AGU PUBLICATIONS

Tectonics

RESEARCH ARTICLE
10.1002/2015TC004109

Key Points:
- Coupled tectonosedimentary evolution of the Pannonian Basin
- Kinematics of extension in back-arc settings
- Migration of tectonic phases

Correspondence to:
A. Balázs,
a.balazs@uu.nl

Citation:
Balázs, A., L. Matenco, I. Magyar, F. Horváth, and S. Cloetingh (2016), The link between tectonics and sedimentation in back-arc basins: New genetic constraints from the analysis of the Pannonian Basin, *Tectonics*, *35*, 1526–1559, doi:10.1002/2015TC004109.

Received 18 DEC 2015
Accepted 5 JUN 2016
Accepted article online 13 JUN 2016
Published online 30 JUN 2016

The link between tectonics and sedimentation in back-arc basins: New genetic constraints from the analysis of the Pannonian Basin

A. Balázs[1,2], L. Matenco[2], I. Magyar[3,4], F. Horváth[1,5], and S. Cloetingh[2]

[1]Department of Geophysics and Space Science, Eötvös Loránd University, Budapest, Hungary, [2]Netherlands Research Centre for Integrated Solid Earth Science, Faculty of Geosciences, Utrecht University, Utrecht, Netherlands, [3]MOL Hungarian Oil and Gas Plc., Budapest, Hungary, [4]MTA-MTM-ELTE Research Group for Paleontology, Budapest, Hungary, [5]Geomega Ltd., Budapest, Hungary

Abstract The architecture of sedimentary basins reflects the relationship between accommodation space and sediment supply, their rates and localization being variable during basin evolution. The mechanisms driving the interplay between tectonics and sedimentation in extensional back-arc basins overlying rheological weak zones inherited from an earlier orogenic evolution are less understood. A typical example is the Pannonian back-arc basin of Central Europe. It is floored by continental lithosphere and was affected by large amounts of extension driven by the subduction rollback that took place in the Carpathians and/or Dinarides. A novel kinematic and seismic sequence stratigraphic interpretation calibrated by wells allows the quantification of the link between the formation of half grabens and coeval sedimentation in the Great Hungarian Plain part of the basin. While the lower order tectonic-induced cycles characterize the main phases of extension in various subbasins, the higher-order cyclicity and associated unconformities define individual moments of fault (re)activation. Our novel interpretation of a temporal and spatial migration of extension during Miocene times explains the contrasting present-day strike of various subbasins as a result of their gradual clockwise rotation. Incorporating the observed asymmetry, in particular the associated footwall exhumation, infers that the amount of extension is much larger than previously thought. The quantitative link between tectonics and sedimentation has allowed the definition of a novel model of sedimentation in asymmetric basins that can be ported to other natural scenarios of similarly hyperextended back-arc basins observed elsewhere.

1. Introduction

The architecture of sedimentary basins reflects the relationship between accommodation space and sediment supply, their rates and localization being variable during basin evolution [e.g., *Cloetingh and Haq*, 2015; *Schlager*, 1993]. The link between tectonics and associated surface processes in terms of erosion and sedimentation has been recognized as the critical feedback interaction influencing the final geometry of sedimentary basins, in particular relevant in extensional settings [e.g., *Burov and Guillou-Frottier*, 2005; *Burov and Poliakov*, 2003]. Extensional back-arc basins, floored by oceanic or continental lithosphere, develop in the hinterland of orogenic arcs when the rate of subduction is higher than the convergence velocity [e.g., *Dewey*, 1980; *Royden and Burchfiel*, 1989; *Uyeda and Kanamori*, 1979]. Although their position behind a magmatic arc is not always very clear, a large number of back-arc basins were defined in the Mediterranean region, formed during Oligocene-Miocene times in response to the subduction retreat of the Aegean, Gibraltar, Calabrian, or Vrancea slabs, in the hinterland of the highly arcuate Hellenides, Rif-Betics, Apennines, and Carpathians orogens, respectively [*Faccenna et al.*, 2005; *Horváth et al.*, 2015; *Jolivet and Brun*, 2010; *Vergés and Fernàndez*, 2012; *Wortel and Spakman*, 2000]. In all these situations extension postdates at relatively short times the contraction and is juxtaposed over an inherited nappe stack, often reactivating thrust contacts and exhuming rocks previously buried, such as in the Apennines or Aegean system [e.g., *Brun and Faccenna*, 2008]. Such reactivations are recognized along major detachments, locally associated with the formation of large core complexes or other extensional domes, or low-angle normal faults, as observed in the Rhodope or the Alboran domain [e.g., *Brun and Sokoutis*, 2007; *Vissers*, 2012]. These structures controlled also the evolution of their hanging wall half grabens grouped along large extensional basins with amounts of cumulated extension in the order of hundreds of kilometers, such as in the Aegean or the Alboran Domain [*Comas et al.*, 1992; *van Hinsbergen et al.*, 2005]. In other geodynamic settings, such large amounts of extension characterize hyperextended passive continental margins

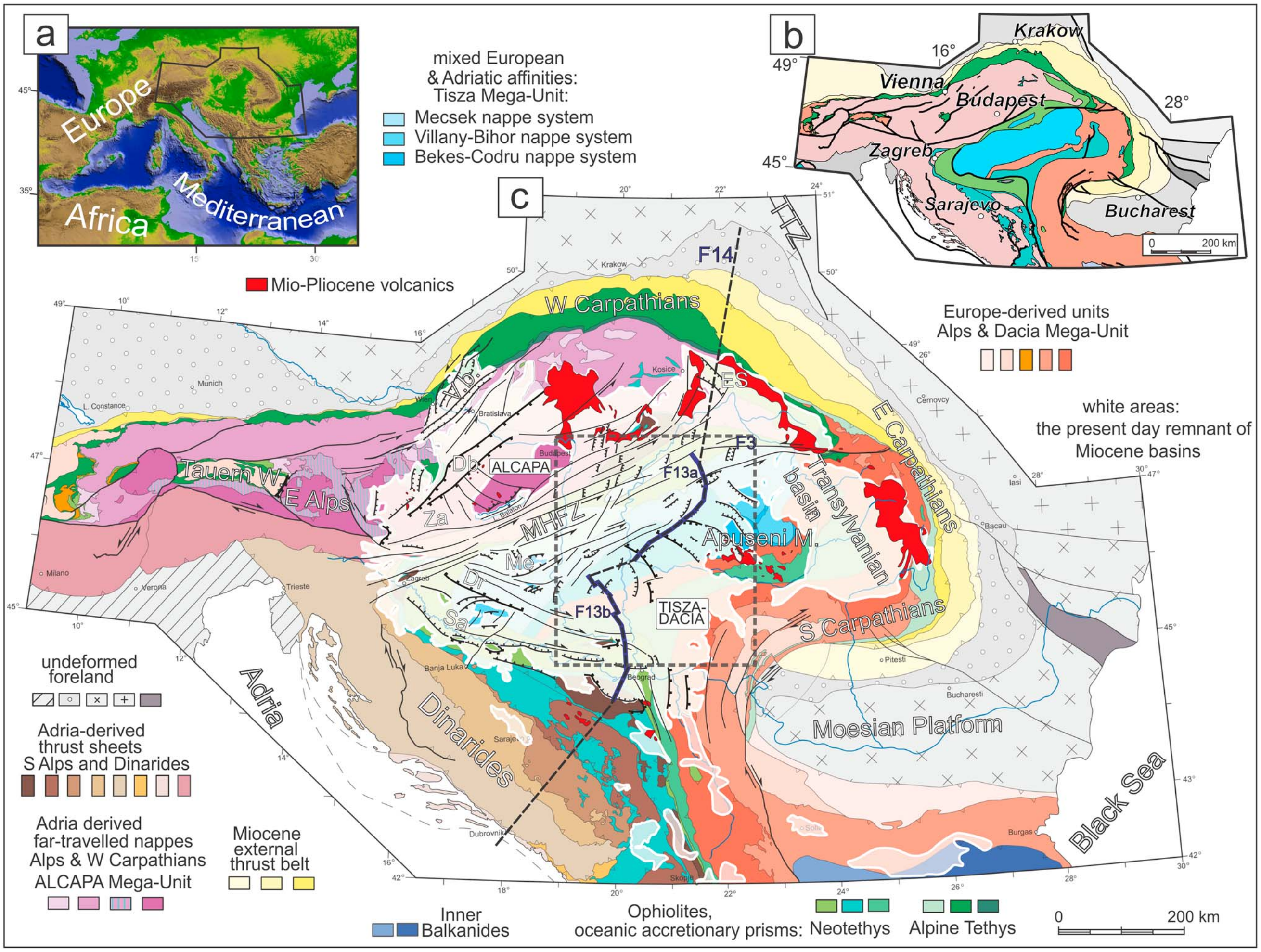

Figure 1. (a) Location of the Pannonian Basin system. (b) Main tectonic units of the Alps-Carpathians and Dinarides [after *Schmid et al.*, 2008]. The Tisza-Dacia contact is slightly modified. Note that the overlying Neogene basins in the intra-Carpathians area are ignored. (c) Miocene–Quaternary tectonic map of the Pannonian Basin and the Alps-Carpathians-Dinarides system showing the present-day extent of the Neogene sediment cover of the Pannonian, Vienna, and Transylvanian basins overlying the pre-Neogene structures and showing the major Miocene to Quaternary faults (modified after *Horváth et al.* [2015], *Schmid et al.* [2008], *Ustaszewski et al.* [2014], and the results of this study). Note that the present-day geometry of these basins does not reflect their original position at the time of formation; some Neogene deposits were eroded in the Pannonian Basin or underthrust beneath the Carpathian nappe stack. The TTZ has served as a rheological contrast zone localizing Miocene tectonics of the Carpathians [e.g., *Matenco et al.*, 2016]. Vb = Vienna Basin, Dr = Dráva subbasin, Sa = Sava subbasin, Za = Zala subbasin, Me = Mecsek hill, Db = Danube basin, ES = East Slovakian basin, TTZ = Teisseyre-Tornquist Zone, MHFZ = Mid-Hungarian Fault Zone.

[e.g., *Tugend et al.*, 2014]. Such definitions are less understood in the case of extensional back-arc basins floored by continental lithosphere characterized by large amounts of extension [e.g., *Huismans and Beaumont*, 2003], in particular when analyzing the link between tectonics and sedimentation. Symmetry or asymmetry in extensional basin is used in many different ways. We use the term of asymmetric extension in a nongenetic way, i.e., for any extensional (sub)basin that shows one major structure, normal fault, or detachment, controlling the coeval sedimentation in the hanging wall and at least comparable uplift of its footwall [*Wernicke*, 1985]. Such a simple definition is independent of the controlling mechanics, from simple shear in single or multilayered lithosphere to complex multistage hyperextended basins [*Huismans and Beaumont*, 2003; *Manatschal et al.*, 2015].

A typical example is the Pannonian Basin of Central Europe (Figure 1), where the large amount of extension is accommodated by the rapid rollback of the Carpathians slab between 20 and 9 Ma [e.g., *Csontos*, 1995; *Fodor et al.*, 1999; *Horváth et al.*, 2006; *Matenco and Radivojević*, 2012; *Merten et al.*, 2010]. The observation of a thin synrift and thick postrift basin fill is usually interpreted as a consequence of depth-dependent stretching, necking depth, or intraplate stresses [e.g., *Horváth and Cloetingh*, 1996; *Lankreijer et al.*, 1995]. The 150–200 km of extension is associated with crustal thinning factors up to ~2.2, while the ones of mantle lithosphere have locally extreme values in the order of 5–50 [*Horváth et al.*, 2006; *Lenkey*, 1999; *Sclater et al.*, 1980]. Such extreme values resemble the geometry controlled by detachments of other extensional Mediterranean back arcs and/or the ones observed in hyperextended continental passive margins [*Huismans and Beaumont*, 2011]. The geometry of the Pannonian Basin is generally dominated by typical half grabens, many of them being flanked by detachments with crustal-scale uplift of footwalls, such as observed in the western part of the basin or inferred along its eastern margin [*Ratschbacher et al.*, 1991; *Tari et al.*, 1992, 1999]. Some of these half grabens show typical synkinematic patterns of deposition that are diagnostic for deriving moments of hanging wall subsidence and, interestingly, of footwall uplift [e.g., *Matenco and Radivojević*, 2012; *Tari et al.*, 1992]. Although the link between kinematics, exhumation, and deposition in these extensional features is well studied near the western and southern margin of the basin, much less is known on such genetic features in the main part of the Pannonian Basin, i.e., the Great Hungarian Plain. Remarkable attention has been devoted recently to this area by seismic interpretation and deep geophysical studies [e.g., *Magyar et al.*, 2006; *Matenco and Radivojević*, 2012; *Sztanó et al.*, 2013; *Windhoffer et al.*, 2005]. In this part of the basin, we have used a large network of regionally distributed 2-D and 3-D seismic surveys to study the link between deformation and coeval sedimentation, in order to quantitatively analyze the genesis of the half grabens filled by Miocene sedimentary and volcanic rocks. Starting from the Pannonian Basin case study, our analysis correlated with previously published data has allowed the definition of a coupled tectonosedimentary model of highly extended back-arc basins.

2. Formation and Evolution of the Pannonian Basin

The Miocene to Quaternary formation of the Pannonian Basin was proceeded by a pre-Neogene orogenic evolution that resulted from the opening and subsequent closure of two oceanic realms (Figure 1), the Triassic-Cretaceous Neotethys and Middle Jurassic–Tertiary Alpine Tethys that separated three continental units [e.g., *Csontos and Vörös*, 2004; *Handy et al.*, 2015; *Schmid et al.*, 2008]. In the NW, the AlCaPa Mega-Unit is an Adriatic-derived block that was sutured to Europe during the northward Cretaceous–Eocene closure of the Alpine Tethys [*Csontos*, 1995; *Schmid et al.*, 2004]. To the east and SE, the Dacia unit separated from Europe during the late Jurassic opening of the Ceahlau-Severin Ocean [*Csontos and Vörös*, 2004; *Săndulescu*, 1988]. In the center (Figure 1), Tisza is a unit with mixed affinities that drifted away from Europe during Middle Jurassic and was sutured to Dacia during the late Jurassic–late Early Cretaceous closure of a NE branch of the Neotethys Ocean (i.e., Eastern Vardar [*Haas and Péró*, 2004; *Schmid et al.*, 2008]). The final closure of the Neotethys Ocean by subduction and collision in latest Cretaceous-Eocene times has juxtaposed the, by now welded, Tisza-Dacia upper tectonic plate with the lower Dinaridic unit, the latter being built up by thick-skinned thrust sheets deforming the former Adriatic continental margin [*Karamata*, 2006; *Schmid et al.*, 2008]. The Neogene formation of the Pannonian Basin, coupled with the extrusion of the Eastern Alps, has created a large amount of translations and opposite sense rotations (i.e., counterclockwise in AlCaPa and clockwise in Tisza-Dacia) accompanying the extension of these continental units [*Balla*, 1987; *Csontos*, 1995; *Márton and Fodor*, 2003]. These units were juxtaposed along a major suture zone (Mid-Hungarian Fault Zone) that possibly accommodated the change in polarity from the southward subduction of the Alpine Tethys in the Alps-Carpathians to the northward subduction of the Neotethys in the Dinarides [*Balla*, 1986; *Csontos and Nagymarosy*, 1998; *Schmid et al.*, 2008].

2.1. Extension of the Pannonian Basin

Similar to other highly arcuated Mediterranean retreating subduction systems [*Faccenna et al.*, 2014], the Neogene extension of the Pannonian Basin was coeval with the contraction recorded at the exterior of the Carpathians [e.g., *Ellouz and Rocca*, 1994; *Roure et al.*, 1993]. Extension in the AlCaPa Mega-Unit was accompanied by lateral extrusion from the Eastern Alps [e.g., *Ratschbacher et al.*, 1991] and large-scale offsets along major transcurrent shear zones such as the Periadriatic Fault system and the Balaton line [e.g., *Balla*, 1987; *Csontos and Nagymarosy*, 1998; *Fodor et al.*, 1998; *Ustaszewski et al.*, 2008]. This extension took place dominantly along extensional detachments exhuming deep crustal rocks in core complexes located near the basin margins [*Tari et al.*, 1992; *Fodor et al.*, 1998]. Much less is known about the extensional kinematics of the largest part of the Pannonian Basin, i.e., the Great Hungarian Plain, where such detachments are indirectly inferred near the Dinaridic or South Carpathians margins [*Matenco and Radivojević*, 2012; *Stojadinovic et al.*, 2013; *Ustaszewski et al*., 2010]. In contrast with the counterclockwise rotations accompanying eastward translations in the AlCaPa unit [*Márton and Fodor*, 2003; *Márton et al*., 2007], the overall clockwise up to 100° Paleogene-Miocene rotation of the Tisza-Dacia took place with a rotation pole situated near the SE junction between the Dinarides and Carpathians Mountains. This latter rotation was accommodated by contraction at the exterior of the east and SE Carpathians and by up to 100 km dextral offset of the curved Cerna-Timok fault system in the South Carpathians [*Balla*, 1987; *de Leeuw et al*., 2013; *Fügenschuh and Schmid*, 2005; *Matenco et al*., 2010; *Ratschbacher et al*., 1993]. The uplift of the Alpine-Himalayan mountainous belt has gradually fragmented the larger Tethys Ocean starting with the late Eocene times. The Pannonian Basin is part of the northern branch, the Paratethys, which evolved in a semienclosed marine to lacustrine basin system. The Paratethys is characterized by a separate endemic biostratigraphy (Figure 2) [see *Báldi*, 1986; *Nagymarosy and Müller*, 1988; *Piller et al*., 2007; *Steininger and Rögl*, 1984].

Timing of the main extensional events of the basin (Figure 2) is constrained by the onset of extensional magmatism, by absolute age dating of exhumation in the footwall of extensional detachments that outcrop near the basin margin, and, more importantly, by the timing of synrift and postrift sediments. In the Hungarian part of the basin, the extension is connected with successive volcanic events, mostly of rhyolites or rhyolitic tuffs (Figure 2) [*Pécskay et al*., 2006]. The base of this volcanic sequence is intercalated in the lowermost part of the synrift sedimentation and was originally dated at ~20 Ma (K-Ar dating [*Hámor*, 1985]). More recent dating of the tuff has indicated younger ages of ~17 Ma (Ar-Ar and U-Pb dating [*Pálfy et al*., 2007]). Widespread extensional magmatism creating intrusions or volcanic successions in the Dinarides or near their border with the Pannonian Basin have indicated ages spanning from 22 to 17 Ma [*Cvetkovic et al*., 2007; *Koroneos et al*., 2011; *Schefer et al*., 2011]. Large-scale calc-alkaline magmatism associated with the subduction at the exterior of the Carpathians followed subsequently by adakitic to alkaline magmatism related to postcollisional slab evolution is also recorded in the Pannonian Basin or its adjacent areas (Figure 2) [*Harangi and Lenkey*, 2007; *Seghedi et al*., 2011]. Low-temperature thermochronology has shown that the peak exhumation occurred in the footwall of extensional detachments situated near the transition between the Alps and the Pannonian Basin (Rechnitz window, Pohorje structure, and the detachments bordering the Tauern Window to the east and west) and the main activity of the Periadriatic Lineament span in the interval 23–10 Ma [*Dunkl et al*., 1998; *Fügenschuh et al*., 1997; *Fodor et al*., 2008; *Scharf et al*., 2013; *Tari et al*., 1992]. Near the margin with the Dinarides, in the southwestern periphery of the Pannonian Basin the exhumation in the footwall of extensional detachments has started already in the Oligocene at ~28 Ma and continued with a main peak at Middle Miocene times ~15–11 Ma [*Stojadinovic et al*., 2013; *Toljić et al*., 2013; *Ustaszewski et al*., 2010]. In agreement with thermochronology studies, the analysis of the basin fill and its synkinematic deposition has constrained the timing of the main extensional events of the basin to Early and Middle Miocene, the latter being the moment of peak extension in the entire basin (Figure 2) [*Fodor et al*., 1999; *Horváth et al*., 2006; *Magyar et al*., 1999a; *Nagymarosy and Hámor*, 2012; *Tari et al*., 1999]. The kinematics indicates that in many situations extensional structures reactivate former pre-Neogene thrust contacts [e.g., *Tari et al*., 1992; *Windhoffer et al*., 2005].

Early Miocene depositional environments are dominantly characterized by fluvial, lacustrine, and other continental sediments in our study area of the Great Hungarian Plain [*Matenco and Radivojević*, 2012; *Pavelic et al*., 2001; *Saftić et al*., 2003], while in the northwestern parts of the basin marine sediments deposited [*Nagymarosy and Hámor*, 2012]. Transgression during the Middle Miocene [*Kovác et al*., 2007] resulted in the deposition of deep basinal sediments in the center of the extensional (half) grabens, while deposition along their margins is dominated by near-shore to shallow-marine sedimentation, including shallow-water algal limestone [*Nagymarosy and Hámor*, 2012]. Interestingly, in the center of the Great Hungarian Plain the late Middle Miocene (Sarmatian s.s.) deposition took place in a near-shore environment overlying basement highs, but such sedimentation is absent in the adjacent deeper areas [*Szepesházy*, 1971; *Kőrössy*, 1992]. This indicates that the basin topography was flatter, lacking significantly deep depressions in the center of the basin [*Magyar et al*., 1999b].

2.2. Postextensional Evolution

There is no consensus on a general onset of postrift deposition across the basin, as this process appears to be diachronous. In several parts of the Pannonian Basin, but dominantly in the NW, an intra-Middle Miocene unconformity marks the cessation of the significant normal faulting (~14.8 Ma, Figure 2 [*Tari et al*., 1999]).

