Morphology of a large paleo-lake: analysis of compaction in the Miocene-Quaternary Pannonian Basin

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Abstract

Lake-floor morphologies may be significantly different from seafloor topographies of other basins, typically observed in passive or active continental margins. The bathymetry of large paleo-lakes is often overwritten by subsequent tectonic evolution, burial beneath thick overburden and inherent compaction effects. We study the evolution of such an initial underfilled, balance fill and finally overfilled large paleo-lake basin by the interpretation of 2D and 3D seismic data set corroborated with calibrating wells in the example of the Neogene Pannonian Basin of Central Europe. Lake Pannon persisted for about 7-8 Myr and was progressively filled by clastic material sourced by the surrounding mountain chains and transported by large rivers, such as the paleo-Danube and paleo-Tisza. We combined sedimentological observations with a backstripping methodology facilitated by well lithology and porosity data to gradually remove the sediment overburden. This approach has resulted in a morphological reconstruction of the former depositional surfaces with special focus on the prograding shelf-margin slopes. Our calculations show that the water depth of the lake was more than 1000 meters in the deepest sub-basins of the Great Hungarian Plain of the Pannonian Basin. The significant compaction associated with lateral variations of Neogene sediment thicknesses has created non-tectonic normal fault offsets and folds. These features have important effects on fluid migration and hydrocarbon trapping. We furthermore compare the geometries and effects of such non-tectonic features with the activity of larger offset sinistral strike-slip zones using 3D seismic attributes.

Keywords: Lake Pannon, bathymetry, compaction, Pannonian Basin, shelf-margin

1. Introduction

Deep lake basins formed in intra-continental settings affected by large amounts of extension can record the deposition of kilometres thick sediments (Katz, 1990). Paleo-water depth and the sedimentary architecture are controlled by several external forcing factors; their effects and interactions show marked differences from open marine environments (Martins-Neto and Catuneanu, 2010; Sztanó et al., 2013). Lakes are more sensitive to regional climate by the primary control of the local balance between precipitation and evapotranspiration (e.g., Carroll and Bohacs, 1999). In contrast to passive margins, the subsidence and/or uplift rates in intra-continental settings are also more variable (Xie and Heller, 2009). Lakes are sensitive to episodic (dis)connections with other neighbouring basins through the separating gateways, which are controlled by tectonics and lake level variations (e.g, Leever et al., 2011; ter Borgh et al., 2013; Matenco et al., 2016). This overall interplay between tectonics, lake level variations, sedimentation rates and transport routing results in spatially and temporally heterogeneous depositional environments (Garcia-Castellanos et al., 2003; de Leeuw et al., 2012; ter Borgh et al., 2015).

A typical example where a high-resolution data set is available for the analysis of the formation and evolution of a paleo-lake is the Pannonian Basin of Central Europe (Fig. 1). The paleo-Danube and paleo-Tisza rivers discharged large volumes of sediments into Lake Pannon during Late Miocene - Early Pliocene times in a sink area that roughly comprised the Vienna, Pannonian and Transylvanian basins. The lake persisted for 7-8 Myrs and was progressively filled by, and buried under, clastic material sourced by the surrounding mountain chains (e.g., Magyar et al. 2013). The long-standing hydrocarbon exploration activity of the basin has resulted in the availability of high-resolution geophysical data including well logs and 2D/3D seismic data (e.g., Bérczi and Phillips, 1985; Royden and Horváth, 1988; Pogácsás et al., 1988; Juhász, 1991; Grow et al., 1994; Vakarcs et al., 1994; Saftic et al., 2003; Magyar et al., 2006;

Sztanó et al., 2013) that allow a high prospectivity for conventional and unconventional geo-resources including geothermal energy (e.g., Cloetingh et al., 2010; Horváth et al., 2015). Lacustrine organic-rich shales define good hydrocarbon source rocks, while deep-water turbidites, deltaic and fluvial sand bodies are important reservoirs (Saftić et al., 2003; Magyar et al., 2006; Tari and Horváth, 2006).

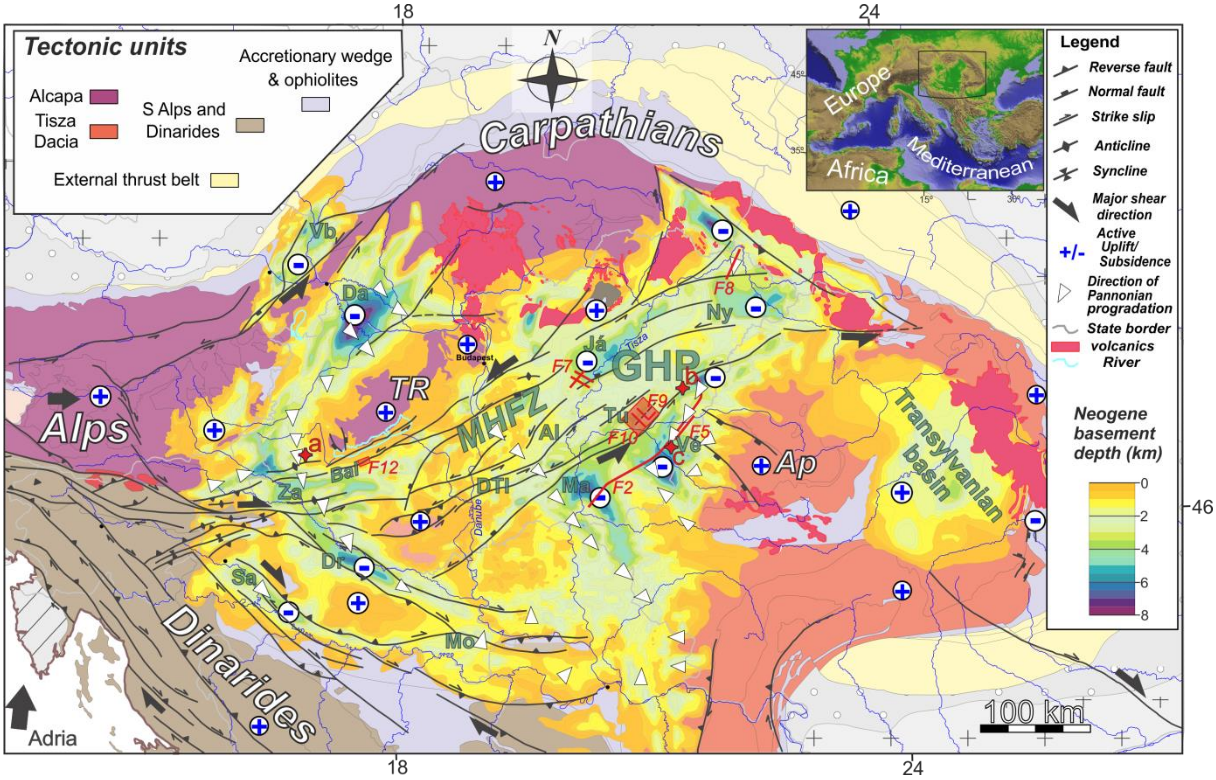


Figure 1. Tectonic map of the Pannonian Basin and adjacent areas showing the neotectonic fault pattern and active differential vertical movements (modified after Bada et al., 2007) overlain by the depth of the pre-Neogene basement. The tectonic units of the pre-Tertiary basement outcropping on the flanks of the basin are simplified after Schmid et al. (2008). GHP – Great Hungarian Plain, Vb – Vienna basin, MHFZ – Mid-Hungarian Fault Zone, Bal – Balaton Fault zone, TR – Transdanubian Range, Ny – Nyírség sub-basin, Já – Jászság sub-basin, Al – Alpár sub-basin, Ma – Makó Trough, Vé – Vésztő Trough, Tú – Túrkeve Trough, DTI – Danube-Tisza interfluve, Da – Danube basin, Za – Zala sub-basin, Dr – Drava Trough,

Sa – Sava Trough, Mo – Morovic depression, Ap – Apuseni Mountains. Well locations of Figure 4 (a,b,c) are marked by red cross symbols.

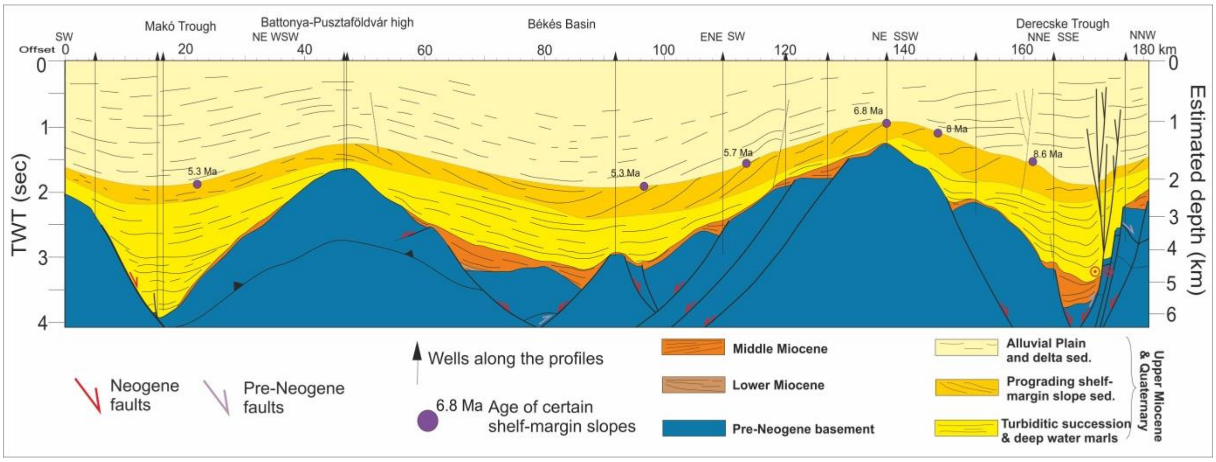


Figure 2. Interpreted composite seismic section from the eastern part of the Great Hungarian Plain (modified after Balázs et al., 2016). For location see Fig. 1. Note the long wavelength folding of the young sediments partly caused by compaction effects.

In order to understand the morphology of depositional surfaces and evolution of such a deeply buried lacustrine system, we have performed 2D and 3D seismic interpretation and backstripping in the up to ~7 km thick Pannonian Neogene sediments (Fig. 2). Paleo-bathymetric estimates were derived by successive decompaction of prograding shelf-margin slope clinoforms based on the available lithology and porosity data from wells in different regions of the Pannonian Basin. We have analysed the spatial and temporal variation of clinoform geometries and shelf-edge trajectories (e.g., Helland-Hansen and Hampson, 2009; Henriksen et al., 2011; Rabineau et al., 2014) controlled by the interplay between high sediment fluxes, inherited pre-Neogene basement geometries, paleo-water depth, the rate of subsidence interrupted by periods of tectonically-induced uplift and climatically controlled lake level variations. We have furthermore analysed the effects of the few kilometres thick overburden and the variable relief of the basin floor in creating significant compaction effects, such as long wavelength folds and differential compaction induced faults (e.g., Magara, 1978; Williams, 1987; Xu et al., 2015).

2. Evolution of the Pannonian Basin and Lake Pannon

The Pannonian basin of Central Europe is a Neogene continental back-arc basin, where the 220-290 km of Miocene extension is accommodated by the roll-back of the Carpathians and Dinaridic slabs (Fig. 1, Ustaszewski et al., 2010; Matenco and Radivojevic, 2012; Faccenna et al., 2014; Horváth et al., 2015 and references therein). Extensional basin formation followed a pre-Neogene orogenic evolution that resulted from the opening and subsequent closure of two oceanic realms, the Triassic-Cretaceous Neotethys and Middle Jurassic – Paleogene Alpine Tethys (e.g., Schmid et al., 2008 and references therein).

