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Invited Review

Evolution of the Pannonian basin and its geothermal resources

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ABSTRACT

The Pannonian basin is an integral part of the convergence zone between the Eurasian and Nubian plates characterized by active subductions of oceanic and continental plates, and formation of backarc basins. The first part of this paper presents an overview of the evolution of the Alpine-Mediterranean region in order to understand the large scale crustal and upper mantle processes in and around the Pannonian basin, resulting a collage of terranes of Alpine and Adriatic origin. It will be followed by a summary of the history of sedimentation, volcanism and tectonic activity. As an illustration, three regional cross sections have been prepared on the base of seismic and borehole data. Reviewing current tectonic ideas and models, we come up with a speculative tectonic scenario depicting Alcapa and Tisza-Dacia as orogenic wedges detached from their mantle lithosphere in the Alpine and Adriatic/Dinaric collision zone during the Late Oligocene to Early Miocene. They suffered a dramatic thermal impact leading to crustal melting during extrusion, when these crustal flakes could have been directly superimposed on the asthenosphere in the Carpathian embayment. Since then, the large part of the Pannonian has been cooling and a new mantle lithosphere growing. Geothermal data show that the Pannonian basin with cessation of volcanic activity in the Late Miocene is still very hot and Miocene to Quaternary clastic basin fill, together with karstified Mesozoic carbonates form good geothermal reservoirs of regional extent. In addition to these gravity-driven aquifer systems, a strongly overpressured reservoir can be found below a regional pressure seal in synrift strata and fractured basement rocks. Eventually, we show maps of geothermal installations in the Pannonian basin and suggest that at the present level of knowledge and geophysical surveying it is easy to find additional resources, however proper water management is a critical issue to avoid harmful drawdown of the groundwater table.

Contents

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1. Introduction

The Alpine-Mediterranean region has been a wide zone of convergence between the Eurasian and Nubian plates during the Late Cretaceous and Cenozoic time. In this overall compressional setting extensional basins have developed from the Oligocene in association with subduction zones, which consumed oceanic and attenuated continental lithosphere. Slab rollback and the related upper mantle flows have been suggested to control extensional disintegration of orogenic terranes (Jolivet et al., 2009; Handy et al., 2010). Mediterranean backarc basins, particularly their continental areas and the islands offer challenging sites for exploration of geothermal resources (Cloetingh et al., 2010). The Pannonian basin in eastern Central Europe constitutes an integral part of the Alpine orogenic system. The Alpine, Carpathian and Dinaric mountains surround this sedimentary basin of Miocene through Quaternary in age. The basin is superimposed on two distinct orogenic terranes (Alcapa and Tisza-Dacia) derived from different paleogeographic position of the Alpine orogenic system (Schmid et al., 2008). The basin is of extensional origin and its formation was accompanied by intensive calc-alkaline magmatism with a paroxysm of silicic volcanism during the Early and Middle Miocene climax of rifting. Attenuated crust and lithosphere, high heat flow and temperature gradient values developed in the Pannonian region (Horváth et al., 2006). This is not the case of the Vienna basin to the West, and the Transylvanian basin to the East, and the mechanism of the formation of these peripheral basins is different from that of the Pannonian basin system (Fig. 1).

Thermal springs has been known and utilized in the Pannonian basin since the Roman times. Drilling exploration for thermal waters dates back as early as the second half of the 19th century. Scientifically and technically, the most notable venture was a 970 m deep well completed in 1878 by Vilmos Zsigmondy at the eastern side of Budapest. The drill hit Triassic karstified carbonates below Tertiary sedimentary rocks and produced thermal water of 74 °C with a yield of 1200 m^3/day. This resource has been supplying since then a popular thermal bath and balneological therapy center in the City Park of Budapest. Starting from the early 20th century, hydrocarbon exploration wells have regularly encountered thermal water-bearing beds in deeper Miocene sedimentary strata. Hungary has become progressively a land of thermal spas and played a pioneering role in the development of balneotherapy. It became quite soon a widely accepted knowledge among hidrogeologists that the secret of successful hot water exploration in Miocene clastic rocks was to find a best yield “thermal water-bearing layer”. However, the origin of this specific layer showing changing depth and thickness, as well as water yielding capacity has been understood only recently in the framework of new results on the structural and stratigraphic evolution of the Pannonian basin. These results will be reviewed in two parts. First, the Alpine-Mediterranean tectonic evolution will be briefly summarized to give a frame for the description of stratigraphic, volcanic and structural history of the Pannonian basin. In the second part, geothermal data and parameters controling regional fluid flow systems in the basin fill and basement will be addressed. In conclusion, we arrive at the final message that complete understanding of basin evolution is a prerequisite in the assessment of geothermal resources and location of high-enthalpy reservoirs.

2. Tectonic evolution of the Pannonian basin

2.1. Alpine-Mediterranean framework

Fig. 2a presents a sketch map to illustrate the early Late Cretaceous (100–80 Ma) paleogeography of Alpine-Mediterranean tectonic units, which took part in the formation of the Pannonian basin and surrounding orogens. Fig. 2b shows a summary diagram with the ages of the tectonostratigraphic units and the timing of the main tectonic events from the beginning of the Triassic. The map and the diagram, as well as the following review of Alpine orogeny rely largely on the results and ideas expressed by Csontos and Vörös (2004), Schmid et al. (2008) and Handy et al. (2010). We accept their definition of tectonic units and use the term *ocean* to describe a deep trough, either narrow or wide, developed by continental rifting and exhibiting subductible lithosphere. Former subduction process can be inferred from rock record including the presence of obducted ophiolites, orogenic flysch in the accretionary wedge, calc-alkaline volcanic arc and HP/LT metamorphic belt. The sketch map shows that two remarkably different oceanic realms existed between the converging European and the Adriatic continental blocks: the older Neotethys and the younger Alpine Tethys (Fig. 2a). The Neotethys located to the East of Adria started to develop as early as Late Paleozoic by rifting the northern margin of Gondwana, followed by northward drifting the Cimmerian superterrane and closure of the Paleotethys (Stampfli and Borel, 2002). Neotethyan rifting propagated towards the West and the Meliata-Maliac ocean developed in the Late Triassic (Fig. 2b). This opening oceanic basin was surrounded by a large shelf region, where production of carbonates was very high, due to Late Triassic climatic conditions. Extensive carbonate platforms developed on the shallow shelves and the slopes transitional to the Neotethys ocean pelagic limestones were deposited (Haas et al., 1995).

Two additional branches of the Neotethys can be recognized in the area: the Western and Eastern Vardar ocean (Schmid et al., 2008). The Western Vardar ocean spread between the Dinaric margin of Adria and the European continent incorporating the Tisza and Dacia terranes. The Eastern Vardar ocean developed at the end of the Triassic, when Tisza rifted apart from Europe. Dacia was also separated from Europe at the beginning of the Jurassic by opening

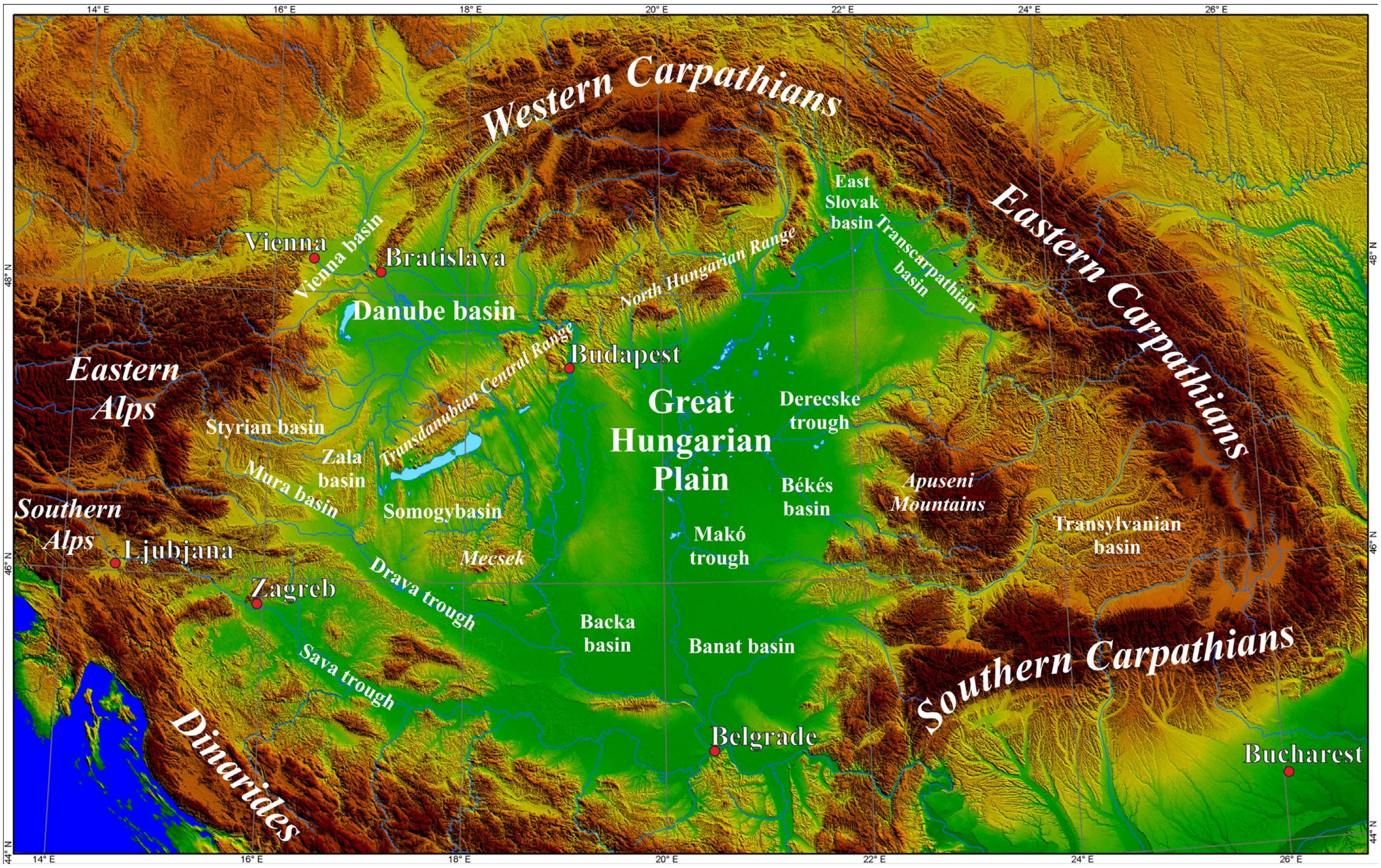

Fig. 1. Digital terrain model of the Pannonian basin to show its position within the Alpine mountain belt and the location of different subunits.

of the Ceahlau-Severin ocean, which is considered a part of the Alpine Tethys, because of its paleogeographic position (Fig. 2a). Closure of this complex system of continental blocks and intervening oceanic troughs in the Neotethyan realm started during the Late Jurassic (Fig. 2b). Obduction of large ophiolitic masses from the Western Vardar onto the Dinaric margin, and the Eastern Vardar onto Dacia margin were initiated at this time. Early Cretaceous ophiolites in the Sava zone are interpreted as derived from a Sava backarc basin generated by intra-oceanic subduction in the West Vardar domain (Schmid et al., 2008). The process of convergence culminated in the late Early Cretaceous (Aptian-Albian) time, and led to complete closure of the Eastern Vardar and Ceahlau-Severin oceans by Late Cretaceous time (Fig. 2a). The Western Vardar ocean became fully consumed as late as the Early Eocene by subduction below Tisza, as well as Dacia. The peak period of continental collision in the Dinarides took place during the Eocene and Oligocene expressed by large-scale thrusting and its propagation towards the external Dinarides (Tomljenovic et al., 2008; Ustaszewski et al., 2010). The Alpine Tethys was born to the North and West of Adria as part of the rifting process that created the Central Atlantic ocean (Stampfli and Borel, 2002). The Ligurian-Piemont-Vahic-Pieniny oceanic branch developed from the beginning of the Jurassic and penetrated as far as the Ceahlau-Severin trough. Another branch, the Valais-Rhenodanubian-Magura formed later, during the Early Cretaceous in a more external position and separated the Iberia-Brianconnais continental block from Europe (Fig. 2a). Between the Alpine Tethyan and Neotethyan realms a broad boundary zone should have existed (Fig. 2a) to accommodate the different kinematics of plate tectonic processes (Handy et al., 2013). Closure of the Alpine Tethys can be divided into three stages (Handy et al., 2010). The first stage (Barremian through Aptian) was an east directed subduction of the Ligurian ocean. It was followed by southeast directed subduction of the Piemont-Vahic-Pieniny branch during the mid-Late Cretaceous to latest Eocene time. Then, the Valais-Rhenodanubian-Magura branch was consumed, including subduction of the Brianconnais continental terrane and the most distal parts of the European continental margin. Middle Eocene saw the onset of continental collision in the Central and Eastern Alps (Schmid et al., 2013). More towards the East, however, the Magura ocean was not closed and together with the rifted East European continental margin (Moldavides) constituted a landlocked basin with subductible lithosphere, what is called the Carpathian embayment (Fig. 3). Figs. 3 and 4 present block models from the surface down to the transition zone at bottom of upper mantle illustrating the Late Oligocene and present position of the Alcapa and Tisza-Dacia terranes, their lithospheric structure and subduction processes. These block models rely on the results and ideas expressed by Ustaszewski et al. (2008). Fig. 3 uses their maps on the top of the block model and Fig. 4 is just a slightly modified version of their Fig. 11. The Middle to Late Eocene continental collision both in the Alps and Dinarides, and subsequent propagation of thrusting towards the external domains was accompanied by delamination of mantle lithosphere and its rollback. Peri-Adriatic plutons with an age of 35–27 Ma (Kovács et al., 2007) are interpreted as a manifestation of slab break off in the Alps (von Blanckenburg and Davies, 1995; Handy et al., 2010). The detatched slabs was dripping into the upper mantle and accumulated on the top of the 660 km phase boundary (Fig. 3), thus contributed to a large lithosphere graveyard in the mantle transition zone beneath Alpine Europe and the central Mediterranean (Piromallo and Morelli, 2003).

The polarity of new subduction beneath the Eastern Alps after the slab break off is a debated issue, particularly because the subducting European mantle lithosphere beneath the Western Alps

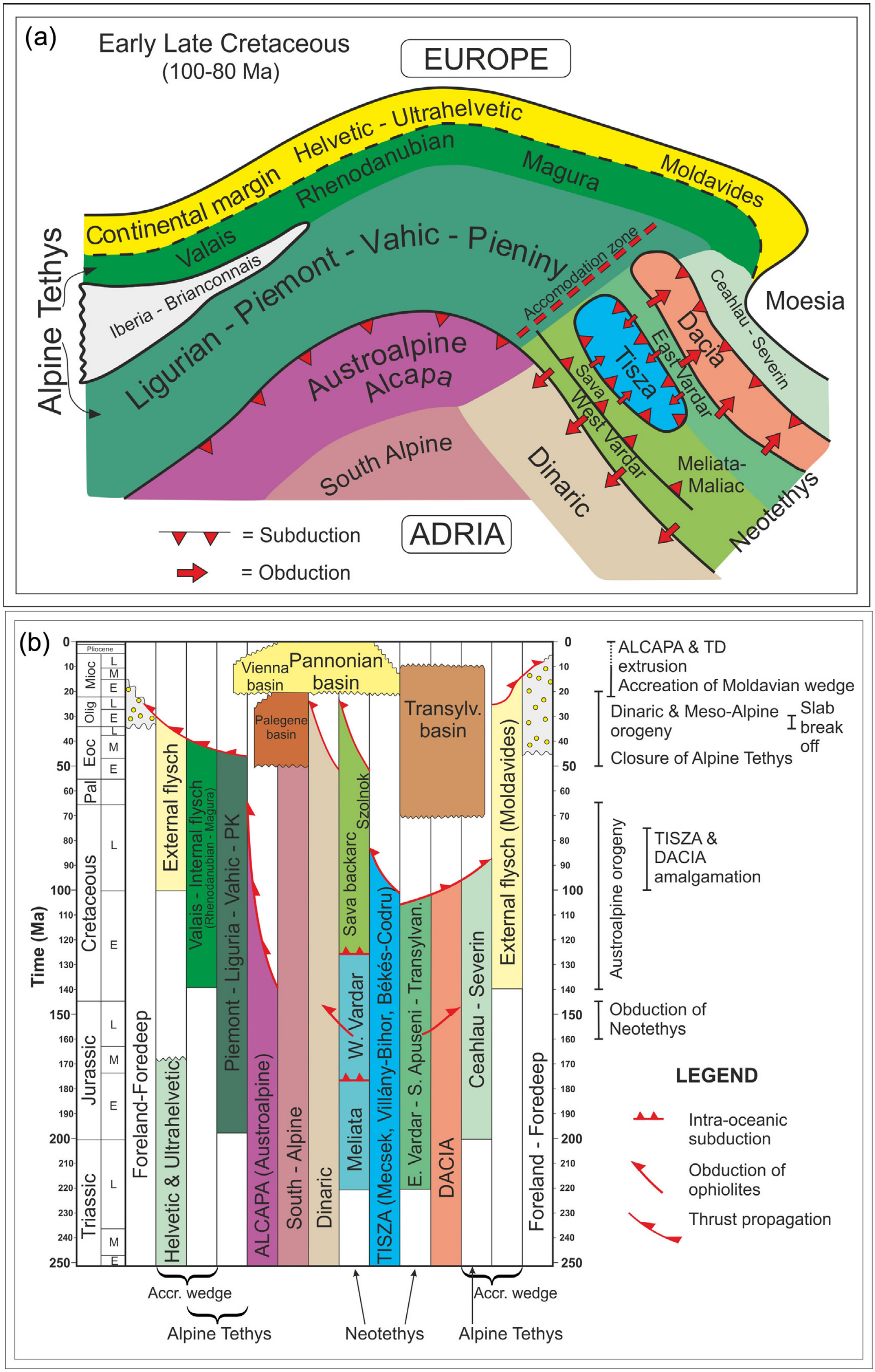

Fig. 2. (a) Schematic paleogeography of the Alpine-Mediterranean region in the early Late Cretaceous (mainly after Csontos and Vörös, 2004; Schmid et al., 2008; Handy et al., 2010). (b) Tectonostratigraphic chart of the megatectonic units in (a) (mainly after Csontos and Vörös, 2004; Schmid et al., 2008; Handy et al., 2010).

