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

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Abstract

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

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

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

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

1. Introduction

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

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

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

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

2. Geological and geomorphological setting

2.1. Quaternary tectonics and drainage pattern evolution

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

3. Material and methods

3.1. Sampling strategy

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

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

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

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

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

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

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

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

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

3.2.1. Theoretical considerations

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

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

eq(1):

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

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

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

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

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

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

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

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

3.2.2. Laboratory analysis

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

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

3.3. Luminescence dating

3.3.1. Theoretical considerations

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

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

3.3.2. Experimental setup

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

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

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

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

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

3.4. Multielectrode resistivity profiles

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

4. Results

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

To simulate this profile, two models can be proposed:

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

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

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

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

4.4.1 Dose-recovery, residuals, fading

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

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

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

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

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

4.4.2 Equivalent dose, incomplete bleaching, saturation, age estimates

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

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

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

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

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

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

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

5. Discussion

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

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

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

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

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

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

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

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

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

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

5.3. Climate control on the onset of terrace formation

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

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

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

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

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

5.4. Denudation of the terrace surfaces

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

Cosmogenic ^{10}Be depth profiles provided well constrained denudation rates of the terrace surfaces (Table 6). The highest denudation rates were calculated at Bana $10.0^{+1.1}_{-0.6}$ m/Ma) and at Grébics ($8.7^{+1.2}_{-0.8}$ m/Ma), at the edge of the terrace remnant and on the small, isolated terrace hill at the most elevated part of the terrace region, respectively. Lower denudation rate was 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 Mocsá. Considering the minimum exposure ages of the terraces these rates suggest 5-8 m of material eroded from the tIV-VI terraces and ~1.0 m denudation from the tIIb-tIIIa at Mocsá.

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

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

5.5. Cryoturbation features

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

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

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

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

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

5.6. Incision rates and tectonic uplift

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

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

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

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

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

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

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

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

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

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

6. Conclusions

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

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

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

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

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

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Figure captions

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

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

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

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

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

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

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

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

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

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

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

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

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

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

Table captions appear in a separate file together with the tables

Table

	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.

** 2 σ 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 ± 0.21	> 924.6	> 410	0.87
Bana-A2	4.6	200-250	2.56 ± 0.15	> 988.8	> 386	0.88
Bana-B3	10.4	150-200	3.40 ± 0.19	> 1119.1	> 320	0.95 ± 0.01
Mocsa-1	1.0	100-150	2.72 ± 0.17	380.0 ± 10.0	139 ± 9	
Mocsa-2	1.6	250-300	2.65 ± 0.17	386.8 ± 9.5	146 ± 10	
Ács-A1	1.0	150-200	3.03 ± 0.16	46.2 ± 3.2	15.2 ± 1.3	
Ács-A2	1.7	200-250	2.68 ± 0.15	253.1 ± 4.7	95 ± 6	
Ács-B3	8.4	150-200	3.51 ± 0.17	> 903.6	> 258	0.93 ± 0.02
Ács-A4	3.9	150-200	2.78 ± 0.15	238.0 ± 12.1	86 ± 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±10*	91±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 1σ confidence window are presented. Minimum exposure ages are Bayesian 2σ 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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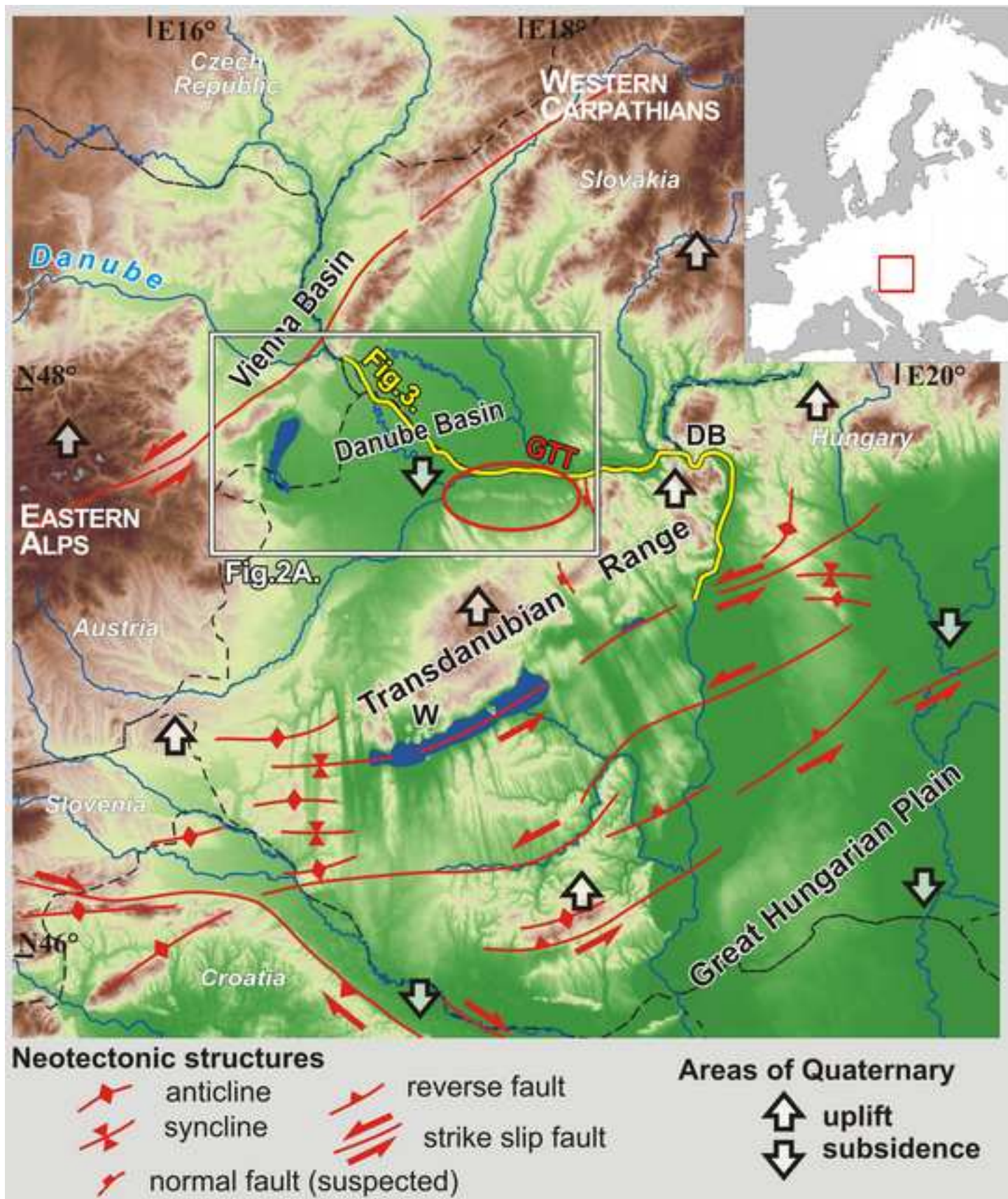


