3. The Cenozoic evolution of the Alpian-Carpathian-Pannonian region – Formation of the Pannonian basin

3.1. The recent setup of the basin

The paleo-biogeographic models of Géczy (1973, 1984), based on Jurassic fauna indicate that two units can be distinguished in the pre-Tertiary basement of the Pannonian basin (Fig. 3), the northern (ALCAPA) which has African affinity and the southern (TISZA) which has European affinity. The presence of these two units is supported by paleomagnetic data of Márton & Márton (1978, 1983) that indicate different rotations of these two mega units.

Fig. 3: Geological sketch map of the Carpathian-Pannonian region indicating the major tectonic units, the extent of Neogene calc-alkaline formations and the 5 major volcanic fields of Neogene-Quaternary alkaline basalt flows containing ultramafic xenoliths.

1 - Styrian basin; 2 - Little Hungarian Plain; 3 - Bakony - Balaton Highland; 4 - Nógrád-Gömör; 5 - Eastern Transylvanian basin, MHL - Middle Hungarian Line
The ALCAPA mega unit includes the Western Carpathians, the Bakony Mts., the Bükk unit (Balla, 1988a, b) and the Eastern Alps. Its southern border could be along the Mid-Hungarian line (Fig. 3) (e.g. Fodor et al., 2000, and references therein). Numerous stratigraphic, seismic and tectonic studies indicate the presence of different nappe structures in the pre-Tertiary basement of the unit, based mostly on seismic and drill hole data in the Little Hungarian Plain (e.g.: Tari, 1996, and references therein) and on stratigraphic and tectonic studies in the Bakony Mts. and Bük Unit. The Mid-Hungarian Line is a right lateral shear zone which was rejuvenated several times during the evolution of the Pannonian basin (e.g.: Csontos & Nagymarosy, 1998 and references therein).

South of the Mid-Hungarian unit, the Tisza and Dacia mega units comprise the pre-Tertiary basement of the recent Pannonian basin (Fig. 3). The basement of this mega-unit, covered by thick Tertiary sediments, is also built up by nappes as shown by drill hole and seismic data. Although the basement of this unit is also strongly folded, its evolution is completely different from that of the northern ALCAPA unit (e.g. Géczy 1973, 1984; Márton & Márton 1978, 1983; Lelkes-Felvári, 1982; Szederkényi, 1984).

3.1.1. Lithospheric thickness in the Carpathian – Pannonian region

Based on seismic S-wave velocities the thickness of the lithosphere is 70-90 km in this region (Fig. 4). Delay times of teleseismic P waves also suggest about 50-70 km thick lithosphere in the Pannonian basin (Ádám et al., 1990), whereas in the Vrancea zone earthquakes at depths of 180 km can be related to a lithospheric slab remnant of Miocene subduction (Royden et al., 1982) this slab remnant is suggested by the results of seismic tomography (Spakman, 1990). The seismic gap between 40 and 70 km depth indicates that the slab has already detached (e.g. Onescu et al, 1984). Other interpretations however depict a deep, continuous lithospheric root (e.g. Chekunov & Sollogub, 1989), as a similar root can be observed in the Eastern Alps.

3.1.2. Crustal thickness in the Carpathian – Pannonian region

Thick crust is an attribute of the mountain range around the Pannonian basin. The western Carpathians are characterized by a 32-36 km thick crust whose structure is well-known from seismic profiles (e.g. Tomek & Thon, 1988). The images show a steeply dipping (40-50°) European foreland which is overlain by autochtonous or allochtonous molasses and a well-developed wedge made up from the Outer Carpathian flysch nappes. This subduction-
accretion complex is cut by horizontal reflectors at depth which can be interpreted as a new MOHO formed after the cessation of subduction in the Early Miocene (Tomek et al., 1989). The Eastern and Southern Carpathians are characterized by very thick crust (more than 40 km) and the crustal root coincides with the axis of the tomographic high, with an extremely deep MOHO in the Vrancea zone which rises steeply towards the Transylvanian basin.

![Map of the Alpian-Carpathian-Pannonian region](image)

**Fig. 4:** Lithosphere thickness map of the Alpian - Carpathian - Pannonian region based on Lenkey (1999). Contours at 20 km intervals.

