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An outline of neotectonic structures and morphotectonics of the western and central Pannonian Basin

László Fodor [a,b,*], Gábor Bada [b,c], Gábor Csillag [a], Erzsébet Horváth [c], Zsófia Ruszkiczay-Rüdiger [b,c], Klára Palotás [a], Ferenc Síkhegyi [a], Gábor Timár [d], Sierd Cloetingh [b], Frank Horváth [c]

[a] Geological Institute of Hungary, 1143 Budapest, Stefánia 14, Hungary
[b] Netherlands Research Centre for Integrated Solid Earth Science (ISES), Vrije Universiteit, Amsterdam, The Netherlands
[c] Institute of Geography and Earth Sciences, Eötvös University, 1117 Budapest, Pázmány Péter Sétány 1/c, Hungary
[d] Space Research Group, Institute of Geography and Earth Sciences, Eötvös University, 1117 Budapest, Pázmány Péter Sétány 1/a, Hungary

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Abstract

Neotectonic deformation in the western and central part of the Pannonian Basin was investigated by means of surface and subsurface structural analyses, and geomorphologic observations. The applied methodology includes the study of outcrops, industrial seismic profiles, digital elevation models, topographic maps, and borehole data. Observations suggest that most of the neotectonic structures in the Pannonian Basin are related to the inverse reactivation of earlier faults formed mainly during the Miocene syn- and post-rift phases. Typical structures are folds, blind reverse faults, and transpressional strike-slip faults, although normal or oblique-normal faults are also present. These structures significantly controlled the evolution of landforms and the drainage pattern by inducing surface upwarping and river deflections. Our analyses do not support the postulated tectonic origin of some landforms, particularly that of the radial valley system in the western Pannonian Basin. The most important neotectonic strike-slip faults are trending to east-northeast and have dextral to sinistral kinematics in the south-western and central-eastern part of the studied area, respectively. The suggested along-strike change of kinematics within the same shear zones is in agreement with the fan-shaped recent stress trajectories and with the present-day motion of crustal blocks derived from GPS data.

Keywords: Neotectonics; Landscape evolution; Drainage patterns; Folds; Pannonian Basin; Quaternary

* Corresponding author. Geological Institute of Hungary, 1143 Budapest, Stefánia 14, Hungary.
E-mail address: fodor@mafi.hu (L. Fodor).

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1. Introduction

The Pannonian Basin is situated within the Alpine, Carpathian, and Dinaric mountain belts (Fig. 1). The basin system was formed due to lithospheric extension during the Late Early to Late Miocene times (Royden and Horváth, 1988). Based on subsidence analysis and geophysical data, Royden et al. (1983) separated the syn- and post-rift phases, although recent structural analyses revealed a more complex evolution (Fodor et al., 1999; Horváth et al., in press). The syn-rift phase of ~18–12 Ma resulted in the formation of numerous grabens filled with relatively thin syn-rift sediments of marine to brackish origin (Royden and Horváth, 1988; Tari, 1994; Csontos, 1995; Fodor et al., 1999). Crustal extension was followed by a post-rift phase characterised by thinning and updoming of the lithospheric mantle, thermal contraction, and related subsidence of the entire basin (Horváth and Royden, 1981). Post-rift subsidence was compensated by intense sedimentation in the brackish to freshwater Lake Pannon during the Late Miocene (Jámbor, 1989; Juhász, 1991). Deltaic to littoral sediments progressively filled up the lake, while all sub-basins became fluvial dominated by the end of the Miocene (Vakarcs et al., 1994).

The cessation of syn-rift faulting was ultimately related to the end of thrusting along the Carpathian arc (Horváth et al., in press), which occurred in the early Late Miocene in the Eastern Carpathian segment (~9–10 Ma, Maţenco and Bertotti, 2000). At the same time, however, the northern push exerted by the Adriatic microplate and juxtaposed Dinaric units continued from the south. As the Pannonian Basin lithosphere had no free space to further extend to the east, a new phase of deformation started. This phase was termed as "inversion of the Pannonian Basin" (Horváth, 1995; Bada et al., 1999), which term is often used as an equivalent for the 'neotectonic phase' of the area and will be adopted in the present paper.

The neotectonic deformation is connected to the uplift of vast areas in the Pannonian Basin (Horváth and Cloetingh, 1996). These regions form rolling hills and low mountains and incorporate the North Hungarian Range, and almost the whole Transdanubia, a region between the Danube and Drava rivers and the Eastern Alps. The uplifting regions surround major plains, which still show subsidence and continuous sedimentation (Great Hungarian Plain, Danube, Drava and Vienna Basins, Fig. 1). The uplifting areas underwent complex denudation processes (Kretzoi and Pécsi, 1982), interrupted only by local terrestrial sedimentation. In consequence, neotectonic deformation and landscape evolution were intimately linked in the Pannonian Basin, and all process can be understood only in a combined way. The presence of deformed and non-deformed landforms offers valuable geomorphologic tools for neotectonic analysis. On the other hand, the lack of syn-deformational marine or lacustrine sediments hampers the determination of structures and their timing. Thus the application of 'classical' structural geological methods alone is not sufficient to establish neotectonic structural pattern.

In our paper we present a multidisciplinary approach for neotectonic research from three different areas within the uplifting ranges. We analysed subsurface structural pattern within the uplifted post-rift sequence, then compared to surface structural and geomorphologic data. Our objectives were to reveal structural meaning of geomorphic indices and to understand the role of structural deformation on landscape evolution. Our combined morphotectonic approach seems to be essential to understand the Quaternary evolution of the Pannonian Basin and may be useful for neotectonic research in other areas as well. In the Zala Hills, located in the southwest, connection of prominent neotectonic folds and their influence on surface morphology can be evaluated. The Vértes Hills are situated at the axis of the elevated Transdanubian Range, and show an example for the transition from post-rift tension to neotectonic compression. Recognition of neotectonic elements in the easterly located Gödöllő Hills are hampered by a thick loess cover but assisted by good subsurface data set and expressed geomorphic indices. The conclusions of our analysis contribute to the more precise timing of the onset of neotectonic phase, which still represent an open question.

2. Geological setting and general aspects of neotectonic deformation

Neotectonics of the Pannonian Basin has been mainly characterised by contractional deformation

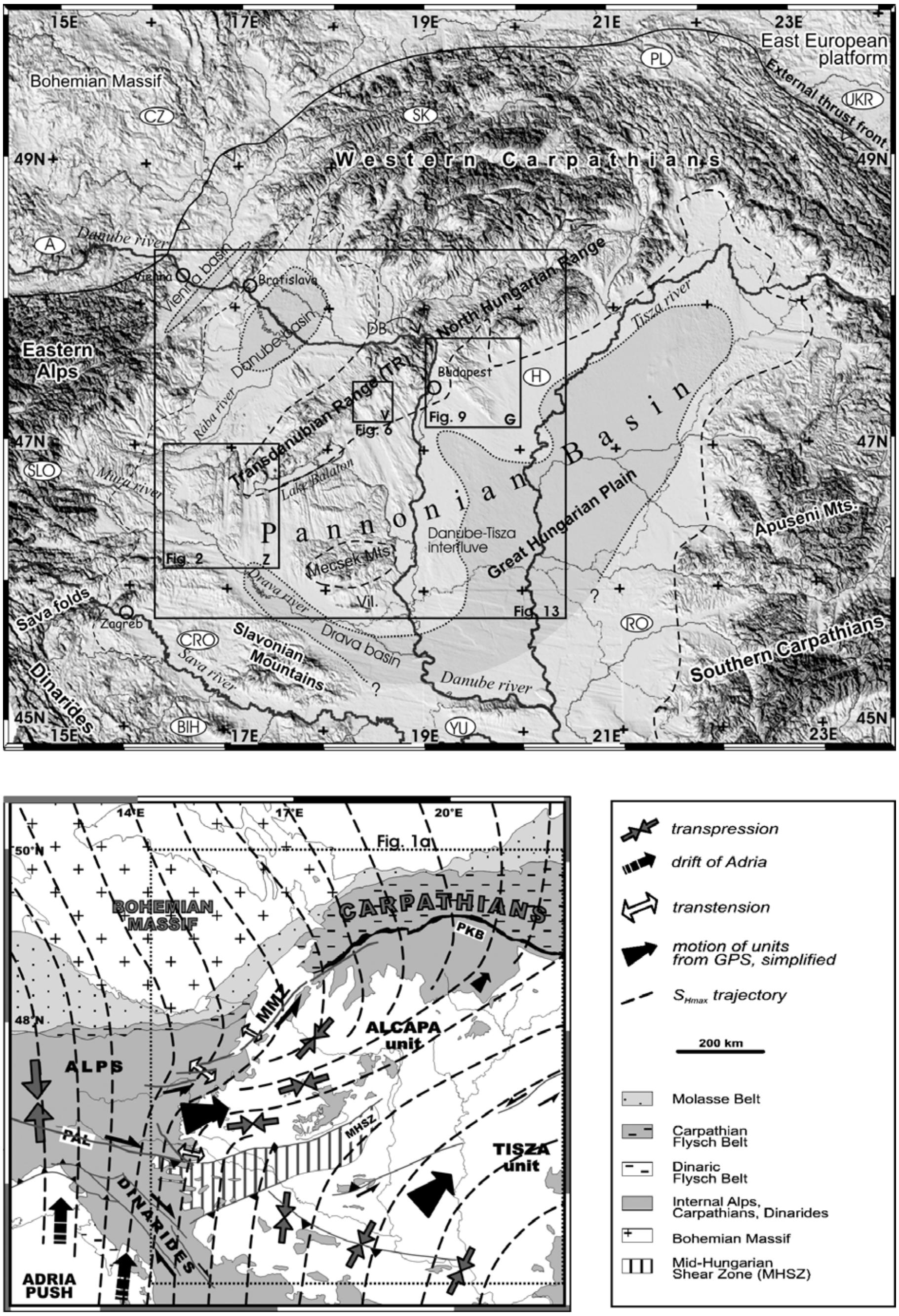

Fig. 1. Digital elevation model of the Pannonian basin and surrounding orogens. DB: Danube Bend area. Dashed line indicates the boundary of the thick post-rift sediments. Dotted line shows the area where Quaternary is more than 100 m thick. Rectangles with Z, V, G indicate the three study areas, the Zala, Vértes and Gödöllő Hills, respectively, shown in detail on Figs. 2, 6 and 9. Vil.: Villány Hills. Fig. 1b shows simplified recent stress trajectories, GPS velocities and some major structures (after Bada et al., 2001; Grenerczy and Kenyeres, 2004). Location of Fig 1a is indicated by dotted frame. MMZ, PAL, PKB: Mur-Mürz Line, Periadriatic Line, Pieniny Klippen Belt.