Figure 2

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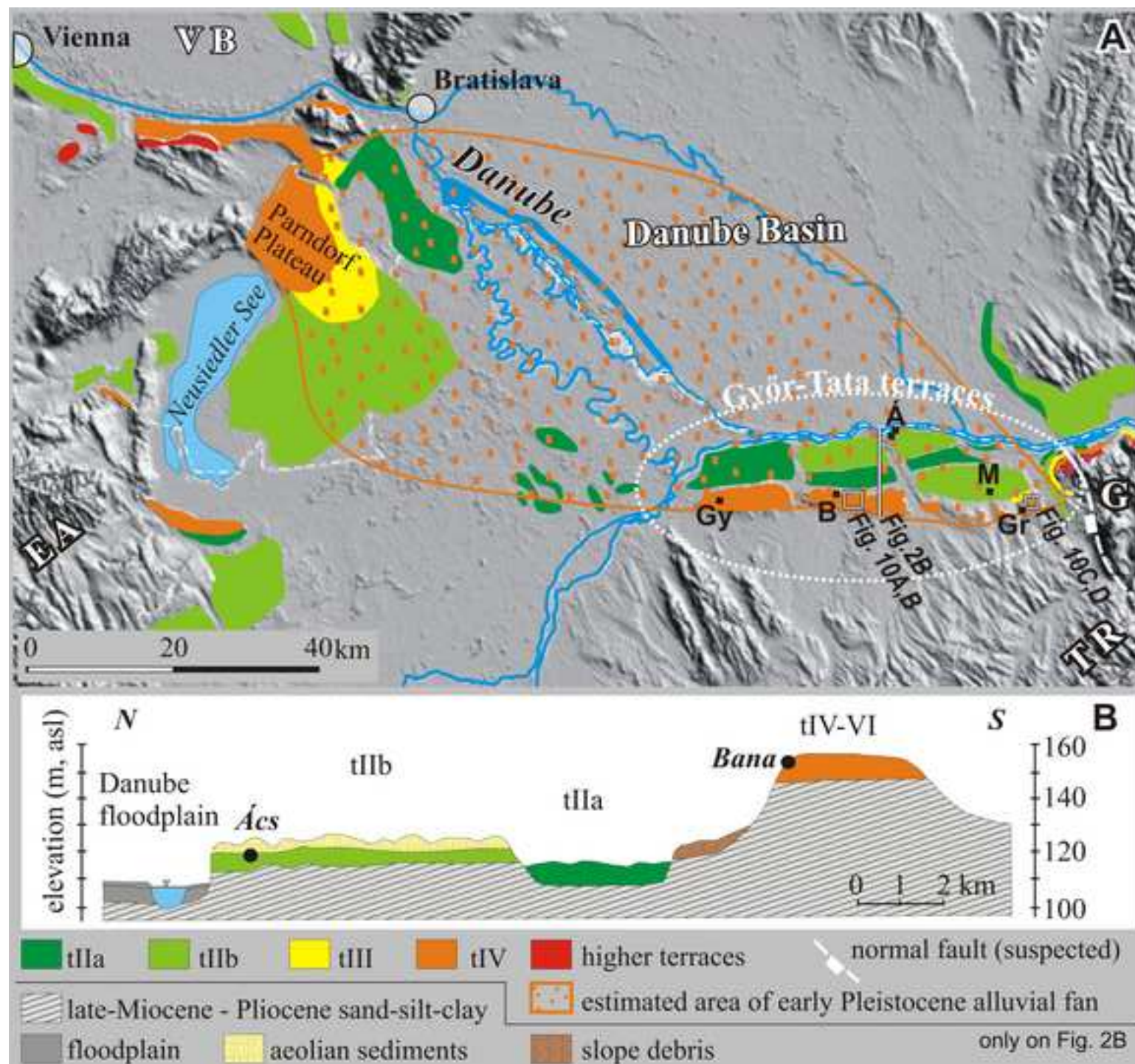