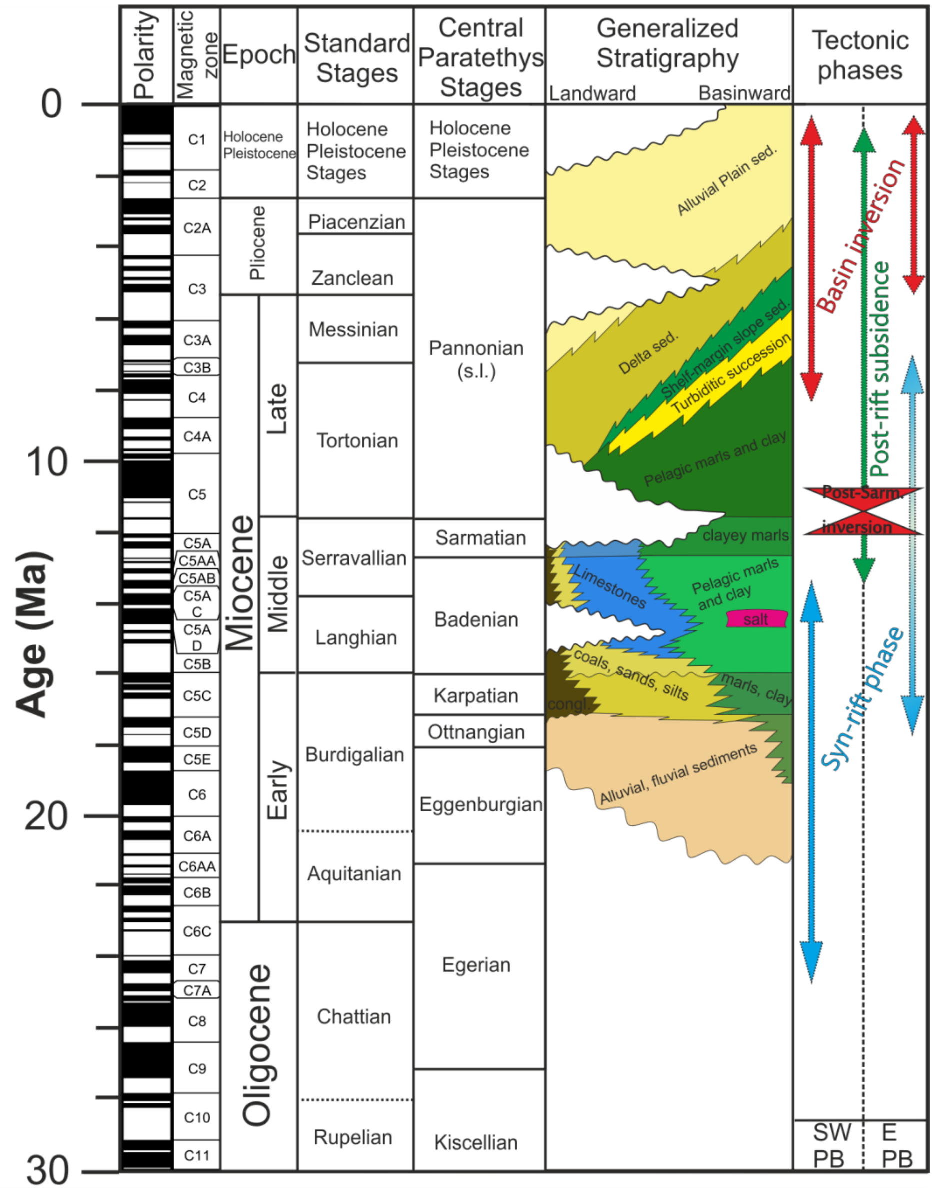


Figure 3. Tectono-stratigraphic chart of the Great Hungarian Plain part of the Pannonian Basin with correlation of the standard and Central Paratethys stages, the generalized Miocene lithostratigraphy of the study area and the main tectonic phases affecting the basin (modified after Balázs et al., 2016). Note that the syn-rift/post-rift boundary and the onset of the latest stage basin inversion are older in the SW and progressively younger E-NE -wards.

Starting from the late Eocene times the uplift of the Alpine – Himalayan mountain belt has gradually fragmented the larger Tethys Ocean and formed the Paratethys branch. The area of the future Pannonian Basin became part of the Central Paratethys, a semi-enclosed marine to lacustrine basin system (Báldi, 1989; Nagymarosy and Müller, 1988; Rögl and Daxner-Höck, 1996). Lower Miocene sediments were deposited in fluvial, lacustrine and locally marine conditions (Báldi, 1986; Nagymarosy and Hámor, 2012). The Middle Miocene is the time when the subsidence associated with extension resulted in the deposition of deep basinal sediments in the centre of extensional (half) grabens, while deposition along their margins was dominated by near-shore to shallow-marine conditions (Kováč et al., 2007; Nagymarosy and Hámor, 2012). The uplift of the Carpathians and Dinarides (ter Borgh et al., 2013) and further mantle dynamics (see Balázs et al., 2016) led to the formation of an unconformity between the Middle and Upper Miocene strata marking the disruption of connections with the Paratethys Sea and development of the large, brackish, isolated Lake Pannon (Fig. 3; Magyar et al., 1999). An up to 7 km thick sedimentary succession was deposited during Late Miocene to recent times in the Great Hungarian Plain, the area with recording most of the stretching in the Pannonian Basin (Figs. 1, 2, Horváth et al., 2015). The basin fill recorded an initial transgression resulting in a period of underfilled stage. It was followed by shelf margin and slope progradation fed by the influx of sediments via fluvial systems resembling the present-day Danube and Tisza rivers. The largest spatial extension of Lake Pannon was at ~9.5 Ma (Magyar et al., 1999), covering the Vienna, Pannonian and Transylvanian basins. The shelf-margin prograded about 500 km in

6 Myrs until the early Pliocene from the NW and NE in a ~S-SE direction, while minor progradation was recorded from other directions (Pogácsás et al., 1988; Vakarcs et al., 1994; Magyar et al., 2013; ter Borgh et al., 2015). The coeval sedimentation reflects the deposition of several diachronous lithostratigraphic formations (Fig. 2) that were deposited in response to the progradation from deep to shallow lake environments (Fig. 3, Bérczi and Phillips, 1985; Juhász, 1991; Sztanó et al., 2013). These associations are laterally variable from deep hemi-pelagic deposition (Endrőd Formation), turbidites (Szolnok Formation), shelf-margin slope (Algyő Formation) and delta (Újfalu Formation) to alluvial plain sediments (Zagyva Formation). Their typical seismic expression provides an excellent lateral correlativity of seismic facies units.

Extension and subsequent thermal subsidence in the Pannonian Basin was followed by a period of basin inversion that started at ~8 Ma (Uhrin et al., 2009), observed by accelerated differential vertical movements and fault reactivations (Horváth and Cloetingh, 1996; Fodor et al., 2005; Bada et al., 2007; Dombrádi et al., 2010). Active sinistral faults with ENE-WSW strike are interpreted in the centre of the basin and dextral shear zones with WNW-ESE strike at its southern margin (Fig. 1, Horváth et al., 2006). Several unconformities are observed during these times in the basin fill (e.g., Vakarcs et al., 1994). One unconformity is dated at ~6.8 Ma. Another unconformity is observed near the boundary between the Miocene and Pliocene (e.g., Vakarcs et al., 1994), being angular and locally erosional near the basin margins and passes to a correlative conformity towards the basin centre. These unconformities are variably interpreted as either related to basin inversion (Sacchi et al., 1999; Magyar and Sztanó, 2008; ter Borgh et al., 2015), or formed in response to major lake level variations (Csató et al., 2015), or representing cross-over zones of different progradational directions reflected by onlap patterns in slope deposits (Magyar and Sztanó, 2008).

3. Data and methods

We have analysed a large array of 2D and 3D seismic data calibrated by a dense network of exploration wells. This analysis is illustrated by the selection of several key seismic lines and wells, generally oriented parallel with the direction of sediment transport (e.g., Fig. 4). The signal/noise ratio and resolution of the seismic sections are variable, and reflect the availability of data, from recent 3D seismic surveys to older 2D seismic lines. The vertical resolution averages 20-30 meters at the depth of 2-3 kilometres. Well-logs were tied to seismic sections using standard VSPs and check-shots, the error-bar is generally below the seismic resolution (see also Mészáros and Zilahi-Sebess, 2001).

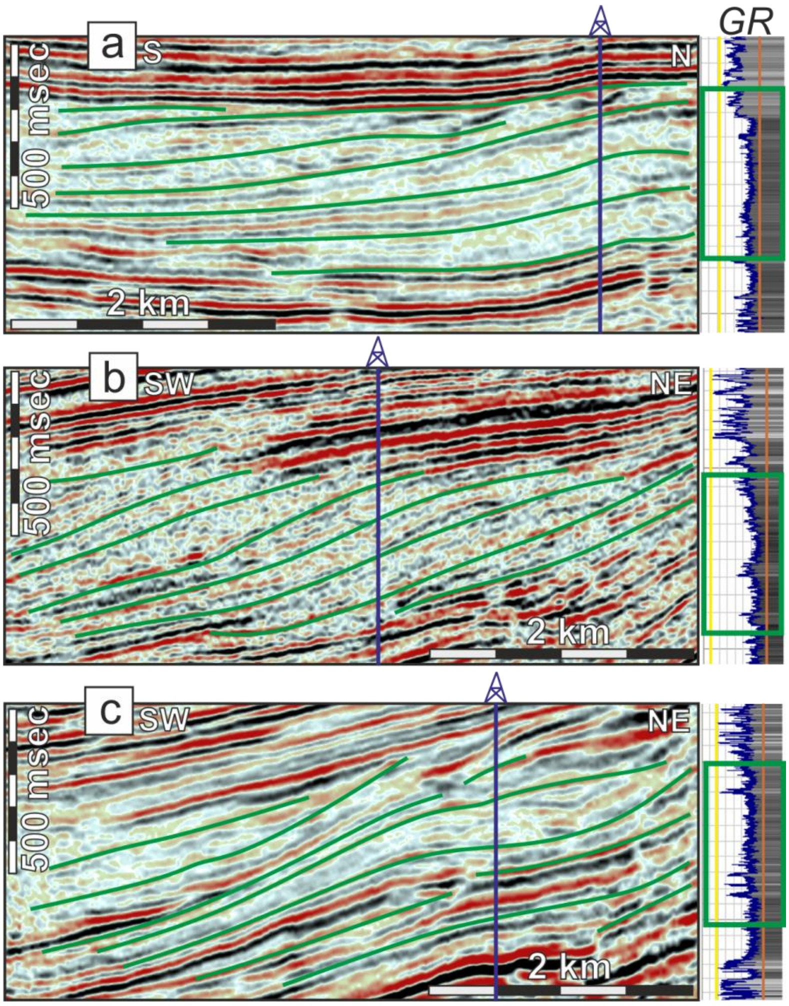


Figure 4. Seismic sections parallel with the direction of progradation and gamma ray logs showing the characteristic seismic facies and lithology of the prograding shelf-margin slope sediments. Slope clinoforms are indicated by green lines, green box indicates the interval of

slope sediments on well logs. Note the low-amplitude seismic facies, fine grained lithology of the unit and the gentle tilting post-dating the deposition of the slope sediments. Well locations (a-c) are displayed in Figure 1.

Our interpretation is focused on the prograding shelf-margin slope clinoforms connecting the shelf with the deep part of the basin. The slope sediments are associated with a medium to low amplitude, continuous-discontinuous alternating, high frequency seismic facies grouped in overall clinoform geometry (Figs. 4, 5 and 6, see also Magyar et al., 2013). In seismic lines oriented perpendicular to the direction of progradation (Fig. 7e) the seismic facies of the slope sediments is rather hummocky to chaotic. They often show incisions or canyons of variable magnitudes near the shelf or along the slope as well as turbidite channels and turbidite channel-levee complexes at the base of slope (see also Juhász et al., 2013; Sztanó et al., 2013).

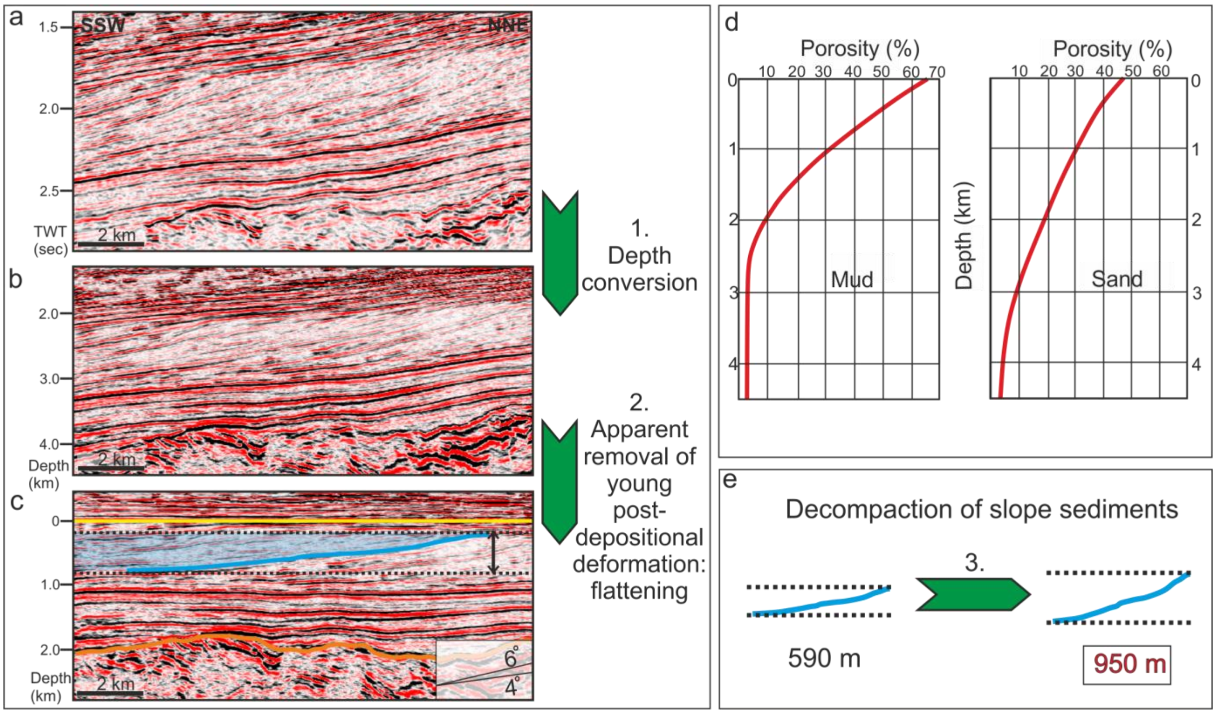


Figure 5. Methodology used for paleobathymetrical calculations. The TWT version of the seismic line (a) is converted to depth (b) and subsequently flattened (c) by using a horizon located immediately above the prograding clinoform sequence (in the overlying delta deposits,

indicated as 0 m depth). (d) We use a lithology dependent porosity-depth function available for these sediments in the Pannonian Basin (after Szalay, 1982) to decompact sediments and calculate the height of shelf-margin slope (e). The distance between topset and bottomset is 590 m and 950 m before and after decompaction, respectively.