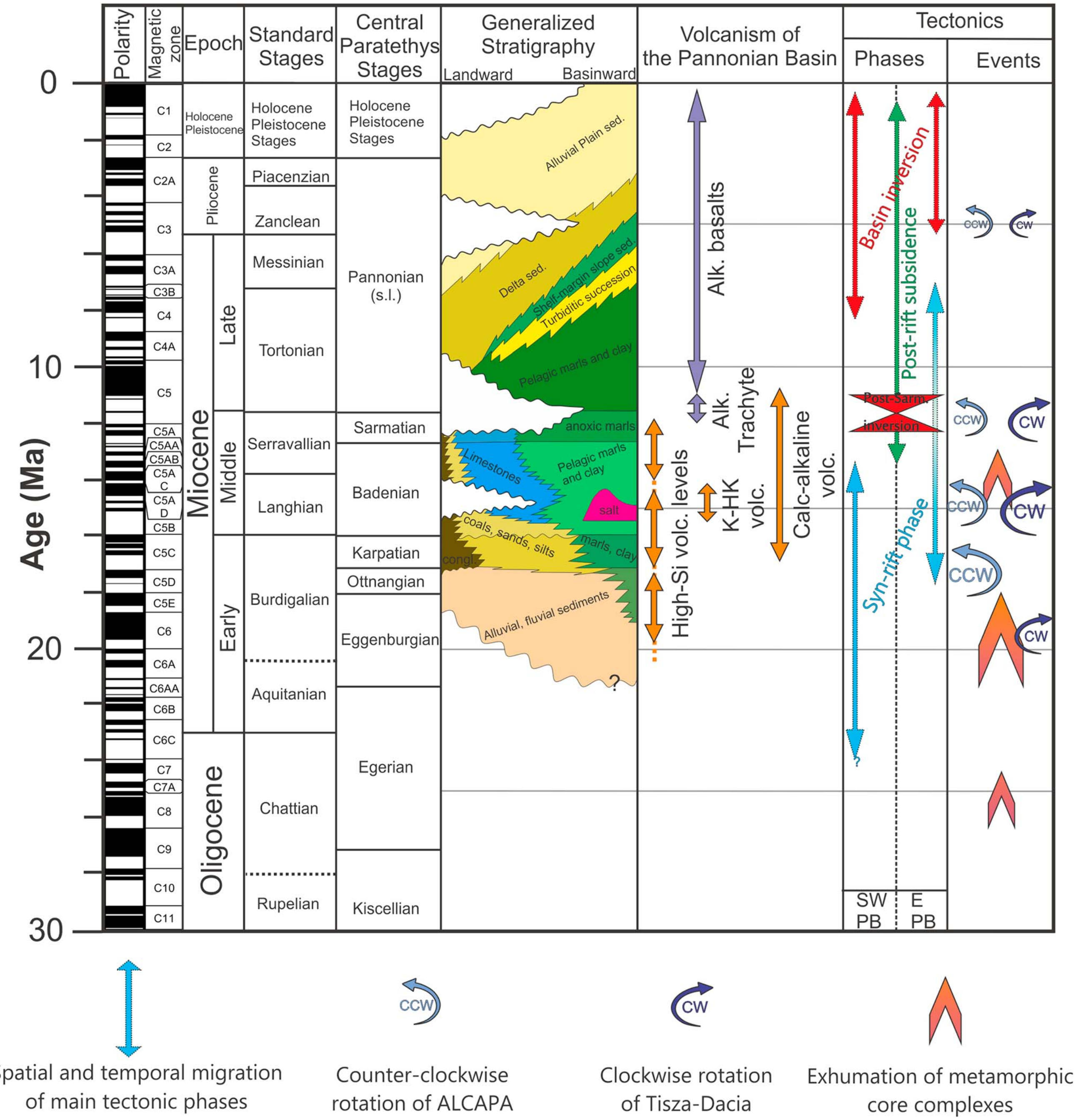

Figure 2. Tectonostratigraphic chart of the Great Hungarian Plain part of the Pannonian Basin with biostratigraphic correlation of the standard and Central Paratethys stages [after *Pezelj et al.*, 2013; *Piller et al.*, 2007] and the generalized Miocene lithostratigraphy of the study area, the volcanic activity of the Pannonian Basin [after *Harangi and Lenkey*, 2007; *Pécskay et al.*, 2006], and the main tectonic phases and events [after *de Leeuw et al.*, 2013; *Horváth et al.*, 2006; *Márton and Fodor*, 2003; *Márton et al.*, 2007]. Latest Miocene-Pliocene minor vertical axis rotations are connected to basin inversion. Note that the synrift/postrift boundary and the onset of last stage of basin inversion appear to be older in the SW and progressively younger east northeastward.

In other parts of the Pannonian Basin, but dominantly in the Great Hungarian Plain, the onset of postrift deposition is interpreted at the boundary between Middle and Late Miocene (11.63 Ma, in the sense of *ter Borgh et al.* [2013]). This was associated with the development of an unconformity interpreted to be coeval with the peak of collision in the East Carpathians (Figure 2, first phase of basin inversion [*Horváth*, 1995]), where thrusting ceased subsequently until 9 Ma [*Matenco and Bertotti*, 2000; *Merten et al.*, 2010]. In the

Pannonian Basin, this unconformity removed parts of the upper Middle Miocene succession (Sarmatian strata [*Magyar et al*., 1999b]). In contrast, near the Pannonian Basin margins (Vienna, East Slovakian, Transylvanian, and Danube basins) the thickness of Sarmatian sediments is significant, locally exceeding 1 km. Other seismic interpretation studies in the SE part of the basin suggested that extensional deformation was diachronous across the basin and migrated in time and space from ~28 Ma near the Dinarides to 8–5.5 Ma northeastward and eastward [*Matenco and Radivojević*, 2012]. This means that in a first stage the Dinaridic nappe contacts were reactivated as extensional detachments or low-angle normal faults. This was followed by a progressive migration of depocenters toward the present-day center and E-NE part of the Great Hungarian Plain. Late Miocene extensional structures with various offsets were observed by outcrop and seismic interpretation studies in the center of the Pannonian Basin [*Balázs et al*., 2013; *Fodor et al*., 2013].

The uplift of the Carpathians and the associated unconformity between the Middle and Late Miocene strata mark the disruption of connections with the remainder of the Paratethys realm. This coincides with the onset of a marked environmental and sedimentological change in the evolution of the Pannonian Basin, which was restricted to the size of a large isolated lake (Lake Pannon [*Magyar et al*., 1999a; *ter Borgh et al*., 2013]). An up to 7 km thick sedimentary succession was deposited in the Great Hungarian Plain during Late Miocene to recent times. The basin fill recorded an initial transgression followed by shelf margin and slope progradation driven by the influx of sediments by a fluvial system resembling the present-day Danube and Tisza rivers. This shelf margin prograded ~400 km in 6 Myr until ~4 Ma from the NW and NE in a ~ S-SE direction, while minor progradation was recorded in other directions [*Magyar et al*., 2013; *ter Borgh et al*., 2015; *Vakarcs et al*., 1994; *Pogácsás et al*., 1988]. The coeval sedimentation reflects the deposition of a number of diachronous lithostratigraphic formations that mirror the various lithofacies associations of a deep lake depositional environment (Figure 2) [*Bérczi et al*., 1987; *Juhász*, 1991; *Sztanó et al*., 2013]. These associations are laterally variable (Figure 2) from deep hemipelagic deposition (Endrőd Formation), turbidites (Szolnok Formation), prograding shelf margin slope (Algyő Formation), and delta (Újfalu Formation) to alluvial plain sediments (Zagyva Formation). Their typical expression in seismic lines provides an excellent lateral correlation of seismic facies. These diachronous associations are also correlated by magnetostratigraphic and biostratigraphic studies calibrated by a limited number of absolute age measurements [*Magyar et al*., 1999b; *Magyar and Sztanó*, 2008; *Magyar et al*., 2013].

The cessation of extension was followed by the onset of large-scale inversion in the Pannonian Basin during late Miocene times (from 8–7.5 Ma) controlled by the counterclockwise rotation and push of the Adriatic microplate, which created large-scale contractional structures near the Dinaridic margin and dominantly transcurrent kinematics elsewhere. This deformation event is still presently active [*Bada et al*., 2007; *Dombrádi et al*., 2010; *Fodor et al*., 2005; *Horváth and Cloetingh*, 1996; *Jarosinski et al*., 2011; *Pinter et al*., 2005; *Uhrin et al*., 2009]. An unconformity is observed in the basin fill near the boundary between the Miocene and Pliocene basin fill [e.g., *Vakarcs et al*., 1994]. This unconformity is angular and locally erosional near the basin margins and passes to a correlative conformity toward the basin center. It is interpreted to be either related to the basin inversion [*Magyar and Sztanó*, 2008; *Sacchi et al*., 1999; *ter Borgh et al*., 2015] or formed in response to the Messinian Sality Crisis of the Paratethys [e.g., *Csató et al*., 2015]. One other slightly older unconformity formed at ~6.8 Ma is observed at depth in the Great Hungarian Plain (Figure 3). This has been interpreted as either the result of a significant water level fall of Lake Pannon [e.g., *Csató*, 1993; *Vakarcs et al*., 1994] associated with the formation of large canyon incision in the center of the Great Hungarian Plain (the Alpar Canyon of *Juhász et al*. [2013]) or a crossover zone of different progradational directions [*Magyar and Sztanó*, 2008].

3. Methodology

We have analyzed the link between structures and sedimentation in the subbasins located in the Great Hungarian Plain part of the Pannonian Basin mostly in Hungary, and also in Serbia, Slovakia, and Romania (Figures 1 and 3). In Hungary, we have analyzed a large array of 2-D and 3-D seismic data calibrated by a dense network of exploration wells distributed regionally in the study area particularly in the studied subbasins and also on their flanks and connecting areas. The signal/noise ratio and resolution of the seismic sections are variable, driven by the variability from recent 3-D seismic surveys to older 2-D seismic lines. Although in this paper we present only a limited number of seismic transects usually oriented across the

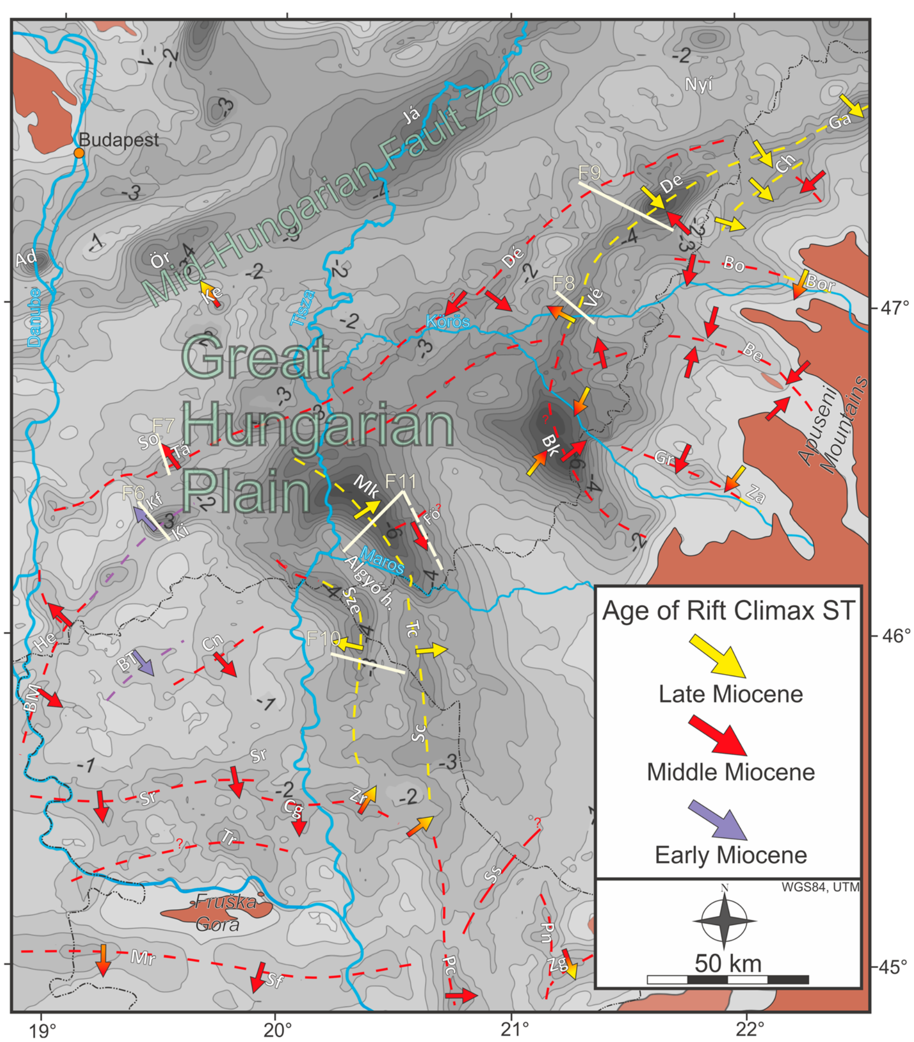

Figure 3. Neogene basement isopach (in kilometers, compiled from *Haas et al.* [2010], *Matenco and Radivojević* [2012], *Răbăgia* [2009], *Tulucan* [2007], and the data of this study) and tectonic map of the Great Hungarian Plain. The map indicates the main depocenters with their strike (dashed lines), age of rift climax system tract, and the main direction of extensional tectonic transport in the various subbasins (the same color-coded arrows). White lines show the locations of the interpreted seismic sections. Main subbasins of the area: Nyí = Nyírség subbasin, Ga = Galospetru-Mecentiu Depression, Ch = Chet Tamaseu Depression, De = Derecske Trough, Bor = Borod Depression, Bo = Bors Depression, Vé = Vésztő Trough, Be = Beius Depression, Dé = Dévaványa Trough, Já = Jászság Basin, Ör = Örkény Trough, Ad = Adony Basin, Ke = Kecskemét Depression, Tá = Tázlár Trough, So = Soltvadkert subbasin, Ki = Kiskunhalas Trough, Kf = Kunfehértó subbasin, Za = Zarand Depression, Gr = Graniceri Depression, Bk = Békés Basin, Mk = Makó Trough, Fö = Földeák subbasin, Sze = Szeged Trough, He = Hercegszántó Trough, BM = Bački Monoštor Depression, BT = Bačka Topola Depression, Cn = Čantavir Depression, Tc = Tomnatec Depression, Sc = Srpska Crnja Depression, Sr = Srbobran Depression, Cg = Čurug Depression, Zr = Zrenjanin Depression, Tr = Temerin Depression, Mr = Morović Depression, Sf = Sefkerin Depression, Ss = Samoš Depression, Pc = Pančevo Depression, Pn = Plandište Depression, Zg = Zagajica Depression.

strike of various subbasins, we used a much larger seismic and well database (including hundreds of 2-D seismic lines and a few 3-D seismic surveys). Well logs were tied to seismic sections using the standard vertical seismic profiling logs and checkshots commonly available in the exploration industry [e.g., *Mészáros and Zilahi-Sebess*, 2001]. The interpretation was also assisted by Bouguer anomaly and vertical component magnetic anomaly data using maps available in the Pannonian Basin [*Kiss*, 2006; *Kiss and Gulyás*, 2006; *Tari et al.*, 1999]. The reflectivity of the seismic data decreases substantially beneath the basin fill sedimentary rocks; thus, the pre-Neogene interpretation relies dominantly on well data [*Haas et al.*, 2010] and the outcropping areas situated on the flanks of few subbasins. The exceptions are the zones with increased reflectivity and correlation in seismic lines, such as the carbonatic cover in the Mesozoic sediments, which were correlated across the basin. This interpretation is in general agreement with the recent pre-Neogene interpretation based of the entire Hungarian well database [*Haas et al*., 2010], but local details may vary. Outside Hungary, in the Pannonian Basin regions of Slovakia, Serbia, and Romania, in the Dinarides of Serbia and Montenegro and in the Carpathians of Poland the regional interpretation used published studies [*Dimitrijević*, 1997; *Ellouz and Rocca*, 1994; *Gagała et al*., 2012; *Matenco and Radivojević*, 2012; *Pigott and Radivojević*, 2010; *Puchnerová et al*., 2002; *Răbăgia*, 2009; *Schmid et al*., 2008; *ter Borgh*, 2013; *Tulucan*, 2007].

The structural analysis followed the typical seismic interpretation methodology in defining reflector terminations, such as truncations, onlaps, toplaps, offlaps, or downlaps. Our seismic stratigraphic interpretation used a combination of classical sequence stratigraphy and applied tectonic system tract methodology that is less known in available literature and requires more explanation. Starting from the readily available principles of sequence stratigraphy that study the relationship between the accommodation space and sediment supply [e.g., *Catuneanu et al*., 2009, 2011; *Posamentier and Allen*, 1993; *Schlager*, 1993; *Vail et al*., 1977], the application to tectonically active basins [e.g., *van Wagoner et al*., 1990] is less known. In active basins, the system tracts and sequences are linked to vertical movements; they can be related to individual events of fault activation and are independent from the known cyclicity timescales of the classical sequence stratigraphy [e.g., *Miall and Miall*, 2001].

In extensional basins, tectonic sequence stratigraphic models are available by correlating the sedimentation and the tectonic subsidence of hanging walls [e.g., *Martins-Neto and Catuneanu*, 2010; *Nottvedt et al*., 1995; *Prosser*, 1993; *Ravnas and Steel*, 1998; *van Wagoner et al*., 1990], but these studies do not quantify the footwall uplift in asymmetric systems. Similar with other studies [e.g., *Hinsken et al*., 2007; *Pereira and Alves*, 2012; *Răbăgia et al*., 2011], we have used a combination of these readily available (seismic) sequence stratigraphic methodologies. The first-order extensional cyclicity has been defined by using a seismic sequence stratigraphic model [*Nottvedt et al*., 1995; *Prosser*, 1993]. Among the limited availability of such approaches in existing literature, this model has the closest similarity in terms of geometries with the ones detected in our study and, therefore, provides the closest meaning of tectonic system tracts. A tectonic system tract is defined by linked depositional systems controlled dominantly by tectonics, bounded by key stratigraphic surfaces [e.g., *Prosser*, 1993]. In such a definition, the rift initiation system tract records the first extensional pulses in the basin and is followed by a rift climax system tract, which reflects the moments of maximum fault activity and subsidence rates. The end of faulting marks the start of an immediate postrift system tract, when the continued thermal sag subsidence resulted in the burial of the inherited rift topography, which is followed by a late postrift system tract, when compaction and gradual slowing of subsidence drives a final stage of regressive basin fill. Differently from the original model, these system tracts have a different meaning by reflecting the specific evolution of half grabens bounded by exhumed footwalls. Each system tract is associated with characteristic seismic facies associations grouping seismic facies units, reflecting the depositional environment. The burial in late postrift stage may create differential compaction effects over the half grabens, such as synclinal geometries or compaction faults with increasing offsets toward the surface, in particular when the thickness of the overburden is high.

The large synkinematic sedimentation rate of the Pannonian Basin increased the time resolution of the seismic lines and enabled us to locally detect a higher-order sequence stratigraphic cyclicity also in the seismic data. This is generally related to individual episodes of normal faulting and was analyzed by defining transgressive-regressive sequences [e.g., *Catuneanu*, 2002; *Catuneanu et al*., 2009; *Embry and Johannessen*, 1992; *Johnson and Murphy*, 1984]. The amplitude, frequency, continuity, terminations, and distribution of reflectors define various seismic facies units (Figure 4), which were subsequently grouped into seismic facies associations defining progradational, retrogradational or aggradational geometries. They are controlled by the rate of accommodation and the sediment supply [*Catuneanu et al*., 2009]. The specificity of asymmetric extensional systems is the footwall exhumation and erosion and the migration of faulting in space and time. As long as the footwall is eroded, the coeval deposition and onlaps in the neighboring hanging wall are coastal, and therefore, a direct interpretation of the progradational, retrogradational, or aggradational geometries in stratigraphic sequences is possible. Footwall erosion, combined with the correlative maximum regression surface interpreted as a classical sequence boundary defined by the geometry of the seismic facies units, is an expression of the composite surface that bounds a transgressive-regressive (TR) sequence [*Embry and Johannessen*, 1992; *Johnson and Murphy*, 1984]. Variations in paleobathymetries are available in published studies of Early and Middle Miocene basin evolution, although in the case of Late Miocene

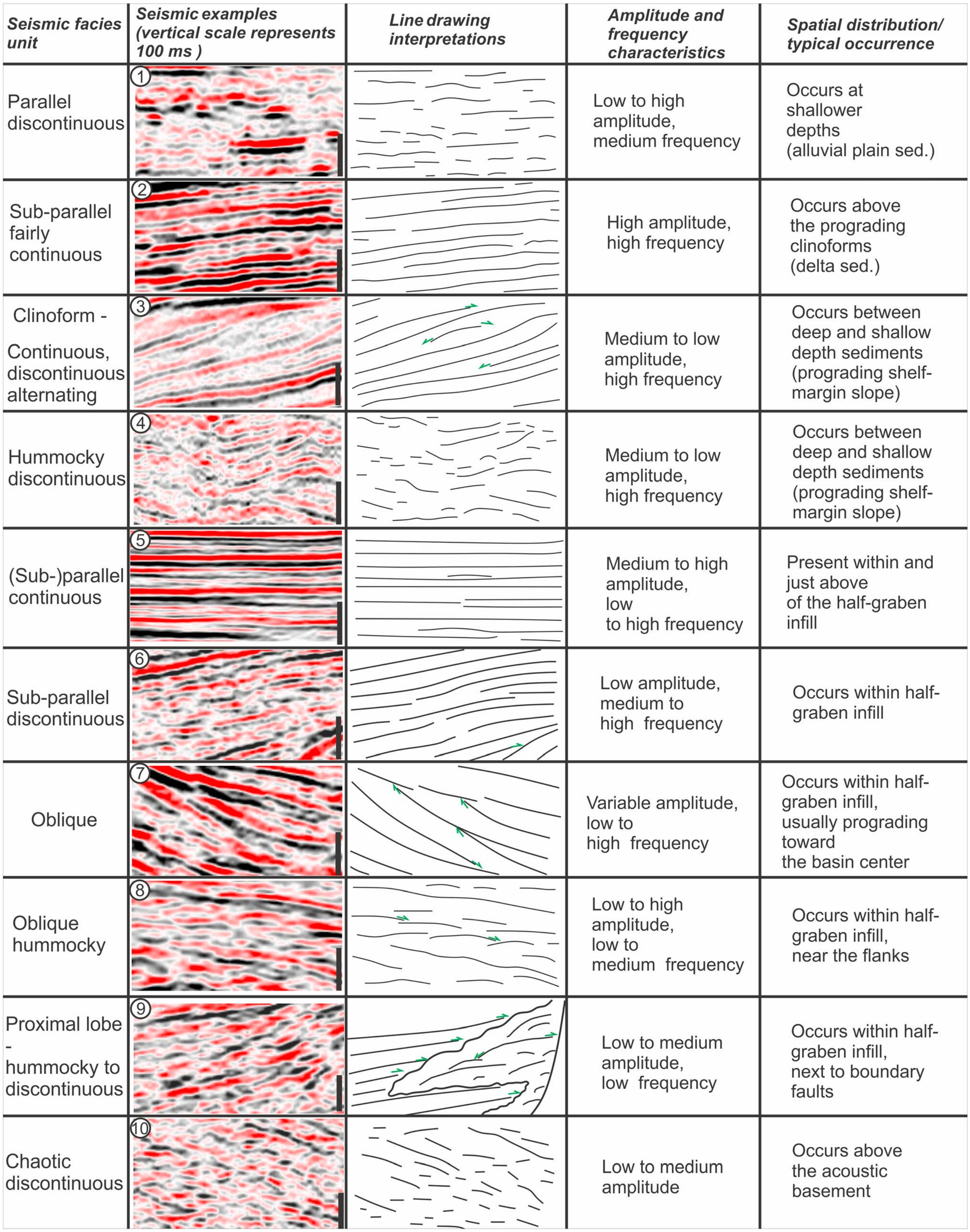

Figure 4. Characteristic seismic facies units used in the seismic sequence stratigraphic interpretation.

paleobathymetry, our model is more speculative [e.g., *Báldi et al*., 2002; *Lemberkovics*, 2014; *Pezelj et al*., 2013; *Sztanó et al*., 2013]. The maximum flooding surface of a TR sequence is less controlled in older sediments located at the center of various subbasins, where wells are less frequent. In these situations, the maximum flooding surface has been taken as the boundary between retrogradational and progradational geometries. Although this is an important approximation in theory, its error does not affect significantly the practical interpretation of individual episodes of normal faulting.