Figure3

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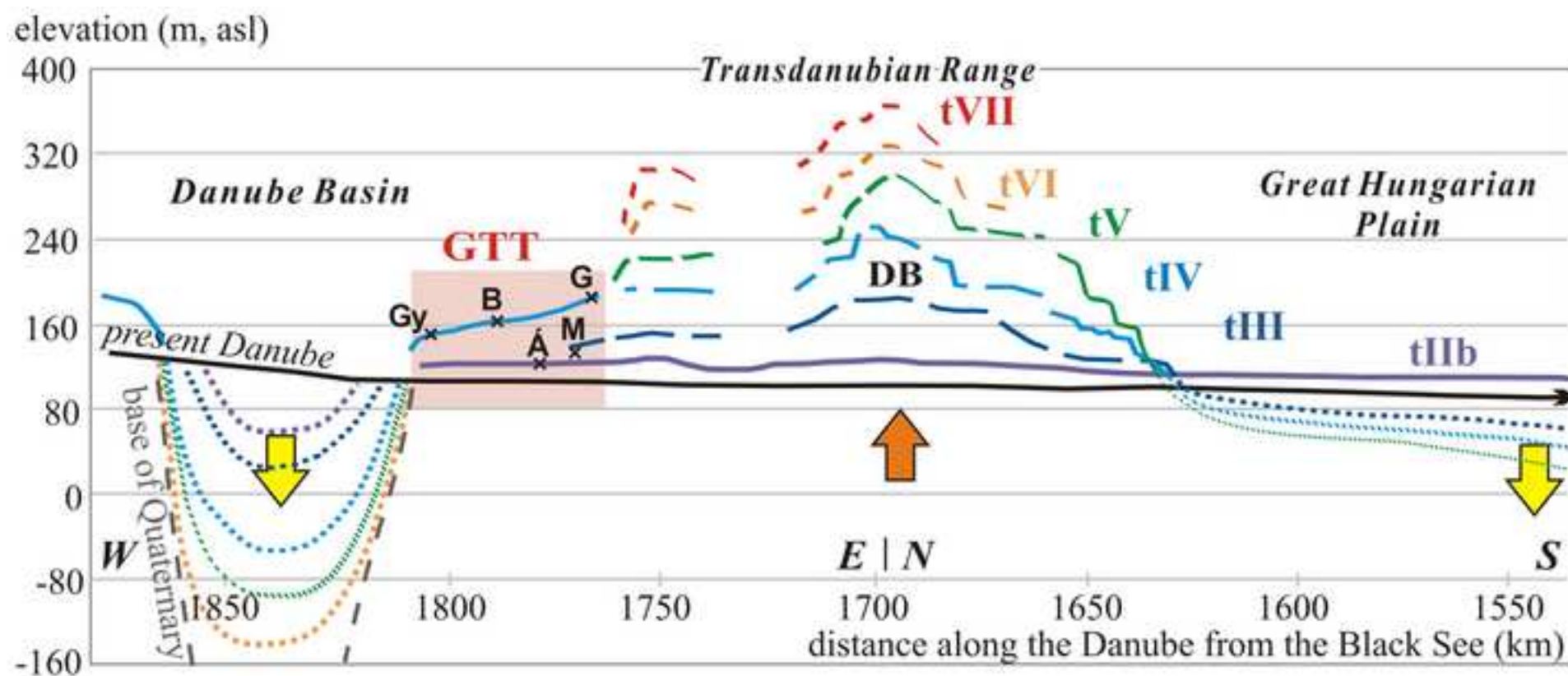


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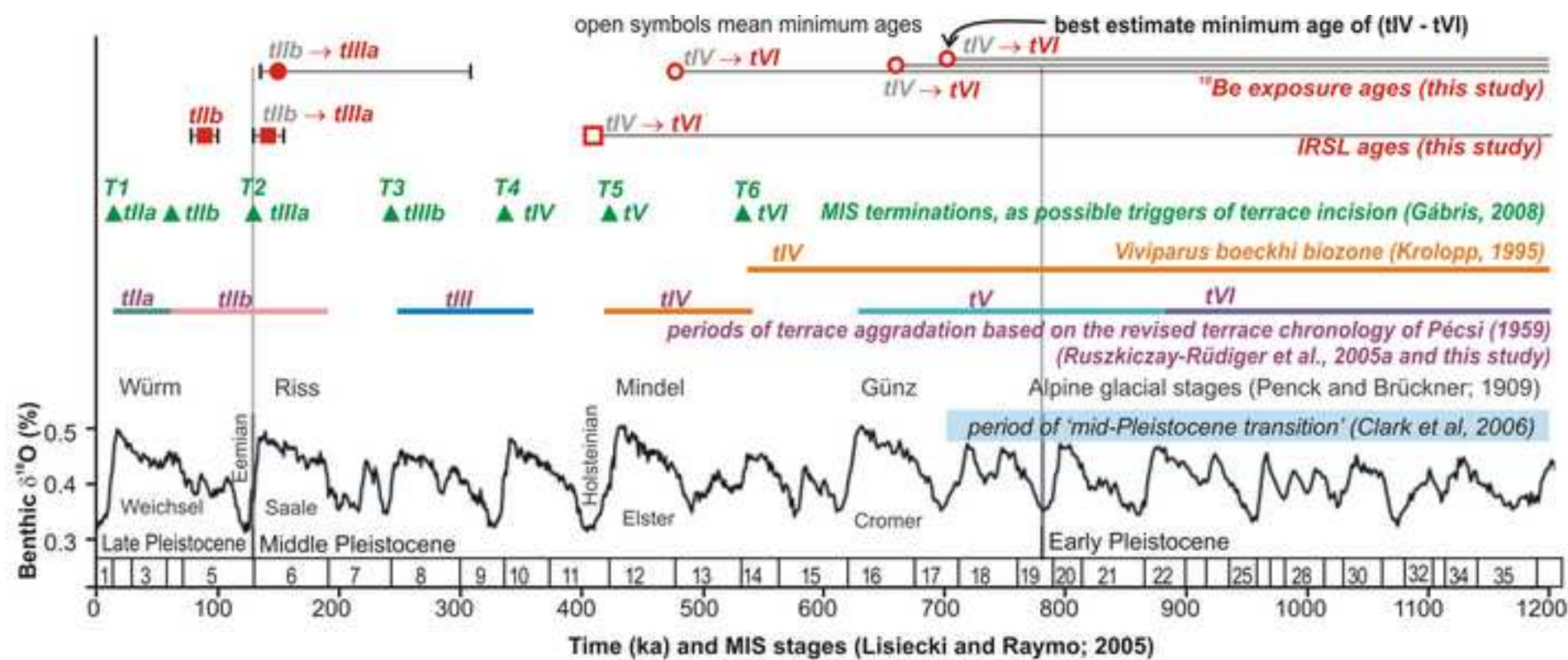