(Horváth, 1995; Bada et al., 1999). This structural style was derived from scattered outcrop and subsurface structural observations, extrapolation of data on recent deformation, and general trends of vertical surface motions. Pure contractional structures (folds and blind reverse faults) occurred in the south-western part of the Pannonian Basin, in the eastern continuation of the 'Sava fold belt' (Pávai Vajna, 1925; Dank, 1962; Horváth and Rumpler, 1984; Horváth et al., in press). Structures are transpressional around the Mecsek–Villány Mts. (Benkovics, 1997; Wórum, 1999; Csontos et al., 2002) and in the Slavonian Mts. (Jamičić, 1995; Tomljenović and Csontos, 2001), as indicated by fault kinematic studies.

The analysis of earthquake distribution, their focal mechanisms, contemporaneous stress directions, and GPS data represent the main tools to describe active deformation of the area. Stress trajectories show a fan-shape pattern of the maximum horizontal stress directions and trend to the N–NNE in the northern Dinarides and southern Pannonian Basin then gradually turning to the NE in the eastern territories and to E–W towards the Eastern Carpathians (Gerner et al., 1999). The Pannonian Basin is characterised by moderate seismicity (Zsíros, 2000). Earthquake focal mechanism solutions indicate that thrust and strike-slip faulting is dominant, although normal faulting also occurs mainly in the central Pannonian Basin (Tóth et al., 2002). Results of numerical stress modelling suggest that an active push from the south (indentation of the Adriatic microplate), the complex geometry and density distribution of the surrounding orogenic belts provide the main boundary conditions for the style of the recent stress field (Bada et al., 1998, 2001). Recent crustal velocities determined by GPS campaigns clearly indicate active deformation in the Pannonian Basin. A differential motion of 1–1.3 mm/year is accommodated within the central-eastern Transdanubian Range (Grenerczy et al., 2000), between the three study areas.

Differential vertical motions resulted in remarkable thickness variations of Plio–Quaternary sediments (Rónai, 1974). Vertical motions are also reflected by the style of sedimentation (Jámbor, 1998; Síkhegyi, 2002) and supported by fission track data (Dunkl and Frisch, 2002). In the Danube, Drava and Sava basins, and below the Great Hungarian Plain subsidence continued after the Miocene, which resulted in a continuous, alluvial Pliocene and Quaternary succession (~1000 m and ~300–500 m thick, respectively, Rónai, 1985).

All other areas underwent uplift and preserve less than 100 m thick Plio–Quaternary formations (Síkhegyi, 2002). In the more elevated Transdanubian and North Hungarian Ranges, as well as in the Mecsek–Villány Mts. the Pliocene is marked by a stratigraphic hiatus. The thickness of Quaternary deposits is only a few metres and they mainly consist of slope deposits or a thin loess cover. It is to note that the Late Miocene post-rift suite was already reduced over these ridges. Between these areas and the still subsiding plains a "transitional zone" can be defined. Here the post-rift thickness reaches 1–2 km, and the Plio–Quaternary suite can be 30–80 m. This sequence locally starts with red clay beds (Tengelic Fm.) (Schweitzer and Szöőr, 1997; Jámbor, 1998), covered by a loess–paleosol sequence with minor fluvial and eolian sand intercalations (Pécsi and Richter, 1996).

The "transitional zones" indicate that the locations of expressed Plio–Quaternary and Late Miocene subsidence do not always coincide. This shift in the location of deposition centres and the results of subsidence modelling led Horváth and Cloetingh (1996) and Cloetingh et al. (1999) to conclude that Plio–Quaternary subsidence is not the continuation of post-rift subsidence. Instead, it was induced by intraplate compression and, hence, the spatial distribution of subsiding and uplifting areas reflect lithospheric folding and compressional neotectonic deformation on a large wavelength.

The onset of uplift of the basin margins (and the related neotectonic folding) is confined between the youngest post-rift fluvial–deltaic sediments and the oldest Pliocene and/or Quaternary cover. However, the age of the topmost pre-Quaternary strata is not easy to define. Post-rift lacustrine and overlying deltaic to fluvial sediments show decreasing age trends towards to southern Pannonian Basin from ca. ~9 to 5 Ma (Vakarcs et al., 1994; Magyar et al., 1999; Sacchi and Horváth, 2002). Scattered bio- and sequence stratigraphic data constraint the end of fluvial–terrestrial sedimentation to Latest Miocene or locally to Early Pliocene (~7–4 Ma). The oldest sediment deposited after the onset of the uplift is the Late Pliocene(?) to Early Quaternary Tengelic red clay (Koloszár and

Marsi, 1999). The overlying loess sedimentation was constrained as 1–1.2 Ma in the "transitional areas", (Pécsi and Pevzner, 1974; Koloszár and Lantos, 2001), but loess sedimentation can be much younger, ca. 0.5–0.02 Ma in the mountain ranges (Frechen et al., 1997). Thus the hiatus between ~7–4 to 1.2 (or 0.5) Ma occur, which includes the possible age of the transition from post-rift to neotectonic phase.

In the uplifting areas, this time span is mainly represented by landforms instead of sediments. The most typical landforms of the western and central Pannonian Basin are the long, linear "meridional" valleys. Their length is variable with a maximum of about 50 km, while width ranges from 0.5 to ~5 km. At a smaller scale, the valleys are rather parallel and are organised with regular spacing. However, on the scale of the whole Transdanubia, they show an almost radial (fan-shaped) pattern (Fig. 1, Lóczy, 1916; Cholnoky, 1920). In south-western Transdanubia valleys are mainly N–S trending. Near Lake Balaton valleys are oriented to NNW–SSE while further to the NE up to the Gödöllő Hills the valley trends change to NW–SE.

Other typical features are the west–northwest trending "longitudinal" valleys (Bulla, 1958; Marosi, 1962) and similarly directed flat depressions often filled with lakes and/or wetlands (like Lake Balaton, Fig. 1). Shape of the "longitudinal" valleys varies from linear to curved or zigzagged. Their origin was attributed to progressive beheading of the drainage elements (Erdélyi, 1961).

The origin of drainage pattern represents a century-long scientific debate in the Hungarian literature. Several authors suggested neotectonic control on the formation of the valleys and lake depressions (Szilárd, 1967; Ádám et al., 1969; Marosi and Szilárd, 1981; Brezsnyánszki and Síkhegyi, 1987; Gábris, 1987; for a discussion see Gerner, 1994). However, exact nature of tectonic control was not described only vaguely attributed to 'faults' or 'fractures'. Because landforms occur in post-Middle Miocene formations, Plio–Quaternary tectonic activity was inferred.

3. Methods

In order to identify neotectonic structures and their possible morphological expression, we applied a multidisciplinary approach. Industrial seismic reflection profiles have been used to recognise structures, which deform the post-rift sediments and, particularly, the highest imaged horizons. These reflectors are located on the studied sections mainly between 100 to 400 m below surface, above which we have no direct evidence for deformation. The thickness of visible post-rift sediments are always much larger than the non-imaged part, which represents only a small portion (less than ~1 Ma) of the post-rift phase. If some structures terminate within the non-imaged sedimentary units, they might have formed ~1 Ma earlier than the last post-rift sediments of the given location. Thus we consider that the reconstructed structures are neotectonic, although a "pre-neotectonic" timing cannot always be excluded, ranging between ~7–4 Ma in the study areas.

We locally used boreholes to fill information gaps between the surface and the highest imaged post-rift reflectors. In the Vértes Hills a new surface geological map was applied to describe tectonic and geomorphic features. We also investigated fractures (mainly outcrop-scale faults) in the post-rift sediments. With the aid of kinematic indicators, paleostress fields have been locally determined.

We compared the constructed sub-surface tectonic maps to the surface morphology. Morphotectonic elements were derived from traditional topographic maps, digital elevation models (DEM) and field morphotectonic observations. Particular attention was paid to anomalies in the drainage network, such as sudden changes in river course, reversal of flow direction, river capture, as these features are indicative of young structural deformation (Holbrook and Schumm, 1999; Timár, 2003). Steep slopes, elevated hills or larger flat-bottomed depressions may also have tectonic significance. Displaced or folded past landforms, like river terraces, planation surfaces also indicates deformation (Pinter and Keller, 1995; Pinter et al., 2003). Coincident subsurface structural pattern and surface morphological features enabled us to validate the tectonic control on landforms. However, it is not always possible to differentiate between active control of ongoing deformation and passive control of older, inactive structures.