Thin crust is found in the Pannonian basin. The minimum thickness is located NW of Belgrade (22.5 km) (Aljanovic, 1987), whereas beneath the Transdanubian Central Range the crust is somewhat thicker than the average 30 km thickness characteristic of the basin. The high heat flow in the region may be compatible with the anomalous updoming of the upper mantle. Positive Bouger anomalies observed in the Pannonian basin also suggest elevated upper mantle (Tomek, 1988), and the highest anomalies are always associated with the deepest part of the sub-basins. Modern reflection seismic profiling (e.g. Posgai et al., 1986) shows strongly layered lower crust in many areas, which is supported by observations of increased electric conductivity in deep crustal levels (Ádám et al. 1989). Warner (1990)
suggests that lower crustal layering is caused by subhorizontal shear zones developed during crustal extension and mostly by related intrusions of mantle derived mafic melts into the lower crust, and this model is also supported by interpretation of gravity anomalies.

3.2. The Cenozoic evolution of the Alpian-Carpathian-Pannonian region

3.2.1. Escape models

Based on the paleobiogeographic models of Géczy (1973, 1984) and paleomagnetic data of Márton & Márton (1978, 1983) described above, Kázmér & Kovács (1985) proposed a model in which the ALCAPA unit escaped eastwards from the Alpine-Dinaric System, moving up to 450 km (Fig. 5). The southern border of this escape could have been the Mid-Hungarian Line. Balla (1984) taken into account the paleomagnetic data and counted with an escape combined with the rotation of the ALCAPA and TISZA units (Fig. 5a) whereas Csontos et al. (1992) modeled the escape without block rotations (Fig. 5b). Below the unit large, low-angle normal faults controlled the movement, i.e. the downwards sliding from the Penninic unit (Ratsbacher et al., 1989, 1991a, b; Neubauer & Genser 1990; Decker & Perreson, 1996).

Fig. 5: Plate tectonic models for the Tertiary evolution of the Carpathian - Pannonian system (Csontos, 1995).

b. Csontos et al. (1992) - escape without rotation.
3.2.2. Subduction and slab break-off model

Nemcok et al. (1998) summarized and synthetized a wide range of diverse geological observations (the available dataset of seismic, gravity, conductivity studies, the nature of the volcanic activity, crustal and lithosphere thickness variations, uplift data, and facies correlations) and from this comprehensive survey they outlined a new model which provided further insights into the evolution of the Carpathian – Pannonian region.

They proposed that the 19-0 Ma calc-alkaline volcanism of the Inner Carpathian Volcanic Chain, as well as the 22.8-19.7 Ma old calc-alkaline magmatism in the Northern Croatian part of the Pannonian Basin (Simunic & Pamic, 1993), are the eastern continuation of the 42-25 Ma old Alpine calc-alkaline magmatism (von Blanckenburg & Davies, 1995) along the Periadriatic Lineament. The younger ages of the Croatian and Inner Carpathian volcanic rocks indicate propagation of slab break-off from west to east. The temporal and spatial development of Neogene foreland basins (e.g. Kovac et al., 1989), as constrained by the timing and direction of the last Flysch Belt thrusting (e.g. Jiricek, 1979) and by structural data (e.g. Nemcok, 1993), indicates that the oceanic plate underlying the remnant of the Carpathian Flisch Basin subducted south-westwards than westwards. This is supported by the seismic data of Tomek & Hall (1993) and Bielik et al. (1998) which point out a flat angle subduction of continental margin beneath the Western Carpathians propagating southwards. The Paleogene basins of the Carpathian-Pannonian region, which have compressional or transpressional origins (Fodor et al., 1992, 1994; Tari et al., 1993; Csontos & Nagymarosy, 1998), were formed in the background of this compressional belt (Tari et al., 1993). The ongoing subduction finally led to the collision between the European platform and the Carpathians at the end of the Early Miocene. The progressive change from the subduction to collision eastwards along the Carpathian arc controlled the break-off of the subducting slab (Seghedi et al., 2001), as indicated both by the calc-alkaline volcanism of the Inner Carpathians that started at 19-17.5 Ma until 0 Ma, and the ongoing detachment process in the Vrancea area (e.g.: Fuchs et al., 1979) (the detachment started in the West during the end of the Early Miocene and ran along the Carpathian arc to its present position in the Brassow area). The detachment-related volcanism was synchronous with collision.
3.2.3. The Neogene evolution of the Pannonian basin in the background of the Carpathian subduction

Extension caused by the role-back effect of the subducted slab was accompanied by the onset of crustally-derived volcanism above an asthenospheric upwelling (Stegena et al., 1975; Royden et al., 1983a, b; Salters et al., 1988; Lexa et al., 1993; Pécskay et al., 1995). Stegena et al. (1975) pointed out that mantle diapirism is the key phenomenon in the formation of the Pannonian basin, assuming that the Carpathian arc is a result of a lithosphere subduction process as suggested by numerous authors.