We have performed first a sedimentological and seismo-stratigraphic interpretation by detecting reflection terminations and separating seismic facies units (e.g., Posamentier and Walker, 2006) It was followed by calculating a number of seismic attributes in 3D seismics that allowed a better differentiation of faults and sedimentary features (e.g., Cartwright and Huuse, 2005; Chopra and Marfurt, 2005). These attributes are particularly suitable to highlight paleo-geomorphological and structural features. We have used seismic amplitude values extracted on mapped horizons to highlight amplitude anomalies related to sharp acoustic impedance contrasts connected, for instance, with sharp lithological changes. We have also used spectral decomposition (e.g., Partyka et al., 1999) to produce amplitude and phase spectra for targeted windows over horizons. Different discrete frequency values were RGB colour blended and displayed on the interpreted horizon. We have calculated coherency attribute cubes based on the cross-correlation of seismic traces in selected windows to highlight structural features.

The bottom morphology of Lake Pannon was derived in a gradual procedure (Fig. 5). Seismic lines were converted to depth (Fig. 5a, b). On top of the lacustrine strata, the upper part of the basin fill is composed by delta and alluvial sediments deposited over a low and flat morphological relief (Sztanó et al., 2007). These sediments show deformation generally characterized by large open folds locally affected by faults with small vertical offsets. The areas affected by local faulting were generally avoided for lake morphology calculations. The effects of the gentle folding were restored by flattening the seismic lines to the first continuous reflector representing the paleo-horizon in the delta and alluvial sediments that is laterally continuous above the clinoforms along the seismic line. The distribution of these sediments (Újfalu and

Zagyva Formations) in seismic lines is very well controlled by available wells, where these have characteristic well-log expressions (Fig. 4, Bérczi and Phillips, 1985; Juhász, 1991). In seismic lines the first deposition of the delta deposits is observed as coherent high amplitude, low frequency continuous reflections facies overlying the topsets and clinoforms of the lacustrine progradation (Fig. 5). Given the resolution of the seismic lines, this type of restoration is a very good approximation of the morphology of Lake Pannon, affected by the subsequent compaction.

The seismo-stratigraphic interpretation has separated seismic facies units and seismic facies associations (e.g., Fig. 7) in the prograding clinoforms, which were converted into lithological facies units based on available well-logs (mostly gamma-rays, e.g., Fig. 4). The shelf-margin slope foresets are built up by about 80% mudstone combined with 20% sandstone (see also Szalay and Szentgyörgyi, 1988), only the upper and lowermost parts contain higher amounts of sand. Decompaction of the progradation geometry to derive the original morphology of Lake Pannon was achieved by a standard modelling technique (e.g., Angevine et al., 1990) based on the lithology dependent porosity-depth data available for the Great Hungarian Plain (Szalay, 1982; Dövényi, 1994). This 1D modelling was performed in successive places in the basin (Table 1). Note that the first continuous reflector of the delta and alluvial seismic facies may be at different depth across one section, due to the progradation/aggradation geometries. In places where a smaller scale delta progradation was detected in the shelf facies, the flattening was performed at the first continuous reflector overlying this secondary progradation. By connecting successive 1D decompacted geometries, the evolution of the lake morphology was reconstructed along each studied seismic line. This lake morphology gives a minimum estimation of the water depth. These calculations have a resolution close to the seismic one in the proximity of the lake shelf-margin slope, while at farther distances these estimates are less precise (Steckler et al., 1999). Based on existing sedimentological interpretations (Juhász, 1991; Sztanó et al., 2013), an additional 0-75 m water-depth characterized the shelf of the lake (where part of the deltaic sedimentation is located), while at farther distances from the progradation our calculation are just minimum estimates, the paleo-bathymetry could have been much deeper. It is likely that the overall paleo-bathymetry decreases with the approaching progradation by the distal infill of deep-water turbidites and more pelagic sedimentation.

4. Paleobathymetry of Lake Pannon

The general extensional geometry of the Pannonian Basin is characterized by individual sub-basins filled by 1 - 3.5 km of Lower to Upper Miocene syn-kinematic deposits, overlain by a 1.5 - 3.5 km thick post-extensional sedimentary cover. Here we focus on the prograding shelf-margin slope clinoforms that post-date the syn-extensional sedimentation to derive the paleobathymetry of the Late Miocene to Pliocene Lake Pannon.

4.1 Paleobathymetric calculations

Based on the flattened height of the Upper Miocene to Pliocene prograding shelf-margin slope clinoforms, paleobathymetric estimations by decompaction have been carried out in 8 representative sub-basins (Fig. 6, Table 1). Seismic section from the Nádudvar sub-basin of the central Great Hungarian Plain (Fig. 6b) shows the initial distribution of Pannonian sediments by prograding shelf-margin slope and delta sediments over deep-water marls and turbidites. This was followed by a base level rise at ~7.5 Ma (Juhász et al., 2007) associated with a major retrogradation and renewed deposition of deep-water sediments over the deltaic succession, overlain by renewed progradation and filling of the basin by deltaic and alluvial sediments in the upper part of the section (Fig. 6b). The calculated evolution of the lake morphology indicates 650 meters for the older clinoforms, up to few tens of metres for the deltaic environment and 200 meters of paleo-bathymetry for the upper, younger clinoforms.

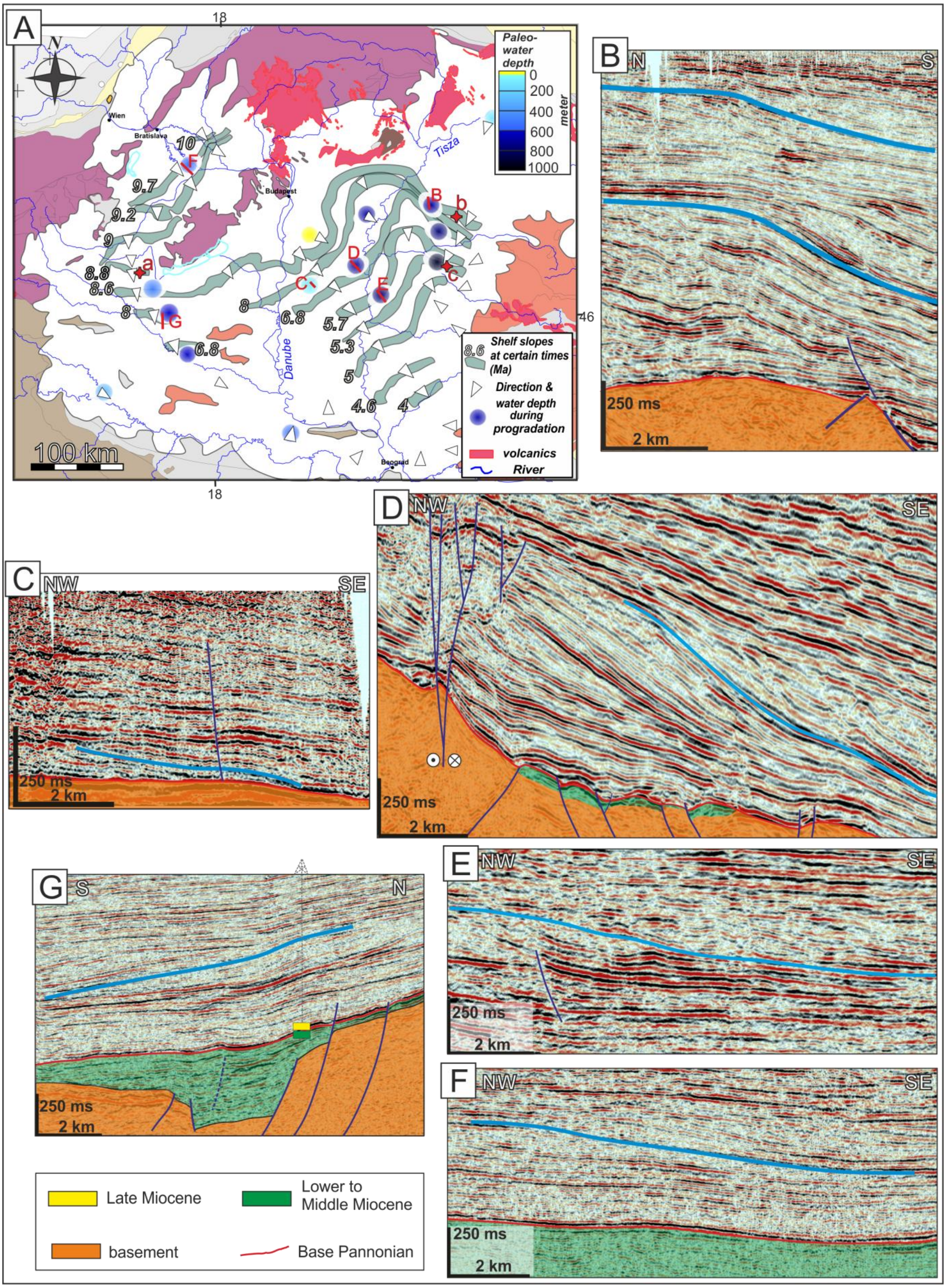


Figure 6. a) Positions of consecutive prograding shelf-margin slopes during the Miocene – Pliocene sedimentation (modified after Magyar et al., 2013). Blue circles indicate our calculated water depth values, different colours correspond to the paleo-water depth scale. Red cross symbols with small letters (a-c) show the well positions of Figure 4. B-F are the locations of the seismic sections in this figure showing the shelf-margin slopes used for paleobathymetric estimations; b) Seismic line in the Nádudvar sub-basin; c) Seismic section in the Danube-Tisza interfluve; d) Seismic section in the Alpár sub-basin; e) Seismic section in the Makó Trough; f) Seismic section in the Danube Basin; g) Seismic section in the Zala Basin.

Location	Age (Ma)	Section	Compacted height (m)	Decompacted height (m)
Jászság sub-basin	~ 7 Ma	Figure 7	440	690
Jászság sub-basin	~ 7 Ma	Figure 7	450	690
Jászság sub-basin	~ 7 Ma	Figure 7	370	580
Túrkeve sub-basin	~ 5.7 Ma	Figure 9, location a)	290	510
Túrkeve sub-basin	~ 5.7 Ma	Figure 9, location b)	350	630
Túrkeve sub-basin	~ 5.7 Ma	Figure 9, location c)	255	470
Túrkeve sub-basin	~ 5.7 Ma	Figure 9, location d)	455	740
N Nyírség sub-basin	~ 10 Ma	Figure 8, delta, location 1	48	70
N Nyírség sub-basin	~ 10 Ma	Figure 8, slope, location 2	91	150
Makó Trough	~ 5.7 Ma	Figure 6e	425	750
Nádudvar sub-basin	~ 7.5 Ma	Figure 6b upper blue line	180	200
Nádudvar sub-basin	~ 8.6 Ma	Figure 6b lower blue line	400	650
Danube Basin	~ 10 Ma	Figure 6f	280	550
Zala Basin	~ 8 Ma	Figure 6g	340	600
Alpár sub-basin	~ 7 Ma	Figure 6d	420	675
Danube-Tisza interfluve	~ 7.5 Ma	Figure 6c	75	130
Vésztő Trough	~ 5.3 Ma	Figure 5	590	950
Sava Trough	~ 6.5? Ma	*	185	275
Morovic depression	~ 4.5? Ma	*	350	525

Table 1. Height of the clinoforms before and after decompaction, the latter represents a minimum estimation of the paleo-water depth. *Seismic data used for our paleobathymetric estimation for the Sava Trough and Morovic depression are from Ustaszewski et al. (2014) and ter Borgh et al. (2015), respectively.

Seismic section from the central part of the Great Hungarian Plain between the present-day Danube and Tisza rivers (Figs. 1 and 6c) shows a thin prograding sequence with decompacted paleo-bathymetries of ~130 m that is laterally slightly higher in the area above the Middle Miocene (half) grabens. To the northeast, seismic sections form the Alpár sub-basin (Fig. 6d) show an inverted basement high to the NW, while the depth of this basement increases SE-wards. The calculated paleo-water depth is ~675 meters, the age of progradation in this area being 7-6.8 Ma (Magyar et al., 2013). We note that multiple phases of inversion and strike-slip deformation observed in the sediments overlying the Alpár sub-basin have also created a large incised canyon system at ca. 6.8 Ma (Juhász et al., 2013). Subsequently it was followed from ~5.3 Ma by continuous differential vertical movements creating the tilting observed in our seismic line (Fig. 6d). To the southeast, the ~5.7 Ma progradation observed in the Makó Trough (Sztanó et al., 2013) has ~750 m calculated paleobathymetries (Fig. 1, 6e), in the centre of this very deep sub-basin, decreasing to 650 m over its flanks (Balázs et al. 2015). In the western part of Lake Pannon, the SE-ward progradation of the paleo-Danube took place between 10-6.8 Ma (Magyar et al., 2013). The calculated paleo-bathymetries for this area are 550 m for the NW in the Danube basin (Figs. 1, 6f), 550 m for the Zala (Fig. 6g) and ~600 m for the Drava sub-basins (see also Balázs et al., 2015). These values are in agreement with earlier predictions in this area (Uhrin et al., 2009). In the southern part of the Pannonian Basin, paleo-bathymetric calculations in the Sava Trough and Morovic Depression (Fig. 1, seismic lines in Ustaszewski et al., 2014 and ter Borgh et al., 2015, respectively), indicating E-wards prograding clinoforms paleobathymetries of 275 m and N-ward prograding clinoforms paleobathymetries of 525 m, respectively.