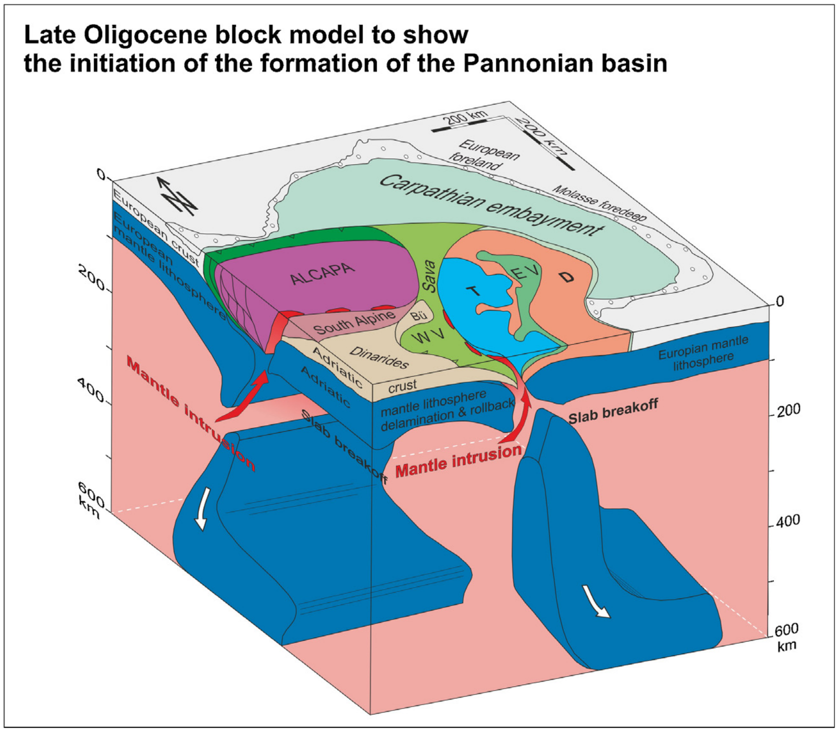

Fig. 3. Block model depicting the Late Oligocene position of the Alcapa and Tisza-Dacia terranes in the Carpathian embayment and the associated lithospheric and asthenospheric processes down to the upper mantle transition zone (inspired after Ustaszewski et al., 2008).

remained unchanged after the Oligocene. In the Eastern Alps, the more than 200 km Miocene to present convergence could have been accommodated either by renewed subduction of the European lithosphere towards the South, or development of a new Adriatic slab subducting towards the Northeast (Mitterbauer et al., 2011; Lippitsch et al., 2003, resp.). We accept Ustaszewski et al. (2008) argumentation and favour the second model as is shown in Fig. 4.

The history of slab beneath the Dinarides after continental collision appears to be less controversial. Similarly to the Alps, the Late Eocene to Oligocene granitoid intrusions in the internal zones are related to the delamination of the Adriatic mantle lid from the overlying continental crust (Schefer et al., 2011). It is suggested that rollback of this mantle lid controlled the exhumation history of the intrusive bodies exhibiting a fast cooling in the 16–10 Ma

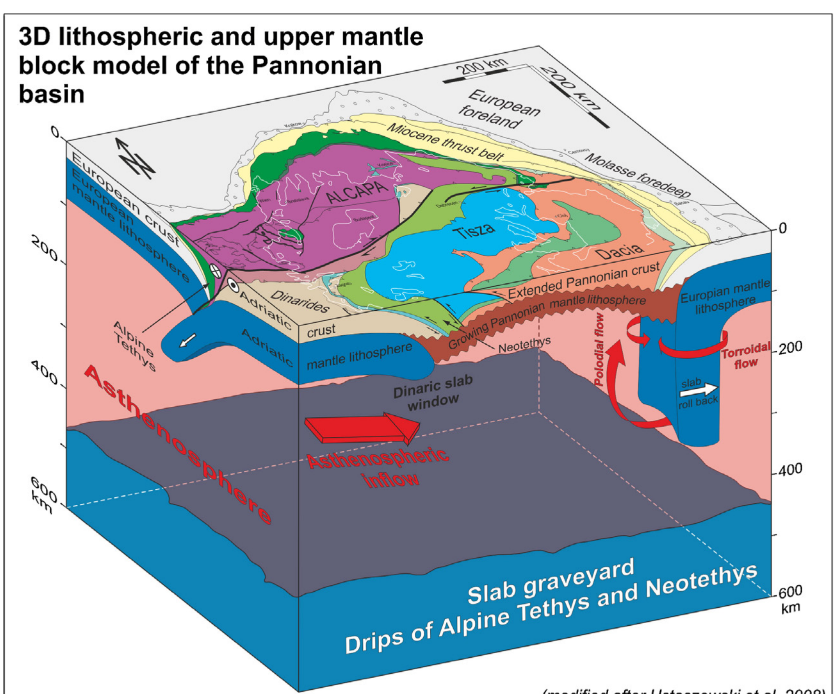

Fig. 4. Block model depicting the present position of the Alcapa and Tisza-Dacia terranes in the Carpathian embayment and the associated lithospheric and asthenospheric processes down to the upper mantle transition zone (inspired after Ustaszewski et al., 2008).

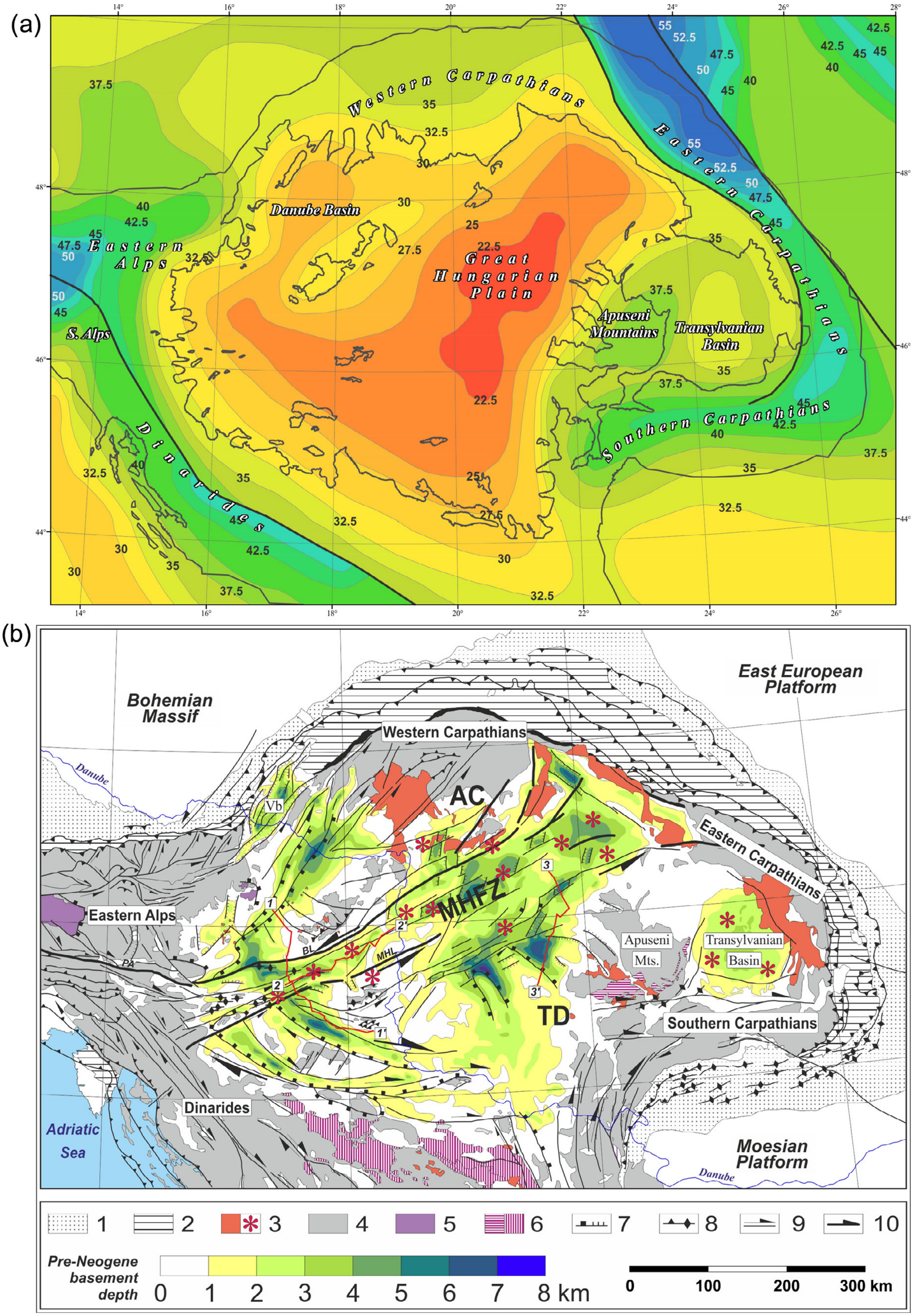

Fig. 5. (a) Crustal thickness map of the Pannonian basin and surrounding regions (modified after Horváth et al., 2006). (b) Map showing the depth to basement of the Pannonian basin and the main faults controlling the basin formation.
1 = Foredeeps; 2 = Flysch belt; 3 = Miocene volcanoes and approximate position of explosive centres erupting rhyolitic volcanoclasts; 4 = Inner Alpine, Carpathian and Dinaric mountains; 5 = Penninic windows; 6 = West- and East-Vardar ophiolites; 7 = Detachment and normal faults; 8 = Thrusts and folded anticlines; 9 = Strike-slip faults; 10 = First order strike-slip faults; AC = Alcapa terrane; MHFZ = Mid-Hungarian Fault Zone; TD = Tisza-Dacia terrane.

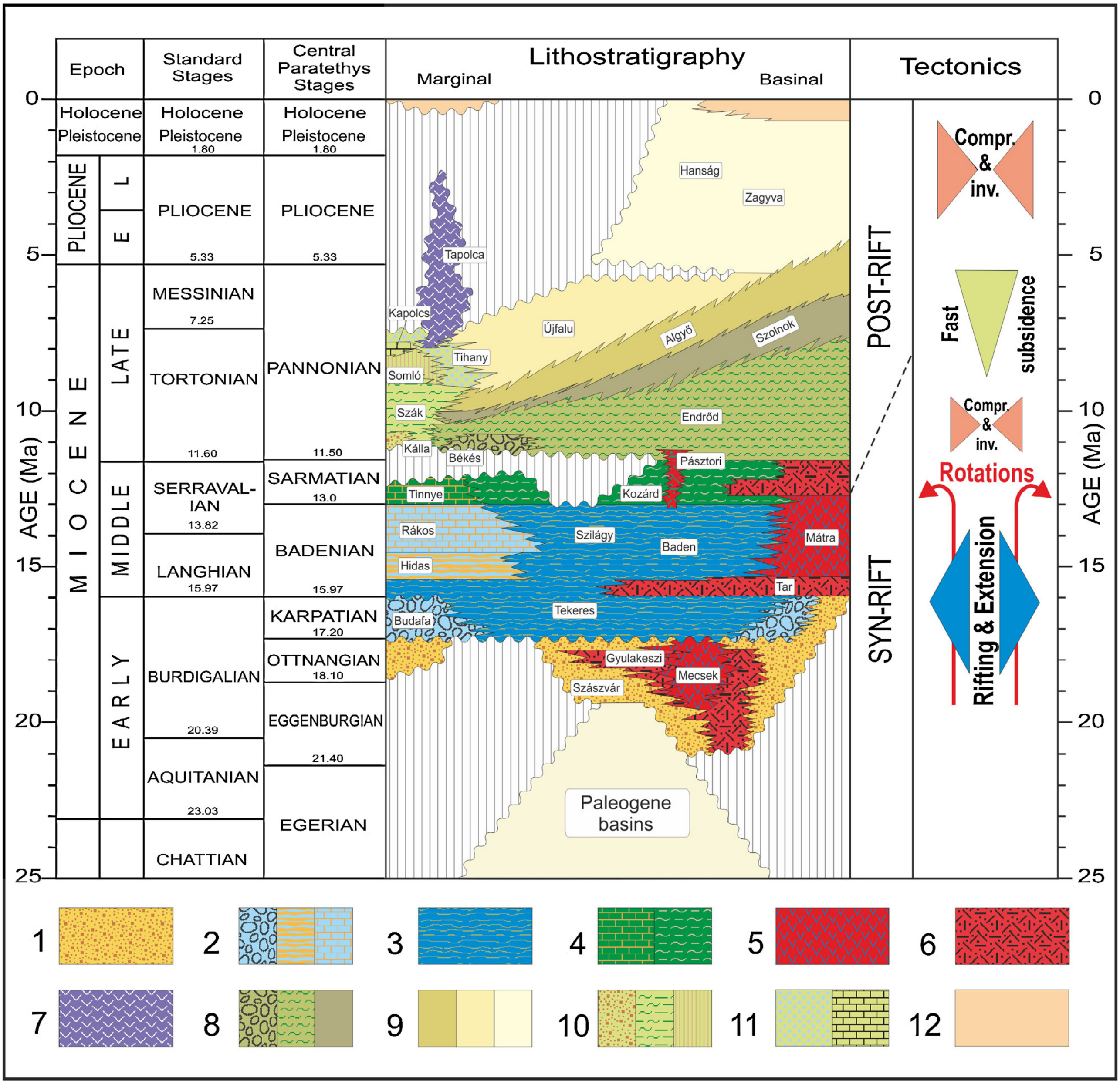

Fig. 6. Stratigraphic chart of the main syn- and postrift formations of the Pannonian basin.
Central Paratethys megasequence: 1 = Fluvial and lacustrine; 2 = Marine basin margin formations; Brackish; 3 = Marine pelagic; 4 = Brackish basin margin; 5 = Volcanics, mostly andesites; 6 = Volcanics, mostly rhyolitic ignimbrites and tuffs;
Lake Pannon megasequence: 7 = Volcanites, alkali basalts; 8 = Lacustrine, deep basin; 9 = Lacustrine, shelf slope to plain; 10 = Lake margin; 11 = Alluvial plain; 12 = Fluvial and aeolian.

