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ORIGINAL PAPER

Evidences for pre-orogenic passive-margin extension in a Cretaceous fold-and-thrust belt on the basis of combined seismic and field data (western Transdanubian Range, Hungary)

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Abstract
Combined sedimentological and structural analysis was carried out in the field and on 2D seismic reflection profiles to recognize pre-orogenic structures in a Cretaceous fold-and-thrust belt. Detailed field observations were made in the Keszthely Hills, Western Hungary, while 2D seismic interpretation was carried out in the neighbouring Zala Basin. As a result, a fault-controlled intraplatform basin system was identified by a detailed analysis of bounding faults, and related outcrop-scale structures. The Norian–Rhaetian (227–201.3 Ma) synsedimentary faulting was associated with talus breccia formation, small-scale faulting, and dyke formation, in addition to slumping and other soft-sediment deformations. Based on the distribution of talus breccia, WNW–ESE-trending map-scale normal faults were identified in the Keszthely Hills, which is in agreement with the directly observed outcrop-scale synsedimentary faults. On seismic sections, similar WNW- or NW-trending Late Triassic normal faults were identified based on thickness variations of the syn-rift sediments and the presence of wedge-shaped bodies of talus breccia. Normal faulting occurred already in the Norian, and extensional tectonics was active through the Early and Middle Jurassic. The Late Triassic grabens of the western Transdanubian Range could be correlated with those in western part of the Southern Alps, and the Bajuvaric nappe system of the Northern Calcareous Alps. These grabens were situated on the proximal Adriatic margin, and they represent the first sign of the Alpine Tethys rifting. The locus of extension was laterally migrated westward, towards the distal Adriatic margin during Early and Middle Jurassic.

Keywords Pre-orogenic extension · Synsedimentary deformation · Norian tectonics · Alpine Tethys rifting · Triassic paleogeography

Introduction

Pre-orogenic structures have an increasing role in the structural interpretation of thrust and fold belts (Butler et al. 2006). Several balanced sections show that the retro-deformed original stratigraphy cannot be considered as a layer-cake and prominent pre-orogenic deformation can be recognized (Perez et al. 2016; Yagupsky et al. 2008; De Vicente et al. 2009). Identification, investigation, and understanding the structural geometry, fault pattern, and deformation style of these pre-orogenic structures have primary importance, since these early structures may have a significant effect on the final geometry of subsequent folding and thrusting.

In many cases, structural inheritance is responsible for the development of backthrusts, young-on-older thrust, and compressional structures non-perpendicular to the shortening, such as oblique or lateral ramps (Bonini et al. 2012; Pace et al. 2014; Ustaszewski and Schmid 2006).

In most cases, pre-orogenic faults develop during the passive-margin evolution, before the onset of shortening. However, normal faults can also be re-activated or newly evolve later, during foreland basin evolution, due to the

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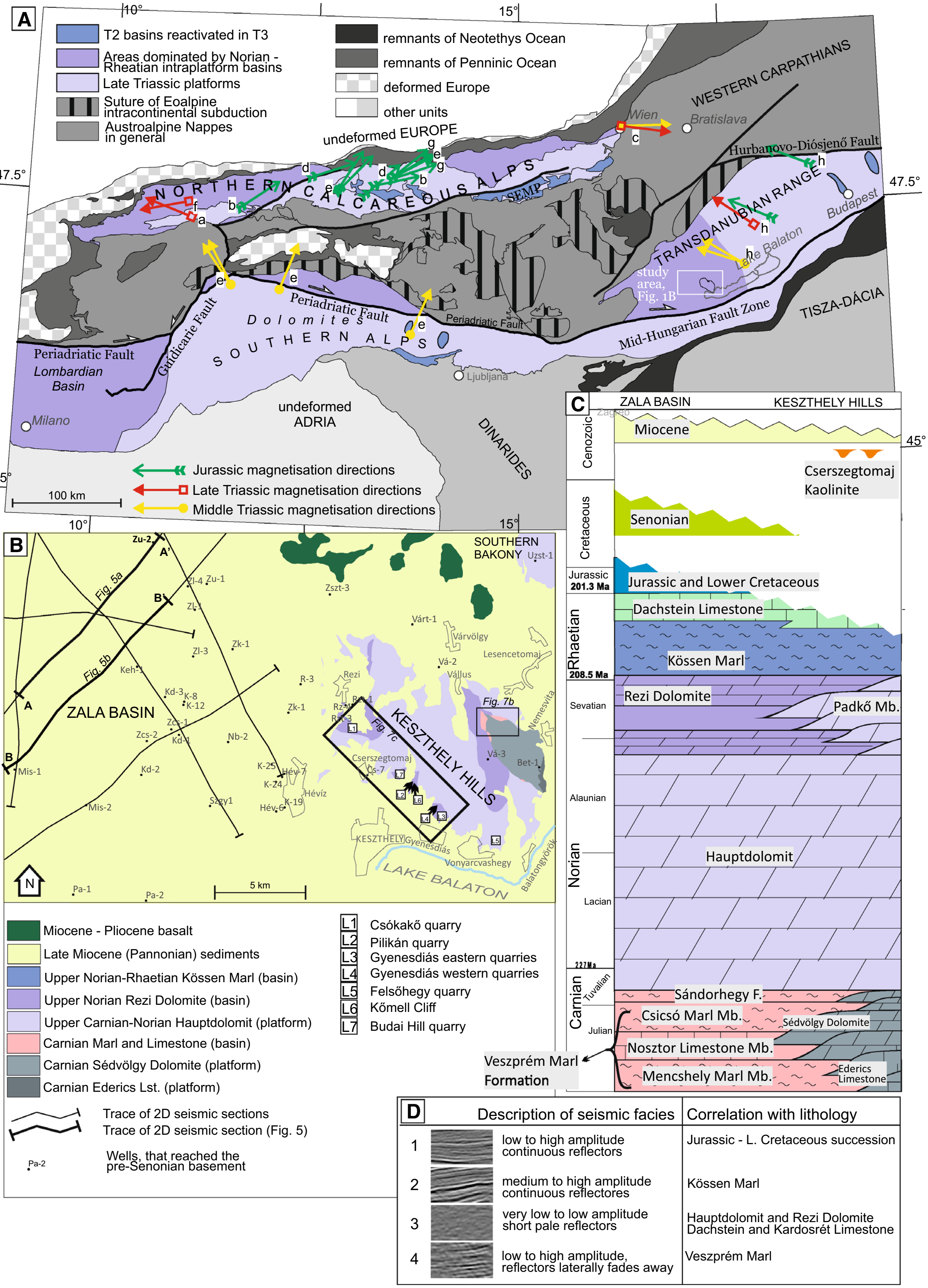

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◂**Fig. 1** **a** Position of the study area on the simplified paleogeographic map of the Alps (after Schmid et al. 2008; Goričan et al. 2012; Haas 2002). Austroalpine cover nappes were coloured on the basis of their dominant Late Triassic depositional environment (basin/platform). Yellow arrows show the Middle, Late Triassic, and Jurassic magnetization directions based on the measurements of **a** Becke and Mauritsch (1985), **b** Channell et al. (1990), **c** Gallet et al. (1998), **d** Heer (1982), **e** Mauritsch (1980), **f** Mauritsch and Becke (1987), **g** Mauritsch and Frisch (1978), and **h** Márton and Márton (1983). **b** Geological map of the study area based on Budai et el. (1999b) and Császár and Gyalog (1982). For coordinates of the investigated outcrops, see Table 1. **c** Stratigraphy of the Keszthely Hills and the Zala Basin based on Csillag et al. (1995) and Kőrössy (1988). Formations are shown in approximately proportion to thickness. **d** Seismic facies types of the pre-Senonian basement of the study area

flexure of the subducting lower plate (Butler et al. 2006; Billi and Salvini 2003).

Investigation of pre-orogenic normal faults is often complicated, since such structures can be strongly overprinted by later compressional deformation in thrust belts. However, pre-orogenic synsedimentary structures are often accompanied by secondary features, which may survive the basin inversion. Such features are abrupt facies changes, reflecting significant changes in depositional environments (e.g., deepening) and characteristic sediments and sedimentary structures related to fault activity. However, synsedimentary extension creates facies change only in those cases, if the rate of extension-related subsidence of the hanging wall is significantly larger than the rate of deposition. If the deposition keeps pace with the subsidence of the hanging wall, both the hanging wall and the footwall can have the same environment, and thus, the fault cannot be identified just on the basis of facies changes. In this case, thickness variation of the pre-orogenic succession can be an indicator of synsedimentary normal faulting. The most characteristic fault-related sediments are coarse-grained talus-cone breccias (Ortner et al. 2008); moreover, synsedimentary fault movements are often associated with soft-sediment deformation (Bergerat et al. 2011).

Pre-orogenic extension in the study area was already supposed by Csillag et al. (1995), based on the presence of coarse breccias and facies distribution. However, they neither determined the exact position of faults nor characterized the fault pattern or the stress field. In our study, we demonstrate how the combined sedimentological and structural observations, fault-slip analysis, geological map interpretations, and 2D seismic sections can be used to identify and characterize pre-orogenic structures in a poorly outcropping area. The study area is the westernmost outcropping part of the Transdanubian Range (Keszthely Hills) and its western continuation submerged below the Cenozoic cover of the Zala basin (Fig. 1a, b). The aim of this paper is to describe the Late Triassic extensional structures, hitherto frequently cited but very rarely characterized. Our results can contribute to understanding the early phase of passive-margin evolution of the study area.

Geological setting

The Transdanubian Range was part of the Adriatic plate, which was situated between the Neotethys and the Alpine Tethys (Mandl 2000; Csontos and Vörös 2004; Schmid et al. 2008). First phase of rifting during Anisian was related to the opening of the western branch of the Neotethys (Haas et al. 1995; Budai and Vörös 2006), while rifting of the Alpine Tethys initiated during Late Triassic and Early Jurassic (Bertotti et al. 1993; Decarlis et al. 2017).

