Deformation, volcanism and geomorphologic evolution in the south-western part of the Transdanubian Range, Pannonian Basin:

Inheritance of pre-orogenic Triassic normal faults in a Cretaceous fold and thrust belt and geomorphologic evolution of the Late Miocene - Pliocene volcanic field in the Bakony-Balaton Highland

Pre-conference Excursion Guide for the 14th ILP Task Force Meeting

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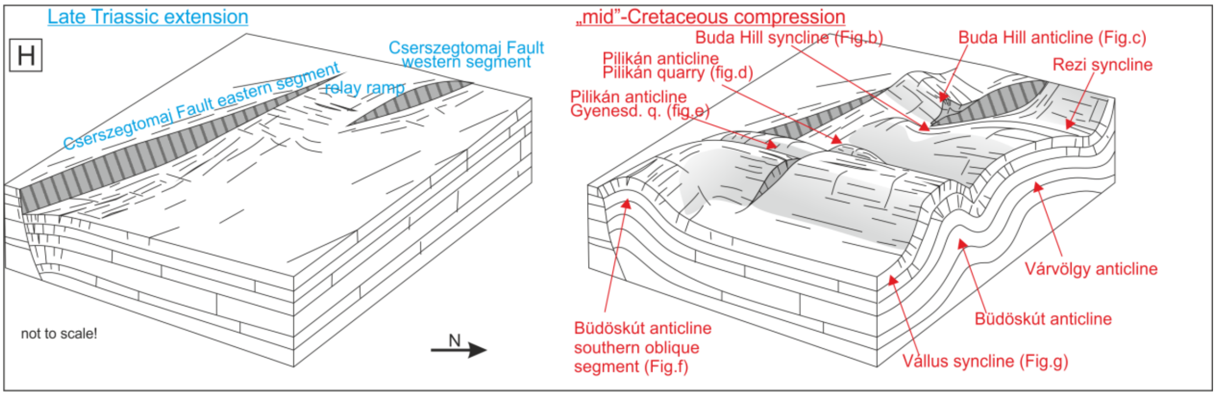

Preamble - Introduction

The pre-conference field trip will focus on three topics: 1) the sedimentological characteristics of Late Triassic pre-orogenic basins and the analysis of Triassic syn-sedimentary faults during the Cretaceous orogeny; 2) morphology and structures of the Miocene deformation of this segment of the Pannonian basin; 3) geomorphology and surface evolution of a Pliocene volcanic field of the Balaton Highland with inferences for neotectonics.

Transdanubia was a province of the Roman Empire located west from the Danube (Danubius) river which was at that time its eastern boundary. The Transdanubian Range (TR) comprises hills and low mountains forming a morphological ridge of SW–NE strike, from the western part of Hungary up to Budapest. This range is composed of Mesozoic and Paleozoic rocks surrounded by Palaeogene to Miocene sediments. The geographical TR is surrounded in the west by the Danube Basin, and in the SW by the Zala basin. The highest of these hills are the Bakony Mts. which passes south-eastward into the "Balaton Highland area" (BH), a geographically and geologically separate region, although the boundary between the two is not precisely defined. At the southernmost outcropping end of the TR, the Keszthely Hills form a separate fault-bounded horst. The excursion will have stops in the Keszthely Hills and in the Balaton Highland.

Fodor et al. (2017) described 13 phases of deformation which affected the TR from the Triassic up to the present. During the excursion, we will deal with only few of them, namely the Triassic extension, the major Cretaceous folding and thrusting, the formation of the Pannonian basin and its neotectonic inversion.

Before the initiation of this Cretaceous orogeny the TRU was situated on the passive margin of the Adriatic plate, which was bounded by two oceanic realms; the Triassic-Jurassic Neotethys and the Jurassic Alpine Tethys. In consequence of this paleogeographic position, the TRU was affected both by the Permian–mid-Triassic and Late Triassic–Jurassic rifting phases. The first one resulted in differentiation of the Early Triassic ramp and early Anisian platform (Fig. 4) (Budai & Vörös 1992, 2003, 2006, Kovács et al. 2011). The deriving basins contained cherty limestone, marlstone, which passes to tuffitic sediments in the late Anisian and Ladinian (Vörös et al. 1997). The basins were coeval with platforms (Budai et al. 1993), which, time to time, prograded into the mixed siliciclastic to carbonate basins (Budai et al. 2001).

In the late Carnian, shallow platform established through the area representing the passive margin of the Neotethys. Renewed faulting occurred in the Norian when isolated small basins developed which hosted organic-rich sediments (cherty dolomites, limestones, marls: Rezi and Kössen Fm.) (Csillag et al. 1995). The depositional environment became deeper in the Early Jurassic which is also marked by reduced sequences. New normal faulting occurred in this time span, and continued during the mid-Jurassic (Galácz 1988). These extensional phases are connected to the opening of Piemont-Ligurian Ocean further to the west.

In this guidebook, we use the term Transdanubian Range Unit (TRU) which is larger than the geographical term Transdanubian Range because it extends below the Neogene. The tectonic position of the TRU was long time debated, mostly due to the fact that its boundaries are always covered by Quaternary and Miocene sediments (Dank & Fülöp 1990). However, a consensus has been reached during the last 20 years (see Horváth 1993; Tari & Horváth 2010). The Transdanubian Range Unit (TRU) is the uppermost thick-skinned nappe of the Austroalpine nappe system (Schmid et al. 2008), which formed during a mid-Cretaceous orogeny (Fig. 1); thus the TRU is a Cretaceous tectonic element. Its nappe position is shown by the cross-section of the area (Fig. 3), by the existence of tectonic boundary below the unit (particularly at the SW side) and the presence of Cretaceous metamorphism below the unit.

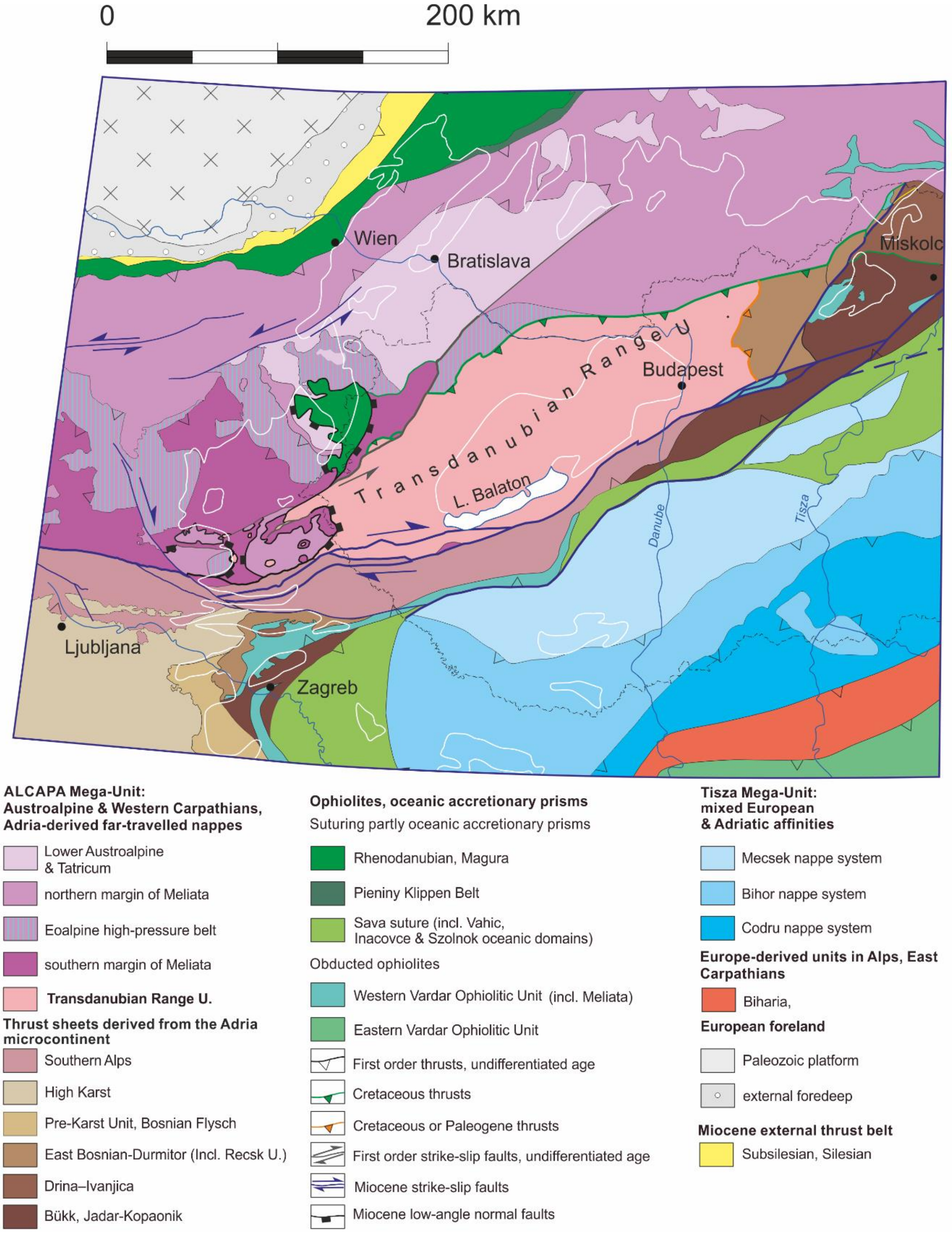

Fig. 1. Regional tectonic frame of the Eastern Alps, Carpathians, Dinarides and the Pannonian basin, after Schmid et al. (2008), modified slightly by Fodor et al. (2017).

The internal structures of the TRU comprises folds and minor thrusts that can be mapped on surface (Budai et al. 1999) and in subsurface (Tari 1994, Dank et al. 1990, Tari & Horváth 2010). The general strike of contractional structures is NE–SW (Fig. 2). On the other hand, in the south-western outcropping part of the TR, the Keszthely Hills show a systematic change in shortening direction (Fig. 2). Both outcrops-scale fold, faults, and map-scale synclines point to E–W shortening (Budai et al. 1999, Héja 2015). Héja (2019) followed folds and reverse faults more to the SW, below Senonian and Cenozoic sediments. The high-quality seismic data point to N–S striking structures, parallel to that on the surface. The detailed observations strengthen the early recognition of Tari (1994) of these structures.

The TRU is bounded by Cretaceous thrusts in the north-west and north; they are only partly reactivated by Miocene normal faulting. The underlying units belong to different Austroalpine nappes, (the highest one is the Graz Paleozoic, correlation of Balázs 1969) and the deeper Penninic units of the Rechnitz windows.

In the south-east, the Balaton Fault is part of the Mid-Hungarian shear zone, a wide belt of strike-slip and transpressional deformation of Oligocene to Early Miocene age (ca. 31–19 Ma) (Fig.1,2) (Balla 1988, Balla & Dudko 1989, Balla et al. 1987, Csontos & Nagymarosy 1998, Fodor et al. 1998). Despite its importance in the structure of Hungary, we will not deal with this zone in detail, simply because it is not outcropping in the TRU.

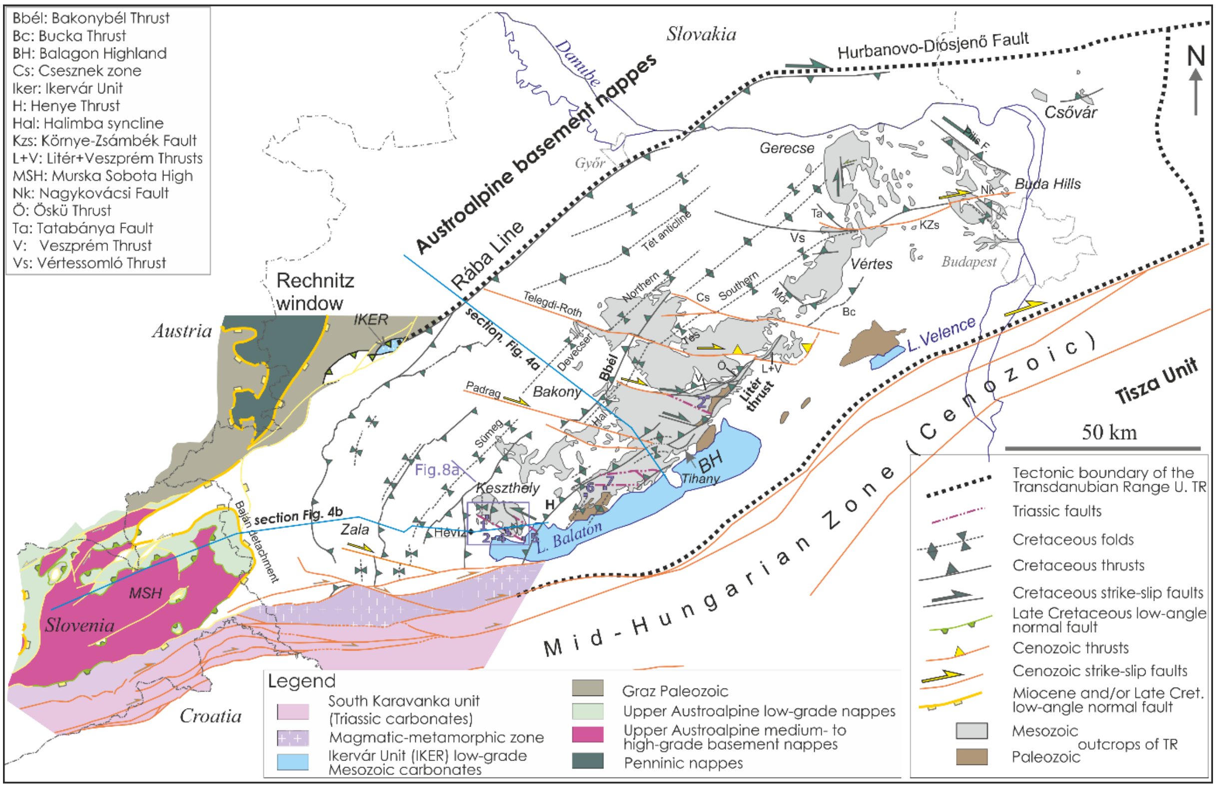

Fig. 2. Simplified structural map of the Transdanubian Range unit, after Fodor et al. 2017.

In the SW, the boundary of the TRU is quite wavy, already shown by Haas et al. (2000) and refined by Fodor et al. (2003b, 2013, 2017) (Fig. 2, 5). This pattern is due to the antiformal shape of the Murska Sobota high, composed of mesograde rocks of the Austroalpine nappe pile (Lelkes-Felvári et al. 2002, Gosar 1995). This unit is overlain by a thick low-angle mylonitic zone of late Cretaceous and Miocene age, termed here as Baján detachment (see also Nyíri et al. 2019). Small slivers of low-grade Paleozoic rocks and also of non-metamorphosed Permo-Triassic sediments occur sporadically on the top of the mesograde rocks both on the surface and in the subsurface; on the Murska Sobota High. They were interpreted as extensional allochtons. Because of a sharp jump in metamorphic degree across the detachment and exhumation of footwall metamorphics, the Murska Sobota High is considered as metamorphic core complex (Fodor et al. 2003a,b, 2013).