time interval (Matenco and Radivojevic, 2012). Seismic tomography shows the absence of the slab under the northwestern part of the Dinarides (slab window between latitudes 46° to 44° N). Further to the southeast, however a continuous and progressively deeper penetrating slab can be seen towards the Hellenides (Bijwaard and Spakman, 2000; Piromallo and Morelli, 2003; Jolivet et al., 2009). Lateral extrusion of the Alcapa and Tisza-Dacia terranes from the East Alpine and Dinaric continental collision zones occurred towards the unconstrained margin of the Carpathian embayment (Figs. 3 and 4). Subduction and rollback of the lithosphere and development the Outer Carpathian accretionary wedge are geo dynamically coupled processes. Formation of the Pannonian basin was initiated in the Early Miocene, at about 21 Ma based on the age of the earliest sedimentary and volcanic rocks in rift grabens (see next chapter). The peak period of extension (early and middle rift climax sensu Prosser, 1993) was the Karpatian to Sarmatian time (Fig. 6), when the main structural pattern of the basin was established (Horváth et al., 2006). Onset and the process of extension can be constrained also by thermochronological studies of exhumed metamorphic rocks. The Tauern window is situated at the western margin of the Alcapa terrane and exposes highly metamorphosed oceanic and continental thrust sheets beneath Austroalpine nappes. Thermochronological data indicate slow exhumation (less than 1 mm/a) from about 30 Ma until 21–20 Ma. Exhumation rate increased up to 4 mm/a, and remained high until 14 Ma, in good agreement with the main rifting phase of the Alcapa terrane (Scharf et al., 2013). Similar history of uplift and fast cooling was determined for the Penninic ophiolites in the Rechnitz window (Dunkl et al., 1998), and the Pohorje pluton (Fodor et al., 2008). The Rechnitz window is located at the eastern boundary of the Eastern Alps and a large part of the exposed metamorphic oceanic rocks is covered by syn- and post-rift sedimentary rocks of the Pannonian basin (Tari et al., 1992). Recent studies have demonstrated a set of exhumed middle crustal rocks at the Dinaric margin of the Pannonian basin (Ustaszewski et al., 2010; Stojadinovic et al., 2013). Detachment and extensional collapse of the hanging wall (Sava and/or Tisza units) exhumed the Dinaric footwall in the Early to Middle Miocene time (18–12 Ma). Interestingly enough, core complex formation affected also the mountainous areas of the internal Dinarides in southern Serbia (e.g. Kopaonik block, Schefer et al., 2011). Over here granitic plutons in the footwall are exposed at the front of the Balkanides, and it makes difficult to imagine that this exhumation was related to the rollback of the Carpathian slab. Instead, ongoing delamination and rollback of the central and southern Dinaric slab has been suggested to control the extension of the southeastern part of the Pannonian basin, as well as the Vardar zone further to the South in the Balkan (Matenco and Radivojevic, 2012). This double control by two different slabs can be related to the process of counterclockwise and clockwise rotations of the two terranes (Alcapa and Tisza-Dacia, resp.) in the Pannonian basin. They became juxtaposed, but not fixed to each other along the Mid-Hungarian Fault Zone (MHFZ, Fig. 4) in the Middle Miocene and since then accommodating the differential displacement between the two terranes (Ustaszewski et al., 2008). This fault zone is actually a suture where Alcapa overthrust Tisza-Dacia and composed from strongly deformed elements of the Sava zone, largely displaced blocks of Dinaric affinity (e.g. Bükk Mts.) and the compressed Eocene to Oligocene deposits in the Szolnok flysch zone of eastern Hungary (Csontos and Nagymarosy, 1998). The first regional review of paleomagnetic data suggested, that during the Miocene 35° counterclockwise, and 100° clockwise rotations of the Alcapa and Tisza-Dacia terranes (resp.) occurred (Balla, 1987). On the basis of larger and more reliable data sets, recent studies corroborated the opposite rotation of the two terranes, but found differential rotations within each terranes and distinct phases of rotational events, with a peak activity in the 18–10 Ma time interval (Márton and Fodor, 2003; Márton et al., 2007; Van Hinsbergen et al., 2008; Tomljenovic et al., 2008; Ustaszewski et al., 2008). This implies that the main phase of rotation of the two terranes and their extensional disintegration occurred at about the same time. After consumption of all the subductible lithosphere in the Carpathian embayment, folding and thrusting of the Moldavian flysch wedge reached its climax and further extension of the Pannonian terranes slowed down and eventually terminated (Fig. 4). The Pliocene through Quaternary has been a period of stress field change towards a compressional regime generated by the continuing counterclockwise rotation of Adria (Grenerczy et al., 2005; Bada et al., 2007). Extension of the Alcapa and Tisza-Dacia terranes and continental collisions in the surrounding mountains resulted in a highly variable crustal thickness pattern of the region (Fig. 5a). Moho depth exceeds 50 km in the Alps and the Eastern Carpathians, 40 km in the Dinarides and Southern Carpathians. The Western Carpathians are characterized by close to normal crust with thicknesses from 30 to 35 km. The outline of the Pannonian basin is indicated by the 30 km isoline, and less than 22.5 km values have been observed at the most attenuated part underlying the Great Hungarian Plain (Janik et al., 2011). This is a surprisingly low value and simple isostatic balancing (Kaban et al., 2010) has shown indeed that the whole basin is "over-elevated", in other words it should be well below the sea level in case of equilibrium.

2.2. Stratigraphy

2.2.1. Base Pannonian unconformity

Paratethys sea was a distinct basin system to the North of the Mediterranean region, which existed from the Late Eocene to the Pliocene (Steininger and Rögl, 1985). This system stretched from the Rhone valley through the Pannonian basin to the Aral lake and composed of three subbasin: the Western, Central and the Eastern Paratethys. The complex history of their isolations, internal and external connectivity led to remarkable faunal endemism and elaboration of specific biostratigraphic time scales for each subbasins (Steininger and Wessely, 2010). Correlation of these scales generated long lasting debates and serious difficulties in understanding their evolutionary history in terms of sequence of events and absolute ages (Sacchi and Horváth, 2002). However, there has been a recent progress in the Central Paratethys realm, and a fairly well established correlation with the global chronostratigraphic scale is available now for the Late Oligocene through Middle Miocene time interval (Mandic et al., 2011). Towards the end of Middle Miocene, during the Sarmatian a brackish water environment with highly endemic fauna developed in the Central Paratethys. Quite uniform sedimentation characterized the whole sea, consisting of shallow water Early Sarmatian siliciclastics passing into biogenic carbonate dominated system during the Late Sarmatian. Lowstand at the beginning and the end of the stage can be correlated with late Serravallian sea-level drops of the glacio-eustatic curve (Ser3 and Ser4/Tor1, resp, Piller et al., 2007). Uplift of mountain belts around the Pannonian basin led to its separation from the rest of the Central Paratethys also at the end of Sarmatian, 11.6–11.3 Ma ago (ter Borgh et al., 2013). This led to the birth of the Lake Pannon, a large isolated lake with a lifetime of 7 to 8 million years (Magyar et al., 1999a). Influx of sediments transported by rivers, particularly the paleo-Danube progressively filled up the lake from the NW towards the SE and a reasonable timing of the shelf edge progradation has been elaborated (Magyar et al., 1999b, 2013). The base of the Pannonian is a prominent stratigraphic horizon due to sharp lithological and biostratigraphic contrasts in large territories of the basin (Magyar et al., 1999b). Furthermore, the uplift of mountain belts around the Pannonian basin apparently involved the central part of the basin system too, resulting in an unconformity characterized by angular discordance and erosional hiatus. However, in a few deep basins, where no erosion occurred, the Sarmatian and the overlying Pannonian marls (*Kozárd* and *Endrőd Formations*, resp.) are very similar and the base Pannonian unconformity cannot be identified in seismic sections and drillhole logs. Apart from these depocentres, Sarmatian strata is missing or present as erosional patches in a thickness of a few tens of metres (Nagymarosy and Hámor, 2012). In this respect, the central part of the Pannonian basin strongly contrasts with the peripheral basins in the West (Vienna and Styrian basins) and the East (Transcarpathian depression and Transylvanian basin), where widespread Sarmatian beds developed and preserved in a thickness of up to 1200 m (Harzhauser and Piller, 2004). The division of the basin fill by the base Pannonian unconformity into two megasequences (Central Paratethys and Lake Pannon) is useful also from a tectonic point of view. Study of seismic sections and borehole data suggested that this unconformity marked the end of main rifting phase (Horváth et al., 2006). However, more recent seismic data leads to a refinement of synrift/postrift boundary in the Pannonian basin (Balázs et al., 2013). Interpretation suggests that formation of growth strata in half grabens (late rift climax sequence of Prosser, 1993) continued in the eastern part of the Pannonian basin during the Late Miocene (Matenco and Radivojevic, 2012; ter Borgh, 2013). We acknowledge this finding and assign the synrift/postrift boundary as time transgressive, as is shown in Fig. 6 together with a stratigraphic chart and a summary diagram of the main tectonic events in the Pannonian basin.

2.2.2. Central Paratethys megasequence

The total thickness of the Neogene through Quaternary basin fill in the Pannonian basin is quite variable: deep troughs exceeding a thickness of 5 km, shallow basin areas typically with a depth of 1 to 2 km and "inselbergs" with exposed substrata of the basin occur (Fig. 5b). This is the final result of wrench fault controlled synrift subsidence, and postrift evolution including events of basin inversion and erosion.

Sedimentary rocks related to initiation of rifting are continental deposits, mostly coarse clastics, fluvial sands, marsh and lagoonal muds and lignite seems (Fig. 6). Recently, a vision of a large Early Miocene (Ottnangian and Karpatian) lake system with wrench fault-controlled depocentres in the inner Dinarides and the southern Pannonian basin has emerged (Mandic et al., 2011; De Leeuw et al., 2012). Coeval lacustrine sedimentary rocks in the central part of the basin (e.g. *Szászvár Formation*, Fig. 6) suggest that this lake system was even larger and a typical environment and tectonic style of the early synrift evolution of the Pannonian basin. The *Szászvár Formation* consists of conglomerates and pebbles, sandstones, mottled clays, marshy siltstones, lignites and carbonaceous clay, deposited in a fluvial and shallow lacustrine environment. The formation is up to 800 m thick and at many places thick interbeddings of rhyolite tuff layers occur. There are sporadic occurrences of calc-alkaline volcanic rocks at the internal boundary of the Dinarides with ages of 23–21 Ma. This initial volcanic activity was followed by massive eruption of felsic calc-alkaline magmas in the 21–17 Ma time intervals (Pécskay et al., 2006). Deposition of thousand metres of rhyolitic ignimbrites and tuff layers (*Gyulakeszi Formation*) took place over large areas, mostly in the Mid-Hungarian Fault Zone and northeastern part of the Alcapa terrane (Fig. 5b). The widespread occurrence of these felsic volcanic products and their interfingering with the earliest synrift strata makes reasonable to assign 21 Ma as the beginning of the formation of the Pannonian basin (Fig. 6).

This initial phase of felsic magmatism was followed by similarly intensive eruptions of rhyolitic ignimbrites, tuffs and occasional outflows of rhyolites and dacites during the Early Karpatian to Late Sarmatian time (Fig. 6). Finding the location of volcanic centres is difficult and the red asterisks in Fig. 5b should be taken as an educated guess. Actually, accumulations of pyroclasts and tuffs in domes of thousands of metres thickness over large territories in the central and northeastern part of the Pannonian basin suggest eruptions via crustal scale fissures (Nagymarosy and Hámor, 2012). The early rift climax sequence of sediment accumulation is represented by Karpatian strata composed from siltstone, claymarl, sandstone and conglomerate beds of the *Budafa Formation* in a total thickness of about 800 m. The continued transgression of the Central Paratethys sea led to the formation of the true marine *Tekeres Schlier* in a thickness of 200–400 m (Fig. 6). It is made up of sandy clay and claymarl with abundant microfauna. Further transgression resulted in open marine conditions, where the *Baden Formation* deposited. It is composed of monotonous grey clay and claymarl, which accumulated in a thickness of up to 1000 m in deep basins. In the Late Badenian, open marine sedimentation continued by the deposition of the *Szilágy Clay Marl Formation* in a thickness of 300–450 m. Badenian strata can represent the middle rift climax sequence of synrift strata. Middle Miocene formations are often intersected by the tuff layers of the *Tar Formation*. In the northeastern part of the Alcapa terrane the andesites of the *Mátra Formation* can be found forming spectacular mountains of the present landscape (Fig. 5b). At the margins of deep basins sediments composed of alteration of brown coal, sand and silt were deposited (*Hidas Formation*). They are overlain by the *Rákos Formation*, which is an algal limestone. At many places it is followed by the very similar Sarmatian biogenic limestone, the *Tinnye Formation*. Elsewhere, accumulation of claymarls (*Kozárd Formation*) occurred in a near shore environment. Sarmatian represents a time period during the late synrift phase with decay of volcanic activity and a landscape without remarkable morphological highs and lows. This is however, the result of post-Sarmatian (early Pannonian) phase of basin inversion and erosion, which left behind a remarkably denudated landmass.

2.2.3. Lake Pannon megasequence

The flat and eroding land started to subside at early Pannonian time and the rising lake level flooded significant land areas (Fig. 6). Rapid subsidence began at places which led to locally 800–1000 m water depths (Magyar et al., 1999, 2007). This large and deep lake was characterized by a well-developed shelf system at the lake margin. During the initial period (until 10 Ma) sediment influx was relatively low, which resulted in the formation of calcareous marl and claymarl (*Endrőd Formation*) in the deep basins. Coarse-grained rocks formed by abrasion only along the basin margins, with a thickness of not more than a few 10 m (*Kálla* and *Kisbér* Formations). Lake Pannon reached its largest extent about 9.8 Ma. Then, rapid infill of clastic material started by inflow of large rivers from two main directions: northwest (paleo-Danube) and northeast (paleo-Tisza). The area of Transdanubia was under the influence of the sediments feed from the northwest, which led to the formation of a sedimentary shelf-slope system (sensu Posamentier and Allen, 1999) prograding southwards (Fig. 7). The main Pannonian lithostratigraphic formations are defined on the basis of this depositional model (Bérczi, 1988; Mattick et al., 1988; Juhász, 1991; Juhász et al., 1997). By definition, these formations are time transgressive (Fig. 6), but integrated application of biostratigraphic, seismic stratigraphic and magnetostratigraphic methods resulted in a good timing of their development (Magyar et al., 1999a, 2013). Furthermore, basic principle of seismic stratigraphy (Vail et al., 1977) holds that reflection horizons in seismic data represent chronostratigraphic surfaces, which are images (with a given resolution) of the successive paleobottoms of depositional basins and their erosional unconformities. This implies that our definition of the Pannonian lithostratigraphic units is adequate (Figs. 6 and 7), and makes possible their identification with the use of seismic and well log data.

