

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

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**Keypoints:**

- Coupled tectono-stratigraphic evolution of the Pannonian Basin
- Kinematics of extension in back-arc settings
- Migration of tectonic phases

**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

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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 roll-back 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 sub-basins, 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 sub-basins 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 hyper-extended back-arc basins observed elsewhere.

**Keywords:** back-arc extension, tectonic system tracts, inversion tectonics, Pannonian Basin, Carpathians

## 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 Fernández, 2012; Wortel and Spakman, 2000*]. In all these situations extension post-dates 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 hyper-extended passive continental margins [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 non-genetic 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. Such a simple definition is independent of the controlling mechanics, from simple-shear in single or multi-layered lithosphere to complex multi-stage hyper-extended basins [Huismans and Beaumont, 2003; Manatschal et al., 2015; Wernicke et al., 1985].

A typical example is the Pannonian Basin of Central Europe (Figure 1), where the large amount of extension is accommodated by the rapid roll-back of the Carpathians slab between 20-9 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 syn-rift and thick post-rift basin fill is usually interpreted as a consequence of either depth dependent stretching, necking depth or intra-plate 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  $\sim 2.2$ , 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 hyper-extended 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; Tari et al., 1999]. Some of these half-grabens show typical syn-kinematic 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 Radivojevic, 2012, Sztanó *et al.*, 2013; Windhoffer *et al.*, 2005]. In this part of the basin, we have used a large network of regionally distributed 2D and 3D 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 tectono-sedimentary 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 preceded 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; Schmid *et al.*, 2008]. In the NW, the ALCAPA Mega-Unit is an Adriatic-derived block that was sutured to Europe during the N-wards Cretaceous – Eocene closure of the Alpine Tethys [Csontos, 1995; Schmid *et al.*, 2004]. To the E 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 Péro, 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 build-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 E-ward 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 semi-enclosed 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, absolute age dating of exhumation in the footwall of extensional detachments that outcrop near the basin margin and, more importantly, by the timing of syn- and post-rift 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 syn-rift 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, Pálffy 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 22-17 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 post-collisional 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 syn-kinematic 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 extensional structures reactivate in many situations 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 Radivojevic, 2012; Pavelić et al., 2001; Saftić et al., 2003*], in the northwestern parts of the basin marine sediments deposited [*Nagymarosy and Hámor, 2012*]. Transgression during the Middle Miocene [*Kováč 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 [Nagyvarosy 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; Kőrössi, 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. Post-extensional evolution

There is no consensus on a general onset of post-rift 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]. In other parts of the Pannonian Basin, but dominantly in the Great Hungarian Plain, the onset of post-rift deposition is interpreted at the boundary between Middle and Late Miocene [11.63 Ma, *sensu 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 9Ma [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 NE- and E-wards [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 towards 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 ~400km in 6My until ~4Ma 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 hemi-pelagic 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 magneto- and bio-stratigraphic

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 micro-plate, 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, 1997]. This unconformity is angular and locally erosional near the basin margins and passes to a correlative conformity towards the basin center. It is interpreted 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 Salinity 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 cross-over zones of different progradational directions [Magyar and Sztanó, 2008].

### **3. Methodology**

We have analyzed the link between structures and sedimentation in the sub-basins located in the Great Hungarian Plain part of the Pannonian Basin mostly in Hungary, but also in Serbia, Slovakia and Romania (Figure 1 and 3). In Hungary, we have analyzed a large

array of 2D and 3D seismic data calibrated by a dense network of exploration wells distributed regionally in the study area particularly in the studied sub-basins, but 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 3D seismic surveys to older 2D seismic lines. Although in this paper we present only a limited number of seismic transects usually oriented across the strike of various sub-basins, we used a much larger seismic and well database (including hundreds of 2D seismic lines and a few 3D seismic surveys). Well logs were tied to seismic sections using the standard VSP 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 sub-basins. 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 [*Dimitrijevic*, 1997; *Ellouz and Rocca*, 1994; *Gagala et al.*, 2012; *Matenco and Radivojevic*, 2012; *Pigott and Radivojević*, 2010; *Puchnerova et al.*, 2002; *Rábăgia*, 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; *Catuneanu et al.*, 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; *Ravnås 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 post-rift system tract, when the continued thermal sag subsidence resulted in the burial of the inherited rift topography, which is followed by a late post-rift 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 post-rift stage may create differential compaction effects over the half grabens, such as synclinal geometries or compaction faults with increasing offsets towards the surface, in particular when the thickness of the overburden is high.

The large syn-kinematic 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 boundaries 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 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 sub-basins, where wells are less frequent. In these situations, the maximum flooding surface (MFS) 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 tectono-sedimentary signature of the syn-kinematic basin fill at different scales [see also 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 sub-basin 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 sub-basins in the Great Hungarian Plain (Figure 3). The main characteristic geometry of these sub-basins 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; Tari *et al.*, 1999]. These structures flank sub-basins with low-angle dipping hanging-walls in overall half-graben geometries. Although antithetic normal faults with significantly lower offsets often cross cut these hanging-walls, the syn-kinematic 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 also 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 GHP 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 sub-basins 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 post-date the extension. The clinoform facies unit contains locally intercalated hummocky ones (i.e. fourth seismic facies unit), which are in fact either clinoforms cross-cut along their strike or local turbiditic bodies (Szolnok Formation). The spatial distribution and significance of these post-rift facies units are well documented elsewhere [Juhász *et al.*, 2007; Magyar *et al.*, 2013; Sztanó *et al.*, 2013]. The syn-kinematic sedimentation in the various sub-basins is characterized by a combination between different seismic facies units (units 5-10, Figure 4). In the lower parts of the syn-kinematic 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., Figure 7 and 9). The sub-parallel 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 inter-fingering 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 sub-basins, in particular well developed 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, either prograding toward the center of the basin, retrograding towards 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.

## 4.2. Structures with peak syn-kinematic deposition during the Early Miocene

The Kiskunhalas Trough, with limited extent along its strike, is a narrow and deep basin, with syn-kinematic sedimentation reaching more than 3 kilometers in thickness (Figure 6). The overall structural geometry is asymmetric, with one large listric normal fault dipping NW-wards 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 syn-kinematic hanging-walls deposition, is in the order of 7 km. Note that 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 sub-aerial (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 (Kiskunhalas Formation). This is correlated with a similar succession found elsewhere that contains poorly developed Lower Miocene (Karpatian) fauna [Rumpler and Horváth, 1988; Kőrössi, 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 [Kőrössi, 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 sub-basin (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 – transgressive sequence boundary. The rift climax system tract contains clastic lobes facies units that decrease gradually upwards in the stratigraphy, most likely as 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 post-rift 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 post-rift system tract (Figure 5). The late post-rift 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 post-rift system tract, which is sub-divided 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 upwards. 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 post-rift system tract is made up just by one transgressive facies association, followed by regression during the onset of the late stage post-rift, interrupted by the uplift, erosion and the formation of the regional Middle-Late Miocene unconformity.

#### **4.3. Structures with peak syn-kinematic 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 SSE-ward located Paleozoic basement of the Villány-Bihor nappe (Figure 7). The overall structural geometry shows larger

offset low-angle normal faults dipping NNW-wards which cross-cut the inherited Late Cretaceous Nappe contact and are associated with smaller offset antithetic normal faults dipping SSE-wards. 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 syn-kinematic 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 cross cuts 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 sub-basin, Figure 7) filled with Lower Miocene (Karpatian) conglomerates and sandstones, which are separated from the overlying Pannonian sediments by an erosional unconformity, likely of sub-aerial origin. This half-graben is also slightly inverted by folding in a gentle anticline-syncline structure in a similar fashion as the Kiskunhalas sub-basin.

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 post-rift system tract is built up by Late Miocene (Pannonian) sediments. Similar with the Kiskunhalas structure, the erosion of the footwall in the Tázlár sub-basin indicates that these are genetic transgressive-regressive sequences. Overlying the rift climax system tract, the Upper Miocene sediments filled the depocenter during the late post-rift system tract. In fact the entire post-rift 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 sub-basins, i.e. Békés, Vésztő and Derecske (Figure 3). The age of the syn-kinematic sediments is younger from SW to NE (i.e., from Békés to Derecske). In the SW, wells in the large Békés sub-basin have penetrated Middle-Miocene syn-kinematic sediments until ~5 km depth [see *Teleki et al.*, 1994]. Beneath these deposits, the age near the base of the syn-kinematic succession in this up to 7 km thick sub-basin is still unknown.

The Vésztő Trough is a half-graben with low-angle normal faults dipping NW-wards (Figure 8). Although the number of wells penetrating this structure is relatively low, the lateral correlation from the neighboring Békés and Derecske sub-basins indicates that the syn-kinematic basin fill is Middle Miocene – earliest Late Miocene. This structure is typical for the overall Pannonian Basin, thin (~1.5 km) syn-kinematic deposition is overlain by remarkably thick (~3.5 km) post-rift 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 post-rift and a thick late post-rift 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 post-rift system tract (Figure 8).

#### **4.4. Structures with peak syn-kinematic 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 (Figure 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 sub-basin 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 E-wards (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 sub-basin to Late Miocene along the ESE flank (Figure 9). The asymmetry of the syn-kinematic basin fill indicates higher cumulated offsets for the ESE-dipping system when compared with the WNW-dipping one. In the upper part of the syn-kinematic 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 sub-basin (Figure 9). This indicates a migration in time ESE-ward 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 sub-cycle completes the rift-climax system tract. Immediate post-rift 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 cross-cut 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 metres 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 sub-basins developed in a relatively small area, i.e., Szeged, Makó and Tomnatec sub-basins (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 westwards and a smaller offset antithetic normal fault dipping eastwards (Figure 10). The highly eroded footwall of the ESE fault bordering the Algyó-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 sub-basin and are associated with Middle and Late Miocene syn-kinematic sedimentation. The minimum value of the horizontal extensional displacement along all these structures is in the order of 8 kilometers. The syn-kinematic sedimentation reached ~2.5 km in the center of the Szeged sub-basin. 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 Radivojevic, 2010].