The following deformations of the Transdanubian Range are related to the closure of these oceans. The partial closure of the western part of the Neotethys led to the folding in the study area during the Albian–Coniacian (Fodor et al. 2017). These structures are discordantly covered by the Senonian strata, and they represent one of the most significant deformations of the Transdanubian Range.

After the final subduction of Alpine Tethys, collision and continental subduction of European plate below Adria occurred during the Late Paleogene. This event led to the eastward extrusion of ALCAPA unit along the Periadriatic Fault (Schmid et al. 2008) (Fig. 1a). The eastern continuation of this structure is the Mid-Hungarian shear-zone, which is located directly south of study area (Balla 1984; Csontos and Nagymarosy 1998; Fodor et al. 1999).

The Pannonian back-arc basin system was formed during the Miocene, in the hinterland of this subduction (Tari 1994; Horváth et al. 2015; Balázs et al. 2016). Therefore, the study area was affected by strong extension, and thus, the studied Keszthely Hills are parts of a Miocene extensional horst bounded by normal faults and grabens, such as the Zala Basin (Fodor et al. 2013).

As it was briefly summarized above, the geodynamic evolution of the study area was affected by the opening and closure of two distinct oceanic systems, and accordingly, it has a complex stratigraphy. In this paper, we focus on the Late Triassic and Jurassic synsedimentary deformations, and therefore, we describe only the coeval sediments in detail.

Late Triassic-to-Early Cretaceous stratigraphy

The oldest known formation of the study area is Carnian in age (Haas et al. 2014). The Carnian basinal marl and limestone (Veszprém Marl and Sándorhegy F.) are laterally interfingering with the coeval carbonate platform (Ederics Formation) (Csillag et al. 1995), which was partly dolomitized (Sédvölgy Dolomite) (Haas et al. 2014).

From the end of the Carnian, the Hauptdolomit Formation was deposited (Fig. 1c). The formation is built up by

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Table 1 Location of the investigated outcrops

Name of the outcrop	Latitude	Longitude
1. Csókakő quarry	46.82219	17.23101
2. Pilikán quarry	46.79337	17.27255
3. Gyenesdiás eastern quarry	46.7790	17.29165
4. Gyenesdiás western quarry	46.77971	17.28958
5. Felsőhegy quarry	46.76542	17.33641
6. Kőmell Cliff	46.7842	17.28789
7. Budai Hill quarry	46.8033	17.26448

Projection datum is WGS84

thin-layered bituminous dolomite in the study area. Occasionally, stromatolite 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), extension-related intraplatform basins were formed, which were filled up by late Middle–Upper Norian Rezi Dolomite and Rhaetian Kössen Marl (Haas 1993; Csillag et al. 1995; Budai and Koloszár 1987; Budai et al. 1999a). Budai and Koloszár (1987) subdivided the Rezi Formation into three members. The lower member is represented by dark-grey cherty bituminous-laminated dolomite. The middle member is made up by the alternation thin-layered and thick-layered dolomite, which often contains re-deposited green algae fragments (Fig. 1c). This middle member was interpreted as a platform progradation tongue of the coeval platform. The upper member is similar to the lower member. Another dolomite breccia lithofacies of the Rezi Dolomit with re-deposited platform-originated blocks were identified in the Csókakő quarry (L1) (Csillag et al. 1995).

According to Csillag et al. (1995), the Rezi Dolomite was deposited in a synsedimentary half-graben in the Keszthely Hills. The dolomite breccia represents the fault-bounded talus breccia of a synsedimentary normal fault. On the tectonically controlled elevated areas, carbonate platform environment still persisted. These areas are represented by the footwall of major normal fault bordering the Rezi half-graben, and the opposite edge of the half-graben (Csillag et al. 1995). From the edge of the half-graben, the propagation of platform occurred. Nevertheless, this model does not specify the exact geometry of the basin and the controlling normal faults.

The Rhaetian Kössen Marl Formation (Fig. 1c) is poorly exposed; therefore, it is rather known from wells (Haas 1993). It is made up by dark-grey-to-black shales with high organic matter content. Thin-bedded limestone intercalations occur frequently within the shale; it is strongly folded due to slumping (Budai and Koloszár 1987).

The younger Mesozoic formations were eroded in the Keszthely Hills, partly, due to the mid-Cretaceous folding (Fig. 1c). However, in the western subsurface continuation of the Keszthely Hills (eastern Zala Basin) and in the Southern Bakony (NE to Keszthely Hills), the younger members of the pre-Senonian succession could be traced. The Kössen Marl is interfingering with the limestone of the coeval Rhaetian Dachstein platform towards NE, based on well data; consequently, the Kössen Marl pinches out NE-ward (Haas 1993, 2002). Platform progradation of the few 100 m-thick Dachstein Formation can also be observed above the Kössen Marl in several wells of the Zala Basin (Kőrössy 1988).

The carbonate platform environment still existed in earliest Jurassic (Kardosrét Fm.); however, it drowned in the beginning of Sinemurian, due to extension-related strong subsidence (Fig. 1c). This extension also created horsts and grabens (Vörös and Galácz 1998). There was hiatus, or just condensed sedimentation on the top of the submarine horsts, while thin, pelagic formations deposited with variable lithology in the grabens (Haas et al. 1984).

The Lower Jurassic succession is characterized by pelagic red nodular limestone and grey cherty limestone (Haas et al. 1984; Vörös and Galácz 1998). During the Middle Jurassic, cherty limestone and radiolarites were deposited. The Upper Jurassic formations seal both the pre-existing horsts and the grabens (Vörös and Galácz 1998; Haas et al. 1984). This Upper Jurassic succession is made up by red nodular limestone and white pelagic cherty limestone. The deposition of the latter formation is lasted till the Early Cretaceous. From the Barremian to the Aptian silty sandy pelagic marl, AQ4 and then shallow marine limestone were deposited (Haas et al. 1984).

Methods

In the eastern part of the study area (Keszthely Hills), Triassic rocks are exposed; their microtectonic and basic carbonate sedimentologic field observations were carried out in dolomite quarries. To help the readers, the quarries were marked by numbers (see after the names of quarries in Sect. 4, and on map Fig. 1b). The measured structural data were illustrated on stereoplots. In the adjacent Zala Basin, Mesozoic basement is under thick Cenozoic cover. There, the investigation of Mesozoic basement is possible based on seismic data.

Stereoplots

Several types of structural data were measured in outcrops of the Keszthely Hills, which were plotted by the software of Angelier (1990) (for legend, see Fig. 2d). Fault-slip data rarely contained slicken lines, in most cases, just the fault planes were measurable. Therefore, fault-slip inversion was not carried out, since at least four slicken line

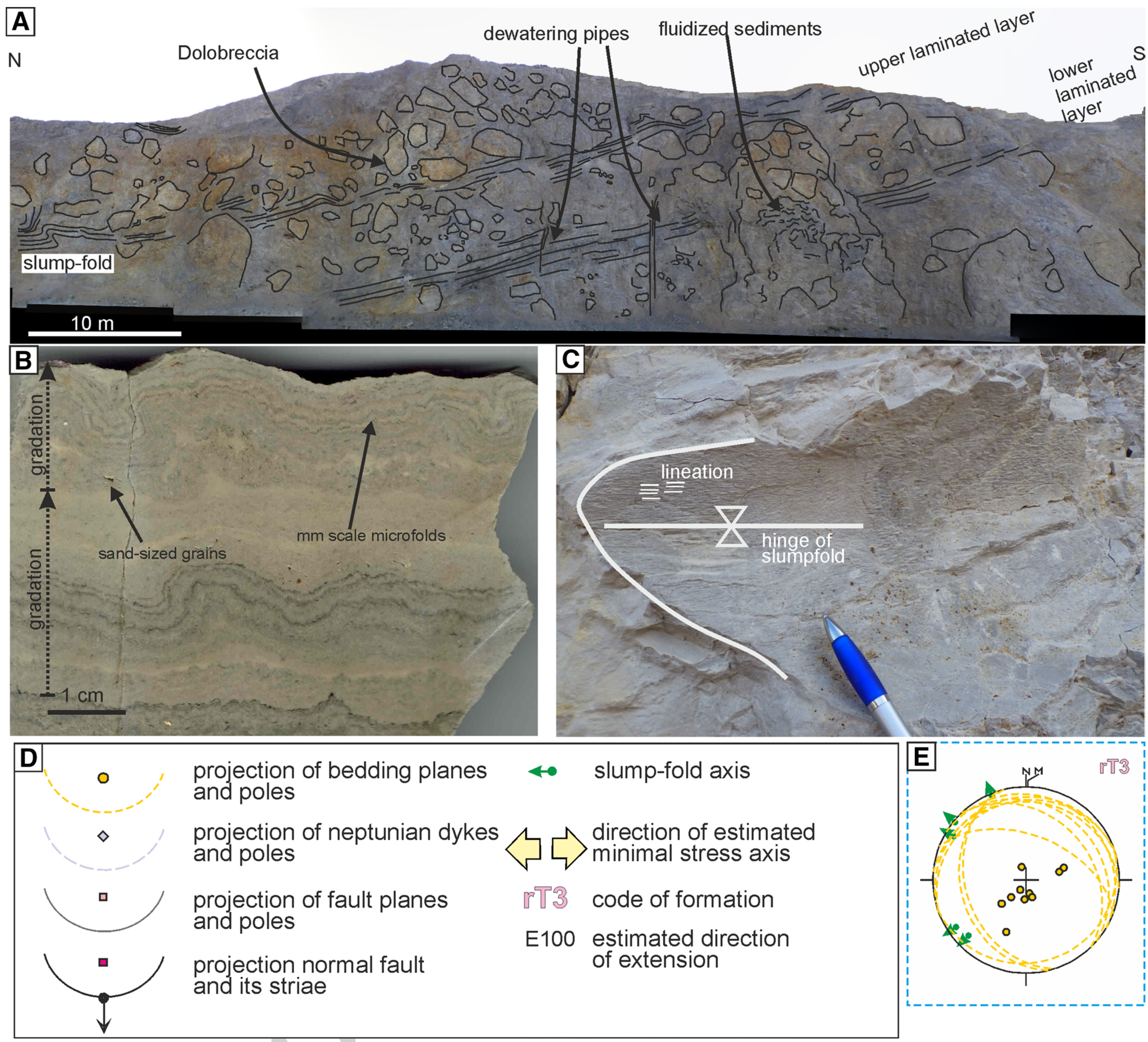

Fig. 2 Eastern wall of the Csókakő quarry (L1). **a** Dolomitic breccia and two laminated dolomite intercalations. Fluidized zones cross cut the dolomite breccia and layers; **b** cm-scale slump folds with mm-scale microfolds; **c** lineation is visible parallel to the slump-fold axis; **d** legend for the stereonets; **e** stereonet of the measured slump-fold axes and bedding