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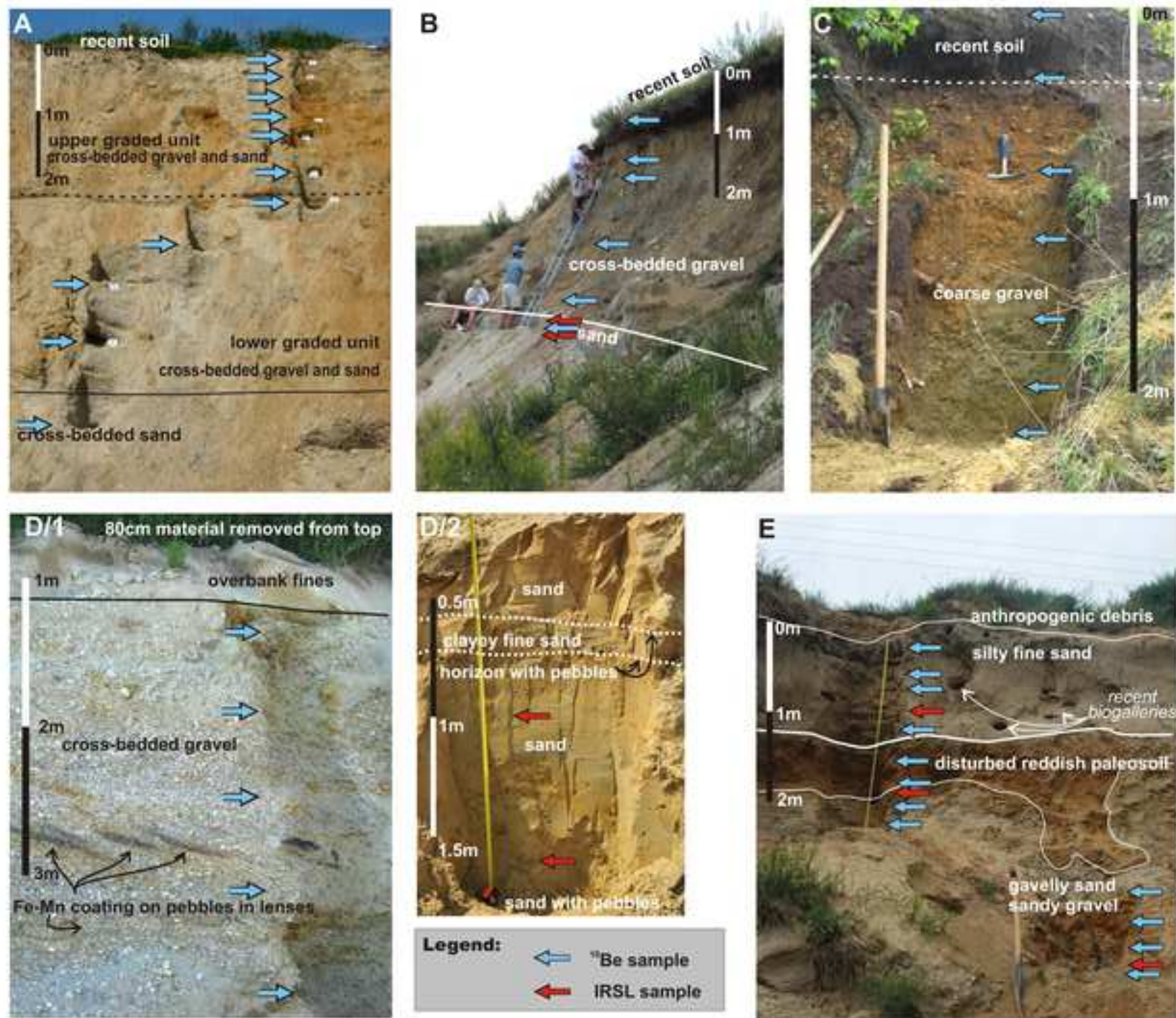