The seismic sequence interpretation of the Szeged sub-basin 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 sub-basin. 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 sub-basin. 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 ESE-ward 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 post-rift system tract has a progradational character that continues also during the late post-rift system tract. In fact the entire post-rift is part of the Pannonian prograding shelf margin – slope clinoform system that started during the immediate post-rift with the deposition of bottomsets at ~5 Ma [Magyar *et al.*, 2013]. The Pliocene late post-rift system tract reaches 1.5 km in thickness. The main fault bordering to the ESE the Szeged sub-basin has an offset upwards in the post-rift 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 post-rift strata eastwards 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 post-rift turbiditic sediments in this sub-basin.

The Makó Trough of SE Hungary (Figure 11) is one of the largest and deepest Neogene sub-basins 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 S with the Tomnatec sub-basin (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 Figure 2 in Sztanó *et al.*, 2013]. On the overall, the geometry of the pre-Neogene sub-basin flanks is remarkably symmetric. However, the real asymmetry is visible in the syn-kinematic 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 sub-basin and the uplift of the adjacent Algyő footwall (Figure 11, see also Tari *et al.*, 1999). This uplift was associated with erosion over the structural culmination of the Algyő footwall. Significantly smaller offsets 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 kilometers.

The rift initiation system tract in the Makó sub-basin 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 sub-basin as overlain by an unconformity (Figure 11). Within the rift climax system tract three higher-order retrogradational sub-cycles 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 sub-cycles. 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 post-rift system tract is interpreted as the first progradational facies association. The Makó sub-basin 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 sub-basin within the alluvial and delta plain environments.

The flanks of the NNW-SSE oriented Makó sub-basin is cross-cut by a smaller WSW-ENE oriented normal fault that created the smaller Földeák sub-basin filled most likely by syn-kinematic 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 sub-basins (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 syn-kinematic succession (e.g., Early Miocene rift initiation of the Kiskunhalas and Middle

Miocene rift initiation of the Szeged sub-basins, Figures 6 and 10, respectively) or over the flanks of the separating highs (e.g., Tázlár and Derecske sub-basins, 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 sub-basin (Figure 12). The sedimentological environment of the rift-initiation system tract in our observed extensional sub-basins is also different when compared with these earlier models. Continental conditions are observed only when the rift-initiation system tract is entirely of Lower Miocene in age (e.g., Soltvadkert sub-basin, 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 sub-basin, 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 sub-basin. 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 criteria to distinguish the high

resolution cyclicality. In more distal positions, (sub-) parallel to divergent seismic facies units characterize the basin fill associated with sigmoidal seismic facies units during periods of progradation. 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 sub-parallel to hummocky turbiditic seismic facies units fill the large accommodation space created by a subsidence that is more regional than the scale of individual sub-basins, followed by the deposition of the cliniform 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 post-rift 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 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, Vesző and Szeged sub-basins (Figures 6, 8 and 10), delta systems can be observed prograding over the hanging-wall towards the center of the sub –basins 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 sub-basins, Figure 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. 10-4 Ma in the Great Hungarian Plain [Magyar *et al.*, 2013]. Previous calculations indicate several hundreds of meters, possibly up to 1 km paleobathymetries [Balázs *et al.*, 2013]. In the peripheral areas of the Pannonian Basin, like the Nyírség sub-basin 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 (Figure 3 and 13). This means that in the deepest Late Miocene half-grabens, situated further away from source areas, shallow water sediments build up only the oldest Pannonian syn-kinematic 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 post-dated the normal faulting, burying some part of the controlling normal fault with deep water deposition during post-kinematic times.

The immediate post-rift 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 sub-parallel, fairly continuous, occasionally gently divergent reflections (Figure 5). The onset of post-rift deposition was diachronous in the various sub-basins and is not marked by a clear erosional unconformity marking a transition to post-rift sedimentation (see also *Tari et al.*, 1999). Driven by regional contractional processes (*Horváth*, 1995), the juxtaposition of the Middle/Late Miocene regional unconformity with a general syn-rift/post-rift boundary would be a coincidence, but this was not observed in every studied sub-basin.

The late post-rift system tract deposits of the Pannonian Basin are made up by turbiditic deposits, prograding shelf margin slope, and alluvial plain and delta sediments (Figure 5 and 12). During the late post-rift system tract the cyclicity was controlled by other regional processes than the local scale of the extensional sub-basins, 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 clinofolds above the former extensional basins, most likely due to differential compaction being active during deposition. This compaction creates also offsets in the post-rift sequence overlying the earlier normal faults [e.g. *Teleki et al.*, 1994], which typically increase upwards 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 sub-basins 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. Tectono-sedimentary evolutionary model of a highly extended back-arc basin

The seismo-stratigraphic interpretation infers a clear interplay between sedimentation and tectonics during the Miocene syn-kinematic 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 sub-basins and associated detachments started near or in the Dinarides already during Early Miocene times, or possibly even earlier during the Oligocene [*e.g.*, Toljic *et al.*, 2013], being accompanied by the deposition of thin continental alluvial to lacustrine sediments [*Krstic et al.*, 2003; *Matenco and Radivojevic*, 2012]. Interestingly, the amplitude of extensional exhumation of the footwalls was much higher than the tectonic subsidence of the associated sub-basins 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 sub-basin.

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 NE to E-ward 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 non-deposition as suggested in areas near the Dinarides [Matenco and Radivojevic, 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 km-thick sequence. Therefore, the unconformity formed likely as a combination between the previously mentioned mechanisms. The N-S oriented direction of contraction is in contrast with the inferred E-ward 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 sub-basins 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 that this extension had major effects by creating or significantly enlarging some of the deepest sub-basins 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 post-rift subsidence that was diachronous and buried most of the sub-basins 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 sub-basin versus the NW-SE strike of Late Miocene Szeged or Makó sub-basins (Figure 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, 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 S and SW margin of the Dinarides N-

wards (i.e. from the Adriatic indenter) towards 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 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 sub-basin, Figure 7), being replaced laterally with a correlative conformity in deeper sub-basins [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  $\sim 20^\circ$  (Makó sub-basin) to  $\sim 30^\circ$  (Szeged sub-basin), or up to  $\sim 40^\circ$  (Derecske sub-basin). 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 sub-basin, Figure 9, see also Windhoffer *et al.*, 2005]. This is in particular obvious for the presently NW-ward 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 pre-existing thrusts (such as the Tazlar sub-basin, 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 sub-basin 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 sub-basins show non-metamorphosed 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 Algyó High that is situated in the footwall of the controlling structure in both the Szeged and Makó sub-basins (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 non-metamorphic 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 sub-basin 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, thrustured probably during Turonian times over the neighboring Bekes-Codru nappe that retains a Variscan age metamorphism and is covered by a non-metamorphosed Mesozoic sequence [Figure 10; Schmid *et al.*, 2008]. The alternative interpretation suggests that Algyó 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 Algyó High pre-dating the Miocene extension. This was subsequently followed by ~7-8km of Miocene exhumation, which is roughly above the 220° ZFT annealing temperature and below the metamorphic threshold, given the high geothermal gradient of the

Pannonian Basin. Alternatively, the Cretaceous metamorphism of the Algyó 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 sub-basins, 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 2-7 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 or higher than thickness of the syn-kinematic basin fill deposited in their hanging-walls. This type of exhumation increases in local detachments and core-complexes towards 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 syn-kinematic 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 syn-kinematic 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 sub-basins (Figure 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 roll-back 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; Horvath et al., 2015*]. Beyond possible heterogeneities created by the inherited nappe structure, the anti-correlation of the basement depth with the Bouguer gravity anomaly (Figure 13a) has been previously observed [*Szafián and Horváth, 2006*]. In our view, such anti-correlation 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 NE- and E- wards 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. Further northwards, the un-stretched parts of the Tisza-Dacia units simply translated NE- and E-wards, 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 dominantly a continental lithospheric domain. These processes were obviously driven by the evolution of inherited subducted slabs, as derived by

teleseismic tomography [e.g., *Benett 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 3D 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 syn-kinematic basin fill (e.g., Derecske Trough, Figure 9), but also their significant burial (such as in the Makó or Szeged sub-basins, 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 sub-basins 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 sub-basins 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 sub-basins) to E-W in the eastern (e.g., Szeged sub-basin) and NW-SE in the northern-central Great Hungarian Plain (e.g., Derecske sub-basin). The gradual and large amount of clockwise rotation of the Tisza-Dacia Mega-Unit significantly modified the original geometry of the sub-basins. The extensional mechanism was controlled by the pre-existence 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 syn-kinematic 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 ca. 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 et al.*, 2009; *Houseman and Gemmer*, 2007; *Horváth et al.*, 2015] have still 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 hyper-extended back-arc, its formation and evolution being strongly controlled by inherited orogenic asymmetries, subsequent slab dynamics and dynamic topography mechanics.