required for the calculation (Angelier 1990). Stress axes were estimated based on conjugate faults. Tilt test was carried out by the module of "Rotilt" (Angelier 1990), if the beds had a significant dip. The basic assumption was that tilting of the strata is mostly the result of Cretaceous folding. Pre-orogenic structures were back-tilted by the dip of the beds, to get a better view on the original geometry. Consequently, tilt test gave a relative chronology with respect to tilting/folding.

Seismic sections

The seismic sections were acquired and processed by GES Geophysical Services Ltd. in 2001 using vibroseis source with 8–90 Hz sweep frequency. Coverage of 100 and 12.5 m distance between CDPs ensured the proper lateral resolution and a good signal/noise ratio. These acquisition parameters represented an advanced technology that time. This facilitated a good image of basement structures. The processing was standard time processing including a challenging static correction due to hilly terrain, DMO correction, and post-stack migration. The featured sections were not time-depth converted and the vertical scales show two ways which travel time is second.

Field observations

Csókakő quarry (L1 on Fig. 1b)

Upper Norian Rezi Dolomite is exposed in the Csókakő quarry (L1 on Fig. 1b), which is covered by Upper Miocene (Pannonian) conglomerate and sand. The most spectacular part of the quarry is its eastern wall (Fig. 2a). Two main facies types of the Rezi Dolomite are visible here: thin-layered–laminated dolomite and dolomite breccia. 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 dolomite layers, further south. These strata dip already moderately toward NNE. Further south, the laminated dolomite intercalations pinch out, and in the southern edge of the quarry, only massive dolomite breccia is present.

The laminated unit is characterized by dark grey, strongly bituminous dolomite (Fig. 2a, b, c). Occasionally, sand-sized dolomite lithoclasts can be observed at the base of the laminated dolomite layers (Fig. 2b). These layers show normal gradation (Fig. 2b). Slide and slumps are common in the laminated dolomite: slumps occur mostly in the northern part, while slide scarps are more common in the middle and southern parts of the eastern wall. Slump folds shows symmetric to slightly asymmetric geometry; the axes of the slump folds show a significant dispersion, nevertheless, WNW trend is most frequent (Fig. 2e). Lineation was observed on the slump folds, which is parallel to the fold axis (Fig. 2b, c).

The clasts of the dolomite breccia are up to few meters in size (Fig. 2a). They are thick-bedded, white- or light-grey boulders, which often contain green algae, molluscs, and gastropods. Stromatolitic, intertidal dolomite represents another clast type of this dolomite breccia.

The matrix of the dolomite breccia is gradually changing south ward. In the northern, distal part of the breccia tongues, the matrix of this dolomite breccia is laminated dolomite. The laminated dolomite matrix is intensively deformed into chaotic folds between the re-deposited large dolomite blocks. In contrast of that, in the southern part of the breccia tongues, the matrix is made up by massive, light-grey dolomite, containing the same platform-originated fossils as the fossil-rich clasts of the dolomite breccia.

Two laminated dolomite intercalations are visible between the dolomite breccia tongues in the middle part of the eastern wall (Fig. 2a). The gently NNE-dipping beds are cross cut by several, few centimetre wide zones (Fig. 2a). The infill of these zones is small dolomite breccia clasts sitting in dolomite matrix.

The lower laminated dolomite layer is interrupted by a 5–10 m-wide sub-vertical collapsed zone, where chaotic, dark-grey dolomite is mixed with huge, light-grey dolomite blocks (Fig. 2a). The dark-grey dolomite probably originally represented the same material as the laminated one, but its original sedimentary features were destroyed (strongly deformed) later; therefore, the original lamination cannot be recognized there (Fig. 2a). The upper laminated dolomite intercalation is down bending and thickened above this zone (Fig. 2a).

Very similar, up to meter-sized dolomite blocks were observed in massive dolomite matrix at the Kőmell Cliff (L6 on Fig. 1b). Poor outcrop conditions does not permit detailed description (Fig. 1b).

Pilikán quarry (L2 on Fig. 1b)

Thinly bedded to laminated, dark-grey, bituminous dolomite (Rezi Fm.) crops out in the Pilikán quarry (L2). In the southeastern corner of the quarry, a 3 m-thick dolomite breccia intercalation was observed (Fig. 3a). The clasts are significantly smaller than those of the previous outcrops; their maximum size is just few dm (Fig. 3b). The contact of the breccia bed and the underlying dolomite is a wavy erosional surface. It is dissected by a number of normal faults (Fig. 3a). The offset of these faults decreasing upward, and finally, they are sealed by cover beds, without any flexure. In the upper part of the eastern wall meter-sized symmetric slump folds occur (Fig. 3a). The thickness variations along the limbs of the slump folds can be observed.

On the southern wall, small faults dissect a dark-grey marker bed, with a few cm offset; the overlying layers seal these structures (Fig. 3c). The NW–SE-trending faults show mostly normal offset; however, some of the faults are steep reverse faults (Fig. 3c, d).

Similar coarse-grained breccia and symmetric slump folds were observed in the southern Buda Hill quarry (L6 on Fig. 1b).

Gyenesdiás, eastern quarry (L4 on Fig. 1b)

ENE ward dipping beds are dominant in this dolomite quarry; therefore, a relatively thick-tilted succession is visible. In the western wall of the quarry thick beds of Hauptdolomit occur, whereas the southern wall exposes the Rezi Dolomite (Fig. 4a). The latter is thin-bedded, laminated, dark-grey bituminous dolomite, in which gentle NW–SE-trending symmetric slump folds were observed. Thick-bedded, light-grey dolomite intercalations occur upward, and

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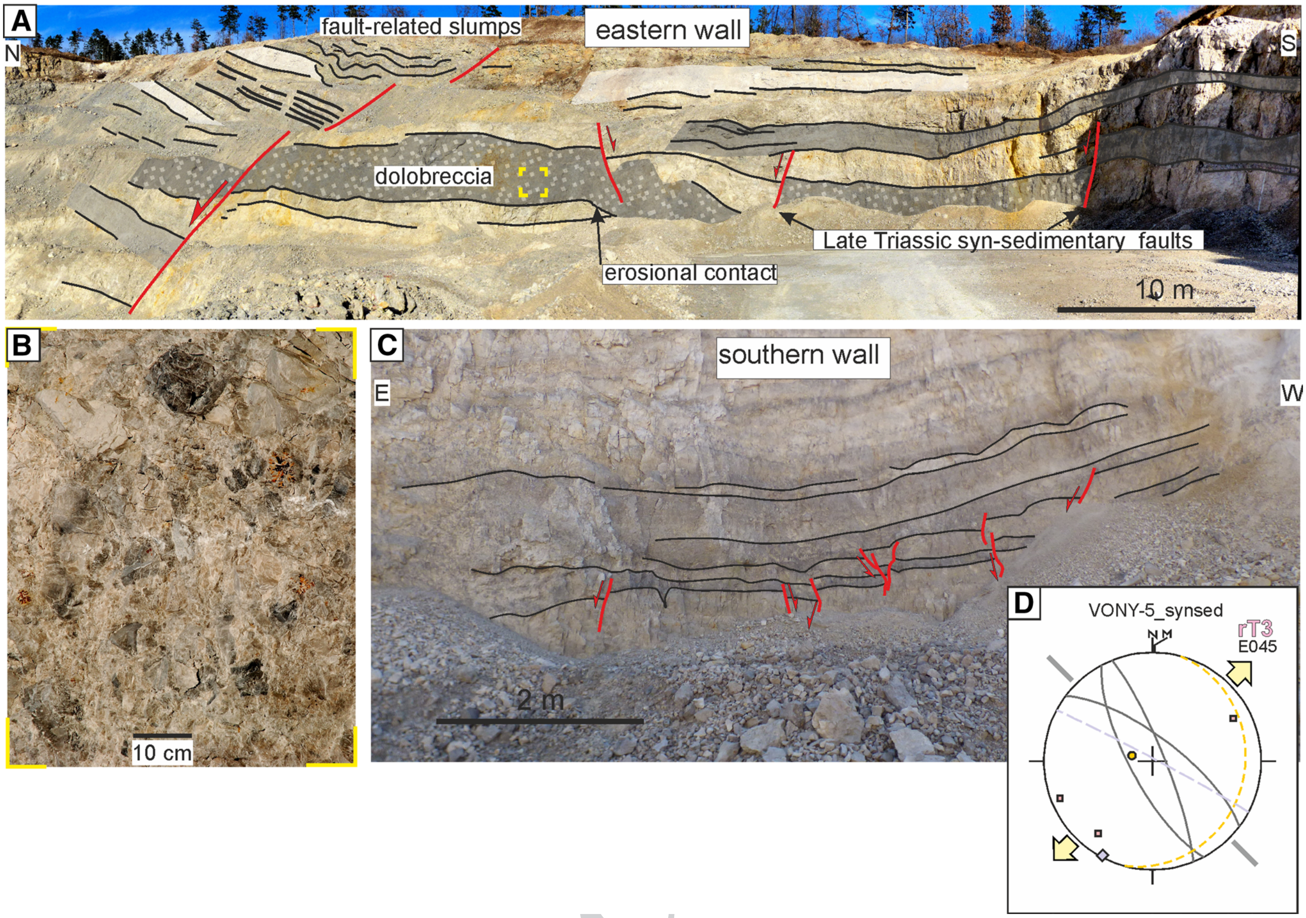