4. New results on neotectonic–geomorphic evolution

4.1. Zala Hills

4.1.1. Morphology and origin of the "meridional" valleys

The Zala Hills occupy the south-western part of Hungary, between the Zala, Mura rivers, the Keszthely Hills and the Marcali ridge (Fig. 2). This area is of key interest for neotectonic study as it contains the best-developed "meridional" valleys and offers the best documented folding of the post-rift sequence (Pávai Vajna, 1925; Dank, 1962; Horváth and Rumpler, 1984). Higher density of valleys and more dissected topography of the Zala Hills with respect to its surroundings (Figs. 1 and 2) were interpreted as signs of active uplift due to Quaternary folding (Kovács, 1999; Síkhegyi, 2002).

Most prominent geomorphic features of the Zala Hills are the ~N–S trending wide "meridional" valleys and intervening ridges (Fig. 2). These valleys were frequently attributed to fluvial erosion associated to tectonic lineaments (Bulla, 1958; Somogyi, 1961;

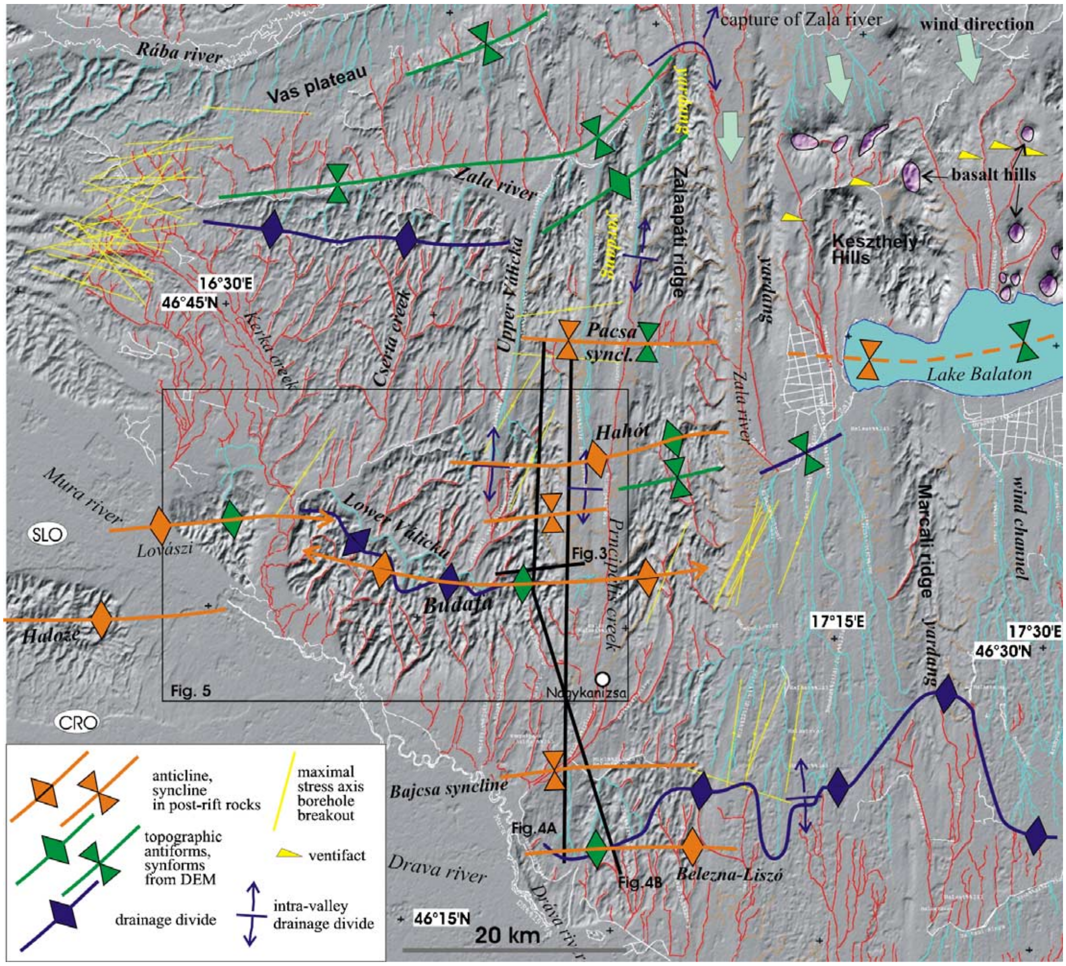

Fig. 2. Post-rift folds, topographic lows and highs, watersheds placed on top of the drainage network and digital elevation model. Colours of drainage indicate flow direction; red: flow to south, blue: flow to north, pale brown: variable flow direction, white: flow to east or west (Gerner, Palotás unpublished data). Yellow sticks indicate maximal horizontal stress direction inferred from borehole breakout data (Windhoffer et al., 2001). Lines A and B show the location of seismic profiles and topographic section of Fig. 4. Inset shows location of Fig. 5. For location see Fig. 1.

Pécsi, 1986). Gravel occurrences on some flat-lying steps of valley slopes were classified into a terrace system, although their fluvial origin was not always demonstrated. Currently the lower Zala River, other small creeks, wetlands and swamps occupy these valleys. Presence of low topography intra-valley drainage divides causes opposite flow directions (Fig. 2).

Tectonic origin of the "meridional" valleys was suggested mainly on the basis of their pronounced linearity (Jámbor et al., 1993; Síkhegyi, 2002). However, several seismic reflection profiles across the valleys and ridges show undisturbed post-rift (Late Miocene) sediments up to the highest imaged horizons (Fig. 3). The resolution of the seismic profiles would permit to establish faults with 20 m offset, but such structures cannot be traced beneath the valley sides. The lack of noticeable faults was already emphasized by mapping geologists Strausz (1942). We cannot exclude the presence of small faults or joint systems along valley margins. Nevertheless, we propose the lack of considerable fault control and suggest a purely erosional origin of the meridional valley system.

Ridges between the "meridional" valleys usually terminate against the upper Zala River towards the north, where they are narrowing to a spearhead shape. Towards the south most ridges are gradually lowering

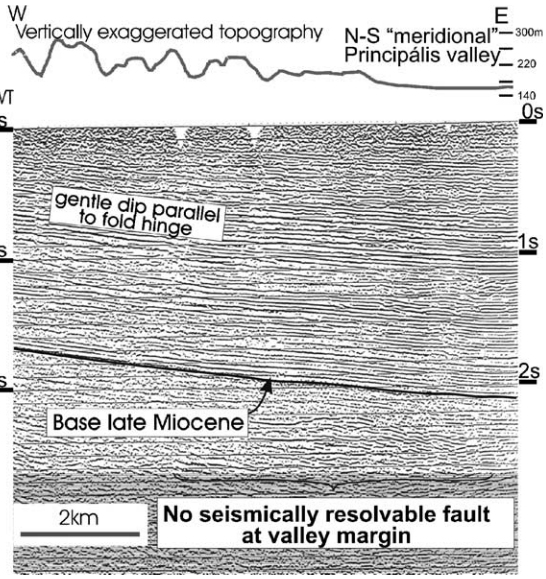

Fig. 3. E–W seismic reflection profile crossing the western boundary of the 'meridional' Principális valley. Note lack of faults near valley boundary.

and thinning (Fig. 2). Boundaries of the ridges are straight or gently curved. These features led Cholnoky (1920, 1936) and Pávai Vajna (1922) to propose that the ridges represent large-scale yardangs, residual landforms of deflational origin, a view what we share with previous authors.

4.1.2. Neotectonic analysis of the Zala Hills

For neotectonic analysis, we analysed some selected seismic reflection profiles and compared the observed structures to digital elevation model and drainage network. In Fig. 4 we present a N–S trending section, typical for the area, where three major anticlines and intervening synclines can be detected. The anticlines have divers origin, connected to transpressive or pure strike-slip faults or to the inversion of a syn-rift half-graben (Belezna–Liszó anticline, Hahót high, Budafa anticline, respectively; Pávai Vajna, 1925; Dank, 1962; Horváth and Rumpler, 1984; Jósvai, 1996; Csontos and Nagymarosy, 1998; Horváth et al., in press). However, all anticlines deform the complete post-rift sequence including the highest imaged horizons thus they represent neotectonic elements. All of them have ~E–W axes and the Budafa and Lovászi anticlines have overlapping periclinal terminations.

Recent stress data measured in southern Zala Hills corroborate neotectonic folding with E–W axis (Fig. 2, Windhoffer et al., 2001). However, an approximately E–W compression prevails NW from the Zala Hills. This latter direction is difficult to reconcile with folding along E–W axes, which occur in the south. One can attribute E–W compression due to gravity forces originated by the uplifting Eastern Alps (Bada et al., 2001) or to the presence of a major shear zone, which bound crustal blocks of different motion characteristics.

We adjusted a smoothed surface (an "envelope surface"), which wraps a topographic profile near the seismic line (location on Fig. 2). Topographic highs and lows of the envelope surface correspond to subsurface anticline and syncline hinges, a fact already observed by Horváth (1995) (Fig. 4). This correspondence suggests that topographic variations are the result of surface folding, and the fold shape can be approximated by the envelope surface.