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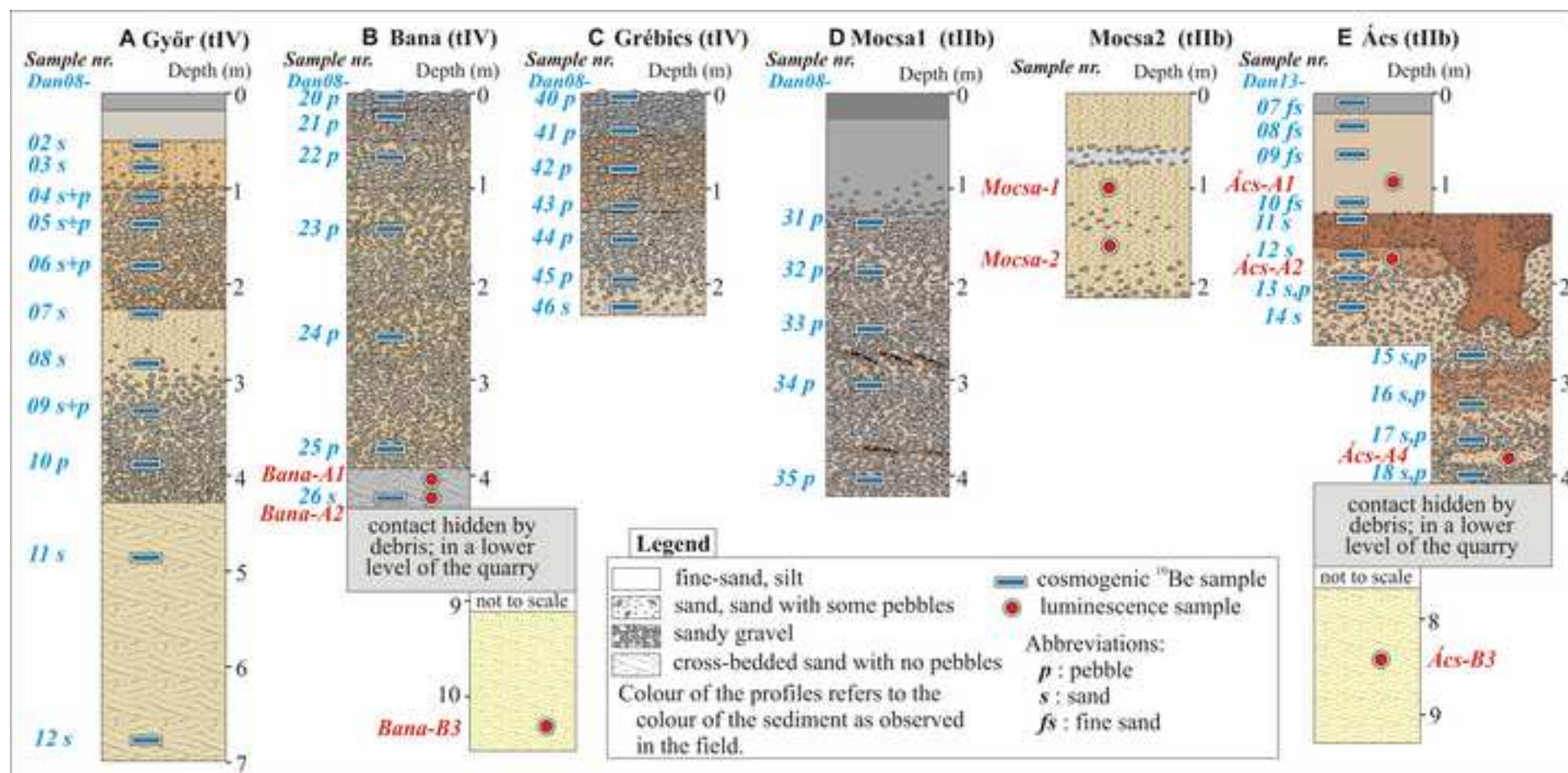


