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published in
Tectonophysics
1996

DOI (link to publisher)
10.1016/S0040-1951(96)00194-1

document version
Publisher's PDF, also known as Version of record

ELSEVIER

Tectonophysics 266 (1996) 287-300

TECTONOPHYSICS

Stress-induced late-stage subsidence anomalies in the Pannonian basin

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Received 21 November 1995; accepted 20 June 1996

Abstract

Subsidence, sedimentation and tectonic quiescence of the Pannonian basin was interrupted a few million years ago by tectonic reactivation. This recent activity has manifested itself in uplift of the western and eastern flanks, and continuing subsidence of the central part of the Pannonian basin. Low- to medium-magnitude earthquakes of the Carpathian–Pannonian region are generated mostly in the upper crust by reverse and wrench fault mechanisms. There is no evidence for earthquakes of extensional origin.

2-D model calculation of the subsidence history shows that a recent increase in magnitude of horizontal compressional intraplate stress can explain fairly well the observed Quaternary uplift and subsidence pattern. We propose that this stress increase is caused by the overall Europe/Africa convergence. In Late Pliocene, consumption of subductible lithosphere at the eastern margin of the Pannonian basin was completed, and the lithosphere underlying the Pannonian basin became locked in a stable continental frame. Consequently extensional basin formation has come to an end, and compressional inversion of the Pannonian basin is in progress.

Keywords: basin evolution; basin modeling; tectonic inversion; intraplate stress

1. Introduction

The thermomechanical model of extensional basin formation (McKenzie, 1978) implies that two contrasting periods of evolution can be distinguished: an initial rifting period followed by a more passive phase of subsidence due to thermal contraction of the crust and underlying mantle lid of the lithosphere. Study of reflection seismic sections in many parts of the Pannonian basin and analysis of fault striations in outcrops have documented the validity of this two-phase division (Horváth and Rumpler, 1984; Bergerat, 1989). The late Early Miocene through Middle Miocene basin fill represents the synrift sequence, and the unconformably overlying Late Miocene to Quaternary sedimentary rocks belong to the postrift period.

A first signal of deviations from this model came from sediment accumulation diagrams constructed by using careful biostratigraphic dating of the basin fill (Nagymarosy, 1981). These diagrams indicated that in the deepest parts of the basin (e.g.,, Makó trough, Fig. 1) subsidence rate increased during the Quaternary.

Elsewhere in the Pannonian basin, opposite vertical movement trends have been inferred. Horváth et al. (1988) carried out subsidence modelling in the Great Hungarian Plain to explain the present temperature