Our seismic stratigraphic interpretation is also constrained by available well logs, such as resistivity and gamma ray logs (Figure 5). These logs provide the required validation of the prograding-retrograding facies associations and evolution of the sedimentary infill in the half grabens of the Great Hungarian Plain. This integrated approach enables the analysis of the episodic tectonosedimentary signature of the synkinematic basin fill at different scales [see *Martins-Neto and Catuneanu*, 2010; *Pereira and Alves*, 2012]. Although such well log interpretations are available in the entire study area of the Great Hungarian Plain, we have chosen to illustrate one well in the center of the Kiskunhalas subbasin and two wells at its flank (Figures 5 and 6). Their interpretation followed a standard well log sequence stratigraphic approach [e.g., *van Wagoner et al*., 1990], the derived cyclicity being correlated and interpreted with the evolution of the seismic facies units and associations.

4. Structural and Seismic Sequence Stratigraphic Interpretations in the Great Hungarian Plain

The Miocene extension created a number of individual subbasins in the Great Hungarian Plain (Figure 3). The main characteristic geometry of these subbasins is the low-angle dip of their flanks (Figures 6–11), which is unusual for typical upper crustal normal faults. This geometry suggests the existence of large-scale low-angle normal faults or detachments, as interpreted in other areas in the upper crust of the Pannonian Basin [*Tari et al*., 1992, 1999]. These structures flank subbasins with low-angle dipping hanging walls in overall half-graben geometries. Although antithetic normal faults with significantly lower offsets often crosscut these hanging walls, the synkinematic basin fill shows that the structure is still highly asymmetric (e.g., Figure 6). These half grabens frequently show an opposite polarity of low-angle normal faults, have decreasing offsets, and are connected by transfer faults along their strikes [see *Tari et al*., 1992]. The overall strike of these structures separated by basement highs is NE-SW in the central part of the study area, changing to E-W in the south near the Dinaridic margin and to NW-SE in the eastern part of the Great Hungarian Plain near the southern Apuseni Mountains (Figure 3). The overall low-angle dipping geometry of the normal faults and the lateral variations along their strike reflect in many situations the geometry and variability of the thrust kinematics in the nappe stack developed during the Cretaceous-Paleogene evolution of the Tisza-Dacia Mega-Unit.

4.1. Seismic Facies Units in the Subbasins of the Great Hungarian Plain

Ten seismic facies units have been differentiated in the Miocene basin fill of the Pannonian Basin (Figure 4). Although the seismic characteristics depend on the seismic acquisition and processing methodology of various seismic data, a general pattern can be established. The first three facies units, large-scale (sub)parallel discontinuous, fairly continuous, and clinoform are characteristic for the large-scale progradation that took place during Pannonian (s.l.) times in the upper part of the basin fill, and in all cases they postdate the extension. The clinoform facies unit contains locally intercalated hummocky ones (i.e., fourth seismic facies unit), which are in fact either clinoforms crosscut along their strike or local turbiditic bodies (Szolnok Formation). The spatial distribution and significance of these postrift facies units are well documented elsewhere [*Juhász et al*., 2007; *Magyar et al*., 2013; *Sztanó et al*., 2013]. The synkinematic sedimentation in the various subbasins is characterized by a combination between different seismic facies units (units 5–10, Figure 4). In the lower parts of the synkinematic basin fill a chaotic seismic facies (unit 10) characterizes the onset of extension. This unit is either buried at high depth in the center of the graben (e.g., Figure 6) or spatially shifted along its flanks (e.g., Figures 7 and 9). The subparallel continuous and discontinuous seismic facies (units 5 and 6) are the most common ones usually observed in the center of the basins at high distance from its flanks by onlapping previous unconformity surfaces and interfingering with other facies units. These units are (sub) parallel or can be gently divergent due to coeval offsets along normal faults. The proximal and distal lobes seismic facies (units 9 and 8) are observed in contact with the flanks of subbasins, in particular well developed

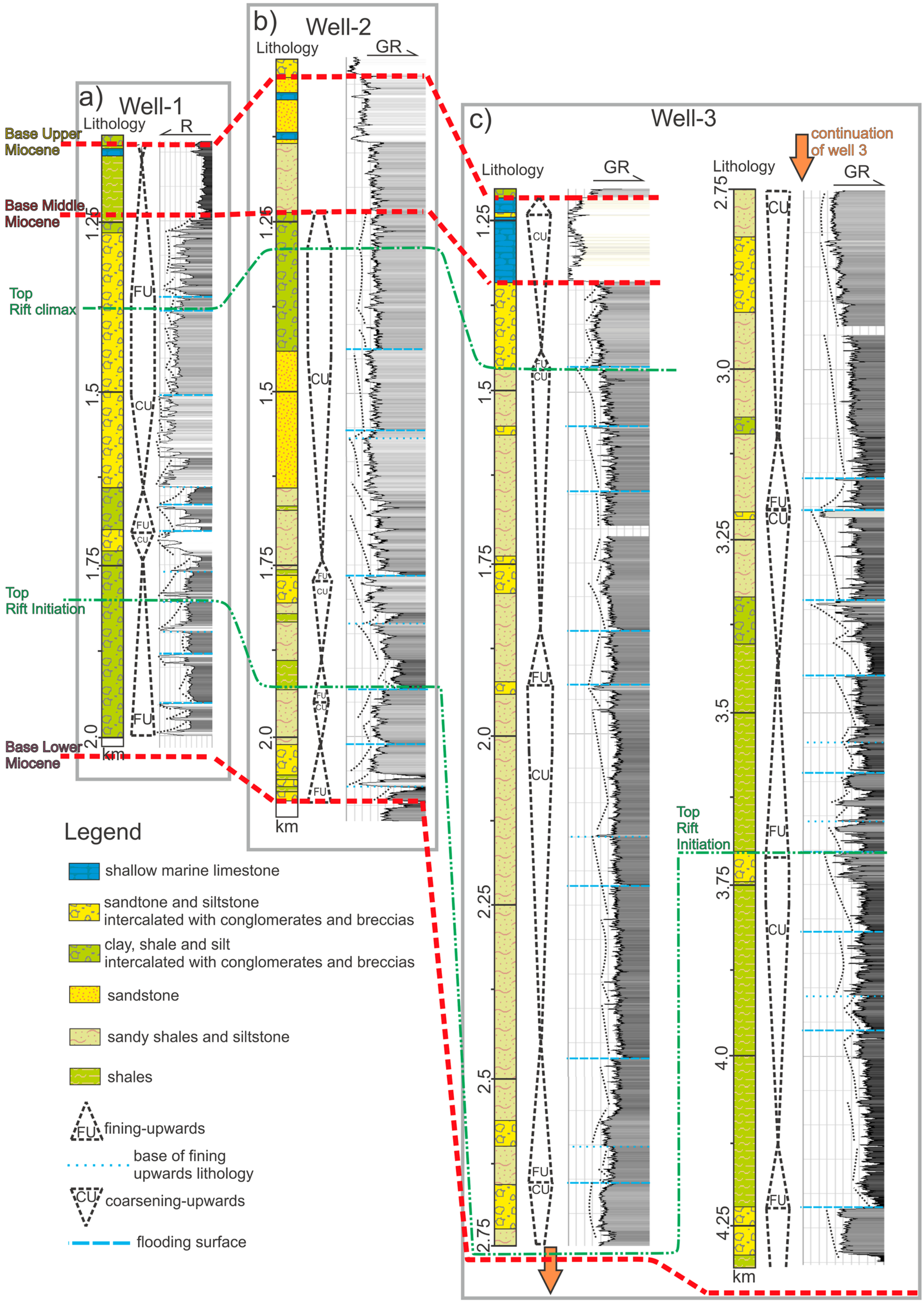

Figure 5. Sequence stratigraphic well log interpretation from the Kiskunhalas subbasin showing cyclicity of the Neogene basin fill. Generalized lithological column is based on well reports. Locations of the wells are displayed in Figure 6.

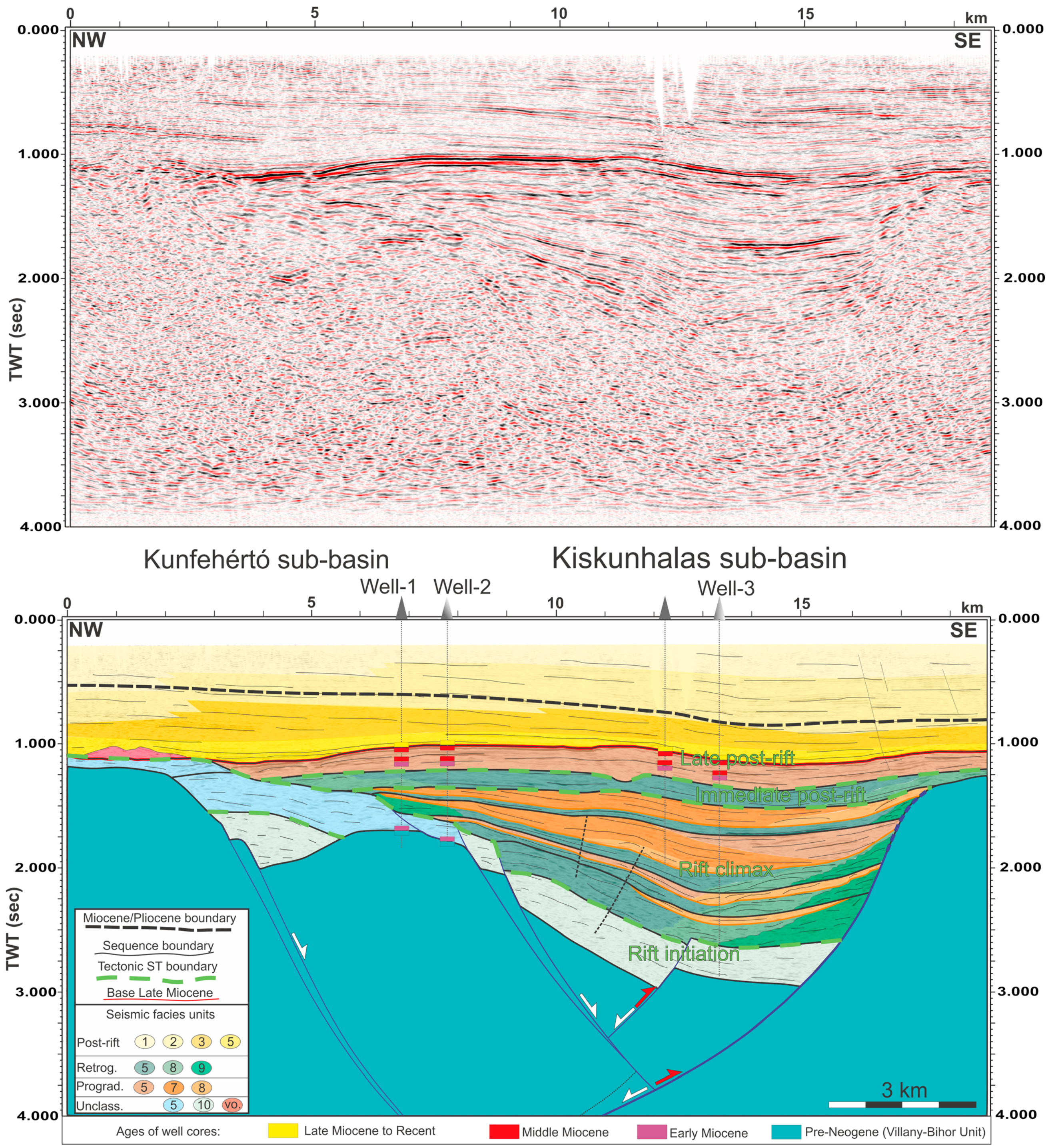

Figure 6. (top) Noninterpreted and (bottom) interpreted reflection seismic section from the Kiskunhalas Trough. Location is displayed in Figure 3. White arrows indicate the Miocene kinematics of faults, red arrows show the late Middle–early Late Miocene inversion of the structure.

against normal faults. The proximal lobes facies unit has a gravitational character with coarse deposition and reflects periods of erosion of the adjacent footwall or hanging wall flank. The lobes are always in close spatial contact, prograding toward the center of the basin, retrograding toward its margins, or aggrading. In the lower parts of the basin fill, this facies shows rapid progradation or retrogradation, while these patterns become more attenuated in its upper part. The oblique facies (unit 7) reflects progradation and is bounded by downlap and toplap reflection terminations.

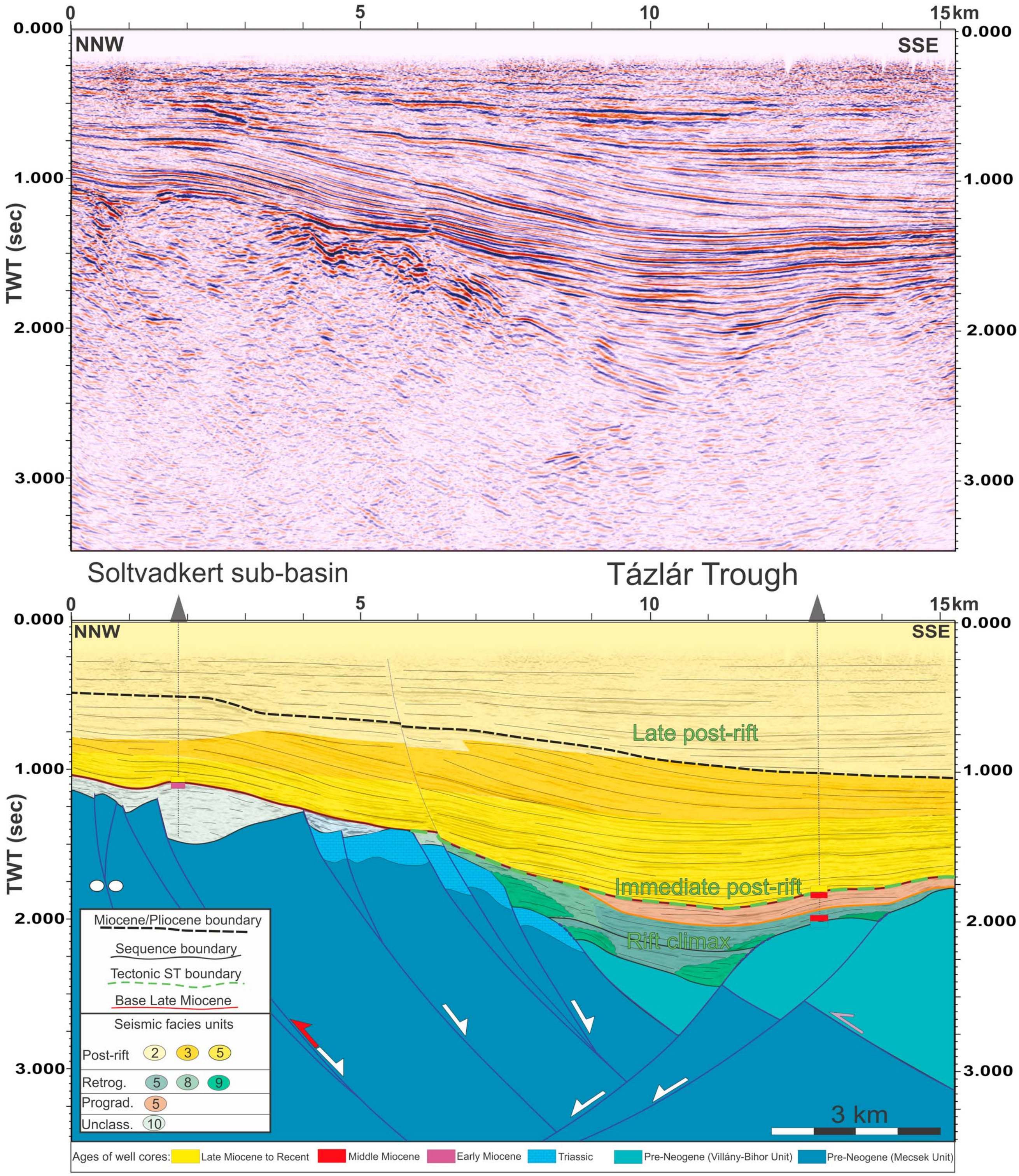

Figure 7. (top) Noninterpreted and (bottom) interpreted reflection seismic section from the Tázlár Trough. For location, see Figure 3. Grey arrow indicates Cretaceous thrust. White arrows indicate the Miocene kinematics of faults. Red arrow indicates the possible inversion of the fault of the small Soltvadkert subbasin.

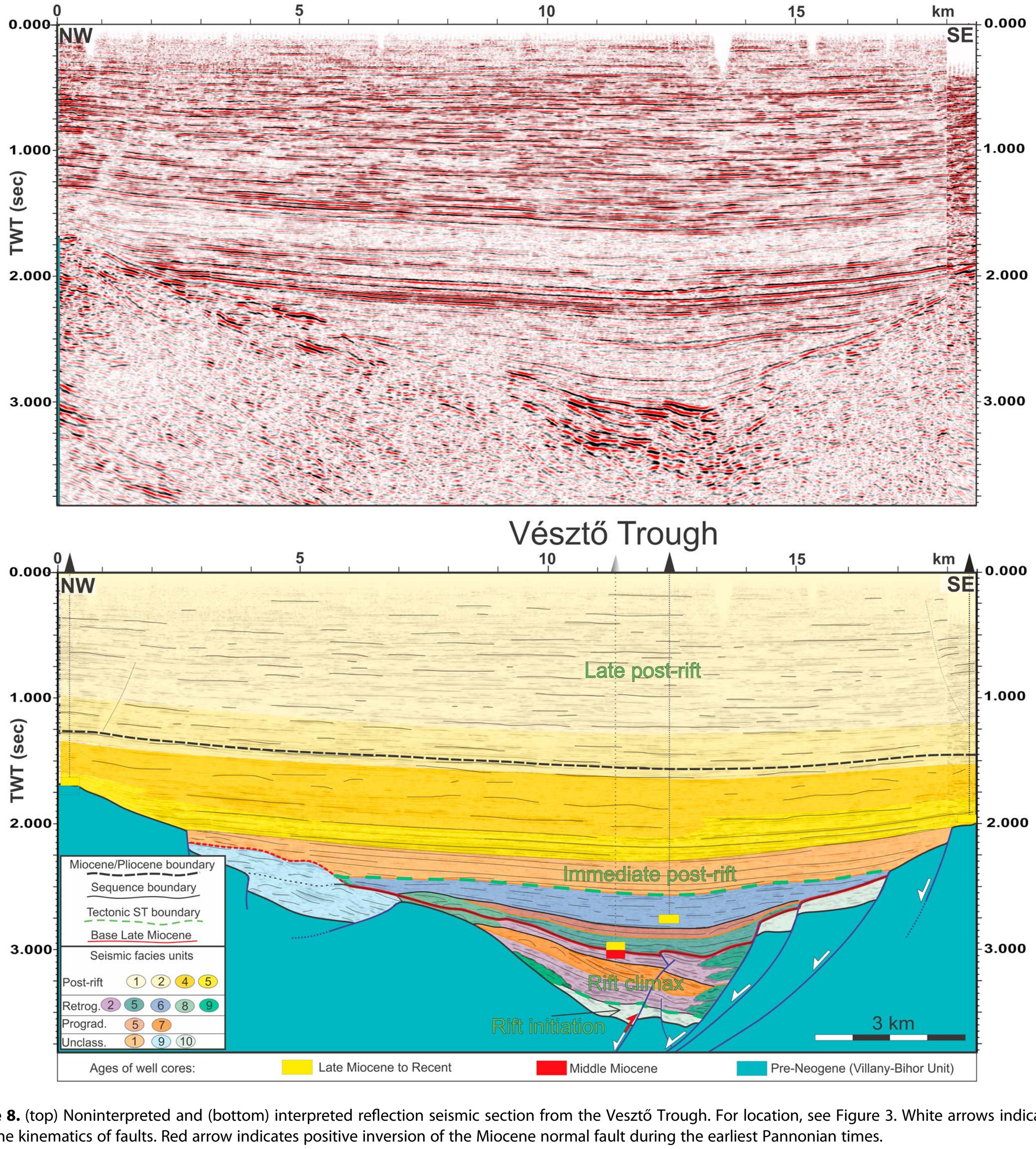

Figure 8. (top) Noninterpreted and (bottom) interpreted reflection seismic section from the Vésztő Trough. For location, see Figure 3. White arrows indicate the Miocene kinematics of faults. Red arrow indicates positive inversion of the Miocene normal fault during the earliest Pannonian times.