Fig. 3 Pilikán quarry (L2). **a** Dolobreccia intercalation of the Upper Norian Rezi F. on the eastern wall; **b** closer view on dolobreccia; **c** late Triassic synsedimentary faults on the southern wall of the quarry. For legend of the stereonet, see Fig. 2d

they dominate the eastern part of the southern wall, whereas, along the easternmost side, laminated dolomite is present, again. On the southern wall tilted, conjugate normal fault pairs were identified (Fig. 4a). The faults have only a few meter offsets. Back-tilted stereonet suggests NE–SW extension (Fig. 4.b, c).

Gyenesdiás, western quarry (L3 on Fig. 1b)

This outcrop is situated in the western vicinity of the previously described quarry (Fig. 1b). It is built up by thin-bedded, laminated dark-grey Rezi Dolomite. Meso-scale synsedimentary normal fault/slide was identified with a few tens of cm offset (Fig. 4d). The beds are thicker in the hanging wall, and the displacement decreases upwards. There is an upward smoothing extensional fault-related fold/flexure above the fault. It is dissected by minor normal faults (Fig. 4e). The discrete fault planes of these small-scale structures are not visible and only the small steps on the bedding planes indicate them. The faults suggest WNW–ESE extension (Fig. 4f).

Felső-hegy quarry (L5 on Fig. 1b)

This quarry exposes the Hauptdolomit Fm. (Fig. 1b). Thick-bedded, light-grey dolomite is the most common, but occasionally, a few cm thick, black, bituminous dolomite interbeds also occur locally. They contain small, angular clasts of light-grey dolomite. The succession is tilted to the NNE. On the western wall pre-tilt normal faults were observed (Fig. 4g). A dissected, bituminous, dark-grey interlayer has increased thickness in the hanging wall. A neptunian dyke running parallel to the fault is present in the footwall. It is filled by dark-grey, bituminous dolomite. These structures suggest NE–SW extension (Fig. 4h).

Seismic section in the western foreland of the Keszthely Hills

Two NE–SW-trending segments of 2D seismic sections are presented in this paper (Fig. 5a, b), which is situated in the northwestern foreland of the Keszthely Hills (Fig. 1b). The

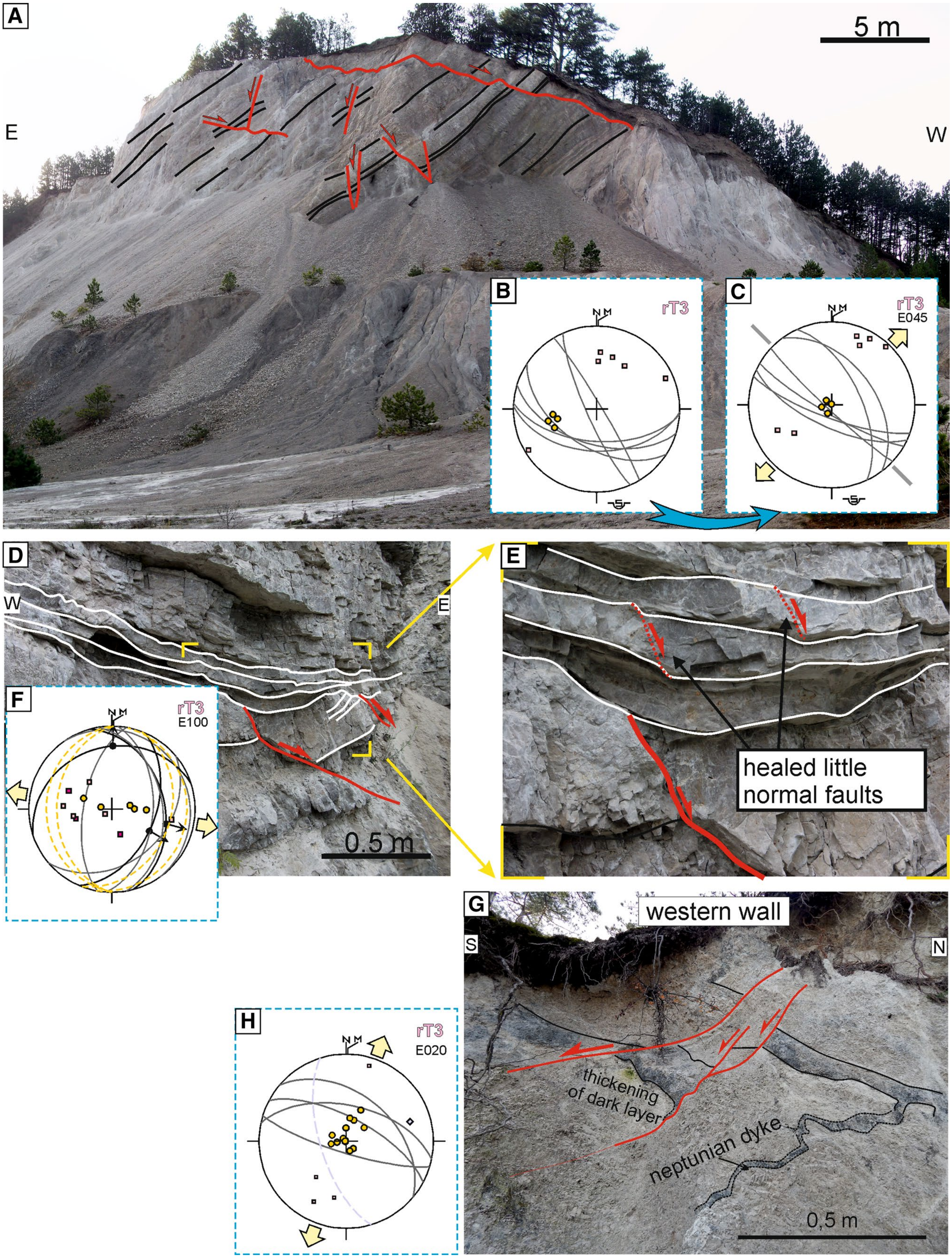

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Fig. 4 **a** Tilted normal faults in the eastern Gyenesdiás quarry (L3), observed in Rezi Dolomite. **b** Stereonet data before tilt test. **c** Stereonet of data after tilt test. **d** Gyenesdiás, western quarry (L4). Late Triassic synsedimentary normal fault or slide in the laminated to thin-layered Rezi Dolomite. **f** Measured fault-slip data **g** Late Triassic synsedimentary structures in the Felsőhegy quarry (L5), observed in Hauptdolomit. **h** Measured fault-slip data. Legend for stereonet: Fig. 2d

sections run parallel to each other, and they add important new information to the pre-orogenic structures.

Seismic facies of the formations

The pre-Senonian basement is represented by characteristic features (Figs. 1d, 5a, b). The relatively strong reflection package between 1.6 and 2.2 s TWT depth is equivalent to the Carnian Veszprém Marl. It is characterized by low-to-high amplitude; reflectors laterally often fade away (Fig. 1d), which may represent interfingering with coeval platforms (Ederics Fm.). Veszprém Marl was drilled by the nearby Kd-3 well under 3.4 km of Upper Triassic dolomite. Above this formation, a significantly thick unit without any strong reflections occurs, which is interpreted as the Hauptdolomit and the Rezi Dolomite, which are considered as one seismic unit in this paper. This seismic unit is represented by very low-to-low amplitudes; occasionally, short, pale reflectors occur (Fig. 1d). The thickness of these two dolomite formations is around 1 s in time, which suggests more than 3 km thickness, applying the VSP data of Dohr (1981). The strong continuous reflections above this unit represent the Kössen Marl. The medium-to-high amplitudes are related to the significant impedance contrast between the marl and the limestone intercalations (Fig. 1d). However, the strong reflections first fade away and then disappear approaching the faults. The thin, non-reflective unit above the Kössen Marl may be correlated to the Upper Rhaetian prograding tongue of the Dachstein and Kardosrét Fm. Consequently, the strong reflections of the Kössen Marl are sandwiched between two, relatively monotonous platform carbonates without reflections. The Jurassic–Early Cretaceous succession, which is made up by thin formations with variable lithologies, shows again relatively strong continuous reflectors on seismic sections, which can be characterized by low-to-high amplitude (Figs. 1d, 5a, b).

This Upper Triassic–Lower Cretaceous succession is unconformably overlain by Senonian shallow marine marl with limestone intercalations and platform limestone (Fig. 5a, b). The variable lithology of the Senonian marl causes again continuous reflectors with high amplitude (Fig. 5a), while the relatively monotonous platform limestone shows low-amplitude reflectors. On the section A–A′, the reflections of the Senonian marl onlap onto the basal surface of the Senonian (Fig. 5a). The Senonian deposits are unconformably overlain by Miocene succession, which was deposited in a prograding delta system. The related clinoforms are well visible, and dip apparently towards SW (Fig. 5a, b).