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## Figures captions

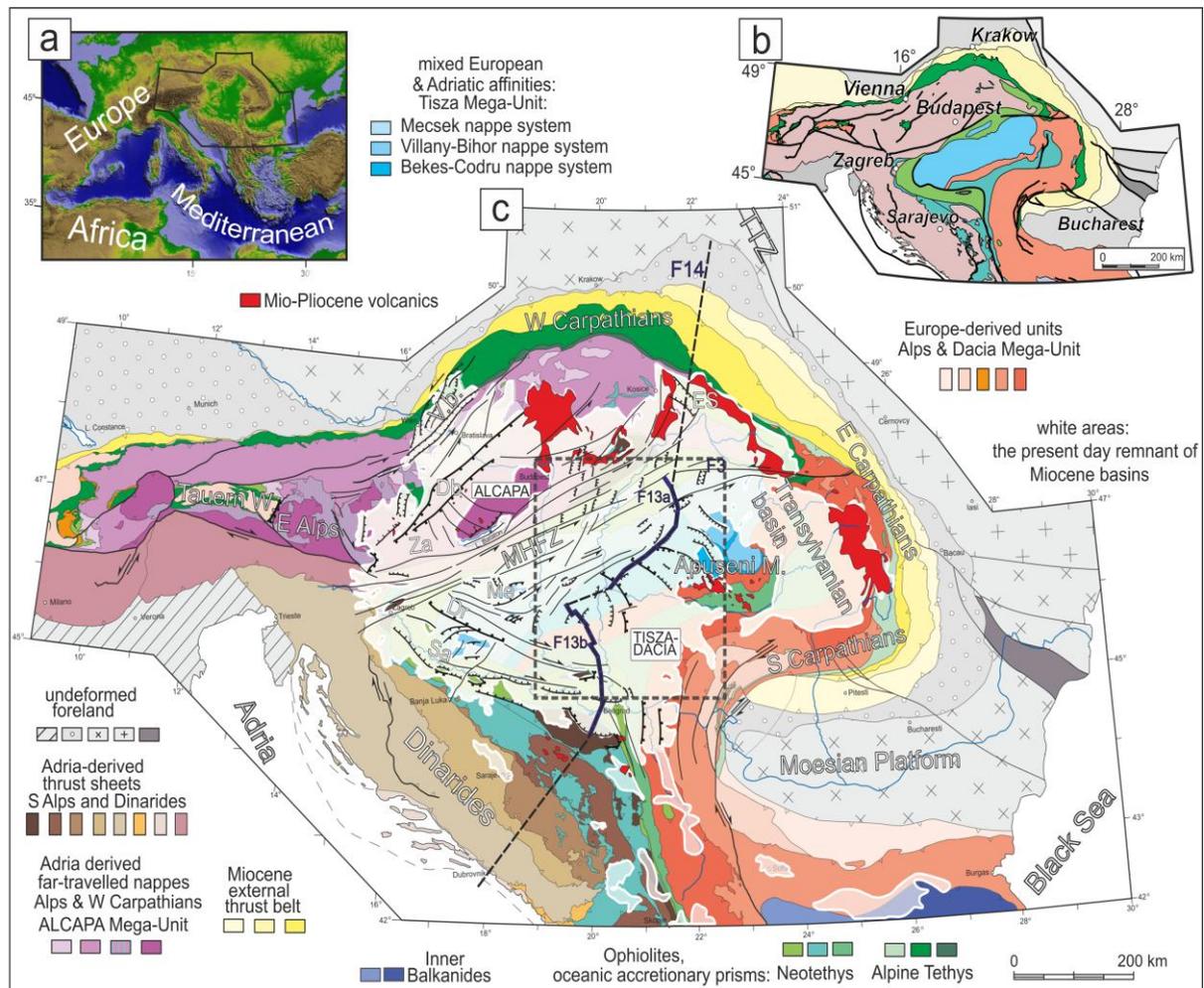


Figure 1 – a) Location of the Pannonian Basin system. b) Main tectonic units of the Alps-Carpathians and Dinarides [simplified after Schmid *et al.*, 2008]. The Tisza-Dacia contact is slightly modified. Note that the overlying Neogene basins in the intra-Carpathians area are ignored. c) Miocene - Quaternary 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 TTTZ has served as a rheological contrast zone localising Miocene tectonics of the Carpathians [e.g., Matenco *et al.*, 2016]. Vb – Vienna Basin, Dr – Dráva sub-basin, Sa – Sava sub-basin, Za – Zala sub-basin, Me – Mecsek hill, Db – Danube basin, ES – East Slovakian basin, TTTZ – Teisseyre-Tornquist Zone, MHFZ – Mid Hungarian Fault Zone.

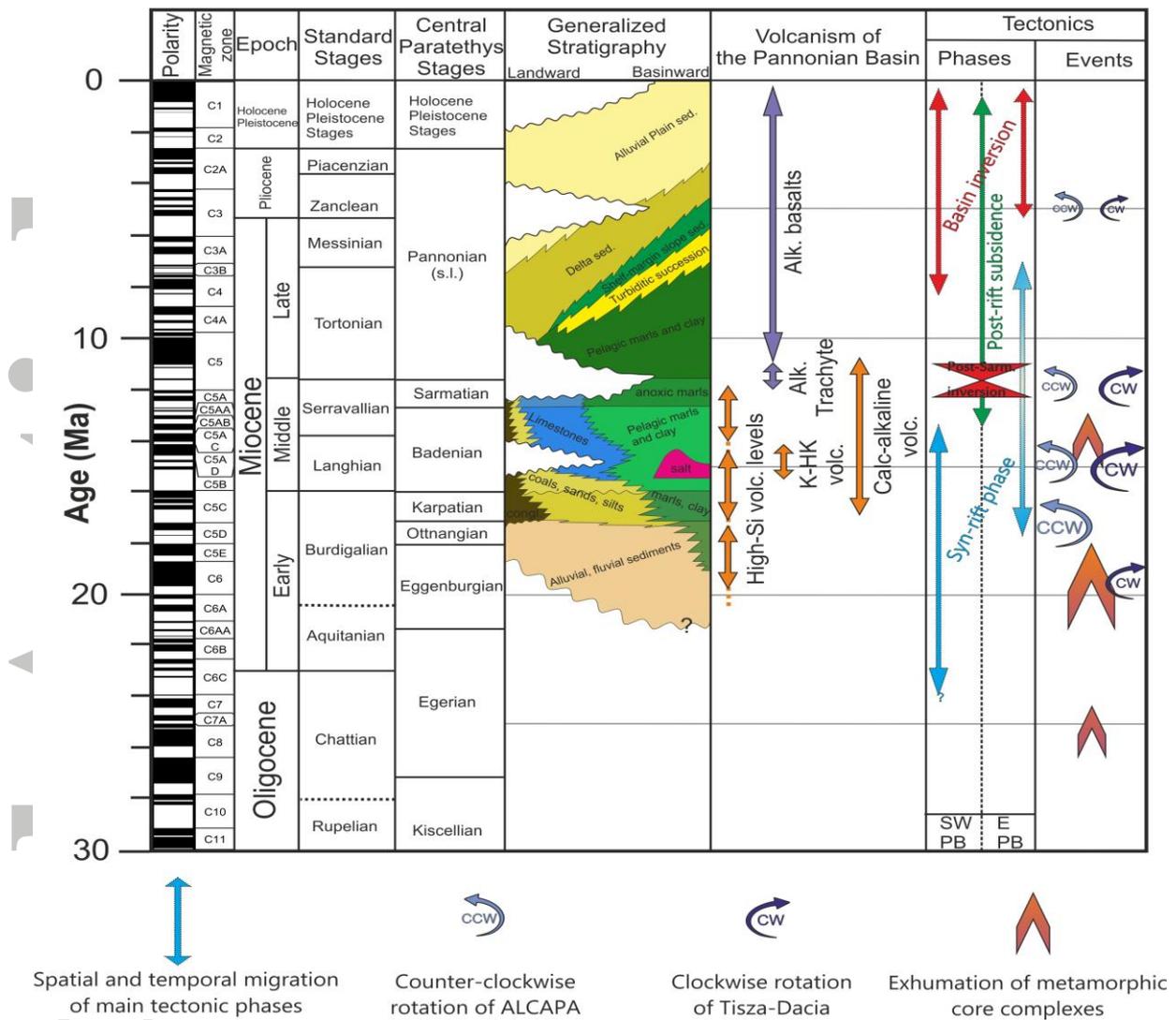


Figure 2 - Tectono-stratigraphic chart of the Great Hungarian Plain part of the Pannonian Basin with biostratigraphic correlation of the standard and Central Paratethys stages [after *Pezelj et al.*, 2013; *Piller et al.*, 2007] and the generalized Miocene lithostratigraphy of the study area, the volcanic activity of the Pannonian Basin [after *Harangi and Lenkey*, 2007; *Pécskay et al.*, 2006] and the main tectonic phases and events [modified 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 syn-rift/post-rift boundary and the onset of last stage of basin inversion appear to be older in the SW and progressively younger E-NE -wards.