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PII S0040-1951(96)00194-1

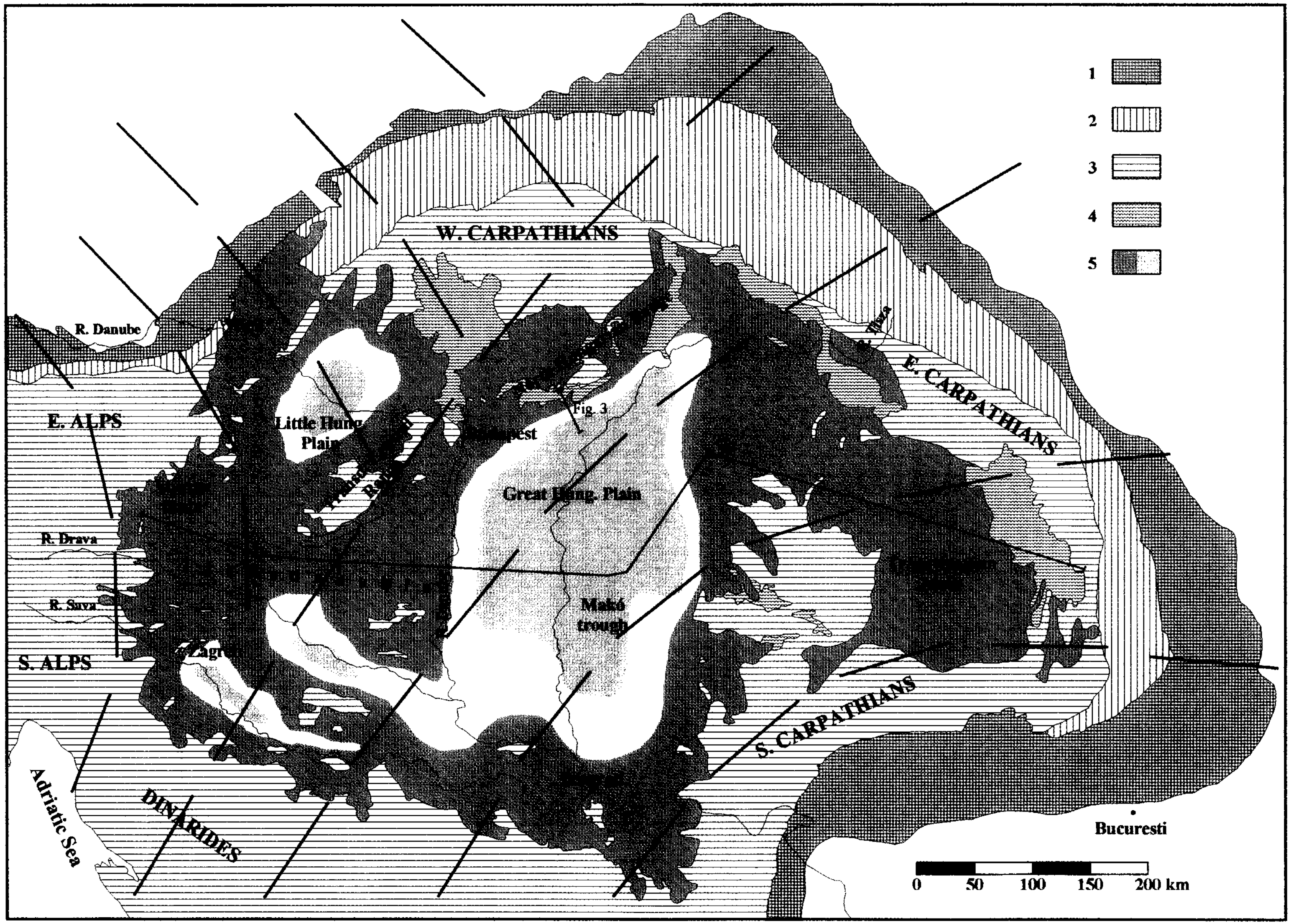

Fig. 1. Simplified tectonic map showing the Pannonian basin and surrounding mountain belt, and the direction of the recent maximum horizontal stress (Rebaï et al., 1992; Müller, 1993). Locations of the long section in Fig. 2 from the Styrian basin to the Transylvanian basin, and the seismic section in Fig. 3 are also indicated. Key: *1* = Molasse foredeep; *2* = Alpine–Carpathian flysch belt; *3* = internal ranges of the Alps, Carpathians and Dinarides; *4* = Miocene calc-alkaline volcanic mountains; *5* = areas of Quaternary uplift and subsidence in the Pannonian basin system. Note that uplift of mountains (*2* and *3*) are not indicated.

and hydrocarbon maturity conditions. Their analysis in Transdanubia showed that model results could be matched to observed data only if a few hundred metres of uplift sometime between 5 Ma and 1 Ma was assumed. Late-stage subsidence anomalies turned out to be a general feature rather than exceptions.

In this paper we demonstrate that the Quaternary has been a period of an important change in tectonic regime. This manifests itself in late-stage subsidence anomalies and structural deformations. Model calculation shows that a late-stage increase of intraplate stress magnitude can explain quite well the observed vertical movement trends along an E–W-oriented crustal section. We discuss the evolution of the Pannonian basin and offer a geodynamic scenario which helps to understand the fate of basin formation.

2. Geomorphological data

Quaternary isopach maps show a very irregular thickness variation from 0 to 700 m in the Pannonian basin, and areas characterized by Quaternary uplift can be recognized (Rónai, 1974). A striking feature is that about 20% of the territory of the Pannonian basin (Fig. 1) is not a depression but is composed of elevated ranges exposing rocks older than the basin fill (mostly Mesozoic and Palaeogene). Early geomorphological observations, and the presence of isolated and eroded flanks of Late Miocene strata at elevated positions have already indicated that these ranges are the results of Late Pliocene through Quaternary uplift (Moldvay, 1965; Jámbor, 1980).

Of particular importance are the data provided by studies of the terraces of the Danube river. The Danube cuts the Transdanubian Central Range and changes direction by about 90° in the 'Danube-bend' area to the north of Budapest (Fig. 1). Pécsi (1959) mapped and correlated terraces of different age at different present-day topographic levels. He concluded that, relative to the present-day water table of the Danube, the Transdanubian Central Range was uplifted by about 200 m during the Quaternary, and most intensively in the Mindel–Riss interglacial period. Further studies of Miocene travertine horizons in the same Range supported these findings and suggested a tectonic uplift of 70 to 80 m for the Middle Pliocene, 80 to 100 m for the Late Pliocene, and 130 to 140 m for the Quaternary (Pécsi et al., 1984).

Geomorphological studies have led to additional interesting results. As opposed to the very flat Hungarian plains (the Little and Great Hungarian Plain, see Fig. 1), most part of Transdanubia is characterized by rather regular surface undulations with height differences up to 300 m. This is because the whole territory has undergone uplift and erosion mostly controlled by wind and river activity (Pécsi, 1986). This is documented by basaltic volcanoes which were erupted during this late period of uplift. Volcanism occurred between 7 and 2 Ma (Balogh et al., 1987), and covered the contemporaneous surface sediments of Pliocene age. However, Late Pliocene sediments have been eroded off during the Quaternary with the exceptions of those covered by basaltic lava flows.

3. Seismicity and recent stress field

The Pannonian basin and its surroundings are characterized by a low- to medium-level of seismic activity (Gutdeutsch and Aric, 1988; Procházkova and Roth, 1993). Earthquakes with magnitude less than 6 can occur in the central part of the basin. Their focal depth is always within the crust, mainly in the upper part with a maximum of occurrence in the depth range of 6 to 10 km (Zsíros et al., 1987). Areal distribution of epicentres is quite irregular and, hence, precise definition of seismoactive faults has always been problematic (e.g., Horváth, 1984). The outer rim of the Pannonian basin, however, exhibits an increased level of seismic activity. More frequent earthquakes up to magnitude 6.5 and again mainly of upper crustal origin clearly delineate a number of seismogenic fault zones, like the Mur–Mürz–Little Carpathian line in Austria and Slovakia, and the Zagreb line in Croatia. The Dinaric rim of the Pannonian basin is the most active zone, particularly along the Adriatic coastline of the mountain belt.

Focal mechanism solution of these crustal events indicates thrust and wrench faulting in the mountain belt around the Pannonian basin. Interestingly enough, this is also the case for the Pannonian basin (Müller et al., 1992). The most recent collection of all available data and new solutions has given further support to this finding by demonstrating the absence of normal faulting earthquakes in the intra-Carpathian basin system (Gerner et al., 1995).

Additional stress indicators in the Carpathian–Pannonian region are provided by in situ stress measurements (Becker, 1993), and borehole break-out studies (Dövényi and Horváth, 1990; Müller et al., 1992). These indicators are compatible with a 'radial' pattern of maximum horizontal stress (S_H) trajectories first inferred by Rebaï et al. (1992) (Fig. 1). Maximum horizontal stress is approximately normal to the Adriatic coastline of the Dinarides. This northeast direction continues toward the Pannonian basin, where S_H trajectories begin to diverge. In the western Pannonian basin S_H directions bend toward the north and northwest, thus taking up the characteristic West European pattern. In the eastern Pannonian basin the bend is toward the northeast, so much as in Transylvanian basin and Eastern Carpathians S_H directions toward the east are predominating (see for recent stress data in the Transylvanian basin Huismans et al., 1997). This radial pattern of regional stress field is presumably generated by superposition of internal body forces (e.g., crustal thickness variations) and boundary forces (e.g., push by the Adriatic promontory).