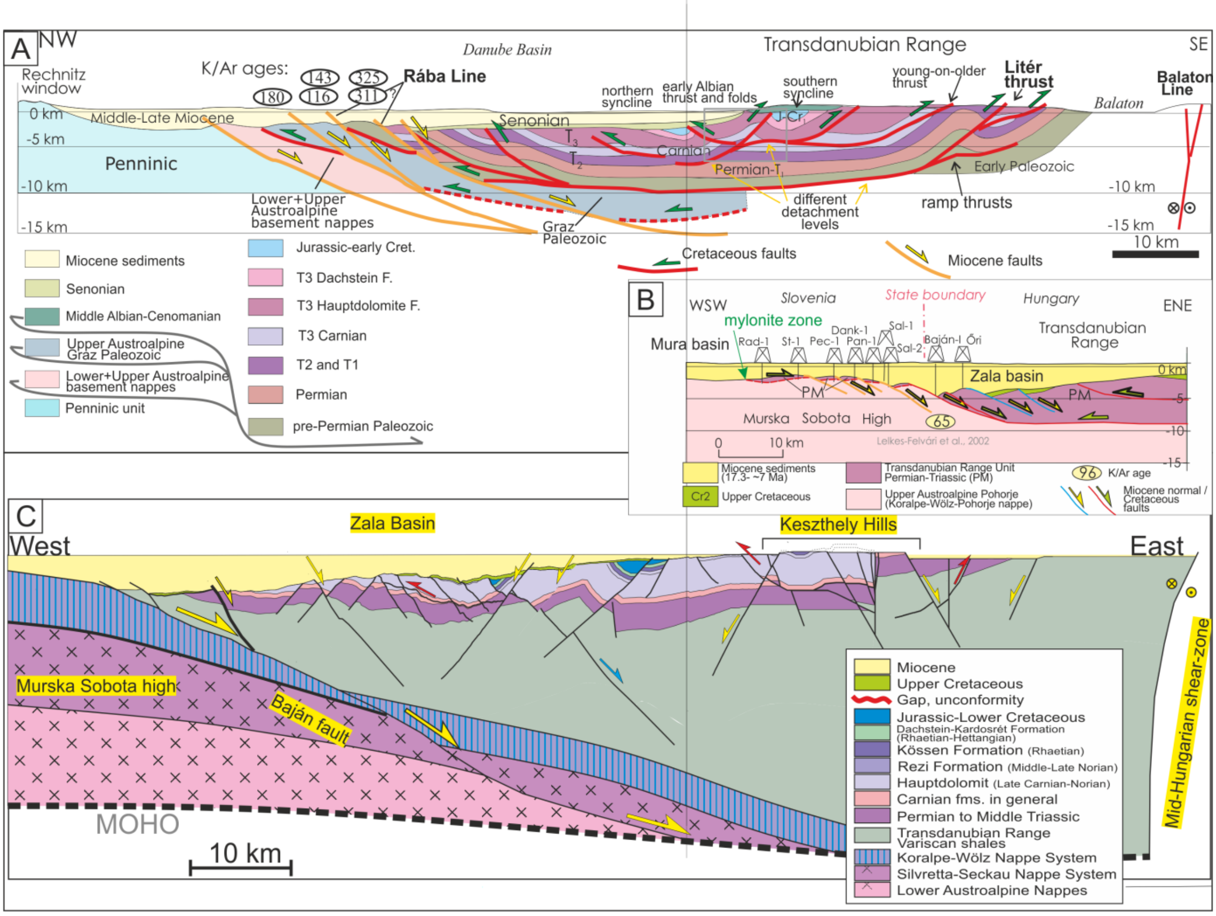

Fig. 3. Cross-sections of the Transdanubian Range unit, and its relationship toward other Alpine units (Fodor et al. 2017, Héja 2019). A) section across the Bakony (central TRU) between the Balaton-highland and the Rehnitz window B-C) sections across the Zala Basin (SW TRU) between the Murska Sobota high (Slovenia) and the Keszthely Hills.

The Baján detachment is already part of the syn-rift structure of the Pannonian basin (Fodor et al. 2013b, Nyíri et al. 2019). In its hanging wall a great number of normal faults are present (Fig. 4). The basin between the Keszthely Hills and the Pohorje Mts. in Slovenia has the common name Zala-Mura basin, with several sub-basins mentioned on fig. 5. Near the field trip area, N-S trending faults bound the Keszthely Hills, and they are connected to NW–SE trending grabens from the Tapolca graben up to the Vasvár graben. The normal faults are connected to NE-SW trending strike-slip faults which played the role of transfer faults between different extensional domains. One of such faults is the Rába fault on the NW and the Periadriatic-Balaton fault zone to the SE from the TRU.

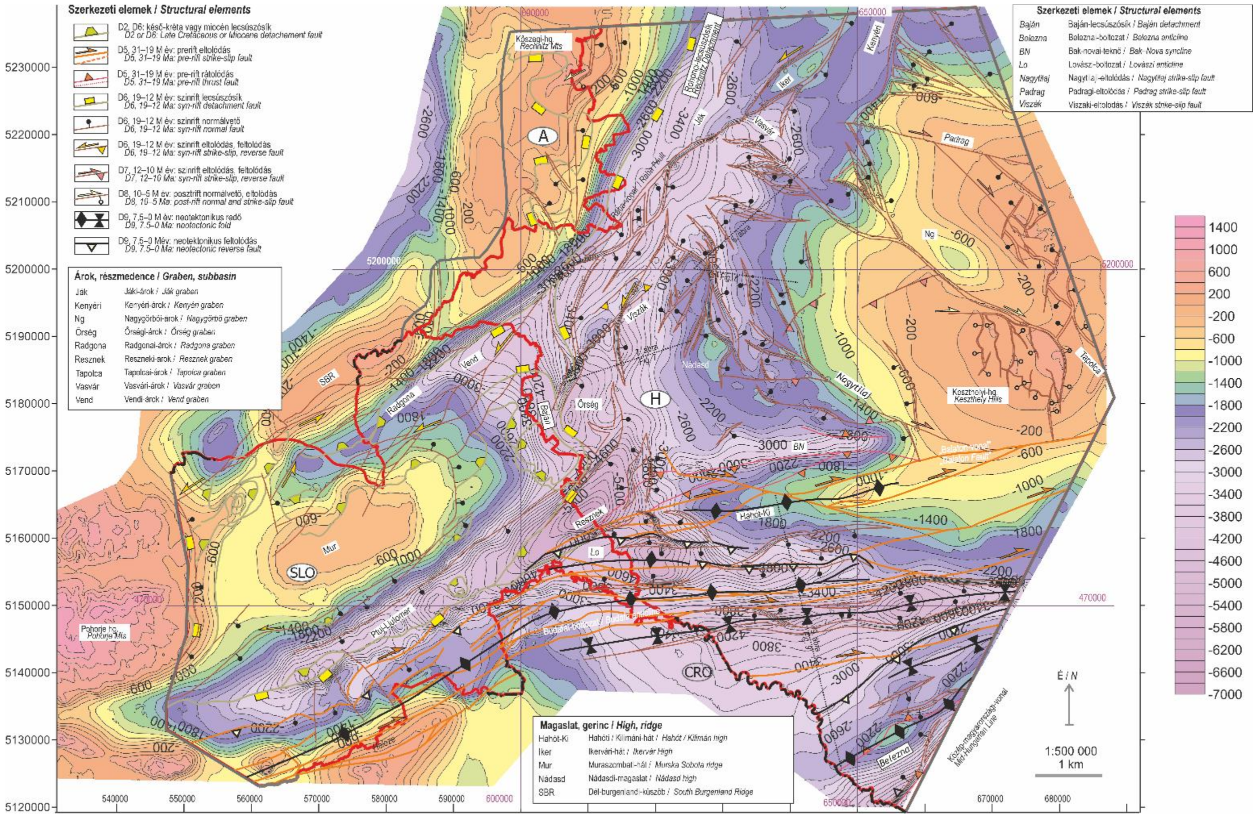

Fig. 4. Top-Mesozoic surface of the SW Pannonian basin showing mostly Miocene normal faults, and neotectonic folds, Oligocene–early Miocene Periadriatic–Balaton fault zone in the SE (Fodor et al. 2013b).

The boundary of syn-rift to post-rift deformation was debated for a long time. The difficulty lies in the fact that fault activity changes in time from place to place, even within the SW Pannonian basin. Tari (1994) correctly recognized a major discordance surface in the Zala-Mura basin, which seals most of the normal faulting. On the other hand, we will try to convince you that faulting was important in the Keszthely Hills up to the early Late Miocene, Here the termination of faulting should be placed in the early Pannonian, ca. 9 Ma.

Latest Miocene to Pliocene alkali basalt volcanism was often connected to a specific deformation phase. However, no convincing evidence was demonstrated about their structural settings. Their distribution may reflect processes in the mantle lithosphere and may not be connected to crustal deformation. On the hand, their topographic distribution can be used for quantification of neotectonic uplift (Németh et al. 1999, Sebe & Csillag 2012, Fodor et al. 2014).

The neotectonic phase of the Pannonian basin was marked by basin inversion, uplift, and denudation. This uplift was a regional phenomenon, encompassing almost the entire Transdanubia, including areas with thin or thick syn-rift to post-rift sequences. Differential uplift should characterize the range, while the Upper Miocene units are dipping away from the elevated range.

We have variable datasets characterising the uplift rate of the TR. First, the reconstruction of volcanic landforms suggests considerable denudation after volcanism. This denudation in the order of several hundreds of metres led to the exhumation of former maars, and deep roots of volcanic necks. The relatively soft Late Miocene sediments were eroded from below and between the volcanic edifices by variable forms of denudation (wind, rivers, sheet wash). Younger volcanoes occupy relatively lower topographic position, which may point to gradual uplift of the entire range, and parallel lowering of the landscape during the late Pliocene and Quaternary.

A simple estimation of the uplift rate takes into account the present height of volcanic remnants, their postulated erosion, and their age (Sebe & Csillag 2012, Fodor et al. 2014). However, the incision rate of former denudation surfaces marked by volcanoes, maars could be very slow, in the order of 0.1 mm/a down to 0.04 mm/a (100-40 m/Ma) (Fig. 6). This is in good agreement with the uplift rates derived from terrace chronological data relevant for the last ~ 3 Ma in the northern part of the TR (Gerecse Hills, Ruszkiczay-Rüdiger, et al., 2018a,b) A series of typical eolian landforms on macro- and meso-scales, like yardangs, ventifacts, wind-polished rock surfaces suggest that wind erosion was the most important way of surface denudation. In situ-produced cosmogenic ^{10}Be exposure ages of wind-abraded landforms suggest considerable wind erosion from at least 1.56 Ma (Ruszkiczay-Rüdiger et al. 2011, see stops 6 and 7).

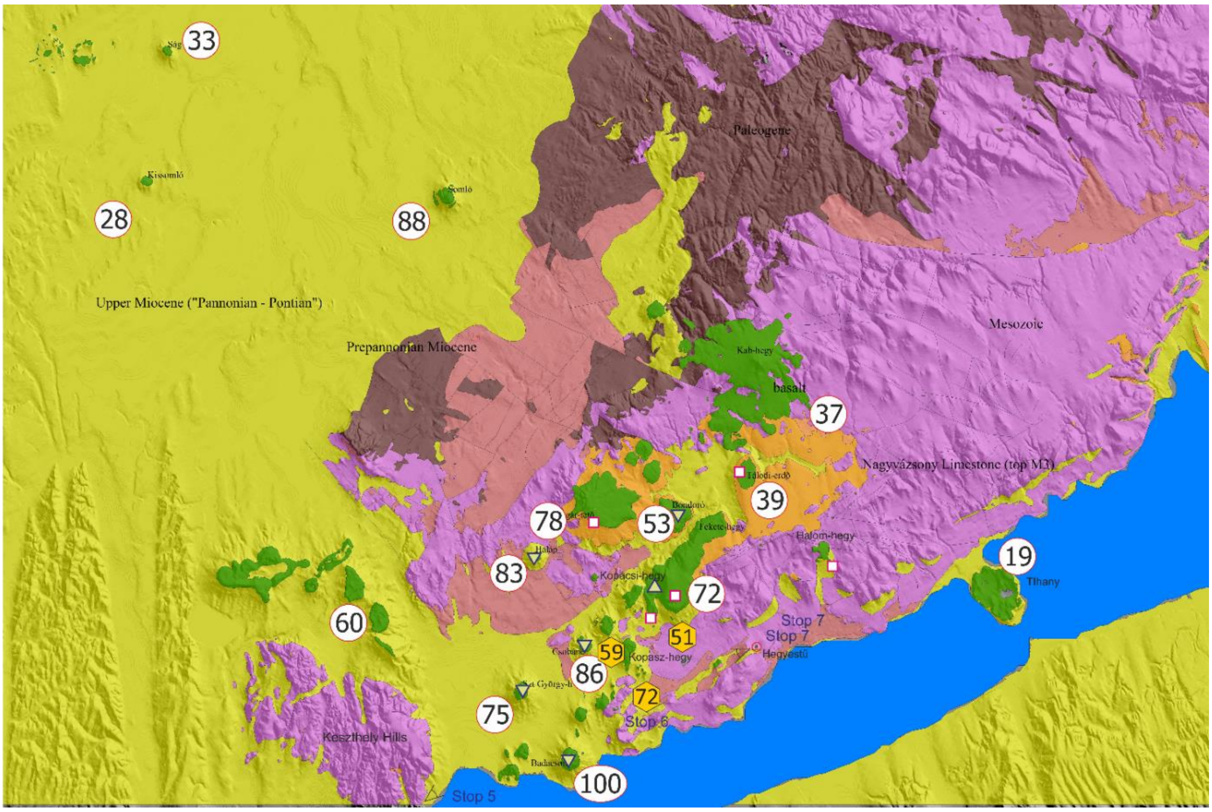

Fig. 5. Incision (lowering) rates of the southern TR based on diverse markers (after Fodor et al. 2014). The values can be taken as approximates of the uplift rate.

The uplift of the TR was long time connected to horizontal shortening and neotectonic inversion tectonics (Horváth 1995). In fact, GPS data indicate 1.3mm/a shortening across the entire Transdanubia (Grenerczy 2005, Bus et al. 2009) and this could be projected back in time. There are few upper crustal structures which could accommodate neotectonic shortening and uplift. The best-known elements are E–W trending folds SW and SE from the Keszthely Hills and Lake Balaton (Fig. 2, 4) (Dank 1962, Horváth & Rumpler 1984, Bada et al. 2007, 2010, Sacchi et al. 1999, Soós et al. 2019). Below the lake, several shallow seismic data acquisition campaigns were organised by Frank Horváth during the past 20 years. The data clearly demonstrate that the Late Miocene sediments were gently folded by ca. N–S shortening and a long, NE-trending sinistral fault zone developed below the Holocene mud of the lake (Bada et al. 2010, Visnovitz et al. 2015). It is possible that the major faults of the Middle Hungarian Shear Zone have been reactivated (Visnovitz et al. 2015, Csontos et al. 2005).

On the other hand, neotectonic structures were only sporadically observed within the TR (Fodor et al. 2005, Bada et al. 2007), or at least, neotectonic reactivation of Miocene or older elements were not unambiguously demonstrated (Ruszkiczay-Rüdiger et al. 2018a, b). This can lead to the conclusion that the neotectonic uplift of the TR cannot be the unique consequence of upper crustal brittle deformation, in other words, models relying of isostasy cannot explain the uplift pattern of the area. In fact, elastic bending of the entire lithosphere, i.e., lithospheric scale folding, was already postulated by Cloething and Horváth (1996). In a similar but somewhat new concept (Balázs et al. 2017), the joint effects of convective support of the mantle, elastic bending of the crust and possible ductile lower crustal deformation can explain the observed large-scale differential vertical movements and the redistribution of surface sediments (Ruszkiczay-Rüdiger et al. 2019) – see discussion at stops 6 and 7.

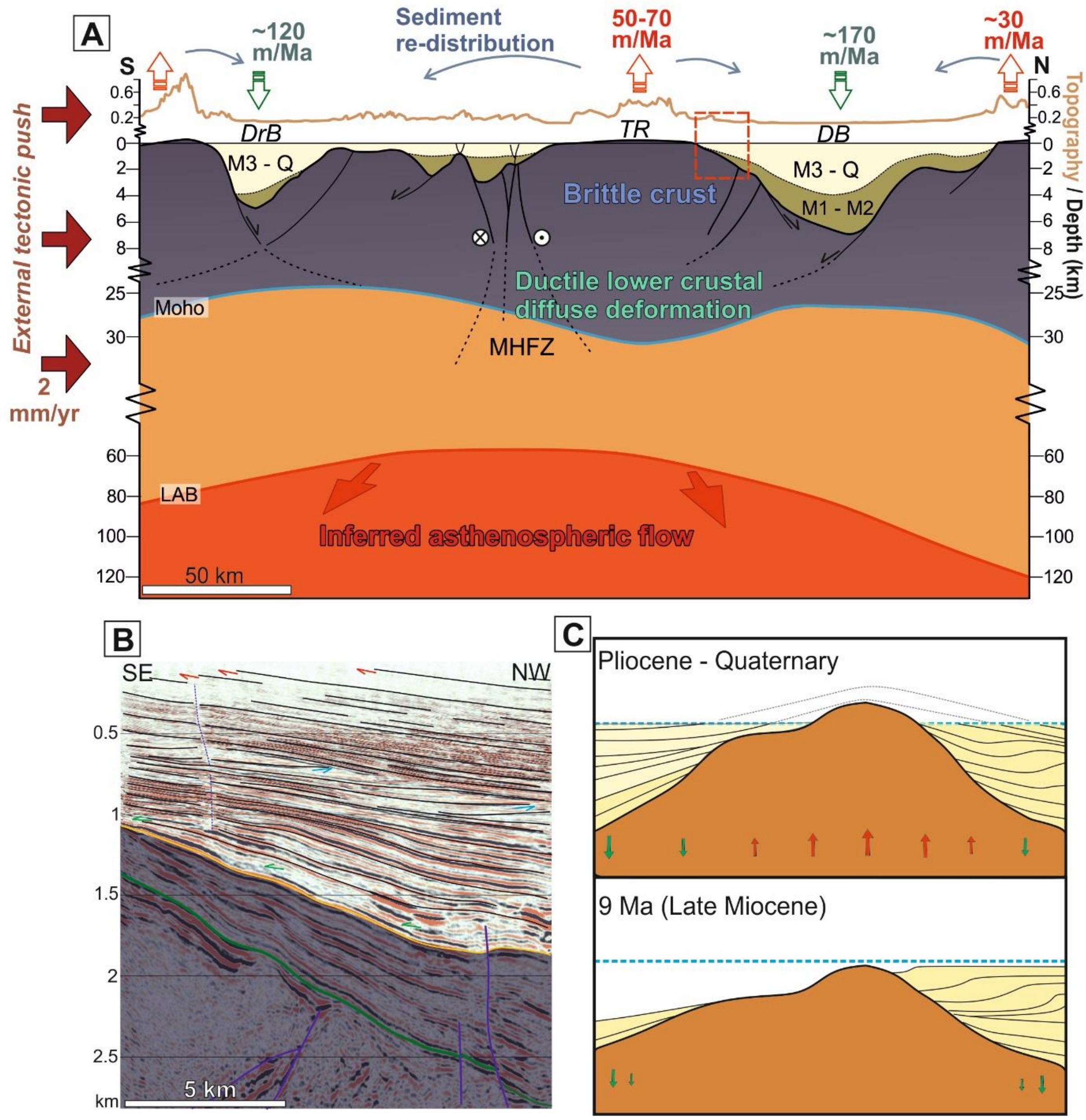

Figure 6. A: Generalized lithospheric-scale cross-section through the Pannonian Basin. B: Depth converted seismic section from the northern periphery of the Transdanubian Range. Note the tilted and eroded Late Miocene sediments. C: Simplified section showing the differential vertical movements in the vicinity of the TR (Ruszkiczay-Rüdiger et al. submitted).