Geological mapping of the Zala Hills revealed the presence of numerous gravel occurrences at different

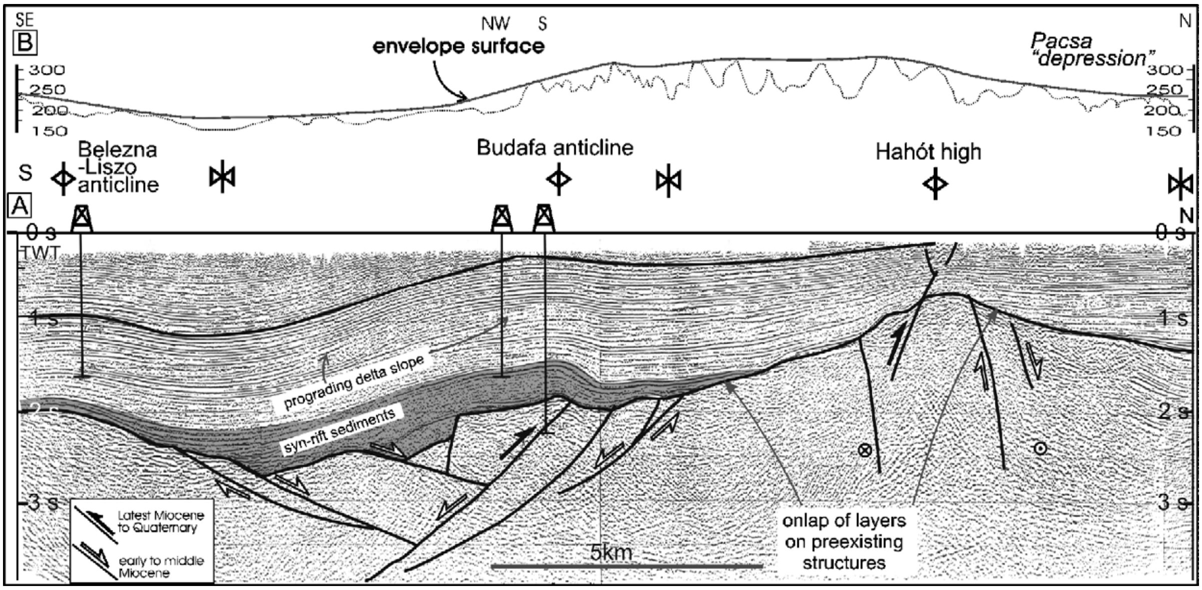

Fig. 4. N–S seismic reflection profile made in the Principális valley showing major folds, which affected the whole post-rift sequence. The topographical section (B) is parallel but slightly shifted westward. Note high vertical exaggeration in topographical section with respect to two-way-travel-time scale of the seismic profile.

topographic positions, but excluding the highest elevations near the crest of the Budafa anticline (Strausz, 1949 and own observation). Pebble composition and roundness data suggest that all gravel occurrences belong to a former gently dipping alluvial fan of the ancient Rába, and/or Drava rivers. Variations in elevation of the remnants of the former gravel carpet reflect deformation since the formation of the alluvial fan. Our qualitative estimates say that the envelope surface of the topography broadly matches a surface adjusted to the locations of the former gravel carpet. Thus we consider the undulations (folding) of the envelope surface as representative for the deformed geometry of the gravel carpet.

The subsurface and topographic folds (and the undulations of the gravel carpet) differ in fold amplitude. Largest subsurface fold amplitude is around 1.5 km (west from Fig. 4, Jósvai, 1996) while maximum topographic folds are around 250 m in amplitude. This difference demonstrates periods of folding before and after the formation of the alluvial fan. The ratios of fold amplitudes also suggest that pre-gravel folding was larger. This time marker, the age of the alluvial fan gravel is poorly constrained, between Late Pliocene and Early Quaternary (Strausz, 1949).

4.1.3. Drainage network and deformation

We also compared the distribution of subsurface and topographic folds to drainage network on map view (Fig. 2). The closely matching subsurface and topographic anticline hinges frequently correspond to drainage divides, both on the ridges and at intravalley position (Gyékényes–Liszó, Budafa, Hahót anticlines). Position of synclines is in accordance with lower topography on the ridges and also matches locations of wetlands, peat accumulations in wider valley sections (Bajcsa, Pacsa synclines). These widened valley segments may indicate lateral migration of creeks due to local fold-related subsidence.

On the other hand, some major creeks like the Principális channel, the Mura River and the Kerka creek continue their course across anticlines without apparent modification or flow reversal. The intravalley drainage divides west and east of the Marcali ridge have no clear corresponding subsurface structure. We attribute these anomalies to the variation in erosion capacity of rivers and/or deflating wind. These processes, alone or combined, can be powerful enough to counterbalance the effect of subtle surface warping.

Another important observation is that the southward flowing Lower Válicka, Cserta and Kerka creeks are deflected against the northern rim of the Budafa anticline and turn westward (Fig. 5). The deflected courses of the creeks follow subsurface structural geometry as they flow in front of the northern limb of the overlapping Budafa and Lovászi anticlines. This diversion was probably caused by the continuous westward propagation of the Budafa anticline, which have asymmetric shape with long southern limb and short, steeper northern limb.

A topographic profile along the anticlinal crests shows several broad valleys crossing the Budafa and Lovászi anticlines. Some of them are too wide to be attributed to erosion of headward propagating small creeks. We interpret these valleys as wind gaps, a term used for dry valleys abandoned due to progressive folding (Burbank and Anderson, 2001). Wind gaps of the Budafa anticline could be incised primary by the former Lower Válicka and Cserta creeks when they still crossed the anticline hinge (Fig. 5). The

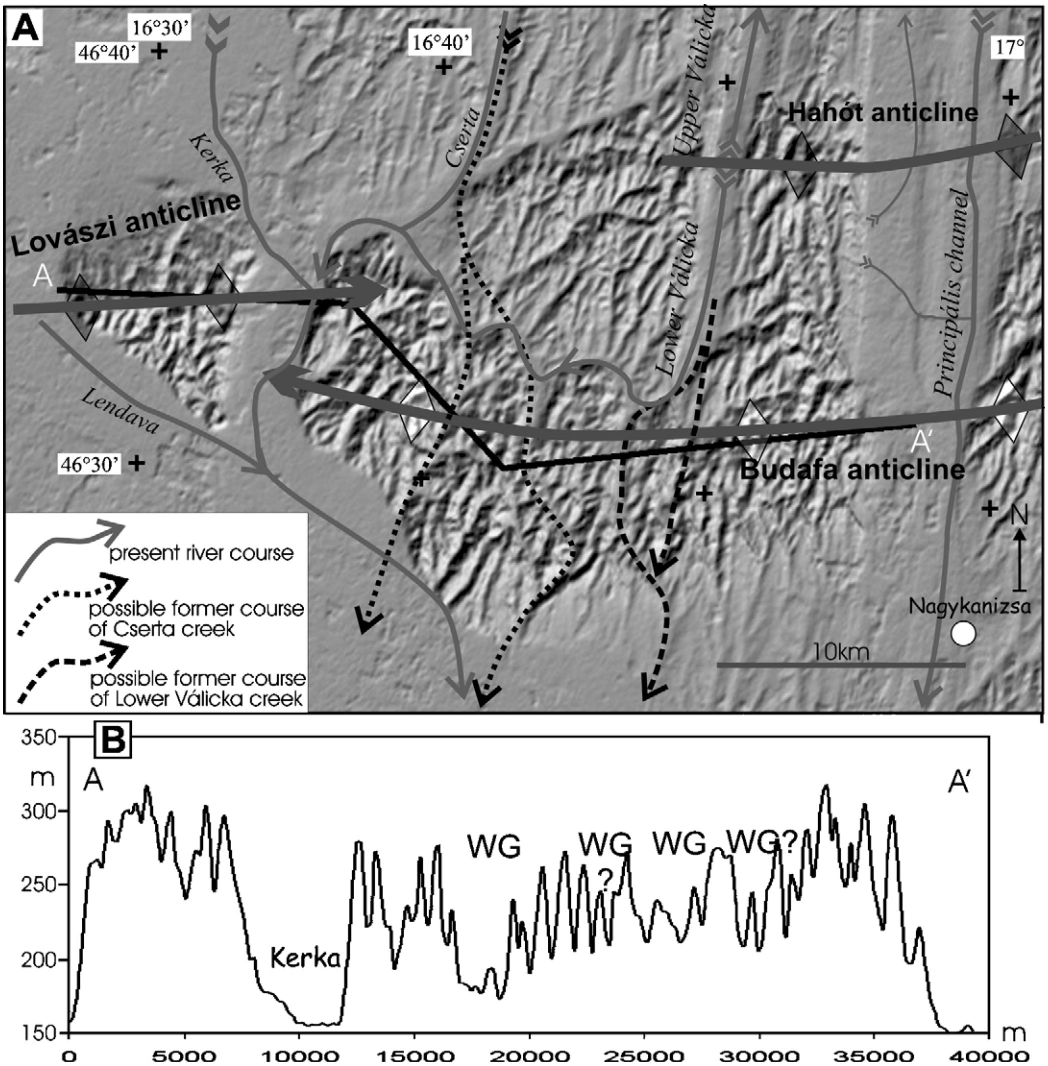

Fig. 5. Digital elevation model of the southern Zala Hills above the Budafa and Lovászi anticlines. Note deflected present-day Lower Válicka and Cserta creeks against uprising anticlines. Different lines indicate possible former courses of the Alsó–Válicka and Cserta creeks. B) Cross-section placed closely along-strike of the fold hinges (A–A′ on Fig. A). Note wide valleys, which are interpreted as wind gaps (WG).

abandonment of straight N–S valleys was likely due to amplification of the Budafa and Lovászi anticlines. Only the integrated stream power of the deflected Kerka, Cserta and Lower Válicka creeks were enough to counterbalance tectonic uplift and to keep incising across the Lovászi anticline. The occurrence of wind gaps and defeated creeks around laterally propagating folds are similar to other well-documented examples of active folds worldwide (Burbank and Anderson, 2001; Pinter et al., 2003; Giamboni et al., 2004).