4.2. Structures With Peak Synkinematic Deposition During the Early Miocene

The Kiskunhalas Trough, with limited extent along its strike, is a narrow and deep basin, with synkinematic sedimentation reaching more than 3 km in thickness (Figure 6). The overall structural geometry is asymmetric, with one large listric normal fault dipping northwestward and several lower offset antithetic normal faults (Figures 3 and 6). The minimum value of the total horizontal offset cumulated along all normal faults, calculated by displacements in the synkinematic hanging walls deposition, is in the order of 7 km. Note that

10.1002/2015TC004109

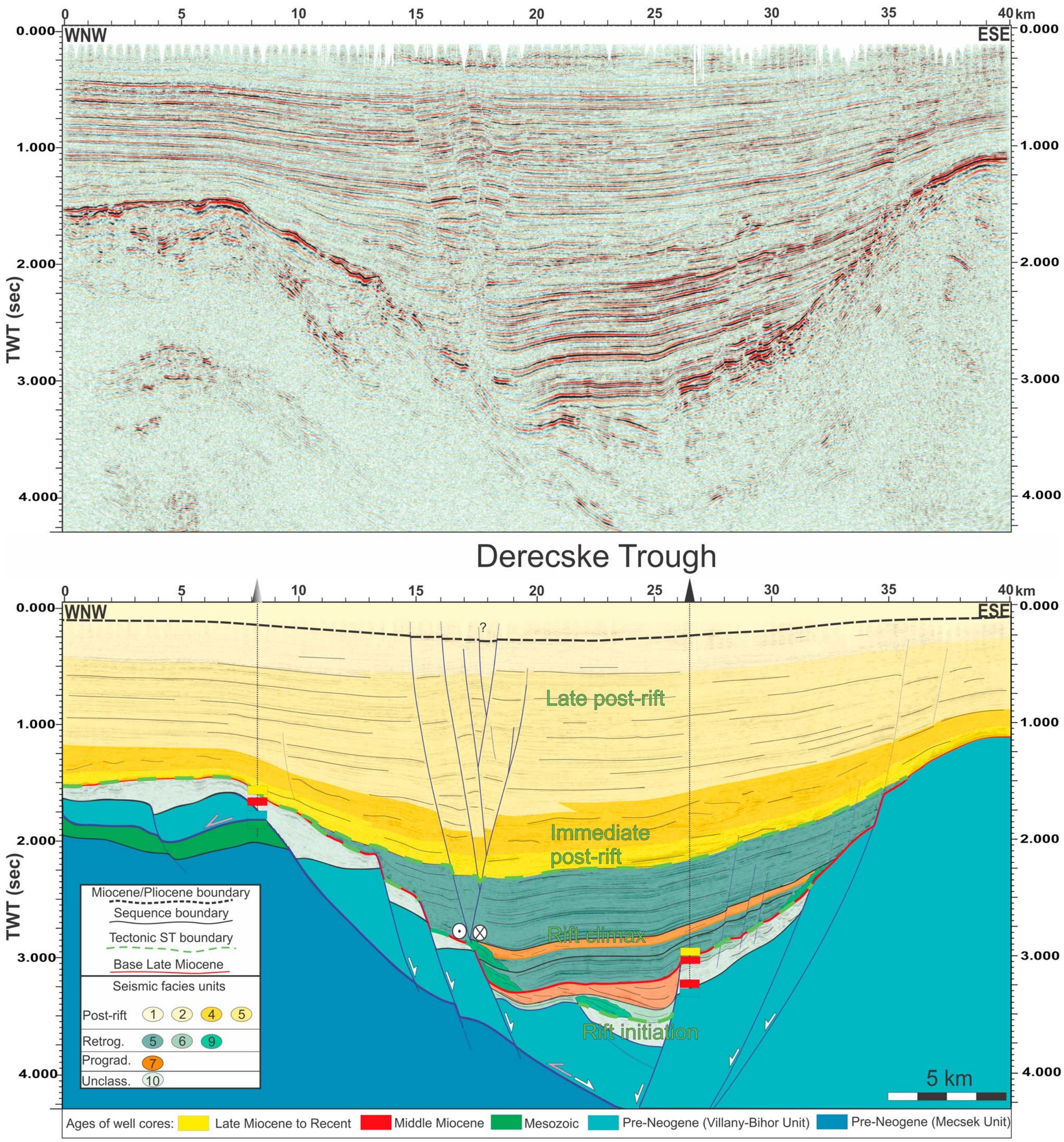

Figure 9. (top) Noninterpreted and (bottom) interpreted reflection seismic section from the Derecske Trough. For location, see Figure 3. Grey arrows indicate Cretaceous thrusts [after *Windhoffer et al.*, 2005]. White arrows indicate the Miocene normal faulting. The fault zone reactivated again during Pliocene–Quaternary times as a sinistral strike-slip zone creating the young negative flower structure.

such offset calculations do not include footwall exhumation affected by erosion and therefore provide a minimum estimate. This erosional geometry is visible in the Kiskunhalas Trough by seismic onlaps and is most likely subaerial (Figure 6). The drilled pre-Neogene sequence includes medium-grade metamorphic rocks, Triassic shallow-water carbonates, and Cretaceous deposits that are diagnostic for the Tisza unit. These pre-Neogene rocks have been attributed to the Villány-Bihor nappe of the Tisza unit [*Haas et al.*, 2010]. Wells have penetrated in the deepest part of the basin, a (volcano-) clastic lacustrine succession

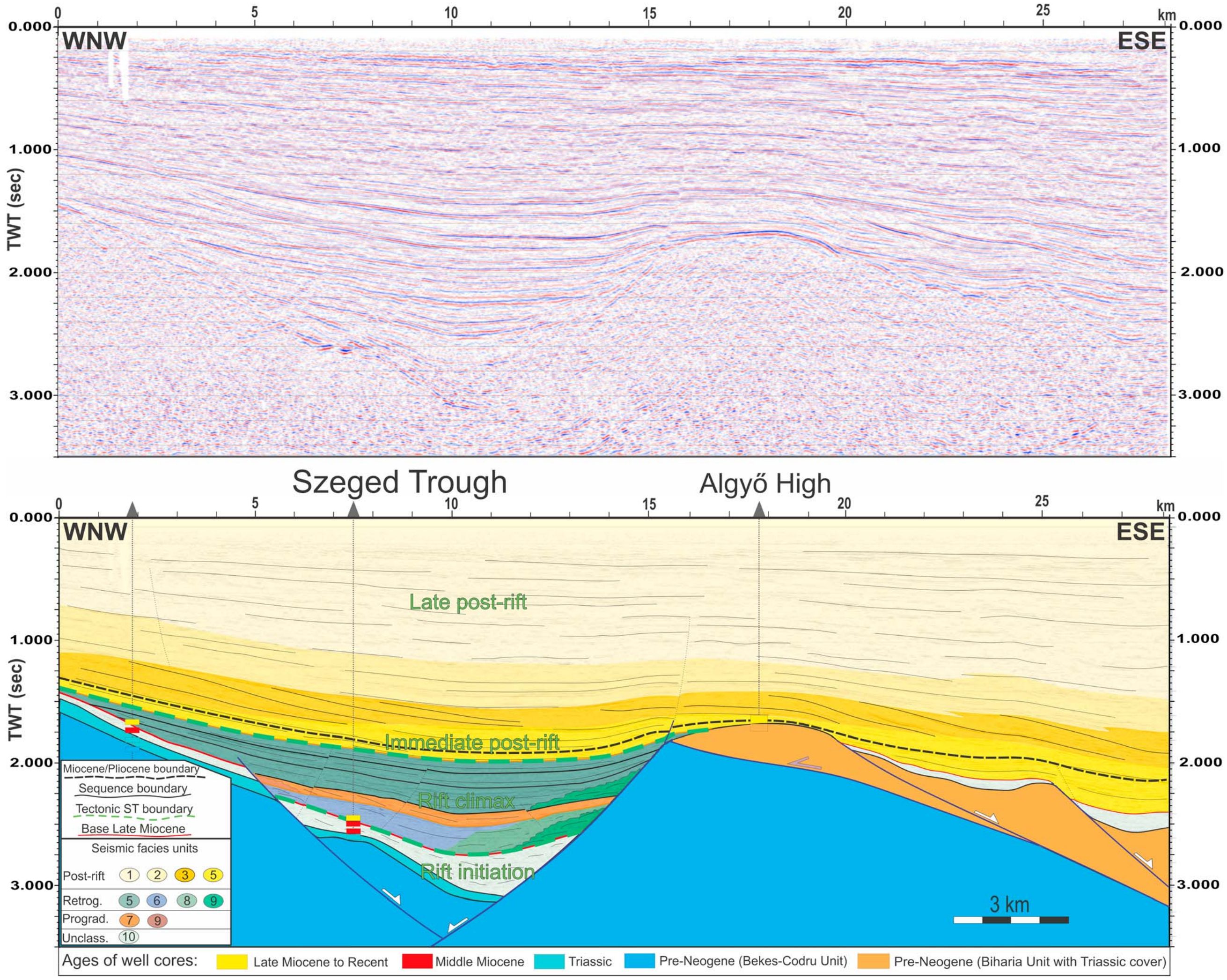

Figure 10. (top) Noninterpreted and (bottom) interpreted reflection seismic section from the Szeged (Banatsko Aranđelovo) Trough. For location, see Figure 3. Grey arrow indicates Cretaceous thrust of a piece of Dacia onto Tisza. White arrows indicate the Miocene kinematics of faults.

(Kiskunhalas Formation). This is correlated with a similar succession found elsewhere that contains poorly developed Lower Miocene (Karpatian) fauna [*Rumpler and Horváth*, 1988; *Kőrössy*, 1992]. It is overlain by lower Middle Miocene (Badenian) carbonate and clastic rocks that are separated by an erosional unconformity from the overlying Pannonian sedimentary rocks. The lateral correlation of biostratigraphic ages observed in wells demonstrates that the age of the normal faults is Early Miocene. This unconformity is folded in a gentle anticline with Pannonian strata onlapping over its flanks. The entire upper Middle Miocene (Sarmatian) is missing or only a few meters thick below the unconformity [*Kőrössy*, 1992]. This folded unconformity is the result of contraction [*Horváth*, 1995] that occurred during the late Middle Miocene or earliest Late Miocene and is associated with very low-offset, high-angle reverse faults with south vergence located in the core of the anticline that reactivate the earlier normal faults (Figure 6). The overlying Late Miocene sequence is gently deformed and has thicknesses of about 1 km.

The seismic sequence interpretation of the Kiskunhalas subbasin (Figure 6) shows a chaotic seismic facies unit in its lower part, interpreted as a rift initiation system tract of Early Miocene in age. The overlying Early Miocene rift climax sedimentation is composed of prograding and retrograding facies association units. Coeval erosion of the footwall indicates that the onlaps are coastal, and the correlative maximum regression surface on the top of the prograding units is a regressive-trangressive sequence boundary. The rift climax system tract contains clastic lobes facies units that decrease gradually upward in the stratigraphy, most likely as

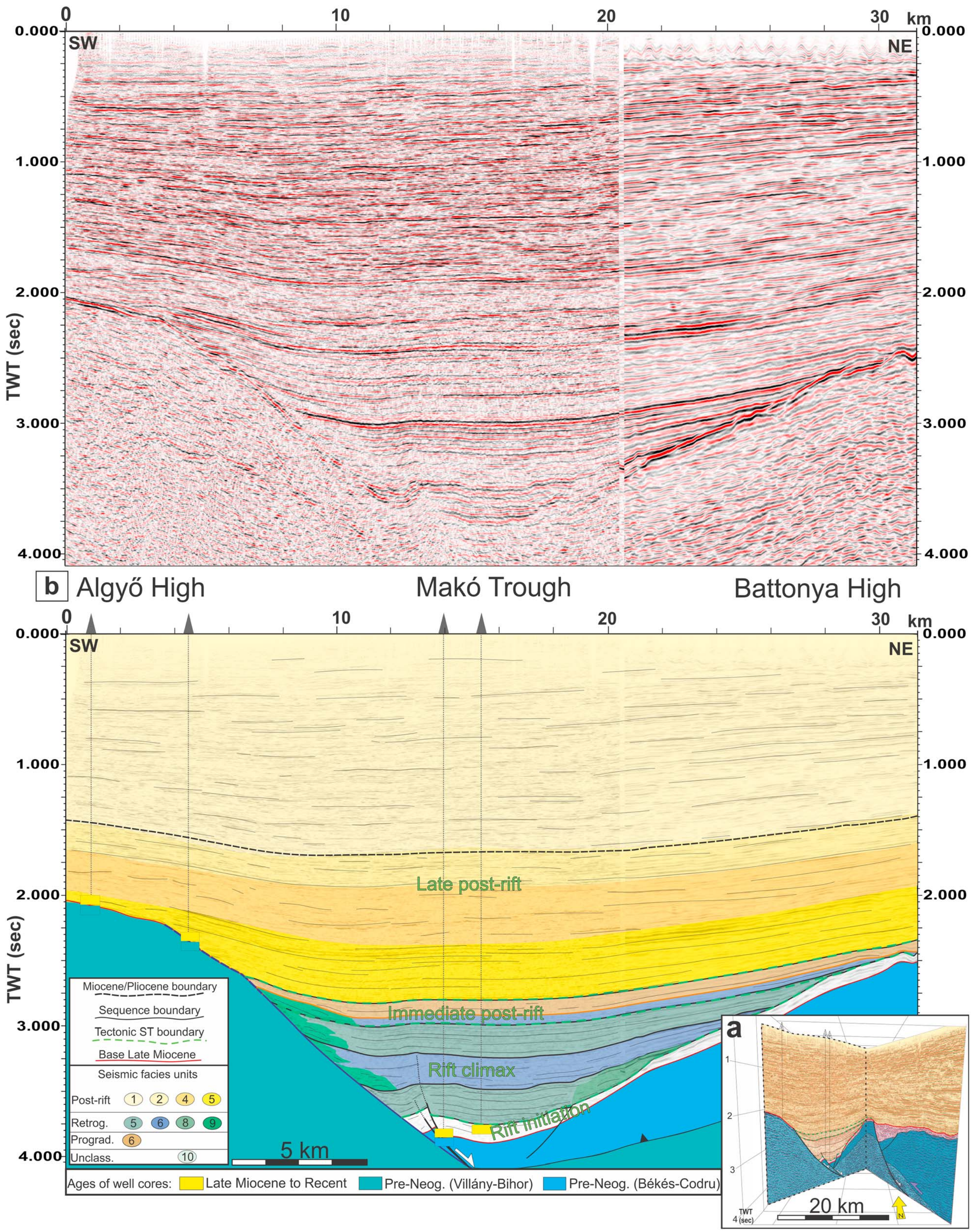

Figure 11. Seismic data from the vicinity of the Makó Trough. (a) Composite section from the subbasins showing perpendicular extensional structures. (b) (top) Noninterpreted and (bottom) interpreted reflection seismic section from the Makó Trough. For location, see Figure 3. In our interpretation of the pre-Neogene basement and cover, the contact between Békés-Codru and Villány-Bihor units is a Late Cretaceous top NNW thrust contact. White arrows indicate the Miocene kinematics of faults.

a result of the gradual footwall burial. The overall interpretation of Lower Miocene (Karpatian) sediments is in agreement with previous studies [*Lemberkovics*, 2014]. The immediate postrift system tract is in fact built up by another Lower Miocene retrograding facies association that was followed by a latest Early–Middle Miocene prograding facies association during the onset of the postrift system tract (Figure 5). The late postrift deposition was interrupted by the contraction creating the erosional unconformity at the base of Late Miocene and its antiformal geometry (Figure 6). The overlying Pannonian sediments are part of the late postrift system tract, which is subdivided by an unconformity separating Miocene and Pliocene sediments [*Magyar and Sztanó*, 2008]. Near the NW margin of the structure hummocky to chaotic seismic facies units overlying the top Middle Miocene are likely Late Miocene basalts that are well studied in the nearby Kecel volcanic field [e.g., *Pécskay et al*., 2006].

Our interpretation of the tectonic system tracts is in agreement with the sequence stratigraphic interpretation of well logs (Figures 5 and 6). The latter demonstrate a higher-resolution cyclicity than the one depicted by the seismic lines, which is visible by high-resolution trends in the well logs (Figure 5). Along this first-order pattern, the high-resolution prograding and/or retrograding trends can be grouped into a lower order cyclicity that matches the ones derived by the seismic interpretation. During the rift initiation, the coarse sedimentation along the NW flank is gradually replaced by lacustrine shales with periodic influxes of coarse alluvial sediments, more frequent near lower order sequence boundaries, which are likely controlled by the increase in accommodation space by normal faulting (Figure 6). One retrograding-prograding cycle can be observed in the rift initiation system tract, likely underlain by another beneath the depth investigation of the deepest well in the basin center (well 3, Figure 5). The rift climax is when coarser sedimentation is distributed throughout the entire basin, its variability matching the cyclicity observed in seismic lines by prograding-retrograding facies associations. Similarly with the seismic interpretation, well logs suggest four retrograding-prograding cycles that are gradually coarser upward. Inside these sequences, the higher-resolution cyclicity may reflect tectonics or, most likely, higher-order climatic or Milankovitch cyclicity. Furthermore, in agreement with the seismic interpretation, the immediate postrift system tract is made up just by one trangressive facies association, followed by regression during the onset of the late stage postrift, interrupted by the uplift, erosion, and the formation of the regional Middle-Late Miocene unconformity.

4.3. Structures With Peak Synkinematic Deposition During the Middle Miocene

To the north, the Kiskunhalas structure is adjacent to the Tázlár Trough (Figure 7). Wells penetrating the basement and the pre-Neogene sedimentary cover have shown that the Tázlár Trough overlies the thrust contact between the Villány-Bihor and Mecsek nappes [*Haas et al*., 2010]. The high-amplitude seismic reflectors observed in the pre-Neogene sequence along the NNW part and center of this structure are interpreted as the Triassic carbonatic cover of the Mecsek Unit, which was thrust by the south southeastward located Paleozoic basement of the Villány-Bihor nappe (Figure 7). The overall structural geometry shows larger offset low-angle normal faults dipping north northwestward which crosscut the inherited Late Cretaceous Nappe contact and are associated with smaller offset antithetic normal faults dipping south southeastward. The footwalls of the main NNW dipping low-angle normal faults are highly eroded, suggesting exhumation and denudation during extension. The minimum value of the total horizontal offset cumulated along all normal faults is in the order of 4–5 km. The oldest stratigraphic age penetrated by wells in the SSE located main depocenter is lower Middle Miocene (Badenian), while the upper Middle Miocene (Sarmatian) is apparently missing. In this main depocenter the entire synkinematic sedimentation is interpreted as lower Middle Miocene (Badenian) in age, based on lateral correlation of well data available in the prolongation of the trough along its strike. One small offset normal fault crosscuts the Late Miocene sediments along the NNW flank of the main depocenter. This depocenter is flanked on the NNW by a smaller half graben (locally named as the Soltvadkert subbasin, Figure 7) filled with Lower Miocene (Karpatian) conglomerates and sandstones, which are separated from the overlying Pannonian sediments by an erosional unconformity, likely of subaerial origin. This half graben is also slightly inverted by folding in a gentle anticline-syncline structure in a similar fashion as the Kiskunhalas subbasin.

No clear rift initiation could be identified in the main depocenter of the Tázlár Trough, where the sedimentation starts directly with a rift climax system tract (Figure 7). A typical rift initiation system tract with chaotic reflectors accompanying continental sedimentation is observed along the NNW located smaller Early Miocene half graben. This geometry shows that the extension started with a rift initiation system tract located on the flank of the later basin followed by a migration of deformation in the main depocenter where the rift climax is located. Such a spatial arrangement demonstrates a migration in space and time of structures and associated system tracts during extension. In the main depocenter, the rift climax system tract is made up by a retrograding facies association and a retrogradational-progradational cycle. The separating progradational facies association cannot be separated at seismic resolution. The immediate postrift system tract is built up by Late Miocene (Pannonian) sediments. Similar with the Kiskunhalas structure, the erosion of the footwall in the Tázlár subbasin indicates that these are genetic transgressive-regressive sequences. Overlying the rift climax system tract, the Upper Miocene sediments filled the depocenter during the late postrift system tract. In fact, the entire postrift is part of the typical upper prograding shelf margin-slope clinoform system observed in the Great Hungarian Plain. A larger unconformity is observed within this latter sequence, where Pliocene sediments onlap the Miocene prograding slope clinoforms. Interpretation of additional seismic sections from this area shows that the height of the prograding clinoform is much higher above the former Middle Miocene half graben than above the flanking structural highs.

In the eastern part of the Great Hungarian Plain, near the Apuseni Mountains, a series of depocenters with arcuated shape connects a number of subbasins, i.e., Békés, Vésztő, and Derecske (Figure 3). The age of the synkinematic sediments is younger from SW to NE (i.e., from Békés to Derecske). In the SW, wells in the large Békés subbasin have penetrated Middle Miocene synkinematic sediments until ~5 km depth [see *Teleki et al.*, 1994]. Beneath these deposits, the age near the base of the synkinematic succession in this up to 7 km thick subbasin is still unknown.

The Vésztő Trough is a half graben with low-angle normal faults dipping northwestward (Figure 8). Although the number of wells penetrating this structure is relatively low, the lateral correlation from the neighboring Békés and Derecske subbasins indicates that the synkinematic basin fill is Middle Miocene–earliest Late Miocene. This structure is typical for the overall Pannonian Basin; thin (~1.5 km) synkinematic deposition is overlain by remarkably thick (~3.5 km) postrift sediments. The minimum total horizontal offset cumulated along all normal faults is in the order of 6 km. In the middle of the basin a low-offset high-angle reverse fault inverted an earlier normal fault and is associated with a small anticline in the Middle Miocene sediments. Similar with elsewhere, the boundary of Middle and Late Miocene is an erosional unconformity although less clearly observed and constrained.

In the Vésztő Trough, the rift initiation system tract can be observed by the typical low-amplitude, chaotic seismic facies both at the base of the main Miocene depocenter and along its flanks, being truncated by faults in the footwall or separated by an erosional unconformity in the hanging wall (Figure 8). The rift climax system tract is built up by Middle and partly Upper Miocene sediments. In the main depocenter, three higher-order transgressive-regressive cycles and a final retrograding cycle build up this rift climax system tract with clear lobe facies units near the flanks of the half graben. Within the thinner second Middle Miocene cycle no progradation was observed, being likely either below seismic resolution or removed by erosion beneath the base Late Miocene unconformity. This unconformity separates the second from the third cycle and is associated with the small inversional structure. During the Late Miocene another retrogradational-progradational and a final retrograding cycle is observed in the rift climax, overlain by the immediate postrift and a thick late postrift system tract. In the latter, the prograding shelf margin slope (Algyő Formation) is observed by a typical low-amplitude seismic facies, with a progradation direction perpendicular to the orientation of the section. The Miocene/Pliocene boundary is interpreted as a correlative conformity within the delta plain sediments of the late postrift system tract (Figure 8).