Topic 1 – Cretaceous inversion of a Late Triassic graben-system

Stop 1: Csókakő (Daw Cliff) quarry (Cserszegtomaj) – talus breccia of a Late Triassic syn-sedimentary normal fault

Late Triassic stratigraphy of the Keszthely Hills

The Keszthely Hills are made up of dominantly Upper Triassic carbonates. From the end of the Carnian the depositional environment became uniform and the more-than-two km thick Hauptdolomit Formation was deposited on a huge carbonate platform (Fig. 7). The formation is built up by well-bedded bituminous dolomite in the study area. Occasionally microbialite intercalations occur. The formation was deposited in ultra-back-reef lagoon environment (Fruth and Scherreiks 1984). From the end of Middle Norian (Budai and Kovács 1986) intraplatform basins opened, and some of them persisted in the Rhaetian. These intraplatform basins were filled up by the late Middle to Upper Norian Rezi Dolomite and the latest Norian–Rhaetian Kössen Marl (Budai & Koloszár 1987; Budai & Kovács 1986; Budai et al. 1999a; Csillag et al. 1995; Haas 1993, 2002). The Rezi Dolomite is composed of the alternation of thin-layered bituminous, cherty dolomite and thick-layered porous dolomites, which often contain redeposited green algae fragments. A special dolomite breccia lithofacies of Rezi Dolomite is exposed in the Csókakő quarry, which is in the focus of the first stop.

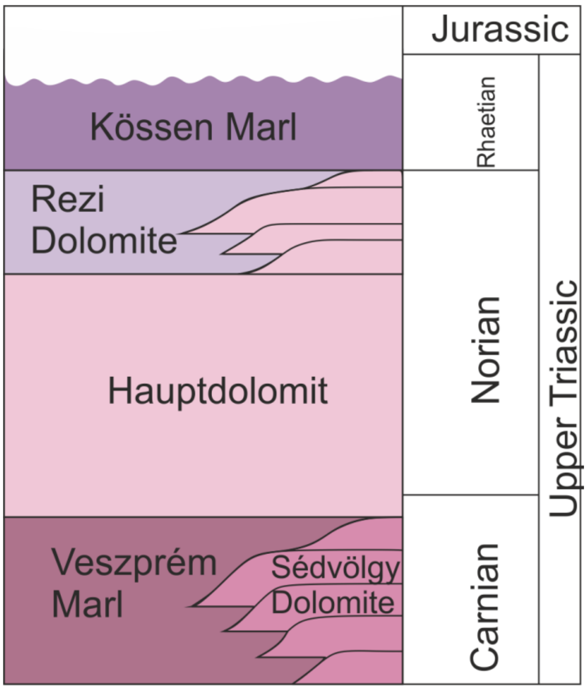

Fig. 7. Simplified Late Triassic stratigraphy of the Keszthely Hills.

Description of the eastern wall of the Csókakő quarry

In the Csókakő quarry (Fig. 8a), the Upper Norian Rezi Dolomite is exposed. The most spectacular part of the quarry is its eastern wall (Fig. 8b). Two main facies types of the Rezi Dolomite are visible here: thinly layered to laminated dolomite and dolomite breccia in a dolomitic matrix. Laminated dolomite occurs in the northern part of the eastern wall with sub-horizontal dip. Southward-thickening dolomite breccia tongues can be observed between the laminated bituminous dolomite layers, further to the south. These strata dip already moderately toward NNE. Further south, the dark grey laminated dolomite intercalations pinch out, and in the southern edge of the quarry only massive dolomite breccia is present (Fig. 8b).

The clasts of the dolomite breccia are up to a few meters in size (Fig. 8b). They are thick-bedded or massive, white or light grey boulders, which often contain green algae, mollusks and gastropods. Stromatolitic, intertidal dolomite represents another clast-type of this dolomite breccia.

The matrix of the dolomite breccia is gradually changing southward. In the northern, distal part of the breccia-tongues, the matrix of this breccia is laminated dolomite. In contrast, in the southern part of the breccia tongues the matrix is made up of massive, light-grey dolomite, containing the same platform-derived fossils as the fossil-rich clasts of the dolomite breccia itself.

Slump folds and other types of soft-sediment structures are frequent in the thin-layered, laminated unit (Fig. 8c, e). These folds have steep or overturned limbs and variable fold hinge directions (Fig. 8f). Weak lineation may occur on the folded bedding planes (Fig. 8d).

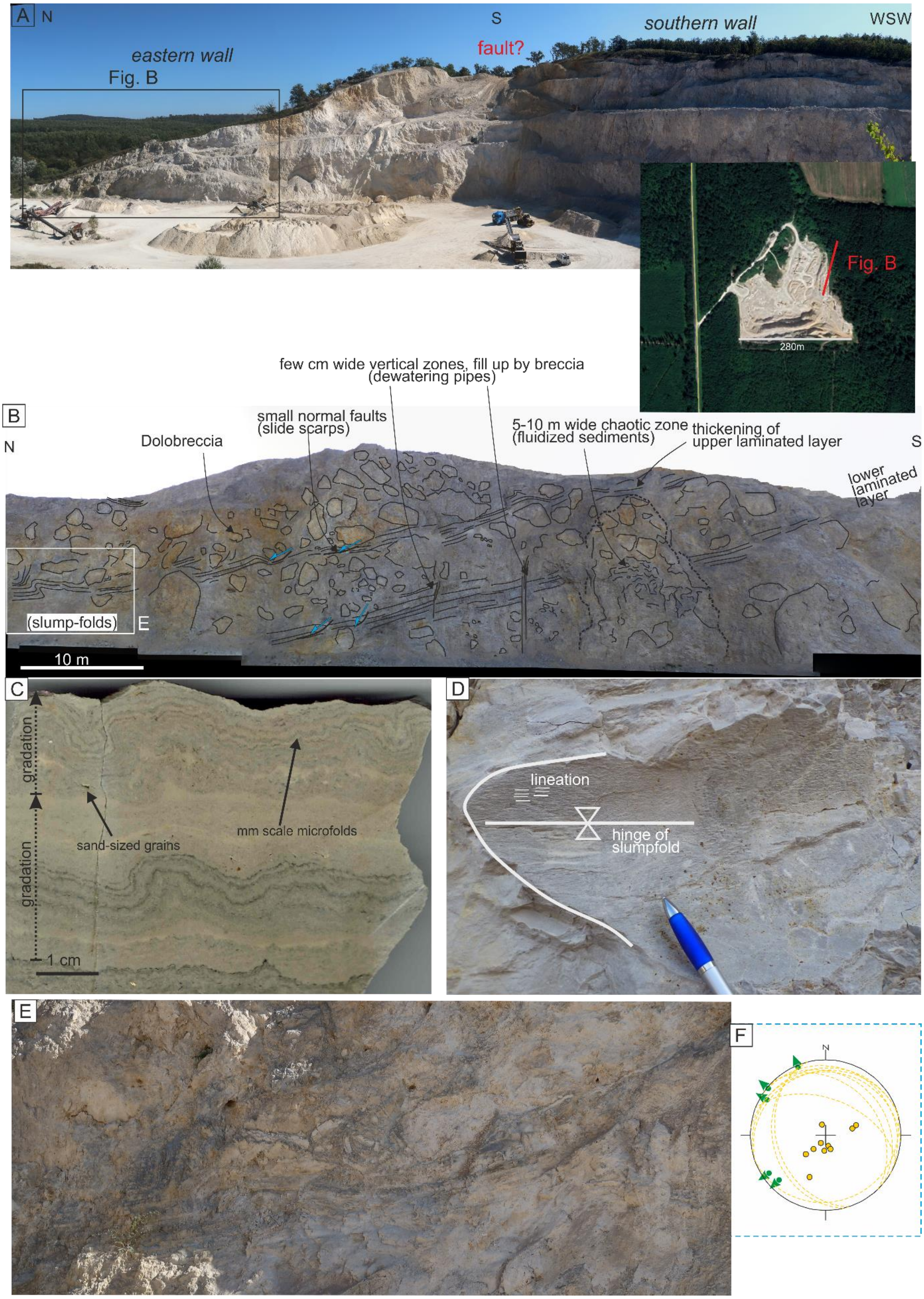

Fig. 8. Late Triassic fault-controlled sedimentation in the Csókakő quarry. A) General view, B) Section of the eastern wall. C) Small-scale folds of the laminated facies, D) Lineations (microfolds?) on sub-vertical bedding plane. E) Slump folds, disrupted laminated layers, embedded dolomite boulders, all showing sediment deformation. F) Stereogram of fold hinges.

Interpretation of the Csókakő quarry

Map-scale syn-sedimentary normal faults can often be outlined based on facies distribution and the presence of coarse-grained sedimentary breccias in the proximal hanging-wall of the fault (e.g. Bertotti et al. 1993). Dolomite breccias of the Rezi Dolomite described in the outcrops of the Keszthely Hills have a dolomite matrix, which suggests that the formation of these breccias pre-dates early dolomitization, and they are likely to be sedimentary breccias. These breccias may have redeposited on a fault-controlled slope (Csillag et al. 1995). In the study area, dolomite breccia outcrops of the Rezi Formation are limited to a WNW-ESE striking belt along the southwestern edge of the Keszthely Hills.

South of the dolomite breccia occurrence of Csókakő quarry the Hauptdolomit is exposed. The WNW-ESE trending contact of the two formations was identified already by former mapping (Budai 1999b), however, it was interpreted as a stratigraphic contact. In our interpretation, this contact represents a Late Triassic syn-sedimentary normal fault, which is referred to as the Cserszegtomaj fault in this guidebook (Fig. 10).

Progradational tongue of dolomitized platform carbonates above the Rezi Dolomite was documented in the eastern part of the Keszthely Hills by Csillag et al. (1995). This pattern, fault-bounded talus breccia on the southwest and prograding platform carbonate on the northeast rather suggests an asymmetric half-graben geometry for the Late Norian basin of the Keszthely Hill (Fig. 9a). Note, that the southward widening outline of the N-trending Rezi syncline also supports this asymmetric geometry, while it can be explained by Cretaceous E–W shortening of a NE-ward shallowing and thinning Late Triassic half-graben (Fig. 9c).

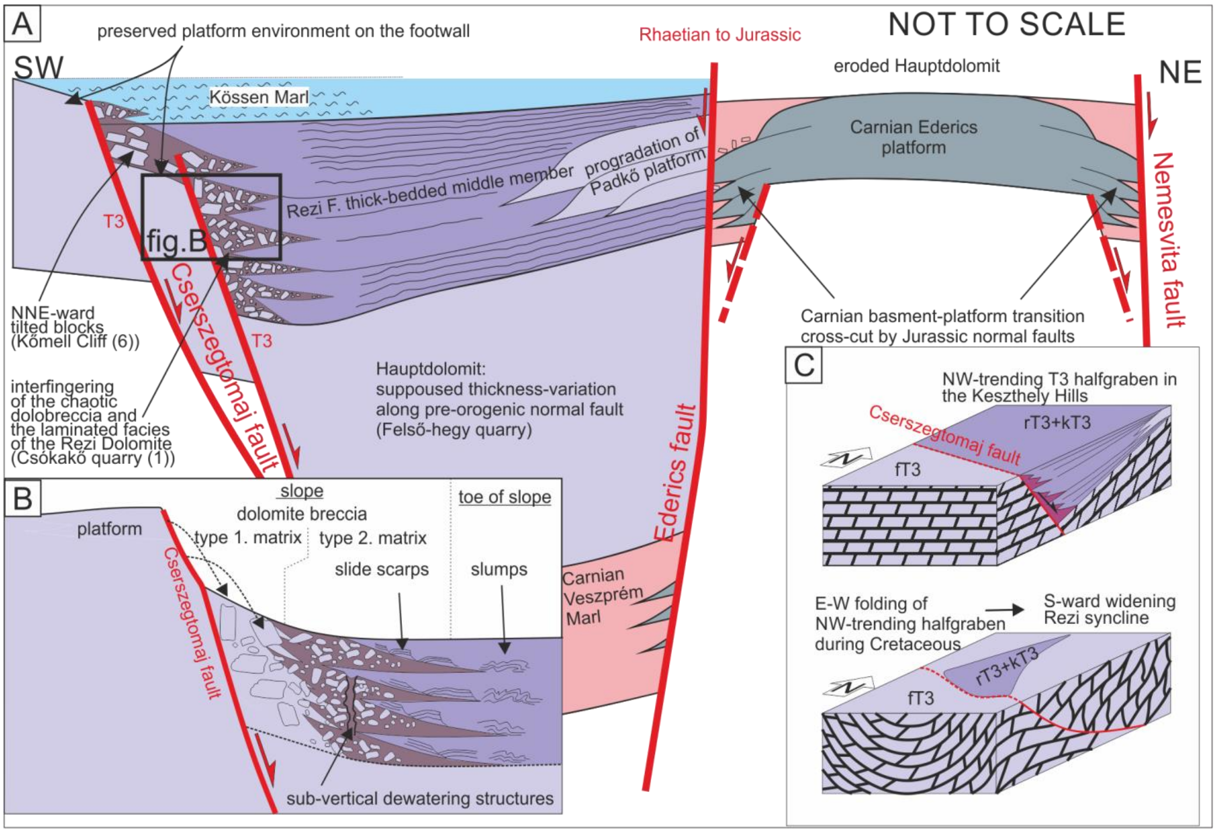

Fig. 9. A) Simplified cross-section of the Late Triassic basin of the Keszthely Hills. B) Interpretation of the talus breccia exposed in the Csókakő quarry. C) Interaction of Triassic half-graben and Cretaceous folding.