Another important boundary condition is related to the presence of a subducted lithospheric slab beneath the Dinarides and Eastern Carpathians (Spakman, 1990). Seismic tomography delineates a high-velocity slab underneath the Dinarides which is dipping towards the northeast. The slab is completely detached from the Adriatic lithosphere and seismically inactive suggesting a relict of a former subduction zone. Similarly extinct subduction can be traced below the Eastern Carpathians with a southwest dip. However, the Vrancea region at the southern termination of the Outer Eastern Carpathians is the only place in the Carpathian–Pannonian region where subcrustal earthquakes down to 200 km depth and with magnitude up to 7.7 occur. Velocity structure and focal depth distribution depict a slab steeply dipping towards the southwest with an angle of about 60° (Oncescu et al., 1984). The slab is probably detached, as suggested by a seismic gap between 40 and 70 km, and sinking under its own weight into the lithosphere. Earthquakes shallower than 40 km are all of the thrust-type and generated by maximum horizontal stress normal to the Carpathian arc (Fuchs et al., 1979).

4. Stratigraphy and basin structure

During the Middle Miocene (synrift phase) several individual extensional/transtensional basins formed, while a significant part of the Pannonian region remained intact and elevated. These local highs were the subject of subaerial erosion and source of clastic influx, together with the supply derived from the Alpine–Carpathian mountains at the periphery of the Pannonian region. The late Middle Miocene (about 12 Ma) marks the onset of the postrift phase which is documented by a regional unconformity, sealing most of the rift-related faults by Upper Miocene strata. Thermal cooling of the asthenospheric dome below the extended lithosphere resulted in a general subsidence of the whole Pannonian area, which became a large single depression, although with uneven bottom morphology during the Late Miocene. Seismic data combined with well-log interpretation and core-sample analysis, demonstrate that the Pannonian depression has been filled up by a fluvial dominated delta-system which prograded from the basin periphery towards the interior (Pogácsás, 1984; Bérczi and Phillips, 1985; Mattick et al., 1988). Sediment accumulation rate was significantly higher than the subsidence rate during Late Miocene time. Infilling continued in the Pliocene, and very shallow lacustrine to terrestrial conditions prevailed all over the basin. This slowly and generally subsiding area has experienced, however, a tectonic reactivation during the Quaternary. This is manifested by accelerated subsidence, uplift and young faulting.

Regional isochrons for the Pannonian basin have been derived recently by combining sequence stratigraphy and age data determined by magnetostratigraphy and radiometric dating of interbedded volcanic rocks (Pogácsás et al., 1988; Elston et al., 1990; Csató, 1993; Vakarcs et al., 1994). These data provide a framework to understand the pattern and amplitude of the Late Pliocene and Quaternary subsidence and uplift.

Fig. 2 shows a regional stratigraphic section across the Pannonian basin in an E–W direction from the East Carpathian Neogene volcanic range to the Styrian basin at the foothills of the Eastern Alps. The very significant undulation of the top of the basement is essentially caused by two factors, i.e. differences in Middle Miocene synrift subsidence due to inhomo-

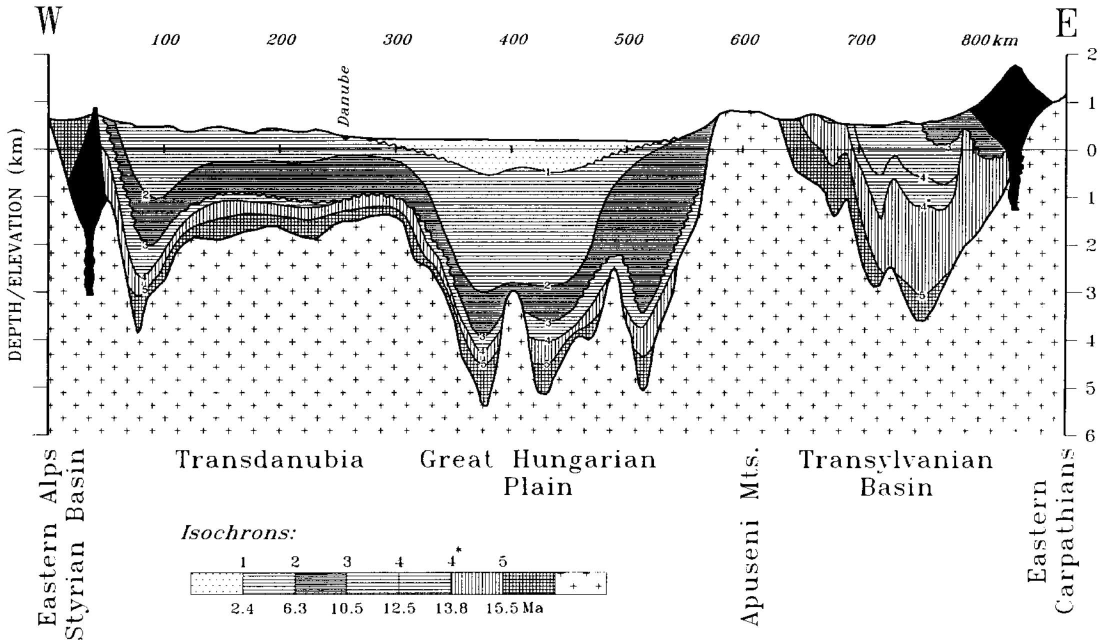

Fig. 2. Generalized cross-section across the Pannonian basin showing the geometry of selected time-lines (isochrons 1, 2, 3, 4, 4* and 5) in the Neogene sedimentary fill (mostly after Ciupagea et al., 1970; Vakarcs et al., 1994; Ebner and Sachsenhofer, 1995). Absolute ages of the time-lines are also indicated in the box below the section. Note that age of the lowermost strata (below isochron 5 and above the basement indicated by + +) is different along the section. The oldest basin fill to the west of the Apuseni Mts. is generally Early Miocene (about 20–22 Ma), but it can be as old as Eocene (about 50 Ma) in the Transylvanian basin. Location of the section can be seen in Fig. 1.