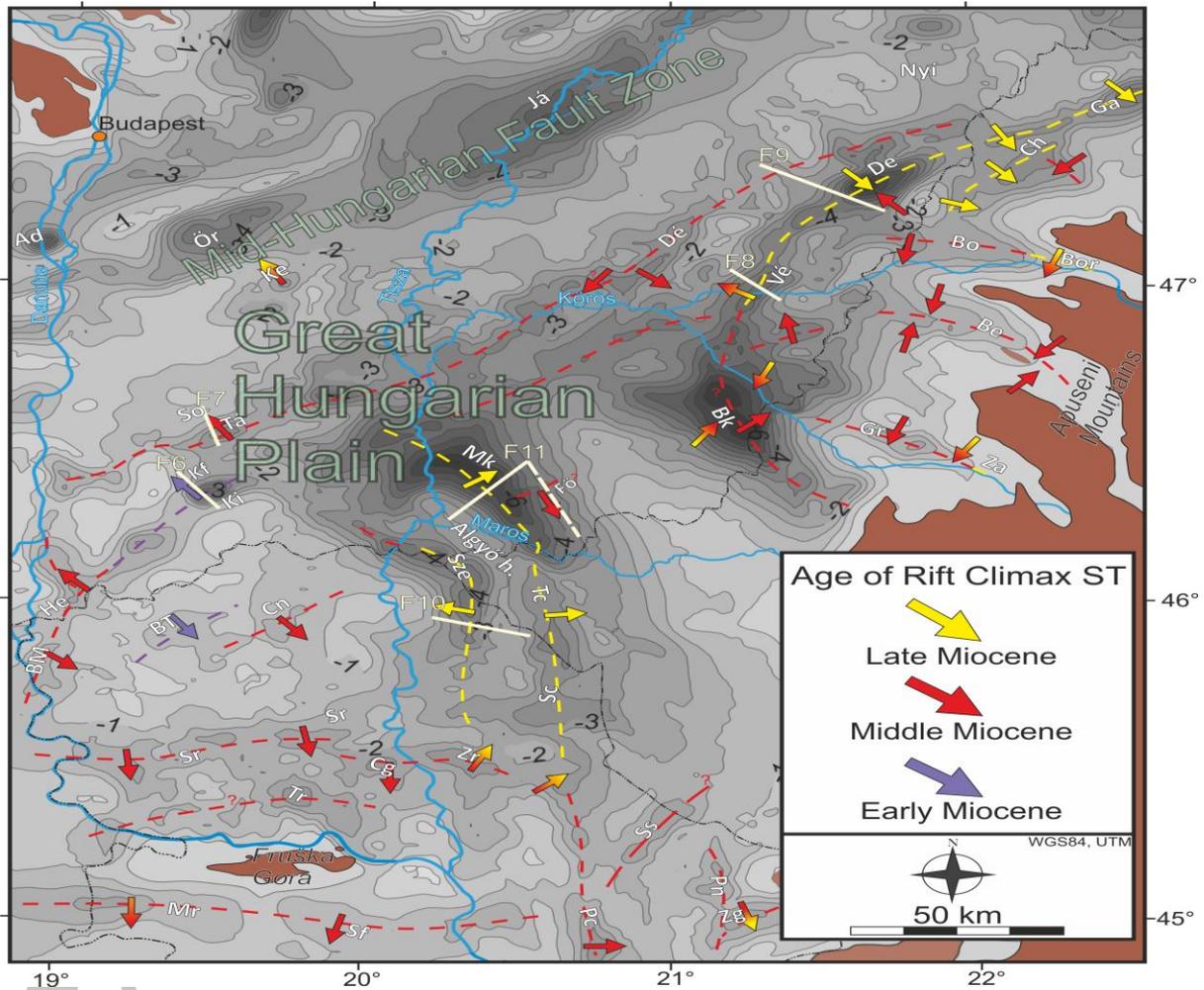


Figure 3 - Neogene basement isopach [in kilometers, compiled from *Haas et al.*, 2010; *Matenco and Radivojevic*, 2012; *Rabagia*, 2009; *Tulucan*, 2007 and the data of this study] and tectonic map of the Great Hungarian Plain. The map indicates the main depocenters with their strike (dashed lines), age of rift climax system tract and the main direction of extensional tectonic transport in the various sub-basins (the same color-coded arrows). White lines show the locations of the interpreted seismic sections. Main sub-basins of the area: Nyí – Nyírség sub-basin, 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 sub-basin, Ki – Kiskunhalas Trough, Kf – Kunfehértó sub-basin, Za – Zarand Depression, Gr – Graniceri Depression, Bk – Békés Basin, Mk – Makó Trough, Fö – Földeák sub-basin, 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.

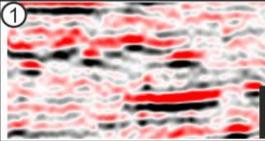
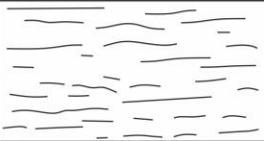
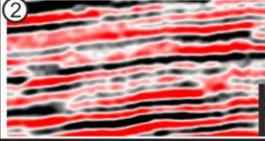
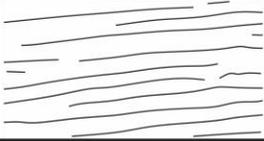
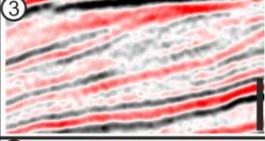
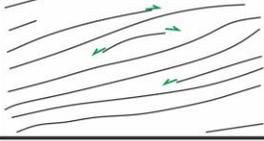
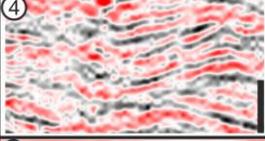
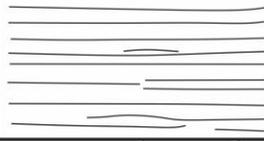
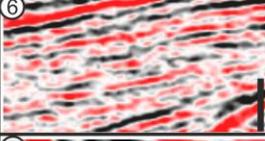
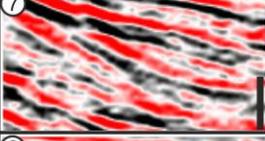
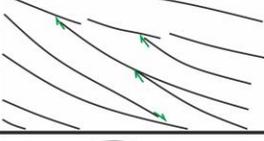
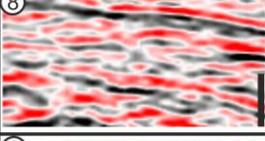
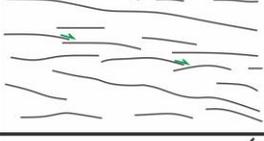
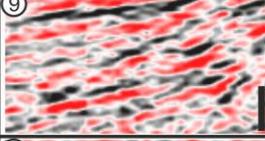
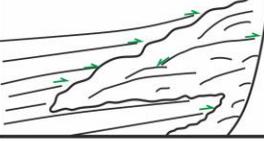
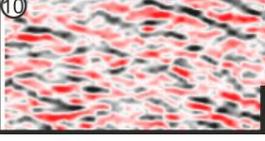
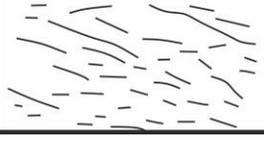
<b>Seismic facies unit</b>	<b>Seismic examples (vertical scale represents 100 ms)</b>	<b>Line drawing interpretations</b>	<b>Amplitude and frequency characteristics</b>	<b>Spatial distribution/ typical occurrence</b>
Parallel discontinuous			Low to high amplitude, medium frequency	Occurs at shallower depths (alluvial plain sed.)
Sub-parallel fairly continuous			High amplitude, high frequency	Occurs above the prograding clinofolds (delta sed.)
Cliniform - Continuous, discontinuous alternating			Medium to low amplitude, high frequency	Occurs between deep and shallow depth sediments (prograding shelf-margin slope)
Hummocky discontinuous			Medium to low amplitude, high frequency	Occurs between deep and shallow depth sediments (prograding shelf-margin slope)
(Sub-)parallel continuous			Medium to high amplitude, low to high frequency	Present within and just above of the half-graben infill
Sub-parallel discontinuous			Low amplitude, medium to high frequency	Occurs within half-graben infill
Oblique			Variable amplitude, low to high frequency	Occurs within half-graben infill, usually prograding toward the basin center
Oblique hummocky			Low to high amplitude, low to medium frequency	Occurs within half-graben infill, near the flanks
Proximal lobe - hummocky to discontinuous			Low to medium amplitude, low frequency	Occurs within half-graben infill, next to boundary faults
Chaotic discontinuous			Low to medium amplitude	Occurs above the acoustic basement

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

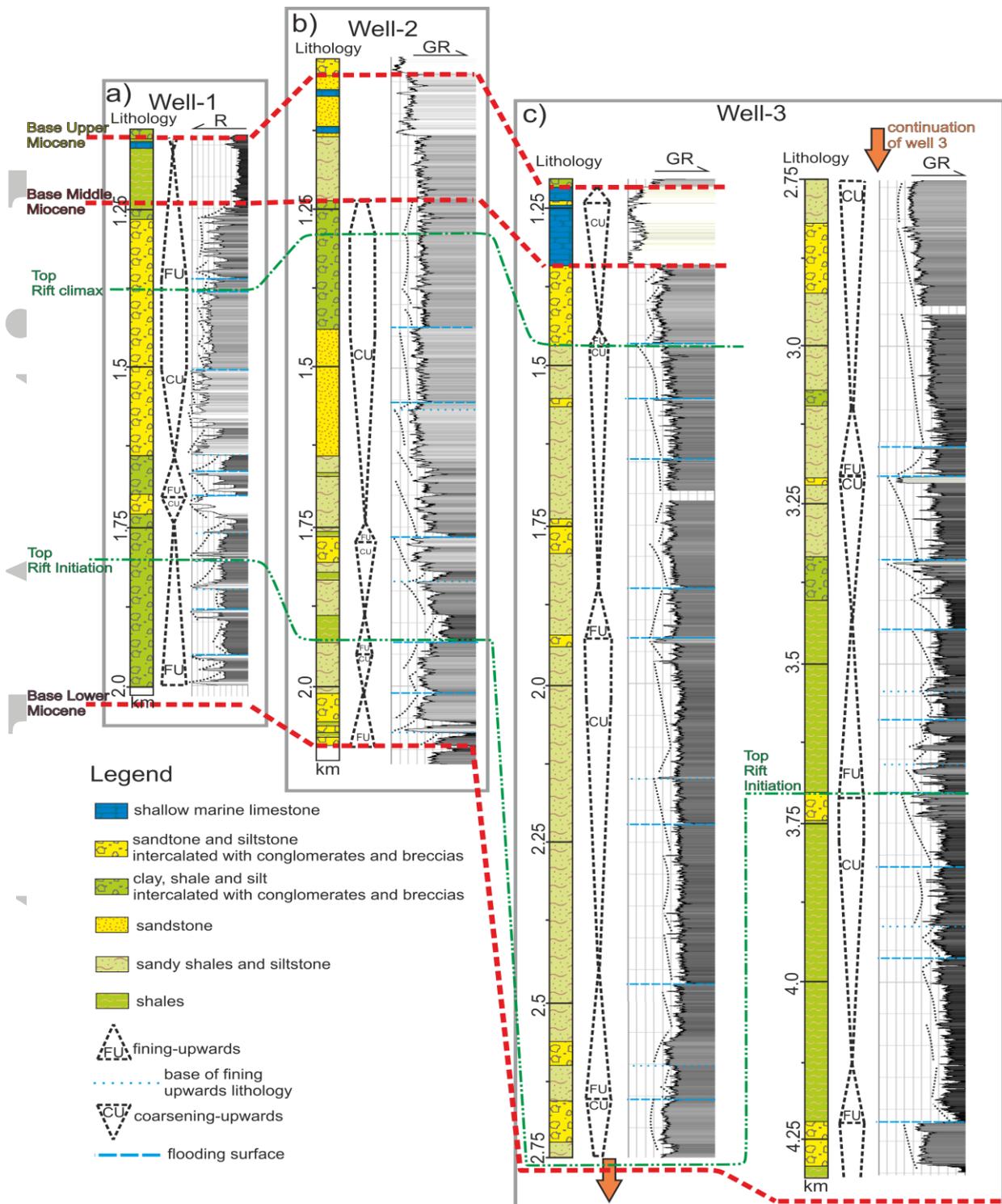


Figure 5 – Sequence stratigraphic well log interpretation from the Kiskunhalas sub-basin showing cyclicity of the Neogene basin fill. Generalized lithological column is based on well reports. Locations of the wells are displayed in Figure 6.

