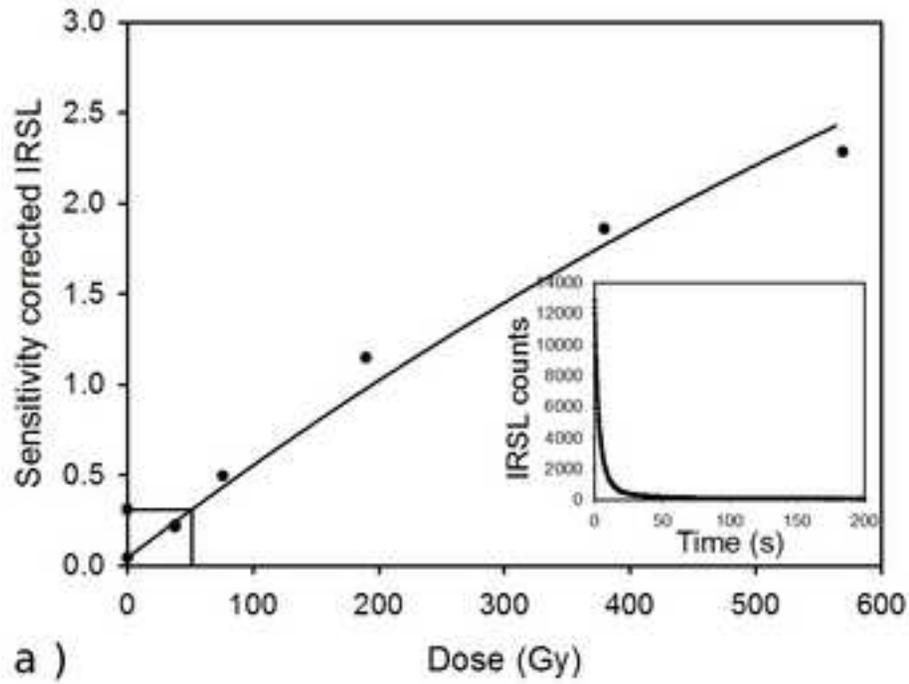


Figure9

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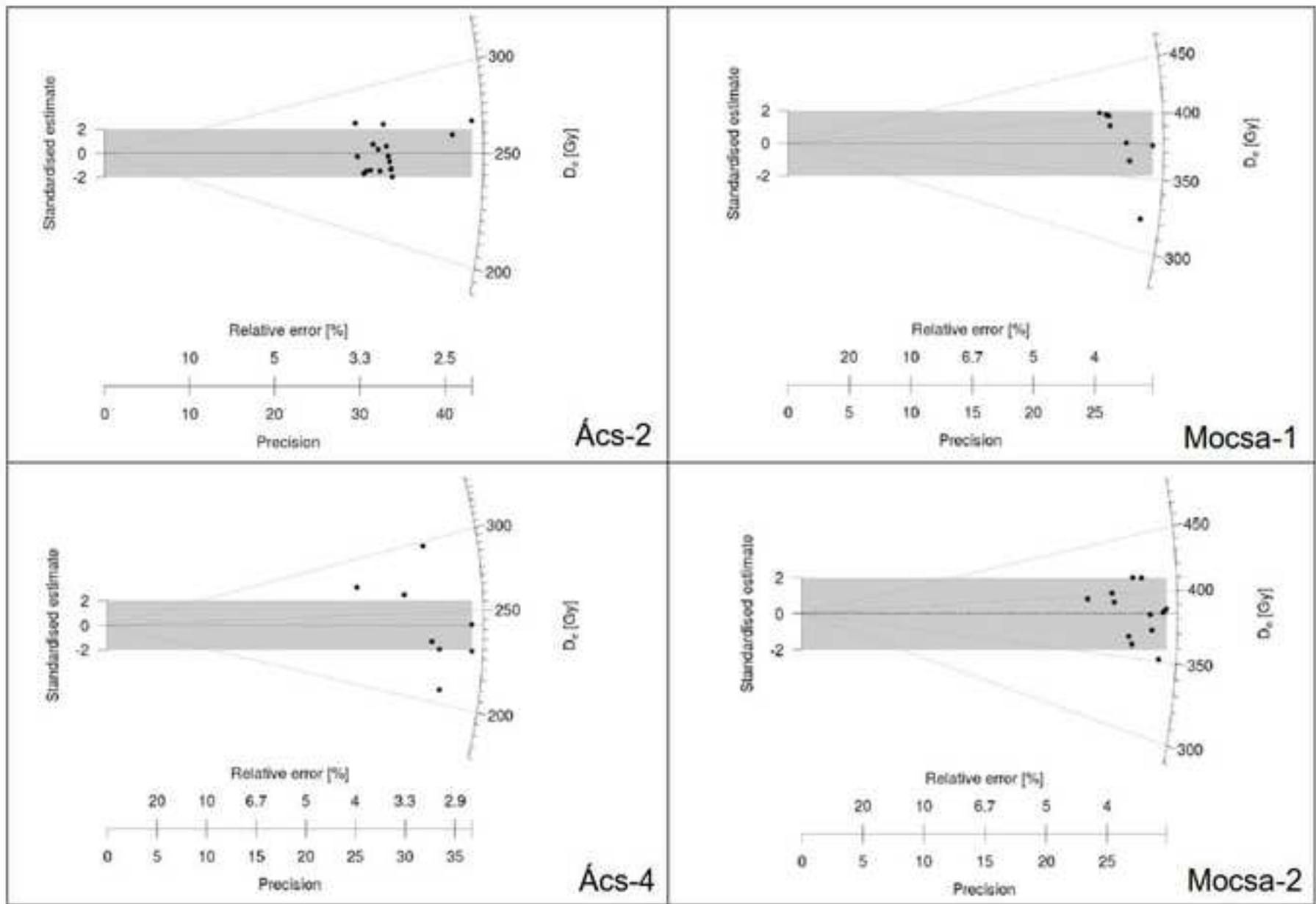