4.4. Structures With Peak Synkinematic Deposition During the Late Miocene

The Derecske Trough overlies the NW verging contact between the Villány–Bihor and Mecsek nappes of the Tisza Mega-Unit emplaced during Cretaceous times (Figures 3 and 9). A deep exploration well has penetrated a structural contact where Mesozoic sediments affected by a Cretaceous greenschists metamorphic degree appear in a tectonic window from beneath the overall Paleozoic of the Villány-Bihor nappe. West of this window, this Paleozoic is also affected by this Cretaceous metamorphism, while eastward it retains only the original Paleozoic high degree of metamorphism [*Árkai et al.*, 1998]. A number of thrust sheets have been identified to be associated with this tectonic contact that were reactivated by the Miocene extension when the ~6.5 km deep Derecske subbasin was created [e.g., *Windhoffer et al.*, 2005]. Interesting is that the Paleozoic and Cretaceous metamorphosed sediments are situated in the footwall of the large-offset normal fault system dipping eastward (Figure 9). This implies that the exhumation of these footwall metamorphics is genetically related to the extension, likely along a low-angle detachment cross cut by a system of low-angle listric normal faults formed at later stages during exhumation. The age of these ESE dipping faults gradually migrated in the same direction from Middle Miocene to Late Miocene times. These ESE dipping faults are associated with antithetic WNW dipping higher angle normal faults that migrate in age from Middle Miocene in the center of the subbasin to Late Miocene along the ESE flank (Figure 9). The asymmetry of the synkinematic basin fill indicates higher cumulated offsets for the ESE dipping system when compared with the WNW dipping one. In the upper part of the synkinematic sediments, tilting of the reflectors gradually ceased within the Pannonian strata, suggesting the termination of the normal faulting during Late Miocene times at ~9 Ma. Minimum horizontal displacement of the extensional structures is ~12 km along the section.

Rift initiation sediments are made up of lower Middle Miocene (Badenian) chaotic seismic facies units drilled by wells in the main depocenter and in particular thicker overlying the WNW flank of the subbasin (Figure 9). This indicates a migration in time east southeastward of the normal faulting during extension. The rift climax system tract is made up by a first retrograding-prograding sequence of hummocky and sigmoid facies units organized in an overall progradation that was interrupted by the Middle-Late Miocene unconformity. A second Late Miocene retrogradational, a retrograding-prograding, and a final retrograding subcycle completes the rift climax system tract. Immediate postrift system tract is built up by progradation made up by the turbidites and clinoform facies units. The footwall of the Middle-Late Miocene normal faults is highly eroded due to the gradual migration in space of normal faulting with time. Basin inversion starting around latest Miocene–Pliocene times has crosscut and reactivated the inherited normal fault system as a sinistral negative flower structure. The horizontal displacement along individual fault segments is in the order of a few hundred meters [*Lemberkovics et al.*, 2005]. A few low-offset normal faults are present near the ESE flank of the basin, most probably induced by differential compaction. The Miocene/Pliocene unconformity is situated only a few hundred meters below the surface indicating that there was limited amount of subsidence during the last ~5 Ma.

In the southeastern part of the Great Hungarian Plain three deep subbasins developed in a relatively small area, i.e., Szeged, Makó, and Tomnatec subbasins (Figure 3). The Szeged Trough (or Banatsko Arandelovo in the Serbian prolongation) has a half-graben structural geometry with a low-angle normal fault dipping westward and a smaller offset antithetic normal fault dipping eastward (Figure 10). The highly eroded footwall of the ESE fault bordering the Algyő-High suggests exhumation during extension. Two smaller offset low-angle normal faults might have inverted the former thrusts on the WSW flank of the neighboring Tomnatec subbasin and are associated with Middle and Late Miocene synkinematic sedimentation. The minimum value of the horizontal extensional displacement along all these structures is in the order of 8 km. The synkinematic sedimentation reached ~2.5 km in the center of the Szeged subbasin. The wells drilled on the flanks of the basin have identified a sequence composed of early Middle Miocene (Badenian) shallow-water limestones and Late Miocene clastics, while the upper Middle Miocene (Sarmatian) is missing [*Pigott and Radivojević*, 2010].

The seismic sequence interpretation of the Szeged subbasin shows a Middle Miocene rift initiation in the main depocenter with chaotic, low-amplitude seismic facies units that are also observed on the WSW flank of the neighboring Tomnatec subbasin. Overlying the basal unconformity, a full retrograding-prograding cycle and three upper retrograding facies associations compose the Late Miocene rift climax system tract, in particular well visible by the evolution of lobe facies along both flanks of the subbasin. The retrograding seismic facies units are separated by unconformities. Overlying the near-shore Middle Miocene limestones and the subsequent Late Miocene lobe facies units, an east southeastward progradation pattern with downlap reflection terminations can be observed in the regressive facies association from WNW direction toward the basin which is coeval in time with footwall erosion along the opposite flank of the basin (Figure 10). This infers that the progradation took place in shallow-water conditions. Overlying the upper retrograding facies associations, the immediate postrift system tract has a progradational character that continues also during the late postrift system tract. In fact, the entire postrift is part of the Pannonian prograding shelf margin-slope clinoform system that started during the immediate postrift with the deposition of bottomsets at ~5 Ma [*Magyar et al.*, 2013]. The Pliocene late postrift system tract reaches 1.5 km in thickness. The main fault bordering ESE of the Szeged subbasin has an upward offset in the postrift sequence. This is only apparent, as the offset is created by differential compaction. There are no contractional structures apparent in our seismic interpretation during the transition between Middle and Late Miocene. The overall thickening of the postrift strata eastward are likely related to the higher amount of extension observed in the neighboring Makó-Tomnatec structure. The Miocene/Pliocene boundary was correlated within the immediate postrift turbiditic sediments in this subbasin.

The Makó Trough of SE Hungary (Figure 11) is one of the largest and deepest Neogene subbasins of the Pannonian Basin, the center of the basin attaining a depth of about 7 km. It continues laterally along its strike to the SE and south with the Tomnatec subbasin (Figure 3). Recent studies have suggested that the entire succession in the center of the basin is Upper Miocene and younger in age [*Magyar et al*., 2006; *Szuromi-Korecz et al*., 2004]. Middle Miocene sediments are observed discontinuously by wells drilled over the neighboring highs and are separated by unconformities in their lower and upper parts [*Magyar et al*., 2006]. In the center of the basin near its base, proximal conglomerates, likely related to normal faults activity, were drilled by the well Makó-6 and are of Late Miocene in age (based on nannoflora and dinoflagellates; see *Sztanó et al*. [2013]). On the overall, the geometry of the pre-Neogene subbasin flanks is remarkably symmetric. However, the real asymmetry is visible in the synkinematic basin fill. The main controlling structure is a low-angle listric normal fault located in the SW side that accommodated the subsidence of the Makó subbasin and the uplift of the adjacent Algyő footwall (Figure 11) [see *Tari et al*., 1999]. This uplift was associated with erosion over the structural culmination of the Algyő footwall. Significantly smaller offset normal faults can be observed near the center of the basin, either antithetic or synthetic. In the middle of the basin, one very gentle dipping anticline is likely associated with a minor inversion along a synthetic normal fault [*Sztanó et al*., 2013]. The minimum value of the horizontal extensional displacement along all these structures is about 10 km.

The rift initiation system tract in the Makó subbasin is Pannonian in age, although the few hundred meters at the base of the depocenter are not yet drilled. It is made up of thin conglomerates and sandy siltstones [*Sztanó et al*., 2013] and can be observed at the base of the subbasin as overlain by an unconformity (Figure 11). Within the rift climax system tract three higher-order retrogradational subcycles can be distinguished based on the variable seismic facies and lobes seismic units near its flanks. Progradational features are absent or below the seismic resolution in the upper part of these subcycles. Sourcing to the half graben with sediments eroded from the exhumed Algyő High during the rift climax system tract took place until ~9–8 Ma, when this high was submerged [*Magyar et al*., 2006]. This means that the higher-order cycles of the rift climax system tract are in fact genetic TR sequences. The onset of the immediate postrift system tract is interpreted as the first progradational facies association. The Makó subbasin was ultimately filled by thick prograding shelf slope sediments at ~5.7 Ma [*Magyar et al*., 2013; *Sztanó et al*., 2013], followed by deltaic and alluvial plain sedimentation. The Miocene/Pliocene boundary is a correlative conformity in the area of this subbasin within the alluvial and delta plain environments.

The flanks of the NNW-SSE oriented Makó subbasin is crosscut by a smaller WSW-ENE oriented normal fault that created the smaller Földeák subbasin filled most likely by synkinematic Middle Miocene sediments overlain unconformably by Upper Miocene deposits (Figure 11). These two orientations may be related to two extensional directions, the older Middle Miocene one reactivating a system of WSW-ENE oriented Cretaceous thrusts with NNW vergence [*Tari et al*., 1999]. In the older basin, the seismic facies is rather uniform and difficult to separate in individual units, characterized by high-amplitude, discontinuous reflectors.

5. Seismic Sequence Stratigraphy in the Great Hungarian Plain and Phenomenological Inferences

The spatial and temporal distribution of seismic facies units is conditioned by episodic activity of normal faults in various subbasins (Figure 12). The chaotic seismic facies make up the rift initiation system tract in the oldest unit observed either at the base of the entire synkinematic succession (e.g., Early Miocene rift initiation of the Kiskunhalas and Middle Miocene rift initiation of the Szeged subbasins, Figures 6 and 10, respectively) or over the flanks of the separating highs (e.g., Tázlár and Derecske subbasins, Figures 7 and 9, respectively). The latter position is absent in typical sequence stratigraphic models of extensional grabens [e.g., *Martins-Neto and Catuneanu*, 2010; *Prosser*, 1993] and is a result of gradual migration of extension in time from the position of the present-day flank to the center of the subbasin (Figure 12). The sedimentological environment of the rift initiation system tract in our observed extensional subbasins is also different when compared with these earlier models. Continental conditions are observed only when the rift initiation system

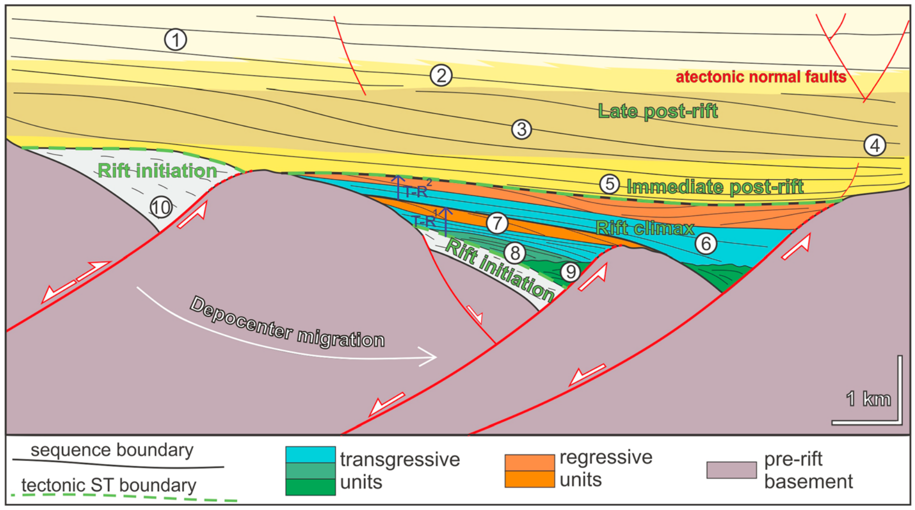

Figure 12. Simplified stratigraphic model of the half grabens of the Pannonian Basin [after *Prosser*, 1993; *Ter Borgh*, 2013]. Numbers represent the characteristic seismic facies units of a half graben; cf. Figure 4. The characteristic retrograding and prograding seismic facies units are grouped into transgressive-regressive cycles. These cycles are grouped into lower order tectonic system tracts, representing phases of basin evolution such as rift initiation, rift climax, immediate postrift, and late postrift. Note the features of these half grabens, such as migration of depocenters and the exhumation and erosion of the footwalls of active low-angle listric normal faults.

tract is entirely of Lower Miocene in age (e.g., Soltvadkert subbasin, Figure 7), when the Central Paratethys domain of the Pannonian Basin was separated from the marine realm. Wherever the onset of extension took place during Middle–Late Miocene, the rift initiation system tract contains the record of a rapid transition from continental to shallow-marine conditions (e.g., the Szeged subbasin, Figure 10). This means that the sedimentological environment of the rift initiation system tract is dependent on the regional presence or absence of connections between the half graben and the marine realm rather than the evolution of the local subbasin. No retrograding-prograding (and therefore no transgressive-regressive) cyclicity could be defined in the rift initiation sequence at the seismic details, most likely because the organization of sediments in these cycles is beneath the seismic resolution. However, such patterns are rather clear in the well logs sequence stratigraphy (Figure 4) that shows cyclicity also within the rift initiation system tract.

The evolution of sedimentation during the rift climax system tract (Figure 12) is directly controlled by normal fault offsets, creating gravity-driven deposits in proximal areas, such as lobes seismic facies units, derived directly from the exhumed hanging wall and, more importantly, from the erosion of the uplifting footwalls. The progradational, retrogradational, or aggradational character of these lobes is one major criterion to distinguish the high-resolution cyclicity. In more distal positions, (sub)parallel to divergent seismic facies units characterize the basin fill associated with sigmoidal seismic facies units during periods of regression. In the final part of the rift climax system tract the footwall subsides and is gradually covered by sediments, decreasing its importance as a source area. Finally, during Late Miocene times the subparallel to hummocky turbiditic seismic facies units fill the large accommodation space created by a subsidence that is more regional than the scale of individual subbasins, followed by the deposition of the clinoform seismic facies units of the prograding shelf margin slope. The (sub)parallel continuous and fairly continuous seismic facies units in the upper part of the basin indicate the gradual fill of the basin and deltaic and alluvial plain sediments (Figure 12).

Footwall erosion was coeval with the deposition of almost all retrogradational-progradational cycles. This indicates that erosion combined with the correlative maximum regression surface defined by the geometry of the seismic facies units is an expression of the composite surface that bounds a transgressive-regressive (TR) sequence (Figure 12). The exception is locally the last retrogradation of the rift climax or the subsequent progradation of the immediate postrift system tracts, where footwall erosion was reduced or absent. The transgressive facies associations are made by (sub)parallel to divergent continuous to hummocky seismic facies units onlapping both the footwall and hanging wall, intercalated with large amounts of lobe seismic

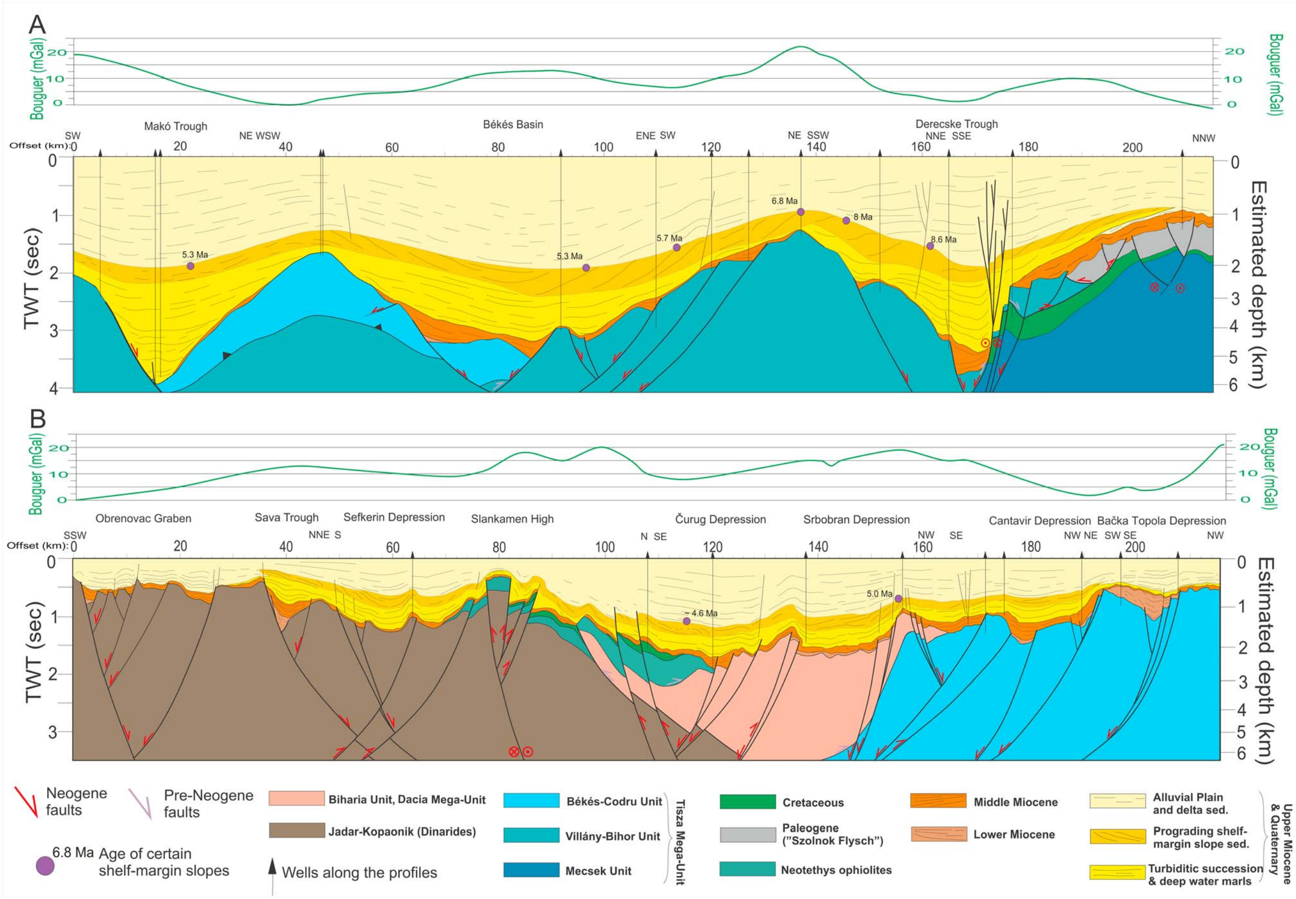

Figure 13. Interpreted composite reflection seismic transects from the Pannonian Basin showing the main tectonic and stratigraphic features of the area. For location, see Figure 1. Note the striking difference of the thickness of Early to Middle Miocene and Late Miocene deposits. Late Miocene succession is subdivided based on the characteristic time transgressive environments of Lake Pannon (age of the prograding shelf edges derives from *Magyar et al*. [2013]). (a) Section from the eastern part of the Pannonian Basin is modified after *Balázs et al*. [2013]. It shows Middle and Upper Miocene synkinematic deposition in the (half) grabens. Bouguer gravity anomalies imply the asymmetry of the highly extended Makó Trough and Békés Basin, where local gravity minimum corresponds to the basement high, while positive values characterize the deep basins (for detailed modeling see *Szafián and Horváth* [2006], *Király et al*. [2012], and *Tari et al*. [1999]). (b) Section from the southern and western parts of the Pannonian Basin is modified after *Matenco and Radivojević* [2012], where syntectonic deposition is Lower to Middle Miocene. Neotectonic inversional structures have formed, for instance, at offset 80, 135, and 170 km in Figure 13b. At the southern part of this section the low Bouguer anomaly values correspond to the crustal "root" of the Dinarides.

facies units. The latter results from the large rate of displacement along normal faults creating high-dip slopes. Periods of reduced rate of offset along normal faults are associated with the regressive facies associations, made up mainly by offlaps and prograding seismic reflection patterns. In the case of Kiskunhalas, Vesztő, and Szeged subbasins (Figures 6, 8, and 10), delta systems can be observed prograding over the hanging wall toward the center of the subbasins during the deposition of the regressive facies associations. The transgressive-regressive sequence boundaries are marked by toplap and overlying downlap reflection terminations. This higher-order transgressive-regressive cyclicity is characteristic for the Early and Middle Miocene rift climax sequences but apparently less developed within Late Miocene rift climax deposits.

In the central part of the Great Hungarian Plain the late Middle Miocene (Sarmatian) sediments are very thin or completely missing beneath the Middle-Late Miocene unconformity, which has an erosional character outside the deepest Middle Miocene (half) grabens. The unconformity is often associated with contractional structures showing N-S compressional direction, such as gentle symmetric anticlines of small inversion of earlier normal faults (e.g., Kiskunhalas or Vésztő subbasins, Figures 6 and 8). This unconformity is also visible in the WNW part of the Pannonian basin [*Tari et al*., 1992]. On the contrary, the late Middle Miocene is thick near the Pannonian basin margins bordering the Serbian Dinarides, Alps, and the Carpathians (Figure 14) [*Magyar et al.*, 1999b; *Pavelić*, 2001; *Kováč et al.*, 1995].

Starting from the erosional unconformity at the Middle/Late Miocene boundary, the water depth of Lake Pannon increased during the early Late Miocene, as indicated by the height of the subsequent prograding Pannonian shelf margin slopes between ca. 10 and 4 Ma in the Great Hungarian Plain [*Magyar et al.*, 2013]. Previous calculations indicate several hundreds of meters, possibly up to 1 km paleobathimetries [*Balázs et al.*, 2013]. In the peripheral areas of the Pannonian Basin, like the Nyírség subbasin in the NE and in the vicinity of the southern coastline of Lake Pannon, delta and alluvial plain environments remained characteristic during the entire Late Miocene–Pliocene evolution of the Great Hungarian Plain, where sedimentation kept pace with the rate of creating accommodation space (Figures 3 and 13). This means that in the deepest Late Miocene half grabens, situated farther away from source areas, shallow-water sediments build up only the oldest Pannonian synkinematic succession. The rate of creating accommodation space (i.e., rapid relative lake level rise) was significantly higher when compared with the rate of sediment supply. Therefore, these thick Late Miocene rift climax system tracts situated in a distal position in the Great Hungarian Plain relative to the source area (e.g., Makó Trough; Figure 11) are dominantly built up by retrograding facies associations. In such situations, it is possible that a large part of the Pannonian sedimentation postdated the normal faulting, burying some part of the controlling normal fault with deep water deposition during postkinematic times.