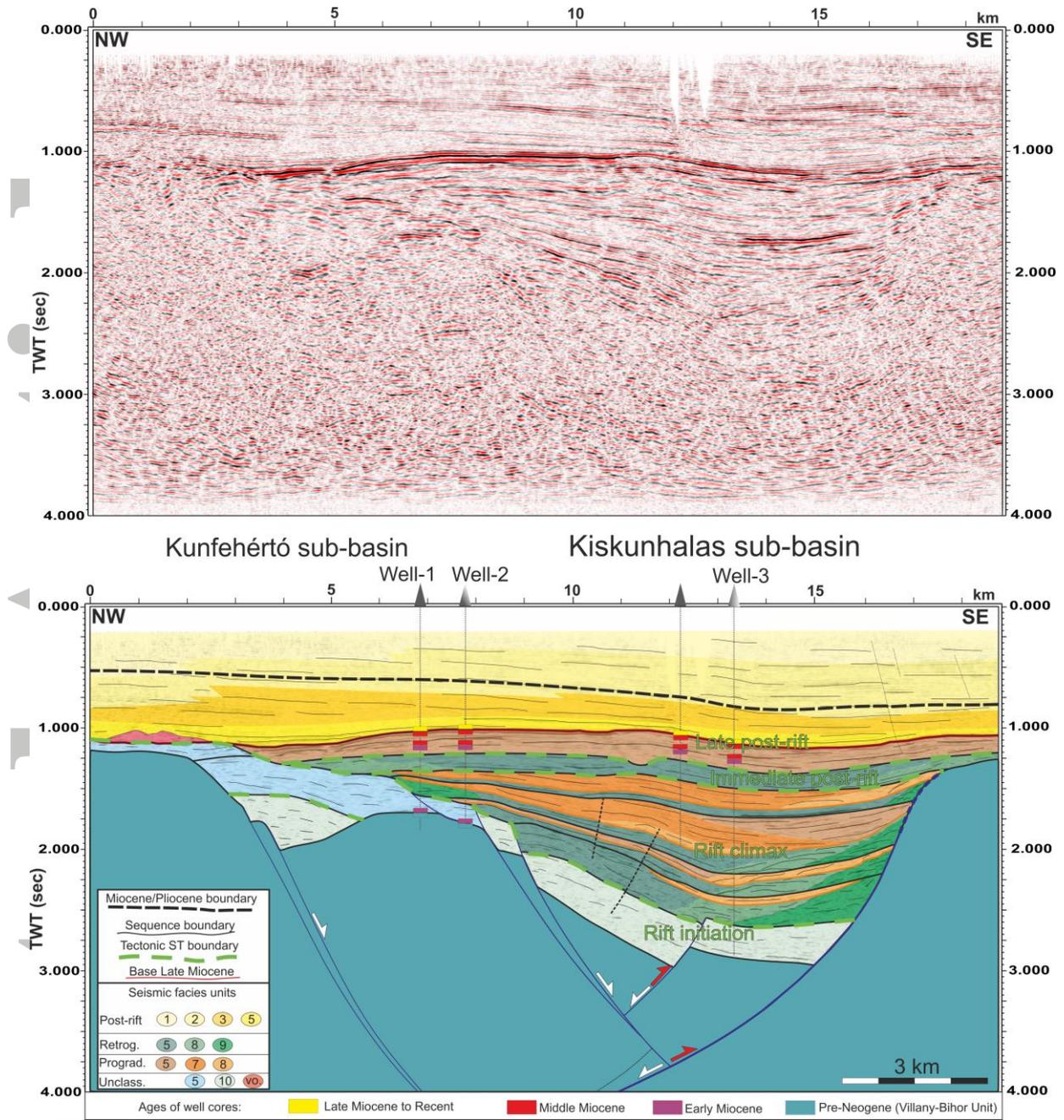


Figure 6 – Non-interpreted (top) and interpreted (bottom) reflection seismic section from the Kiskunhalas Trough. Location is displayed in Figure 3. White arrows indicate the Miocene kinematics of faults, red arrows show the late Middle - early Late Miocene inversion of the structure.

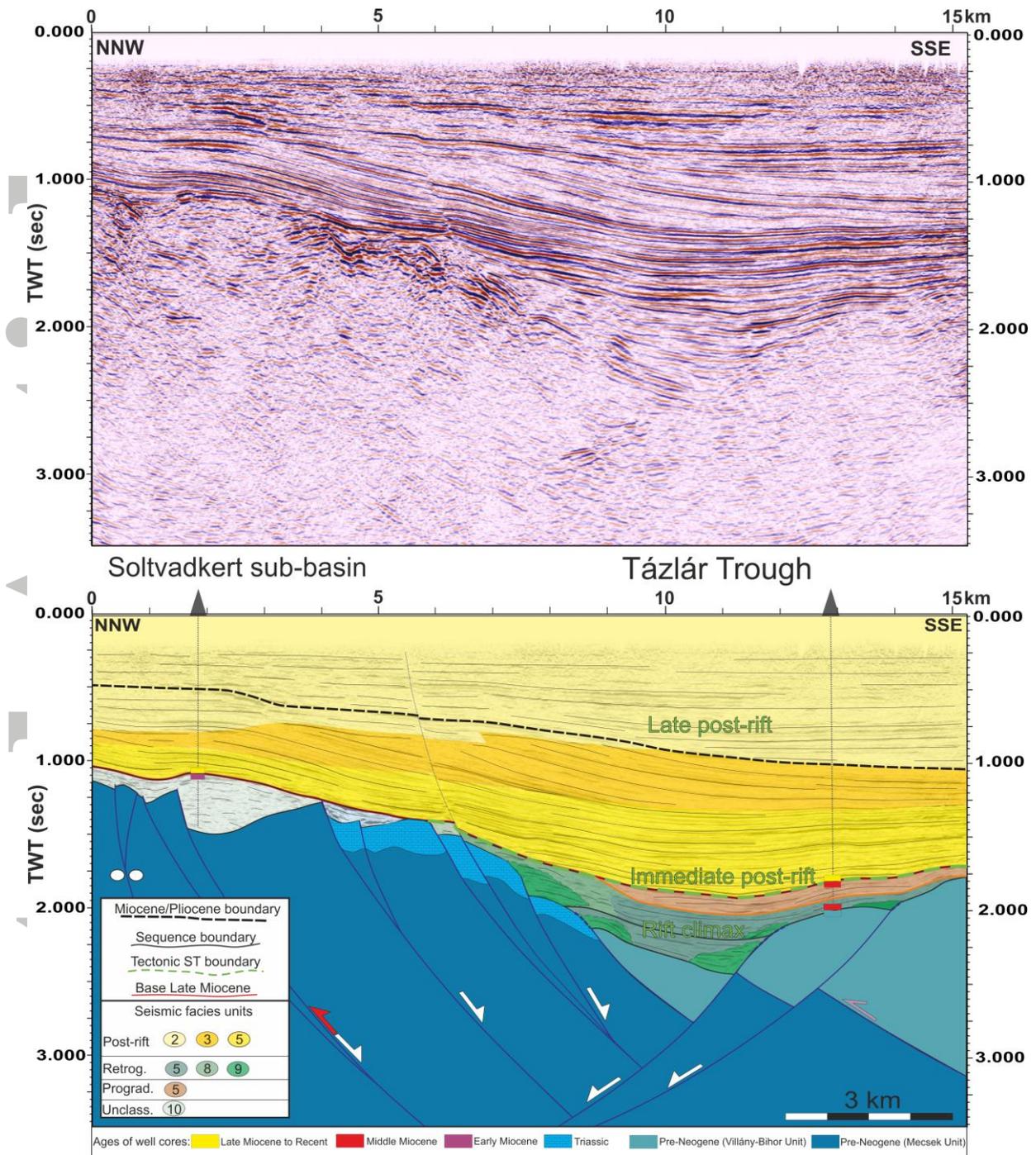


Figure 7 – Non-interpreted (top) and interpreted (bottom) reflection seismic section from the Tázlár Trough. For location, see Figure 3. Grey arrow indicates Cretaceous thrust. White arrows indicate the Miocene kinematics of faults. Red arrow indicates the possible inversion of the fault of the small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