Structural geometry

The most prominent structure of the section A–A′ is an extensional graben, which is sealed by the Senonian deposits (Fig. 5a). This graben can be traced on the northeastern part of the section B–B′, but it is much narrower there (Fig. 5b). The graben has a segmented southwestern, NE-dipping boundary fault (Fault A) and a northeastern, SW-dipping boundary fault (Fault B). The graben is dissected by an additional NE-dipping fault (Fault C) creating two sub-grabens. These faults are post-dated by Senonian; however, Fault A and C show minor Senonian re-activation (Fig. 5a). On section B–B′, Fault C seems to be cut by a younger, probably Senonian fault (Fig. 5b). The Kössen Marl forms SW-ward thickening half-grabens above the gently SW-ward tilted blocks, which are pronounced on section A–A′ (Fig. 5a). On the section B–B′, the graben shows more symmetric geometry (Fig. 5b). In the southwestern sub-graben, off-lap surface within the Kössen Marl occurs (Fig. 5a). The contact between the Dachstein Limestone and the Kössen Marl is also an off-lap surface. In the vicinity of the major faults, the seismic image of the Kössen Marl shows poor quality, and in the hanging wall of the faults, wedge-shaped bodies are outlined (Fig. 5a, b). We interpret these bodies as fault-bounded talus breccia. The thickening trends of the Kössen Marl suggest that Fault A and C were dominantly active during its deposition (Fig. 5a, b). The Dachstein and Kardosrét Limestone are gradually thickening towards Fault B, which is well illustrated in section B–B′ (Fig. 5b). Only minor offset of these formations can be observed along the other two faults (Fault A and C). Therefore, the fault activity retreated onto Fault B during the deposition of the Dachstein and Kardosrét Limestone. If we restore the Senonian re-activation of Fault A and C, it seems that Jurassic strata sealed these faults. However, Jurassic deposits occur only in the hanging wall of Fault B, which suggests that the fault was active during or after deposition, but before the deposition of Senonian rocks.

On the southwestern part of the section B–B′, another pre-orogenic graben is enclosed by Fault D and Fault E (Fig. 5b). The Kössen, Dachstein, and Kardosrét Formations do not have any thickness changes related to these faults. On the other hand, Jurassic succession is thicker in the graben, and reflections in the Jurassic seals Fault D, that suggest Jurassic synsedimentary movement. Fault E is re-activated by post-orogenic extension probably during Senonian and Miocene; nevertheless, it shows significantly bigger pre-Senonian offset (Fig. 5b).

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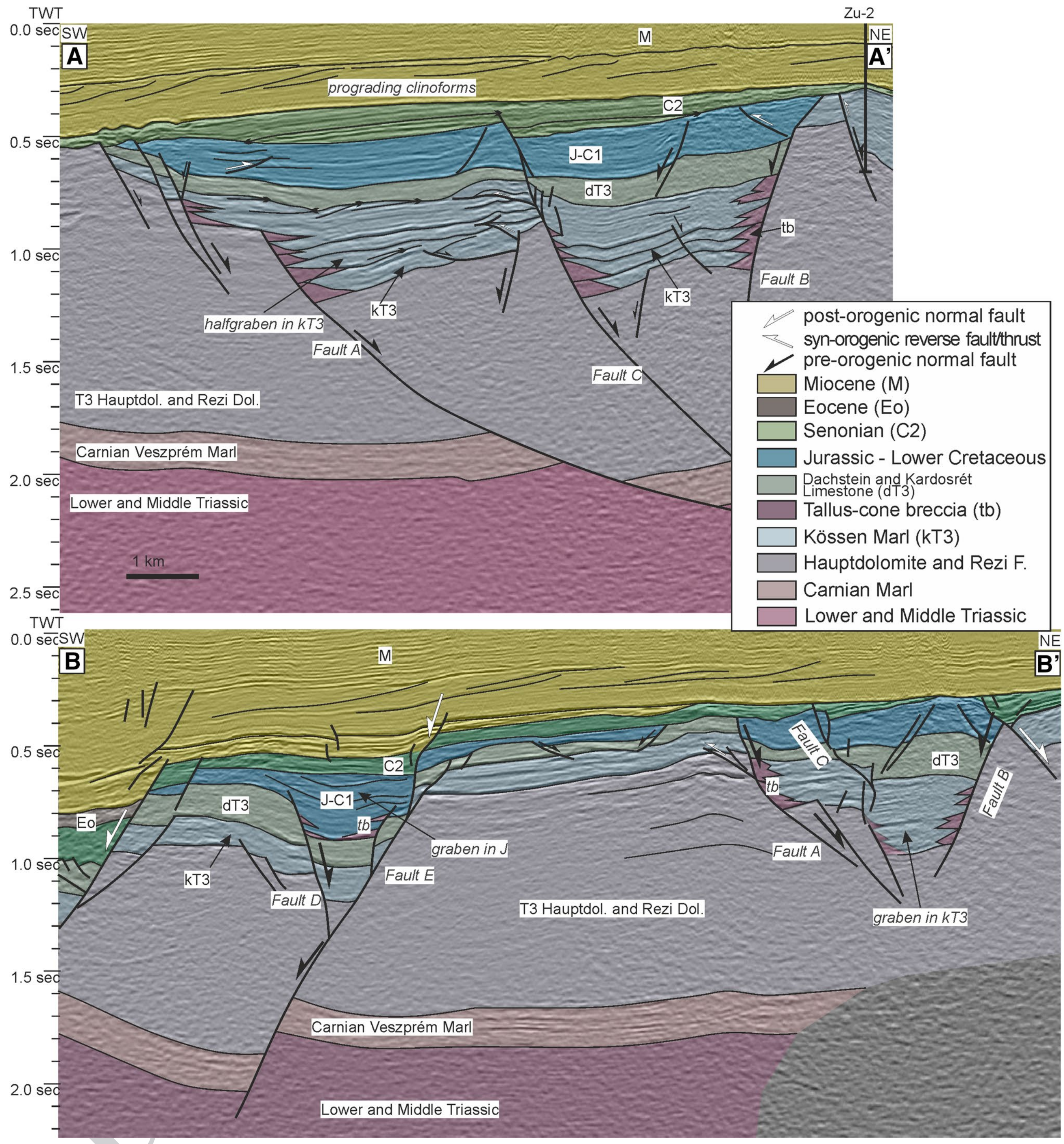

Fig. 5 **a** A–A′ and **b** B–B′ interpreted 2D seismic sections in the NW foreland of Keszthely Hills. Approximate location of the section is indicated in Figs. 1b and 6a. Blank version of the sections is visible in Supplementary Appendix I

Small normal faults are visible on the horst between Fault E and Fault A. These faults cross cut only the Dachstein and Kardosrét Limestone, and they probably detach on the Kössen Marl. These structures possibly represent mega-slides (Fig. 5b).

The sections are situated in the core of the "mid"-Cretaceous Sümeg-Devecser syncline (Tari 1994), and they are subparallel to the axis of syncline. Therefore, no major "mid"-Cretaceous contractional structures are visible on these sections. However, minor thrusts and related faults

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occur in the pre-Senonian rocks (Fig. 5a). It is interesting that these structures are localized by the early pre-orogenic faults. In the case of Fault A and B, antithetic small thrusts developed in the proximal hanging wall of the faults within the Jurassic succession. Above/near the Fault C on top of a small horst small double verging thrusts developed in the Kössen Marl, which made a gentle anticline in the Dachstein Limestone. These compressional structures formed due to moderate Cretaceous shortening.

The above-described extensional faults can be traced on other seismic sections, as well. On the bases of these 2D seismic lines, the strike of these structures is between WNW–ESE and NW–SE.

Interpretation of field observations

Coarse breccias along the Late Triassic Cserszegtomaj Fault

Map-view synsedimentary normal faults can be often outlined based on facies distribution and the presence of coarse 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 dolomite matrix, which suggests that these are sedimentary breccias. These breccias re-deposited on a most probably fault-controlled slope (Csillag et al. 1995). Such breccias could be alternatively formed after deposition, due to seismic shock of semi-unconsolidated mud (Hips et al. 2016). In the study area, dolomite breccia outcrops of the Rezi Formation are limited to an NW–SE-trending belt along the southwestern edge of the Keszthely Hills (Fig. 6c). Coarse breccias of the Csókakő quarry (L1) and the Kőmell cliff (L6) could be interpreted as proximal talus breccia (Fig. 2a), while the more fine-grained breccia intercalation in the Pilikán quarry (L2) could be interpreted as a more distal lobe of fault-related mass movements (Fig. 3b). Platform environment is suggested as the source of fossil-rich blocks in the Csókakő quarry (L1) (Csillag et al. 1995).