Turbidity currents originating from the slopes led to a gradual transition of the marls of the *Endrőd Formation* to the turbiditic *Szolnok Formation*, which is built up by alternating fine grained sandstone and pelitic layers. Usually, the individual sand bodies cannot be correlated between distant localities, suggesting that they are laterally confined. The sediments of the prograding slope itself are defined as *Algyő Formation*, comprising chiefly silt and argillaceous marl, with very subordinate, lense-like sandy intercalations of mass flow origin. The thickness of this formation (generally 200–600 m between the topsets and bottomsets) reflects the depth of the water body in which the progradation took place. Above the slope deposits, there is an abrupt change from pelitic to sandy lithology, marking the base of *Újfalu Formation*. The lower part of Újfalu Formation contains thick (up to 20–30 m), coarsening-upward sand sheets, which are occasionally separated by a few metres thick silt beds. Most of the sand sheets, however are in connection with each other over long distances, which make this unit a very efficient fluid reservoir of regional extent. This basal part of the *Újfalu Formation* is actually the legendary "thermal water-bearing layer" in the Pannonian basin, the main target of thermal water exploration. Its thickness is 100–300 m in deep basins, while it pinches out along the basin margins and above basement highs. The upper part of *Újfalu Formation* (reaching a thickness of 500–1500 m) consists of sediments deposited on a coastal plain, formed on the shallow shelf. Laterally restricted, fining-upward channel sand bodies are predominant in this unit, interbedded with siltstones, claystones and marls, locally lignite beds. The uppermost part of the Pannonian strata is characterized by the cessation of the lacustrine influence. As the shelf margin of Lake Pannon moved further towards the South, the coastal plain environment was replaced by an alluvial plain across a larger and larger area. Sediments of

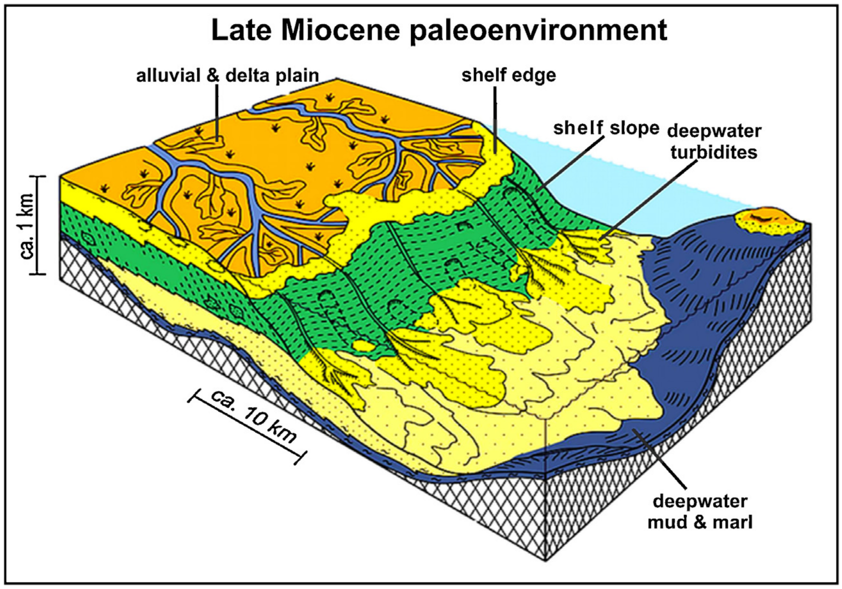

Fig. 7. Generalized cross-section of Lake Pannon in the Late Miocene showing the characteristic depositional environments in front and on the top of a prograding shelf margin (after Juhász, 1991; Bérczi, 1988).

the *Zagyva and Hanság Formations* are channel fills, floodplain and marsh deposits and paleosols, and they are passing continuously into recent sediments. From a litho- and biostratigraphical point of view these formations are very similar to each other. Alkaline basaltic magmatism took place in the 8–2 Ma time interval (Fig. 6), and resulted in small monogenetic volcanic fields distributed sporadically in the Pannonian basin (Wijbrans et al., 2007; Harangi and Lenkey, 2007). The presence of a Messinian event in the Pannonian basin is a hotly debated issue. Seismic data show an angular unconformity at the upper part of Pannonian strata in the deeper subbasins (Sacchi et al., 1999). Magnetostratigraphic profiles available in a couple of boreholes if correlated by seismic data suggest an age of 5 to 6 Ma for this unconformity, which makes tempting to relate it to the Messinian salinity crises. However, the school of thought on the complete isolation of the Lake Pannon explains this coincidence in terms of tectonics, suggesting that the unconformity was brought about by the onset of basin inversion (Magyar and Sztanó, 2008). Another school applies stratigraphic modelings and concludes that tectonic inversion was accompanied by an absolute lake-level drop with amplitude of up to 200 metres, followed by a lake level rise in concert with the Messinian history of the Mediterranean (Csató et al., 2007, 2013). Leever et al. (2010) argue that there was a permanent connection between the Pannonian and Dacic basins after the Sarmatian through the Danube gateway in the Southern Carpathians. Messinian sea level fall in the Dacic basin resulted in deeper fluvial incision into the sill of the gateway, and a lowering of the Pannonian lake level. It has been suggested that fall of the water level of about 50 to 100 metre occurred at 5.8 Ma in the Dacic basin, and major change in Eurasian climate to warmer and more humid resulted in a widespread transgression from 5.5 Ma onward (Krijgsman et al., 2010). Even if we are not taking a stand in this debate, we attempted to map this Miocene/Pliocene unconformity on the profiles constructed to illustrate the stratigraphic architecture and tectonics of the Pannonian basin.

2.3. *Regional geophysical-geological profiles*

Fig. 5b shows the location of three profiles, which can be seen in Figs. 8a, 9 and 10. Fig. 8b gives the legend to the profiles. Profile 1 is located in the western part of the basin and passing from the Alcapa terrane through the MHFZ into the Tisza terrane. Profile 2 is a perpendicular section and located completely in the MHFZ. Profile 3 is situated in the eastern Pannonian basin and starting from the southern edge of the MHFZ traverses the Tisza-Dacia terrane. They were composed from numerous individual 2D seismic sections of different vintages and quality. The time-depth conversion of the interpreted horizons has been carried out by the use of average velocities for the main stratigraphic units. Average velocities were derived from check-shot surveys from the study area. It is to be noted that the profiles are vertically over exaggerated by a factor of 8 to 10.

The selected wells along the profiles are all not farther than 5 km from the seismic lines. In the course of interpretation of the basement units and the most important faults we relied on, but not necessarily agreed with the pre-Cenozoic geological map and textbook of Hungary (Haas et al., 2010, 2012). For convenience, we call all of the rock masses on this map *basement*, including the Mesozoic cover sequences of the Paleozoic crystalline rocks. This is because these pre-Cenozoic rocks constitute the substrata of the two superimposed basin systems in the Pannoian region, i.e. the Paleogene and Neogene basins (Fig. 2b). During our interpretation of the profiles Bouguer-anomaly and vertical component magnetic anomaly maps of Hungary were also taken into consideration (www.mfgi.hu).

2.3.1. *Profiles 1 and 2*

Profile 1 starts NW at the southern part of the Danube basin (Fig. 1). The basement, at a depth of 2.5–3 km, is composed of various low-grade metamorphic rocks of Early Paleozoic age and they are structurally assigned to the Upper Austroalpine nappe system (Schmid et al., 2008). Moving towards SE the profile crosses the NE–SW striking "*Rába line*" that separates the structurally deeper Austroalpine unit from the overlying Transdanubian Range (TR) unit. The "*Rába line*" represents a complex Miocene normal and strike-slip fault which strongly reworked the primary Cretaceous nappe contact of the Austroalpine and the Transdanubian Range units (Tari, 1996). Lithologically, the TR consists of a low-grade metamorphic Variscan basement and an unconformably overlying, non-metamorphosed Permian–Cretaceous sedimentary cover made up predominantly of massive Triassic carbonates and mixed carbonatic siliciclastic rocks (Szafián et al., 1999). Morphologically,

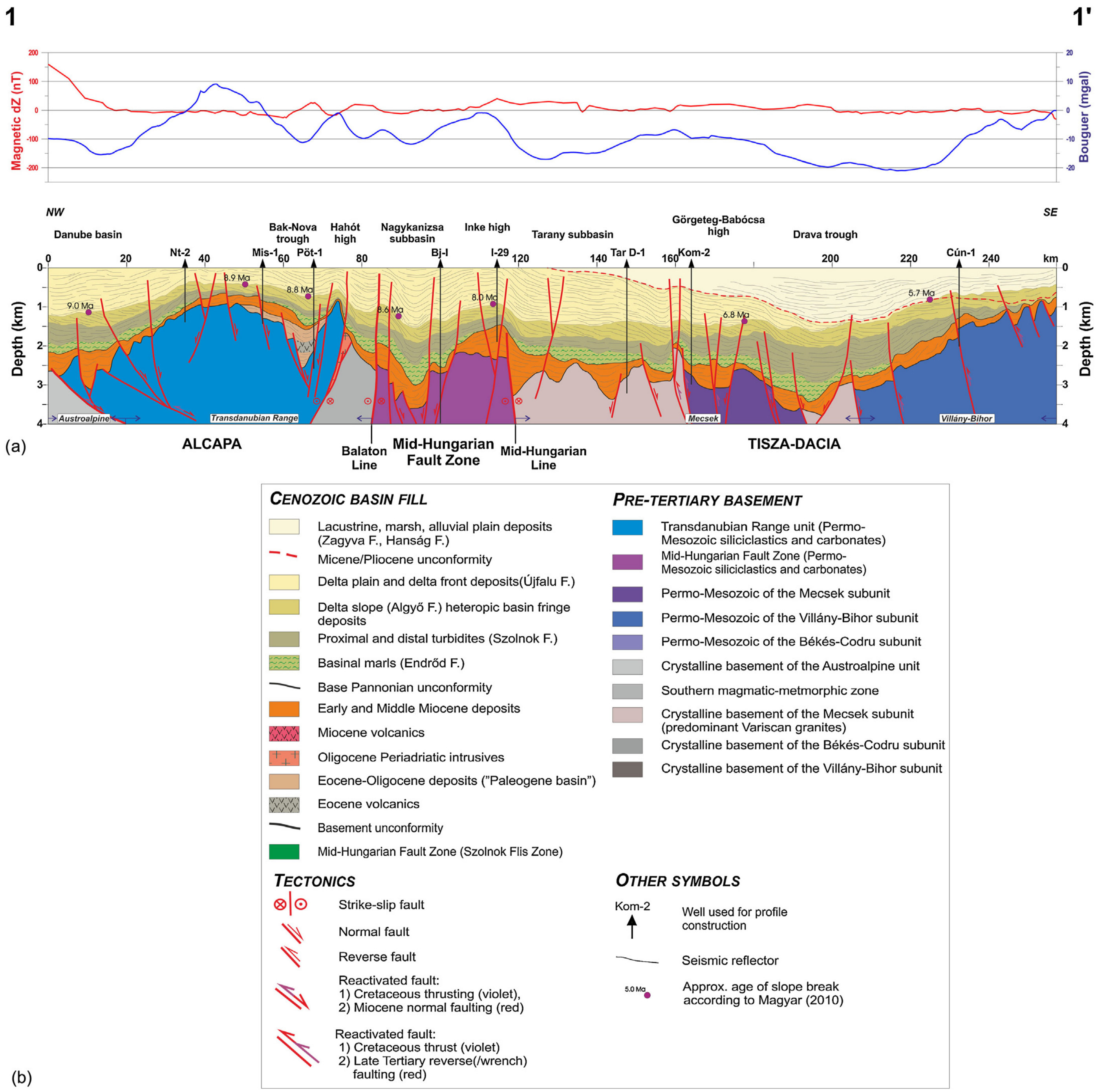

Fig. 8. (a) NW-SE directed composite geological profile across the western Pannonian basin (Profile 1, see location in Fig. 5b). (b) Legend to geological profiles.

it forms a dome-like feature in the subsurface continuation of the exposed Transdanubian Range (Fig. 5b).

On its western flank, the narrow Bak–Nova trough with about 1500 metres of Eocene volcanic and sedimentary rocks represents a strike-slip fault controlled basin. It is bordered on the South by the characteristic Hahót basement high. This narrow high marks the southern boundary of the thick Upper Triassic platform carbonates that form the main thermal karst aquifer of the Transdanubian Range Unit. Triassic platform carbonates often make up a common hydrodynamic system with unconformably overlying Upper Cretaceous reef limestones and/or Eocene/Miocene carbonates. South of the Hahót high a narrow zone made up from metamorphic rocks of variable grade and an Oligocene intrusive body, belonging to the Periadriatic plutons follow (Benedek, 2002). Further to the South profile 1 traverses the Balaton line, which is the continuation Periadriatic fault, hence it represent the southern boundary of the Alcapa terrane (Schmid et al., 2008). MHFZ is the suture between the Alcapa and Tisza-Dacia terranes that is known only from boreholes in the territory of Hungary (Haas et al., 2010). Its structural complexity is the results of multiphase deformations including subduction and obduction of the Sava backarc lithosphere in the Cretaceous, shortening and right-lateral wrenching in the Paleogene and the left-lateral wrench faulting from the Miocene to present time (Csontos and Nagymarosy, 1998, Ustaszewski et al., 2008). This young phase of deformation is indicated in Fig. 8a, which presents a highly oversimplified scheme of the actual structural

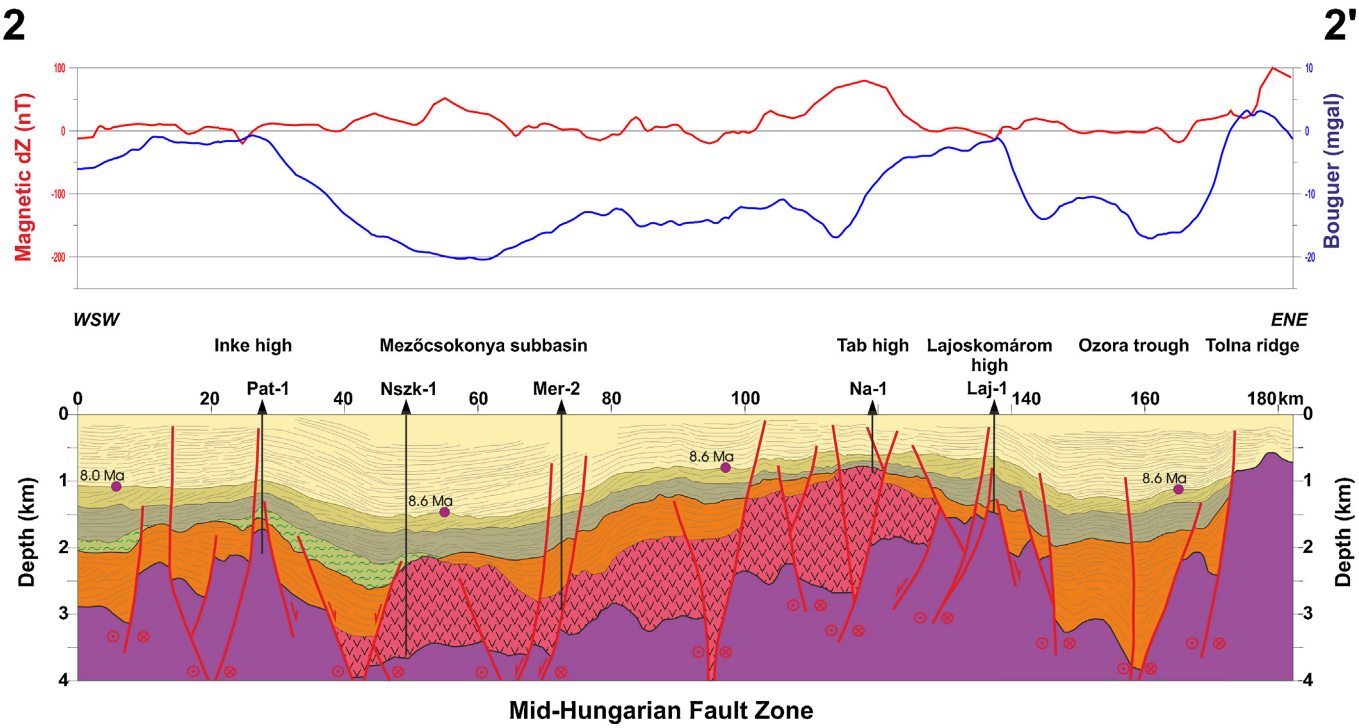

Fig. 9. W-E directed composite geological profile in the MHFZ (Profile 2, see location in Fig. 5b).

conditions. The basement is deeply buried by Neogene sediments and volcanics within the whole MHFZ unit. The deep Nagykanizsa subbasin reaches a depth of about 4 km along the section and is bounded by the Inke high at the South, where several wells reached the basement. These wells exposed mainly Middle- to Upper Triassic carbonates that are widespread in the whole MHFZ, and show a clear South Alpine/Dinaric affinity (Csontos and Vörös, 2004). The deeply buried Triassic platform carbonates are the most important geothermal reservoirs in the unit. In contrast to the Transdanubian Range, these carbonates have no direct surface recharge area, therefore they contain stagnant thermal waters with high salinity. The NE–SW oriented profile 2 (Fig. 9) passes all the way along the MHFZ and shows the Triassic carbonates in the basement. The next megatectonic unit at the southeastern part of profile 1 is the Tisza terrane separated structurally by the Mid-Hungarian line from the MHFZ (Figs. 5b and 8a). The internal structure of the Tisza unit