4.2. Late Miocene to Quaternary faulting in the Vértes Hills

The Vértes Hills are located in the central part of the Transdanubian Range (Fig. 1). Tilted Mesozoic carbonates, locally covered by sub-horizontal Eocene sediments, form the elevated central part of this area. During the Early and Middle Miocene the Vértes Hills were dissected by normal and strike-slip faults (Fodor et al., 1999). During these deformation phases the Vértes Hills suffered denudation under sub-aerial conditions (Kaiser, 1997). Consequently, a sub-horizontal denudation surface originally of Cretaceous age was exhumed and slightly modified. These flat landforms dominate the highest part of the Vértes Hills and marked in this study as P5 surface (Fig. 6).

Small occurrences of Late Miocene sediments suggest a temporal lacustrine transgression over the entire area (Csillag et al., 2004). Terrestrial sediments of poorly constrained Late Miocene to Early Pliocene age (Vértesacsa Formation) covered the lacustrine to deltaic Pannonian sequence. Post-Miocene denudation wrapped off the Pannonian to Early Pliocene (?) cover from the Vértes Hills and re-exhumed the complex P5 Cretaceous–Miocene denudation surfaces. P5 surfaces on Triassic dolomite remained relatively intact. On the other hand, at least four pediment surfaces (P4–P1) were cut into Late Miocene sediments in the foreland of Vértes below the level of regional P5 surface (Kaiser, 1997; Csillag et al., 2002). P2 and P3 pediments covered by a thin dolomite colluvium or loess can be traced over several kilometres while P4 and P1 have limited aerial extent (Fig. 6).

Two major fault zones deform the Triassic to Late Miocene rocks (Fig. 6). The western zone separates the Eastern Vértes Ridge (EVR) from the central part. The eastern one represents the eastern fault scarp facing the Vértes foreland. The main fault strands are trending north to north-east, and have curved geometry. They have normal kinematics according to the observed small-scale faults and rare slickensides (Fig. 6b, c). The main fault segments are connected by NW trending fault splays that had dextral-normal kinematics as shown by slickensides (Fodor et al., 2005).

Some major faults are also located in the Vértes foreland buried by Late Pleistocene loess or colluvial sediments. The buried faults and the exposed scarp together form the Eastern Vértes Fault Zone (EVFZ). A borehole-based geological cross-section shows the temporal evolution of the different fault branches (Fig. 7). Activity along the Eastern Vértes Fault Zone initiated in the Late Middle Miocene as indicated by facies changes across fault branches. During the Late Miocene lacustrine transgression, the EVFZ evolved into a fault-controlled abrasional cliff indicated by conglomerates and breccias draping the fault scarp and by some syn-sedimentary dykes (Site Cs on Fig. 6). Some fault branches displace the Vértesacsa Fm. and may have Pliocene to Early Quaternary age. Young Quaternary slip can be excluded, because denudation surfaces, e.g., the P2 pediment surface clearly cover some of the fault branches without displacement. More to the south-east, the Csaplár Fault runs along the north-western slope of the Nyárjas Hill. Two phases of displacement were reconstructed along this structure. The older normal (or oblique-normal) slip downfaulted the Vértesacsa Fm. that now occurs only in hanging wall position. During this first phase, kinematics of the Csaplár Fault can be sinistral-normal extrapolating data from the EVFZ.

On the other hand, geomorphic observations suggest a second phase of motion along the Csaplár Fault. Its hanging wall, the Nyárjas Hill is somewhat higher (20–50 m) than its surroundings and disturbs the drainage pattern in the Vértes foreland, where most of the linear creeks flow south-eastward. However, some creeks near the Nyárjas Hill turn to the north-west. Dolomite clasts derived from the eastern Vértes scarp were found on top of the Nyárjas Hill (Fig. 6). These clasts have likely been transported along a denudation surface. We assume that this surface belongs to the oldest surface P5 found also on top of the dolomites of the Vértes Hills.

We reconstructed the possible geometry of the old P5 denudation surface descending from the Vértes

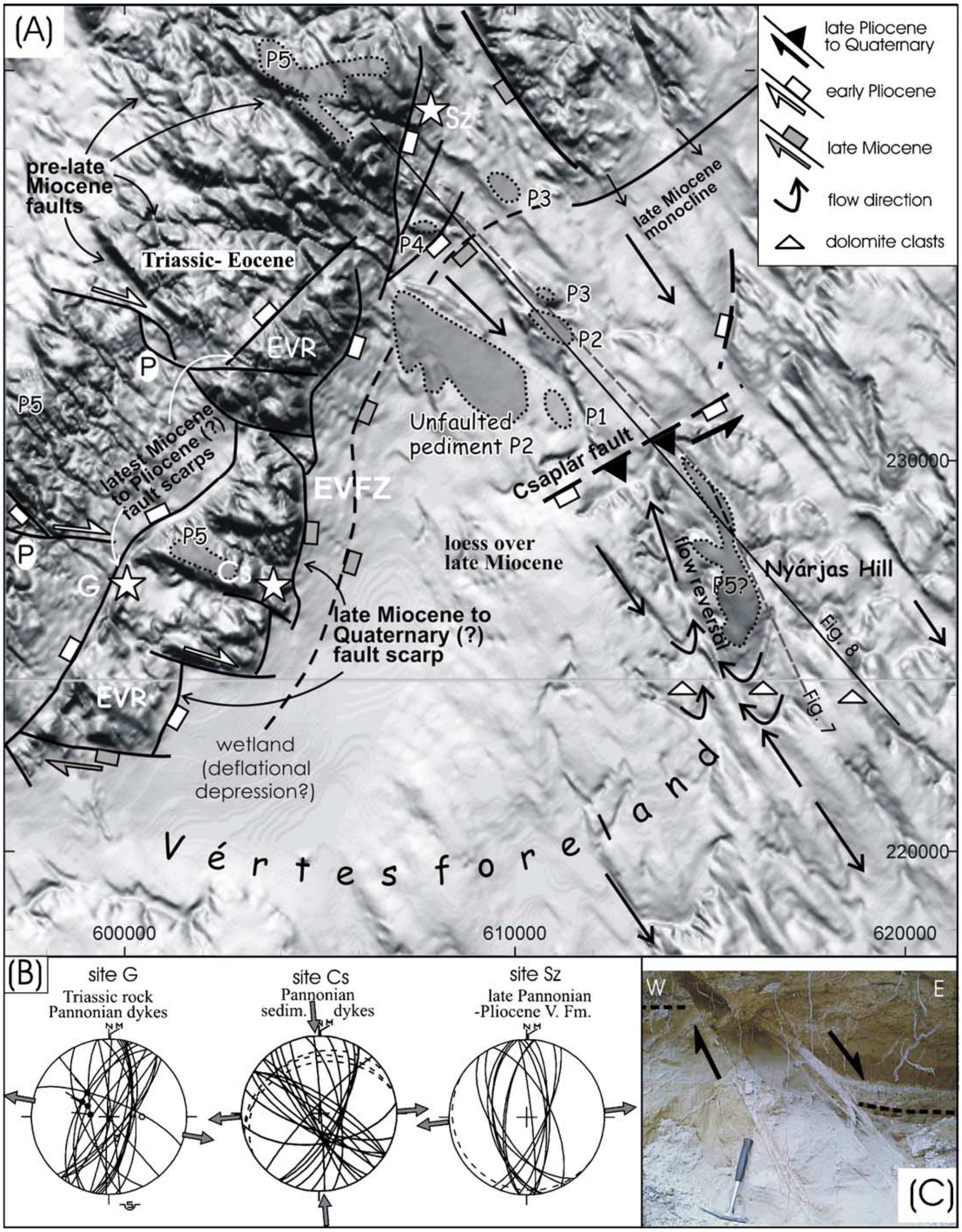

Fig. 6. Digital elevation model of the Vértes Hills and its SE foreland (For location see Fig. 1). Note range-bounding fault system, which displaces the post-rift Late Miocene to Early Pliocene(?) sediments (after Csillag et al., 2004; Fodor et al., 2005). Numbers at margins indicate distances in metres using the Hungarian grid EOV. EVFZ: Eastern Vértes Fault Zone; EVR: eastern Vértes ridge; P: pull-apart basins; P1–P5: pediment surfaces. G., Cs, Sz with stars indicate location of fault data shown on Fig. 6B. B) Stereograms showing lower hemisphere projections (Schmid net) of outcrop-scale faults with estimated stress axes (grey arrows). Age of deformed rock is indicated. Fractures at site Cs and partly G are syn-sedimentary. C) Field example of outcrop-scale faults in site Sz, affecting the Vértesacsa Fm. Dashed lines show displaced bed across fault.