4.2 The paleo-morphology of Lake Pannon

An appropriate place where the fine interplay between the creation of accommodation and sedimentation can be observed is the Jászság sub-basin of the northern part of the Great Hungarian Plain (Fig. 1). This sub-basin was filled by sediments transported by both the paleo-Danube from WNW and the paleo-Tisza from NE directions between ~8-6.8 Ma (Magyar et al., 2013). Two seismic lines (Figs. 7a,b) parallel with the local direction of progradation show the geometries of the shelf-margin slope, delta progradation on the shelf and toe of slope sediment complexes. An angular unconformity between the Miocene and Pliocene sediments (green line in Figs. 7a,b) marks the boundary of delta and alluvial environments as well. Shelf-edge trajectories (Fig. 7f) infer interplay between normal regression, forced regression (reflecting base-level drop of ~80 m), transgression and retrogradation (reflecting a base-level rise of ~200 m). Seismic lines perpendicular to the direction of progradation show that shelf incisions took place during both relative water-level rise and water-level fall (Fig. 7e). This means that such observed incisions or canyons are not necessarily subaerial. They can be also the result of slope failure during rapid transgression (cf., Fongngern et al., 2016).

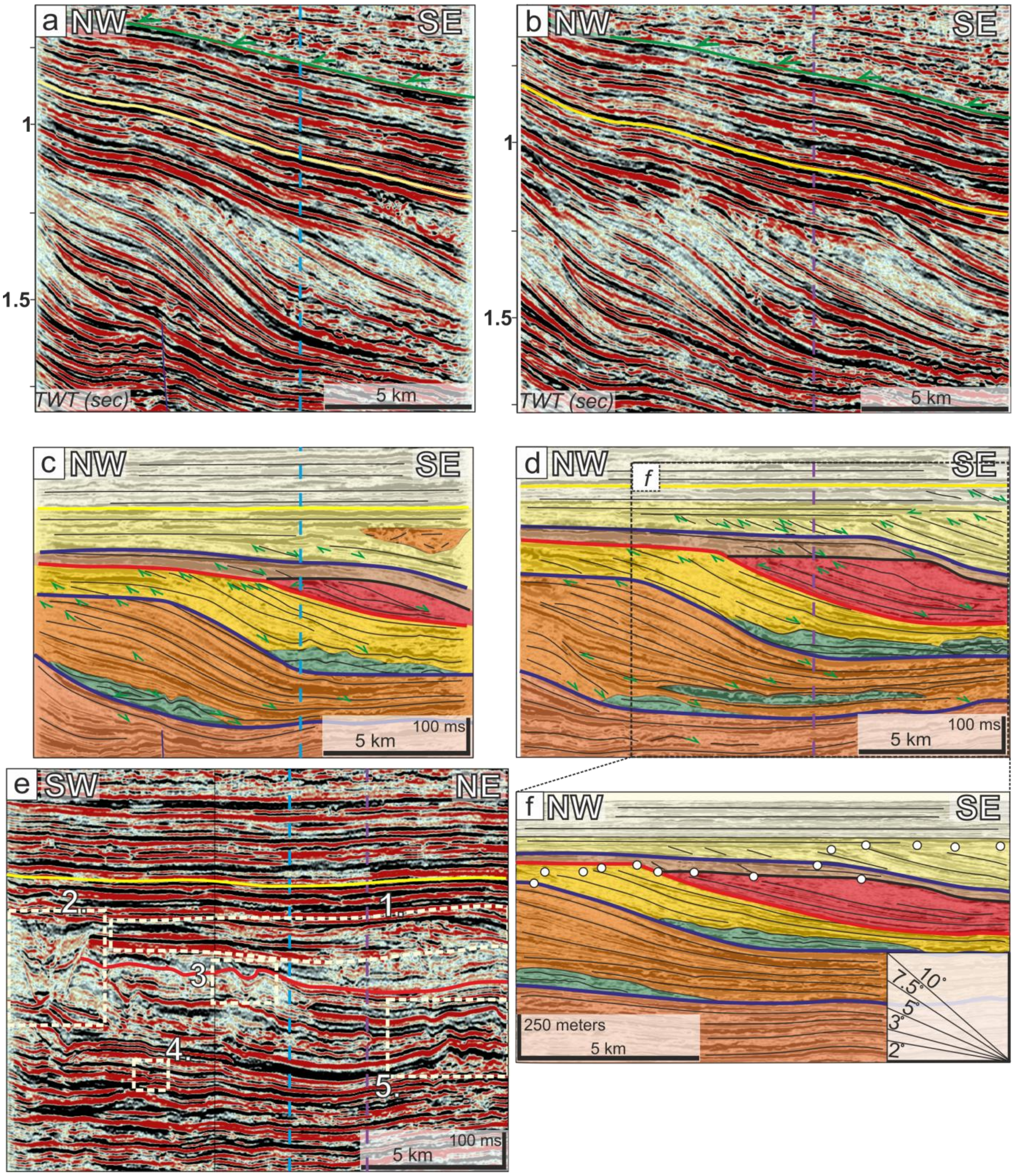


Figure 7. Seismic sections a) and b) oriented parallel with the direction of progradation in the Jászság sub-basin. Interpretation (c, d) shows typical progradational (in orange), aggradational (in yellow), forced regression (red), and retrogradational patterns in the Great Hungarian Plain (location in Figure 1). Note the turbidite complexes at the toe of slopes (in green). Green half-arrows are reflection terminations. Green line is the unconformity between Miocene and

Pliocene sediments, yellow line is the flattening level; c) and d) are the flattened version of the seismic lines above; e) seismic section in the same area perpendicular to the direction of progradation. Vertical dashed lines are intersections with profiles displayed in Figures 7a and 7b. This seismic line shows (1) delta progradation over the shelf area; (2,3) large-scale incisions (~ up to 200 meters deep) near the transition between the shelf and the slope; (4) small turbidite channel within deep water sediments; (5) stacked channel-levee systems (~30-60 m thick); see also Sztanó et al.(2013) and Juhász et al. (2013); f) depth converted version of part of the seismic line in Fig. 7d. Small circles denote the evolution of the shelf margin (i.e., shelf edge trajectories).

4.3 Effects of inherited extensional half-grabens on the paleobathymetry of Lake Pannon

The analysis of the syn-kinematic sedimentation in the diachronous extensional half-grabens is available in previous studies (e.g., Matenco and Radivojevic, 2012, ter Borgh et al., 2015; Balázs et al., 2016). Here we only illustrate the structural history characterizing the evolution of the Pannonian Basin by choosing two specific zones as examples: the Nyírség sub-basin in the north-eastern margin of the Great Hungarian Plain and the Túrkeve sub-basin from the deep central part.

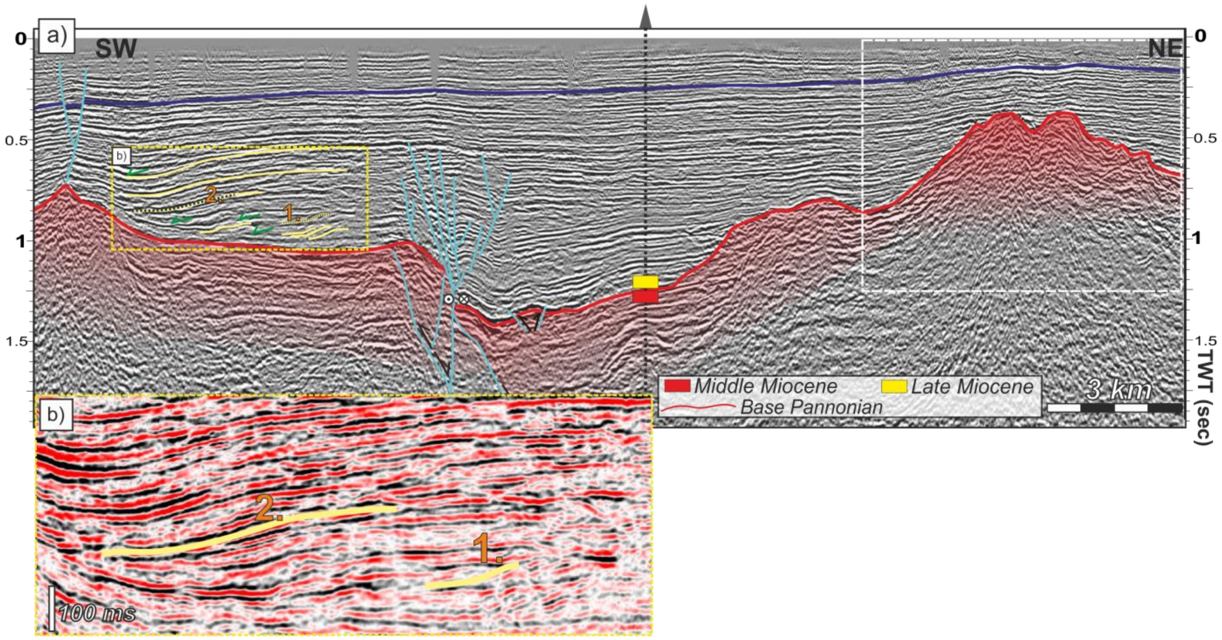


Figure 8. a) Interpreted seismic section located in the NE part of the Pannonian Basin. Location in Figure 1. The interpretation demonstrates a buried late Middle Miocene (Sarmatian) volcano beneath the subsequent Late Miocene (Pannonian) sediments. The gentle anticline geometry of the overlying Pannonian sediments (see arbitrary blue horizon above the volcano highlighted in the white rectangle) is created by differential compaction, see chapter 5 for details. The red line is the Middle- Upper Miocene unconformity. Red area indicates further volcanic edifices and volcano-clastic sediments. Small deltas indicated by numbers are used in Table 1; b) Detail of the same seismic line showing a prograding delta, observed by downlap reflection terminations (green arrows).

The region of the Nyírség sub-basin in the NE part of the Great Hungarian Plain (Fig. 1) contains significant amounts of upper Middle Miocene (Sarmatian) rhyolites, rhyodacites and dacites, domes, lava flows and related tuffs (Pécskay et al., 2006). The analysis of a seismic line in this sub-basin (Fig. 8) shows such a buried Sarmatian volcanic geometry made up by extrusive lava flows, sills, and other intrusive complexes, surrounded by high amplitude reflectors located beneath the base Late Miocene unconformity. The Miocene depocentre is filled with Middle Miocene volcano-clastic sediments, which is a typical feature observed in many other sub-basins located in a similar tectonic position along the Mid-Hungarian Fault Zone (e.g., Horváth et al., 2015). These Middle Miocene syn-kinematic wedges were deposited against the controlling NE-dipping normal faults (Fig. 8). The fault zone was likely reactivated during earliest Late Miocene times with low reverse and strike-slip offsets creating a flower structure (Fig. 8). Available interpretation infers that the Upper Miocene succession in the Nyírség sub-basin is characterized by geometries of deltaic and alluvial environments with no observed deeper shelf-margin slope clinoform geometries. This suggests that the rate of sedimentation has always kept pace with the local subsidence rate and the absence of inherited paleobathymetries. This is in agreement with observations in one well (Fig. 8) penetrating the entire Upper Miocene succession reporting frequent coal intercalations (Székyné et al., 1985). Laterally to the SW in the analysed seismic lines 10s of meters thick deltaic clinoforms prograding over a shallow shelf are observed within the lowermost Pannonian sediments. The height of these clinoforms increases further SW-ward in the direction of progradation where they become the much larger shelf-margin slope clinoforms observed almost everywhere in the basin. In other words, these clinoforms show the older onset of Late Miocene progradation in the basin that started with low amplitude clinoforms and that gradually increase in height with time indicating an increase of the paleo-water depth. The decompacted height of these initial clinoforms (Fig. 8b, Table 1) is in the order of 70 to 150 meters and were deposited between 11.6 - 9 Ma. Their geometries are similar to the transitional slopes interpreted in the western parts of the Pannonian Basin (cf., Sztanó et al., 2015). Note the gentle anticline geometry of the younger Pannonian to Quaternary horizons above the buried volcano. The present-day Tisza river is changing its course significantly around this anticline (Fig. 1).