Horváth (1993) proposed a three layer extension model for the Pannonian basin based on the relations between the shallow and deeper lithospheric structure beneath the Little Hungarian Plain, which showed that:

1) In the brittle upper crust extension is accommodated by low-angle normal faults.
2) The lower crust is extended by subhorizontal ductile shear.
3) The mantle lid of the lithosphere is characterized again by high strength and extension is taken up primarily by movement along mantle fault and internal deformation of the mantle footwall.

Furthermore, Horváth & Cloetingh (1996) modelled the formation of the Pannonian basin by a two-layer finite stretching model of the lithosphere incorporating lateral heat flow. They adopted crustal stretching parameters varying from 1.3 in Transdanubia to a maximum value of 1.8 in the Great Hungarian Plain. The stretching event was taken to be from 17-12 Ma taking into account the compaction of sediments. The basin formation process would be in association with thick postrift Alpine crust and high T regime. An important feature in the model is the incorporation of an increase in the level of stress during the last 2 Ma from 0 to 400 MPa compression to get the best fit between the predicted and the real stratigraphy in the basin. The modelling also suggests the notable absence of lithospheric strength in the mantle part. The strength profiles are characterized by a contraction of strength in the upper 15 km. Furthermore, the model points out that a dramatic change in the system occurred a few million years ago, when slab pull became inefficient as roll-back of the subducted slab could not proceed further. This occurred because the attenuated crust of the former marine basin had been fully consumed and the accretionary wedge had reached the Tornquist-Teisseyre zone of the East European foreland. From this time the Pannonian basin became completely locked in a stable continental environment without any chance of further extension in any direction.
This tectonic setting implies the establishment of a new stress field controlled by the Europe/Africa convergence.

Quantitative models of extensional basin formation have explored the possible role of lower crustal flow toward the rift flanks in response to MOHO uplift (Moretti & Pinet, 1987) or in response to surface processes where erosion of the rift flanks may enhance lower crustal thinning in the basin area (Burov & Cloetingh, 1997). Such processes potentially explain the observed differential thinning of the lower crust with respect to the upper crust. Modeling shows that small-scale convective upwelling following a first phase of passive rifting may explain the late synrift to postrift mantle lithosphere thinning. It should be emphasized that no external heat source (i.e. mantle plume) is needed to drive the convective flow.

Huismans et al. (2001) applied a two-dimensional thermomechanical finite element model to the Pannonian basin history to investigate the conditions of small-scale convective upwelling in the mantle following the first passive rift phase. On the basis of tectonic evolution the following model provides the most reasonable scenario: in the model the thickness of the initial lithosphere was 150 km and the total amount of extension is 100 km (Fig. 6a). The first phase of extension happened between 17.5 and 14 Ma due to the role-back effect of the subducted slab. At 14 Ma an ~150 km wide surface depression and ~200 km wide upwelling of the asthenosphere developed. The subsidence of the center of the basin area was about 1800 m, whereas the asthenosphere had risen 40 km (Fig. 6b). At the end of the synrift the stretching factor for the crust was 1.3 whereas the mantle lithosphere was thinned by a factor of 2.2. In the postrift phase the asthenosphere has risen an additional 7-8 km, with a total mantle lithosphere thinning at 0 Ma of 3.0 (Fig. 6c). The asthenospheric dome obtained its peak rise velocity 7.6 km/my at 2.3 Ma following the initiation of the second rift phase.

Fig. 6: The Miocene to Pliocene evolution of the Pannonian basin system based on Huismans et al. (2001).

a. Prerift thickening,
b. Passive rifting driven by slab roll-back
c. Active rifting due to asthenosphere doming.