South of the dolomite breccia occurrence of Csókakő quarry (L1) Hauptdolomit is exposed (Figs. 6a, 7a). The WNW–ESE-trending contact of the two formations was identified already by former mapping (Bohn 1979; Budai

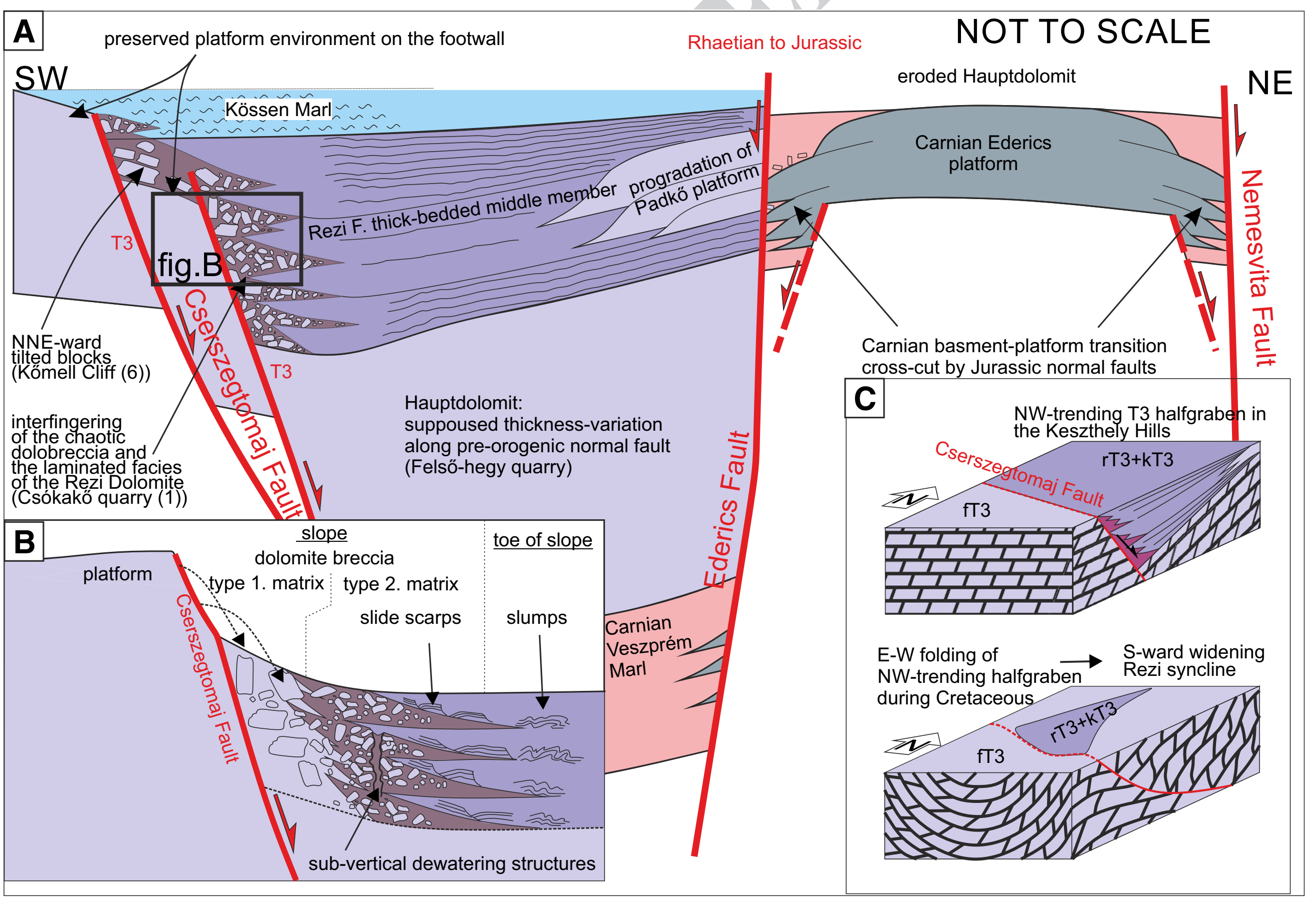
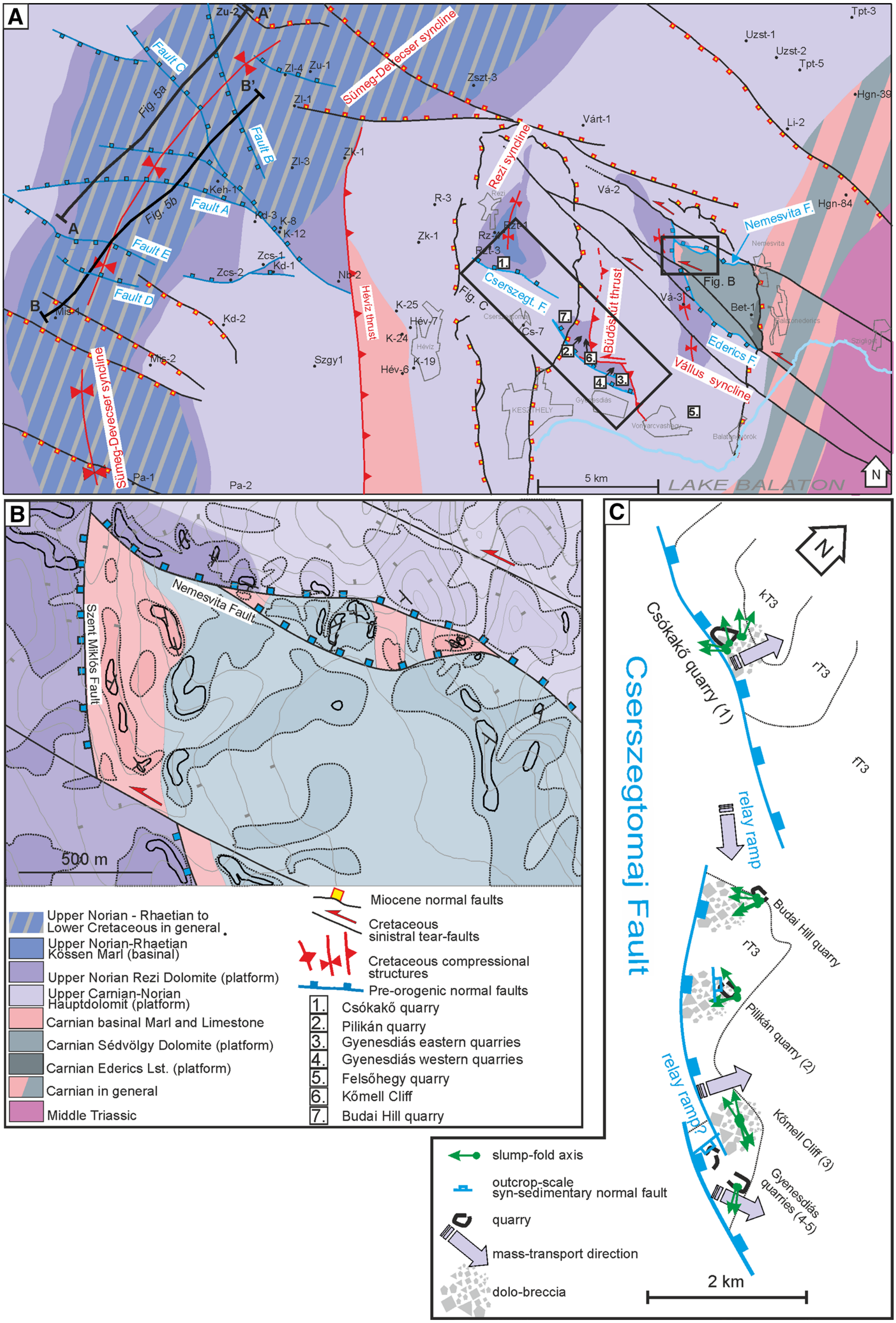


Fig. 6 **a** Pre-Miocene geologic map of the Keszthely Hills, and its NW foreland. **b** Detailed map of the Eastern Keszthely Hills. **c** Late Triassic synsedimentary structures along the Cserszegtomaj Fault



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◂**Fig. 7** **a** Model cross section across the Keszthely Hills, showing the pre-orogenic basin geometry. **b** Position of the Csókakő quarry in relationship with Cserszegtomaj fault. **c** Simplified cartoon explaining the formation of the southward widening Rezi syncline, as a folded Late Triassic half-graben

et al. 1999b); however, it was interpreted as a stratigraphic contact (Fig. 1b). In our interpretation, this contact represents a Late Triassic synsedimentary normal fault, which is referred as the Cserszegtomaj Fault in this paper (Fig. 6a). Probably, the southern WNW-trending border of the Rezi Dolomite occurrences near Gyenesdiás represents further NNE-dipping segments of Cserszegtomaj Fault, which was connected by ESE-dipping relay ramps (Fig. 6a, c).

Progradational tounge 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 suggests rather asymmetric half-graben geometry for the Late Norian basin of the Keszthely Hill (Fig. 7a). Note that the southward widening geometry of the N-trending Rezi syncline could be also explained by the Cretaceous folding of a WNW–ESE-trending Late Triassic half-graben (Fig. 7c).

The Cserszegtomaj Fault can be correlated with Fault A and C introduced on seismic sections. The activity of these NE–NNE-dipping faults is evidenced by main syntectonik deposits, which is represented by Rezi Dolomite in the Keszthely Hills (Fig. 7a, b), and Kössen Marl on the seismic section (Fig. 5a, b). Probably, all of these faults were active simultaneously, although the resolution of seismic sections does not allow the observation of synsedimentary deformation of the Rezi Dolomite. The lack of Kössen Marl outcrop in the Keszthely Hills made it problematic to observe its deformation. Nevertheless, the presence of slumps and sedimentary breccias in the Kössen Marl was documented based on wells in the Keszthely Hills (Haas 1993).

Rhaetian? To Jurassic extensional horst of the eastern Keszthely Hills

Reinterpretation of former geologic map (Budai et al. 1999b) suggests the presence of further map-view pre-orogenic structure in the Eastern Keszthely Hills. The easternmost part of the Keszthely Hills (Fig. 1b) is built up by Carnian formations (Csillag et al. 1995; Budai et al. 1999b). These formations partly dolomitized platform carbonates (Ederics Fm.) intercalating with the basinal Veszprém Marl (Csillag et al. 1995; Budai et al. 1999b; Haas et al. 2014). The Carnian formations have tectonic contact with the Rezi Dolomite and the Hauptdolomit. The fault system bounding Carnian formations has a northern WNW–ESE-trending segment (Nemesvita Fault), a western N–S trending segment (Szent Miklós Fault), and a southern NW–SE-trending segment (Ederics Fault) (Fig. 6a, b). The whole area is dominated by western dips, which formed during the "mid-Cretaceous" E–W shortening. There are areas (e.g., along the Szent Miklós Fault), where the Carnian Veszprém Marl is in direct contact with the Upper Norian Rezi F., and thus, the whole Hauptdolomit, which is more than 1 km thick, is tectonically omitted.