Figure7

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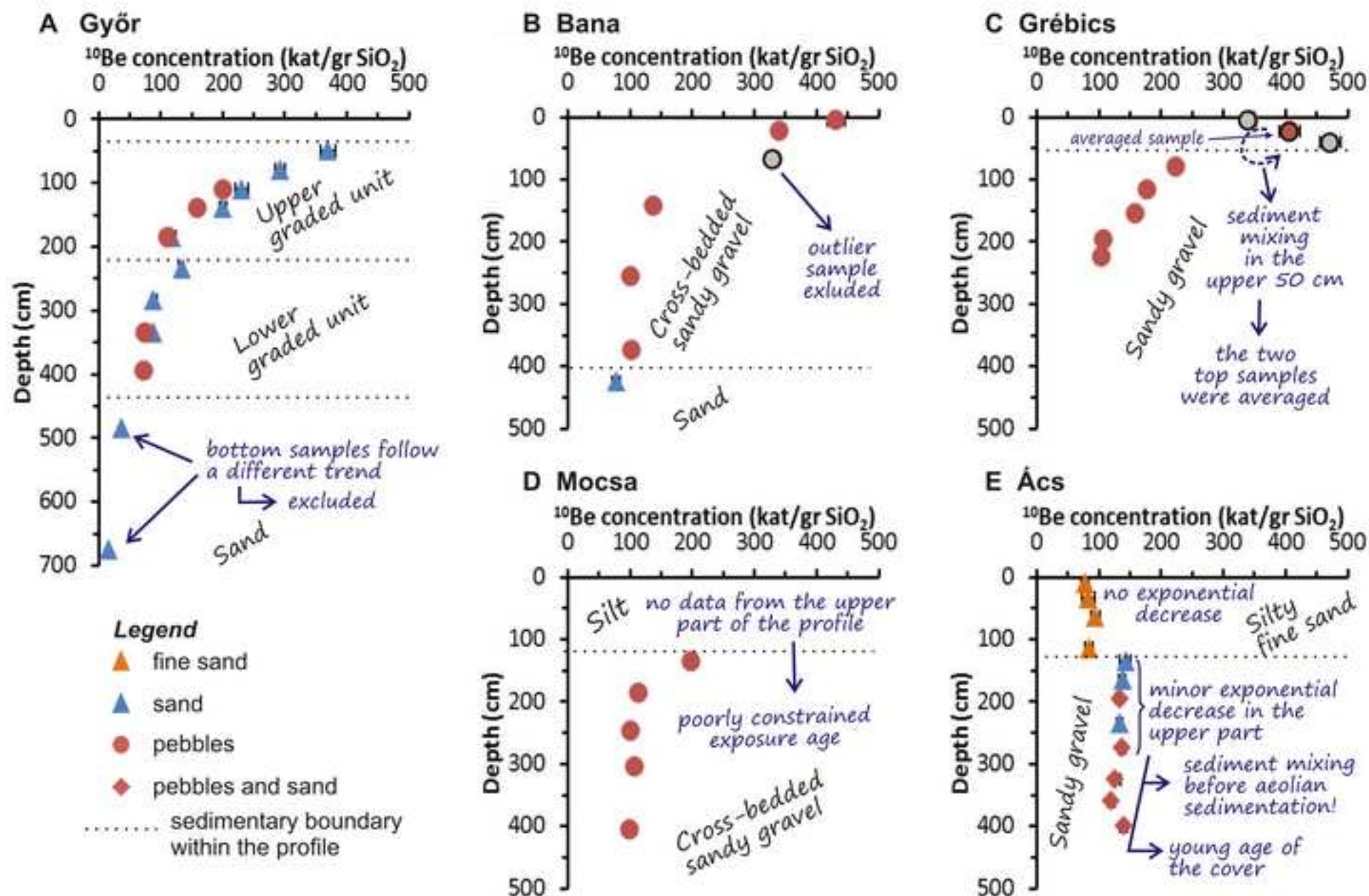


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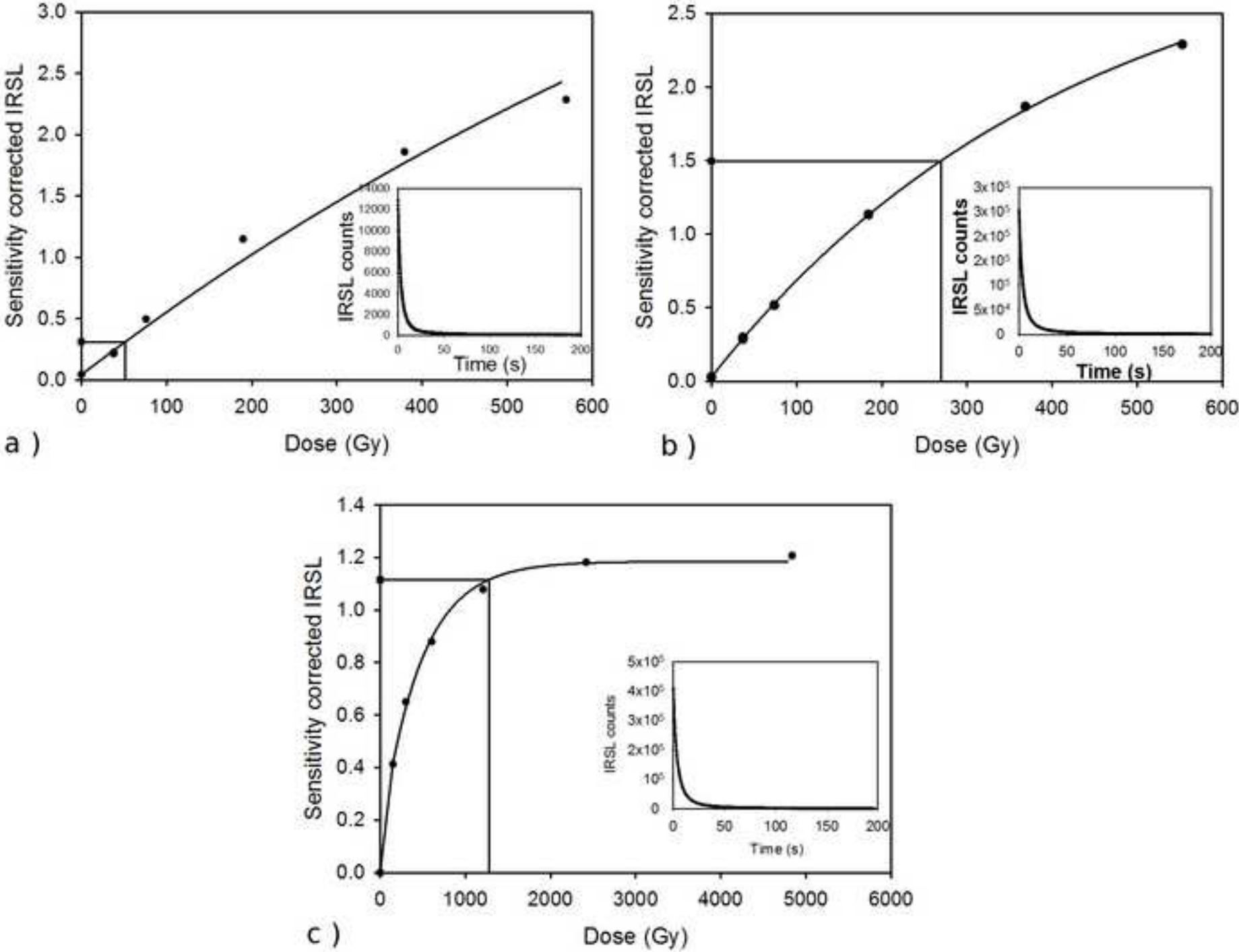