geneous extensional strain, and Quaternary vertical movements. It is important to note that Late Pliocene and Quaternary strata (0–2.4 Ma) can only be found in the Great Hungarian Plain with a thickness up to 700 m. The surface of this area is indeed a plain and with an elevation between 90–120 m above sea level. In contrast, the western part of the Pannonian basin (Transdanubia and the Styrian basin) are more elevated, surface morphology is more variable (200 to 700 m elevations), and Quaternary deposits are practically missing. There is further evidence of Quaternary subsidence in the Southwestern Vienna basin (Mitterndorfer Senke, see Tollmann, 1984), where a minimum Quaternary subsidence of 150 m has been observed. Between this region and the Little Hungarian Plain there is an area with young uplift (Leithagebirge). Striking is the asymmetry of strata in the Styrian basin, which is mostly the consequence of uplift, tilting and erosional truncation (Ebner and Sachsenhofer, 1995; Sachsenhofer et al., 1997). The same pattern can be observed even more dramatically in the Transylvanian basin (Ciupagea et al., 1970). Inspection of the section leaves little doubt that the Apuseni Mountains were also covered by Neogene sedimentary blanket, and older rocks have become exposed due to recent uplift and erosion.

Fig. 3 shows a seismic section and its interpretation in Hungary just to the south of the North Hungarian Range, illustrating a typical stratigraphic pattern at the transition from the area of recent uplift (north) to subsidence (south). A few kilometres to the north from the end of section the Triassic carbonate rocks are already exposed on the surface and elevated up to 900 m in the Bükk Mountains. A recent fission track study (Dunkl et al., 1994) indicates a burial of these Mesozoic rocks in the Bükk by about 1000 m of Tertiary strata as late as 5 Ma, and their rapid unroofing since then.

A synthesis of available geomorphological and seismic data has led to the generalized map in Fig. 1 showing areas of Quaternary subsidence and uplift in the intra-Carpathian basin system.