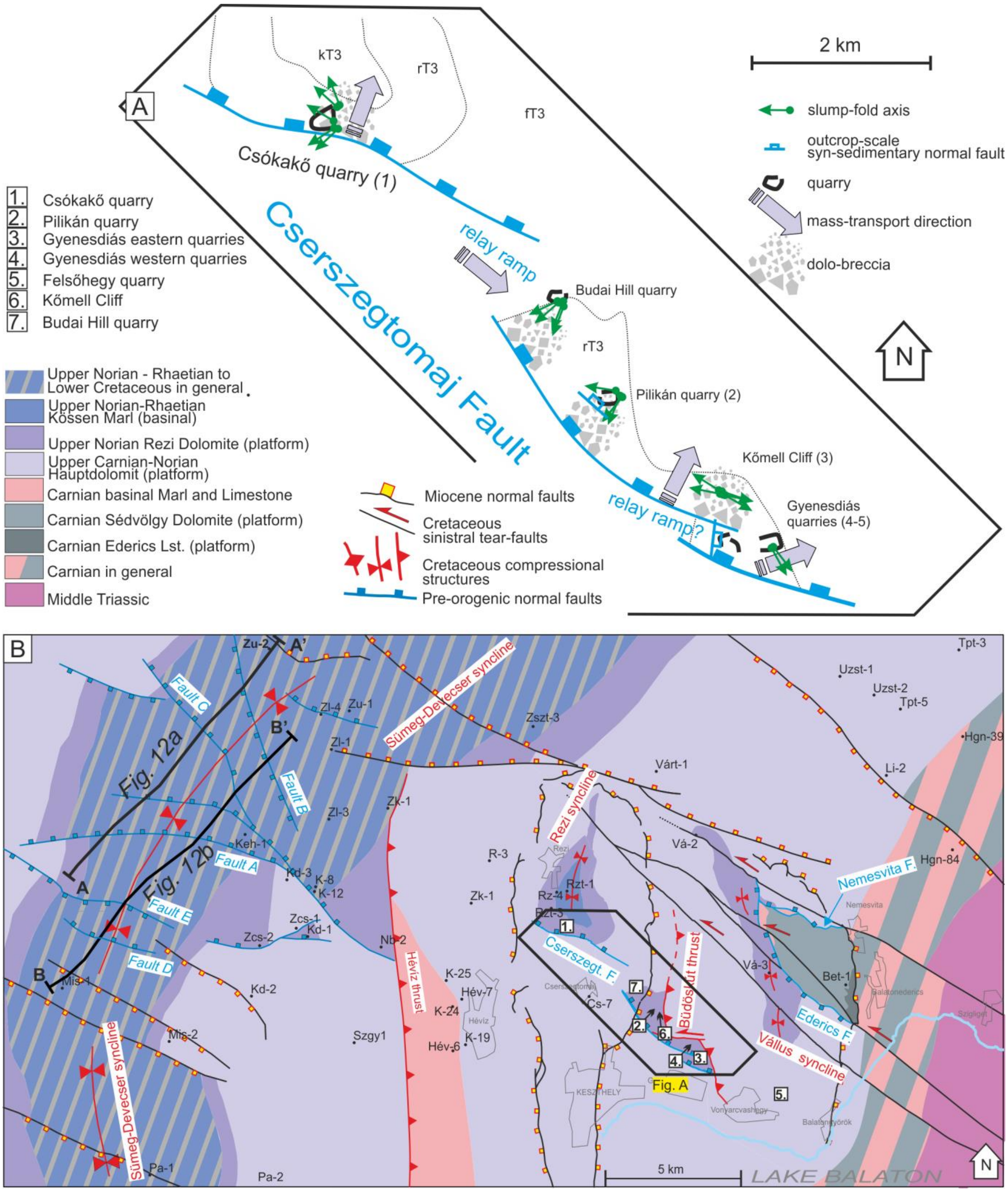

Fig. 10. A) Late Triassic syn-sedimentary structures along the Cserszegtomaj Fault. B) Pre-Upper Cretaceous map of the Keszthely Hills and its neighbouring area, after Héja (2019).

ESE from the Csókakő quarry, the southern WNW-striking border of the Rezi Dolomite occurrences near Gyenesdiás represents further NNE-dipping segments of the Cserszegtomaj Fault (Fig. 10). In map view, the fault segments do not constitute a continuous fracture, but are dissected into two or three segments. The overlapping fault segments of 1-3 km width were connected by ESE dipping relay ramps (Fig. 10). WNW-ESE strike of the Cserszegtomaj fault is in accordance with small-scale syn-sedimentary normal faults, measured in several outcrops of the Keszthely Hills (Fig. 11).

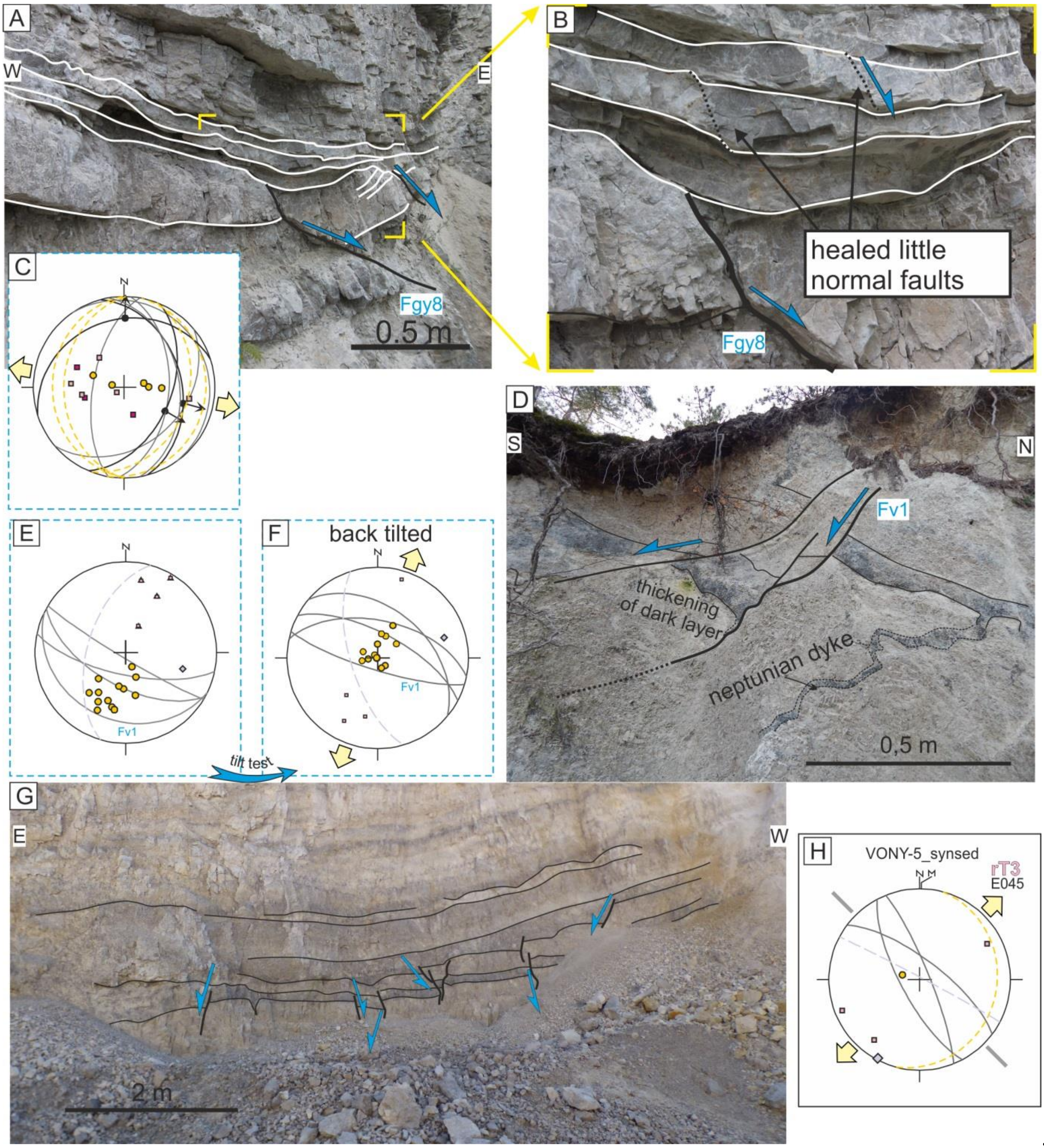

Fig. 11. Outcrop-scale examples of syn-sedimentary normal faults in the SW Keszthely Hills. A-B) Gyenesdiás, western quarry; Late Triassic syn-sedimentary normal fault or slide in the laminated to thin-layered Rezi Dolomite. C) The measured fault slip data D) Late Triassic syn-sedimentary structures in the Felsőhegy quarry, observed in Hauptdolomit. E-F) The measured fault slip-data. G) syn-sedimentary normal faults in the Pilikán quarry. H) The measured fault slip-data. For localities of the quarries see Fig. 10b

Syn-sedimentary normal faults with similar strikes were reported by Héja et al. (2018) based on seismic sections in the western foreland of the Keszthely Hills (Fig. 12). These normal faults were active during the deposition of Rhaetian Kössen Marl, and “Dachstein Limestone”; and the extension was probably continuous during the Jurassic as well.

The surface and subsurface grabens form a coherent system of Late Triassic normal faults (Fig. 12).

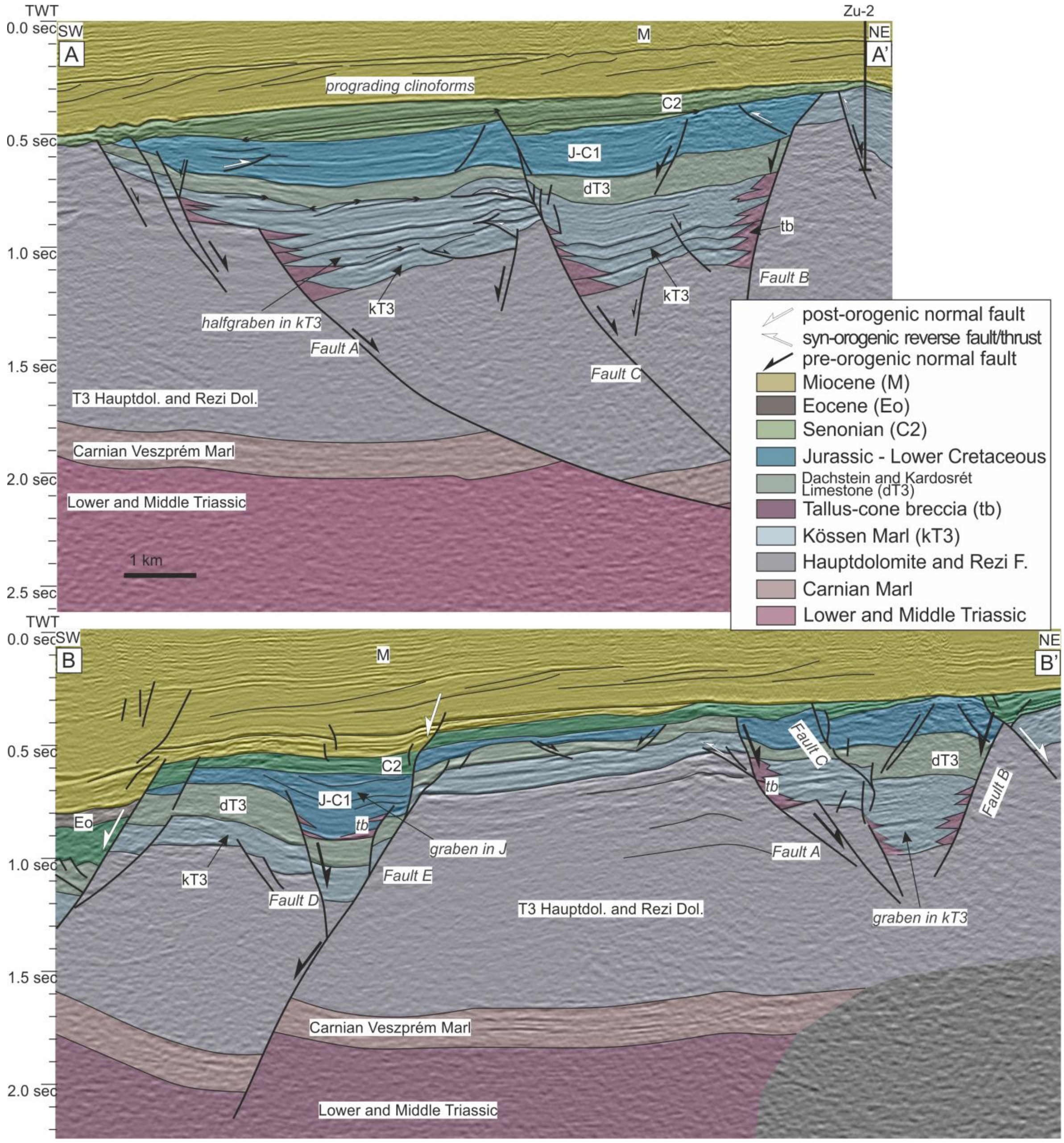

Fig. 12. Seismic examples of late Triassic to Jurassic normal faults in the Zala basin (west from the Keszthely Hills).

Paleogeographic implications

The introduced Late Triassic normal faults represent most probably the first sign of continental rifting related to the Alpine Tethys (Piemont-Ligurian Ocean). Similar extensional grabens with very similar stratigraphy are known from the Northern Calcareous Alps (Behrmann and Tanner 2006), and from the western Southern Alps (Lombardy, Bertotti et al. 1993).

On the basis of Late Triassic facies boundaries, several authors argue that the Transdanubian Range was located more to the west, between the Northern Calcareous Alps, the Drau Range and Southern Alps (Haas et al. 1995; Mandl 2000), and these Late Triassic extensional basins formed a continuous graben system, which is referred to as the Kössen Basin in this guide (Fig. 13a).

The Late Triassic Kössen basin was situated significantly to the east of the future Alpine Tethys, on the proximal Adriatic passive margin (Fig. 13). During Jurassic, the westward migration of extensional tectonism was pointed out in the case of Austroalpine nappes (Froitzheim and Manatschal 1996). The proximal Adriatic margin was subject to dominantly Hettangian-Sinemurian extension, whereas in the distal Adriatic margin Pliensbachian-Callovian extension occurred (Fig. 13.a). A similar situation was interpreted for the Southern Alps, west of Lombardy. In the Cusio-Biella-Canavese Zone extensional grabens formed just during the Early Jurassic (Decarlis et al. 2017). According to Froitzheim and Manatschal (1996) this westward migration of the locus of extension is due to the asymmetric rifting of the Alpine Tethys, where the Adriatic plate represented the lower plate, whereas the European plate was in an upper plate position (Fig. 13.b).

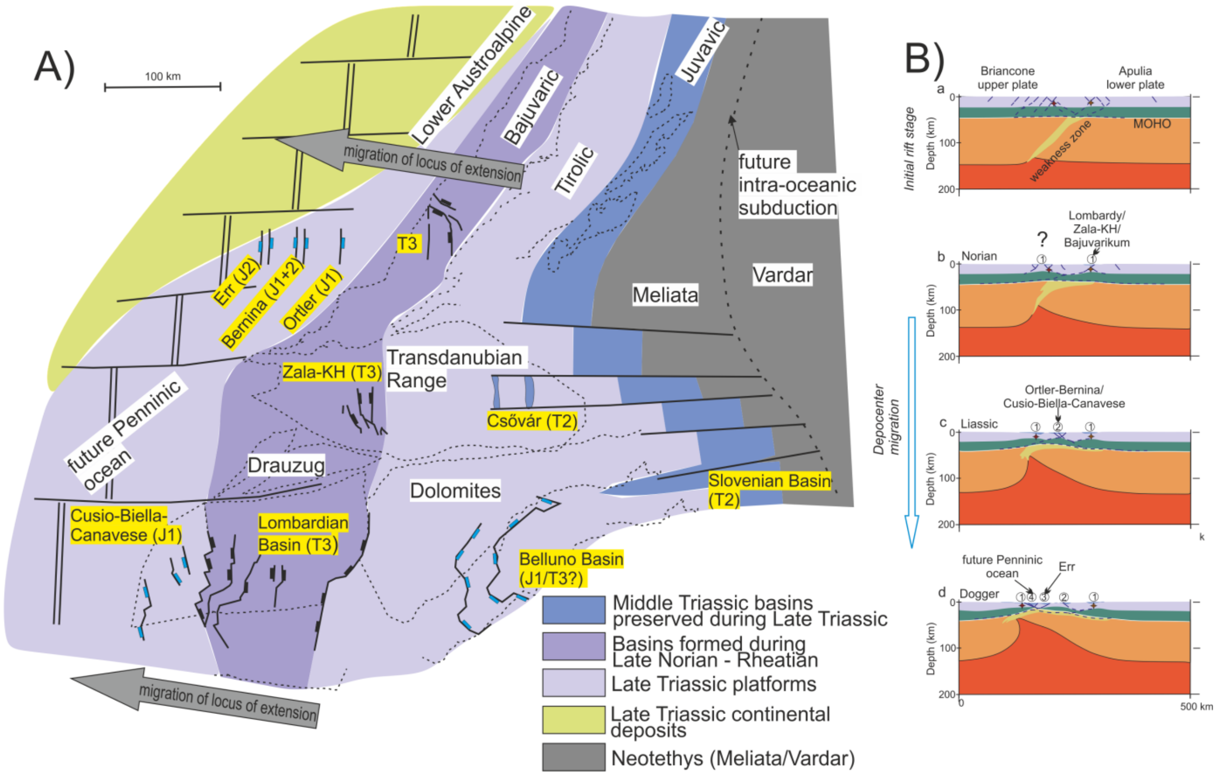

Fig. 13. A) Tentative Late Triassic paleogeographic reconstruction of the Transdanubian Range and neighbour units based on Haas et al. (1995), Froitzheim and Manatschal (1996), Mandl (2000), Héja et al. (2018) B) Migration of locus of extension on the lower Apulian plate during the opening of Piemont-Liguria ocean, applying the numerical model of Balázs et al. (2017).