The immediate postrift system tract was coeval with the cessation of fault offsets, with the rate of sediment supply being generally higher than the rate of creating accommodation space. This can be recognized by subparallel, fairly continuous, occasionally gently divergent reflections (Figure 5). The onset of postrift deposition was diachronous in the various subbasins and is not marked by a clear erosional unconformity marking a transition to postrift sedimentation [see *Tari et al.*, 1999]. Driven by regional contractional processes [*Horváth*, 1995], the juxtaposition of the Middle/Late Miocene regional unconformity with a general synrift/postrift boundary would be a coincidence, but this was not observed in every studied subbasin.

The late postrift system tract deposits of the Pannonian Basin are made up by turbiditic deposits, prograding shelf margin slope, and alluvial plain and delta sediments (Figures 5 and 12). During the late postrift system tract the cyclicity was controlled by other regional processes than the local scale of the extensional subbasins, such as absolute lake level variations or the late stage basin inversion [*Csató et al.*, 2015; *Magyar and Sztanó*, 2008; *Sacchi et al.*, 1999]. Interesting is that the large-scale Pannonian progradation contains thicker prograding clinoforms above the former extensional basins, most likely due to differential compaction being active during deposition. This compaction creates also offsets in the postrift sequence overlying the earlier normal faults [e.g., *Teleki et al.*, 1994], which typically increase upward in the stratigraphy.

6. Discussion

Our study demonstrates that the Pannonian Basin system has undergone multiple phases of extension and basin inversion during its Miocene evolutionary history. Different subbasins were affected by different amounts of deformation at various times, as a function of rheological variations, inherited weakness zones, and degree of extensional asymmetry (Figure 14).

6.1. Tectonosedimentary Evolutionary Model of a Highly Extended Back-Arc Basin

The seismostratigraphic interpretation infers a clear interplay between sedimentation and tectonics during the Miocene synkinematic deposition in the Great Hungarian Plain part of the Pannonian Basin. The type and amount of sedimentation was variable in time and space and resulted in different sedimentary facies units deposited at first in continental alluvial to lacustrine, shallow-marine to pelagic, and locally deep water environments, and ultimately back to continental lacustrine to alluvial sedimentation. Although discussed in previous studies, our analysis is the first basin-scale demonstration that the extension in the Great Hungarian Plain part of the Pannonian Basin migrated in time and space throughout the Miocene.

The activity of the extensional subbasins and associated detachments started near or in the Dinarides already during Early Miocene times, or possibly even earlier during the Oligocene [e.g., *Toljić et al.*, 2013], being accompanied by the deposition of thin continental alluvial to lacustrine sediments [*Krstić et al.*, 2003; *Matenco and Radivojević*, 2012]. Interestingly, the amplitude of extensional exhumation of the footwalls

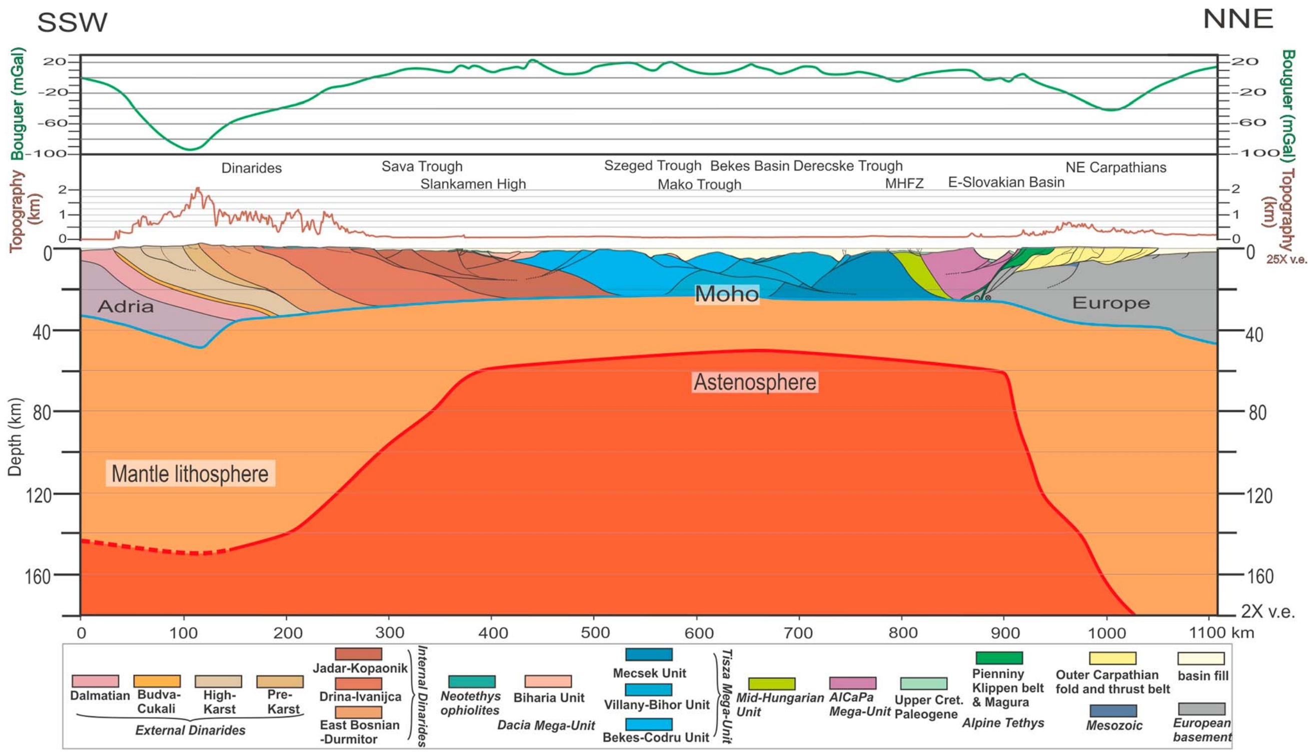

Figure 14. Generalized lithospheric scale cross section over the Dinarides-Pannonian Basin-NW Carpathians. Lithosphere-asthenosphere boundary compiled from *Artemieva et al.* [2006], *Tari et al.* [1999], and *Tašárová et al.* [2009]. Moho depth derives from *Horváth et al.* [2006], *Janik et al.* [2011], and *Sumanovac* [2010]. Tertiary tectonic interpretation is based on this study, *Csontos and Vörös* [2004], *Gagała et al.* [2012], *Matenco and Radivojević* [2012], *Roure et al.* [1993], and *Schmid et al.* [2008].

was much higher than the tectonic subsidence of the associated subbasins during the Early Miocene. Our data show that the Early Miocene direction of extension was NW-SE oriented (Figure 3), such as observed in the Kiskunhalas subbasin.

In agreement with all previous studies, our data show that the Middle Miocene was the peak period of extensional subsidence of the Pannonian Basin, when most of the half grabens of the Great Hungarian Plain accommodated maximum hanging wall deposition. The extension had variable offsets and extensional directions, in general N-S in the southern and westernmost parts and NW-SE in the central part (Figure 3). The variable extensional directions were likely related to coeval vertical axis clockwise rotations during the overall northeastward to eastward translations (Figure 15).

The anomalous pattern of the distribution of late Middle Miocene (Sarmatian) sediments might be related to different processes, but one interesting feature is its wavelength of ~400 km. This geometry can be interpreted either as complete removal of late Middle Miocene strata by the inversion in the central part of the basin [*Horváth*, 1995] or by gradual footwall exhumation leading to erosion and/or nondeposition as suggested in areas near the Dinarides [*Matenco and Radivojević*, 2012]. This can be also interpreted as an effect of basin-scale uplift in the center of the Great Hungarian Plain and coeval subsidence of its peripheral areas. This interpretation is in agreement with the observation that in the center part of the Pannonian Basin shallow-water environment was dominant in contrast to the early Middle Miocene higher water depth. Contractional structures are also present, but their amplitude cannot justify the erosional removal of a kilometer-thick sequence. Therefore, the unconformity formed likely as a combination of the previously mentioned mechanisms. The N-S oriented direction of contraction is in contrast with the inferred eastward coeval movement of Carpathian units. This is also the peak moment of coeval clockwise rotations in the Tisza-Dacia units (Figure 2), whose effects in the various subbasins of the Great Hungarian Plain are largely unknown.

The extension continued during the Late Miocene times. Although various normal faults with variable offsets have been previously described [*Fodor et al.*, 2013; *Balázs et al.*, 2013; *ter Borgh*, 2013], our study demonstrates

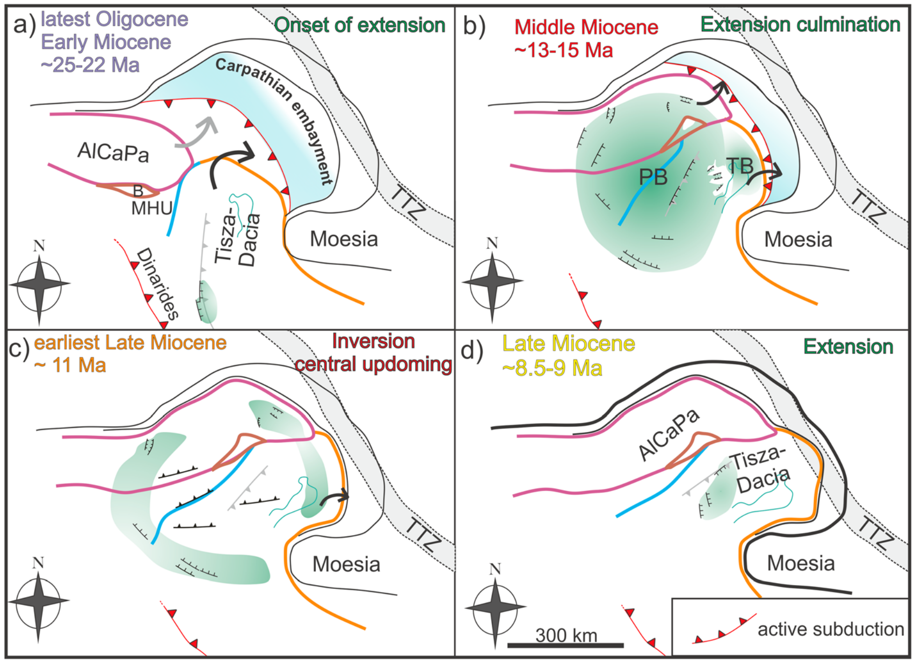

Figure 15. Simplified sketch showing the synrift evolution of the Pannonian Basin [after *Csontos and Nagymarosy*, 1998; *Faccenna et al.*, 2014; *Fodor et al.*, 1999; *Ustaszewski et al.*, 2008]. Green color shows the area of synrift subsidence. Coeval vertical axis clockwise rotation of the Tisza-Dacia and CCW rotation of the AlCaPa Mega-Units and the Carpathians and Dinaridic slab rollback-related extension resulted in several deep half grabens in the Great Hungarian Plain with variable strikes during the Miocene. Internal deformation accommodated the different amounts of rotation in various parts of the mega-units. A short interruption in extension is inferred at the onset of the Late Miocene times, when N-S compressional structures are observed. Note the striking difference of subsiding areas during the early Middle Miocene (ca. 15–13 Ma) and the latest Middle Miocene to earliest Late Miocene (ca. 12–11 Ma). PB = Pannonian Basin, TB = Transylvanian Basin, B = Bükk Unit, MHU = Mid-Hungarian Unit, TTZ = Teisseyre-Tornquist Zone (Trans-European Suture Zone, location after *Malinowski et al.* [2013] and *Pharaoh* [1999]).

that this extension had major effects by creating or significantly enlarging some of the deepest subbasins in the Great Hungarian Plain, such as Makó, Szeged, or Derecske with an average E-W extensional transport direction. This was followed by large-scale postrift subsidence that was diachronous and buried most of the subbasins beneath 2–3 km thick deposition of sediments.

The overall extensional directions presently observed in the Great Hungarian Plain were obviously affected by the gradual clockwise rotation of the Tisza-Dacia block [e.g., *Balla*, 1987]. Earlier Miocene structures recorded larger amounts of rotation when compared with the later ones. In fact, there is just one main ENE-WSW extensional direction in the entire Great Hungarian Plain that becomes more N-S near the Dinarides and more E-W near the Apuseni Mountains [*Csontos and Nagymarosy*, 1998; *Fodor et al.*, 1999] following the clockwise geometry of their relative rotations. All other orientations reflect subsequent rotations after deformation (Figure 15). This also explains why different orientations are observed in the same area, such as the early Miocene NE-SW strike of Kiskunhalas or Földeák subbasin versus the NW-SE strike of Late Miocene Szeged or Makó subbasins (Figures 3 and 11a). Such high degrees of vertical axis rotation during extension are also observed in other back-arc basins, for instance, at the Alboran domain of the Gibraltar arc, the Aegean Sea at the Hellenic Trench, or the Caribbean region at the Lesser Antilles Trench [*Faccenna et al.*, 2014; *Govers and Wortel*, 2005].

The first structures of the last stage of tectonic inversion formed during the latest Miocene times near the Dinarides such as in the SW Zala Basin (Figure 1) [*Uhrin et al.*, 2009] or in the southern Serbian part (Figure 13b) at ~7.5–8 Ma. This means that extension was still active in the eastern part of the Pannonian Basin (Szeged, Makó, and Derecske), when the onset of contractional deformation took place near the Dinarides. Therefore, the last stage of inversion also migrated in space and time from the south and SW margin of the Dinarides northward (i.e., from the Adriatic indenter) toward the central Pannonian Basin.

The peak contractional event took place at the end of Miocene or earliest Pliocene, caused likely by the northward drift and counterclockwise (CCW) rotation of the Adriatic microplate [*Pinter et al*., 2005]. It has resulted in a clear unconformity near the Miocene/Pliocene boundary with major angular aspect in various places in the basin (e.g., Tázlár subbasin, Figure 7), being replaced laterally with a correlative conformity in deeper subbasins [*Magyar and Sztanó*, 2008].

6.2. Extensional Detachments Versus Low-Angle Normal Faults in the Great Hungarian Plain

One interesting feature of the Pannonian extensional structures is the relative low-angle dip of fault planes that is variable from ~20° (Makó subbasin) to ~30° (Szeged subbasin), or up to ~40° (Derecske subbasin). This was facilitated by the reactivation of the former low-angle Cretaceous thrusts as observed by seismic and well data in our or previous studies (e.g., the Derecske subbasin, Figure 9; see also *Windhoffer et al*. [2005]). This is in particular obvious for the presently northwestward vergent Turonian nappe contacts of Tisza unit. The extensional mechanism is clearly asymmetric, being linked with the activity of controlling low-angle normal faults or extensional detachments that resulted in significant erosion during the relative uplift of footwalls. In other situations, the low-angle normal faults crosscut preexisting thrusts (such as the Tazlar subbasin, Figure 7), most likely because the inherited thrusting geometry was not favorable for reactivation. Extensional detachments are widely known near the Pannonian Basin margins along or inside the Eastern Alps or the Dinarides [e.g., *Ustaszewski et al*., 2010; *van Gelder et al*., 2015]. Such detachments exhumed in their footwall previously metamorphosed Mesozoic rocks during the Cretaceous–Paleogene nappe stacking. In the studied area of the Great Hungarian Plain, only the WNW part of the Derecske subbasin shows such Paleozoic–Mesozoic basement and sediments affected by a Cretaceous greenschist metamorphic degree [*Árkai et al*., 1998] (Figure 9). This area may satisfy the possible conditions of a ductile shear zone exposed in the footwall of an extensional detachment that could eventually define a core-complex type of structure. All other subbasins show nonmetamorphosed Mesozoic or Paleogene sediments in the immediate footwall of the controlling structures and, therefore, such structures should be considered as low-angle normal faults with footwall exhumation controlling half grabens.

Such an interpretation includes likely the Algyő High that is situated in the footwall of the controlling structure in both the Szeged and Makó subbasins (e.g., Figure 10), previously interpreted as a Miocene metamorphic core complex based on a preliminary Early Miocene zircon fission track age [*Tari et al*., 1999]. Wells penetrating this high have identified a nonmetamorphic Triassic carbonatic sequence that is in structural contact with rocks metamorphosed during Cretaceous times [*Lelkes-Felvári et al*., 2005], while the wells penetrating the pre-Neogene sequence on the western flank of the Szeged subbasin have encountered Paleozoic metamorphics and a Triassic sequence of the Békés-Codru nappe of the Tisza unit. The Cretaceous metamorphosed rocks were assigned to the Biharia nappe of the Apuseni Mountains, thrusted probably during Turonian times over the neighboring Bekes-Codru nappe that retains a Variscan age metamorphism and is covered by a nonmetamorphosed Mesozoic sequence (Figure 10) [*Schmid et al*., 2008]. The alternative interpretation suggests that Algyő High contains a part of Tisza overprinted by a pressure-dominated eo-Alpine, amphibolite facies metamorphism, likely a window of a deeper nappe unit [*Lelkes-Felvári et al*., 2005], probably a tectonic window of Villány-Bihor unit (Figure 11). Whichever interpretation is favored, these all assume large exhumation of metamorphic units in the Algyő High predating the Miocene extension. This was subsequently followed by ~7–8 km of Miocene exhumation, which is roughly above the 220° zircon fission track annealing temperature and below the metamorphic threshold, given the high geothermal gradient of the Pannonian Basin. Alternatively, the Cretaceous metamorphism of the Algyő High may be explained as a Miocene extensional tectonic window beneath a detachment, but such an interpretation is difficult to accommodate in the current image of the basement structure [*Haas et al*., 2010].

Such significant exhumation can be derived in the footwall of all main controlling structures of the studied subbasins, obviously controlled by the asymmetry of extension. Given the amounts of footwall erosion, block tilting, and correlation markers across these structures, we can estimate the amount of exhumation between 2 and 7 km, significantly higher in the case of the Derecske detachment, where a minimum of ~10 km is a reasonable estimation. These amounts of exhumation are locally comparable with or higher than thickness of the synkinematic basin fill deposited in their hanging walls. This type of exhumation increases in local detachments and core complexes toward the Eastern Alps and Dinarides margins of the Pannonian Basin. Therefore, the amount of extension along various controlling structures is in reality much higher than looking solely on the synkinematic basin fill. The 150–180 km of extension estimates in the Tisza-Dacia part of the Great Hungarian Plain are based on a subsidence restoration procedure that took into account only the synkinematic sedimentation combined with crustal and lithospheric attenuation [*Lenkey*, 1999]. This means that the amount of crustal extension in the Pannonian basin is much higher than previously thought, at least with 50% given the 20–40° average normal faults dip. This leads to a total amount in the order of 220–270 km along a NE-SW transect that crosses the deepest subbasins (Figures 13 and 14). Such rough calculations can become quantitative whenever thermochronology would become more widely available in the footwall of controlling extensional structures.

6.3. Inferences for the Regional Geodynamics

The subduction rollback of the Carpathian and possibly the Dinaridic slab was associated with dynamic mantle evolution and associated topography in the Pannonian Basin [e.g., *Burov and Cloetingh*, 2009; *Horváth et al*., 2015]. Beyond possible heterogeneities created by the inherited nappe structure, the anticorrelation of the basement depth with the Bouguer gravity anomaly (Figure 13a) has been previously observed [*Szafián and Horváth*, 2006]. In our view, such anticorrelation does not reflect a heterogeneous mantle structure but rather the lateral shift of stretching at various crustal or upper lithospheric mantle depths, controlled by detachments and/or low-angle normal faults. Such a geometry is anyway required at the crustal level by the observed detachments near the Eastern Alps and Dinarides [*Tari et al*., 1992; *Ustaszewski et al*., 2010].

When combining our study of the Great Hungarian Plain with the regional structure of the Dinarides and Carpathians along a NE-SW oriented transect (Figure 14), a number of critical inferences can be derived. The Carpathians kinematics assume that the upper units translated northeastward and eastward during Miocene times in the absence of absolute plate motions, the shortening at the exterior being accommodated entirely by back-arc extension [*Horváth et al*., 2015; *Matenco et al*., 2016]. In other words, the Africa-Europe convergence in the Tisza-Dacia sector of the Pannonian Basin was retained entirely in the Dinarides during the extension of the Pannonian Basin. Farther northward, the unstretched parts of the Tisza-Dacia units simply translated northeastward and eastward, shortening the external Carpathians nappes at the exterior and collapsing by extension of the Pannonian back arc. In such a restricted lithospheric configuration (Figure 14) there is no space for significant amounts of oceanic subduction accommodating absolute plate convergence. The overall shortening, collision, and extension must have affected dominatly a continental lithospheric domain. These processes were obviously driven by the evolution of inherited subducted slabs, as derived by teleseismic tomography [e.g., *Bennett et al*., 2008; *Martin and Wenzel*, 2006; *Wortel and Spakman*, 2000].

The overall pattern of large-scale erosion in the center of the Pannonian Basin and significant subsidence and deposition near its margins during the transition from Middle to Late Miocene has a large-scale wavelength with hundreds of kilometers that suggests deep lithospheric mechanism. One can speculatively link this with dynamic topography mechanisms, such as an active mantle upwelling creating a 3-D lithospheric folding in the Pannonian basin. This would result in accelerated subsidence rates in the peripheral areas and updoming in the central parts, thus decreasing the accommodation space in the Great Hungarian Plain. Such a mantle upwelling can be related to subduction-induced poloidal mantle flow, created by the Carpathian slab [e.g., *Funiciello et al*., 2006], or a passive extension induced upper mantle upwelling [*Huismans et al*., 2001] or can be the effect of a deep mantle plume [e.g., *Burov and Cloetingh*, 2009]. A clear discrimination of these mechanisms requires further process-oriented modeling.

The latest Miocene–Quaternary inversion of the Pannonian Basin has higher amplitudes on individual fault structures in the western part of the Tisza-Dacia block, near the basin margins (Figure 13b). However, the regional pattern is the one of wide open antiforms and synforms at the scale of the entire Great Hungarian Plain (Figure 13a). The induced vertical movements created shallow positions of the Miocene synkinematic basin fill (e.g., Derecske Trough, Figure 9) and also their significant burial (such as in the Makó or Szeged subbasins, Figures 11 and 10). Our study in the Great Hungarian Plain is in agreement with the previously inferred large-scale lithospheric folding due to the northward push of the Adriatic indentation into the rheologically weak Pannonian Basin that created various wavelengths in contrasting rheologies [e.g., *Dombrádi et al*., 2010; *Horváth and Cloetingh*, 1996; *Jarosinski et al*., 2011].