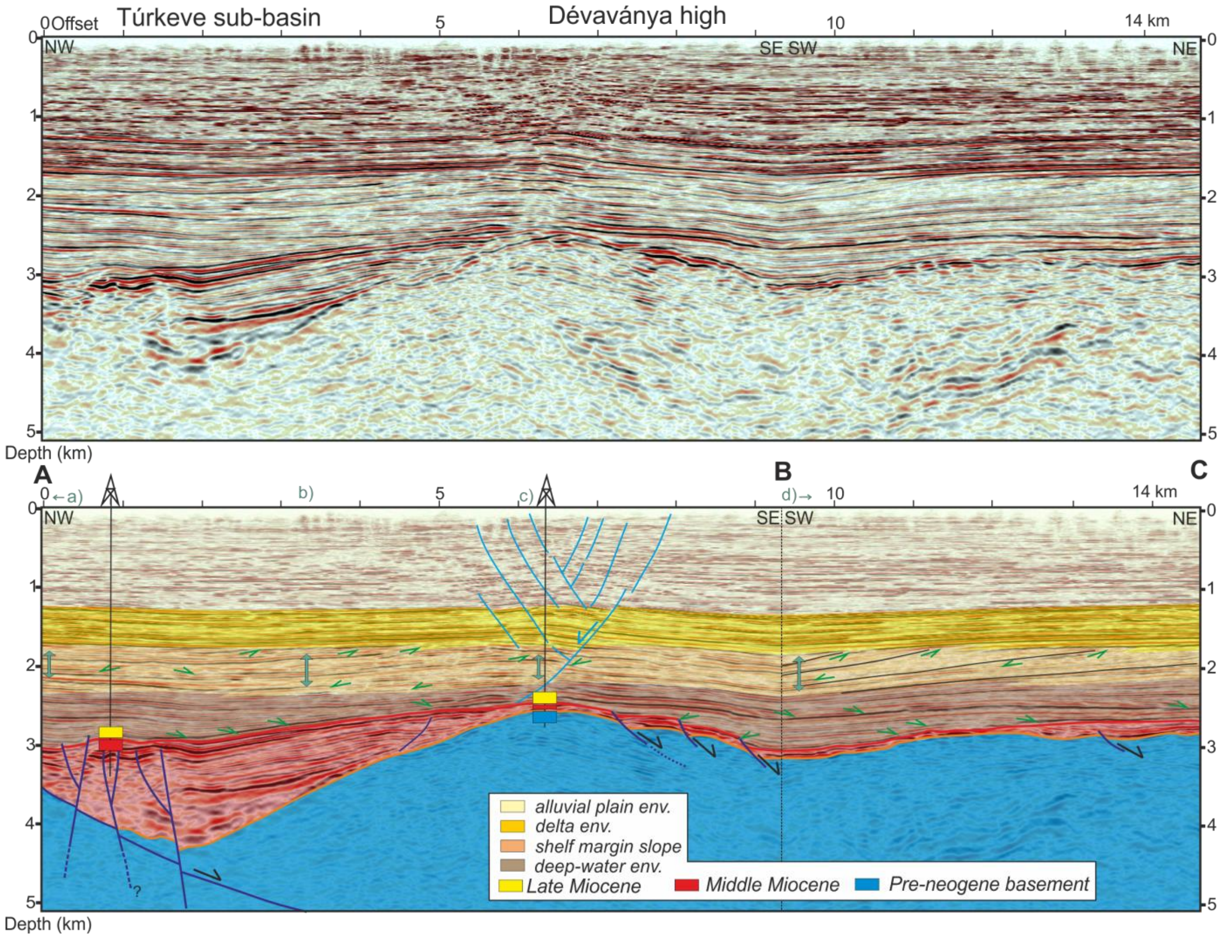


Figure 9. Non-interpreted (up) and interpreted (down) seismic depth-converted section from the central part of the Great Hungarian Plain (location in Figure 1) showing an Early to Middle Miocene sub-basin and typical Late Miocene seismic facies. Note that the trace of the section is a composite between a segment parallel with and one perpendicular to the direction of progradation (BC and AB, respectively). Note the wide segmented fault zone above the basement high that reflects differential compaction in the sub-basins above the basement high (see chapter 5 for details). Green arrows are reflection terminations. The four blue double arrows are locations of paleo-water depth calculations (a-d) within the slope environment.

A depth converted seismic section from the Túrkeve sub-basin (Figs. 1, 9) shows a typical structure for the central part of the Great Hungarian Plain: a half-graben filled with Early to Middle Miocene syn-extensional sediments is overlain by ~3 km of Late Miocene post-kinematic deposits. The half-graben is controlled by a large offset SE-dipping low-angle normal fault, which is accompanied by lower offset normal faults. The half-graben is also slightly inverted by a positive flower structure and shows the typical unconformity also observed elsewhere in the Pannonian Basin at the transition between Middle and Late Miocene (Sarmatian to Pannonian). The structural style is otherwise similar to other such Early – Middle Miocene sub-basins (e.g., Kiskunhalas or Vésztő, see also Balazs et al., 2016). The overlying strata show gentle anticline geometry over the Dévaványa basement high (Fig. 9). The typical progradation of shelf-margin slope clinoforms is observed in the overlying Upper Miocene (Pannonian) sediments. The prograding shelf-margin slope reached this sub-basin by prograding SW-wards at about 5.7 Ma (Magyar et al., 2013). The paleo-water depth was calculated in four points along the same spatially correlated timeline (or reflector, Fig. 9, Table 1). These calculations show a variable bathymetry at the base of the slope ranging from 475m over the Dévaványa basement high to 630 and 740 m in the region overlying the depocentres. Our calculations thus demonstrate that paleo-bathymetries were controlled by the inherited extensional basin geometries, the base of the slope showing higher values over the various sub-basins when compared with intervening basement highs. This means that the deposition of deep-water pelagic sediments and turbidites was unable to compensate the inherited morphological differences from extensional times before the shelf-margin slope progradation arrived to a proximal position.

5. Compaction-induced folds and faults

Seismic sections from different sub-basins of the Pannonian Basin system show the lateral variation of basement depth created by the variable Miocene crustal thinning and subsequent ongoing differential vertical movements (Figs. 5-9). Late Miocene sediments up to ~6 km thick are affected by different amounts of compaction resulting in gentle fold geometries (Fig. 2) or differential compaction induced faults, such as the fault system above the Dévaványa basement high. Compaction induced anticlines are interpreted, for instance, above the buried volcano in the Nyírség sub-basin (Fig. 8) or above the Battonya basement high (Fig. 2).

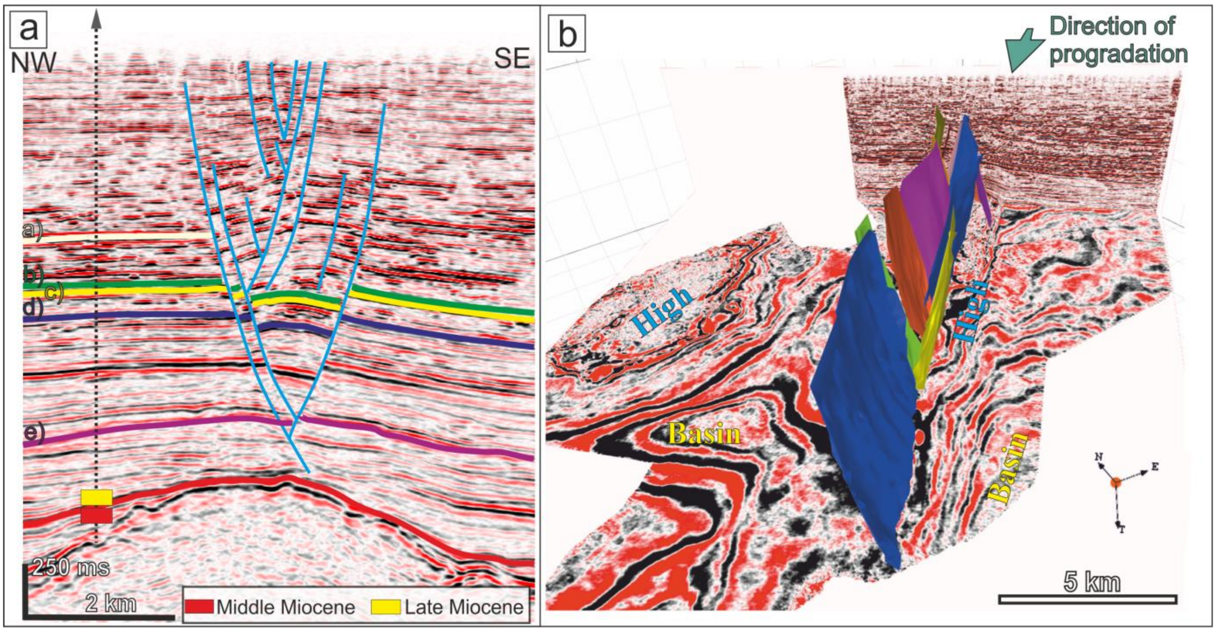


Figure 10. Interpreted seismic data from the central part of the Great Hungarian Plain. Location in Figure 1. a) The segmented Túrkeve Fault zone affecting Late Miocene to Quaternary sediments. Note the increase in offsets upwards in the stratigraphy. The horizons a) to e) were mapped in the seismic cube and are displayed as horizon maps in Figure 11. b) 3D image of the same Túrkeve Fault segments above the basement high, surrounded by deeper basins on either side.

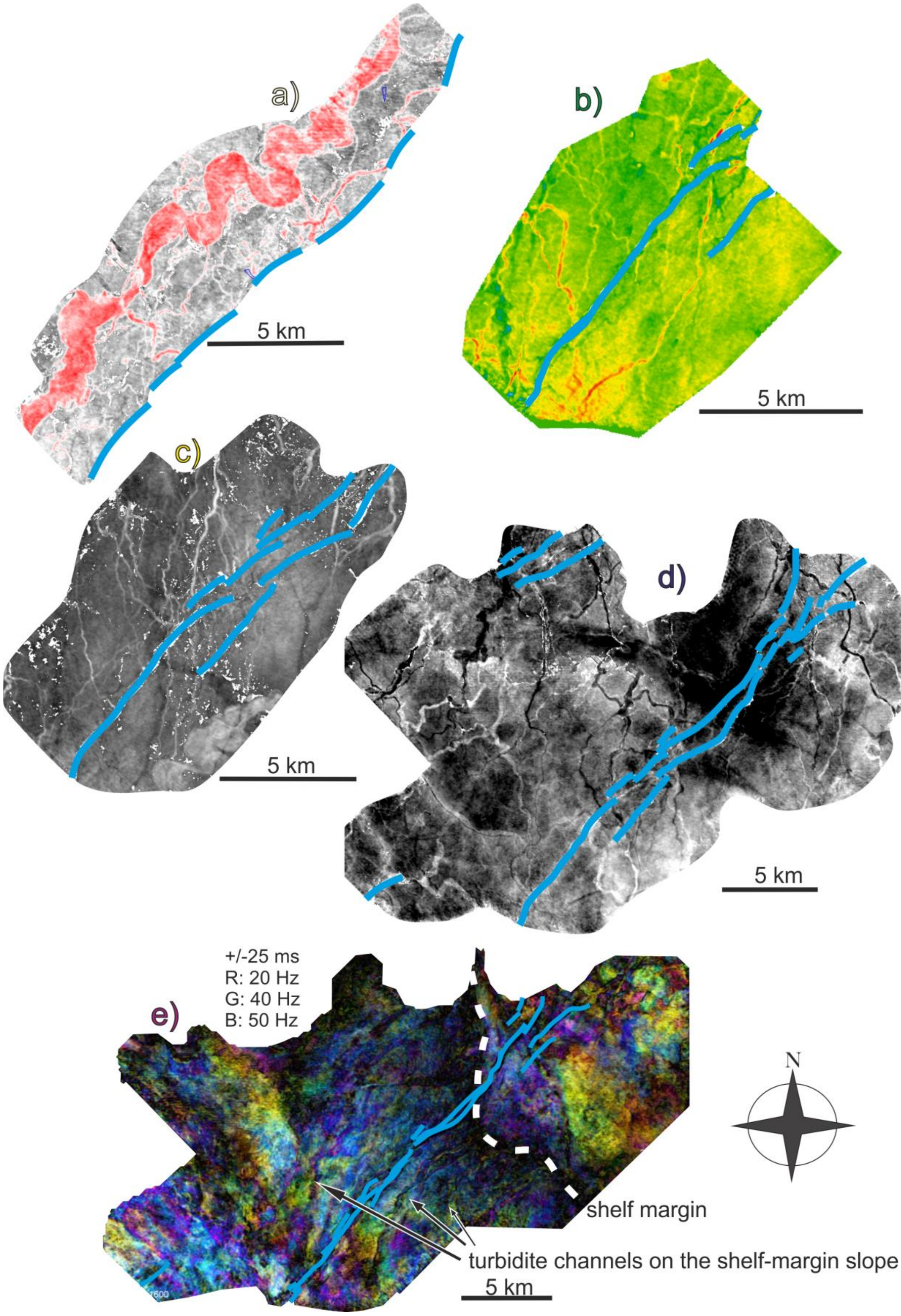


Figure 11. Interpreted seismic horizon maps. Blue lines indicate fault segments cross-cutting horizons. Figures a) to e) correspond to horizons a-e shown in Figure 10a. (a) Amplitude map of a horizon from the alluvial plain, illustrating a meandering river and smaller distributaries.

(b-d) Amplitude maps of horizons from anastomosing delta plain environments showing meandering and anastomosing channels. (e) Spectral decomposition attribute map of a horizon through the shelf, shelf-margin, slope and deep water environments (see text for further explanations). Note that no channel is displaced horizontally by the fault segments and, therefore, these faults do not show strike-slip displacements.

The effects of compaction can be also analysed by the example of the 3-4 km thick sediments overlying the Dévaványa basement high (Figs. 9, 10). The centre of the gentle anticline overlying this basement high (Fig. 9) is cross-cut by a wide normal fault zone truncating the Late Miocene - Quaternary sediments. This fault zone has been previously interpreted either as normal growth fault (e.g., Grow et al., 1994), or a wide strike slip fault zone that is similar to other negative flower structures commonly interpreted in 2D seismic lines in many other areas of the Pannonian Basin (Horváth et al., 2006; Bada et al., 2007). Interestingly, the detailed analysis in this Túrkeve area shows that fault offsets gradually increase upwards from the basement high and furthermore decrease in the uppermost part of the section.