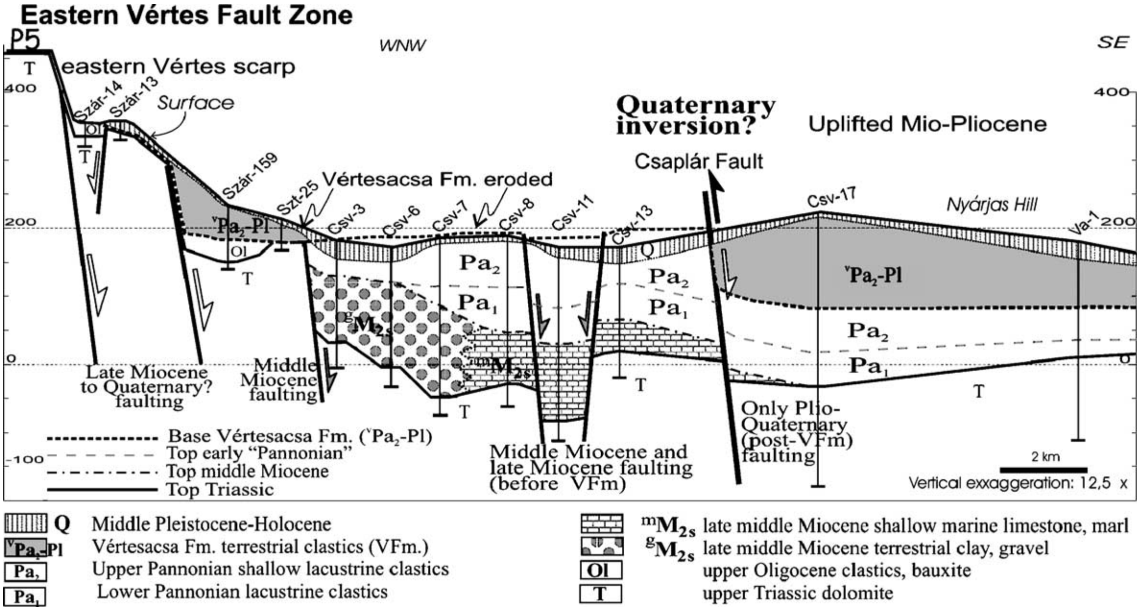

Fig. 7. Geological cross-section of the SE foreland of the Vértes Hills parallel to older sediment transport direction, after Csillag et al. (2002), modified. Grey, white and black arrows indicate Mid- to Late Miocene, early Pliocene and late Pliocene to Quaternary slip, respectively.

scarp to the Nyárjas Hill. We measured dips of the youngest P3–P1 pediment surfaces and using these values we projected surfaces from the Nyárjas Hill north-westward (Fig. 8). P1 surface may not represent a real pediment because occurs only in some wide valleys incised below pediment P2. It is more likely that the dip values derived from P2 and P3 provide valid approximations of the original geometry of the older P5 denudation surface topping the Nyárjas Hill. Surfaces having the same dip as pediments P2 or P3 clearly project above the highest P5 surface of the Vértes Hills (Fig. 8). Because this geometry would not allow sediment transport from the source, we assume that the Nyárjas Hill was uplifted after the transport

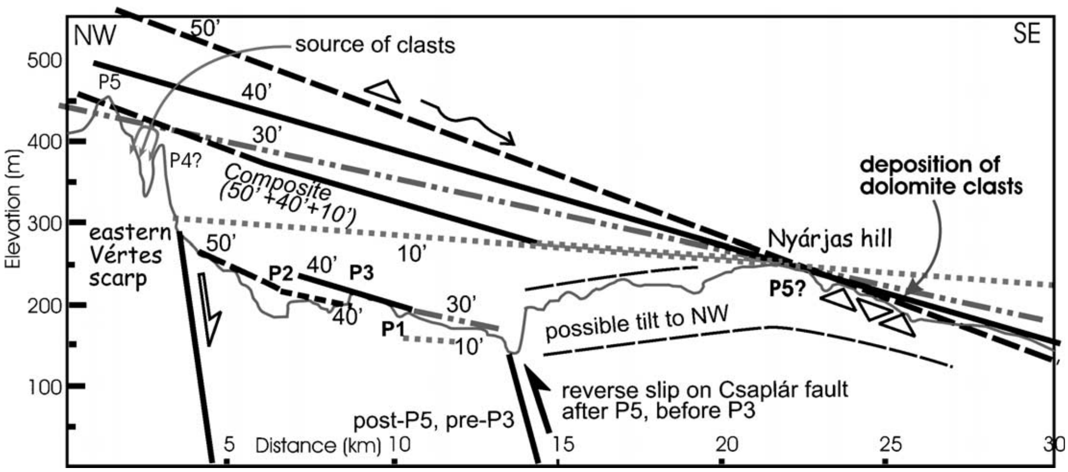

Fig. 8. Reconstruction of supposed denudation surface upon which dolomite debris was transported from the Vértes scarp to the Nyárjas hill. Lines indicate possible projections using different dip angles measured on surfaces P1–P3. Triangles indicate Triassic dolomite clasts derived from the eastern Vértes scarp. Vertical exaggeration is 25 times.

and deposition of dolomite clasts. The uplift was likely caused by the inverse reactivation of the Csaplár Fault associated with slight northward tilt of the Nyárjas hill (Figs. 7 and 8). This uplift can reverse flow direction and explain drainage anomalies around the hill.

The P5 surface truncates the youngest post-rift sediment (Vértesacsa Fm.) and post-dates normal faulting of the Csaplár Fault, thus likely has Late Pliocene (?) to Early Quaternary age. Accordingly the uplift of the Nyárjas Hill was probably caused by latest Pliocene or Quaternary inverse reactivation of the Csaplár Fault associated with slight northward tilt of the Nyárjas Hill block (Figs. 7 and 8). This event indicates a change of faulting style from tensional to compressional or transpressional. Moderate earthquake activity in the Vértes foreland (Kiszely, 2001) may indicate ongoing activity as well.

4.3. Neotectonic deformation of the Gödöllő Hills

4.3.1. Geological and geomorphologic settings

The Gödöllő Hills are located in the central-northern part of the Pannonian Basin, east of the Danube River. The hills form ranges gradually lowering southeastward, towards the Great Hungarian Plain (from 340 to 100 m asl) (Fig. 9). The subsurface Mesozoic, Palaeogene–Early Miocene and syn-rift sediments are covered by an extensive post-rift ("Pannonian") sequence, with thickness up to 1.5 km. Overlying the thin Early Pannonian lacustrine silt or calcareous clay, the upper part of the post-rift suite is a large prograding delta formation containing intercalated sand, silt, clay sequences. The shift from lacustrine–deltaic to fluvial sedimentation might have occurred between 8–6 Ma (Müller and Magyar, 1992) while mammal bones and teeth prove fluvial sedimentation up to the Middle Pliocene (~4 Ma, Mottl, 1939).

The post-rift unit is covered by up to 30 m thick Quaternary sediments, mainly different loess units and intercalated paleosols (Balla, 1959; Frechen et al., 1997). Recently luminescence dating, radiocarbon ages of charcoal and fossil gastropods demonstrated that the upper part of the loess has an age around 100 ka and younger (Frechen et al., 1997; Novothny et al., 2002). The lower part of the loess sequence was not numerically calibrated but was tentatively correlated to glacial phases and may represent loess up to ~400 ka (Horváth, 2001). The hills are dominated by slope sediments formed mainly by the gravitational re-deposition of the loess units (Balla, 1959; Fodor et al., 2001).

The slope distribution map, derived from DEM, suggests that the Gödöllő Hills are formed by the Valkó and Úri ridges, and the Isaszeg channel in between (Fig. 9). The area is characterised by a system of southeastward flowing creeks. These are tributaries of larger creeks, which dissect the two ridges (Rákos, Alsó-Tápió, Kókai creeks). West from the Úri ridge, the Pest plain was formed by the fluvial erosion of the Danube.

Two wedge shaped ridges can be recognised on the relative relief map of the area (Fig. 9). These ridges are very narrow and relatively high at their NW part and are gradually widening and lowering south-eastward. Loess occurs on the ridges but is almost completely missing in the Isaszeg channel where aeolian sand dunes directly cover fluvial post-rift sediments. At the northwestern termination of the Úri ridge and the Isaszeg channel, wind-abraded pebbles (ventifacts) were observed at a number of locations (Jámbor, 1992, 2002; and own data). Pebbles with desert varnish indicate (semi)desert environment, favourable for deflation (Schweitzer, 1997). It is therefore suggested that the Isaszeg channel represents a wind-deflated valley, which acted as a wind channel throughout the Quaternary. On the other hand, the ridges may represent yardangs, where their NW tips resisted against deflation, protected post-rift sediments and enabled loess deposition on the lee side of the ridges. The Isaszeg channel is the easternmost "meridional valley" and they share a common deflational origin.

4.3.2. Morphological and tectonic observations in the Gödöllő Hills

Using industrial seismic reflection profiles we constructed a subsurface map of structures (Fig. 9). Major Miocene structural features of the Gödöllő Hills are (N)NW–(S)SE trending tilted blocks and related asymmetric half grabens (Tari et al., 1992; Fodor et al., 1999). Seismic sections demonstrate that normal faulting mainly occurred during the syn-rift and early post-rift stages (18–9 Ma, Fig. 10). In the northern part of the Gödöllő Hills, however, a number of NNW–SSE trending faults were slightly reactivated. At the highest imaged level (at 100–300 ms TWT that is between 200 and 0 m below sea level) the normal