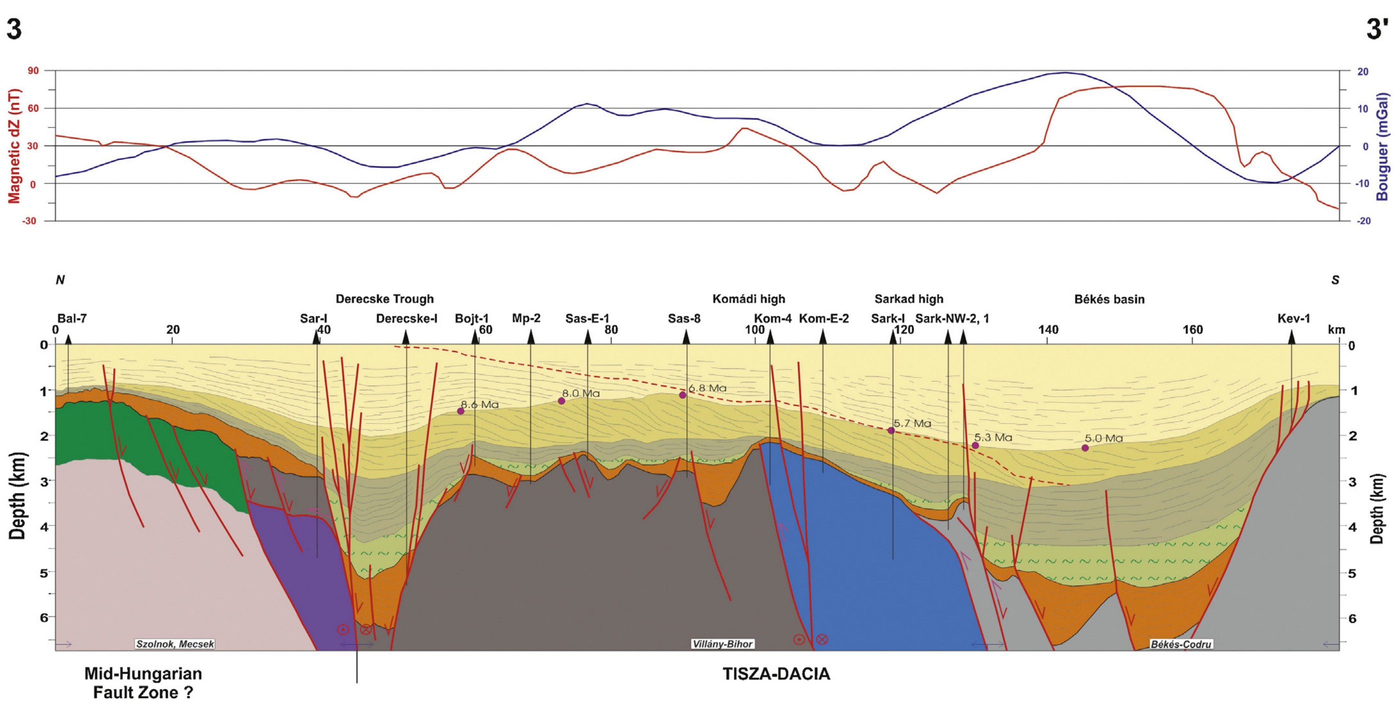

Fig. 10. NE-SW directed composite geological profile across the eastern Pannonian basin (Profile 3, modified after Balázs et al., 2013, see location in Fig. 5b).

is characterized by a WNW vergent nappe system formed during Cretaceous tectogenesis, when Tisza as well as Dacia were at the front of the Dinaric-Vardar subduction-obduction system (Fig. 2a). Indeed, if Tisza-Dacia is rotated counterclockwise by about 100°to its early Cretaceous position, the geometry of the thrust system becomes the same as that of the Dinarides. The Tisza-Dacia nappe pile was later heavily reworked by postorogenic uplift and erosion, and normal and strike-slip faulting during the formation of the Pannonian basin. The basement is situated mostly in a deep position except the Görgeteg-Babócsa crystalline high and the subsurface continuation of the Villány Mts. at the end of the profile. The basement can be divided into two main groups:.

(1) Variscan medium-grade metamorphic rocks reached by several boreholes in the Tarany subbasin and the adjoining Görgeteg-Babócsa high, as well as in the Central Dráva trough.
(2) Non-metamorphic Late Paleozoic–Mesozoic cover sequences in the western subsurface extension of the Villány Mountains. This complex represents again a large fractured/karstified carbonate geothermal reservoir.

The basin fill along the profiles is divided into two megasequences by the base Pannonian unconformity (Figs. 8b and 9). This is a remarkable unconformity in the large part of the Western Pannonian basin, because Sarmatian strata are either strongly reduced (less than 100 m in the Danube basin) or completely missing. Lack of Sarmatian in the Tarany subbasin and the Drava trough has been explained in terms of regional post-Sarmatian basin inversion and erosion (Saftić et al., 2003; Nagymarosy and Hámor, 2012). It is inferred, that this unconformity assigns also the synrift/postrift boundary in the western part of the Pannonian basin. It can be seen in Figs. 8a and 9 that this is a reasonable division as the *Endrőd Formation* immediately overlying the unconformity does not show syntectonic growth faulting, as opposed to the underlying Early and Middle Miocene strata. Synrift beds have accumulated in transtensional half grabens, where they reach a thickness of 500 to 800 metres. Elsewhere, synrift strata cover the basement rocks as a drape in a thickness of less than 300 metres. The Inke high in the middle of Fig. 8a is an apparent exception, showing thick synrift beds in an elevated position due to post-Sarmatian inversion. The Ozora trough at the eastern termination of profile 2 (Fig. 9) offers a nice example of a well-developed fault-bounded trough with up to 2000 metres of synrift strata. Such features are relatively rare, and only a few wells penetrated such a thick complex, showing the complete sequence of synrift strata as summarized in Fig. 6. The oldest sedimentary rocks indicating the initiation of rifting is the fluviatile to lacustrine *Szászvár Formation* with intercalations of the products of widespread rhyolitic volcanic activity with radiometric ages of 21 to 17 Ma (*Gyulakeszi Formation*, Fig. 6). The first marine transgression on terrestrial beds occurred at the beginning of the Karpatian with deposition of the Tekeres schlier, followed by the open marine Badenian claymarls. Very thick synrift strata can be observed at places connected to Ottnangian and Badenian volcanic centres. Dacites and rhyolites, but mostly explosive tuffs and ignimbrites created massive accumulations with a thickness of more than 2 km, like in the middle part of profile 2 (Fig. 9). Similar features can be observed at other places in the MHFZ further to the East. Understandably, there are very few wells (like Nszk-1 in Fig. 9) which deeply penetrated into volcanic masses. As a matter of fact, precise delineation of volcanic masses is quite difficult, but seismic interpretation can be assisted by gravity and magnetic anomaly data. Interpretation of postrift sedimentary rocks can be performed in a more detailed way and all the defined lithostratigraphic units identified. The base Pannonian unconformity usually appears with high amplitude. The overlying *Endrőd Formation* is characterized usually by low amplitude reflectors, because it is practically a homogenous marl. Going upwards, strong amplitude reflectors follow due to the large acoustic impedance contrast between the sandy and marly beds of the turbiditic *Szolnok Formation*. The *Algyő Formation* is an unconformably dipping unit. It is composed of generally low-amplitude sigmoidal reflections (clinoforms), showing slope progradation. Most clinoforms have bottomsets downlapping at the base of this unit. The topsets of the clinoforms can be followed in the overlying unit, which represents the deposits of the shelf plain (*Újfalu Formation)*. A sharp break in the topset reflectors corresponds to the location of the shelf edge. Timing of shelf edge progradation from the northwestern periphery of the Pannonian basin (Vienna basin) towards the Serbian basin around Belgrade in the southeast (Fig. 1) has been derived from integrated stratigraphy (Magyar et al., 2013) and shown on our profiles by a series of red dots with ages (Figs. 8a, 9 and 10). The deposits of the originally flat sedimentary shelf appear as tilted reflections, but nearly parallel with each other. The lowermost horizons of *Újfalu Formation* generally have high amplitude and strong continuity, showing the presence of thick, laterally extensive sand sheets intercalating with thin marly beds. This is the thermal water-bearing layer, the optimal object of thermal water exploration in porous sedimentary rocks. Reflections in the overlying part of the *Újfalu Formation* are characterized by medium to low amplitude and medium to weak continuity, indicating a more heterogeneous mixture of sand and clay beds deposited on the coastal to alluvial plain of the Lake Pannon. The actual geometry of the different lithostratigraphic units as compared to the original depositional geometry (Fig. 7) demonstrates remarkable neotectonic deformations. This is most spectacularly shown by the Mio/Pliocene unconformity (a suspect Messinian signal in the Pannonian basin) indicated by broken red line in profiles 1 and 3. Identification of this unconformity is the result of wide scale correlation using a large seismic and well log data base, as well as magneto- and biostratigraphic informations (Magyar and Sztanó 2007, Sztanó et al., 2013). It can be seen in Fig. 8a that the unconformity is truncated at about the Inke high and towards the South it merges into the Drava trough, slightly unconformably following shelf plain reflectors of the *Újfalu Formation*. At the shelf edge with an age of 5.7 Ma it enters into the Algyő Formation. This geometry clearly suggests a large differential vertical movement during the Pliocene-Quaternary between the emerging and eroding Transdanubian Range and the Drava trough with ongoing subsidence and sediment accumulation. If the compaction and sediment loading corrections are taken into account, one can arrive at about 600 m tectonic subsidence in the axial part of the Drava basin. It is compared to some 500 to 600 m of uplift of the Transdanubian range, one can calculates a 1000 m scale neotectonic differential movements. Furthermore, one can imagine that before the onset of this differential movement the Pannonian basin was a large alluvial plain with flat morphology (just like the present day Great Hungarian Plain, Fig. 1), and a reduced Lake Pannon at the southern periphery, where the formation of the Paludina Lake was in progress (Müller et al., 1999).

2.3.2. *Profile 3*

Profile 3 is located in the eastern part of the Pannonian basin and passes from the Szolnok flysch zone to SW across the Tisza terrane (Fig. 10). The Tisza terrane is covered by 2 to 6 km thick fill of the Pannonian basin here in the Great Hungarian Plain, and all the knowledge comes from drillhole and seismic data. Outcrops are only available at its western edge in the Mecsek, Villány and Slavonian Mts. and eastern edge in the Apuseni Mts. (Fig. 1). Correlation of the outcrops and subcrops leads to the following division of the Tisza terrane from the North to the South, i.e. the lower towards the upper nappes in the tectonic hierarchy (Csontos and Vörös, 2004): Mecsek, Villány-Bihor and Békés-Codru subunits (Fig. 4). Formation of this nappe pile and overthrust by the East Vardar ophiolites took place during a series of tectonic events including Late Jurassic obduction of ophiolites and subsequent mid-Cretaceous compressional events, with a paroxysm of W to NW vergent thrusting in the Turonian (Schmid et al., 2008). We call the attention to the fact that the structure of this nappe system below the Pannonian basin is poorly known due to the limitation of seismic resolution and shallow penetration of boreholes. Therefore, the basement structure shown in Fig. 10 (just like in Figs. 8a and 9) is highly schematic and speculative. Similar profiles can be found in Tari et al. (1999a,b) and Schmid et al. (2008). The Szolnok flysch on the northwestern edge of profile 3 is a few kilometre thick and strongly deformed complex, composed of rhythmically alternating shale and fine to coarse grained sandstone layers with turbiditic character. Its age ranges from the Campanian to the Oligocene and most widespread are the Middle and Late Eocene sequence (Szederkényi et al., 2012). The flysch zone stretches from the central part of the Great Hungarian Plain to the Maramures area of the internal East Carpathians, where it is exposed in association with the easternmost occurrences of the Pieniny Klippen belt that is attributed to the Piemont-Liguria ocean (Fig. 2b). Accordingly, the Szolnok flysch zone is considered to represent the continuation of the of the Sava zone, the suture between the Alcapa and Tisza-Dacia terranes (Schmid et al., 2008; Tischler et al., 2007). As a matter of fact, this is difficult to verify by direct observations from the Pannonian basin, because available drillhole data show that the Szolnok flysch is underlain by the rocks of the Mecsek zone, which is the lowermost tectonic unit of the Tisza terrane (Fig. 10). Furthermore, Alcapa most probably overthrusts Tisza–Dacia in this part of the contact zone, just like in the Maramures area (Csontos and Nagymarosy, 1998; Tischler et al., 2007). In other words, the ophiolite-bearing suture between the Alcapa and Tisza terranes (i.e. Sava zone proper) should have been completely consumed and overprinted by multiphase tectonism (compression, dextral and sinistral wrenching) characteristic for the Mid-Hungarian Fault Zone. The Szolnok flysch is tectonically separated to the South from a block where the Mesozoic cover sequence of the Mecsek unit is overthrust by the crystalline nappe of the Villány-Bihor zone (Fig. 10). Further to the South, the basement of the Pannonian basin is made up from this medium grade Variscan metamorphic complex in a width of about 50 km. It consists of gneisses, micaschists and granites (Szederkényi et al., 2012). Thermochronological studies in the Apuseni Mts. show an Alpine overprint related to mid-Cretaceous nappe stacking (Merten et al., 2011; Kounov and Schmid, 2013). Fast cooling, rapid exhumation and massive erosion of the uplifting footwall strata took place in the latest Cretaceous to Middle Eocene time. Extension related to the formation of the Pannonian basin did not induce significant regional exhumation in the Apuseni Mts. In the middle of Profile 3 the crystalline complex is overthrust by a nappe composed from the Permian through Mesozoic cover sequence of the Villány-Bihor subunit. Further to the South this Mesozoic sequence is overthrust by a nappe consisting of the Variscan basement of the Békés-Codru subunit (Fig. 10). The Permian-Middle Triassic cover sequences display strong similarities in the different subunits of the Tisza terrane, with a transitional pattern of Triassic facies from the external Mecsek towards the internal Algyő-Biharia subunit (Haas and Péró, 2004). Permian and Early Triassic red sandstones and conglomerates, together with products of rhyolitic volcanism indicate terrestrial conditions. Marine transgression started in the Middle Triassic and massive carbonate beds formed in shallow ramp and platform environments. Towards the end of Triassic terrestrial-lagoonal environment was reestablished in the Mecsek subunit and a thick coal-bearing sequence developed. In more internal zones pelagic limestones and other basinal facies prevailed. Rifting and opening of the Neothetyan oceanic basins from the end of Triassic (Fig. 2a and b) led to a dramatic change in depositional environment from the Middle Jurassic showing that Tisza became part of the Adriatic paleogeographic realm (Schmid et al., 2008). Longlasting rifting process led to basaltic volcanism with a peak period in the Early Cretaceous. Compressional phases at the Adriatic margin started in the Aptian and culminated in the Turonian. Postorogenic uplift and erosion produced coarse clastics in the Late Cretaceous and Gosau-type basins seal the thrust faults in the Tisza terrane. The basin fill along profile 3 is quite compatible with that of the profile 1. The base Pannonian unconformity can be well recognized and the underlying Middle and Lower Miocene strata attain a thickness of up to 1.2 km in the extensional half-grabens of the Békés basin (Fig. 10). The shape of the Derecske trough in section and map view (Fig. 5b) offers a nice example of a pull-apart basin. It is an interesting feature of the profile that the synritf/postrift boundary appears to be younger than in the western Pannonian basin, because the lower Pannonian *Endrőd Formation* exhibits features of late rift climax strata (Prosser, 1993). However, initiation of rifting is about the same as the synrift sedimentation starts with the fluvial to lacustrine deposits of the *Szászvár Formation* (Nagymarosy and Hámor, 2012). This is overlain by different sedimentary rocks ranging from terrestrial to marine Karpatian schlier and mostly Badenian marls. Sarmatian beds are missing in the northern part of the section and in the wider area, but thin (up to 30 metres) brackish marls, locally biogenic limestones and sandstones were encountered in the Békés basin (Szentgyörgyi and Teleki, 1994). A post-Sarmatian basin inversion and erosion is a plausible mechanism to explain these findings, as well as the remarkable vertical offset of pre-Pannonian strata along profile 3. In addition to similarities between the western and eastern part of the Pannonian basin, there is a sharp contrast in volcanic activity. Along profile 3 and in its immediate surroundings hardly any pyroclastics and lava flows were encountered in the Miocene strata. It is quite interesting, however, that further to the East, in the non-extensional Transylvanian basin Lower Badenian and Lower Sarmatian rhyolitic tuff layers are of general extent, even if their thickness does not exceed 50 metres (De Leeuw et al., 2013). The Lake Pannon megasequence in Profile 3 exhibits the Pannonian lithostratigraphic units which are younging towards the South due to progradation (Juhász et al., 2007). Again, the geometry of the Miocene/Pliocene boundary bears important information on the amount of Pliocene-Quaternary differential vertical movements which is larger by a few hundred metres here in the Békés than the Drava trough (Figs. 8a and 10).