The mechanism of formation of this system of normal faults with variable offsets observed above the Dévaványa basement high can be studied in more details on a 3D seismic cube, where individual fault segments and marker horizons cross-cut by faults were mapped (Figs. 10, 11). These faults truncate and offset Pannonian post-rift sediments. The fault with the largest offset dips SE-wards in the southern part of the studied area and changes to a NW-ward dip in the north, where the fault zone is wider (Fig. 10). The maximum throw of the fault is reached within the delta sediments of the Újfalu Formation by ~100 meters. The offset analysis in the 3D cube confirms the observation of the 2D seismic lines of a gradual increase of offset upwards from the oldest Late Miocene deep-water sediments and furthermore a decrease in the uppermost part of the section (Fig. 10). This pattern is a typical attribute of faults related to salt movement and/or differential compaction effects (e.g., Magara, 1978; Williams, 1987; Xu et al., 2015). The absence of salt bodies in our seismic observations and previous studies infers differential compaction effects. A much clearer discrimination from strike-slip deformation is provided by the analysis of horizontal offsets. We have calculated a large number of attribute maps that all show excellent expressions of the faults system and the sedimentology of variable fluvial-alluvial to deltaic environments, from meandering rivers (Fig. 11a) to turbiditic channels on slopes (Fig. 11e). Most of the larger channels are oriented parallel with the normal faults, which is also the strike of the neighbouring older extensional basins and the strike of the basement high. However, smaller channels often cross the various branches of the normal fault system but none of these sedimentary channels indicate any horizontal offset when crossing the various fault branches. As a consequence, the strike-slip kinematics of this zone can be ruled out. We conclude that differential compaction is the primary mechanism creating such structures. This interpretation is also supported by the lack inversion of the underlying basement structure. Initiation of similar extensional faults otherwise can be also associated with the inversion of the underlying basement structures.

6. Discussion

6.1 Controls on water depth variations and progressive infill of Lake Pannon

The main observed mechanism of Late Miocene – Early Pliocene basin infill is the shelf-margin slope progradation. Because the processes controlling the balance between the accommodation space and sediment supply in Lake Pannon have similar orders of amplitude, they create a local fine interplay with aggradational and progradational geometries superimposed on the overall prograding pattern (e.g., Juhász et al., 2007; Sztanó et al., 2013). The controlling factors are coeval thermal subsidence, climatic variations, massive sourcing of sediments from the paleo-Danube and paleo-Tisza rivers, inherited extensional morphology determining bathymetrics and eventual connection at the separating gateways with other basins (e.g., Leever et al., 2010). The relative importance of these forcing factors varied spatially through time.

The widespread erosional unconformity at the base of the Late Miocene sediments and the very thin or absent late Middle Miocene (Sarmatian) succession in the centre of the Pannonian Basin (Magyar et al., 1999) suggests an overall shallow water depth or even subaerial environment during the onset of Late Miocene (Pannonian) times. Exceptions are recorded in the deepest Middle Miocene (half-)grabens, such as the Békés basin, and areas near the margins of the Pannonian Basin, like the Danube basin, where Middle Miocene subsidence outpaced sediment supply and, therefore, deep water environments could have continued. This asymmetry of shallow in the centre and deep bathymetry near the margins of the Pannonian Basin was created by the overall variability of the extensional dynamics (Balázs et al., 2016; 2017). The base Pannonian unconformity probably also marks the final disconnection of the Lake from the remnant of the Paratethys (Magyar et al., 1999), although the later (Messinian) connectivity of the Pannonian and Dacian and then the Black Sea basins is still under debate (c.f., Magyar and Sztanó, 2008; Leever et al., 2010; Csató et al., 2015; Matenco et al., 2016).

After a short break in extension during earliest Pannonian times, rapid subsidence continued in the Pannonian Basin (Horváth et al., 2015; Balázs et al., 2016). This subsidence was locally enhanced by the formation of other Late Miocene half-grabens, mostly concentrated in the E and SE parts of the Pannonian Basin until about 9-8 Ma. Subsidence has created a rapid transgression associated with the deposition of a deep-water facies recorded in most of the Pannonian and Transylvanian basins (e.g., Krézsek et al., 2010). These processes have resulted in highly variable paleo-bathymetries during the evolution of Lake Pannon, as reflected by our calculated heights of the subsequent shelf-margin slope progradation (Fig. 6). In the NW

Danube basin, the water depth increased to a minimum of 550 m and the basin was subsequently filled by 9 Ma with ~1.5 km thick deep water sediments. In the NE (e.g., the Nyírség sub-basin) the subsidence and water level rise kept pace with sedimentation, resulting in a small paleo-bathymetrical variability of consistently shallow water prograding – aggrading – retrograding delta and alluvial environments during the entire Late Miocene - Quaternary basin evolution with only a few localized exceptions (Fig. 8). Southwards (near the Nádudvar sub-basin, Fig. 6b), the general progradation was interrupted by a major flooding and retrogradation at ~ 7.5 Ma. In contrast, the rate of tectonic subsidence was lower in the western parts of the Pannonian Basin resulting in a gradual basin fill by aggradation and progradation. In other words the rates of sediment supply and creation of accommodation space were roughly in balance there. In the centre of the Pannonian Basin (the Danube-Tisza interfluve, Fig. 1) the subsidence rates were low during the entire evolution and represented a basement high, therefore, the paleo-bathymetry has never reached the few hundreds of meters observed elsewhere (Fig. 6).

Because subsidence rates and consequently creation of accommodation continuously decreased with time after the initial Late Miocene transgression, while sediment input remained high or even increased, therefore, the entire Lake Pannon was finally filled by ~4 Ma (Magyar et al., 2013). Smaller scale water level variations are observed by the analysis of the shelf-edge trajectories. Such an isolated lacustrine system is more sensitive to regional climate and therefore lake level variations are interpreted to be climatically driven (Uhrin and Sztanó, 2011; Sztanó et al. 2013) or could be controlled by local subsidence and uplift pulses associated with the late stage inversion of the Pannonian Basin. Our reconstructed paleobathymetries between 6.8 Ma and 5 Ma show that the highest water depth values of the lake reached and most probably exceeded values of 1000 m (Table 1, see also Balázs et al., 2015). The asymmetry of the transport direction dominant from the NE and NW during the continuous subsidence has created higher paleo-bathymetries in the SE where progradation was recorded at later times (Fig. 6a).

By the same reasoning, these basins contain the largest thicknesses of deep water pelagic sediments and distal turbidites reaching up to 3.5 km (e.g., Sztanó et al., 2013).

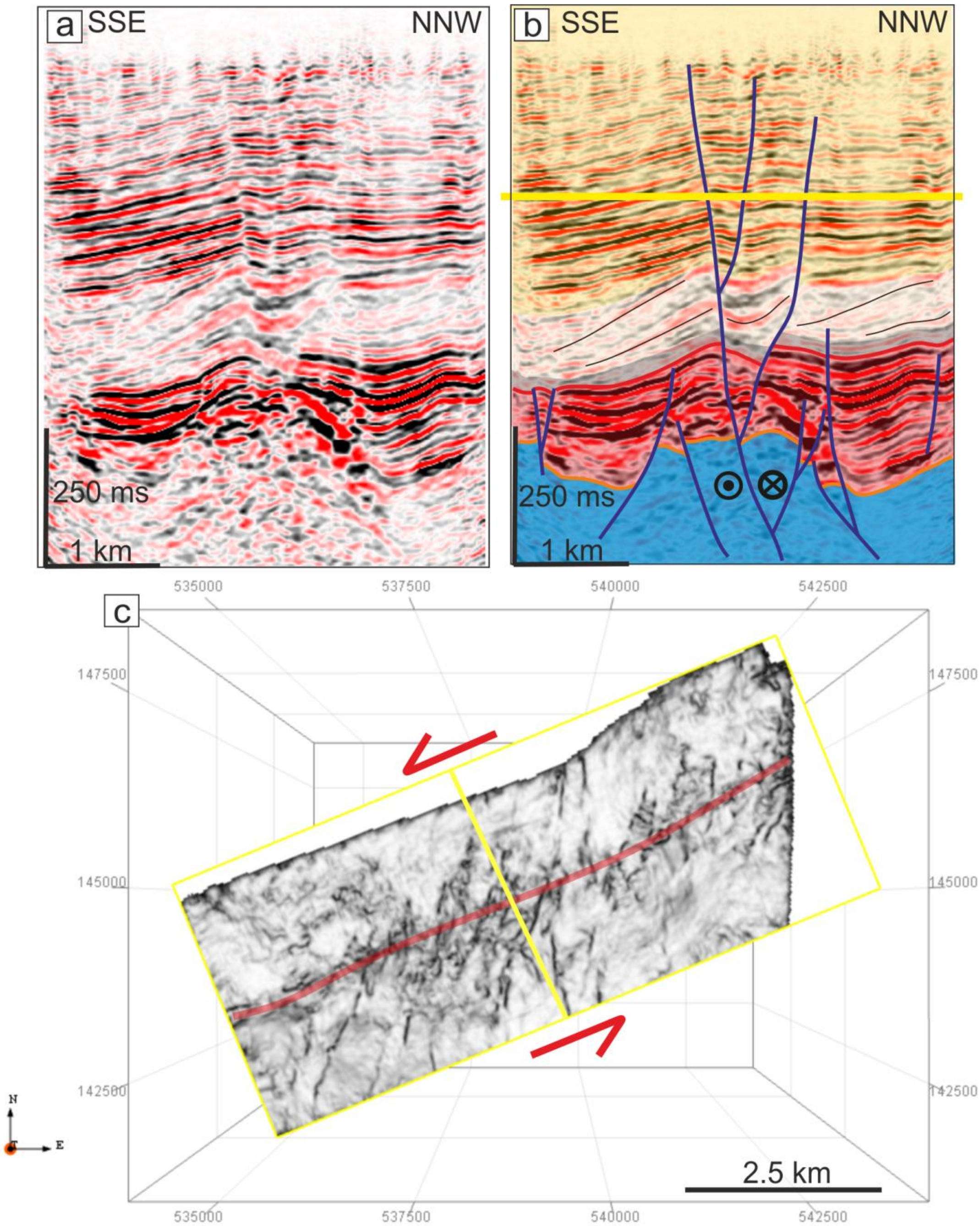
6.2 Shelf-margin morphology and basin evolution

Miocene-Pliocene sediments deposited on the slope connecting the shelf with the deep-water basin of Lake Pannon are presently deeply buried in the Pannonian Basin. Our analysis shows that the width of the slope between the shelf-edge to the toe-slope varies between 5 and 15 km at decompacted heights between 200 and 1000 m. This results in slope angles between 3^{o} and 8^{o}. Such values are similar to dip angles of marine slopes (Porebski and Steel, 2003; Johanessen and Steel 2009; Gong et al., 2016) that are controlled by lithology, grain size distribution or sediment influx from the source area (e.g, Gvirtzman et al., 2014).

Our calculations demonstrate that paleo-bathymetries were controlled by the inherited extensional geometries, with higher values (600-700 m at the base of the slope) over the various sub-basins than over the intervening basement highs (400-500 m). This means that the deposition of deep-water pelagic sediments and turbidites was unable to compensate all the inherited morphological differences from extensional times before the shelf-margin slope progradation arrived (see also Törő et al., 2012).

Our analysis of the shelf sedimentation (Fig. 7) shows progradation of tens of meters thick deltas (Uhrin and Sztanó, 2011). Their position on the inner or outer shelf is controlled by lake water level variations that typically reach ~100 m during highstands, as observed in marine domains or semi-enclosed seas, such as the Mediterranean (Rabineau et al., 2006) or the Black Sea (Porebski and Steel, 2003; Matenco et al., 2016). Our interpretation of water-level variations infer periods of ascending, descending and stationary shelf-edge trajectories (Fig. 7). Such an analysis does not necessarily take into account the small-scale variations of accommodation on the shelf (cf., Sztanó et al., 2013), but in basins characterized by ongoing tectonic subsidence, such as the Miocene Pannonian Basin, even stationary shelf-edge trajectory indicates periods of climatically-driven water-level fall. Their amplitudes are similar to the rate of basin subsidence. However, in our case their local amplitude is only in the order of tens of meters usually. In contrast with typical passive margin settings, back-arc extension has resulted in highly variable basement morphology, such as deep half-grabens, like for instance the Makó Trough (Fig. 2) or basement highs, like the Transdanubian Range (Fig. 1). These structures also control locally the direction of sediment transport, such as in the Túrkeve sub-area, where the direction of progradation followed the strike of the inherited Middle Miocene sub-basin (Fig., 1, 9) such as in the Sava Trough.

The inherited relief, spatially variable subsidence rates and lake water level variations controlled the paleo-bathymetries and created tens of metres high deltaic clinoforms over the shelf and up to 1000 meters high shelf-margin slope clinoforms (c.f., Leroux et al., 2014; Rabineau et al., 2014). Of course, between such end members the balance between the rate of sedimentation and progressively increasing base-level rise could result in the continuous transition from small scale deltas to high shelf-margin slopes (cf. Sztanó et al., 2015). Such transitional slopes are observed in the Nyírség sub-basin (Fig. 8) and its prolongation towards the deep Derecske Trough (Balázs et al., 2016), or in the Danube-Tisza interfluve. Water depths are in general higher above the former half (grabens) and lower above the separating basement highs.