Stop 2: Eastern pit of the Gyenesdiás quarries – on the limb of an obliquely trending Cretaceous anticline (17° 17' 50.7917" E — 46° 46' 43.9991" N)

Description of the eastern Gyenesdiás quarry – folds of the Keszthely Hills

Late Triassic succession is exposed in this quarry. In the western wall of the eastern pit relatively thick, the tilted succession of the Hauptdolomit is visible (Fig. 14.b). It is overlain by a dark-grey, laminated, thin-layered dolomite (Rezi Formation), which is interrupted by an intermediate thick-bedded dolomite intercalation (southern wall; Fig. 14.a). On the eastern wall of the pit beds are sub-horizontal again (Fig. 14.b). East to north-eastward dipping beds of the Gyenesdiás quarry represents the north-eastern limb of the NNW-SSE trending Gyenesdiás anticline (Fig.14.d).

At the easternmost part of the quarry, several N-S trending normal fault crosscut the Triassic layers, they are probably of Miocene age. In fact, they are parallel to a more prominent fault scarp seen from the belvedere (Stop 3). E-W extension could be calculated based on these faults, which fits well with several map-scale faults and outcrop-scale fault-slip data.

On the steeply tilted fold limb pre-tilt conjugate normal faults occur. These faults suggest NE-SW extension according to the back-tilted stereonet (Fig. 14.a), and they represent an older phase of pre-orogenic extension, possibly late Triassic or Jurassic. The contact of beds are frequently represented by powdered dolomite (Fig. 14.a); according to our interpretation this powdered zones can be considered as layer-parallel shear zones, which developed due to flexural slip folding.

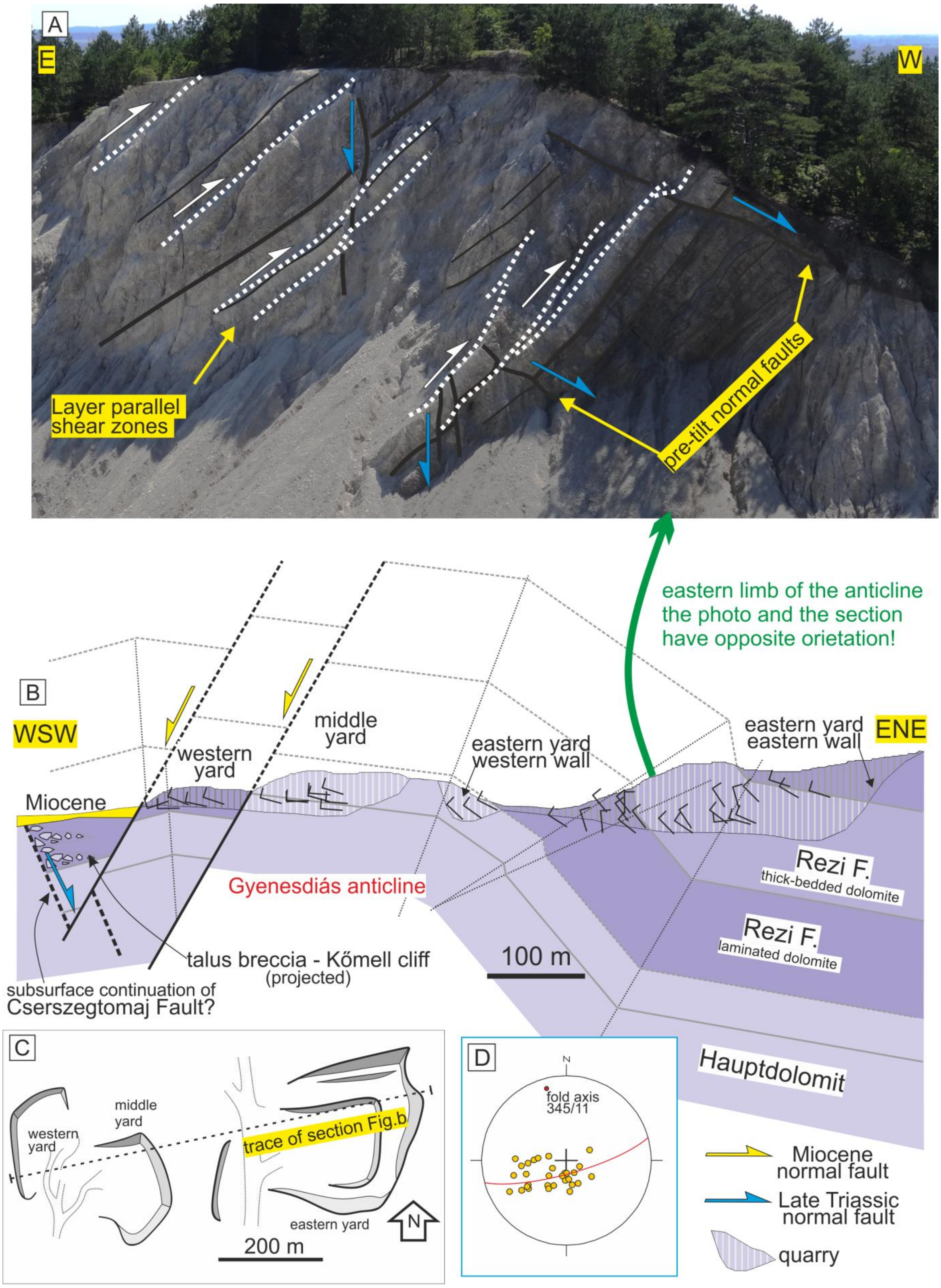

Fig. 14. A) The southern wall of the astern Gyenesdiás quarry, where pre-tilt normal faults, and the laminated Rezi Formation is exposed. B) Section across the Gyenesdiás quarries showing the NNW-SSE trending Gyenes anticline. C) simplified map of the Gyenesdiás quarries. D) Poles of bedding planes, and the best-fitting great circle.

Age of the folding

The time interval of folding is poorly constrained in the Keszthely Hills, since folded Upper Triassic rocks are discordantly sealed by Miocene deposits. However, based on projection of field and seismic data from the middle segment of the TRU it is evident, that Early Cretaceous (Aptian to earliest Early Albian) rocks are involved in folding, and contractional structures are discordantly sealed by sub-horizontal Upper Cretaceous (Santonian) beds. The time span of folding thus could be 113–84 Ma, with several episodes of deformation.

Interpretation: oblique folds due to structural inheritance of Late Triassic normal faults

It is clear from the geological map of the Keszthely Hills that the uppermost Triassic deposits (Rezi and Kössen Fm.) preserved only in the core of two N-S trending synclines (Rezi and Vállus synclines; Fig. 15.a). However outcrop-scale folds of the Keszthely Hills with hectometric wavelength show various trend; the NNW-SSE trending anticline which is exposed in the Gyenesdiás quarry has also such oblique trend (Fig. 14). However, the northern continuation of the "Gyenesdiás anticline", exposed in the Pilikán quarry, has N-S strike, which is corresponding to the general trend (Fig. 15.d). In other quarries, further west, a quarry-scale anticline and a syncline show totally different NE-SW trend (Figs. 15.b and c).

On one hand, these three fold trends can be interpreted as the result of poly-phase folding, where the individual fold groups represent individual folding phases. On the other hand it is obvious that deviation in fold trend occurs along major pre-existing normal faults, namely the Cserszegtomaj fault, therefore the obliquity of folds can be explained by rather structural in heritance of pre-existing normal faults (Figs. 15.a and h). The way of this oblique inversion is described below.

The map-view pattern of the Pilikán-Gyenesdiás and Büdöskút anticlines shows collinear change (Fig. 15a); while they are N-S trending near the Pilikán quarry, their postulated southern continuation near the Gyenesdiás quarries is NNW-SSE trending meaning that the anticlines change their trend just in the vicinity of the Cserszegtomaj fault (Fig. 15a). This geometry also suggests the buttressing of the Pilikán-Gyenesdiás anticline against the pre-existing Cserszegtomaj fault.

The two main segments of the Late Triassic Cserszegtomaj fault are most likely connected by a relay ramp (Fig. 10; Héja et al., 2018). The NE-SW trending anticline of the Molnárkő quarry and the ENE-SWS trending Pajtika syncline developed directly on this structure (Figs. 15.a and h). The relay ramp between the two NNE-dipping normal faults probably had a south-eastern dip before the shortening, which might have promoted low-angle layer-parallel detachment below those folds which strike obliquely to the general trend (Figs. 15.a and h). It is also possible, that the formation of these oblique folds was due to the inversion of minor faults that initially breached the relay ramp.

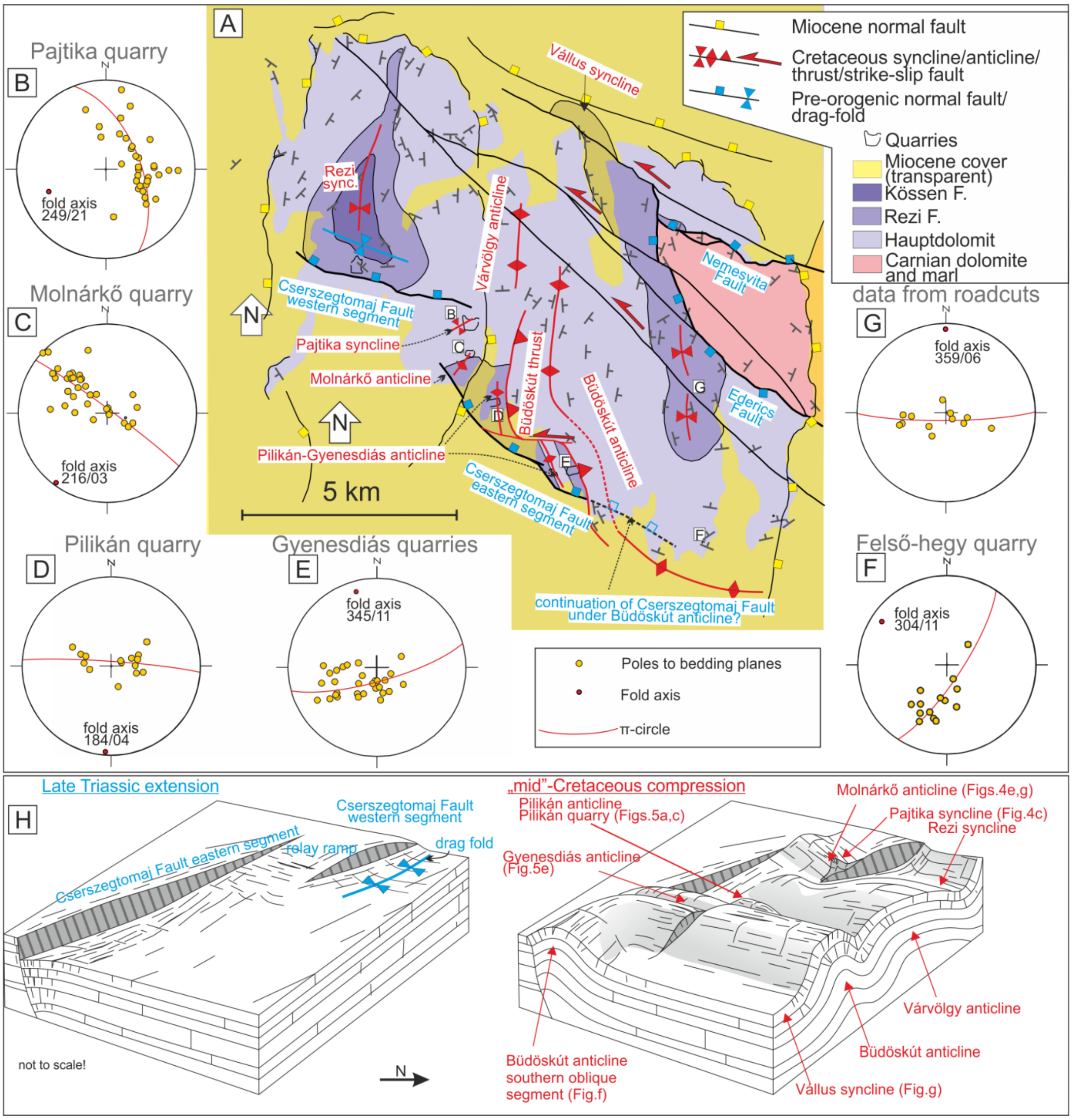

Fig. 15. A) Oblique folds along the pre-orogenic Cserszegtomaj fault (Keszthely Hills). (B-G) Stereoplots of the measured folds in different quarries. Localities are indicated on Fig.a. (h) Schematic 3D model of structural inheritance in the Keszthely Hills: the left figure shows the Late Triassic geometry of the Cserszegtomaj fault; the right figure shows the geometry of the structures related to mid-Cretaceous inversion. The block model incorporates approximately the area shown on Fig.a.

Topic 2 – Miocene topography and basin formation

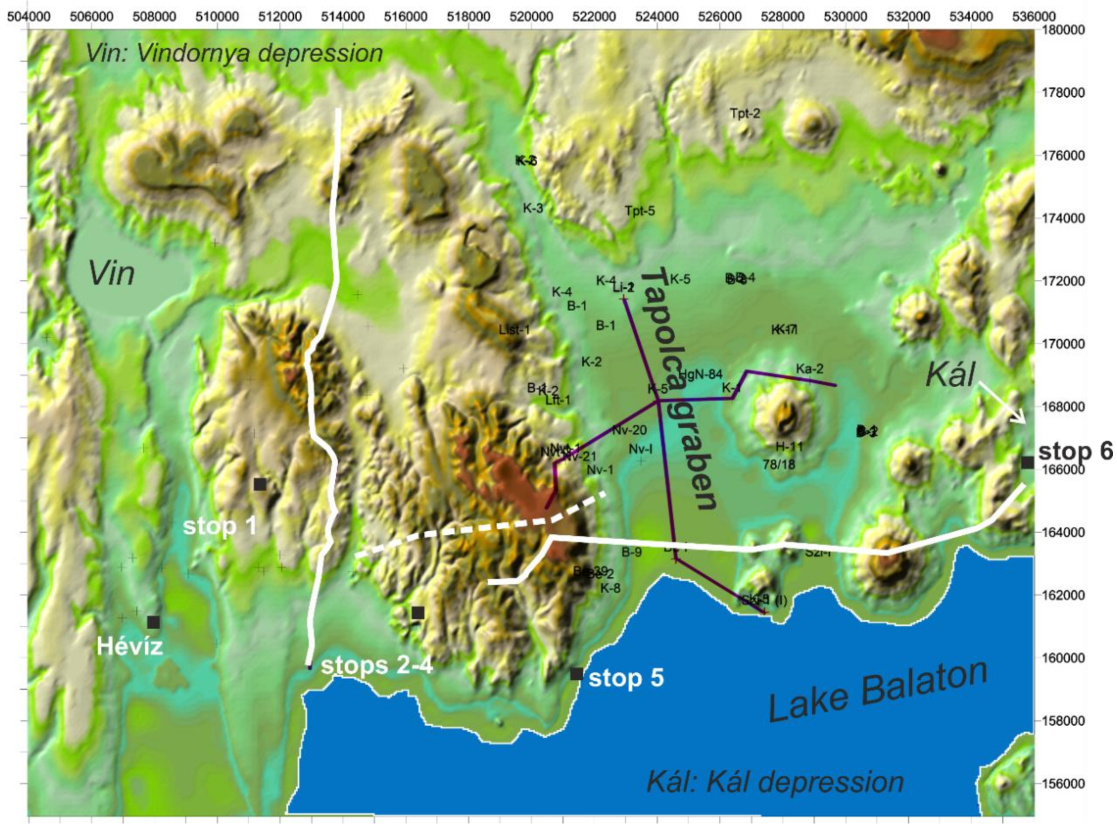

Fig. 16. Morphology of the Keszthely Hills and Tapolca graben with geological (solid line) and topographic (dashed) sections and boreholes

Stop 3. Gyenesdiás, Festetics belvedere (viewpoint)

Structures and morphology of the southern Keszthely Hills – the story of rifting "on land"

The 360° panoramic view permits the discussion on the morphologic, tectonic and sedimentological evolution of this segment of the Pannonian basin. Looking to the N, we can see a ~W–E cross-section of the Keszthely Hills divided into three morphologically distinct parts (as described in Csillag & Nádor 1997).