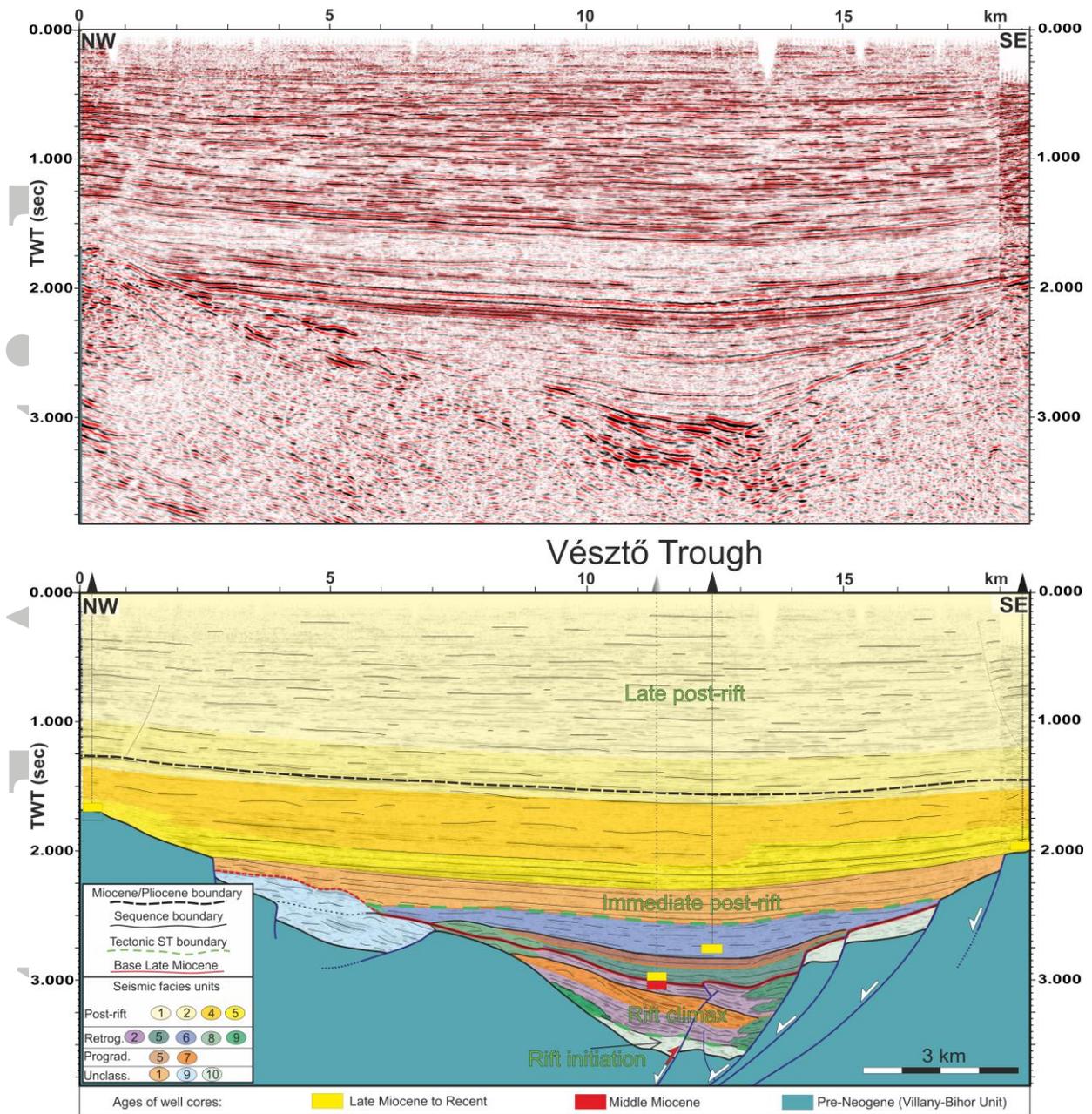


Figure 8 – Non-interpreted (top) and interpreted (bottom) reflection seismic section from the Vésztő Trough. For location, see Figure 3. White arrows indicate the Miocene kinematics of faults. Red arrow indicates positive inversion of the Miocene normal fault during the earliest Pannonian times.

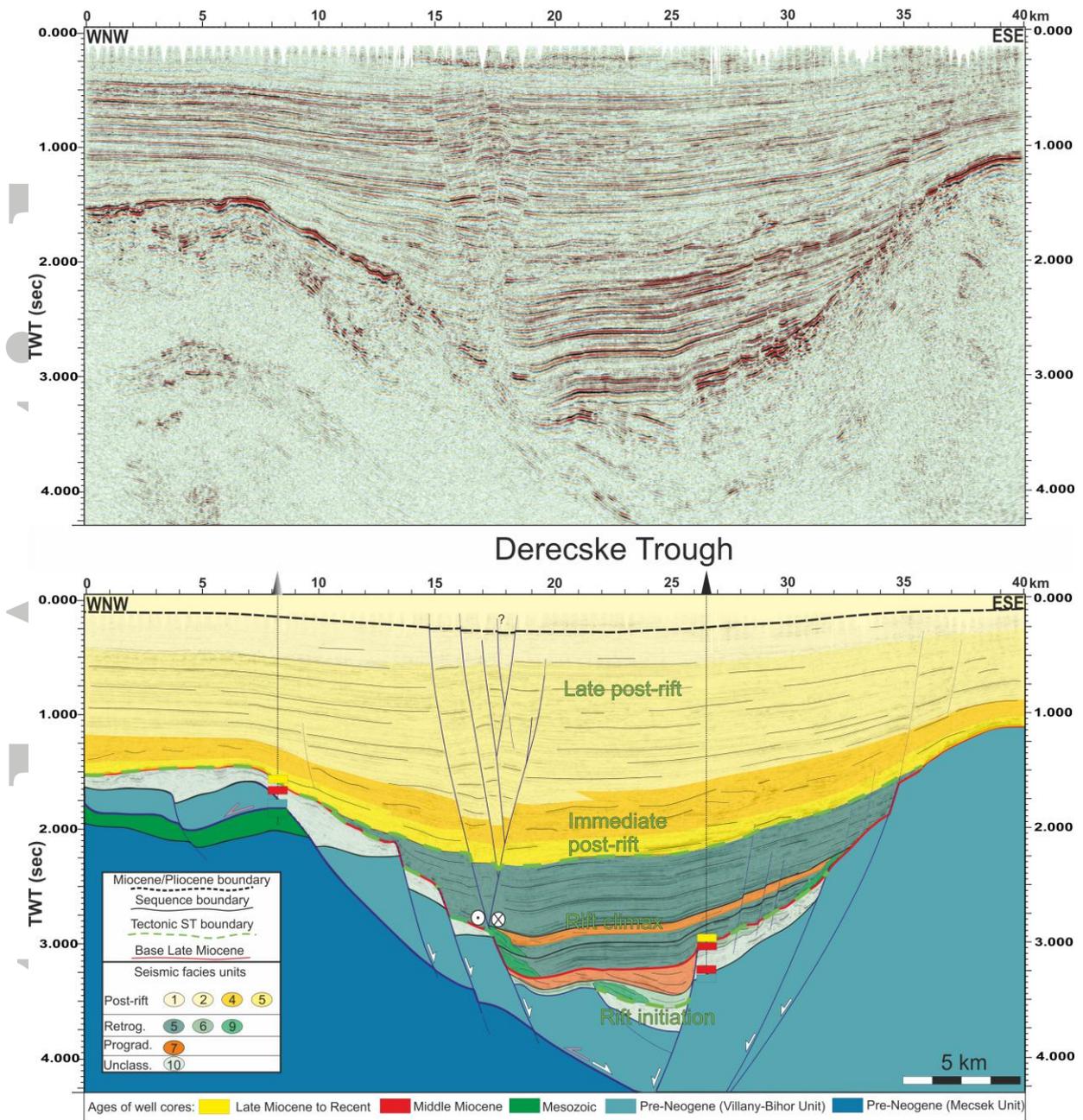


Figure 9 – Non-interpreted (top) and interpreted (bottom) reflection seismic section from the Derecske Trough. For location, see Figure 3. Grey arrows indicate Cretaceous thrusts [after Windhoffer *et al.*, 2005]. White arrows indicate the Miocene normal faulting. The fault zone reactivated again during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