574 **Figure 12.** General geometry of a strike-slip fault zone. Non-interpreted (a) and interpreted (b) seismic section crossing the Balaton Fault zone, location in Figure 1; c) Coherency cube time slice highlighting the geometry of synthetic Riedel faults and demonstrating the sinistral strike-slip offset of this fault zone (see also Várkonyi et al., 2013 and Visnovitz et al., 2015).

The syn-rift extension and half-graben formation in the Pannonian Basin ultimately ceased at 8-9 Ma (e.g., Matenco and Radivojevic, 2012; Balázs et al., 2017). The subsequent evolution was controlled by post-rift thermal cooling and the basin-wide inversion during the Adriatic indentation creating differential vertical movements (Fig. 1; Sacchi et al., 1999; Bada et al., 2007). Such inverted structures are well documented from earliest late Miocene times in the western part of the Pannonian Basin (Fodor et al., 2005; Uhrin et al., 2009), at 6-8 Ma along the Mid-Hungarian Fault zone (Fig. 6d, see also Juhász et al., 2013). Our results show that effects of basin inversion should be taken into account significantly during the calculation of the paleo-bathymetries in the entire basin. The observed contraction reached a peak at the transition between Miocene and Pliocene times, caused likely by the northward drift and CCW rotation of the Adriatic microplate (Pinter et al., 2005). This peak contraction is the main mechanism creating the widespread unconformity observed at the transition between the Miocene and Pliocene in the Pannonian Basin (e.g., Fig. 7), being replaced laterally by a correlative conformity in deeper sub-basins (Magyar and Sztanó, 2008). Our observations confirm that the Late Miocene to Recent evolution of the Pannonian Basin and associated subsidence/uplift pattern is mainly controlled by basin scale flexural effects superimposed on post-rift thermal sagging (Horváth and Cloetingh, 1996; Dombrádi et al., 2010; Jarosinski et al., 2011).

Previous interpretations assumed that the inversion was also associated with the (re)activation of strike slip zones along former structures (Fig. 1; Horváth et al., 2006, Bada et al., 2007; Visnovitz et al., 2015). Strike-slip kinematics are certainly significant in many parts of the Pannonian Basin, demonstrated by the observation of offsets and Riedel shears in 2D or 3D seismics (e.g., Fig. 12, see also Várkonyi et al., 2013). However, our study demonstrates for the first time that compaction effects creating fault systems such as the one quantified above the Dévaványa basement high are certainly significant in the sediments of the Pannonian Basin. The effects should be similar elsewhere: faults with variable offsets, increasing and subsequently decreasing towards the surface, reaching a maximum in the order of 150 m (Figs. 9, 13).

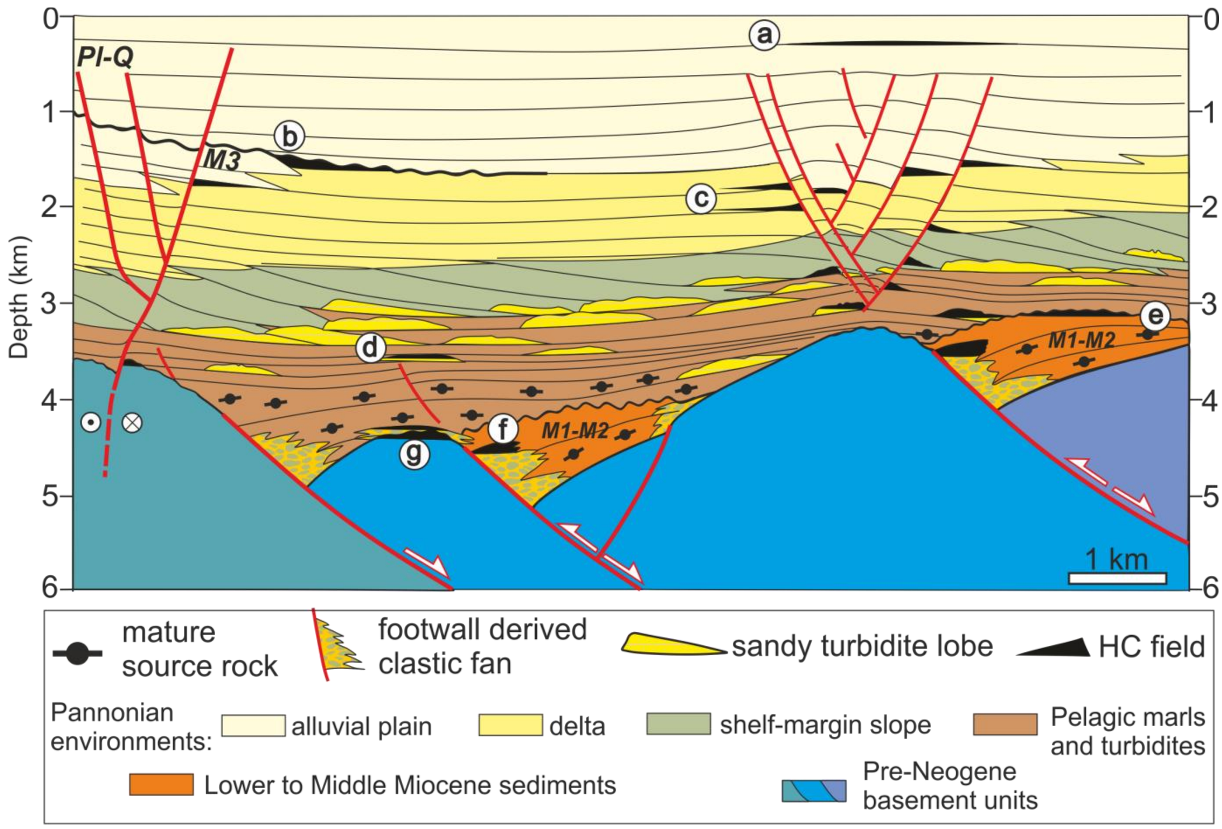


Figure 13. Conceptual model of the Neogene basin fill in the Great Hungarian Plain that takes into account the morphology of the bed of Lake Pannon, including inherited extensional structures and further effects, such as differential compaction over basement highs and neotectonic strike-slip fault zones (modified after Tari and Horváth, 2006). Figure also shows the main hydrocarbon play types of the basin: a) Biogenic and thermogenic fields in drape-folds above basement highs; b) stratigraphic traps in delta sandstones or connected to unconformities; c) delta or deep-water turbiditic sandstones affected by compaction induced normal faults; d) stratigraphic or structural traps in deep-water turbiditic sandstones. e) Sandstones and conglomerates at the unconformity of inverted Early-, Middle- Miocene basins; f) footwall-derived fans along the boundary faults of half-grabens; g) basal conglomerates deposited onto

the basement or the fractured Neogene-basement itself. Differential compaction induced faults at the periphery of half grabens may provide migration pathways.

7. Conclusions

Our interpretation of 2D and 3D seismic data correlated with well logs from the Late-Miocene to Quaternary sedimentary succession filling the Pannonian Basin shows a transition from an initially underfilled to a finally overfilled large lacustrine basin. Spatial and temporal variations of the external and internal forcing factors resulted in lateral changes of prograding – aggrading – retrograding shelf-margin slope geometries and paleo-water depths. Using decompacted thicknesses of the prograding shelf-margin slope clinoforms, our calculations indicate the variation of water depth values from ~75 m up to ~1 km. Highest water depth values characterized the SE part of the basin as a consequence of higher subsidence rates and more distal position from the source areas. The shelf had paleo-bathymetries of up to 75 m with a high order variability controlled by climate. Both water depth and sedimentary transport routes were primarily determined by inherited and/or active local tectonics; they controlled Late Miocene shelf-margin progradation directions as well as Recent fluvial transport routes.

Latest Miocene to Recent tectonic topography appears to be basin scale folding process. Areas of uplift were subject to denudation and the eroded material continuously overfilled the generated accommodation space. Sediments up to ~6 km have been affected by this still ongoing differential vertical movement and compaction creating gentle fold geometries and differential compaction induced fault offsets, playing a major role in hydrocarbon migration and trapping (Fig. 13). Geometries of such non-tectonic faults above basement highs can be clearly distinguished from extensional or strike-slip fault geometries by the calculation and analysis of 3D seismic attributes.

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References

Angevine, C.L., Heller, P.L., Paola, C., 1990. Quantitative sedimentary basin modelling. AAPG Course Note Series 32, 133 pp.

Bada, G., Horváth, F., Dövényi, P., Szafián, P., Windhoffer, G., Cloetingh, S., 2007. Present-day stress field and tectonic inversion in the Pannonian basin. Global and Planetary Change 58, 165-180.

Balázs, A., Tőkés, L., Magyar, I., 2015. 3D analysis of compaction related tectonic and stratigraphic features of the Late Miocene succession from the Pannonian Basin. Abstract Book of 31st IAS Meeting of Sedimentology 1–52.

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

Balázs, A., Burov, E., Matenco, L., Vogt, K., Francois, T., Cloetingh, S., 2017. Symmetry during the syn- and post-rift evolution of extensional back-arc basins: the role of inherited orogenic structures. Earth and Planetary Science Letters 462, 86-98.

Báldi, T., 1986. Mid-Tertiary stratigraphy and Paleogeographic evolution of Hungary. Budapest, Akadémiai Kiadó 201.

Bérczi, I., Phillips, R.L., 1985. Process and depositional environments within Neogene deltaic–lacustrine sediments, Pannonian Basin, Southeast Hungary. Geophysical Transactions 31, 55–74.

Carroll, A. R., Bohacs, K. M., 1999. Stratigraphic classification of ancient lakes: balancing tectonic and climatic controls. Geology 27, 99-102.

Cartwright, J., Huuse, M., 2005. 3D seismic technology: the geological ’Hubble’. Basin research 17, 1-20.

Chopra, S., Marfurt, K. J., 2005. Seismic attributes. – A historical perspective. Geophysics 70, 3.

Cloetingh, S., Van Wees, J. D., Ziegler, P., Lenkey, L., Beekman, F., Tesauro, M., Förster, A., Norden, B., Kaban, M., Hardebol, N., Bonté, D., Genter, A., Guillou-Frottier, L., Voorde, M. T., Sokoutis, D., Willingshofer, E., Cornu, T., and Worum, G., 2010. Lithosphere tectonics and thermo-mechanical properties: An integrated modelling approach for Enhanced Geothermal Systems exploration in Europe. Earth-Sci. Rev. 102, 159–206, 2010.

Csató, I., Tóth, S., Catuneanu, O., Granjeon, D., 2015. A sequence stratigraphic model for the Upper Miocene–Pliocene basin fill of the Pannonian Basin, eastern Hungary. Marine and Petroleum Geology 66, 117-134. doi: 10.1016/j.marpetgeo.2015.02.010

Csontos, L., Nagymarosy, A., 1998. The Mid-Hungarian line: a zone of repeated tectonic inversions. Tectonophysics 29, 51-71.

De Leeuw, A., Mandic, O., Krijgsman, W., Kuiper, K., Hrvatović, H., 2012. Paleomagnetic and geochronologic constraints on the geodynamic evolution of the Central Dinarides. Tectonophysics 530—531, 286-298.

Dombrádi, E., Sokoutis, D., Bada, G., Cloetingh, S., Horváth, F., 2010. Modelling recent deformation of the Pannonian lithosphere: Lithospheric folding and tectonic topography. Tectonophysics 484, 103-118.

Dövényi, P., 1994. Geofizikai vizsgálatok a Pannon-medence litoszférafejlődésének megértéséhez. MTA thesis, Hungarian Academy of Sciences, Budapest, 127 p. (in Hungarian).

Faccenna, C., Becker, T. W., Auer, L., Billi, A., Boschi, L., Brun, J. P., Capitanio, F. A., Funiciello, F., Horvàth, F., Jolivet, L., 2014. Mantle dynamics in the Mediterranean. Rev. Geophys. 52, 283–332, doi:10.1002/2013RG000444.

Fodor, L., Bada, G., Csillag, G., Horvath, E., Ruszkiczay-Rudiger, Z., Palotas, K., Sikhegyi, F., Timar, G., Cloetingh, S., 2005. An outline of neotectonic structures and morphotectonics of the western and central Pannonian Basin. Tectonophysics 410, 15-41.

Fongngern, R., Olariu, C., Steel, R. J., Krézsek, Cs., 2015. Clinoform growth in a Miocene, Para-tethyan deep lake basin: Thin topsets, irregular foresets and thick bottomsets. Basin Research 28, 770–795.

Garcia-Castellanos, D., Verges, J., Gaspar-Escribano, J., Cloetingh, S., 2003. Interplay between tectonics, climate, and fluvial transport during the Cenozoic evolution of the Ebro Basin (NE Iberia). J. Geophys. Res. 108, 2347.

Gong, C., Steel, R., Wang, Y., Lin, C., Olariu, C., 2016. Shelf-margin architecture variability and its role in sediment-budget partitioning into deep-water areas. Earth-Science Reviews 154, 72-101.