Figure10
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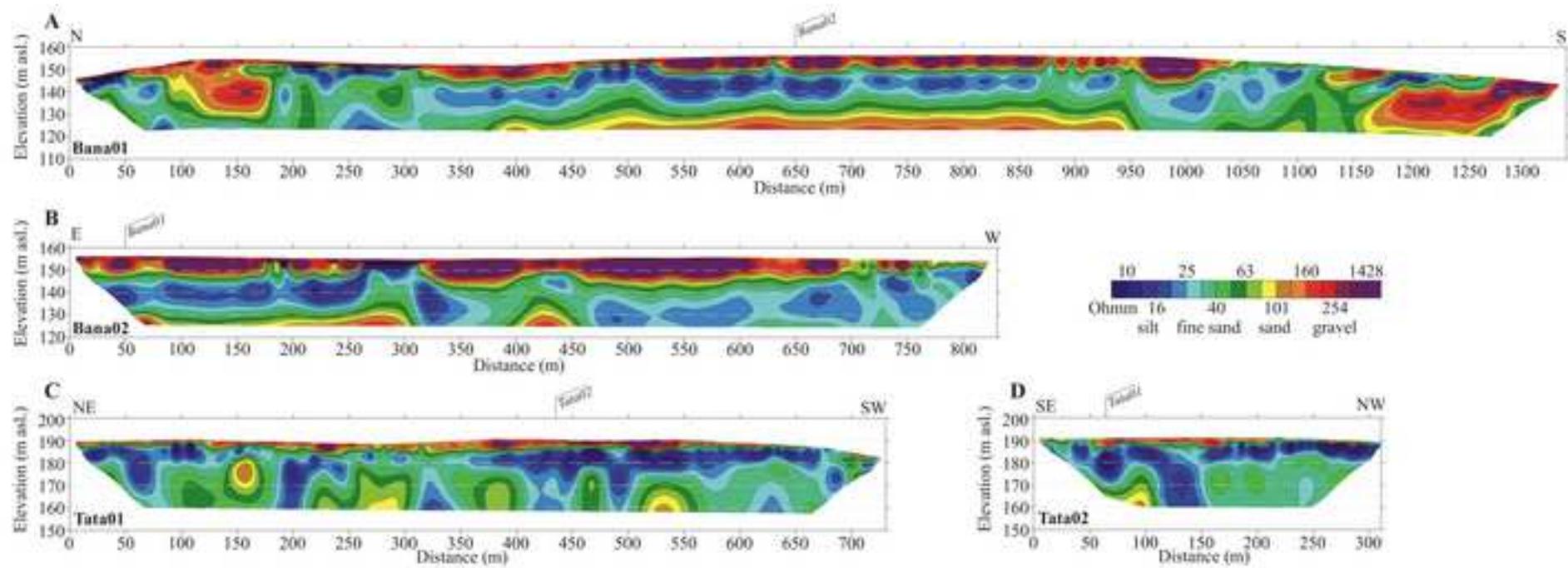
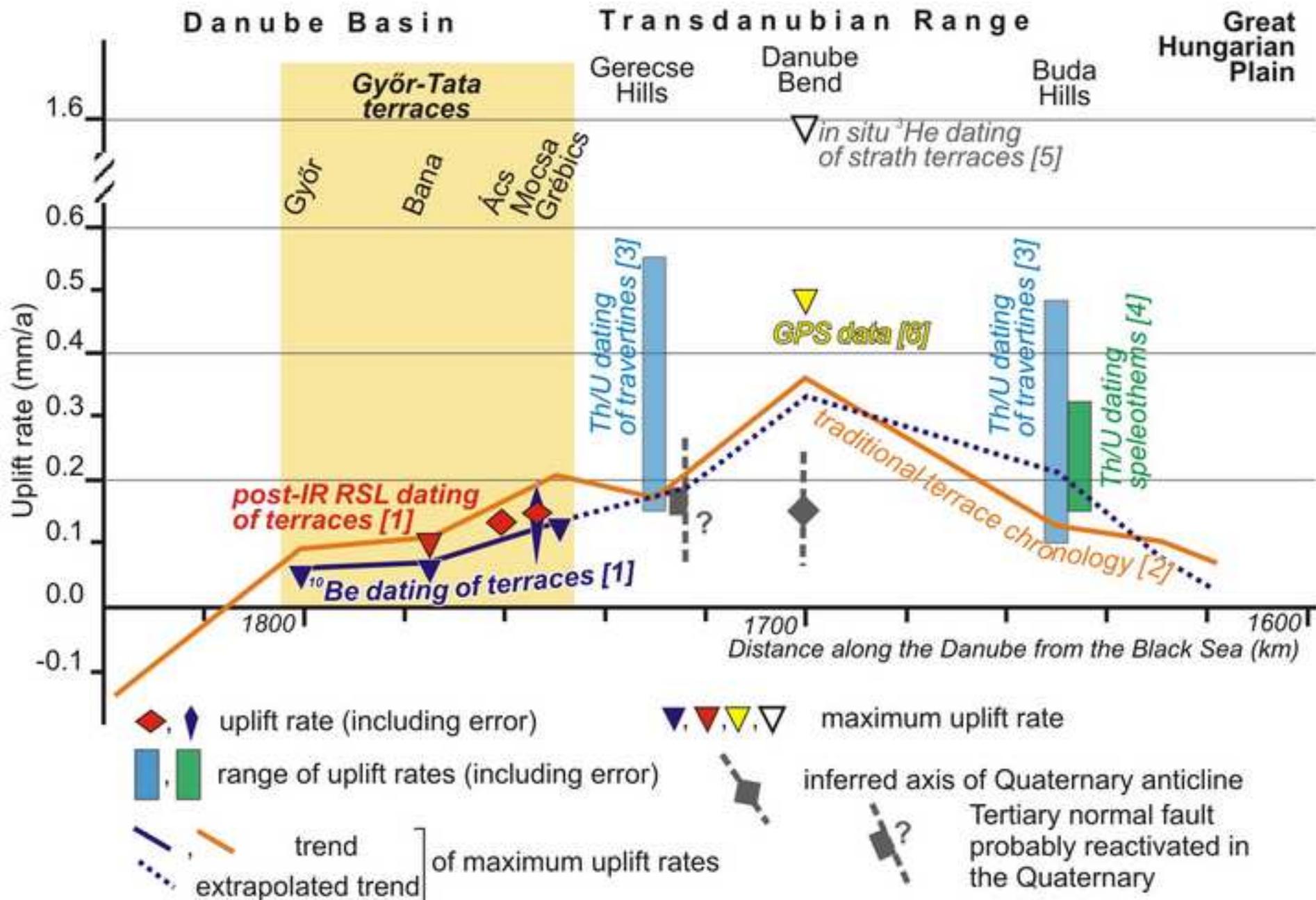


Figure11

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Supplementary information

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matlab code 2

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matlab code 3

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