Figure9

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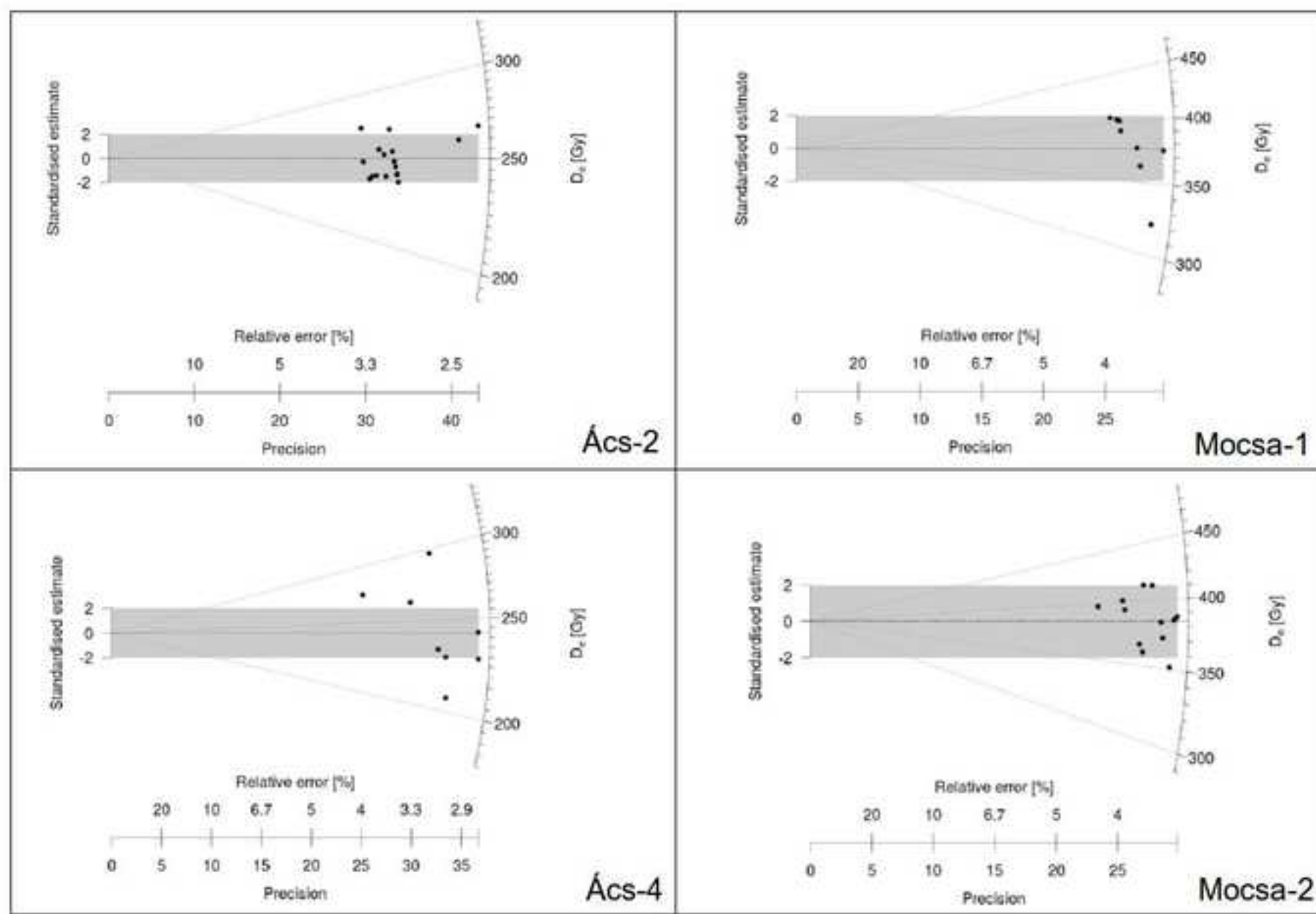