As frequently in the Pannonian basin, the Permo-Mesozoic successions suffered terrestrial denudation during several phases; this denudation resulted in sub-horizontal surfaces. In the study area, an important denudation phase took place during the late early to early Middle Miocene ca from 18 to 14 Ma. In the Keszthely Hills, it resulted in the formation of a sub-horizontal denudation surface, which was dissected by vertical dolines (sinkholes) reaching max. 50m depth and valleys. The sinkholes were filled with sediments, mostly kaolinitic clays and underwent in-situ weathering during the Middle Miocene Climatic Optimum (17 to 14 Ma). The infill was dated by zircon crystals deriving from airborne volcanic ash originated from the Carpathian-Pannonian Neogene volcanism; the most probable source is NE Hungary (Lukács et al. 2018). The age spectrum of Miocene zircons is between 16 to 14 Ma in a Keszthely Hills borehole (Kht-4) while is somewhat older in a nearby borehole (18–16 Ma, Uzsa-2, samples UZB, UZM, UZT) (Fig. 17) (Kelemen et al. in prep and 2019).

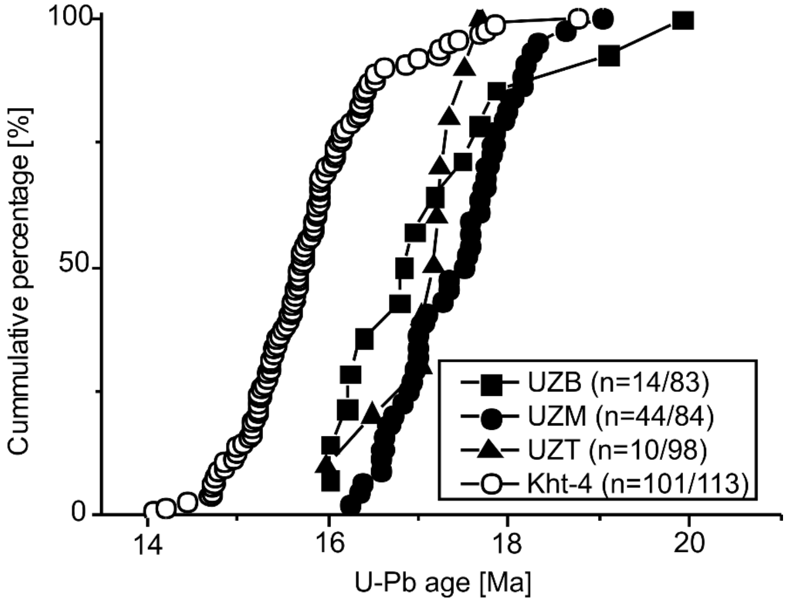

Fig. 17. Miocene part of the U-Pb age spectrum of zircon crystals in the sinkholes near the Keszthely Hills. View from Stop 3 encompasses the location of borehole Kht-4.

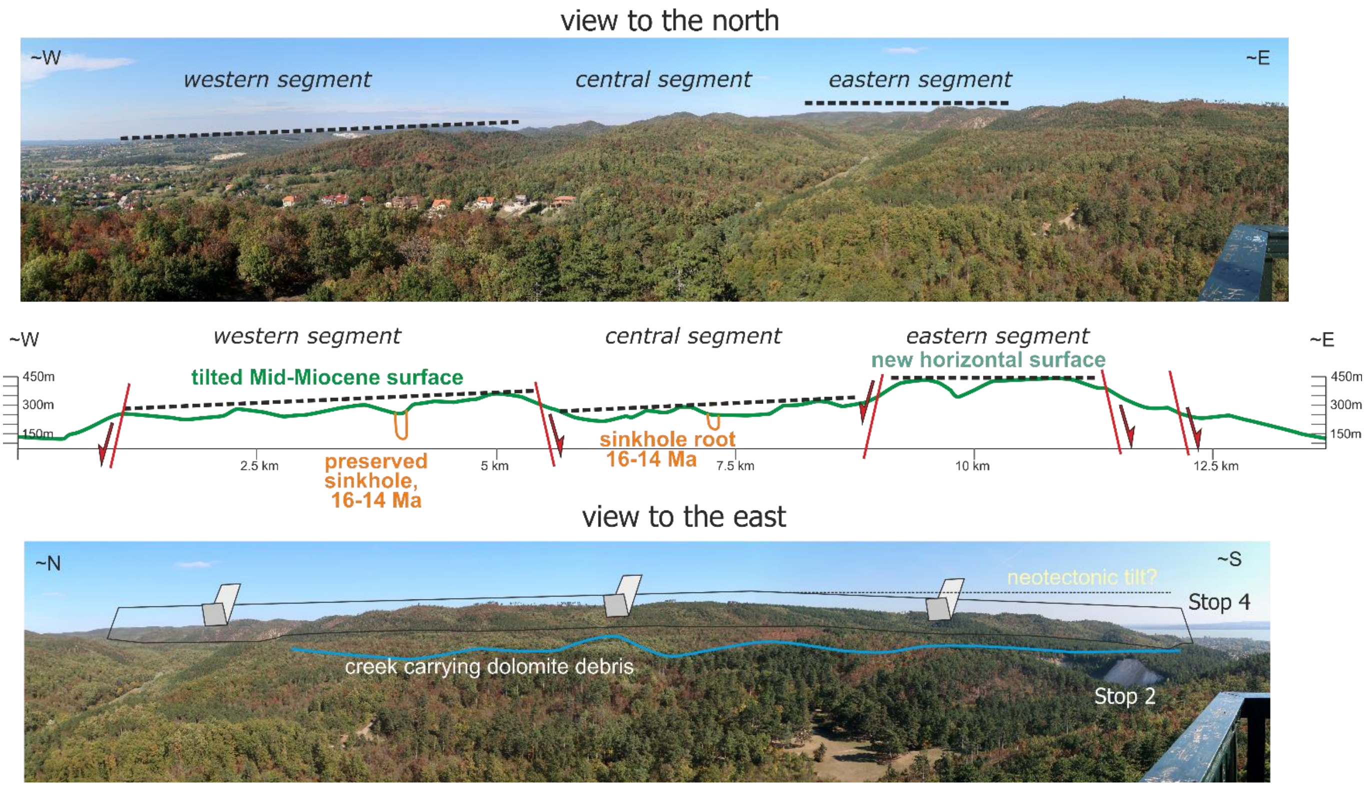

Fig. 18. Miocene part of the U-Pb age spectrum of zircon crystals in the sinkholes near the Keszthely Hills. View from Stop 3 encompasses the location of borehole Kht-4.

The existence of the deep dolines proves that this area represented and elevated karstic plateau, at least 50m higher than the surrounding area. To the north and west, boreholes demonstrate marine sedimentation during most part of the middle Miocene, and fault systems can be detected around the Keszthely Hills. This fault system can be followed to the NW, where syn-rift sediment thickness reached several hundreds of metres.

This time span corresponds to the most active deformation in the SW Pannonian basin (Zala-Mura basin), where crustal extension extended in time from ~ 18.5 to ~ 15 Ma (Fig. 21). This means that the Keszthely Hills represented an elevated footwall block of the major syn-rift normal faults, namely the Baján detachment (Fig. 3,4). While 1-1.5 km of sediments were accumulated in the southwestern syn-rift grabens (Zala basin), the Keszthely Hills were in terrestrial condition, and under denudation.

After the initial phase of the rifting, the whole area is gently tilted to the west, and possibly during this tilting dissected by new (or reactivated?) normal faults. From the viewpoint, we can see this tilted surface (Fig. 18, 21). The western segment of the Keszthely Hills still preserves almost intact the Mid-Miocene denudation surface, where the infilled dolines preserved in their original depth. In the middle part of the hills the terrain is rugged, and only the peaks of the hills correspond to the former mid-Miocene denudation surface. Here the valley incision possibly continued during the tilting, but the steep dolines were eroded to their bottoms. (Of course, Plio-Quaternary exhumation also contributed to the dissection of the area). In the eastern segment, see in the background, no Miocene dolines can be detected, probably because they were eroded (Fig. 18, 21). Instead, a sub-horizontal new denudation surface developed which remained intact and was not dissected by valleys (Fig. 18). We interpret this surface as a new denudation surface having cut the former tilted surface. This could form in the same 14–10 Ma time span as the tilting, but evidently somewhat later.

Csillag & Nádor (1997) estimated the number and displacement of these faults, using the results of geological mapping and geomorphic markers. The displacements could reach 40–100 metres, so the crustal stretching remained modest. This process could happen between 14 and ca. 10 Ma, before the transgression of the Lake Pannon onto the Keszthely horst.

During this time span, the basins around the Keszthely Hills subsided, although only a thin late Middle Miocene sediment pile was formed. This sedimentation can be proved by boreholes and the resulting grabens are shown in cross-sections. Boreholes demonstrate that fine-grained sedimentation continued into the earliest Late Miocene (Pannonian) from 11.6(?) Ma up to ca. 10 Ma. (cPa on fig. 19.).

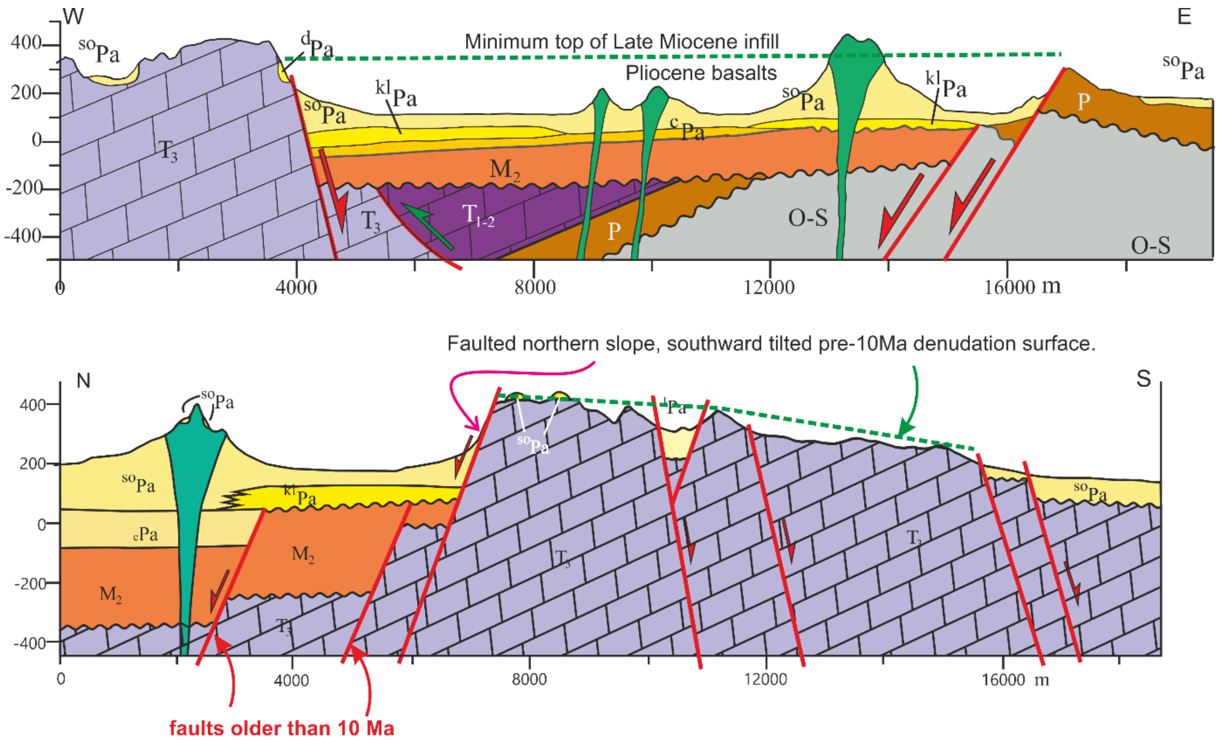

Fig. 19. Cross sections in the Tapolca graben and across the Keszhely Hills. Location of sections see on Fig. 16.

The timing of the faulting can be summarized as follows:

(1) We have indirect evidence that the Keszthely Hills were surrounded by the syn-rift fault system. It is to note, however, that these faults had modest separation, not reaching 1 km as largest.
(2) The tilting and the dissection of the Mid-Miocene ("syn-rift") denudation surface argue for considerable deformation after 14 Ma, before 10 Ma flooding
(3) Continuation of faulting in the early late Miocene.

Phase 2 and 3 are already post-dating the faulting in the deep part of the SW Pannonian basin (Mura-Zala basin). There we can see km-scale faulting, low-angle detachment faulting in operation only up to ca. 15 Ma (intra-Badenian unconformity), the younger sequences, and particularly the Late Miocene sediments are almost completely undeformed (Tari 1994, Fodor et al. 2013b).

This may suggest that the post-15 Ma deformation was concentrated to the "shallow part" of the Pannonian basin, to the formerly elevated Transdanubian Range. In fact, active faulting up to ca. 9 or 8.5 Ma was widespread, all over the range (Magyar et al. 2017, Fodor et al. 2013a, b) so this scenario is not an exception but the rule for the TR.

The 10 Ma old flooding was associated with the formation of small-scale, locally sourced deltas (Sztanó et al. 2010, referred to as Kálla Fm., klPa on Fig. 19). The transgression was soon followed by normal regression, because the distally-sourced large-scale deltas and shelf slope of the western Pannonian basin reached the elevated TR. Biostratigraphic data suggest to 9.7–8.7 Ma for this process (Hably 2013, Magyar et al. 1999). The deltas and subsequent fluvial sedimentation had aggradational character so the area continued to be subsided. This regional character led scientists to consider Late Miocene as post-rift in character, but persisting faulting may suggest another view.

Building on the above described observational data makes the Pannonian Basin an ideal natural laboratory for understanding the coupling between deep Earth and surface processes. All those data point to a clear shift of active extension and syn-tectonic sedimentation area from the SW (Mura-Zala basin) toward the Pannonian basin centre. Such shift, maybe even in larger amount, is predicted by thermodynamic modelling of asymmetric extension (Balázs et al. 2018). Such models are compared to observed subsidence (and uplift) data in the Pannonian Basin, which can be compared with other extensional basins with similar asymmetric extensional history. Geophysical and geochemical data implies the thinned lower crust underneath the TR as a result of Middle Miocene syn-rift extension that controlled local modest subsidence. Late Middle Miocene and Late Miocene subsidence of the area is caused by the overall basin-scale sagging related to the final thinning of the lithosphere. The late stage differential vertical movements of the TR and the Recent observed residual topography is the results of the indentation and pushing of the Adriatic micro-continent and further lithospheric bending and asthenospheric support.

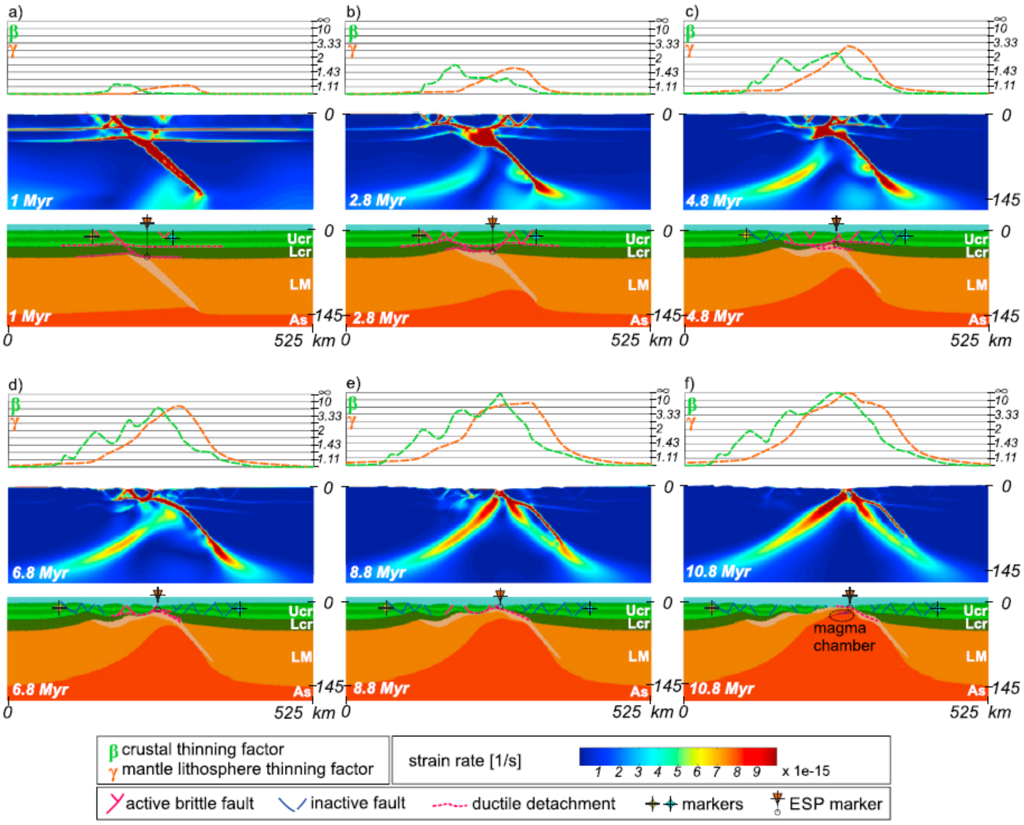

Figure 20. Thermo-mechnical tectonic evolutionary model of asymmetric extension (Balázs et al. 2018). Crustal and lithospheric mantle thinning values underlain by strain rate figures. (bottom row). The phase configuration of the model indicating the active and inactive structures. Two marker points are plotted at the sides of the extending domain. Syn-rift extension of the Pannonian Basin corresponds to figure 20d), while further extension results in crustal break-up and oceanic spreaing (figure 20f).