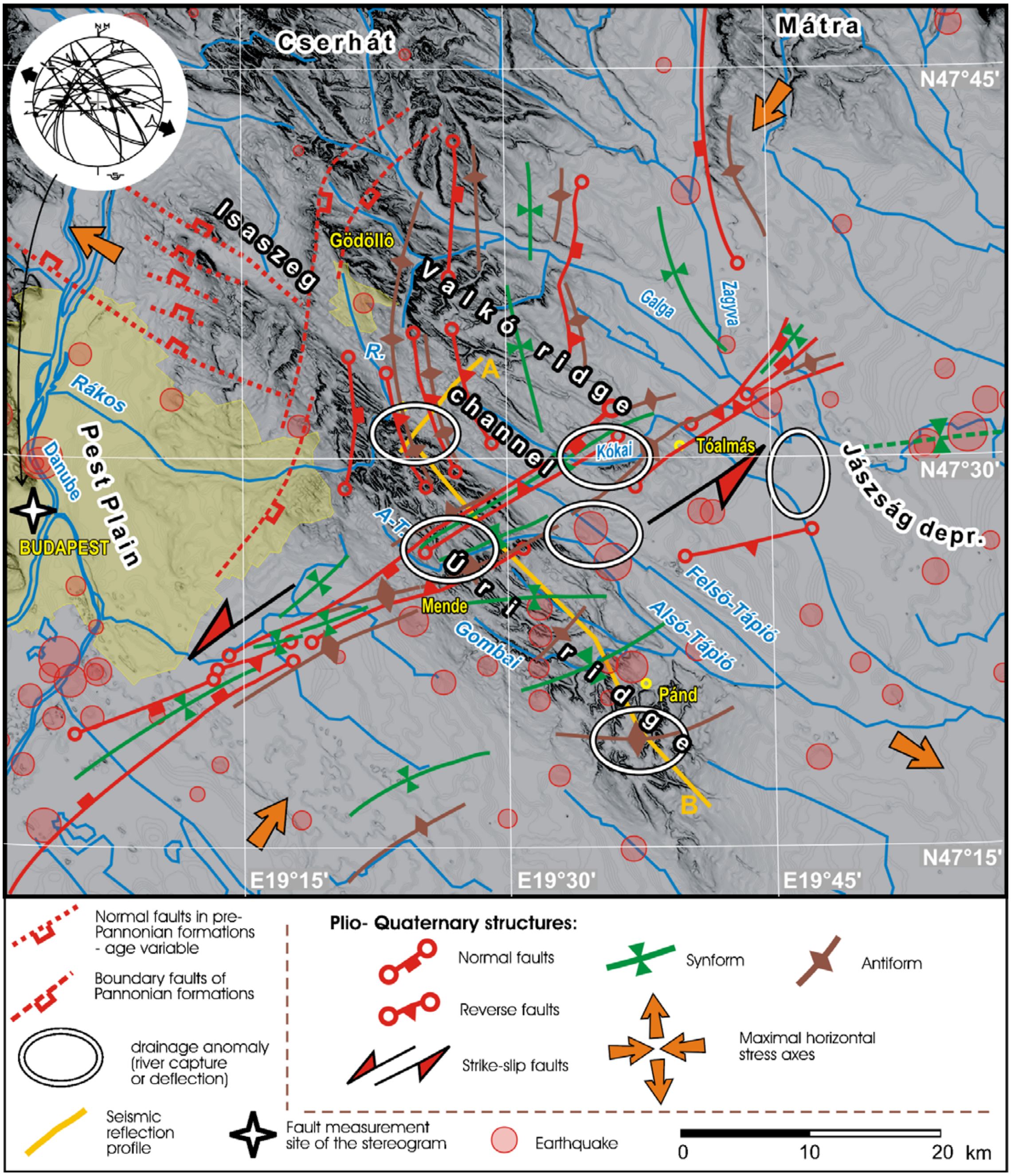

Fig. 9. Digital slope model of the Gödöllő Hills, the steepest slopes are black, and the flat surfaces are grey colour. Lines A and B indicate the locations of seismic sections shown on Figs. 10 and 11. Earthquake distribution from Zsíros (2000). Stress axes are simplified from Gerner et al. (1999) and Tóth et al. (2002). A-T.: Alsó Tápió creek. R: upper Rákos creek.

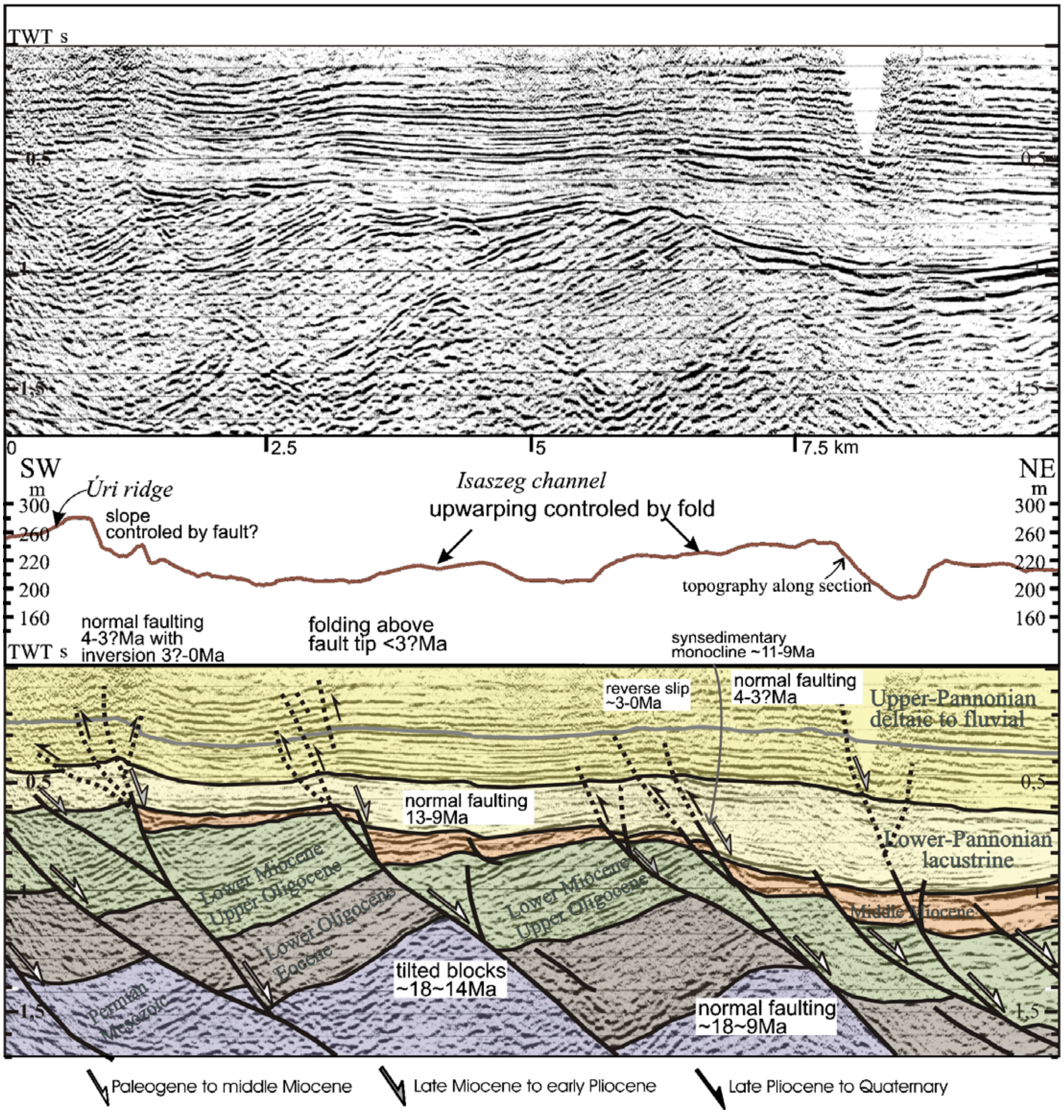

Fig. 10. NE–SW trending seismic reflection profile shows tilted syn-rift blocks, slight motion during late syn-rift and early post-rift phase (grey arrows) and neotectonic faults (black arrows and dotted lines). The latters have small reverse motion and associated folding above the inverted normal faults. Note tectonic control of some landforms.

offset is very small, close to the resolution of the seismic profiles (~10–20 m). Small anticlines parallel to these faults can be considered as fault-related folds. Amplitude of such deformation is small, less than 50 ms (ca. 50 m).

On the other hand, small anticlines occur exactly above the tip of syn-rift faults and not in their hanging wall below the Isaszeg channel (Fig. 10). We interpret such anticlines as the result of inversion of earlier normal faults. Reverse displacement was not sufficient to compensate the earlier normal separation at the level of basal post-rift layers, but produced distributed deformation, small folds and very modest reverse offsets in the upper post-rift sediments.

In the northern part of the Gödöllő Hills most of the structures do not seem to influence the landscape and drainage network. We suggest that the deformation could be of late Pliocene to earliest Quaternary

(?) in age, pre-dating a denudation event and the deposition of the middle to late Quaternary loess series. On the other hand, the anticlines related to the slip reversal could influence the drainage pattern. Surface upwarping could have induced the disintegration of a former SE flowing creek within the Isaszeg channel. Consequently an intra-valley drainage divide formed by the separation of the Felső–Tápió and upper Rákos creeks. The Rákos creek was possibly captured, and currently it drains into the Danube.

In the southern Gödöllő Hills, an ENE trending zone of drainage anomalies can be followed from the Danube up to the Zagyva River (Fig. 9, Tóalmás zone). Within this zone the Alsó–Tápió and Kókai creeks cross the Úri and Valkó ridges. Seismic sections clearly demonstrate the presence of a broad fault zone located beneath the 'anomalous' creeks (Figs. 9 and 11) (Fodor et al., 2001). This Tóalmás zone had a complex Miocene deformation history (e.g., Tari et al., 1992; Csontos and Nagymarosy, 1998); here we focus on Late Miocene and younger events. At the lower, lacustrine part of the post-rift sequence ('Lower Pannonian'), syn-sedimentary thickening occurs within the fault zone (Fig. 11). Individual segments of the Tóalmás zone change their fault polarity along strike from north to south and again to north. These features outline a flower structure and suggest transtensional strike-slip faulting during the early Late Miocene. Main structural features within the upper part of the post-rift sequence are folds with 30–250 m amplitude following en echelon geometry (Fig. 9). Some faults show reverse slip or reverse drag fold at higher post-rift reflections, while at depth they still show normal separation. We interpret these features as transpressive reactivation of the early Pannonian transtensional fault system.

The diversion of the Alsó–Tápió creek at Mende can be explained by the raising anticline above the southern branch of the Tóalmás zone. The fold amplitude is the largest at the location of creek diversion (i.e., at Mende) and gradually decreases eastward where the creek could turn again to the SE. Further northeast the diversion of the Kókai creek has similar structural explanation, whereas the smaller structural amplitude is expressed by less developed morphological indications.