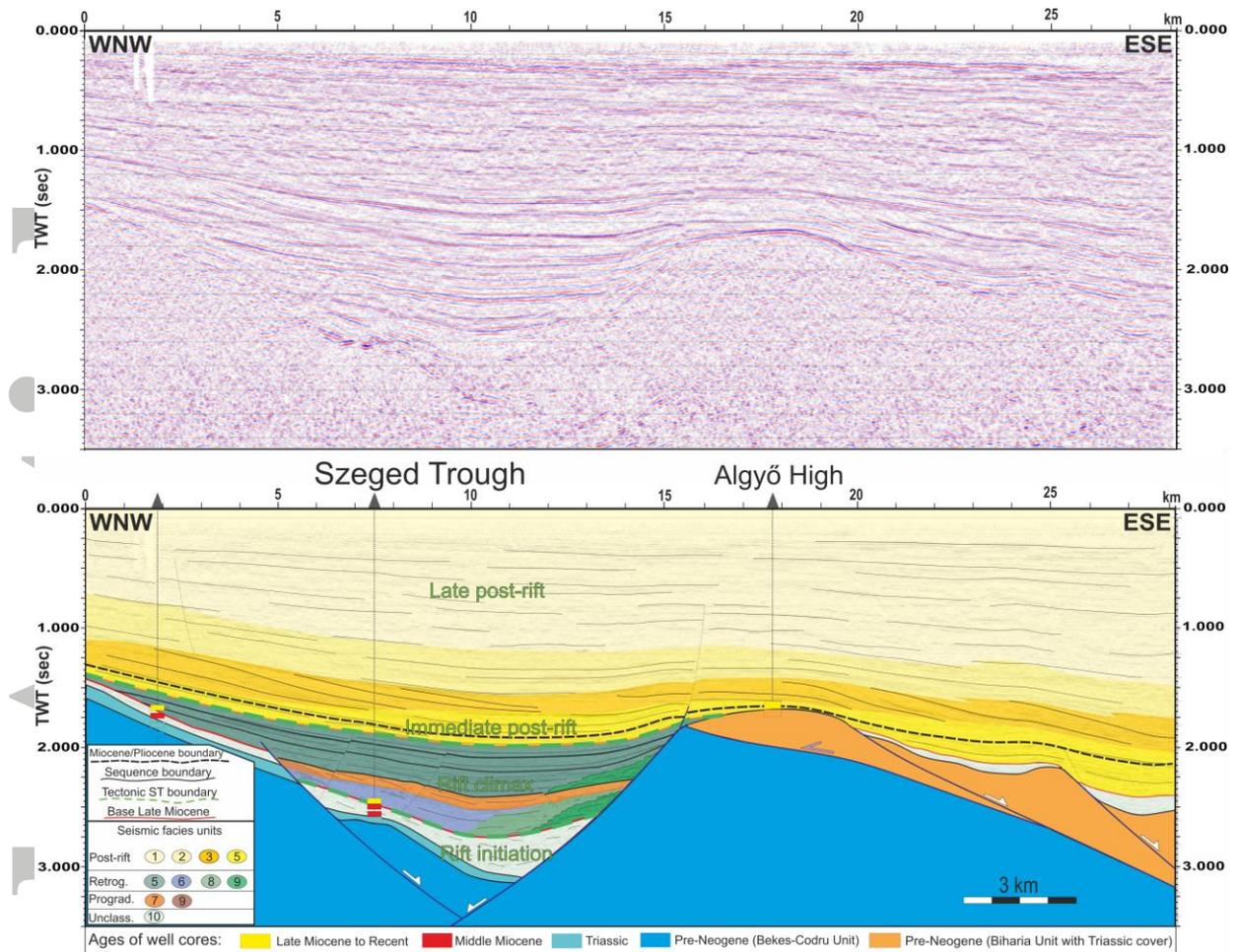


Figure 10 – Non-interpreted (top) and interpreted (bottom) reflection seismic section from the Szeged (Banatsko Arandelovo) Trough. For location, see Figure 3. Grey arrow indicates Cretaceous thrust of a piece of Dacia onto Tisza. White arrows indicate the Miocene kinematics of faults.

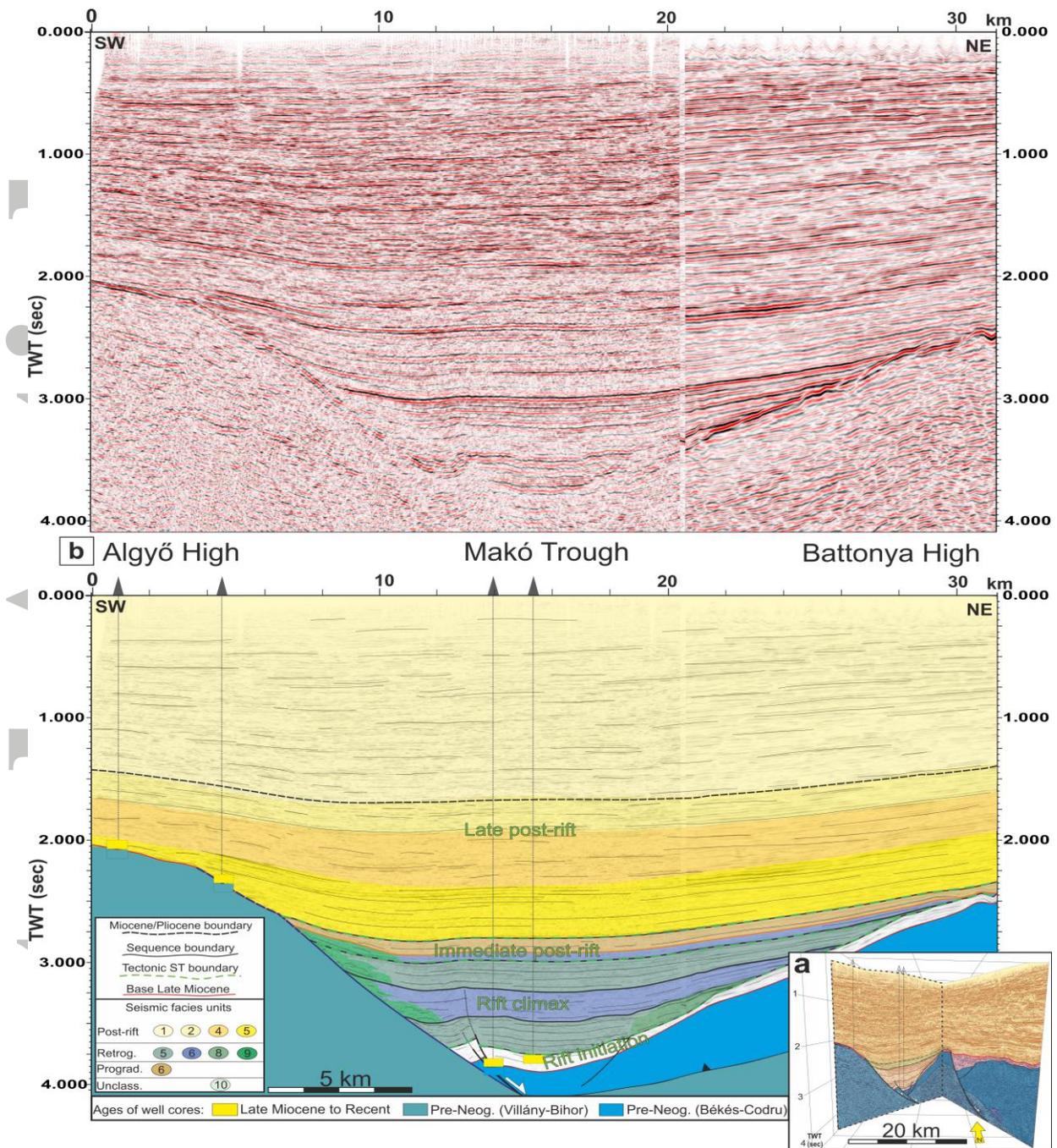


Figure 11 – Seismic data from the vicinity of the Makó Trough. (a) Composite section from the sub-basins showing perpendicular extensional structures. (b) Non-interpreted (top) and interpreted (bottom) reflection seismic section from the Makó Trough. For location, see Figure 3. In our interpretation 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.

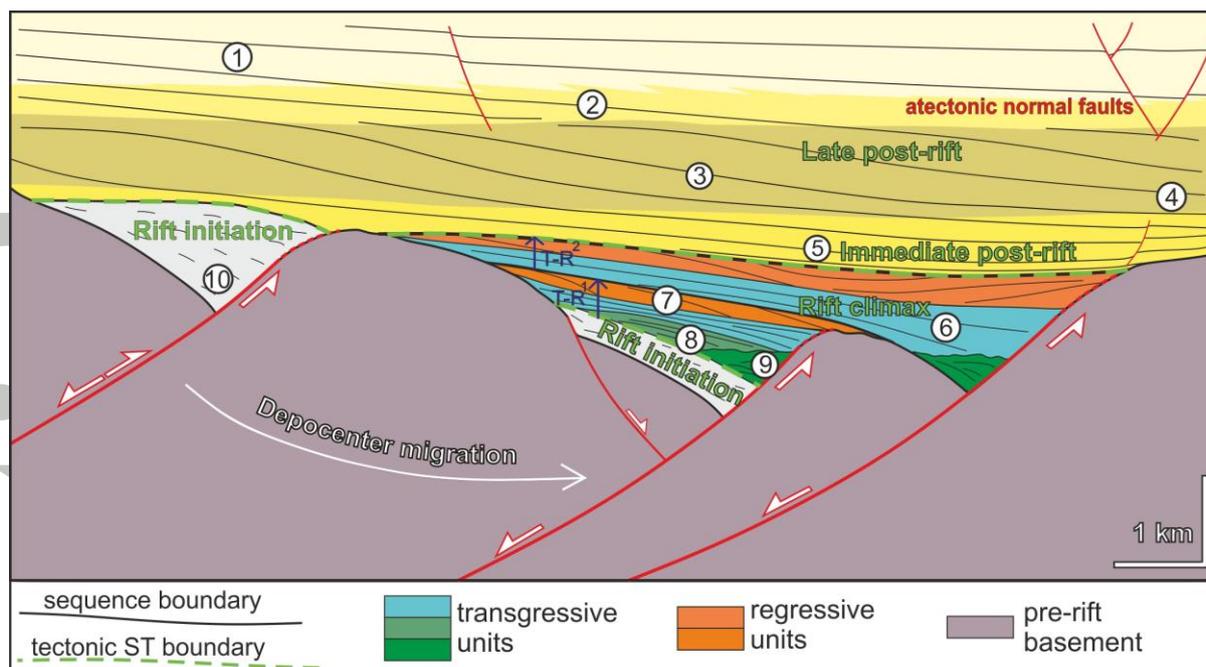


Figure 12 – Simplified stratigraphic model of the half-grabens of the Pannonian Basin [modified after *Prosser, 1993; Ter Borgh, 2013*]. Numbers represent the characteristic seismic facies units of a half-graben, cf. Figure 4. The characteristic retrograding and prograding seismic facies units are grouped 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 post-rift and late post-rift. 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.