Grow, J.A., Mattick, R.E., Bérczi-Makk, A., Péró, Cs., Hajdú, D., Pogácsás, Gy., Várnai, P., Varga, E., 1994. Structure of the Békés Basin Inferred from Seismic Reflection, Well and Gravity Data. In: Teleki, et al. (Eds.), Basin Analysis in PetroleumExploration. Kluwer Academic Publishers, Dordecht, 1–37.

Gvirtzman, Z., Csató, I., Granjeon, D., 2014. Constraining sediment transport to deep marine basins through submarine channels: the Levant margin in the Late Cenozoic. Mar. Geol. 347, 12–26.

Helland-Hansen, W., Hampson, G.J., 2009. Trajectory analysis: concepts and applications. Basin Research 21, 454–483.

Henriksen, S., Helland-Hansen, W., Bullimore, S., 2011. Relationship between shelfedge trajectories and sediment dispersal along depositional dip and strike: a different approach to sequence stratigraphy. Basin Research 23, 3–21.

Horváth, F., Bada, G., Windhoffer, G., Csontos, L., Dombrádi, E., Dövényi, P., Fodor, L., Grenerczy, Gy, Síkhegyi, F., Szafián, P., Székely, B., Timár, G., Tóth, L., Tóth, T., 2006. A Pannon-medence jelenkori geodinamikájának atlasza: Euro-konform térképsorozat és magyarázó. Magyar Geofizika 47, 133–137 (in Hungarian, with English abstract).

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

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

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

Johannessen, E.P., Steel, R.J., 2005. Shelf-margin clinoforms and prediction of deepwater sands. Basin Research 17, 521–550.

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

Juhász, Gy, Pogácsás, Gy, Magyar, I., Vakarcs, G., 2007. Tectonic versus climatic control on the evolution of fluvio–deltaic systems in a lake basin, Eastern Pannonian Basin. Sedimentary Geology 202, 72–95.

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

Katz, B.J., 1990. Lacustrine basin exploration – Case Studies and Modern Analogues. American Association of Petroleum Geologists Memoir 50, 340 pp. AAPG, Tulsa.

Krézsek, C., Filipescu, S., Silye, L., Matenco, L., Doust, H., 2010. Miocene facies associations and sedimentary evolution of the Southern Transylvanian Basin (Romania): Implications for hydrocarbon exploration. Marine and Petroleum Geology 27, 191-214.

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

Leever, K.A., Matenco, L., Rabagia, T., Cloetingh, S., Krijgsman, W., Stoica, M., 2010. Messinian sea level fall in the Dacic Basin (Eastern Paratethys): palaeogeographical implications from seismic sequence stratigraphy. Terra Nova 22, 12–17.

Leever, K.A., Matenco, L., Garcia-Castellanos, D., Cloetingh, S.A.P.L., 2011. The evolution of the Danube gateway between Central and Eastern Paratethys (SE Europe): insight from numerical modelling of the causes and effects of connectivity between basins and its expression in the sedimentary record. Tectonophysics 502, 175–195.

Leroux, E., Rabineau, M., Aslanian, D., Granjeon, D., Droz, L., Gorini, C., 2014. Stratigraphic simulations of the shelf of the Gulf of Lions: testing subsidence rates and sea-level curves during the Pliocene and Quaternary. Terra Nova 26, 230-238.

Magara, K., 1978. Compaction and fluid migration. Elsevier, 349 pp. Amsterdam/New York.

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

Magyar, I., Fogarasi, A., Vakarcs, G., Bukó, L., Tari, G.C., 2006. The largest hydrocarbon field discovered to date in Hungary: Algyő. In: Golonka, J., Picha, F.J. (Eds.), The Carpathians and their foreland: geology and hydrocarbon resources. AAPG Memoir 84, pp. 619–632. AAPG, Tulsa.

Magyar, I., Geary, D. H., Müller, P., 1999. Paleogeographic evolution of the Late Miocene Lake Pannon in Central Europe. Palaeogeogr. Palaeoclimatol. Palaeoecol. 147, 151-167.

Magyar, I., Sztanó, O., 2008. Is there a Messinian unconformity in the Central Paratethys? Stratigraphy 5, 245-255.

Martins-Neto, M.A., Catuneanu, O., 2010. Rift Sequence Stratigraphy. Marine and Petroleum Geology 27, 247-253.

Márton, E., Fodor, L., 2003. Tertiary paleomagnetic results and structural analysis from the Transdanubian Range (Hungary): rotational disintegration of the Alcapa unit. Tectonophysics 363, 201-224.

Matenco, L., Munteanu, I., ter Borgh, M., Stanica, A., Tilita, M., Lericolais, G., Dinu, C., Oaie, G., 2016. The interplay between tectonics, sediment dynamics and gateways evolution in the Danube system from the Pannonian Basin to the western Black Sea. Science of The Total Environment 543, 807-827. doi: 10.1016/j.scitotenv.2015.10.081

Matenco, L., Radivojević, D., 2012. On the formation and evolution of the Pannonian Basin: Constraints derived from the structure of the junction area between the Carpathians and Dinarides. Tectonics 31(6), TC6007.

Mészáros, F. Zilahi-Sebess, L., 2001. Compaction of sediments with great thickness in the Pannonian Basin. Geophysical Transactions 44, 21-48.

Nagymarosy, A., Hámor, G., 2012. Genesis and evolution of the Pannonian Basin, in Geology of Hungary, Regional Geology Reviews, edited by J. Haas, pp. 149-200, Springer.

Nagymarosy, A., Müller, P., 1988. Some aspects of the Neogene biostratigraphy in the Pannonian Basin. in: The Pannonian Basin, a study in basin evolution, edited by L. H. Royden and F. Horvath, AAPG Memoir 45, pp. 58-68. AAPG, Tulsa.

Partyka, G., Gridley, J., Lopez, J., 1999. Interpretational applications of spectral decomposition in reservoir characterization. The Leading Edge 18, 353–360.

Pécskay, Z., Lexa, J., Szakács, A., Seghedi, I., Balogh, K., Konecny, V., Zelenka, T., Kovacs, M., Póka, T., Fülöp, A., Márton, E., Panatiotu, C., Cvetkovic, V., 2006. Geochronology of Neogene magmatism in the Carpathian arc and intra-Carpathian area. Geol. Carpathica 57, 511–530.

Pinter, N., Grenerczy, Gy., Weber, J., Stein, S., Medak D., (Eds.), 2005. The Adria Microplate: GPS Geodesy, Tectonics and Hazards (Nato Science Series: IV: Earth and Environmental Sciences), 413 pp., Springer.

Pogácsás, Gy., Lakatos, L., Révész, I., Ujszászi, K., Vakarcs, G., Várkonyi, L., Várnai, P., 1988. Seismic facies, electro facies and Neogene sequence chronology of the Pannonian Basin. Acta Geologica Hungarica 31, 175–207.

Porebski, S.J., Steel, R.J., 2003. Shelf-margin deltas: their stratigraphic significance and relation to deepwater sands. Earth-Science Reviews 62, 283–326.

Posamentier, H.W., Walker, R.G., 2006. Deep water turbidites and turbiditic fans, in: Posamentier, H.W., Walker, R.G. (Eds.), Facies models revisited. Society for Sedimentary Geology pp. 339-527.

Rabineau, M., Berné, S., Olivet, J.-L., Aslanian, D., Guillocheau, F., Joseph, P., 2006. Paleo sea levels reconsidered from direct observation of paleoshoreline position during Glacial Maxima (for the last 500,000 yr). Earth Planet. Sci. Lett. 252, 119–137.

Rabineau, M., Leroux, E., Aslanian, D., Bache, F., Gorini, C., Moulin, M., Molliex, S., Droz, L., Reis, A.D., Rubino, J.-L., Guillocheau, F., Olivet, J.-L., 2014. Quantifying subsidence and isostatic readjustment using sedimentary pale-omarkers, example from the Gulf of Lions. Earth Planet. Sci.Lett. 388, 1–14.

Rögl, F., Daxner-Höck, G., 1996. Late Miocene Paratethys Correlations. In: Bernor R.L., Fahlbusch, V., Mittmann, H-W. (eds.), The evolution of western Eurasian Neogene mammal faunas. Columbia University Press, New York, 47-55

Royden, L. H., Horváth, F. (eds) 1988. The Pannonian Basin: a Case Study in Basin Evolution. American Association of Petroleum Geologists Memoir 45. 394 p. AAPG, Tulsa.

Sacchi, M., Horváth, F., Magyari, O., 1999. Role of unconformity-bounded units in the stratigraphy of the continental record: A case study from the late Miocene of the western Pannonian Basin, Hungary, in The Mediterranean Basins: Extension Within the Alpine Orogen, edited by B. Durand et al., Geol. Soc. Spec. Publ., 156, 357–390.

Saftic, B., Velic, J., Sztanó, O., Juhász, Gy., Ivkovic, Z., 2003. Tertiary subsurface facies, source rocks and hydrocarbon reservoirs in the SW part of the Pannonian Basin (Northern Croatia and South-Western Hungary). Geologica Croatica 56, 101–122.

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

Steckler, M.S., Mountain, G.S., Miller, K.G., Christie-Blick, N., 1999. Reconstruction of Tertiary progradation and clinoform development on the New Jersey passive margin by 2-D backstripping. Marine Geology 154, 399-420.

Szalay, Á., 1982. A rekonstrukciós szemléletű földtani kutatás lehetőségei a szénhidrogénperspektívák előrejelzésében. Kandidátusi dolgozat, Hungarian Academy of Sciences, Budapest, 146 p. (in Hungarian).

Szalay, Á., Szentgyörgyi, K., 1988. A Method for Lithogenetic Subdivision of Pannonian (s. l.) Sedimentary Rocks. in: The Pannonian Basin, a study in basin evolution, edited by L. H. Royden and F. Horváth, pp. 89-96. AAPG Mem., AAPG, Tulsa.

Székyné Fux, V., Pap, S., Barta, I., 1985. A nyírségi Nagyecsed-I. és Komoró-I. fúrások földtani eredményei. Földtani Közlöny 115, 63-77 (in Hungarian, with English abstract).

Sztanó, O., Magyar, I., Horváth, F., 2007. Changes of water depth in the Late Miocene Lake Pannon revisited: the end of an old legend. Geophysical Research Abstracts 9, 38897836.

Sztanó, O., Sebe, K., Csillag, G., Magyar, I., 2015. Turbidites as indicators of paleotopography, Upper Miocene Lake Pannon, Western Mecsek Mountains (Hungary). Geologica Carpathica 66, 331-344.

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

Tari, G.C., Horváth, F., 2006. Alpine evolution and hydrocarbon geology of the Pannonian Basin: An overview. In: Golonka, J., Picha, F.J. (Eds.), The Carphathians and their foreland: Geology and hydrocarbon resources. AAPG Memoir 84, 605–618.

ter Borgh, M., Radivojević, D., Matenco, L., 2015. Constraining forcing factors and relative sea-level fluctuations in semi-enclosed basins: the Late Neogene demise of Lake Pannon. Basin Research 27, 681-695. doi: 10.1111/bre.12094.

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

Törő, B., Sztanó, O., Fodor L., 2012. Aljzatmorfológia és aktív deformáció által befolyásolt pannó-niai lejtőépülés Észak-Somogyban. Földtani Közlöny 142, 445-468 (in Hungarian, with English abstract).

Uhrin, A., Magyar, I., Sztanó, O., 2009. Effect of basement deformation on the Pannonian sedimentation of the Zala Basin, SW Hungary. Földtani Közlöny 139, 273-282 (in Hungarian, with English abstract).

Uhrin, A., Sztanó, O., 2011. Water-level changes and their effect on deepwater sand accumulation in a lacustrine system: a case study from the Late Miocene of western Pannonian Basin, Hungary. International Journal of Earth Sciences 101, 1427-1440. dx.doi.org/10.1007/s00531-011-0741-4.

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

Ustaszewski, K., Kounov, A., Schmid, S.M., Schaltegger, U., Krenn, E., Frank, W., Fügenschuh, B., 2010. Evolution of the Adria-Europe plate boundary in the northern Dinarides: From continent-continent collision to back-arc extension. Tectonics 29, TC6017, doi: 6010.1029/2010tc002668.

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

Várkonyi, A, Törő, B, Sztanó, O, Fodor, L., 2013. Late Cenozoic deformation and tectonically controlled sedimentation near the Balaton zone (central Pannonian basin, Hungary). Occasional Papers of the Geological and Geophysical Institute of Hungary 72–73. ISSN 2064-0293, ISBN 978-963-671-294-5

Visnovitz, F., Horváth, F., Fekete, N., Spiess, V., 2015. Strike-slip tectonics in the Pannonian Basin based on seismic surveys at Lake Balaton. Int. J. Earth Sci. 104, 2273-2285.

Williams, S.R.J., 1987. Faulting in abyssal-plain sediments, Great Meteor East, Madeira Abyssal Plain. Geol. Soc. Lond. Spec. Publ. 31, 87–104.

Xie, X., Heller, P.L., 2009. Plate tectonics and basin subsidence history. GSA Bulletin 121, 55–64. doi:10.1130/B26398.1.

Xu, S., Hao, F., Xu, C.,Wang, Y., Zou, H., Gong, C., 2015. Differential compaction faults and their implications for fluid expulsion in the northern Bozhong Sub-basin, Bohai Bay Basin, China. Marine and Petroleum Geology 63, 1–16.