2.4. Tectonic ideas and models

The pioneering work of Stegena et al. (1975) first explained the evolution of the Pannonian basin in terms of subduction along the Carpathian arc and mantle diapirism resulting in crustal attenuation, subsidence and volcanism. Lexa and Konecny (1974) argued that volcanism of the region should be divided into two main types. Calc-alkaline volcanism associated with the East Carpathian arc (Fig. 5b) exhibited geochemical features of subduction generated melts in a mantle wedge. Its age is 16 Ma (Upper Badenian) in the NW and progressively younging towards the SE, where the relict slab in the Vrancea region is just breaking off. The areally distributed volcanism in the Pannonian basin was related to lithospheric attenuation and mantle upwellings resulting in crustal anatexis and eruption of dominantly felsic magmas during the time interval of 21 to 11 Ma. This simple model and its more advanced version (Konecny et al., 2002) are widely accepted as they represent a straightforward interpretation of petrological and geochemical data. More recent geochemical studies have made attempts to refine this picture, however their results are contrasting, which reflect the complexity of chemical geodynamics (Harangi and Lenkey, 2007; Kovács and Szabó, 2008; Seghedi and Downes, 2011). The first quatitative model for the evolution of extensional sedimentary basins was suggested by McKenzie

(1978). His thermomechanical model assumes the mechanism of pure shear for the lithospheric deformation which occurs rapidly and is characterized by a single stretching parameter. The initial phase of isostatic subsidence is followed by a long period of slow subsidence due to thermal relaxation. An analysis by Sclater et al. (1980) demonstrated the inadequacy of this uniform lithospheric stretching to explain the subsidence and thermal evolution history of the Pannonian basin. Namely, the Pannonian basin was too hot and the postrift subsidence too large to arrive at a reasonable fit between predicted and observed subsidence histories. More heat was required to introduce into the system and it was possible by a non-uniform model of lithospheric deformation. It was inferred that moderate crustal extension was accompanied by large attenuation of the mantle lithosphere during the synrift phase. The vital role of slab rollback in generation of the backarc extension in the Pannonian region and related upper mantle flows were also envisaged in this evolutionary scenario. Formation of the non-extensional Transylvanian basin was explained by crustal downwarping due to pull of the retreating slab. It was thought that after slab breakoff crustal rebound elevated Transylvania by 600–700 m above sea level (Royden et al., 1982, 1983). The non-uniform stretching model was successfully applied to explain a wide range of observations in the Pannonian basin (Royden and Horváth, 1988). Particularly important were the subsidence, thermal and maturity history simulations using data from a number of hydrocarbon exploration wells in Hungary (Szalay, 1988; Horváth et al., 1988). It has become well constrained that a major heat impact during the synrift phase was indeed required in the Pannonian basin to explain the present high maturity level of organic matter and heat flow. However, the mechanism which resulted in major attenuation of lithospheric mantle and transfer of extra heat into the system remained obscure. A new input came from Alpine geology, when Ratschbacher et al. (1991a,b) presented a novel idea on the extrusion tectonics in the Eastern Alps. They defined extrusion as an interaction of extensional collapse due to gravity forces and lateral motion (escape) of an orogenic wedge by forces applied to wedge edges. Boundary conditions for extrusion in the Eastern Alps were given by transpression between the Adriatic and European plates, a gravitationally unstable (overthickenned) orogenic wedge, and a free passage towards the Carpathian embayment with subductible lithosphere. In addition to drafting the strike-slip fault system in the Eastern Alps facilitating the eastward escape of the Alcapa terrane, these authors took a stand in the issue of the vertical dimension of the orogenic wedge. They argued that Alcapa terrane was a crustal flake, composed from a pile of Austroalpine nappes detached from its ductile Penninic substrata. Most recent studies agree with this concept, however different detachment horizons were specified at middle to lower crustal levels above the rheologically strong upper mantle (Frisch et al., 2000; Kummerow et al., 2004; Schmid et al., 2013). Anyhow, this concept suggests that in the course of escape crustal wedge overrides directly the hot asthenosphere emerging in the hanging wall of the retreating slab (Horváth et al., 2006). This process can offer the required heat impact and crustal melting during the synrift evolution of Alcapa. Similar model can be applied to Mid-Hungarian Fault Zone, where felsic volcanism was the most intensive. However, the case of the composite Tisza-Dacia terrane appears to be more complex and possible detachment from its mantle lithosphere needs supporting data. Anyhow, if the block model in Fig. 3 is analyzed from a simple kinematic point of view, one can imagine that the complex interaction of terranes in the Carpathian embayment was a crustal scale process rather than interaction of lithospheric blocks. To quantify lithospheric deformations Lenkey (1999) calculated the variation of crustal thinning factor for the Pannonian basin. This has been used recently by Ustaszewski et al. (2008) for palinspastic restoration of the Alcapa and Tisza-Dacia terranes before the onset of rifting in Early Miocene (Fig. 3). They have shown that the eastern edge of Alcapa was located west of the present-day Budapest and affected by 290 km extension and coeval counterclockwise rotation. Tisza-Dacia experienced 180 km extension during clockwise rotation into the Carpathian embayment around a pole in western Moesia. This kinematic history and juxtaposition of the two terranes were associated with dramatic volcanism and deformation of the intervening Sava zone and correlated units (e.g. Bükk, Szolnok flysch) including significant shortening by compressional events, large scale wrenching with changing sense of displacements in space and time, and extension (Figs. 3 and 4). This complex structural evolution also speaks for thin-skinned (crustal) tectonics. The issue of synrift evolution was addressed by quantitative subsidence analyses incorporating the concept of necking depth and finite strength of the lithosphere during and after rifting (Lankreijer et al., 1995, 1997; van Balen et al., 1999). Furthermore, Huismans et al. (2001) performed 2D thermo-mechanical finite element modelling, using pressure and temperature dependent visco-elasto-plastic rheology. A first phase of passive rifting due to extensional stresses generated by slab rollback was followed by an active mantle lithosphere thinning as a result of buoyancy induced asthenospheric uprise beneath the rift. Model results offered a good description of the extension in the Pannonian basin and coeval compression in the neighbouring orogens. The important contribution of asthenospheric flow to the tectonic evolution of the Pannonian basin received a new dimension when it was viewed from a Mediterranean perspective. Horváth and Faccenna (2011) put forward the idea that asthenospheric inflow from below the Adriatic region through the northern Dinaric slab window played an important role in controlling the rollback of the Carpathian slab, and driving the opposite rotation and extension of the Alcapa and Tisza-Dacia terranes. An alternative source of asthenospheric inflow can be the East Alpine upper mantle as is indicated by SKS anisotropy directions (Kovács et al., 2012; Bokelmann et al., 2013). Upper mantle flow system has been shown to be responsible for generating dynamic topography in the Mediterranean region (Faccenna and Becker, 2010; Faccenna et al., 2014). The static component of the topography relative to a reference level can be calculated by the assumption that a lithospheric column floats freely within the asthenosphere. The difference between the actual and calculated topography in the Pannonian basin turns out to be a robust feature with values as high as 1000 m. This residual topography is thought to be a dynamic feature and explained in terms of instantaneous mantle flow due to temperature anomalies as inferred from regional P and S wave tomography (Fig. 11). Dynamic topography is derived from the radial tractions acting upon a free-slip surface boundary in a Newtonian-type fluid (Becker et al., 2014). Results show a remarkably good fit between dynamic and residual topography pattern suggesting a marked convective support of the over-elevated Pannonian basin (Horváth et al., 2014). The Carpathian-Pannonian system shows a remarkable variation of thermo-mechanical properties. Results of rheology calculations can be expressed in terms of effective elastic thickness (EET) by integrating the thickness of the mechanically strong layers of the lithosphere. Lankreijer (1998) constructed a map which shows very low EET values for the weak central part of the Pannonian basin (5–10 km), and an increase towards the surrounding orogens (15–30 km). Such a system reacts very sensitively for changes of intraplate stress and deformations of different wavelengths can be predicted (Cloetingh and Kooi, 1992). Differential movements during the Pliocene and Quaternary and their spatial distribution in the Pannonian basin have been interpreted as a consequence of stress field change from extension to compression (Horváth and Cloetingh, 1996). Modelling results predicted quite well the observed spatial pattern, but the amplitude of stress induced deflections was much smaller than the observed thousand metre scale movements (Figs. 8a and 10). More

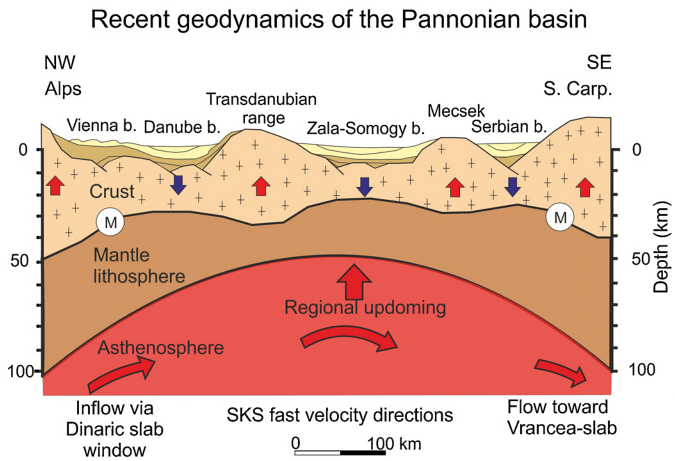

Fig. 11. Sketch to show the neotectonic deformation of the Pannonian basin. The attenuated crust is under compression, which generates differential vertical movements. Lithosphere is thin and asthenospheric flow system sustains it in an elevated position relative to its isostatic equilibrium position.

sophisticated numerical simulations included flexural isostasy and necking of the lithosphere during extension (van Balen et al., 1999), and viscoelastoplastic rheology (Jarosinski et al., 2011) have led to better predictions, but still far from a reasonably good fit with observations. Dombrádi et al. (2010) addressed this problem by analogue modelings examining the folding mechanism of the hot Pannonian lithosphere characterized by very low strength apart from a thin layer of brittle upper crust. They arrived at the conclusion that this extremely weak lithosphere and crustal thickness variations are the key factors governing the regional deformation pattern and influence the timing and extent of basin inversion (Fig. 11).

Finally, we pay attention to a most recent attempt to explain in a unified way the evolution of the Pannonian and Transylvanian basins (Matenco and Radivojevic, 2012; Tilita et al., 2013). These authors have presented detailed studies of the Serbian part of the Pannonian basin (Backa and Banat basins in Fig. 1) and Transylvanian basin and shown the strong connections and parallelism in their evolution. They recall the early idea of Tari et al. (1999a,b) who tested the applicability of the simple shear lithospheric deformation mechanism of Wernicke (1985) to the western part of the Pannonian basin. The point of the simple shear mechanism lies in the asymmetry of the developed structure: region of crustal extension and region of lithospheric thinning are separated by an offset on the scale of lithospheric thickness. This separation involves that the region of crustal extension and subsidence is not associated with increased heat flow, while the area with increased heat flow is not subsiding but uplifting (Wernicke, 1985). Tari et al. (1999a,b) application was reasonable as the Danube basin and the neighbouring Transdanubian Range exhibited these asymmetric characteristics. Application of a simple shear, which is in the crust beneath the Pannonian basin and truncates the mantle lithosphere under the Transylvanian basin is a more difficult exercise. Indeed, a two phase evolutionary scenario is envisaged, starting with a simple shear and followed by a pure shear lithospheric deformation (Tilita et al., 2013). In the Early and Middle Miocene during the retreat of the Carpathian slab, there was crustal extension and subsidence in the Pannonian region, and lower lithospheric stretching in the Transylvanian region resulting in uplift and erosion. From the Late Miocene, the whole lithosphere of the Pannonian basin extended in a uniform way, while the Transylvanian basin passed into a phase of thermal sag development.

3. Geothermics and regional water flow systems

3.1. Temperature and heat flow data

The temperature database of Hungary contains more than 14 000 reliable data derived from measurements in about 4800 boreholes (Dövényi and Horváth, 1988; Dövényi et al., 2002; Horváth et al., 2012). Drill-stem test results in hydrocarbon exploration wells and logs from stabilized wells were accepted as stationary temperature values, while bottom hole and outflowing water temperatures were properly corrected. Different kinds of isothermal maps were constructed in Hungary and fitted into a European framework (Dövényi et al., 2002). Fig. 12a shows average temperature versus depth diagram for the central (Hungarian) part of the Pannonian basin, as well as the range of temperature profiles. Average temperature gradients calculated for every 250 m depth intervals can also be seen. As a brief summary of the geothermal conditions of the Pannonian basin one can conclude that 100 °C and 200 °C temperatures can be reached in a depth of 1900 m and 5000 m (resp.), and the temperature gradient varies between 40 and 50 mK/m. More adequate characterization of the geothermal conditions can be received by preparation of terrestrial heat flow maps. Primary heat flow determinations in the central Pannonian basin have been made in 28 deep boreholes with reliable temperature data at different depths and access to core samples for laboratory determination of the thermal conductivity. Several hundred measured data on sedimentary rocks allowed the construction of conductivity versus depth diagrams for coarse-grained and fine-grained rocks (Dövényi and Horváth, 1988). These main lithologies in the basin fill can be determined from well logs and combined with the diagrams and temperature data, heat flow can be easily calculated. This determination was performed in 120 boreholes to arrive at a good heat flow data base of Hungary. This has been completed by data from the neighbouring countries and a heat flow map prepared for the Pannonian basin and the wider surroundings (Tari et al., 1999a,b; Dövényi et al., 2002).

Heat flow can be disturbed by fast sedimentation and erosion, and water flow systems. The thermal effect of sedimentation was calculated by a one-dimensional numerical model, which took into account the variation in the sedimentation rate and the change in the thermal properties of sediments due to compaction (Lenkey, 1999). The heat flow map corrected for sedimentation is shown in Fig. 12b. Data coverage is quite good for the whole area except the Apuseni Mountains and the Eastern Alps, where only few measurements were available. The heat flow distribution in the Pannonian basin shows values ranging from 50 to 130 mW/m^2, with a mean value of about 100 mW/m^2. The average heat flow in the basin is considerably higher than in the surrounding regions. The Ukrainian and Moesian Platforms are characterized by low heat flow values of 40–50 mW/m^2, which are typical for the stable continental crust. The Carpathians and the Bohemian Massif show heat flow values of 50–70 mW/m^2, which are close to the worldwide mean for continental crust. The Outer Dinarides exhibit extremely low heat flow (about 30 mW/m^2) due to cooling by meteoric water inflowing at the high karst plateau. The peripheral Vienna basin is characterized by low (50–70 mW/m^2) and the Danube basin fairly high values (80–90 mW/m^2). The Transylvanian basin is remarkably cold except its eastern periphery, where the volcanic activity ceased in the Pleistocene and a local asthenospheric uprise can be present (Tilita et al., 2013). The hottest part of the Pannonian basin with heat flow values above 100 mW/m^2 are the East Slovakian basin and the Great Hungarian Plain including its continuation into the Serbian basins and the Vardar zone. The Transdanubian Central Range and the Hungarian mountains in the NE (e.g. Bükk and Aggtelek-Gemer), are characterized by low heat flow (50–60 mW/m^2). Just like the Outer Dinarides, these mountains are built up mostly from