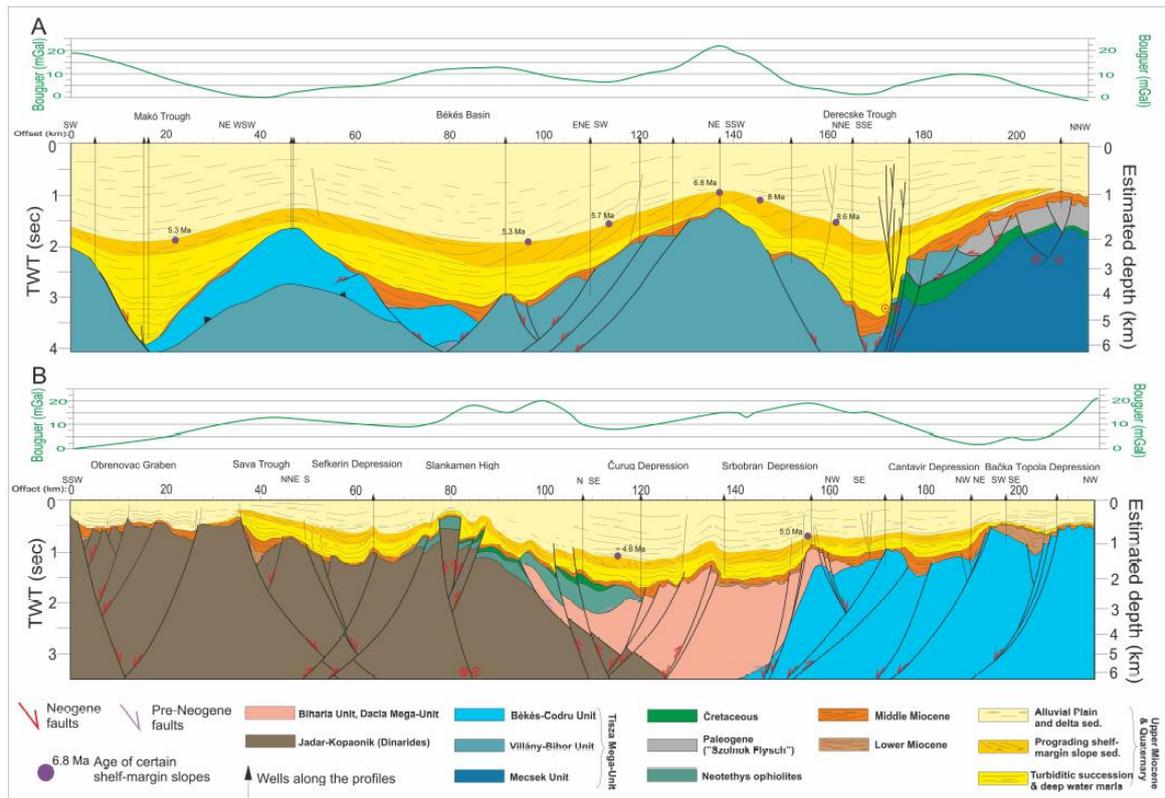


Figure 13 – Interpreted composite reflection seismic transects from the Pannonian Basin showing the main tectonic and stratigraphic features of the area. For location, see Figure 1. Note the striking difference of the thickness of Early to Middle Miocene and Late Miocene deposits. Late Miocene succession 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 syn-kinematic 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; *Tari et al.*, 1999]. b) Section from the southern and western parts of the Pannonian Basin is modified after *Matenco and Radivojevic* [2012], where syn-tectonic deposition is Lower to Middle Miocene. Neotectonic inversional structures have formed, for instance, at offset 80, 135 and 170 kilometers in section b. At the southern part of this section the low Bouguer anomaly values correspond to the crustal “root” of the Dinarides.

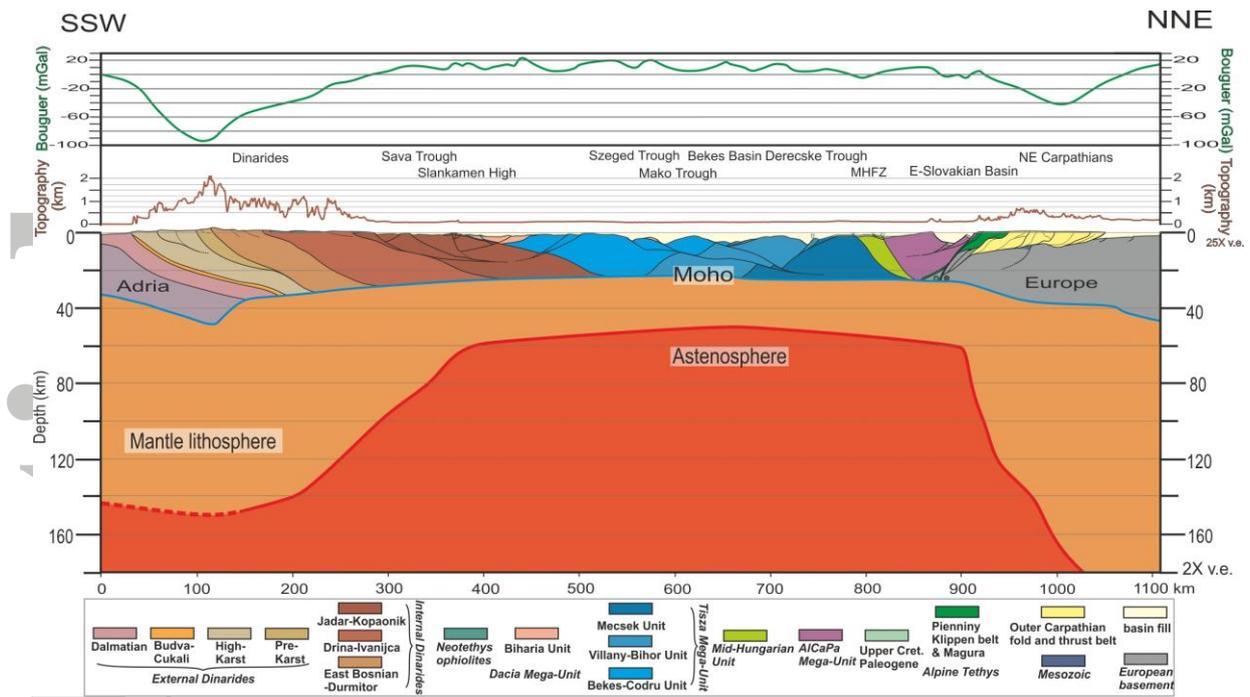


Figure 14 – Generalized lithospheric scale cross section over the Dinarides – Pannonian Basin – NW Carpathians. Lithosphere-asthenosphere boundary compiled from Artemieva *et al.* [2006], Tari *et al.* [1999], Tasarova *et al.* [2009]. Moho depth derives from Horváth *et al.* [2006], Janik *et al.* [2011]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

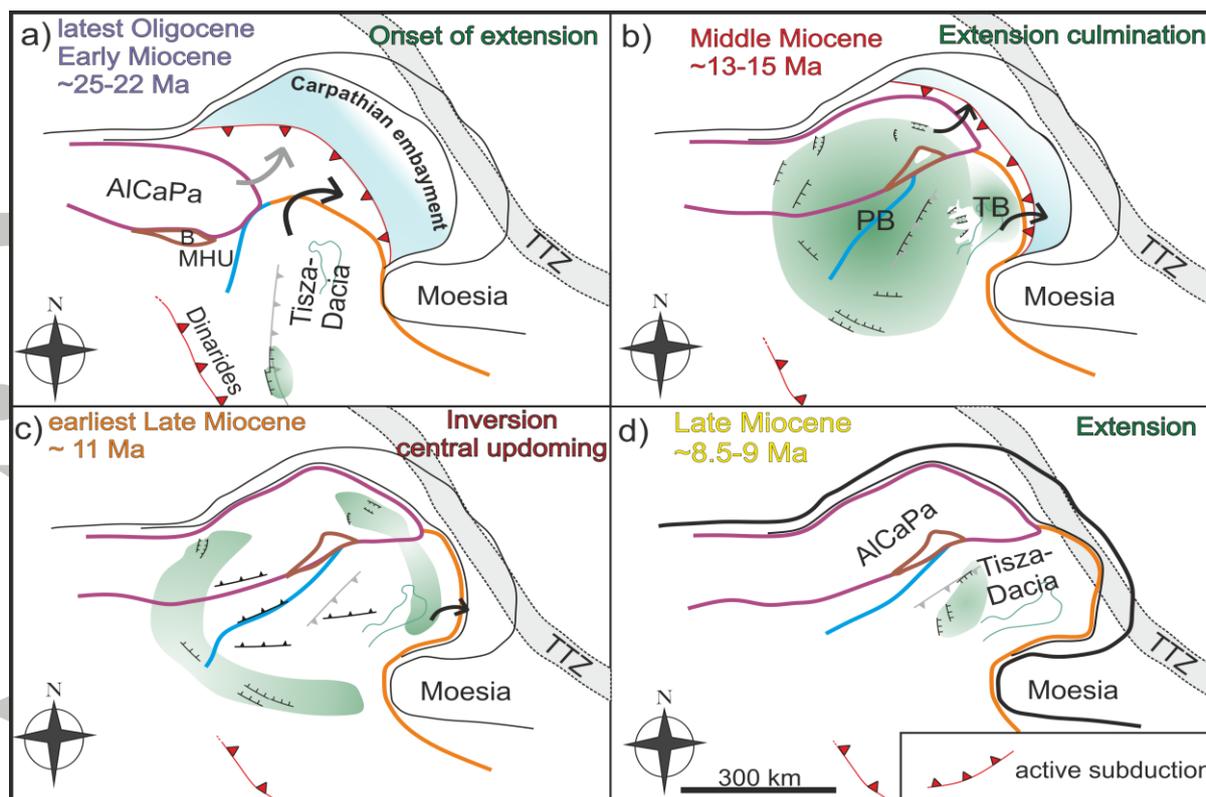


Figure 15 – Simplified sketch showing the syn-rift evolution of the Pannonian Basin [after Csontos and Nagymarosy, 1998; Faccenna et al., 2014, Fodor et al., 1999; Ustaszewski et al., 2008]. Green color shows the area of syn-rift subsidence. Coeval vertical axis CW rotation of the Tisza-Dacia and CCW rotation of the AICaPa Mega-Units and the Carpathians and Dinaridic slab roll back 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; Pharaoh, 1999).