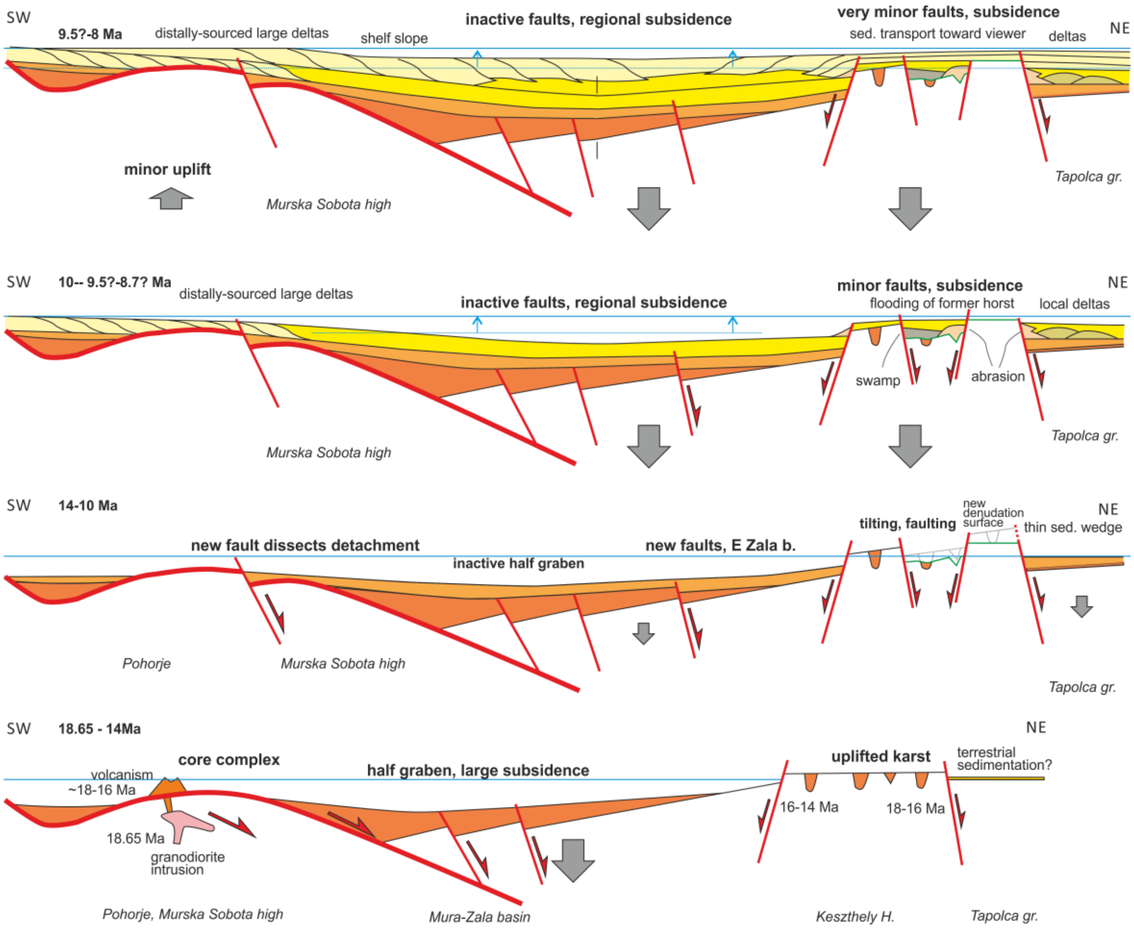

Fig. 21. A model for subsidence and uplift in the southwestern Pannonian basin; comparison of the Mura-Zala basin part and the Keszthely Hills- Tapolca graben

Stop 4. Gyenesdiás, Vadlány-barlang ("wild girl cave"): late Miocene transgressive sediments

This stop exposes the basal part of the transgressive sequence of the late Miocene (Pannonian) sedimentary sequence. The tilted denudation surfaces seen from the previous outcrop, the incised valleys, and the variably eroded dolines were flooded by the brackish Lake Pannon. The first sediments on the horst are two types. The very fine-grained clastics, often with lignite intercalations were formed in bays protected from wave action and coarse clastic inputs, while abrasional gravels deposited along steep slopes and on wave-dominated coastal segments. In this latter environment, small creeks arriving from the incised valleys formed deltas with coarse clastic material; an outcrop exposing such sediment is the stop 4. Abrasional gravels were mapped all around the Keszthely Hills. They are not dated directly, but projection of biostratigraphic data from nearby sites suggest a transgression around 10 Ma (Csillag et al. 2010).

The cliff exposes breccia and conglomerate units. Grain size is variable, from cm to decimetre, this later clasts are outsized form average. 1-5cm layer thickness is frequent, probably referring to sheetwash sedimentation. Clasts are completely angular (breccias) or well-rounded (derived from abrasion). Units are frequently clast-supported. Rounded gravels show imbrication dipping to the south, toward the lake. We can interpret these signs that angular breccia clasts arrived from the elevated background, possibly by temporal creeks, and shed breccias to the Lake Pannon. The creek could follow 14 to 10 Ma faults, Sediments arriving to the lake could form small fans. With slight water level rise, clasts were reworked by abrasion and became rounded. Because the elevation of abrasional gravel are variable, we can prove gradual flooding of the Keszthely horst almost completely (Fig. 21).

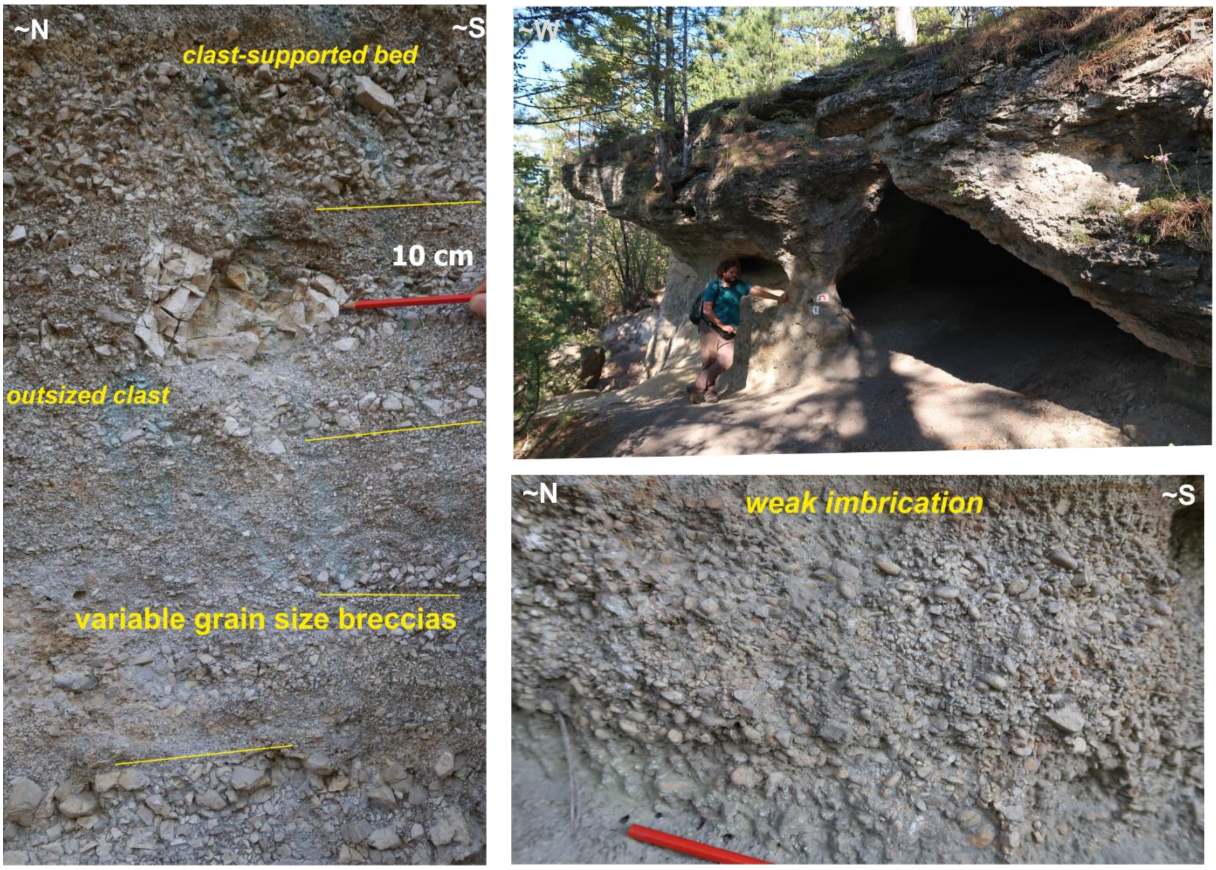

Fig. 22. 10 Ma old abrasional gravel and breccia of the base of the transgressive sedimentary cycle which covered the Keszthely horst.

Topic 2 and 3 – Miocene basin formation, Pliocene volcanism and Miocene to Quaternary landscape evolution

Stop 5. Balatongyörök/Balaton Ederics, Szépkilátó Belvedere

We have a view to the footwall of the eastern boundary fault of the Keszthely horst, to the Tapolca graben and to its spectacular landscape spotted with remnants of Pliocene basalt volcanoes.

As we described earlier, the subsidence of the Tapolca graben started during the late Middle Miocene (Fig. 21). It is not clear if sedimentation continued across the Middle to Late Miocene boundary, (or there is a hiatus) but the earliest late Miocene clastic sediments (ca. ~11-10 Ma) are present in the deepest part of the graben. This sequence is missing from the Keszthely horst!

A new wave of transgression started around 10 Ma, and the Pannonian Lake flooded the horst (see previous stop, and Fig. 21). We can see the boundary fault of the Tapolca graben, which bounded the Keszthely hills to the east. The visible steep slopes correspond to eroded fault scarps, partly inherited from the mid-Miocene, partly reactivated during the Late Miocene. Coarse talus breccias interfingering with abrasional gravels, the presence of outcrop-scale faults in the Late Miocene sediments argue for the continuation of faulting during the Late Miocene. However, displacements ar small, always less than 400m, so the extension remained also minor.

Deltas and rivers filled the graben completely by 8.7 Ma (Magyar et al. 1999, Hably 2013, Sztanó et al. 2010). By that time, the Zala basin was also filled, although there the deltas were connected to high shelf slopes prograding into the deep lake (Fig. 21). The western part of the basin could have been uplifted by that time (Fig. 21). We can thus conclude that the subsidence was regional character, and not only related to compaction. Faulting was too small, and could not cause the increase of accommodation space.

Use of volcanic landforms for uplift estimates

The spectacular part of the landscape are the basalt volcanoes. Basalt volcanism started here around 4.5 Ma. We can see the remnants of maar lava lakes (Szt György Hill, Badacsony), Szigliget diatreme.

The volcanic edifices determine the minimum upper surface of the Late Miocene sediments. However, we do not know, how much of Pannonian sediments were deposited above the highest preserved level, and eroded completely before the onset of volcanism at ca. 4.5 Ma. Because of the long time span, and the general aggradation type of Pannonian sedimentation, we can postulate several hundreds of metres of such disappeared sediments. The thickness depends on the uplift rate, and on the time of the change from subsidence to uplift, the onset of neotectonic inversion phase. Placing the onset of inversion to 6 or 7 Ma, and using the maximum uplift rate (0.1 mm/a) we can arrive to maximum surface of Late Miocene sediments up to 600 or 700m, being 200 to 300m higher than the highest preserved Late Miocene sediments (ca. 400m).

Volcanic landforms could be used for the reconstruction of past morphological surfaces of the TR (Németh & Martin 1999, Martin & Németh 2004). Lava lakes and maars could be compared to the former surface considering that the preserved maar surface was somewhat below the surrounding topography (ca. 30m, Fig. 23, Németh & Martin 1999, Martin & Németh 2004). Necks are more difficult to be involved in such estimation, while the position of the paleosurface can just be tentatively estimated (Fig. 23).

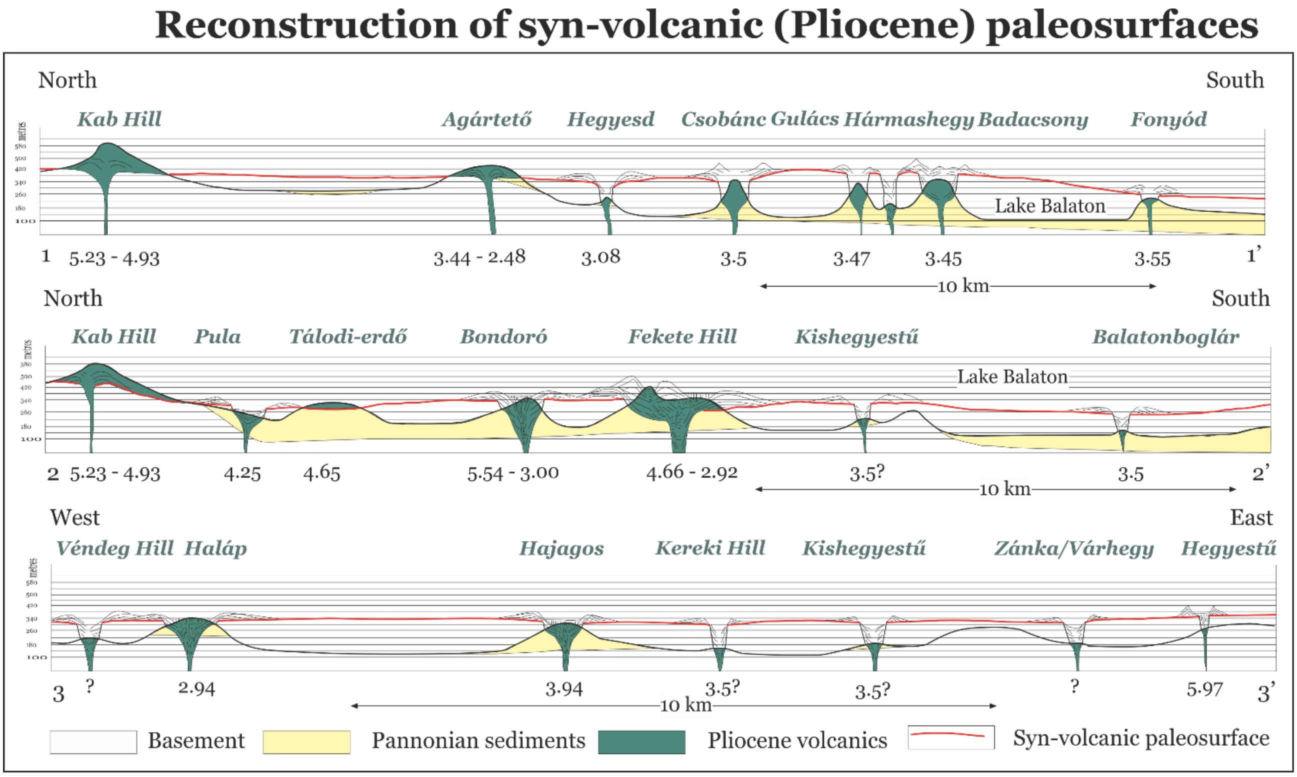

Fig. 23. Use of remnants of volcanic landforms to reconstruct former surfaces. Note important denudation.

From the age of the volcanic rocks and the reconstructed paleo-position of the adjacent morphological surface, a denudation rate can be calculated down to the level of the Lake Balaton, as to the lowest topography (local erosional base level). While the erosion of the basaltic rocks remained modest, the host Pannonian sediments were considerably lowered. Martin and Németh (2004) calculated the post-volcanic erosion rate to be between 20 and 100 m/Ma. A repeated calculation issued similar results, in the order of 39-100m/Ma suggesting a quite slow incision rate (lowering rate) in the Tapolca and Kál basin (Fig. 5) (Sebe & Csillag 2012, Fodor et al. 2014).