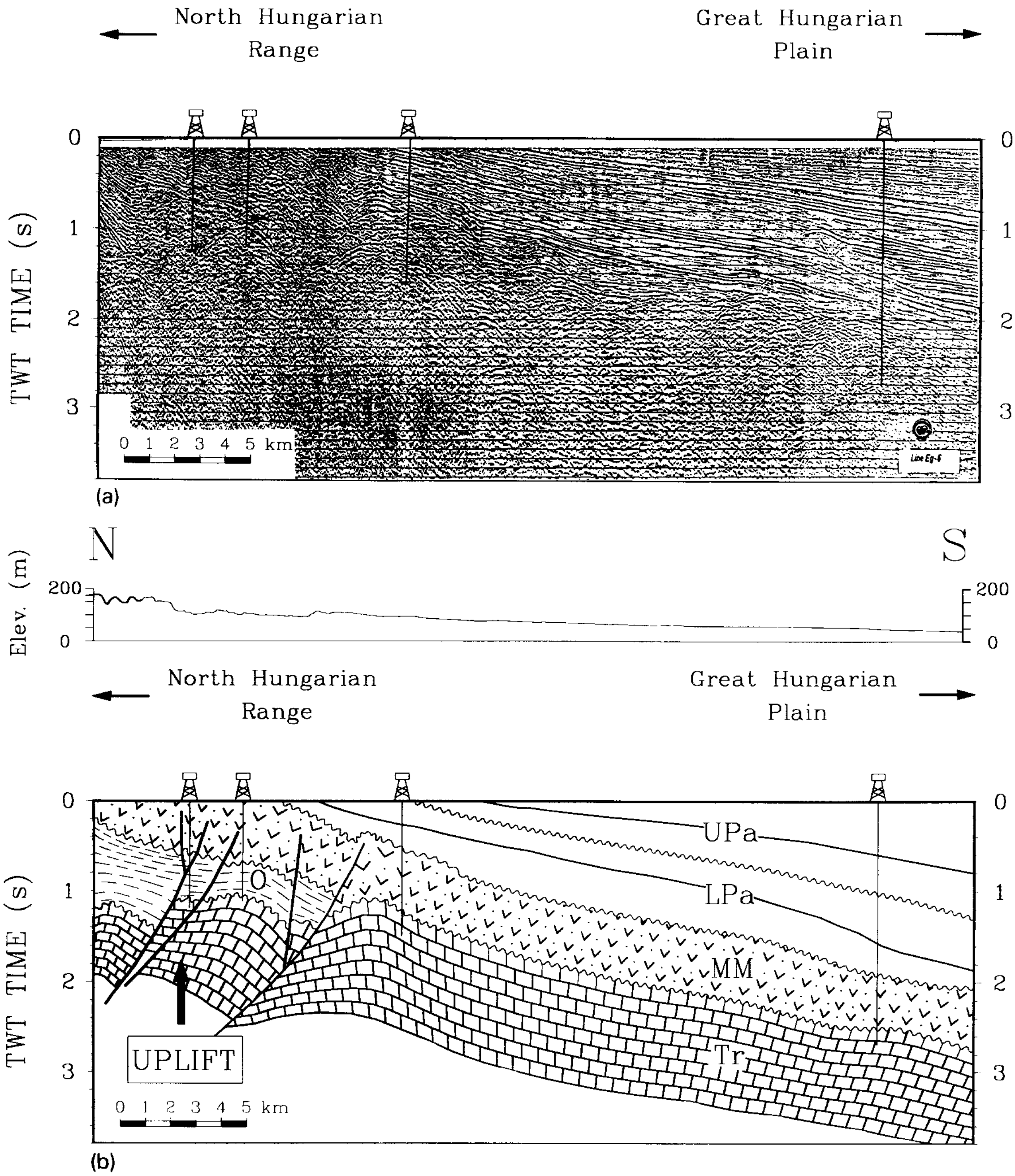

Fig. 3. A 35-km-long reflection seismic section (a) and its interpretation (b) to show the characteristic stuctural–stratigraphic features in the transition zone between the North Hungarian Range (area of Quaternary uplift) and the Great Hungarian Plain (area of continuous subsidence). Location of the section can be seen in Fig. 1. Substrata of the basin are made up of Triassic limestones (*Tr*). They are overlain by Oligocene marls and sandstones (*O*) on the north and directly by Middle Miocene volcanoclastics (*MM*) in the south. This is followed by Lower and Upper Pannonian (*LPa* and *UPa*, resp.) clastic deposits of Late Miocene trough Pliocene in age. Faults at the left side of the section are Middle Miocene strike-slips associated with the opening of the Pannonian basin and calc-alkaline volcanic activity. Strongly tilted strata and their erosional truncation indicate recent uplift at the left side of the section in the North Hungarian Range.

5. Modelling of the basin evolution

To investigate the basin response to a variation in the stress field during its late-stage evolution, we have performed forward modelling of the basin stratigraphy and associated vertical motions. We adopt a two-layer finite stretching model of the lithosphere incorporating lateral heat flow employed earlier by us for the modelling of a number of Mediterranean basins, including the Gulf of Lions (Kooi et al., 1992), the Valencia Trough (Janssen et al., 1993) and the Tyrrhenian Sea (Spadini et al., 1995).

Based on results of earlier modelling studies of the Pannonian basin (Royden and Dövényi, 1988), we have adopted crustal stretching parameters ranging from 1.3 in Transdanubia to a maximum value of 1.8 at the Great Hungarian Plain. The stretching event is taken to be from 17 Ma to 12 Ma. Palaeowaterdepths have been inferred from well data and seismic stratigraphy (Mattick et al., 1988). Compaction of the sediments has been taken into account using standard equations from Bond and Kominz (1984). The flexural strength of the lithosphere is taken to be controlled by the 300°C isotherm.

An important feature in the modelling is the incorporation of finite strength of the lithosphere during extensional basin formation. The basin formation process is described in terms of the mechanism of lithospheric necking (Braun and Beaumont, 1989; Kooi et al., 1992). The level of necking can be connected with the zone of maximum lithospheric strength (Kooi et al., 1992), or alternatively with the depth of intra-lithosphere detachment (Van der Beek et al., 1994). The basin formation processes in the Pannonian basin would by its association with thick pre-rift Alpine crust and a high-temperature regime qualify for a low-strength lithosphere and a level of necking at shallow depth (see Cloetingh et al., 1995 for discussion). The basin formation processes operating in the Pannonian basin have been modelled by us (Van Balen and Cloetingh, 1995; Van Balen et al., 1996) using constraints from a large dataset of wells and seismic profiles acquired by petroleum exploration in the basin support a necking depth level between 5 and 10 km. Fig. 4. displays strength profiles through the lithosphere constructed on the base of extrapolation of rock mechanics data, incorporating constraints on crustal and lithospheric structure and present-day heatflow for two sites along the modelled cross-section. Inspection of the lithospheric strength distribution for the western part of the Pannonian basin (Fig. 4a, see Sachsenhofer et al., 1997 for further details) and the strength profile for the eastern part of the Pannonian basin (Fig. 4b, see

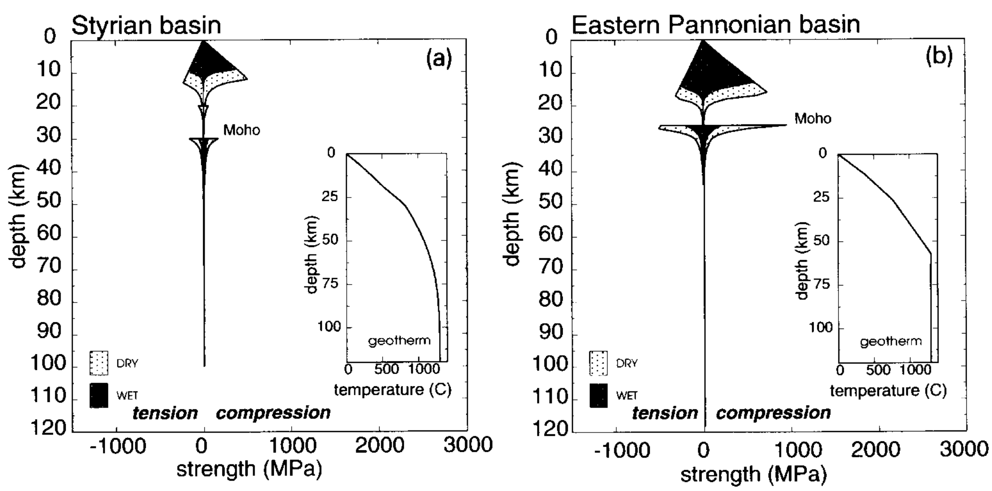

Fig. 4. Strength profiles for two locations along the modelled section through the Pannonian basin (see for location Fig. 1.). The strength envelopes have been calculated adopting constraints on crustal and lithospheric thickness, petrological stratification of the lithosphere (quartzitic/granitic upper crust, dioritic lower crust, olivine mantle) and surface heatflow. A strain rate of 10^{-15} s^{-1} is adopted. Noticeble is the absence of mantle strength predicted by the model. (a) Predicted strength profile characteristic for the Styrian part of the Pannonian basin (Sachsenhofer et al., 1997). (b) Strength distribution for the Romanian part of the Pannonian basin (Lankreijer et al., 1997).

Lankreijer et al., 1997 for discussion) demonstrates the noticeable absence of lithospheric strength in the mantle part of the lithosphere in the Pannonian basin. The strength profiles are characterized by a concentration of strength in the upper 15 km of the lithosphere. These findings are consistent with the absence of seismicity in the Pannonian basin at a level deeper than 15–20 km (Zsíros et al., 1987). We have adopted a lithospheric necking depth of 10 km as an average representation of the bulk strength of the lithosphere in the Pannonian basin.