The Szent Miklós Fault is sub-vertical based on the vertical electric sounding of Gulyás (1991). That is why, it was interpreted by Dudko (1996) as a syn-orogenic, syn-folding strike–slip fault. In our interpretation, the large (km-scale) vertical displacement can be explained rather by normal or oblique-slip faulting (Fig. 6a, b). The actual sub-vertical dip of the fault (Gulyás 1991) can be the result of later, moderate tilting, associated with syn-orogenic Cretaceous folding, which steepened, but not overturned the original west-dipping fault.

It is clear from map view that the Szent Miklós Fault is dissected by NW–SE-trending sinistral faults with few 100 m of offset (Fig. 6a, b) (Budai et al. 1999b; Dudko 1996). These sinistral faults can be considered as syn-folding tear faults, since they have significant offset on the eastern "mid-Cretaceous" syncline (Vállus syncline); on the other hand, they die out towards northwest, and do not crosscut the western syncline (Rezi syncline). These sinistral faults were probably re-activated during the Late-Oligocene–Early Miocene, when a very similar stress field was present (Fodor et al. 1999). These sinistral faults also prove that the Szent Miklós Fault is an older, pre-orogenic fault, which was overprinted by the structures of Cretaceous compression. On the other hand, no coarse breccia was observed along the Szent Miklós Fault, which may suggest that it is younger than Rezi Dolomite.

Although there are no data on the age of Nemesvita and Ederics Fault, we suggest that these faults are coeval with the Szent Miklós Fault, and they represent a pre-orogenic extensional horst. The Ederics Fault shows many similarities to Fault B, which is slightly younger than Fault A and C. Fault B was moderately active during the deposition of Kössen Marl, but it was still active later, during the deposition of Dachstein and Kardosrét Limestone when the southwestern boundary faults (Fault A and C) were inactive (Fig. 5a). The presence of Jurassic deposits in the hanging wall and the Senonian seal suggests that Fault B was slightly active during the Jurassic, too. The same situation is suggested for the pre-orogenic horst of the eastern Keszthely Hills (Fig. 7a).

Pattern of outcrop-scale pre-orogenic normal faults

Many of the described faults in the Keszthely Hills can be interpreted as synsedimentary Late Triassic structures, based on several features. Such features are thickness variations of the beds along faults (Fig. 3c, 4d, e, g), the presence of

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wedge-shaped syntectonic beds (Fig. 4d, e), or faults sealed by younger beds of Late Triassic succession (Fig. 3a, c). In the Pilikán quarry (L2), small synsedimentary reverse faults occur besides normal faults. Certainly, all of these faults formed in the same extensional stress field, and the reverse faults formed due to space problems related to the movement along a non-planar normal fault plane.

Small steps of the beds in the western Gyenesdiás quarry (L4) (Fig. 4e) can be interpreted as healed normal faults, where the discrete faults disappeared due to diagenetic processes. These structures can be considered as pre-diagenetic faults. Tilted normal faults of the eastern Gyenesdiás quarry (L3) represent structures which are postdate deposition and diagenesis; on the other hand, they developed before tilting/folding, which age is Early Albian based on projection of structural data from the central Transdanubian Range (Fodor et al. 2017).

The strike of outcrop-scale synsedimentary normal faults is in accordance with map-scale pattern (Fig. 6a), since most of these structures shows NNE–SSW or NE–SW extension (Figs. 3d, 4c, h,). This direction of extension is in accordance with the trend of other pre-orogenic normal faults, described in the central and northeastern Transdanubian Range. The ages of such structures are Middle Triassic (Budai and Vörös 2006) or Early and Middle Jurassic (Vörös and Galácz 1998; Lantos 1997; Fodor 2008). Perpendicular, WNW–ESE extension (Fig. 4f) was estimated based on the fault/slide of the western Gyenesdiás quarry (L4), which is situated most probably on a relay ramp which connects two segments of the Cserszegtomaj Faults (Fig. 6a).

Pattern of slumps and slides

The presence of slumps and slides is widespread in the laminated Rezi Dolomite. In the Csókakő quarry (L1), an extensional and a compressional domain can be separated, similar to many case studies (e.g., Farrell 1984; Debacker et al. 2009; Alsop and Marco 2011). We suggest that the NNE-ward-dipping beds of the southern part of the quarry represent the original dip of the slope (see next chapter). Slide scarps are present mostly in this part of the quarry (extensional domain). The northern side of the quarry, which can be characterized by horizontal dips, is dominated mostly by slumps. This part of the quarry situated on the toe of the slope where compressional domain developed.

The strike of slide scarps and slump-fold axes may allow to determine the sedimentary transport direction. The strikes of slide scarps are theoretically parallel to the strike of the slope. On the other hand, slump-fold axes can suffer notable rotation, after a significant transport (Alsop and Marco 2011). During the early stage of slump formation, the slump fold shows symmetric geometry. In that stage, the axis of the slump is perpendicular to the dip direction of the slope (Bradley and Hanson 1998). Slumps observed in Rezi Dolomite show symmetric or slightly asymmetric geometry suggesting minor transport (e.g., Fig. 2b). Therefore, the transport direction is supposed to be sub-perpendicular to the fold axis. While most of the observed slump axes are NW–SE directed (Fig. 6c), the gravity slide transport direction is toward NE in the hanging wall of the fault segments [Csókakő quarry (L1), Kőmell Cliff (L6)] (Fig. 6c). On the other hand, the slump-fold axes pronouncedly different on relay ramps (NE-trending) which may suggest SE-ward mass transport [Budai Hill quarry (L7), Gyenesdiás quarries (L3) and (L4)] (Fig. 6c).

Slump folds often show features resembling metamorphic ductile structures; for example, the presence of stretching lineation is common in the case of soft-sediment deformation (Ortner 2013). However, in our case, a completely different type of lineation was observed. On the polished surface of samples, it is visible that this lineation derives from fold hinges of microfolds (Fig. 2b, c).

Dewatering structures

An episode of talus-cone breccia formation probably provided considerable volume of sediments. The sudden load made the underlying unconsolidated thin-layered carbonate mud compacted, and de-watered. The chaotic zone of the Csókakő quarry (L1) may indicate dewatering and fluidisation of originally laminated sediments, similar to examples of Ortner (2007). Sediment fluidization occurred where the original sedimentary features of the laminated dolomite were completely destroyed (Knipe 1986; Ortner 2007). The dewatering related compaction could be responsible for the collapse and subsidence (down bending) of overlying beds (upper laminated layer). The vertically arranged few cm zones can be interpreted as dewatering pipes, where water was released from a deeper beds (Fig. 2a). The tilted beds of the Csókakő quarry (L1) are dissected by sub-vertical dewatering pipes and fluidized zone (Figs. 2a, 7b). It confirms that the tilting in the Csókakő quarry (L1) pre-dates diagenetic processes, such as dewatering, and the tilted strata there represent the original dip of the tectonically controlled slope.

The style of the above-mentioned early deformation was probably highly influenced by the early diagenetic process such as dolomitization (Meister et al. 2013). Platform-originated re-deposited blocks observed in the Csókakő quarry (L1) (Fig. 2a) probably underwent early dolomitization, as well (Haas et al. 2012), which redound the “brittle” re-deposition, represented by blocks. On the other hand, the laminated Rezi Dolomite—which deposited in deeper marine environment—probably dolomitized later, during burial; therefore, dolomitization did not obstruct soft-sediment deformation, such as slumping or soft-sediment fluidization.

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Discussion: regional outlook

Comparison with other Alpine basins

A newly defined Late Triassic extensional graben system was identified in this paper, which can be correlated to other areas of the Alpine region. Similar Late Triassic extensional back-platform basins are known from the western Southern Alps (Lombardy) and from the Bajuvaric nappe system of the Northern Calcareous Alps. 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 of 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 as Kössen Basin in this paper (Fig. 8a). The correlation of the related succession was the topic of several publications (Haas et al. 1995; Gale et al. 2015; Rožič et al. 2009); therefore, in this paper, we compare these basins from a structural point of view.

The geometry of the Late Triassic extensional basins is well reconstructed in the Lombardy; notwithstanding, it is strongly overprinted by southvergent Cenozoic thrusts (Bertotti et al. 1993; Carminati et al. 2010; Jadoul et al. 2005). Approximately 10 km wide, N–S trending horsts and grabens were formed there, which show a similar geometry like the Late Triassic grabens of the southwestern Transdanubian Range (Bertotti et al. 1993). Bally et al. (1981) documented listric geometry for some of these faults, similar to Fault A in the present study. According to Bertotti et al. (1993), some of the Norian–Rhaetian faults in the Lombardian region were active during the Jurassic, which is also a common feature, compared to our observations. Back-rotating the units with the Mesozoic paleomagnetic data (Fig. 1a; references therein), the pre-orogenic normal faults of Lombardy and the study area have similar N–S strike (Fig. 8a).