7. Conclusions

Our interpretation of seismic data from the Great Hungarian Plain of the Pannonian Basin corroborated with calibrating wells and correlated with available studies has led to a novel image of the extensional mechanism in the Pannonian Basin. The back-arc extension took place at high rates with dominantly asymmetric mechanism and has resulted in the formation of a significant number of subbasins separated by uplifted basement highs. Our study demonstrates for the first time that the extension of the entire Great Hungarian Plain was diachronous and migrated in space and time across the basin. It started during the Early Miocene, and significant deformation was still taking place until 9 Ma. The evolution of these subbasins was controlled by low-angle normal faults or detachments that resulted in significant footwall exhumation. The overall extensional direction remained roughly constant through time, presently observed as NNE-SSW near the Dinarides (e.g., Dráva and Sava subbasins) to E-W in the eastern (e.g., Szeged subbasin) and NW-SE in the northern central Great Hungarian Plain (e.g., Derecske subbasin). The gradual and large amount of clockwise rotation of the Tisza-Dacia Mega-Unit significantly modified the original geometry of the subbasins. The extensional mechanism was controlled by the preexistence of the Cretaceous nappe stack that exerted a fundamental control on the (re)activation of detachments or low-angle normal faults.

The combined seismic and well logs sequence stratigraphy has resulted in the novel definition of a tectonic system tracts model in the Pannonian asymmetric extensional basins. This model is able to detect the succession of higher-resolution individual offsets in the synkinematic basin fill along major controlling extensional structures.

The combined kinematic and depositional model at the scale of the entire basin infers that the cumulated amounts of Miocene extension were much higher than previously thought, reaching approximately 220–270 km. The contraction and associated vertical movements observed near the boundary between the Middle and Late Miocene show differential distribution patterns in the upper crustal structure of the Pannonian Basin, indicating a dynamic topography response. Similar potential dynamic topography mechanisms [e.g., *Burov and Cloetingh*, 2009; *Houseman and Gemmer*, 2007; *Horváth et al*., 2015] still have to be quantified. The effects of the subsequent latest Miocene–Pliocene contraction migrated in space and time from the Adriatic to the Carpathian margins of the basin.

All these inferences show that in terms of the upper crustal geometry correlated with the overall lithospheric configuration, the Pannonian Basin should be considered rather a hyperextended back arc, its formation and evolution being strongly controlled by inherited orogenic asymmetries, subsequent slab dynamics, and dynamic topography mechanics.

Acknowledgments
This study was financed by the Netherlands Centre for Integrated Solid Earth Science (ISES) and Eötvös Loránd University, Budapest, in a collaborative study of the Pannonian Basin. MOL Plc., TXM Ltd., and RAG Ltd. are acknowledged for providing seismic and well data. The first and fourth authors (A.B. and F.H.) are grateful to the academic support of the Hungarian Science Foundation (OTKA NK83400). This is MTA-MTM-ELTE Paleo contribution No. 221. László Fodor, Gábor Bada, and Endre Dombrádi are thanked for the stimulating discussions about the tectonics of the area. Orsolya Sztanó is thanked for insights on the sedimentological evolution of the basins. Viktor Lemberkovics is acknowledged for discussions on the Kiskunhalas subbasin. We thank Bruno Saftić and an anonymous reviewer for their detailed and constructive remarks. The data supporting this paper are available by contacting the corresponding author.

References

Árkai, P., A. Bérczi-Makk, and D. Hajdu (1998), Alpine prograde and retrograde metamorphism in an overthrusted part of the basement, Great Plain, Pannonian Basin, Eastern Hungary, *Acta Geol. Hung.*, *41*(2), 179–210.

Artemieva, I., H. Thybo, and M. K. Kaban (2006), Deep Europe today: Geophysical synthesis of the upper mantle structure and lithospheric processes over 3.5 Ga, in *European Lithosphere Dynamics*, *Geol. Soc. Mem.*, vol. 32, edited by D. Gee and R. Stephenson, pp. 11–41.

Bada, G., F. Horváth, P. Dövényi, P. Szafián, G. Windhoffer, and S. Cloetingh (2007), Present-day stress field and tectonic inversion in the Pannonian basin, *Global Planet. Change*, *58*(1–4), 165–180.

Balázs, A., I. Magyar, and F. Horváth (2013), Stratigraphic and structural interpretation of regional seismic sections from the Pannonian Basin, 14th RCMNS Congress, Istanbul, Abstract Book p. 191.

Báldi, K., L. Benkovics, and O. Sztanó (2002), Badenian (Middle Miocene) basin development in SW Hungary: Subsidence history based on quantitative paleobathymetry of foraminifera, *Int. J. Earth. Sci.*, *91*, 490–504.

Báldi, T. (1986), *Mid-Tertiary Stratigraphy and Paleogeographic Evolution of Hungary*, 201 pp., Akadémiai Kiadó, Budapest.

Balla, Z. (1986), Palaeotectonic reconstruction of the central Alpine-Mediterranean belt for the Neogene, *Tectonophysics*, *127*, 213–243.

Balla, Z. (1987), Tertiary palaeomagnetic data for the Carpatho-Pannonian region in the light of Miocene rotation kinematics, *Tectonophysics*, *139*, 67–98.

Bennett, R. A., S. Hreinsdóttir, G. Buble, T. Bašic, Ž. Bacic, M. Marjanović, G. Casale, A. Gendaszek, and D. Cowan (2008), Eocene to present subduction of southern Adria mantle lithosphere beneath the Dinarides, *Geology*, *36*, 3–6.

Bérczi, I., V. Dank, S. Gajdos, S. Pap, I. Révész, K. Szentgyörgyi, and L. Völgyi (1987), Ablagerungen der Kunság-Stufe (Pannonien s. str.) auf der Grossen Ungarischen Tiefebene (Translated title: Depositional environments of the Kunság stage of the Great Hungarian Plain), *Ann. Hung. Geol. Inst.* , *69*, 179–211.

Brun, J.-P., and C. Faccenna (2008), Exhumation of high-pressure rocks driven by slab rollback, *Earth Planet. Sci. Lett.*, *272*, 1–7.

Brun, J.-P., and D. Sokoutis (2007), Kinematics of the Southern Rhodope Core Complex (North Greece), *Int. J. Earth Sci.*, *96*, 1079–1099.

Burov, E., and L. Guillou-Frottier (2005), The plume-head continental lithosphere interaction using a tectonically realistic formulation for the lithosphere, *Geophys. J. Int.*, *58*, doi:10.1111/j.136.-246x.2005.02588.x.

Burov, E., and A. Poliakov (2003), Erosional forcing of basin dynamics: New aspects of syn- and post-rift evolution, *Geol. Soc. London Spec. Publ.*, *212*, 209–223.
Burov, E. B., and S. A. P. L. Cloetingh (2009), Controls of mantle plumes and lithospheric folding on modes of intraplate continental tectonics: Differences and similarities, *Geophys. J. Int.*, *178*, 1691–1722.
Catuneanu, O. (2002), Sequence stratigraphy of clastic systems: Concepts, merits, and pitfalls, *J. Afr. Earth Sci.*, *35*, 1–43.
Catuneanu, O., et al. (2009), Towards the standardization of sequence stratigraphy, *Earth Sci. Rev.*, *92*, 1–33.
Catuneanu, O., W. E. Galloway, C. G. S. C. Kendall, A. D. Miall, H. W. Posamentier, A. Strasser, and M. E. Tucker (2011), Sequence stratigraphy: Methodology and nomenclature, *Newslett. Stratigr.*, *44*, 173–245.
Cloetingh, S., and B. U. Haq (2015), Inherited landscapes and sea level change, *Science*, *347*(6220).
Comas, M. C., V. Garcias, and M. J. Jurado (1992), Neogene extensional tectonic evolution of the Alboran Basin from MCS data, *Geomar. Lett.*, *12*, 157–164.
Csató, I. (1993), Neogene sequences in the Pannonian Basin, Hungary, *Tectonophysics*, *226*, 377–400.
Csató, I., S. Tóth, O. Catuneanu, and D. Granjeon (2015), A sequence stratigraphic model for the Upper Miocene–Pliocene basin fill of the Pannonian Basin, eastern Hungary, *Mar. Pet. Geol.*, *66*, 117–134, doi:10.1016/j.marpetgeo.2015.02.010.
Csontos, L. (1995), Tertiary tectonic evolution of the intra-Carpathian area: A review, *Acta Vulcanol.*, *7*, 1–13.
Csontos, L., and A. Nagymarosy (1998), The mid-Hungarian line: A zone of repeated tectonic inversions, *Tectonophysics*, *29*, 51–71.
Csontos, L., and A. Vörös (2004), Mesozoic plate tectonic reconstruction of the Carpathian region, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *210*, 1–56.
Cvetkovic, V., H. Downes, D. Prelevic, M. Lazarov, and K. Resimic-Saric (2007), Geodynamic significance of ultramafic xenoliths from Eastern Serbia: Relics of sub-arc oceanic mantle?, *J. Geodyn.*, *43*, 504–527.
de Leeuw, A., S. Filipescu, L. Maţenco, W. Krijgsman, K. Kuiper, and M. Stoica (2013), Paleomagnetic and chronostratigraphic constraints on the Middle to Late Miocene evolution of the Transylvanian Basin (Romania): Implications for Central Paratethys stratigraphy and emplacement of the Tisza–Dacia plate, *Global Planet. Change*, *103*, 82–98.
Dewey, J. F. (1980), Episodicity, sequence and style at convergent plate boundaries, in *The Continental Crust and Its Mineral Deposits*, edited by D. W. Strangway, pp. 553–573, Geol. Assoc. Canada, Waterloo, Ontario.
Dimitrijević, M. D. (1997), *Geology of Yugoslavia*, 2nd ed., 187 pp., Geoinstitute, Belgrade.
Dombrádi, E., D. Sokoutis, G. Bada, S. Cloetingh, and F. Horváth (2010), Modelling recent deformation of the Pannonian lithosphere: Lithospheric folding and tectonic topography, *Tectonophysics*, *484*(1–4), 103–118.
Dunkl, I., B. Grasemann, and W. Frisch (1998), Thermal effects of exhumation of a metamorphic core complex on hanging wall syn-rift sediments: An example from the Rechnitz Window, Eastern Alps, *Tectonophysics*, *297*, 31–50.
Ellouz, N., and E. Rocca (1994), Palinspastic reconstruction of the Carpathians and adjacent areas since the Cretaceous: A quantitative approach, in *Peri-Tethyan Platforms*, edited by F. Roure, pp. 51–78, Editions Technip, Paris.
Embry, A. F., and E. Johannessen (1992), T-R sequence stratigraphy, facies analysis and reservoir distribution in the uppermost Triassic-Lower Jurassic succession, western Sverdrup Basin, Arctic Canada, in *Arctic Geology and Petroleum Potential*, *Norwegian Petrol. Soc. Spec. Publ.*, vol. 2, edited by T. Vorren, Elsevier, Amsterdam.
Faccenna, C., L. Civetta, M. D'Antonio, F. Funiciello, L. Margheriti, and C. Piromallo (2005), Constraints on mantle circulation around the deforming Calabrian slab, *Geophys. Res. Lett.*, *32*, L06311, doi:10.1029/02004GL021874.
Faccenna, C., T. W. Becker, L. Auer, A. Billi, L. Boschi, J. P. Brun, F. A. Capitanio, F. Funiciello, F. Horvàth, and L. Jolivet (2014), Mantle dynamics in the Mediterranean, *Rev. Geophys.*, *52*, 283–332, doi:10.1002/2013RG000444.
Fodor, L., B. Jelen, E. Márton, D. Skaberne, J. Čar, and M. Vrabec (1998), Miocene-Pliocene 1089 tectonic evolution of the Slovenian Periadriatic fault: Implications for Alpine-Carpathian 1090 extrusion models, *Tectonics*, *17*, 690–709, doi:10.1029/98TC01605.
Fodor, L., L. Csontos, G. Bada, I. Györfi, and L. Benkovics (1999), Tertiary tectonic evolution of the Pannonian basin system and neighbouring orogens: A new synthesis of paleostress data, in *The Mediteranean Basins: Tertiary Extension Within the Alpine Orogen*, edited by B. Durand et al., pp. 295–334, Geol. Soc., London.
Fodor, L., G. Bada, G. Csillag, E. Horvath, Z. Ruszkiczay-Rudiger, K. Palotas, F. Sikhegyi, G. Timar, and S. Cloetingh (2005), An outline of neotectonic structures and morphotectonics of the western and central Pannonian Basin, *Tectonophysics*, *410*, 15–41.
Fodor, L., et al. (2008), Miocene emplacement and rapid cooling of the Pohorje pluton at the Alpine-Pannonian-Dinaridic junction, Slovenia, *Swiss J. Geosci.*, *101*, 255–271.
Fodor, L., O. Sztanó, I. Magyar, B. Törő, A. Uhrin, A. Várkonyi, G. Csillag, S. Kövér, Z. Lantos, and L. Tőkés (2013), Late Miocene depositional units and syn-sedimentary deformation in the western Pannonian basin, Hungary, in *11th Workshop on Alpine Geological Studies Abstract Book*, *Berichte der Geologischen Bundesanstalt*, vol. 99, edited by R. Schuster, pp. 33–34, Abstracts & Field Guides, Schladming.
Fügenschuh, B., and S. M. Schmid (2005), Age and significance of core complex formation in a very curved orogen: Evidence from fission track studies in the South Carpathians (Romania), *Tectonophysics*, *404*, 33–53, doi:10.1016/j.tecto.2005.03.019.
Fügenschuh, B., D. Seward, and N. Mancktelow (1997), Exhumation in a convergent orogen: The western Tauern window, *Terra Nova*, *9*, 213–217.
Funiciello, F., M. Moroni, C. Piromallo, C. Faccenna, A. Cenedese, and H. A. Bui (2006), Mapping mantle flow during retreating subduction: Laboratory models analyzed by feature tracking, *J. Geophys. Res.*, *111*, B03402, doi:10.1029/2005JB003792.
Gagała, Ł., J. Verges, E. Saura, T. Malata, J. C. Ringenbach, P. Werner, and P. Krzywiec (2012), Architecture and orogenic evolution of the northeastern Outer Carpathians from cross-section balancing and forward modeling, *Tectonophysics*, *532–535*, 223–241.
Govers, R., and M. J. R. Wortel (2005), Lithosphere tearing at STEP faults: Response to edges of subduction zones, *Earth Planet. Sci. Lett.*, *236*, 505–523.
Haas, J., and C. Péró (2004), Mesozoic evolution of the Tisza Mega-unit, *Int. J. Earth Sci.*, *93*, 297–313.
Haas, J., T. Budai, L. Csontos, L. Fodor, and G. Konrád (Eds.) (2010), Magyarországprekainozoos földtani térképe 1:500000 (Translated title: Pre-Cenozoic geological map of Hungary 1:500000), Geological Institute of Hungary (in Hungarian).
Hámor, G. (1985), Geology of the Nógrád–Cserhát area, *Geol. Hung.*, *22*, 1–307.
Handy, M. R., K. Ustaszewski, and E. Kissling (2015), Reconstructing the Alps–Carpathians– Dinarides as a key to understanding switches in subduction polarity, slab gaps and surface motion, *Int. J. Earth Sci.*, *104*, 1–26, doi:10.1007/s00531-014-11291060-3.
Harangi, S., and L. Lenkey (2007), Genesis of the Neogene to Quaternary volcanism in the Carpathian-Pannonian region: Role of subduction, extension, and mantle plume, *Geol. Soc. Am. Spec. Pap.*, *418*, 67–92.
Hinsken, S., K. Ustaszewski, and A. Wetzel (2007), Graben width controlling syn-rift sedimentation: The Palaeogene southern Upper Rhine graben as an example, *Int. J. Earth Sci.*, *96*, 979–1002.
Horváth, F. (1995), Phases of compression during the evolution of the Pannonian basin and its bearing on hydrocarbon exploration, *Mar. Pet. Geol.*, *12*, 837–844.

Horváth, F., and S. Cloetingh (1996), Stress-induced late-stage subsidence anomalies in the Pannonian basin, *Tectonophysics*, *266*, 287–300.

Horváth, F., G. Bada, P. Szafian, G. Tari, A. Adam, and S. Cloetingh (2006), Formation and deformation of the Pannonian Basin: Constraints from observational data, *Geol. Soc. London Mem.*, *32*(1), 191–206.

Horváth, F., B. Musitz, A. Balázs, A. Végh, A. Uhrin, A. Nádor, B. Koroknai, N. Pap, T. Tóth, and G. Wórum (2015), Evolution of the Pannonian basin and its geothermal resources, *Geothermics*, *53*, 328–352.

Houseman, G. A., and L. Gemmer (2007), Intra-orogenic extension driven by gravitational instability: Carpathian–Pannonian orogeny, *Geology*, *35*, 1135–1138.

Huismans, R., and C. Beaumont (2011), Depth-dependent extension, two-stage breakup and cratonic underplating at rifted margins, *Nature*, *473*(7345), 74–78.

Huismans, R. S., and C. Beaumont (2003), Symmetric and asymmetric lithospheric extension: Relative effects of frictional-plastic and viscous strain softening, *J. Geophys. Res.*, *108*(B10), 2496, doi:10.1029/2002JB002026.

Huismans, R. S., Y. Y. Podladchikov, and S. Cloetingh (2001), Dynamic modelling of the transition from passive to active rifting, application to the Pannonian basin, *Tectonics*, *20*, 1021–1039, doi:10.1029/2001TC900010.

Janik, T., M. Grad, A. Guterch, J. Vozár, M. Bielik, A. Vozárova, E. Hegedus, A. Kovacs, I. Kovács, and G. R. Keller (2011), CELEBRATION 2000 Working Group, Crustalstructure of the Western Carpathians and Pannonian Basin: Seismic modelsfrom CELEBRATION 2000 data and geological implications, *J. Geodyn.*, *52*, 97–113.

Jarosinski, M., F. Beekman, L. Matenco, and S. Cloetingh (2011), Mechanics of basin inversion: Finite element modelling of the Pannonian Basin System, *Tectonophysics*, *502*, 121–145.

Johnson, J. G., and M. A. Murphy (1984), Time-rock model for Siluro-Devonian continental shelf, Western United States, *Geol. Soc. Am. Bull.*, *95*, 1349–1359.

Jolivet, L., and J.-P. Brun (2010), Cenozoic geodynamic evolution of the Aegean, *Int. J. Earth Sci.*, *99*, 109–138.

Juhász, G. (1991), Lithostratigraphical and sedimentological framework of the Pannonian (s.l.) sedimentary sequence in the Hungarian Plain (Alföld), Eastern Hungary, *Acta Geol. Hung.*, *34*, 53–72.

Juhász, G., G. Pogacsas, I. Magyar, and G. Vakarcs (2007), Tectonic versus climatic control on the evolution of fluvio-deltaic systems in a lake basin, Eastern Pannonian Basin, *Sediment. Geol.*, *202*, 72–95.

Juhász, G., G. Pogácsás, I. Magyar, and P. Hatalyák (2013), The Alpar canyon system in the Pannonian Basin, Hungary – Its morphology, infill and development, *Global Planet. Change*, *103*, 174–192.

Karamata, S. (2006), The geological development of the Balkan Peninsula related to the approach, collision and compression of Gondwanan and Eurasian units, in *Tectonic Development of the Eastern Mediterranean Region*, *Geol. Soc. Spec. Publ.*, edited by A. H. F. Robertson and D. Mountrakis, pp. 155–178.

Király, Á., F. Horvath, and G. Tari (2012), Revisiting the anomalous crustal structure of the Pannonian Basin, 29th IAS Meeting of Sedimentology, Schladming.

Kiss, J. (2006), Magyarország gravitációs Bouguer-anomália térképe, Bouguer gravity map of Hungary, *Geophys. Trans.*, *45*(2), 99–104.

Kiss, J., and A. Gulyás (2006), Magyarország mágneses DZ-anomália térképe, 1:500 000 (Translated title: Magnetic DZ anomaly map of Hungary 1:500000), *ELGI*, www.mfgi.hu (in Hungarian).

Koroneos, A., G. Poli, V. Cvetkovi, G. Christofides, D. Krsti, and Z. Pécskay (2011), Petrogenetic and tectonic inferences from the study of the Mt Cer pluton (West Serbia), *Geol. Mag.*, *148*, 89–111.

Kőrössy, L. (1992), A Duna-Tisza koze koolaj- es foldgazkutatasanak foldtani eredmenyei (Translated title: Hydrocarbon geology of the Duna–Tisza interfluve, Hungary) [in Hungarian], *Általános Földtani Szemle*, *26*, 3–162.

Kováč, M., P. Kováč, F. Marko, S. Karoli, and J. Janočko (1995), The East Slovakian Basin: A complex back-arc basin, *Tectonophysics*, *252*, 453–466, doi:10.1016/0040-1951(95)00183-2.

Kovác, M., A. Andreyeva-Grigorovich, Z. Bajraktarevic, R. Brzobohatý, S. Filipescu, L. Fodor, M. Harzhauser, A. Nagymarosy, N. Oszczypko, and D. Pavelic (2007), Badenian evolution of the Central Paratethys Sea: Paleogeography, climate and eustatic sea-level changes, *Geol. Carpathica*, *58*, 579–606.

Krstić, N., L. Savić, G. Jovanović, and E. Bodor (2003), Lower Miocene lakes of the Balkan Land, *Acta Geol. Hung.*, *46*, 291–299.

Lankreijer, A., M. Kovac, S. Cloetingh, P. Pitonak, M. Hloska, and C. Biermann (1995), Quantitative subsidence analysis and forward modelling of the Vienna and Danube basins: Thin-skinned versus thick-skinned extension, *Tectonophysics*, *252*, 1–4.

Lelkes-Felvári, G., R. Schuster, W. Frank, and R. Sassi (2005), Metamorphic history of the Algyo High (Tisza Mega-unit, basement of Great Hungarian Plain)—A counterpart of crystalline units of the Koralpe–Wölz nappe system (Austroalpine, Eastern Alps), *Acta Geol. Hung.*, *48*(4), 371–394, doi:10.1556/AGeol.48.2005.4.2.

Lemberkovics, V. (2014), Lessons from combination of modern 3D seismic and historical well data: Development of a gilbert-delta, late Early Miocene Trough, Hungary, *CEMSEG onf. Abstr. Book*.

Lemberkovics, V., Á. Bárány, I. Gajdos, and M. Vincze (2005), A szekvencia-sztratigrafiai esemenyek es a tektonika kapcsolata a Derecskei-arok pannoniai retegsoraban (Translated title: Connection of sequence stratigraphy and tectonics in the Pannonian strata of the Derecske Trough) [in Hungarian], *Földtani Kutatás*, *42*, 16–24.