Although formed in an extensional regime, as discussed above, the present-day Pannonian basin has been strongly affected by compressional neotectonics (Gerner et al., 1995; Bada et al., 1996). An important feature in our model is the incorporation of an increase in the level of stress during the last two million years from zero to 400 MPa compression (equivalent to a force of 6×10^{12} N/m in a 15-km-thick elastic plate). This stress value should be considered as an upper estimate, based on an elastic plate analogue of the mechanical properties of the lithosphere. As discussed by Cloetingh et al. (1989), the incorporation of a depth-dependent brittle–ductile rheology would significantly enhance the effectivity of the horizontal stresses, in particular for stresses with a level approaching the integrated strength of the lithosphere. Under such circumstances lithospheric folds can be developed with vertical motions of the order of kilometres (Cloetingh and Burov, 1996). The incorporation of 3-D modelling approaches (see Van Wees and Cloetingh, 1996) would also add drastically to reduction of the level of the horizontal forces required to lead to a significant deflection of the lithosphere.

Fig. 5 shows the result of the forward modelling. In the model we focus on the late-stage evolution of the basin and therefore simplify the stratigraphy into four sequences (Fig. 5a): a synrift sequence, two postrift sequences and a Plio–Quaternary sequence. Fig. 5b displays the predicted stratigraphy of the modelled section in the absence of an intraplate stress field. This model shows a moderately good fit with the observed stratigraphy. Fig. 5c illustrates the effect of a late-stage intraplate stress on the stratigraphy. The predicted stratigraphy in the central part of the profile in the area of the Great Hungarian Plain shows a thickening of the Plio–Quaternary sequence. At the same time the model predicts an uplift of the flanks in Transdanubia and the Apuseni Mountains. These stress-induced perturbations in basin shape and associated vertical motions lead to a better fit of the predicted stratigraphy with the observed stratigraphy.

Fig. 5d shows the stress-induced deflection along the profile. The differential stress-induced subsidence and uplift has an amplitude of a few hundred metres and a wavelength of about 400 km. The wavelength is in the same range as observed by other studies of compression-induced deflection of continental lithosphere in the North Sea area (Kooi et al., 1991; Cloetingh and Kooi, 1992), Arctic Canada (Stephenson and Cloetingh, 1991), and Central Asia (Nikishin et al., 1993; Burov et al., 1993). At first sight the large wavelengths for stress-induced deflection in the Pannonian basin, characterized by a decoupled mode of upper crustal undulations, is surprising in the light of the low rigidity of the lithosphere in this area. It should be noticed, however, that in basins such as the Pannonian basin and the North Sea, the wavelength of 400 km is primarily controlled by the configuration of the rift basin and its associated sedimentary loads prior to the onset of the late-stage compression.

From the foregoing, it appears that the model invoking finite strength of the lithosphere during extension and an increase in the level of stress during the late-stage evolution of the Pannonian basin explains some of the essential features of the basin configuration and stratigraphic architecture. It is important to realise, however, that a number of other features are not explained very well with the presented 2-D model. Noticeable deviations remain between observed and predicted topography and stratigraphy at both ends of the section in the Styrian and Transylvanian basins. In the following section we discuss these deviations in the context of the geodynamics of the late orogenic evolution of the region.

6. Geodynamics of the late orogenic evolution

Fig. 6 presents a simple three-dimensional lithospheric block model of the Alpine/Carpathian/Pannonian system, summarising the basic kinematic and structural features controlling the geodynamics of the late orogenic evolution.