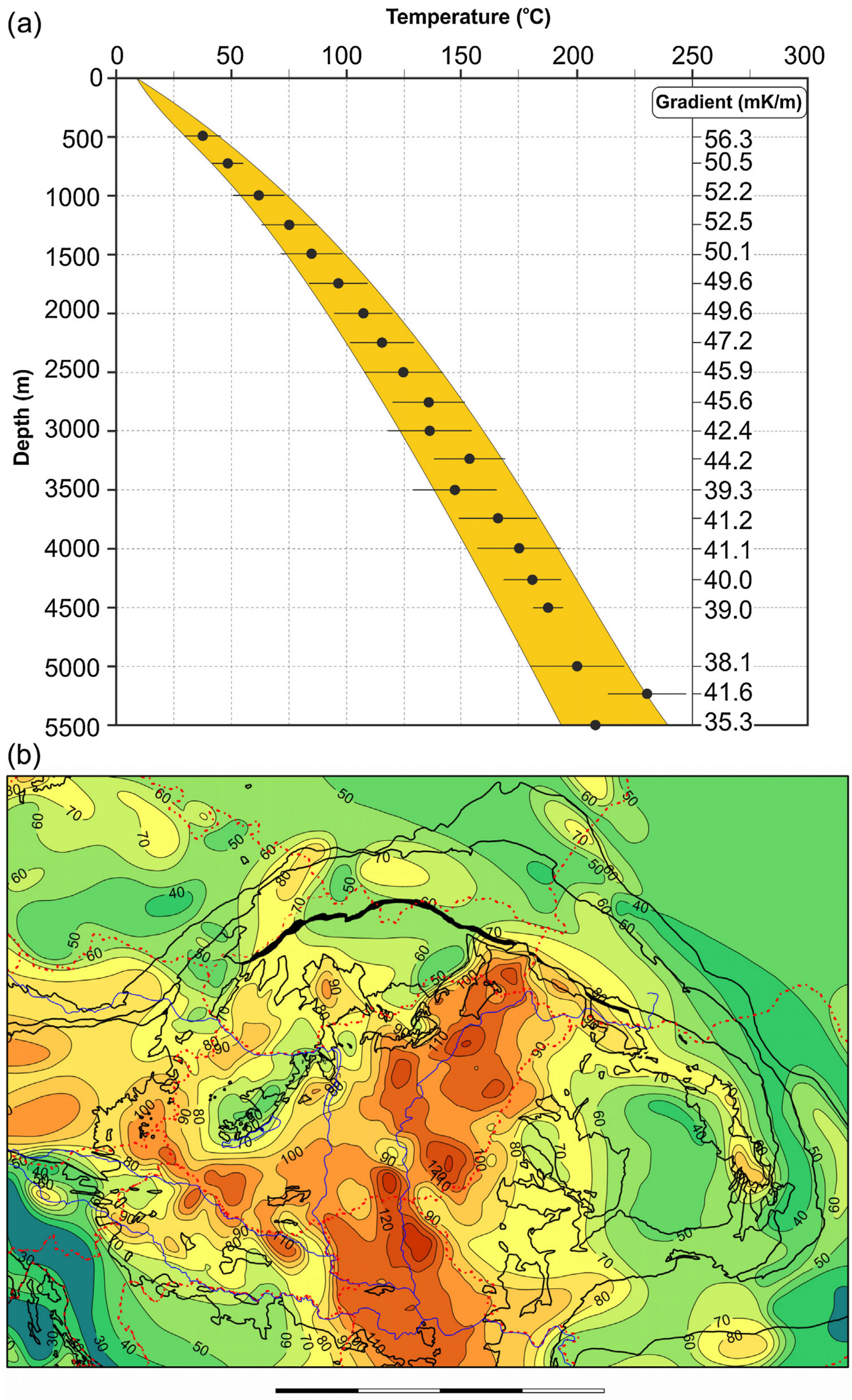

Fig. 12. (a) Average temperature versus depth profile in the central part of the Pannonian basin (modified after Dövényi and Horváth, 1988). (b) Heat flow map of the Pannonian basin and surrounding regions (heat flow values in mW/m^2), corrected for (where appropriate) the cooling effect of fast sedimentation (Lenkey et al., 2002).

Mesozoic carbonates, which are exposed at the surface. These fractured and karstified rocks have large permeability, which allows infiltration of the cold meteoric water. It is heated up at depth and can return to the surface along faults at the feet of the mountains in thermal springs. The total thermal energy output of the thermal springs can be calculated giving an estimate of the convective heat transfer. Adding this value to the observed heat flow we obtain the "undisturbed" heat flow of these karstic mountains, which is in the range of 80–100 mW/m^2 in Hungary (Lenkey et al., 2002). Finally, we note that it is worthwhile to compare the first order features

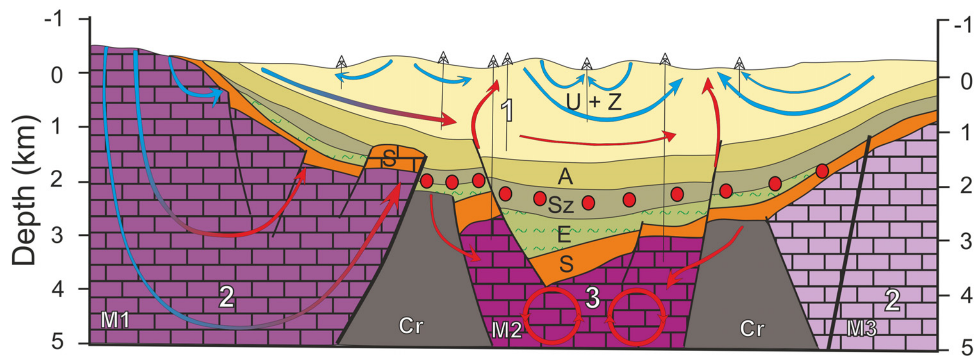

Fig. 13. Generalized model of the Pannonian basin showing the main water flow systems and the related geothermal resources in a N-S oriented section. Blue to red arrows illustrates cold to warm water. 1 and 2 indicate gravity-driven flow system in the porous basin fill rocks, and Mesozoic (locally Eocene and/or Miocene) carbonates, resp. 3 indicates an overpressured system below a pressure seal (indicated by red ellipses) including lower Pannonian and Early to Middle Miocene basin fill, and fractured basement rocks.
U + Z = Újfalu and Zagyva; A = Algyő; Sz = Szolnok; E = Endrőd; S = Synrift; M1,2,3 = Mesozoic rocks belonging to the Transdanubian Range unit, the MHFZ and Villány - Mecsek unit; Cr = Crystalline basement.

of the crustal thickness, the basement depth and heat flow maps (Figs. 5a, b and 12b, resp.). The good correlation of the amount of crustal attenuation, basement subsidence and heat flow highs suggests that pure shear is a reasonable mechanism to describe lithospheric deformations in the Pannonian basin.

3.2. *Regional water flow systems*

Two superimposed hydraulic systems can be distinguished in the Pannonian basin representing an upper and a lower domain (Tóth and Almási, 2001). In the upper domain a gravity-driven water flow system exists, which is regionally unconfined, hydrostatically pressured and recharged from precipitation. Such a system prevails in the porous sedimentary rocks of the basin fill from the surface down to a depth of about 2000 m. In addition, exposed older permeable rocks (mostly Mesozoic carbonates) and their subsurface continuations represent another type of gravity-driven flow system in the Pannonian basin (Fig. 13).

The lower hydraulic domain is a regionally confined system characterized by remarkable overpressures below a low permeability pressure seal and contains highly saline waters expelled from sedimentary rocks during compaction. Generally, the *Endrőd Formation* represents the upper lid of this domain, which includes the underlying Early and Middle Miocene synrift strata, as well as fractured Mesozoic and crystalline rocks (Fig. 13). The two systems are spatially separated but not isolated, because pressure dissipation can take place through the imperfect pressure seal conducting fluids from the lower to the upper aquifer system mostly along neotectonic faults (Mádl-Szőnyi and Tóth, 2009; Czauner and Mádl-Szőnyi, 2011).

3.2.1. *Gravity-driven flow system in porous basin fill*

In regions where the groundwater table is conform to the land surface and the rock framework is hydraulically continuous, flow is directed downward in areas of high elevation (recharge area), upward in areas of low elevation (discharge area), and laterally in regions of intermediate elevation (midline area). Groundwater flow is spatially distributed in systems of different hierarchical order (local, intermediate, regional) as a function of the complexity of the relief (Tóth, 1963). In the eastern part of the Pannonian basin (Great Hungarian Plain) the relief is very smooth: local variations with less than 10 m are superimposed on a subbasin scale topography with an amplitude of 50 m (Fig. 1). Local flow systems is the case when the streamlines at shallow depth connect directly adjacent recharge and discharge areas (Fig. 13). A relatively deep penetrating regional flow system develops between the major water divide and a distant discharge area. Regions of inflow and outflow are characterized by small deviation of fluid pressure gradient from the normal hydrostatic gradient. Subhydrostatic gradient is associated with descending water flow and superhydrostatic gradient characterizes regions of water ascent. The porous basin fill in the Pannonian basin is composed mostly of clastic rocks of different grain size (sandstones, marls and siltstones). The porosity and permeability decrease with depth due to compaction as is shown in Fig. 14a and b.

Porosity and permeability of sedimentary strata depend primarily on depth and lithological composition. It is most convenient to accept the well-defined lithostratigraphic units as hydrostratigraphic units and give their hydrological characterization (Table 1).

The *Újfalu* and *Zagyva* Formations together constitute the largest hydrostratigraphic unit in the Pannonian basin with general extent and a thickness of up to 2 km as illustrated by profiles 1 to 3 (Figs. 8a, 9 and 10). This is made up of unconsolidated sediments deposited on a large alluvial plain and is characterized by large to very large permeability values, with limited interbeddings of poorly permeable muds and other fine-grained beds. It represents a major aquifer utilized for drinking and communal water supply of several million people. Furthermore, the deeper part of the aquifer, particularly the layer directly overlying the low permeability *Algyő Formation* is an optimal source of thermal water exploitation. Fig. 13 shows a schematic flow model of the Pannonian basin. The *Algyő Formation* is the progradational and aggradational unit deposited on the shelf slope with lithologies dominated by poorly- to well-consolidated marls and siltstones, with frequent interbeddings of more sandy mouth bar and channel fill sequences. As a whole it can be considered as a leaky aquitard with permeabilities in the range

Table 1
Characteristic permeabilities of main hydrostratigraphic units in the Pannonian basin (after Tóth and Almási, 2001).

Hydrostratigraphic unit	Permeability range (mD)	Classification
Újfalu and Zagyva Fm.	100–500	Aquifer
Algyő Fm.	1–10	Aquitard
Szolnok Fm.	10–100	Aquifer
Endrőd Fm.	0.1–1	Aquitard
Early and Middle Miocene clastics	10–100	Aquifer
Early and Middle Miocene tuffs	10–50	Aquifer
Carbonates (fractured and karstified)	50–500	Aquifer
Crystalline (fracture)	10–100	Aquifer
Crystalline (intact)	0.001–0.1	Aquitard

1 mD = 10^{-15} m^2, which corresponds to 10^{-8} m/s conductivity.

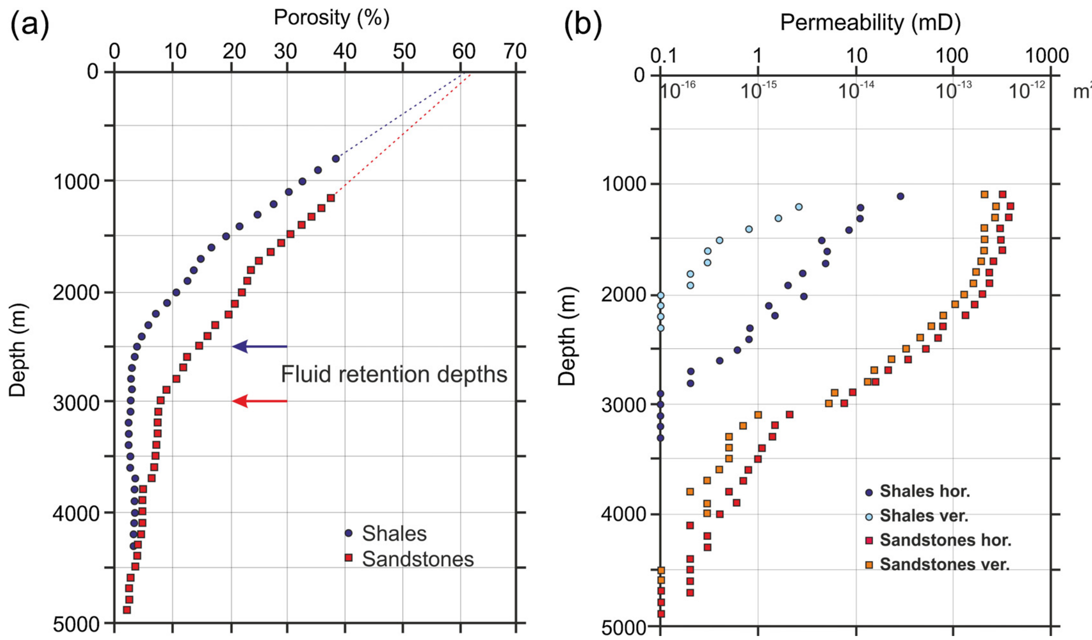

Fig. 14. (a) Results of porosity determinations from well logs for shales and sandstones (modified after Spencer et al., 1994). From about 2500 to 3000 m depth disequilibrium compaction leads to isolation of pores and fluid retention. (b) Horizontal and vertical permeability of shales and sandstones determined on rock samples in laboratory (modified after Spencer et al., 1994).

of 1–10 mD. The *Szolnok Formation* is made up of a cyclic alternation of consolidated sandstones and siltstones of turbiditic origin. The laterally distinct turbiditic bodies can be covered by a continuous drape of marls, which results in a hydrostatic unit with moderate permeability. Its thickness can be as large as 1.2 km in the deep basins, which is thinning towards basement highs, or even pinching out on its flank (Fig. 8a and 10).

3.2.2. Gravity-driven flow system in carbonate rock aquifers

We have already seen in Fig. 12b that large exposed carbonate plateaus exhibit decreased heat flow values, because karstification creates large porosity and permeability to facilitate infiltration of cold meteoric water. Karstification is an epigenic process of calcite dissolution in the course of infiltration of water including CO_2 from the atmosphere and soil. There are additional processes summarized under the term of “hypogenic speleogenesis”, which lead to the development of the void system at depth (Goldscheider et al., 2010). This is usually combined with extensive development of fracture system during tectogenesis and, as a consequence, exposed carbonates with a large subsurface extent (Fig. 13) represent the second largest thermal resources in the world after regions of active volcanoes. Water circulation is gravity-driven caused by the topography gradient and also hierarchical as a response to variation of the relief. In addition to hydraulic head drive, ascent of hot water in areas of discharge is facilitated by temperature-induced decrease of density and viscosity (Goldscheider et al., 2010).

3.2.3. Overpressure-driven flow system

Fluid pressures in excess of hydrostatic are commonly observed in sedimentary basin with ongoing subsidence and sedimentation, as well as active hydrocarbon generation (Gordon and Flemings, 1998). Great attention has been paid to determine these pressures and explain their origin and evolution. There are a set of different mechanism that can create overpressure: i) disequilibrium compaction, ii) diagenetic reactions which release pore fluids, iii) aquathermal pressuring, iv) gas generation and v) tectonic processes. It has been demonstrated, that the most important mechanism is disequilibrium compaction, which occurs in sedimentary basins during rapid burial, when the pores are getting isolated at depth and the retained fluid overpressured (Osborne and Swarbrick, 1997). Overpressures are present in deeper parts of the Pannonian basin, including the Vienna basin and excluding the Transylvanian basin. Fig. 13a shows a simplified model of the overpressured hydraulic system in the Pannonian. Fig. 15a and b presents pore pressure data for the southwestern and the south-eastern part of the Pannonian basin derived from drill-stem tests in hydrocarbon exploration wells (modified after Tóth and Almási, 2001). Both diagrams shows the hydrostatic pressure conditions in the upper hydraulic system of the basin fill down to a depth of about 2200 m. In the southwestern Pannonian basin higher than normal pressure values appear with a pressure gradient slightly above normal between 2200 and 3000 m. The transition is taking place in synrift strata and Mesozoic carbonates. Further down, pressure values are more scattered, but obviously indicate the presence of a highly overpressured domain, with excess pressures of 20–30 MPa in the 3000–4500 m depth range (Fig. 15a). In the southeastern Pannonian basin the increase of pressures seems to be very sharp in the 2200–2900 m depth range (Fig. 15b).