The pre-orogenic basins in the Northern Calcareous Alps were strongly overprinted by Cretaceous and Cenozoic nappe stacking. Therefore, the Late Triassic basin

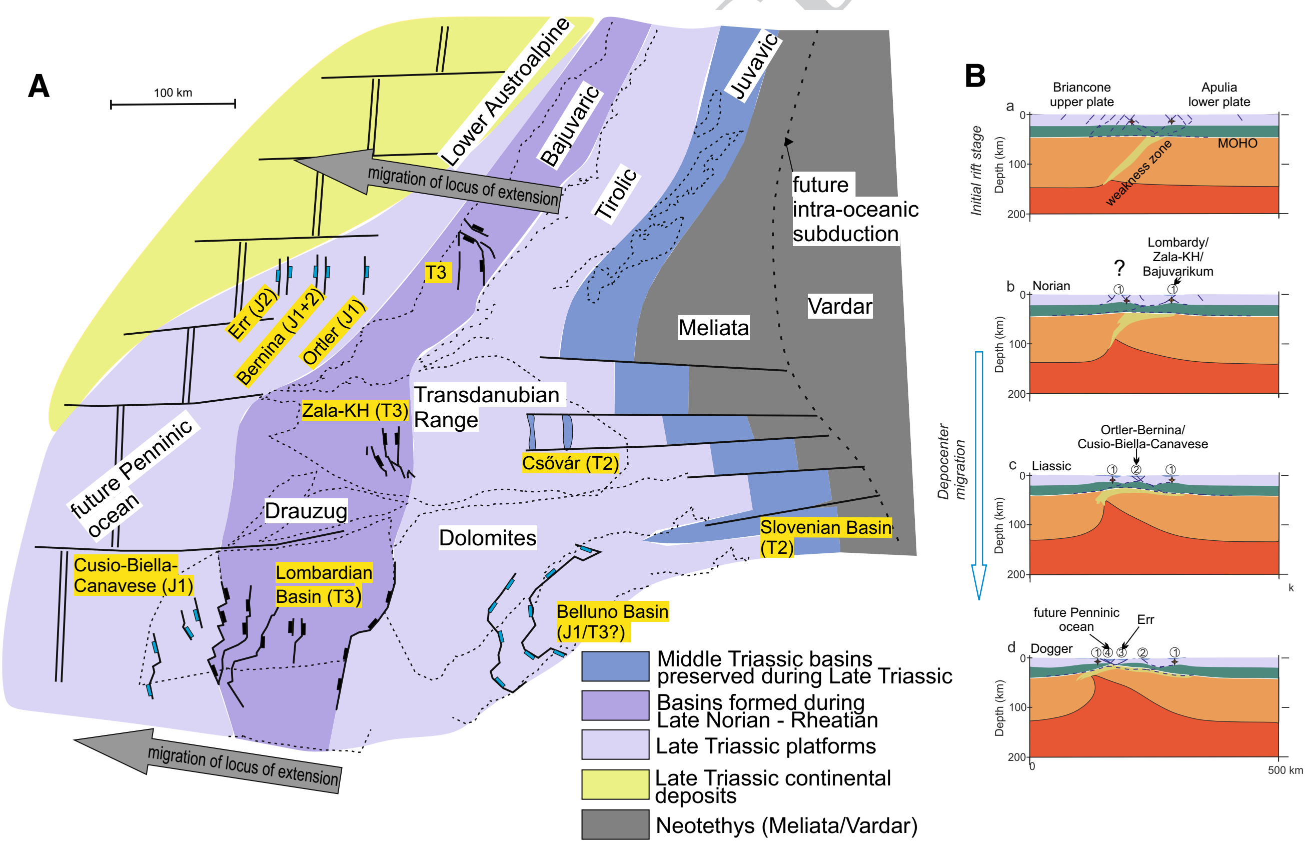

Fig. 8 **a** Late Triassic paleogeographic reconstruction of the Transdanubian Range and neighbour units based on Haas et al. (1995), Froitzneim and Manatschal (1996), Mandl (2000), and Gawlick et al. (1999). Fault patterns are adopted from Cazzini et al. (2015) and Behrmann and Tanner (2006). The faults were back rotated based on the magnetization direction of Fig. 1a (see references there). **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)

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geometry has not been reconstructed, yet from the structural point of view. However, several authors propose the presence of Norian extensional deformation based on facies distribution and other sedimentological evidences (Satterley and Brandner 1995; Gawlick and Missoni 2013). An exception is the work of Behrmann and Tanner (2006), which reported the thickness variation of the Hauptdolomit along N–S and NW–SE-trending faults based on restored cross-sections. This also confirms that a significant deformation initiated already during the deposition of the Hauptdolomit. Considering paleomagnetic data (Fig. 1a; references therein), the NW–SE-trending segment also shows N–S paleo-trend, like in the study area (Fig. 8a).

Most of the Anisian basins, which were formed due to opening of the Neotethys, were filled up during the Carnian. However, in those basins, which were situated along the distal Adriatic passive margin, deep-water sedimentation was continuous during Late Triassic. Probably, Norian–Rhaetian extension contributed to the preservation of these basins. Such basins are the Hallstatt basin of the Northern Calcareous Alps (Lein 1985; Gawlick and Böhm 2000), the Csővár and Mátyáshegy basin in the NE Transdanubian Range (Haas et al. 2010), and the Slovenian basin in the Southern Alps (Goričan 2012; Gale et al. 2015; Celarc et al. 2013; Oprčkal et al. 2012). According to Missoni et al. (2008), the Slovenian basin was controlled by strike–slip movement during Late Triassic times (Fig. 8a).

Geodynamic implications

Geodynamic background of Norian deformation of the Adriatic plate is still under debate. Bertotti et al. (1993) considered this deformation as the first sign of Alpine Tethys rifting. According to Cozzi (2000), Norian faults of the Southern Alps can be related rather to the opening of the Neotethys. Based on the recent works, continental rifting of the Alpine Tethys started just during Early Jurassic (Froitzneim and Manatschal 1996; Berra et al. 2009; Decarlis et al. 2015).

On the basis of the Triassic evolution of the Transdanubian Range, the Anisian extension, related to the opening of the Neotethys (Vörös and Budai 2006), can be clearly separated from Norian extension. Our results show that the main syn-rift sediments (Rezi and Kössen Fm.) are Late Triassic in age in the study area, and extension was continuous in the Jurassic. This observation may link the formation of these basins rather to the continental rifting of the Alpine Tethys. It suggests that the rifting should have started on the proximal Adriatic margin even during Norian (Bertotti et al. 1993).

Asymmetry of the Alpine Tethys rift

The Late Triassic Kössen basin was situated significantly to the east of the future Alpine Tethys, on the proximal Adriatic passive margin (Fig. 8a). During Jurassic, westward migration of extensional tectonism was pointed out in the case of Austroalpine nappes (Froitzneim and Manatschal 1996). The proximal Adriatic margin was subject of dominantly Hettangian–Sinemurian extension, whereas, in the distal Adriatic margin, Pliensbachian–Callovian extension occurred (Fig. 8a). 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).

Nevertheless, Jurassic normal faults and grabens are known east of the Lombardian basin. In the Belluno Basin, deep-water sedimentation and facies differentiation started just during the Early Jurassic. However, thickness changes suggest that a significant extension initiated also in the Belluno Basin during the Norian, but, in contrast with Lombardian Basin, the sedimentation could keep pace with extension-related subsidence (Masetti et al. 2012).

Most authors agree that the opening of the Piemont–Ligurian Ocean is the result of asymmetric rifting, where the Adriatic plate represents the lower plate, while the European plate is the upper plate (Froitzneim and Manatschal 1996). Alternatively, it is also possible that the rift system changed polarity along a major transform fault, such as paleo-Periadriatic Fault (Decarlis et al. 2017). According to Lavier and Manatschal (2006) and Decarlis et al. (2017), the rift system became asymmetric only after necking of the lithosphere, when the residual crust did not contain any ductile level.

According to our opinion, the rifting of the Alpine Tethys was the initial asymmetric, since we connect the formation of the Late Triassic basins to the initial continental extension within the upper crust. Thus, westward migration of extensional deformation started during Late Triassic (Fig. 8b). This feature is asymmetric, while such processes are not present on the adjacent European margin. According to a numerical model of Balázs et al. (2017), the development of initially asymmetric rift zones can be triggered by inherited weakness zones (e.g., inherited suture). In the present case, the role of Variscan orogeny or its Permian collapse can arise (Manatschal et al. 2015). On the other hand, this relationship needs further investigation, since the built up of Variscan orogeny has been poorly reconstructed yet in the study and neighbour areas, due to strong Alpine overprint.

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Conclusions

Late Triassic and Jurassic map-scale normal faults were defined in the southeastern part of the Transdanubian Range, based on field observation and 2D seismic data. The faults show NW–SE-to-WNW–ESE trend. The extension was active already during the deposition of the Hauptdolomit (Norian). Initially, the sedimentation was able to keep pace with the extension-derived subsidence, and thus, platform environment was present both in the footwall and hanging wall. At the end of Middle Norian, extension-related intraplatform basin of the Rezi Dolomite and the Rheatian Kössen Marl formed. The extensional deformation was active during the Late Rhaetian progradation of the Dachstein Limestone, and through the Lower-to-Middle Jurassic. Consequently, we propose continuous extension during the Late Triassic and Early–Middle Jurassic.

In the Keszthely Hills, the late Middle Norian–Upper Norian Rezi Dolomite is proved to be the main syntectonic deposit. The synsedimentary faulting was associated with the development of slides and slumps, and the formation of fault-bounded talus-cone breccia.

Other type of pre-orogenic extensional faults post-dates the deposition of Rezi Dolomite, but pre-dates Albian folding. These structures formed probably during the Jurassic. The Szent Miklós Fault represents one map-view example of these structures in the Keszthely Hills.

On 2D seismic section, normal faulting proved to be coeval with the deposition of Kössen Marl and Dachstein Limestone. The main reasons are thickness variations due to normal faulting, and the presence of talus-cone breccia. Jurassic activity of fault B was also proposed in this case.

The Late Triassic extension was the first sign of continental rifting of Alpine Tethys, which represents an initially asymmetric rift, where at least the northern part of the Adriatic plate was in lower plate position. The Lombardian Basin–Zala Basin–future Bajuvaric nappe system was the first locus of rift-related extension on the proximal Adriatic magin. Later on, during the Early and Middle Jurassic, the axis of extensional deformation was migrated westward, towards the future Alpine Tethys.

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