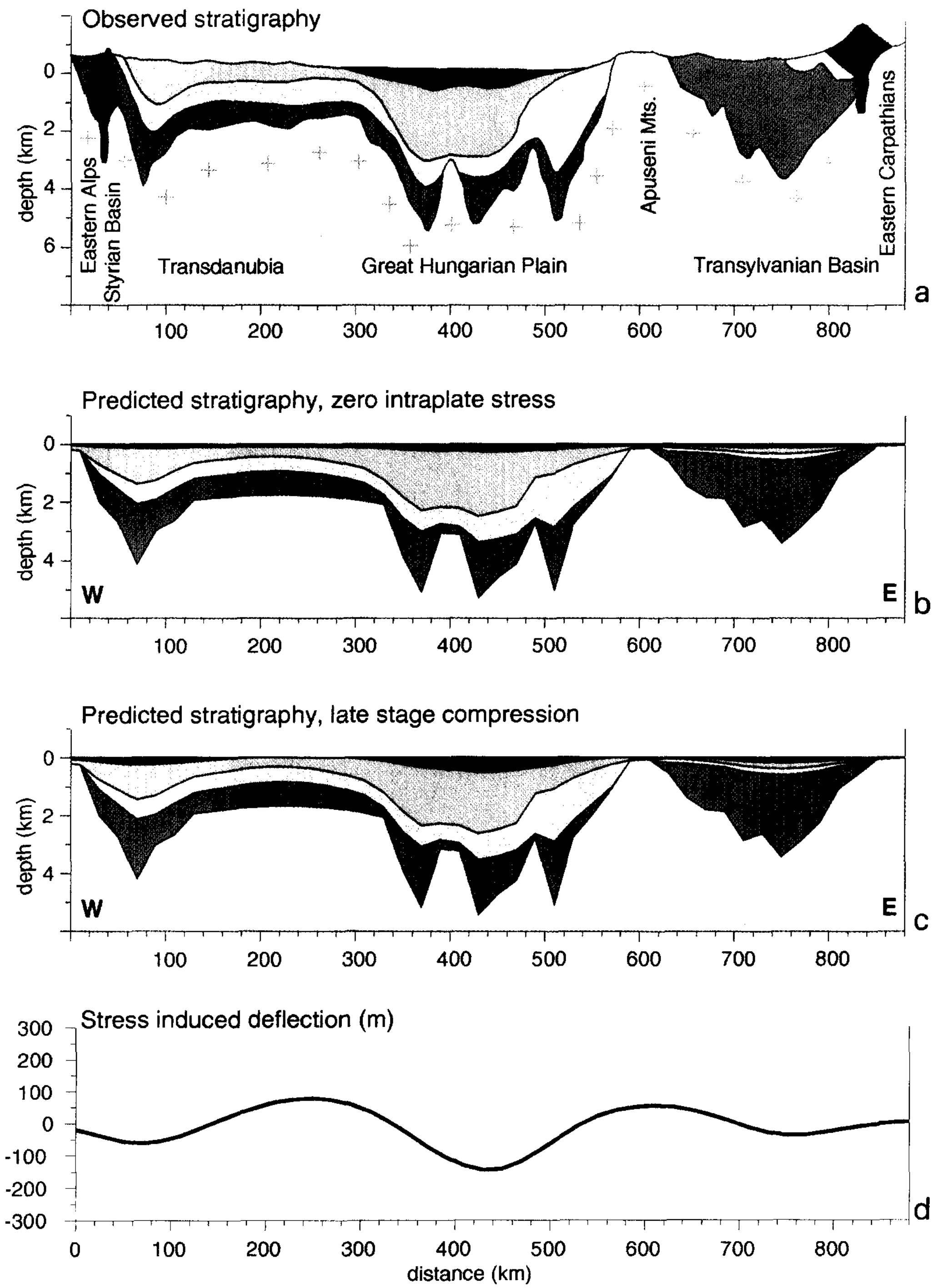

Fig. 5. Panels showing stratigraphy and basin configuration for the section through the Pannonian basin in Fig. 2. (a) Observed and simplified stratigraphy. (b) Predicted stratigraphy for a model incorporating lithospheric necking and assuming a zero stress field in the lithosphere. (c) Predicted stratigraphy for a model incorporating necking and a stress field which changes from zero stress at 2 Ma to a magnitude of 400 MPa for the present-day . (d) Stress-induced deflection at the sediment basement interface.

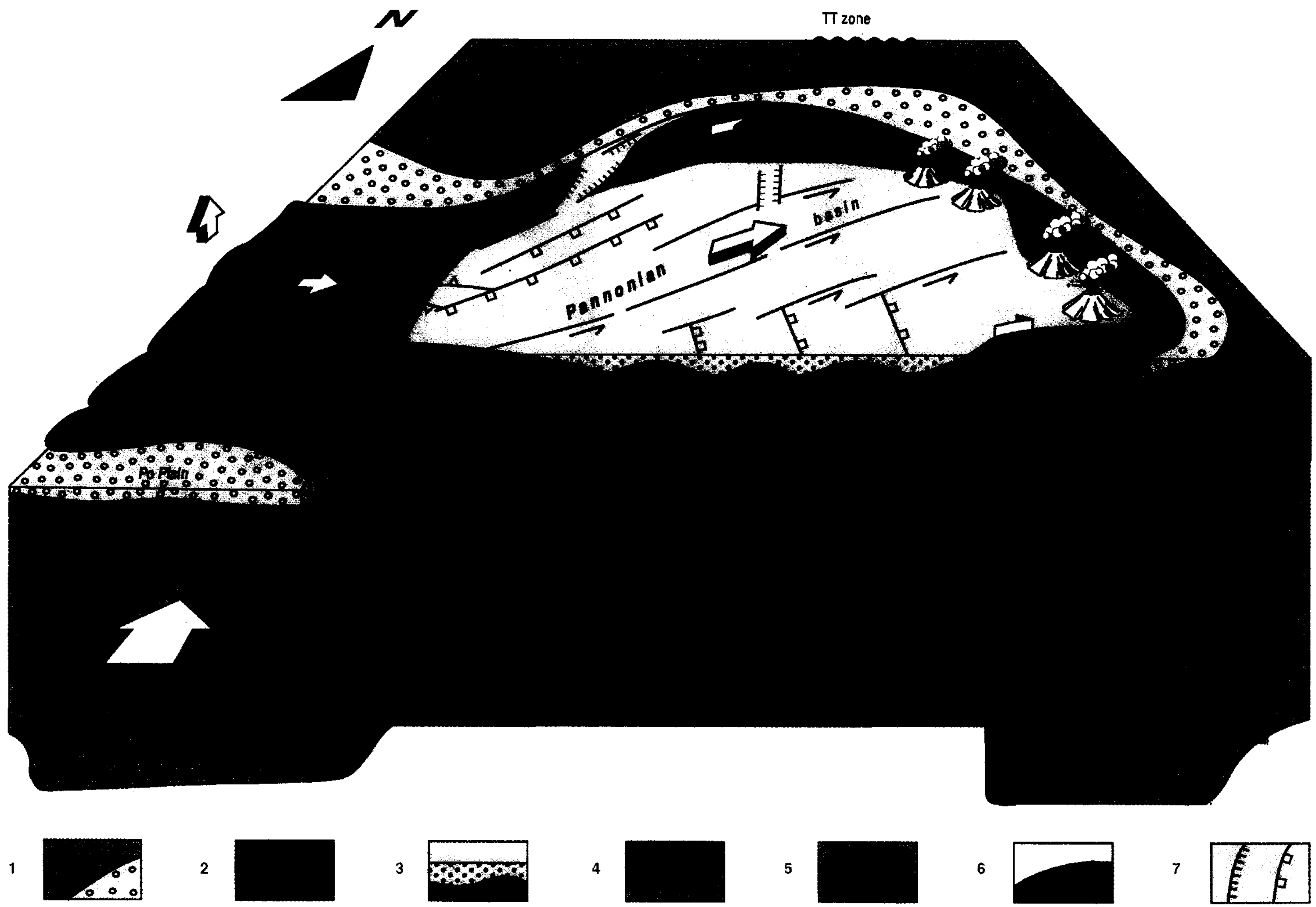

Fig. 6. Generalized block diagram to illustrate the geodynamic scenario of the Pannonian basin and surrounding orogenic belts a few million years ago. Subduction and rollback of the foreland lithosphere stopped, the Pannonian basin lithosphere became locked in a stable continental frame and, thus, the possibility for further extension ceased.