This dramatic increase takes place in the Szolnok hidrostratigraphic unit and the whole Endrőd unit is already strongly overpressured. In the fractured Mesozoic basement 25–30 MPa excess pressures were measured in the depth range of 3000–4500 m. Fluid flow in this system is driven by the overpressure gradient, which has a maximum at the inflection point of the pressure curve. This defines a pressure seal separating the upper and lower flow system (Magara, 1975). Due to the high permeability, temperatures and pressures in the lower system, conditions are met to develop free convection systems (Lenkey, 2002). There are two schools of thoughts in explaining the origin of overpressures in the Pannonian basin. The majority of the researchers favours the disequilibrium compaction mechanism (Szalay, 1988; Spencer et al., 1994). They argue that porosity versus depth plots (Fig. 14a) show isolation of pore space in a depth of about 2000 m. Fast sedimentation and burial following the deposition of the Endrőd marl hampered its dewatering and generated overpressure. Numerical modelings support this idea (van Balen et al., 1999). The first notion about the

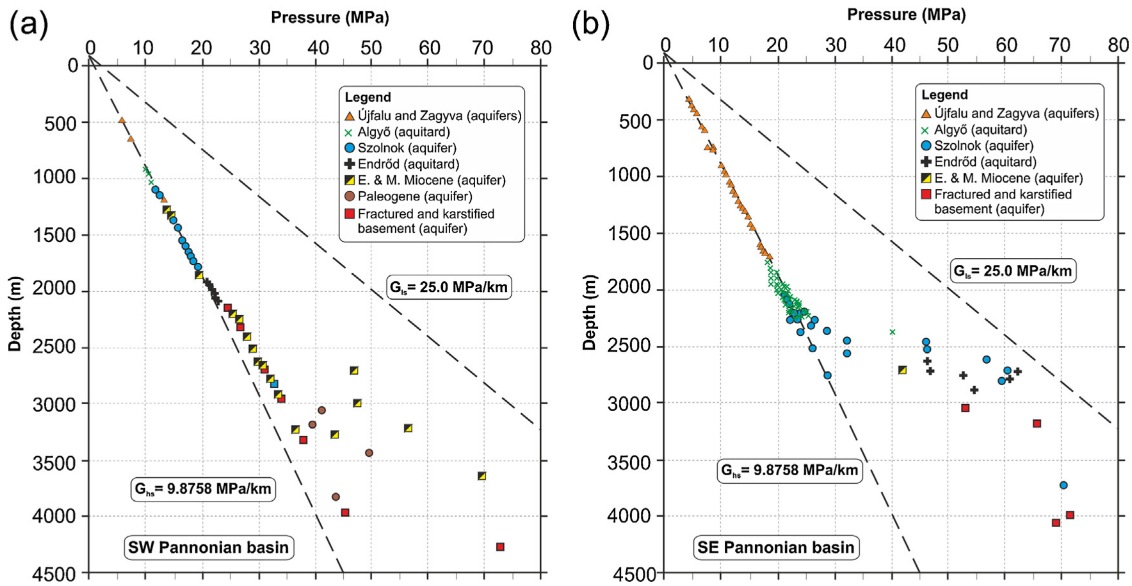

Fig. 15. (a) Change of pore pressure versus depth in the southwestern Pannonian basin and indication of the hydrostratigraphic units with symbols shown in the Figure. Broken lines give the increase of hydrostatic and the lithostatic pressure with depth. (b) Change of pore pressure versus depth in the southeastern Pannonian basin and indication of the hydrostratigraphic units with symbols shown in the Figure. Broken lines give the increase of hydrostatic and the lithostatic pressure with depth.

possible role of tectonics in generating overpressure was raised by Horváth (1995). There is set of overpressured hydrocarbon fields at the eastern margin of the Pannonian basin with reservoirs in the fractured crystalline basement of the Villány-Bihor zone (see the middle part of Fig. 10). They are located in an area uplifted by 500–800 m during the neotectonic inversion. It was suggested that these fields were originally hydrostatic and had a good seal, which retained their pressure during uplift, leading to apparent overpressure. Tóth and Almási (2001) went further in emphasizing the importance of the neotectonic change from extensional to compressional stress. They put forward the idea that this compressive stress is responsible for the generation of general overpressure in the Pannonian basin. Their main argument is based on a few pressure profiles they constructed to show that largest overpressures developed at the elevated parts of the basement, rather than in the deepest grabens, where it should be if disequilibrium compaction were the main overpressure generating mechanism. They draw the bold conclusion that fluid flow directed from the highs towards the lows of the basin. Their argumentation, however can be challenged, because of the obvious fluid flow directions in the 6 Ma to present time period: most of the hydrocarbon fields in the Pannonian basin are found at basement highs and sourced from mature source rocks in deeper troughs (Spencer et al., 1994; Tari and Horváth, 2006).

3.3. Geothermal resources

Fig. 16a and b presents new maps of geothermal installations producing thermal waters from porous Pannonian beds and Mesozoic carbonate reservoirs (resp.) in the Pannonian basin. The basin belongs to the territory of eight different countries and the reservoirs are crossing the political boundaries. Accordingly, their utilization is a joint scientific matter and responsibility. This has been the rationale behind the recently completed European Union geothermal project of "Transenergy" performed in the cooperation of Slovenia, Austria Hungary and Slovakia (http://transenergy-eu.geologie.ac.at/results). A previous outstanding cooperation has been the preparation of the "Atlas of geothermal resources in Europe" (Hurter and Haenel, 2002), which presents basic geothermal data country by country in Europe. They were our main sources of data completed with national reports presented at different conferences (Kolbah, 2010; Fendek and Fendekova, 2005; Jelić et al., 2005; Goldbrunner, 2005).

Fig. 16a shows 261 localities of geothermal installations all over the Pannonian basin producing thermal waters from the porous aquifer with yields typically of 200–300 m^3/day, and temperature in the range of 30–100 °C. More than one well belongs to each locality as the total number of active geothermal wells in the region is between 1200 and 1300. The main source of the hot water is regularly the 100–300 m thick thermal water-bearing layer at the bottom of the *Újfalu Formation*. Temperature of water depends on the depth of this layer: the deeper the layer the higher the temperature. In the Békes basin (Fig. 10) for example, the thermal water-bearing layer reaches a depth of 2000 m and yields water with temperature of 90–100 °C. It is to be emphasized that the porous reservoir represents one single and continuous geothermal resource laterally as well as vertically. Thermal water is utilized for bathing in swimming pools, space heating and balneological therapy in wellness hotels and hospitals. This makes about 40% of the total water. About 45% is used for space heating of special districts, and in agriculture for heating of green houses and other spaces (Szanyi et al., 2009). The rest is contributing to drinking water supply after mixing with cold water. Unfortunately, injection of water back to the porous reservoir occurs only in a very few places. Szanyi and Kovács (2010) collected reliable data from Hungary to show that the total thermal water production in the central part of the Pannonian Basin amounts to 80–90 million m^3/year. It has resulted in together with 815 m^3/year drinking water extraction a 4–16 m drawdown of the hydraulic heads in the past decades. This obviously speaks for a more strict regulation of thermal water production. Fig. 16b shows the more complex geometry of thermal water resources in (mostly Mesozoic) fractured-karstified carbonate rocks specifying the exposed regions of water recharge and their subsurface extension. These rock complexes represent upper nappes in the Alcapa terrane (e.g. Northern Calcareous Alps and Transdanubian Central Range), and Tisza-Dacia terrane (e.g. Mecsek, Villány-Bihor and Békés-Codru), and dismembered Dinaric units in-between (e.g. Mid-Hungarian Fault Zone and Bükk). Fig. 16b indicates 56 installations in and another 21 outside of Hungary. They utilize thermal water from shallow to 2000 m depth with temperatures of 30–100 °C, and the

(a)

(b)

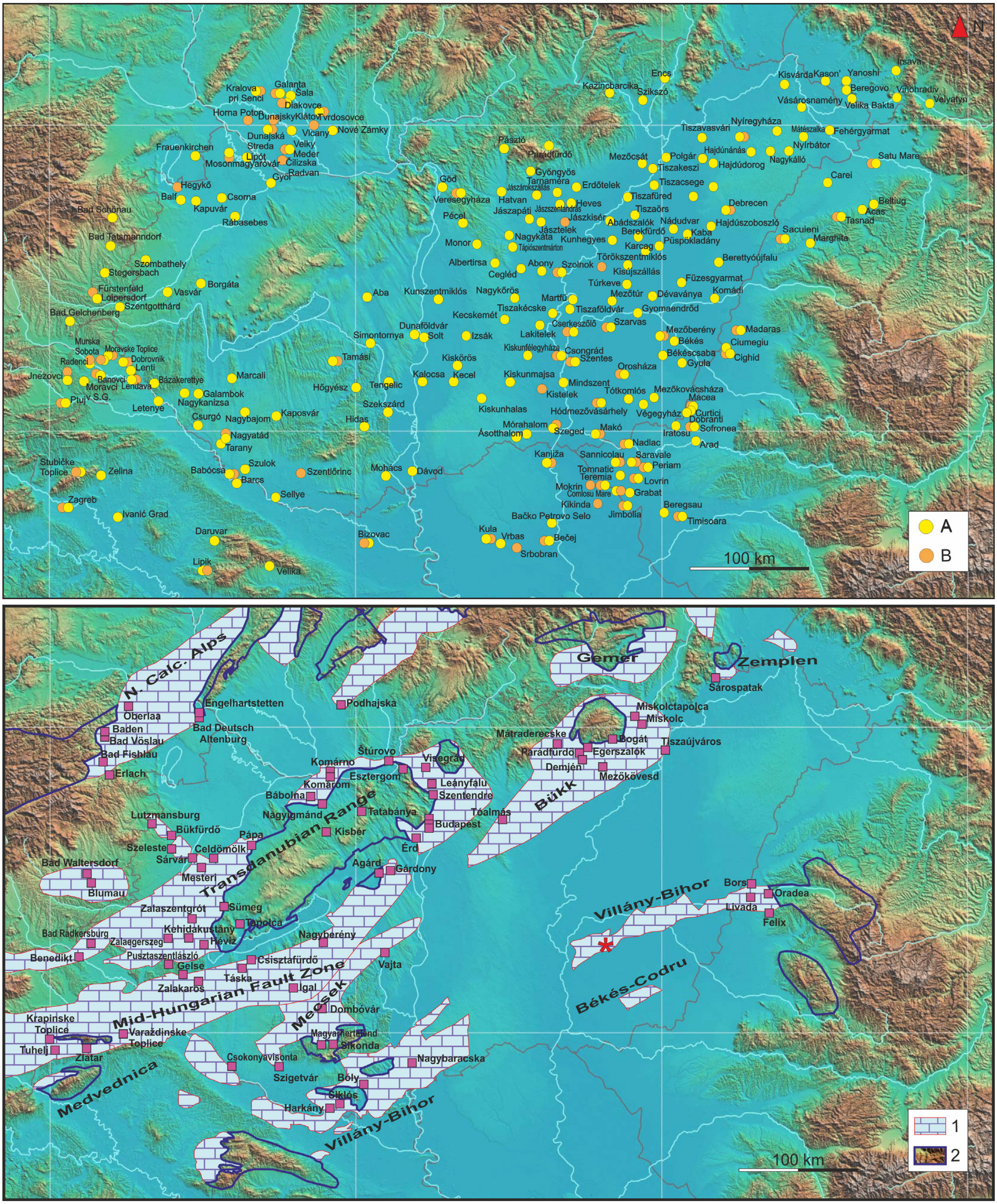

Fig. 16. (a) Location of main geothermal installations in the Pannonian basin utilizing thermal water from the porous Pannonian reservoir (preferably from the "thermal water-bearing layer", see text for discussion).
1 = Thermal bathes and wellness centres; 2 = Space heating included.
(b) Map showing the main Mesozoic karstic reservoirs below the surface (symbol 1) in the Pannonian basin and their recharge areas (symbol 2) bordered by blue lines. Location of geothermal installations is shown by red squares and they are dominantly thermal spas and balneological therapy centres. The red asterisk at the western tip of the Villány-Bihor shows the location of the blowout of the overpressured system.

yield of individual wells can be as high as 1000–1500 m^3/day. It is an interesting feature of the karstic system that shallow rock bodies in the proximity of the recharge region can supply remarkably warm waters (40–60 °C) demonstrating the importance of faults as fluid conduits (Goldscheider et al., 2010). This is the case of the famous thermal bathes in Budapest and Hévíz to the immediate east and west of the Transdanubian Range, resp. (Fig. 16b). As opposed to the large extraction of thermal waters from the two unconfined aquifer systems in the Pannonian basin, there has been no utilization of the high temperature-high pressure geothermal system developed beneath a pressure seal in the deeper part of the basin and fractured basement. This system is, however quite well known as a consequence of an accident during drilling of the Fáb-4 well (Stegena et al., 1994). In the year of 1985 a drilling was started in the central part of the Great Hungarian Plain, above the western tip of the Villány-Bihor Mesozoic strip (Fig. 16b). The well penetrated the following layers: 0–2960 m Pannonian sandstone and claystone, 2960–3153 m Middle Miocene marl, 3153–3750 m Upper Cretaceous sandstone and siltstone, 3750–4034 m Middle Triassic dolomite breccia, 4034–4239 m Lower Triassic sandstone. 70 mD permeability was measured on core samples from the dolomite breccia. 202 °C temperature and 71 MPa pressure was observed during drill-stem test at a depth of 4226 m. In the course of further completion a dramatic blowout occurred from a depth of 3800–4000 m. Wet steam jet destroyed the drilling tower and the well head, and it took more than a month killing the well. During this interval the well-head pressure did not decrease and the steam production was equivalent to 5000–8500 m^3/day water yield. Silica-solubility gave a reservoir temperature of 254 °C, much higher than the in situ value (202 °C), suggesting a convective uprise and mixing of over heated water in the reservoir. It was shown that under the given conditions free convection of overpressured water can take place (Lenkey, 1999). Magnetotelluric soundings indicated a high conductivity anomaly in association with the overpressured system and its extending downward to depths of 8 km (Nagy et al., 1992). The presence and characteristics of this huge geothermal resource has been corroborated by further drillings and geophysical modelling, still utilization is in a very preliminary planning phase.

4. Conclusions

More than a century of academic research and hydrocarbon exploration in the Pannonian basin have resulted in a good knowledge of subsurface geology, crustal and lithospheric structure and upper mantle processes. At the same time spectacular progress can be seen in understanding the tectonic development of the surrounding Alpine-Carpathian-Dinaric mountain system and its relationship to the formation and evolution of the Pannonian basin. Data and results from very diverse fields have been reviewed in this paper with the aim to offer a general problem statement and a coherent tectonic picture. Furthermore, it was our intention to illustrate the use of this general knowledge in exploration and utilization of geothermal resources.

The most important observations and their interpretation led to the following main conclusions:

i) The Pannonian basin is a backarc basin developed from the beginning of the Miocene and formed by extensional disintegration of orogenic terranes and subsequent events of basin inversion. These deformations resulted in variable basin morhology characterized by deep half grabens, relative basement highs and island mountains exposing the substrata of the basin;
ii) Basin fill can be divided into two megasequences by the base Pannonian (early Late Miocene) unconformity. Synrift/postrift boundary coincides with this unconformity in the western part of the basin, but can be younger in the eastern part;
iii) In the central part of the basin relatively thin synrift sedimentary complex is overlain by thick postrift strata, as opposed to peripheral basins (Vienna, Styrian, Transcarpathian and Transylvanian) which are overwhelmed by synrift strata. The synrift complex can be very thick in regions, where explosive eruption resulted in thousand metres of rhyolitic ignimbrites and tuffs;
iv) Primary structures developed during rifting and opposite sense rotations of the Alcapa and Tisza-Dacia terranes extruded after continental collision from the East Alpine and internal Dinaric orogenic system (resp.) into the Carpathian embayment, a relict oceanic branch of the Alpine Tethys;
v) Extrusion was facilitated by subduction rollback of oceanic lithosphere and subsequent break off of the subvertical slab. This progressed from the North towards the South along the Eastern Carpathians and was accompanied by the formation of a volcanic arc.
vi) The present basin exhibits high heat flow in concert with the asthenospheric dome elevated as high as 50–60 km beneath the Great Hungarian Plain and attenuated crust. Depth of Moho varies in the range of 32–22 km, which fairly well mirror imaging the first order pattern of basement subsidence. This geometry favours a pure shear mechanism of extensional basin evolution;
vii) Slab breakoff halted further extension in the basin, which passed into a neotectonic phase of inversion. A Mio/Pliocene unconformity can be recognized in the basin and its position indicates thousand metre scale differential movements during the Pliocene-Quaternary. The present smooth topography of the basin suggests fast erosion of uplifting areas and sourcing the sinking basins;
viii) Neotectonic inversion has been shaping surface relief, and thus controlling the gravity-driven flow systems in the basin fill, and Mesozoic carbonates. Both systems are used widely to supply thermal waters. Regional decrease of hydraulic heads in the basin fill due to thermal water withdrawal has been observed.
ix) A rapid burial of the low permeability *Endrőd Formation* has generated large overpressure in basin fill below 2000 m and fractured basement rocks, thus created a huge, confined and high-enthalpy geothermal resource.

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