Lenkey, L. (1999), Geothermics of the Pannonian Basin and its bearing on the tectonics of basin evolution, PhD thesis, 215 pp., Vrije Universiteit Amsterdam.

Magyar, I., and O. Sztanó (2008), Is there a Messinian unconformity in the Central Paratethys?, *Stratigraphy*, *5*(3–4), 245–255.

Magyar, I., D. H. Geary, and P. Müller (1999a), Paleogeographic evolution of the Late Miocene Lake Pannon in Central Europe, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *147*, 151–167.

Magyar, I., D. H. Geary, M. Sütő-Szentai, and M. Lantos (1999b), Integrated biostratigraphic, magnetostratigraphic and chronostratigraphic correlations of the Late Miocene Lake Panon deposits, *Acta Geol. Hung.*, *42*(1), 5–31.

Magyar, I., A. Fogarasi, G. Vakarcs, L. Bukó, and G. C. Tari (2006), The largest hydrocarbon field discovered to date in Hungary: Algyő, in *The Carpathians and Their Foreland: Geology and Hydrocarbon Resources: AAPG Memoir*, vol. 84, edited by J. Golonka and F. J. Picha, pp. 619–632, AAPG, Tulsa, Okla.

Magyar, I., D. Radivojević, O. Sztanó, R. Synak, K. Ujszászi, and M. Pócsik (2013), Progradation of the paleo-Danube shelf margin across the Pannonian Basin during the Late Miocene and Early Pliocene, *Global Planet. Change*, *103*, 168–173.

Malinowski, M., et al. (2013), Deep seismic reflection profile in Central Europe reveals complex pattern of Paleozoic and Alpine accretion at the East European Craton margin, *Geophys. Res. Lett.*, *40*, 3841–3846, doi:10.1002/grl.50746.

Manatschal, G., L. Lavier, and P. Chenin (2015), The role of inheritance in structuring hyperextended rift systems: Some considerations based on observations and numerical modeling, *Gondwana Res.*, *27*, 140–164.

Martin, M., and F. Wenzel (2006), High-resolution teleseismic body wave tomography beneath SE-Romania—II. Imaging of a slab detachment scenario, *Geophys. J. Int.*, *164*, 579–595.

Martins-Neto, M. A., and O. Catuneanu (2010), Rift sequence stratigraphy, *Mar. Pet. Geol.*, *27*, 247–253.
Márton, E., and L. Fodor (2003), Tertiary paleomagnetic results and structural analysis from the Transdanubian Range (Hungary): Rotational disintegration of the AlCaPa unit, *Tectonophysics*, *363*(3–4), 201–224.
Márton, E., M. Tischler, L. Csontos, B. Fügenschuh, and S. Schmid (2007), The contact zone between the AlCaPa and Tisza-Dacia mega-tectonic units of Northern Romania in the light of new paleomagnetic data, *Swiss J. Geosci.*, *100*, 109–124.
Matenco, L., and G. Bertotti (2000), Tertiary tectonic evolution of the external East Carpathians (Romania), *Tectonophysics*, *316*, 255–286.
Matenco, L., and D. Radivojević (2012), On the formation and evolution of the Pannonian Basin: Constraints derived from the structure of the junction area between the Carpathians and Dinarides, *Tectonics*, *31*, TC6007, doi:10.1029/2012TC003206.
Matenco, L., C. Krézsek, S. Merten, S. Schmid, S. Cloetingh, and P. Andriessen (2010), Characteristics of collisional orogens with low topographic build-up: An example from the Carpathians, *Terra Nova*, *22*, 155–165.
Matenco, L., I. Munteanu, M. ter Borgh, A. Stanica, M. Tilita, G. Lericolais, C. Dinu, and G. Oaie (2016), The interplay between tectonics, sediment dynamics and gateways evolution in the Danube system from the Pannonian Basin to the western Black Sea, *Sci. Total Environ.*, *543*, 807–827, doi:10.1016/j.scitotenv.2015.10.081.
Merten, S., L. Matenco, J. P. T. Foeken, F. M. Stuart, and P. A. M. Andriessen (2010), From nappe stacking to out-of-sequence postcollisional deformations: Cretaceous to Quaternary exhumation history of the SE Carpathians assessed by low-temperature thermochronology, *Tectonics*, *29*, TC3013, doi:10.1029/2009TC002550.
Mészáros, F., and L. Zilahi-Sebess (2001), Compaction of sediments with great thickness in the Pannonian Basin, *Geophys. Trans.*, *44*, 21–48.
Miall, A. D., and C. E. Miall (2001), Sequence stratigraphy as a scientific enterprise: The evolution and persistence of conflicting paradigms, *Earth Sci. Rev.*, *54*, 321–348.
Nagymarosy, A., and G. Hámor (2012), Genesis and evolution of the Pannonian Basin, in *Geology of Hungary, Regional Geology Reviews*, edited by J. Haas, pp. 149–200, Springer, Heidlelberg, doi:10.1007/978-3-642-21910-8_3.
Nagymarosy, A., and P. Müller (1988), Some aspects of the Neogene biostratigraphy in the Pannonian Basin, in *The Pannonian Basin, A Study in Basin Evolution*, edited by L. H. Royden and F. Horvath, pp. 58–68, AAPG Mem., AAPG, Tulsa, Okla.
Nottvedt, A., R. H. Gabrielsen, and R. J. Steel (1995), Tectonostratigraphy and sedimentary architecture of rift basins, with reference to the northern North Sea, *Mar. Pet. Geol.*, *12*, 881–901.
Pálfy, J., R. Mundil, P. R. Renne, R. L. Bernor, L. Kordos, and M. Gasparik (2007), U–Pb and 40Ar/39Ar dating of the Miocene fossil track site at Ipolytarnóc (Hungary) and its implications, *Earth Planet. Sci. Lett.*, *258*, 160–174.
Pavelić, D. (2001), Tectonostratigraphic model for the North Croatian and North Bosnian sector of the Miocene Pannonian Basin System, *Basin Res.*, *13*, 359–376.
Pavelic, D., R. Avanic, K. Bakrac, and D. Vrsaljko (2001), Early Miocene braided river and lacustrine sedimentation in the Kalnik Mountain area (Pannonian Basin System, NW Croatia), *Geol. Carpathica*, *52*, 375–386.
Pécskay, Z., et al. (2006), Geochronology of Neogene magmatism in the Carpathian arc and intra-Carpathian area, *Geol. Carpathica*, *57*, 511–530.
Pereira, R., and T. M. Alves (2012), Tectono-stratigraphic signature of multiphased rifting on divergent margins (deep-offshore southwest Iberia, North Atlantic), *Tectonics*, *31*, TC4001, doi:10.1029/2011TC003001.
Pezelj, D., O. Mandic, and S. Coric (2013), Paleoenvironmental dynamics in the southern Pannonian Basin during initial Middle Miocene marine flooding, *Geol. Carpathica*, *64*(1), 81–100, doi:10.2478/geoca-2013-0006.
Pharaoh, T. C. (1999), Paleozoic terranes and their lithospheric boundaries within the Trans-European Suture Zone (TESZ): A review, *Tectonophysics*, *314*, 17–41.
Pigott, J., and D. Radivojević (2010), Seismic stratigraphy based chronostratigraphy (SSBC) of the Serbian Banat region of the Pannonian Basin, *Cent. Eur. J. Geosci.*, *2*(4), 481–500, doi:10.2478/v10085-010-0027-2.
Piller, W. E., M. Harzhauser, and O. Mandic (2007), Miocene Central Paratethys stratigraphy—Current status and future directions, *Stratigraphy*, *4*(2–3), 151–168.
Pinter, N., G. Grenerczy, J. Weber, and S. Stein (2005), *The Adria Microplate: GPS Geodesy, Tectonics and Hazards (Nato Science Series: IV: Earth and Environmental Sciences)*, edited by D. Medak, 413 pp., Springer, Dodrecht, Netherlands.
Pogácsás, G., L. Lakatos, I. Révész, K. Ujszászi, G. Vakarcs, L. Várkonyi, and P. Várnai (1988), Seismic facies, electro facies and Neogene sequence chronology of the Pannonian Basin, *Acta Geol. Hung.*, *31*, 175–207.
Posamentier, H. W., and G. P. Allen (1993), Variability of the sequence stratigraphic model: Effects of local basin factors, *Sediment. Geol.*, *86*, 91–109.
Prosser, S. (1993), Rift related depositional system and their seismic expression, in *Tectonics and Seismic Sequence Stratigraphy*, edited by G. D. Williams and A. Dobb, *Geol. Soc. Spec. Publ*., 71, 35–66, doi:10.1144/GSL.SP.1993.071.01.03.
Puchnerová, M., J. Šefara, V. Szalaiová, H. Tkáčová, J. Gretsch, M. Bielik, J. Kiss, and S. Kovácsvölgyi (2002), Results of the geophysical survey within the framework of the Triberg project, *Geol. Carpathica*, *53*.
Răbăgia, A., L. Matenco, and S. Cloetingh (2011), The interplay between eustacy, tectonics and surface processes suring the growth of a fault-related structure as derived from sequence stratigraphy: The Govora–Ocnele Mari Antiform, South Carpathians, *Tectonophysics*, *502*, 196–220.
Răbăgia, A.-M. (2009), Sequential stratigraphic studies in the northern part of the Pannonian Basin for deriving the tectono-stratigraphic evolution, PhD thesis, 98 pp., Fac. of Geol. and Geophys., Univ. of Bucharest, Bucharest.
Ratschbacher, L., W. Frisch, H. G. Linzer, and O. Merle (1991), Lateral extrusion in the Eastern Alps; Part 2, Structural analysis, *Tectonics*, *10*, 257–271, doi:10.1029/90TC02623.
Ratschbacher, L., H. G. Linzer, F. Moser, R. O. Strusievicz, H. Bedelean, N. Har, and P. A. Mogos (1993), Cretaceous to Miocene thrusting and wrenching along the central South Carpathians due to a corner effect during collision and orocline formation, *Tectonics*, *12*, 855–873, doi:10.1029/93TC00232.
Ravnas, R., and R. J. Steel (1998), Architecture of marine rift-basin successions, *AAPG Bull.*, *82*, 110–146.
Roure, F., E. Roca, and W. Sassi (1993), The Neogene evolution of the outer Carpathian flysch units (Poland, Ukraine and Romania): Kinematics of a foreland/fold-and-thrust belt system, *Sediment. Geol.*, *86*, 177–201.
Royden, L., and B. C. Burchfiel (1989), Are systematic variations in thrust belt style related to plate boundary processes? (the Western Alps versus the Carpathians), *Tectonics*, *8*, 51–61, doi:10.1029/TC008i001p00051.
Rumpler, J., and F. Horváth (1988), Some representative seismic reflection lines from the Pannonian basin and their structural interpretation, in *The Pannonian Basin, A Study in Basin Evolution*, edited by L. H. Royden and F. Horváth, pp. 153–169, AAPG Memoir, AAPG, Tulsa, Okla.
Sacchi, M., F. Horváth, and O. Magyari (1999), Role of unconformity-bounded units in the stratigraphy of the continental record: A case study from the late Miocene of the western Pannonian Basin, Hungary, in *The Mediterranean Basins: Extension Within the Alpine Orogen*, edited by B. Durand et al., *Geol. Soc. Spec. Publ.*, *156*, 357–390.

Saftić, B., J. Velic, O. Sztanó, G. Juhász, and Ž. Ivkovic (2003), Tertiary subsurface facies, source rocks and hydrocarbon reservoirs in the SW part of the Pannonian Basin (northern Croatia and south-western Hungary), *Geol. Croat.*, *56*, 101–122.

Săndulescu, M. (1988), Cenozoic tectonic history of the Carpathians, in *The Pannonian Basin, A Study in Basin Evolution*, *AAPG Mem.*, vol. 45, edited by L. H. Royden and F. Horváth, pp. 17–25.

Scharf, A., M. R. Handy, S. Favaro, S. M. Schmid, and A. Bertrand (2013), Modes of orogen-parallel stretching and extensional exhumation in response to microplate indentation and roll-back subduction (Tauern Window, Eastern Alps), *Int. J. Earth Sci.*, *102*, 1627–1654.

Schefer, S., V. Cvetković, B. Fügenschuh, A. Kounov, M. Ovtcharova, U. Schaltegger, and S. Schmid (2011), Cenozoic granitoids in the Dinarides of southern Serbia: Age of intrusion, isotope geochemistry, exhumation history and significance for the geodynamic evolution of the Balkan Peninsula, *Int. J. Earth Sci.*, *100*, 1181–1206.

Schlager, W. (1993), Accommodation and supply—A dual control on stratigraphic sequences, *Sediment. Geol.*, *86*, 111–136.

Schmid, S., D. Bernoulli, B. Fügenschuh, L. Matenco, S. Schefer, R. Schuster, M. Tischler, and K. Ustaszewski (2008), The Alpine-Carpathian-Dinaridic orogenic system: Correlation and evolution of tectonic units, *Swiss J. Geosci.*, *101*, 139–183.

Schmid, S. M., B. Fugenschuh, E. Kissling, and R. Schuster (2004), Tectonic map and overall architecture of the Alpine orogen, *Eclogae Geol. Herv.*, *97*, 92–117.

Sclater, J. G., L. Royden, F. Horvath, B. C. Burchfiel, S. Semken, and L. Stegena (1980), The formation of the intra-Carpathian basins as determined from subsidence data, *Earth Planet. Sci. Lett.*, *51*, 139–162.

Seghedi, I., L. Matenco, H. Downes, P. R. D. Mason, A. Szakács, and Z. Pécskay (2011), Tectonic significance of changes in post-subduction Pliocene-Quaternary magmatism in the south east part of the Carpathian-Pannonian Region, *Tectonophysics*, *502*, 146–157.

Steininger, F. F., and F. Rögl (1984), Paleogeography and palinspastic reconstruction of the Neogene of the Mediterranean and Paratethys, *Geol. Soc. Lond. Spec. Publ.*, *17*, 659–668.

Stojadinovic, U., L. Matenco, P. A. M. Andriessen, M. Toljić, and J. P. T. Foeken (2013), The balance between orogenic building and subsequent extension during the Tertiary evolution of the NE Dinarides: Constraints from low-temperature thermochronology, *Global Planet. Change*, *103*, 19–38.

Sumanovac, F. (2010), Lithosphere structure at the contact of the Adriatic microplate and the Pannonian segment based on the gravity modeling, *Tectonophysics*, *485*, 94–106.

Szafián, P., and F. Horváth (2006), Crustal structure in the Carpatho-Pannonian region: Insights from three-dimensional gravity modelling and their geodynamic significance, *Int. J. Earth Sci.*, *95*, 50–67.

Szepesházy, K. (1971), A Tiszantul kozepso reszenek miocen kepzodmenyei a szenhidrogenkutato melyfurasok adatai alapjan (Translated title: Miocene formations of the central Great Hungarian Plain. Inferences from well data) [in Hungarian], *Annu. Rep. Hung. Geol. Inst.*, *1968*, 297–325.

Sztanó, O., P. Szafián, I. Magyar, A. Horányi, G. Bada, D. W. Hughes, D. L. Hoyer, and R. J. Wallis (2013), Aggradation and progradation controlled clinothems and deep-water sand delivery model in the Neogene Lake Pannon, Makó Trough, Pannonian Basin, SE Hungary, *Global Planet. Change*, *103*, 149–167.

Szuromi-Korecz, A., M. Sütő-Szentai, and I. Magyar (2004), Biostratigraphic revision of the Hód-I well: Hungary's deepest borehole failed to reach the base of the Upper Miocene Pannonian Stage, *Geol. Carpathica*, *55*, 475–485.

Tari, G., F. Horváth, and J. Rumpler (1992), Styles of extension in the Pannonian basin, *Tectonophysics*, *208*, 203–219.

Tari, G., P. Dovenyi, I. Dunkl, F. Horvath, L. Lenkey, M. Ştefănescu, P. Szafian, and T. Toth (1999), Lithospheric structure of the Pannonian basin derived from seismic, gravity and geothermal data, in *The Mediterranean Basins: Extension Within the Alpine Orogen*, *Geol. Soc. Spec. Publ.*, vol. 156, edited by B. Durand et al., pp. 215–250.

Tašárová, A., J. C. Afonso, M. Bielik, H.-J. Götze, and J. Hók (2009), The lithospheric structure of the Western Carpathian–Pannonian Basin region based on the CELEBRATION 2000 seismic experiment and gravity modeling, *Tectonophysics*, *475*, 454–469.

Teleki, P. G., R. E. Mattick, and J. Kokai (Eds.) (1994), *Basin Analysis in Petroleum Exploration. A Case Study From the Bekes Basin, Hungary*, 330 pp., Kluwer Acad., Dordecht.

ter Borgh, M. (2013), Connections between sedimentary basins during continental collision: How tectonic, surface and sedimentary processes shaped the Paratethys, PhD thesis, 212 pp., Utrecht Univ., Utrecht, Netherlands.

ter Borgh, M., I. Vasiliev, M. Stoica, S. Knežević, L. Matenco, W. Krijgsman, L. Rundić, and S. Cloetingh (2013), The isolation of the Pannonian basin (Central Paratethys): New constraints from magnetostratigraphy and biostratigraphy, *Global Planet. Change*, *103*, 99–118.

ter Borgh, M., D. Radivojević, and L. Matenco (2015), Constraining forcing factors and relative sea-level fluctuations in semi-enclosed basins: The Late Neogene demise of Lake Pannon, *Basin Res.*, doi:10.1111/bre.12094, in press.

Toljić, M., L. Matenco, M. N. Ducea, U. Stojadinović, J. Milivojević, and N. Đerić (2013), The evolution of a key segment in the Europe–Adria collision: The Fruška Gora of northern Serbia, *Global Planet. Change*, *103*, 39–62.

Tugend, J., G. Manatschal, N. J. Kusznir, E. Masini, G. Mohn, and I. Thinon (2014), Formation and deformation of hyperextended rift systems: Insights from rift domain mapping in the Bay of Biscay-Pyrenees, *Tectonics*, *33*, 1239–1276, doi:10.1002/2014TC003529.

Tulucan, A. (2007), Complex geological study of the Romanian sector of the Pannonian Depression with special regard to hydrocarbon accumulation, PhD thesis, 220 pp., Fac. of Geol. and Geophys., Univ. of Bucharest, Bucharest.

Uhrin, A., I. Magyar, and O. Sztanó (2009), Effect of basement deformation on the Pannonian sedimentation of the Zala Basin, SW Hungary, *Földtani Közlöny*, *139*, 273–282.

Ustaszewski, K., S. Schmid, B. Fügenschuh, M. Tischler, E. Kissling, and W. Spakman (2008), A map-view restoration of the Alpine-Carpathian-Dinaridic system for the early Miocene, *Swiss, J. Geosci. Prague*, *101*, 273–294, doi:10.1007/s00015-008-1288-7.

Ustaszewski, K., A. Kounov, S. M. Schmid, U. Schaltegger, E. Krenn, W. Frank, and B. Fügenschuh (2010), Evolution of the Adria-Europe plate boundary in the northern Dinarides: From continent-continent collision to back-arc extension, *Tectonics*, *29*, TC6017, doi:10.1029/2010TC002668.

Ustaszewski, K., M. Herak, B. Tomljenović, D. Herak, and S. Matej (2014), Neotectonics of the Dinarides-Pannonian Basin transition and possible earthquake sources in the Banja Luka epicentral area, *J. Geodyn.*, *82*, 52–68.

Uyeda, S., and H. Kanamori (1979), Back-arc opening and the mode of subduction, *J. Geophys. Res.*, *84*, 1049–1061, doi:10.1029/JB084iB03p01049.

Vail, P. R., R. M. Mitchum, K. M. Campion, and V. D. Rahmanian (1977), Seismic stratigraphy and global changes of sea level. Part 3: Relative changes of sea level from coastal onlap, in *Seismic Stratigraphy—Application to Hydrocarbon Exploration*, *AAPG Mem.*, vol. 26, edited by C. E. Payton, pp. 83–97.

Vakarcs, G., P. R. Vail, G. Tari, G. Pogácsás, R. E. Mattick, and A. Szabó (1994), Third-order Middle Miocene-Early Pliocene depositional sequences in the prograding delta complex of the Pannonian Basin, *Tectonophysics*, *240*, 81–106.

van Gelder, I. E., L. Matenco, E. Willingshofer, B. Tomljenović, P. A. M. Andriessen, M. N. Ducea, A. Beniest, and A. Gruié (2015), The tectonic evolution of a critical segment of the Dinarides-Alps connection: Kinematic and geochronological inferences from the Medvednica Mountains, *NE Croatia, Tectonics*, *34*, 1952–1978, doi:10.1002/2015TC003937.

van Hinsbergen, D. J. J., E. Hafkenscheid, W. Spakman, J. E. Meulenkamp, and R. Wortel (2005), Nappe stacking resulting from subduction of oceanic and continental lithosphere below Greece, *Geology*, *33*, 325–328.

Van Wagoner, J. C., R. M. Mitchum, K. M. Campion, and V. D. Rahmanian (1990), *Siliciclastic Sequence Stratigraphy in Well Logs, Core, and Outcrops: Concepts for High-Resolution Correlation of Time and Facies*, *Am. Assoc. Petrol. Geol. Methods Explor. Ser.*, vol. 7, 55 pp., AAPG, Tulsa, Okla.

Vergés, J., and M. Fernàndez (2012), Tethys–Atlantic interaction along the Iberia–Africa plate boundary: The Betic–Rif orogenic system, *Tectonophysics*, *579*, 144–172.

Vissers, R. L. M. (2012), Extension in a convergent tectonic setting: A lithospheric view on the Alboran system of SW Europe, *Geol. Belg.*, *15*, 53–72.

Wernicke, B. (1985), Uniform-sense normal simple shear of the continental lithosphere, *Can. J. Earth Sci.*, *22*, 108–125, doi:10.1139/e85-009.

Windhoffer, G., G. Bada, D. Nieuwland, G. Worum, F. Horvath, and S. Cloetingh (2005), On the mechanics of basin formation in the Pannonian basin: Inferences from analogue and numerical modelling, *Tectonophysics*, *410*, 389–415.

Wortel, M. J. R., and W. Spakman (2000), Subduction and slab detachment in the Mediterranean-Carpathian region, *Science*, *209*, 1910–1917.