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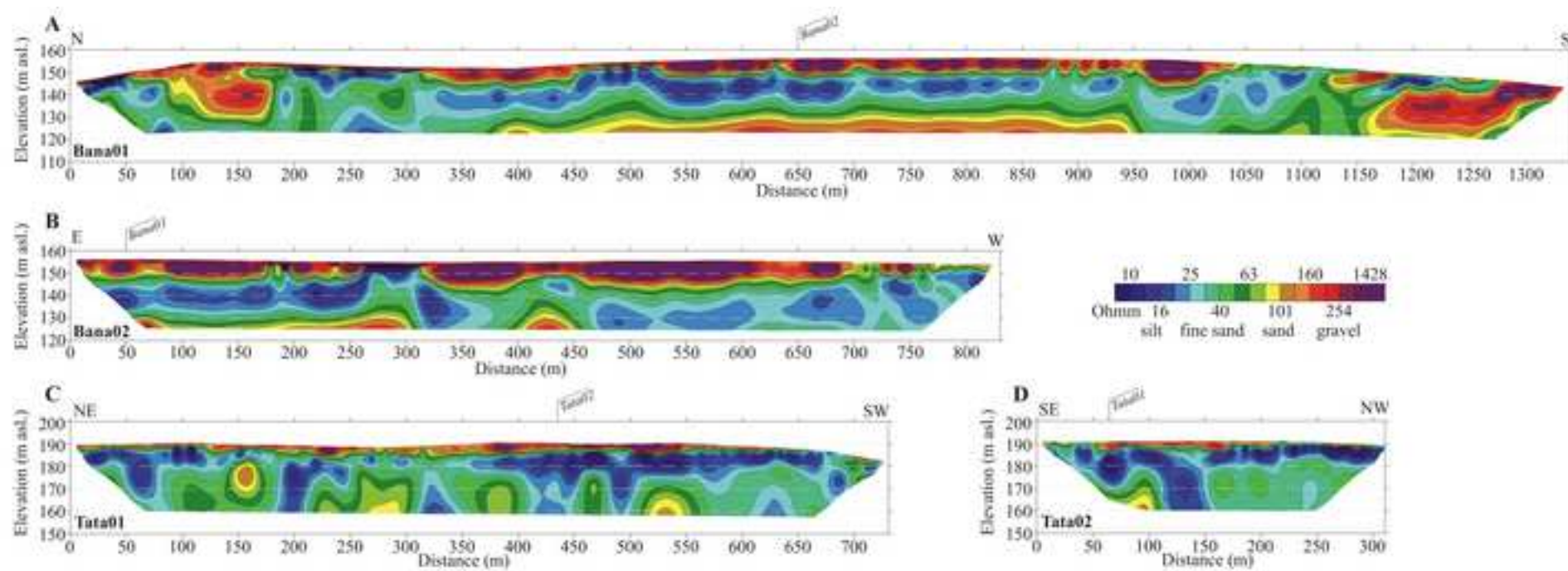
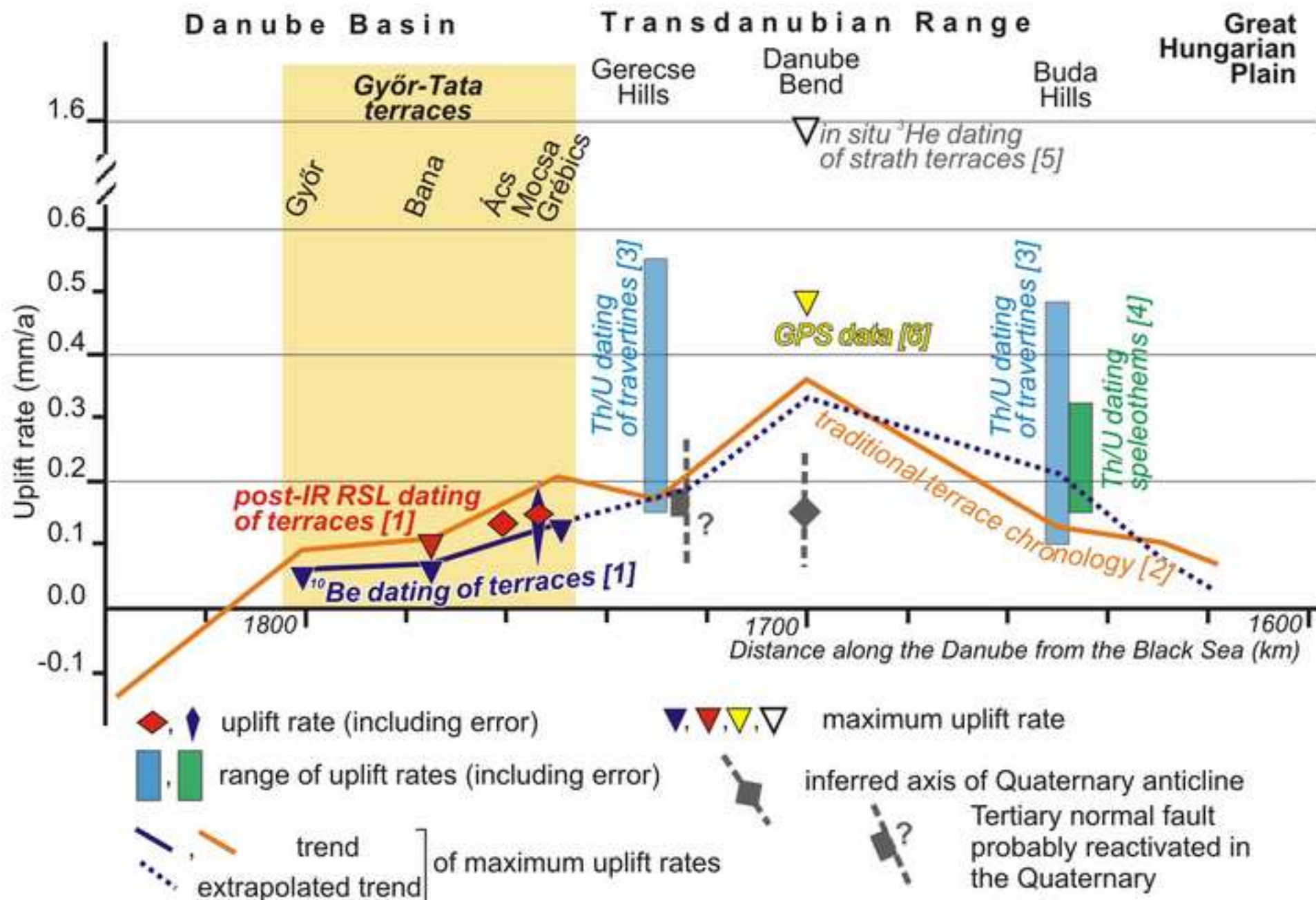


Figure11
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