Nature of denudation after volcanism

The dominant means of surface denudation has been a longtime debated question in the western Pannonian Basin including the Kál and Tapolca Basins. At the beginning of the 20th century Lóczy (1913) and Cholnoky (1918) attributed the formation of these depressions to Pliocene deflation. Later authors suggested that fluvial erosion was of major importance (Bulla 1958, Góczán 1960). More recent geomorphological investigations of Borsy et al. (1986) suggested that present shape of the Tapolca Basin is most probably the result of wind erosion active during the Pleistocene glacial periods. Our observations strongly suggest deflation origin of these depressions (Fig. 24) (Sebe et al. 2011, Ruszkiczay-Rüdiger et al. 2011). Other arguments will be visited and discussed at stop 6 and 7.

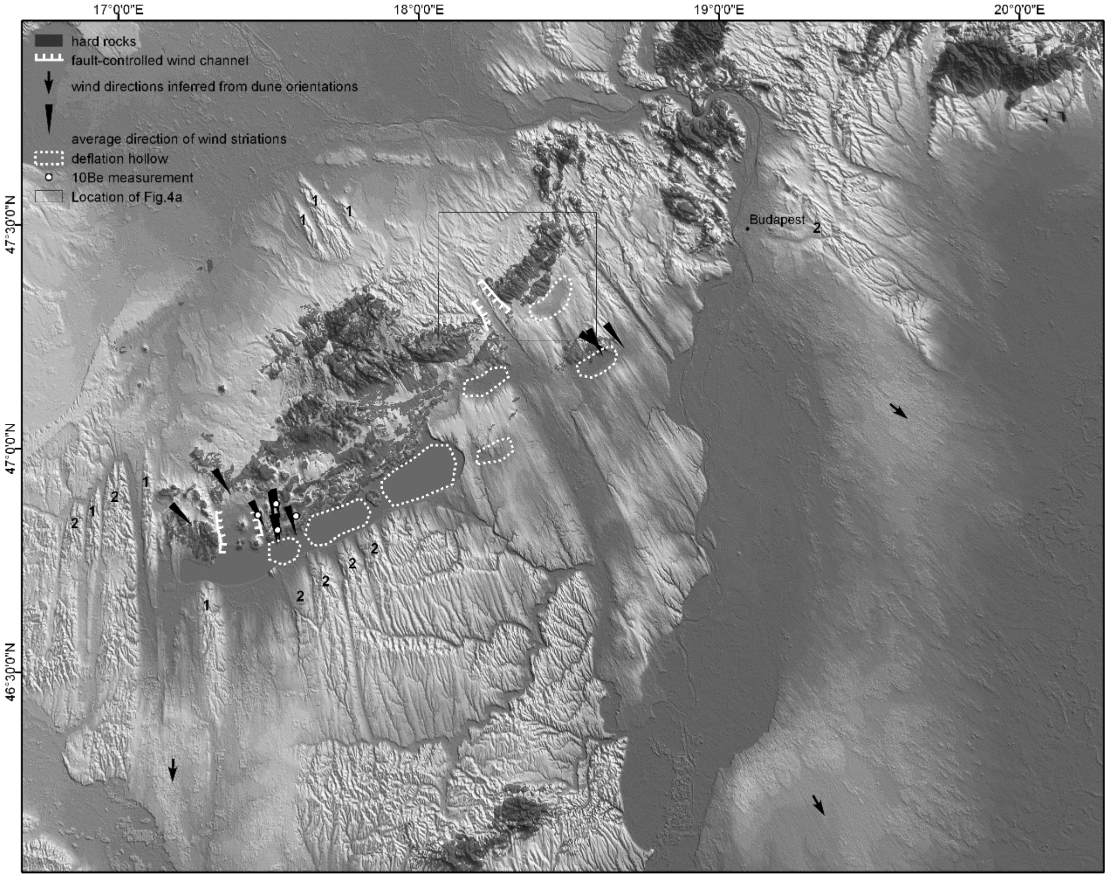

Fig. 24. Morphology of Transdanubia, and the deflational landforms (yardangs marked by 1 and 2, playa lakes, marked dotted lines, and measured wind-polished grooves

Stop 6. Salföld, "stone sea"

This outcrop serves evidence for strong deflation in the Balaton Highland, and in western Hungary in general. Wind-polished surfaces of cemented Late Miocene sandstone are widespread in this "stone sea", as popular name mention this site (Fig. 25). Grooves are trending NW-SE, which correspond to present and paleo-wind direction. The cementation of the loose Late Miocene sand occurred during Quaternary, during the exhumation of the sediment, when it was affected by groundwater silicification (Budai et al. 1999a, András 2012).

The age of this surface at 140 ma.s.l. and ca. 15m higher than the base of the local depression, Kál Basin, and 35m higher than the lake level, was determined using a cosmogenic ^{10}Be depth profile sampled the nearby quarry (Ruszkiczay-Rüdiger et al., 2011). Its exposure age is 287±23ka at an elevation of 140m, Ruszkiczay-Rüdiger et al. (2011) also dated the exposure age of other sites, yielding exposure ages of 0.87±0.07 Ma and 1.56±0.09 Ma. The well-constrained age of these sites show a gradual lowering of the surface with a rate of 43–53m/Ma. Other estimates of erosion rate (lowering rate) fall in the range of 40-75m/Ma (Fig 5, 26). All these estimates on denudation rate (incision rate) are in agreement with values coming from the denudation of paleosurfaces preserved by the volcanic edifices. Thus, the **low rate of 40–100m/Ma can be considered as realistic** for the southern part of the TR.

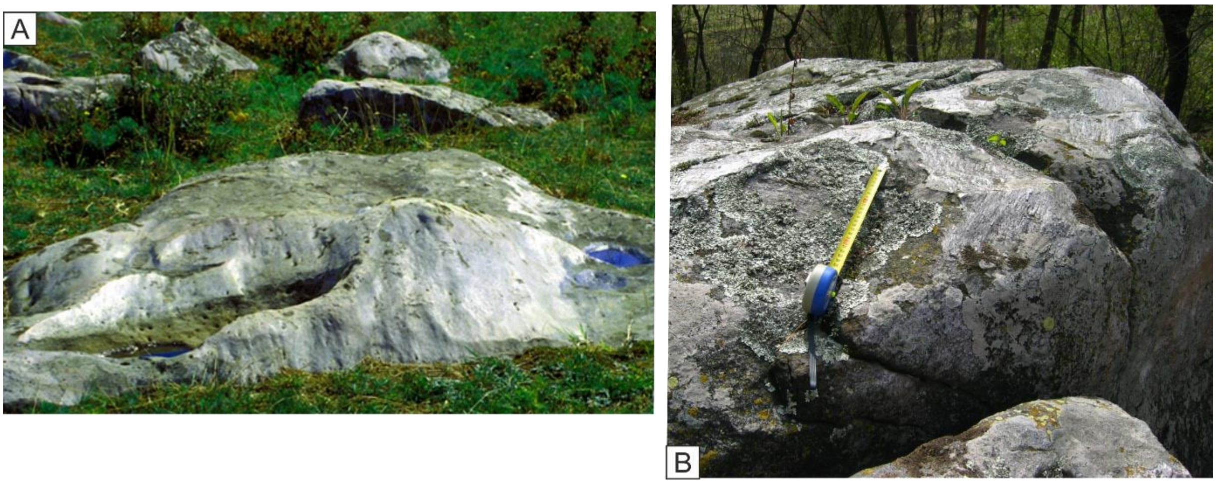

Fig. 25. Wind-polished surfaces at Salföld (A) and Kőmags (B). The Pannonian sand was cemented during Quaternary.

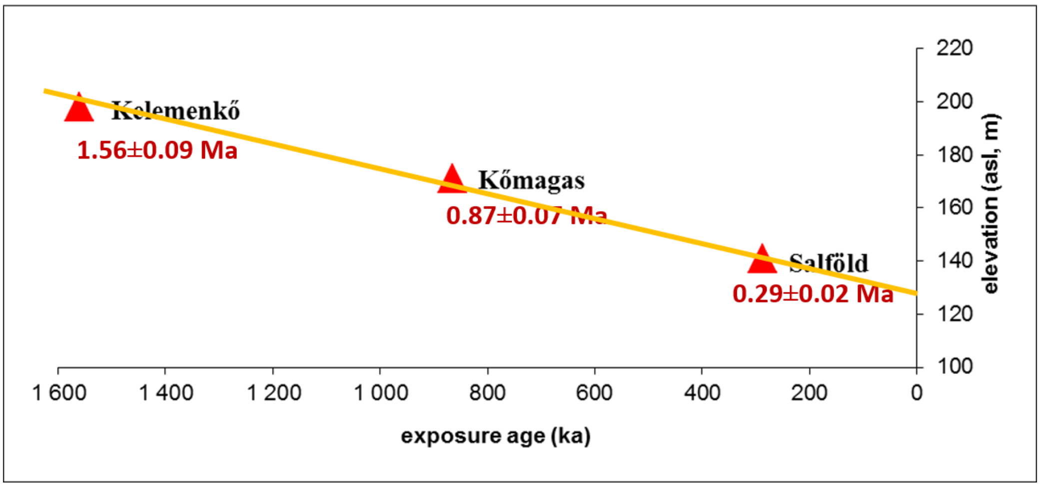

Fig. 26. Exposure ages of three sites in the Kál and Tapolca basins, and the denudation rate derived from the ages of gradually lowering morphological surfaces. (Ruszkiczay-Rüdiger et al., 2011). The slope of the trend-line fitted to the data suggest a surface lowering rate of 46±3 m/Ma.

Stop 7. Zánka, Hegyestű (17° 38' 53.4303" E 46° 53' 21.6885" N)

GPS: 46.889872, 17.646881

https://www.bfnp.hu/en/hegyestu-geological-visitor-site-monoszlo

The 336 m high basanite plug rises more, than 200 m above the level of Lake Balaton. The mixture of siliciclastic material and highly vesicular basanite tuff-breccia along the edges of the basalt body refers to magma – water interaction. The characteristic vertical columnar jointing of the basalt can be interpreted as sub-horizontal position of the lava body within a neck. The top of the Triassic strata is situated around 290–300m. The higher elevation level of the siliciclastic tuff-breccia refers to mixing of the eroded Pannonian sediments and the magma. Present-day morphology reflects the eroded remnants of pre-existing craters, which were formed in the Triassic and Pannonian strata (Martin & Németh 2004).

The radiometric age of the magmatic body is 7.94 +/-0.03 Ma with 40Ar/ 39Ar method (Wijbrans et al. 2007). K/Ar data of Keresztúri et al. (2011) suggests slightly younger time constraint: 7.56+/-0.17 Ma.

Panoramic view from the top of the Hegyestű Cliff gives a perfect overview on the geology and geomorphology of the Balaton Highland. Eroded Triassic rocks form the characteristic flat-lying surfaces in the western background (Keszthely Hills) and in the closer surrounding (Fig. 27). Even the most important Cretaceous structural element of the area (Litér Thrust) is expressed by the morphology only on those areas, where the footwall and hanging-wall is formed by remarkably different lithologies. Individual steeper hills emerge from this flat-lying, old erosional surface. There is a great number of volcanic edifices that were formed during the latest Miocene – Early Pleistocene continental basalt magmatic event (Fig. 27). Prior to the volcanism, Lake Pannon had flooded almost the whole Pannonian Basin, and had deposited siliciclastic sediments in varying thickness. The volcanic edifices were built onto this covered surface.

Two larger lava floods were formed: the Agártető and the Kab Hill. However, the most characteristic morphological features are the maars which are products of phreatomagmatic eruptions. Within these maars, lava lakes were formed. The surface of these previous lakes are equal to the present-day top of the volcanic peaks; only minor amount of erosion is supposed. In contrast with it, considerable amount of the Pannonian sediments has been eroded from the vicinity of the maars and lava flows resulted in this geomorphological inversion (Fig. 22). In consequence, maars actually form elevated peaks and plateaus.

Wind erosion played an important role among the denudation processes (Fig. 25, 28). Pannonian sediments of the Kál Basin (right at the foothill of Hegyestű Cliff) were transported away mainly by the wind (Ruszkiczay-Rüdiger et al. 2011). Dated aeolian surfaces suggest wind erosion during the last 1.5 Ma (Fig. 28). The denudation of the wind-polished surfaces was negligible, only 4-6m/Ma. However, the lowering of the regional morphological surface was around 50m/Ma, similar to what we obtained from the exhumation of volcanic landforms. While the general denudation was quite slow, the uplift rate of this part of the TR should also be relatively slow, possibly not exceeding 0.1mm/a (see Fig. 5).

Deflation origin is true for the Lake Balaton; it is in fact a playa lake. Several similar depressions exist in the near vicinity and in other part of Transdanubia (Fig. 24). All these lakes lie in the leeward side of a morphological ridge, which can be the TR, the basalt hills north from the Keszthely horst. They are all shallow, flat-bottomed. The turbulent winds had strong capacity to deflate the loose Late Miocene sand. The only protected landforms are silicified sandstones, volcano remnants and Triassic carbonate horst. Wind deflation repeated several times during the cold periods of the Quaternary.

Fig. 27. Panoramic view form the Hegyestű to the east. Note flat-bottomed depression of the Kál basin, which is similar in origin as the Lake Balaton.

Acknowledgment

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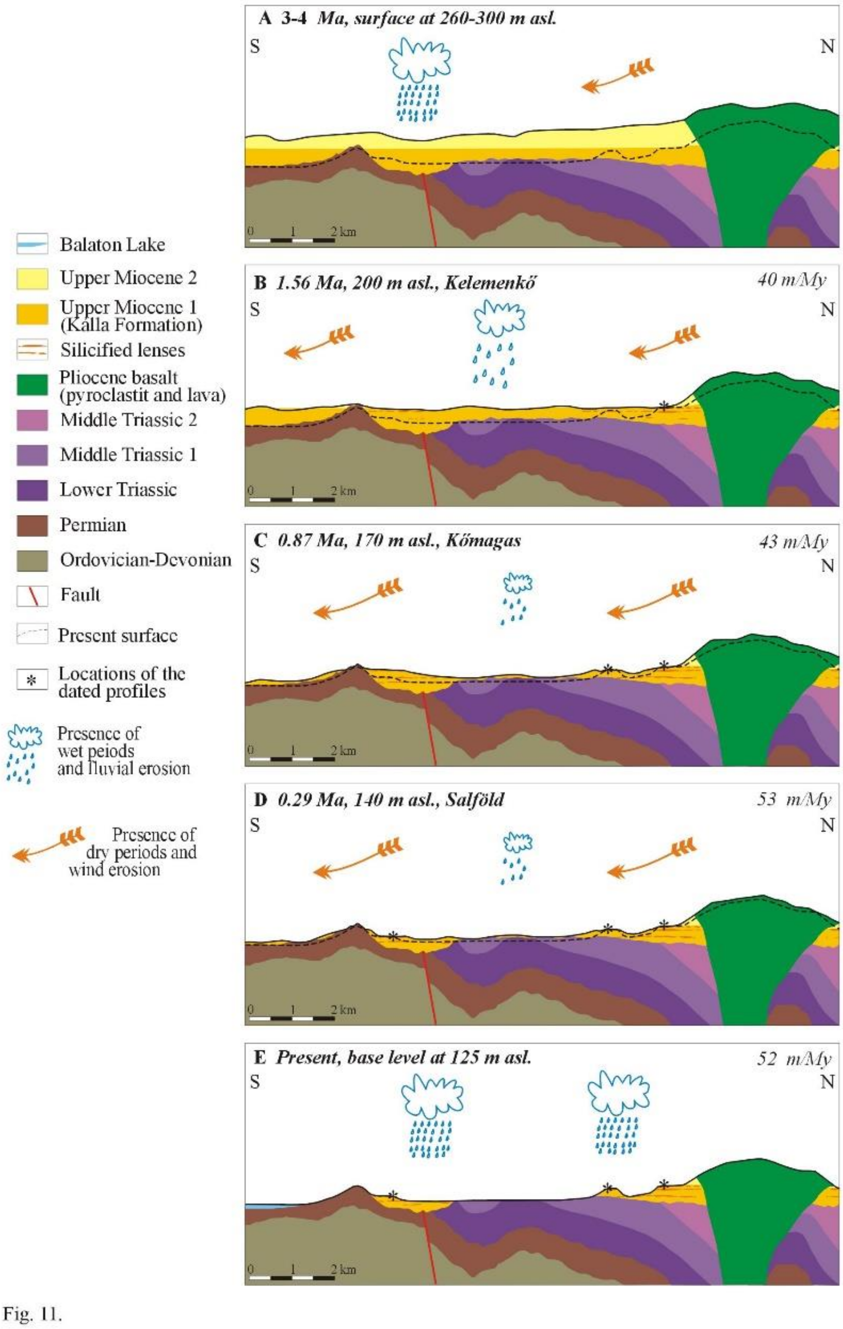

Fig. 28. Morphological evolution of the Kál depression (Ruszkiczay-Rüdiger et al. 2011)

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