Ratschbacher et al. (1993) have verified by field observations and scaled-model experiments that in the Oligo–Miocene history of the Eastern Alps crustal thickening interacts with subhorizontal extension centred at topographic highs and acting toward the weakly constrained boundary of the Pannonian basin at the Eastern Carpathians. Extensional collapse of the Eastern Alps has been driven by gravity flow of the overthickened Alpine orogenic wedge. Wedge thickening and lateral extrusion has been caused and maintained by ongoing indentation and convergence between the European foreland and Adriatic promontory (Africa push in Fig. 6).

The weakly constrained boundary of the Pannonian basin at the Eastern Carpathians was actually a broad marine area where the Moldavian flysch and its transitional facies towards the foredeep molasse were deposited. This large marine basin was created by rifting and significant attenuation of a continental crust in Late Cretaceous through Oligocene time (Stefănescu, 1983). Its weakness was given by the fact that this area was easy to override and consume its lithosphere by westward-dipping subduction (Fig. 6). Royden et al. (1982) visualised this process as a passive subduction which occurred by lower plate rollback and steepening due to its negative buoyancy. Rollback was accompanied by formation of a frontal accretionary wedge from the sedimentary cover (flysch and molasse) and locally slices of the basement detached from the downgoing plate (Roure et al., 1993).

Extension of the Pannonian basin and Eastern Alps has been facilitated by slab pull along the Eastern Carpathians (see e.g., Wortel and Spakman, 1992, and Zoetemeijer and Cloetingh, 1996) and affected former orogenic terranes which have been squeezed towards the east from the Alpine collision zone. A dramatic change in this dynamic system occurred a few million years ago. Slab pull became inefficient as rollback of the subducted slab could hardly proceed further. This occurred because the attenuated crust of the former marine basin had been fully consumed, and the accretionary wedge reached the Tornquist–Teisseyre zone (TTZ) of the East European foreland (Fig. 6).

The TTZ is characterized by 50–55-km-thick continental crust (Guterch et al., 1983), and formed as a major Late Cretaceous/Early Tertiary suture between Variscan Western Europe and the Precambrian massifs of Eastern Europe (Ziegler, 1987). This preexisting structure acted as a barrier and halted the advance of the accretionary wedge and the flexural downbending of the foreland. From this time on, the Pannonian basin became completely locked in a stable continental environment without chance for further extension in any direction. This tectonic setting implies the development of a new stress field controlled by the Europe/Africa convergence. It seems reasonable to suppose that this stress field is the one we can observe now in the Pannonian basin (Fig. 1). 2-D modelling of stress-induced subsidence and uplift along an approximately E–W-oriented section has given a fair description of the observed subsidence and uplift pattern in the Pannonian basin, but is insufficient to explain the observed remarkable uplift in the Styrian and Transylvanian basins. This feature of the model is probably a consequence of the 2-D approximation, adopting a stress in the plane of the section which is obviously not the case in the western part of the Pannonian basin. Uplift in the Styrian basin (Sachsenhofer et al., 1997) has been generated presumably after locking of extension by the proximal collision between the Alps and the Adriatic promontory. There is an obvious relationship between the uplift of the Tauern Window, of which the continuing uplift is constrained by geodetic measurements. This can also be observed in fission track cooling ages within the Tauern Window. Late Pliocene convergence also occurred along the Periadriatic line, where the Post-Sarmatian Klagenfurt basin has been overridden by the tectonic elements of a positive flower structure (F. Neubauer, pers. commun., 1996). A different explanation can be offered for the observed uplift in the Transylvanian basin. Recent detachment of the subducted slab could have occurred with unloading followed by rapid isostatic uplift (Zoetemeijer and Cloetingh, 1996).

7. Conclusions

Analysis of recent data and 2-D modelling support the following tectonic scenario for the main phases of the evolution of the Pannonian basin:

(1) Lateral extrusion of orogenic wedges from the Alpine collision zone towards the east during the Oligocene through Early Miocene, and their overrid-

ing of the attenuated continental lithosphere of the Eastern Carpathian flysch basin.

(2) Extensional collapse of the upper plate during the Middle Miocene, subduction and rollback of the lower plate.

(3) Thermal phase of subsidence of the upper plate and steepening of the subducted slab during the Late Miocene and Pliocene.

(4) Termination of rollback and progressive detachment of subducted slab along the Carpathian arc, locking of the intra-Carpathian basin system and development of a new compressive stress field during the latest Pliocene and Quaternary.

This tectonic scenario makes it clear that evolution of the Pannonian basin, and most probably of many other intermontane basins, is largely controlled by stresses generated by lithosphere subduction and continental collision. These stresses can significantly modulate the subsidence and deformation history of the basins. Most remarkable features are accelerated subsidence and uplift during the post-rift phase, when the thin and hot lithosphere is rheologically weak and, hence, easy to deform by fluctuating intraplate stress. 2-D model calculations can give a fair description of the observed broad-scale stratigraphic architecture of the basin. Future 3-D simulations, incorporating new constraints on spatial variations of stress and rheological properties in the lithosphere, will probably be able to further fine-tune the predictions of the models for the effect of neotectonics on the Pannonian basin area.

Acknowledgements

The authors are grateful to Albert W. Bally for sharing with them his views on the recent compressional inversion of the Pannonian basin at an early stage of this work. Paul Reemst, Anco Lankreijer and Peter Szafián are gratefully acknowledged for technical assistance during the preparation of this paper. The authors are grateful to Giorgo Ranalli, Franz Neubauer, Matthias Gölke and Evgene Burov for their constructive comments on the manuscript of this paper. This research was supported by grants from the JOULE-II IBS programme and the National Science Foundation of Hungary (OTKA T 002131). Publication number 960604 of the Netherlands Research School of Sedimentary Geology (NSG).

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