South from the Alsó–Tápió creek the thick Pannonian sequence shows a series of syncline–anticline of

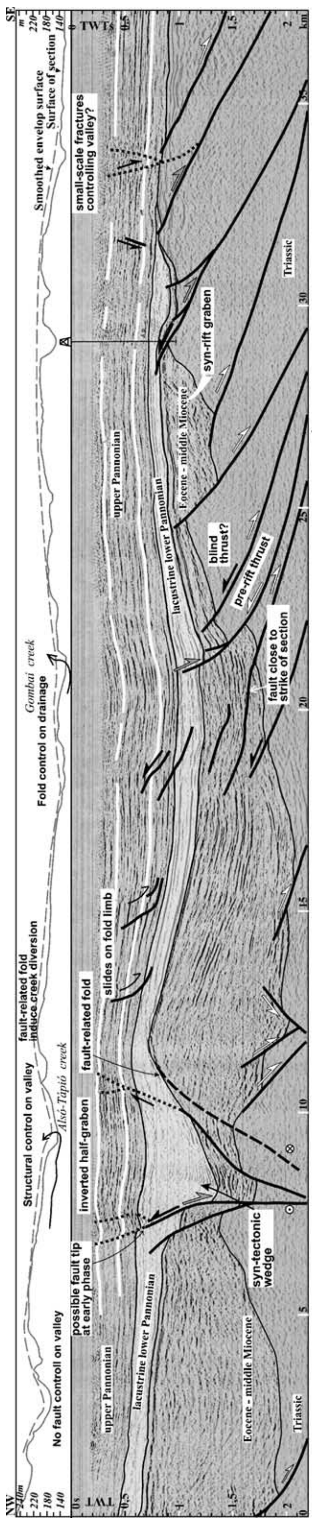

Fig. 11. NW–SE trending seismic reflection profile showing the Tóalmás zone and the structure of the southern Gödöllő Hills. Note corresponding topographic and subsurface folds, and the diverted drainage.

250–320 ms TWT amplitude, with smaller folds on major fold limbs (Fig. 11). The shape of the smoothed envelope surface of the topography is similar to the folded Pannonian strata, while elevations on the north and south bound a wide valley, the Gombai creek north of Pánd. The E–W trending axis of the major syncline closely coincides with the east flowing reach of the Gombai creek (Fig. 9). Tributaries arriving from the northern and southern slopes of the depression flow over flanks of the syncline. The Gombai creek has a wide alluvial plain indicating subsidence probably because of its position over an actively deforming syncline hinge. The anticline south of Pánd corresponds to a local high characterised by a radial drainage network indicative of active uplift during valley incision.

4.3.3. Neotectonic significance of the morphotectonic analysis

Based on its en echelon geometry, the Tóalmás strike-slip zone is likely to have sinistral kinematics. Large synclines in the southern Gödöllő Hills and Jászság basin can also represent en echelon folds. These structures, together with the inversely reactivated NNW–SSE trending faults in the north suggest NE–SW oriented maximal and NW–SE oriented minimal stress axes (Fig. 9). These stress axes agree well with the recent stress field of the central Pannonian Basin (Gerner et al., 1999), with earthquake focal mechanism solutions (Tóth et al., 2002), and with post Mid-Miocene fault slip data from the neighbourhood of the Gödöllő Hills.

In the southern Gödöllő Hills, the presented subsurface and surface observations suggest that folding affected the whole post-rift sequence including the highest fluvial unit and could start in the Late Pliocene. Because fold amplitude is larger in the post-rift suite than in the smoothed topographic surface, several episodes of folding and denudation could be envisaged during the Late Pliocene and Quaternary.

The distribution of historical earthquakes (Tóth et al., 2002) suggests active deformation in the southern part of the Gödöllő Hills (Fig. 9). Earthquakes occur along the Gombai creek where post-rift sediments do not show faulting. We propose that these shocks are related to folding of the E–W trending major syncline and to slip on related blind reverse faults at depth. A WSW–ENE trending cluster of earthquakes stretches from the Danube up to the Zagyva River. These events can be correlated to strike-slip motion along the Tóalmás zone. Another important earthquake zone is located east of the Gödöllő Hills, in the Jászság depression. Latest Pleistocene to Holocene onset of subsidence is indicated by the capture of the Zagyva River from its original southward flow toward this depression (Gábris, 2001; Timár et al., 2005—this volume). We tentatively interpret earthquakes and subsidence as sign of an east-trending syncline, similar to that of the southern Gödöllő Hills. Because the Jászság fold lacks marked expression on seismic sections, the structure is probably at early stage of formation. This recent folding may indicate northeastward propagation of the Tóalmás zone.

5. Discussion

5.1. Structural styles and their geomorphic expression

Our results allow depicting a general model for the style of neotectonic deformation and its relation to surface processes within the uplifting western part of the Pannonian Basin (Fig. 12). Syn- to post-rift faults impose an important although passive control on the landscape in form of steep fault scarps. They have formed due to the presence of soft hanging wall sediments contrasting to erosion-resistant footwall rocks (Fig. 12). However, neotectonic activity of such predominantly normal faults is difficult to assess. The most common style of neotectonic deformation is the inversion of earlier syn- to post-rift faults (Fig. 12) with change of fault kinematics from normal to reverse (Zala Hills) or from transtensional to transpressional strike-slip faulting (Gödöllő Hills). In most cases reverse faults do not propagate to the surface: the slip on reactivated single syn-rift fault strand at depth was distributed into small-scale reverse fault branches upwards. Apart from small-scale faulting, the most characteristic neotectonic feature is folding and related surface undulations. Consequently, surface rupturing is scarce. In addition, the relatively fast late Pleistocene loess or fluvial sedimentation could also erase traces of surface ruptures. Folds can appear as surface undulations, phenomenon also observed by other studies (Horváth, 1995; Horváth et al., in press; Csontos et al., 2005—this volume).

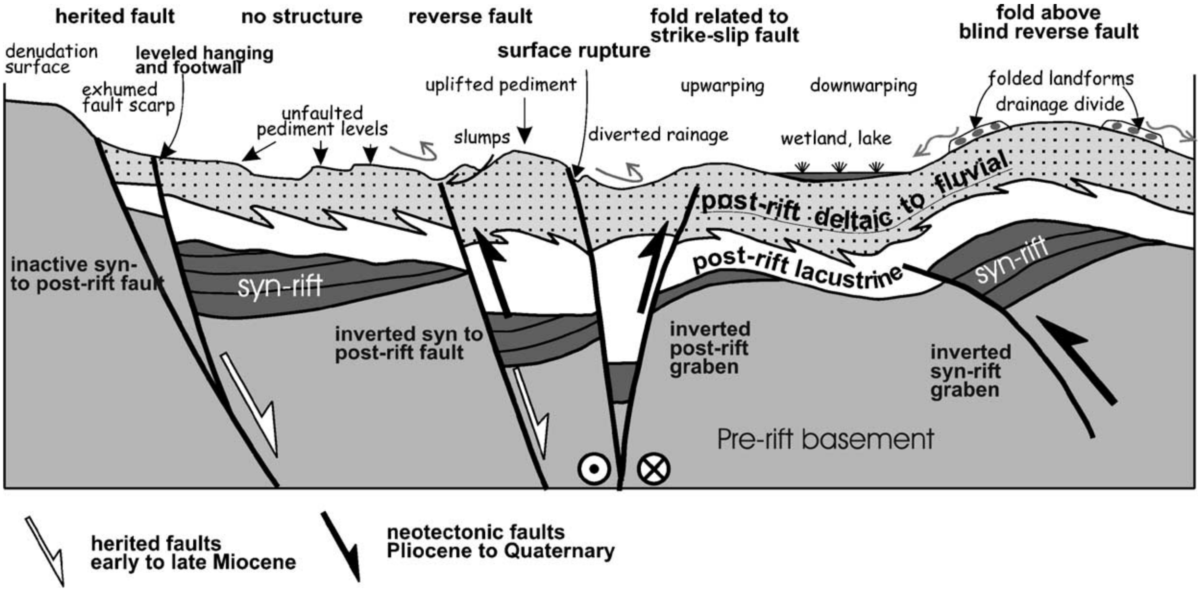

Fig. 12. Idealised cross-section modelling styles of neotectonic deformation and related landforms in the Pannonian Basin, broadly in N–S direction. Not to scale.

At several sites in the Gödöllő, Vértes and Zala Hills neotectonic surface warping induced river piracy, formation of new watersheds along anticlinal hinges, and the development of wind gaps (Fig. 12). Gravitational sliding can be abundant on flanks of uplift but they are only signs but not proof for neotectonic deformation. Some denudation surfaces (e.g., pediments of the Vértes Hills) are unfaulted and thus indicate the lack of surface rupturing faults. On the other hand, deformed planation surfaces are good candidates to image and quantify surface deformation. Examples are folded alluvial fans (Zala Hills), uplifted pediment surfaces (Vértes Hills) and river terraces (Danube Bend area, Pécsi, 1959; Ruszkiczay-Rüdiger et al., 2005—this volume a,b).

5.2. Origin of some landforms in the western and central Pannonian Basin

Our observations enabled us to provide new insights on the questionable tectonic origin of some landforms, particularly on the “meridional” and “longitudinal” valleys and lakes. Some ENE trending faults can be well correlated with “longitudinal” valleys. The best-documented examples are the Kapos and Tamási Lines (Fig. 13, Síkhegyi, 2002; Bada et al., 2003), which together follow an Early to Mid-Miocene shear zone of complex origin (Csontos et al., 2005—this volume). Earthquake epicentre distribution (Tóth et al., 2002) confirms recent faulting, while landforms extend the activity back to the Quaternary (Síkhegyi, 2002).

The depression corresponding to the Pacsa syncline seems to continue eastward towards Lake Balaton (Fig. 2). This observation may suggest that the location of the large SW–NE tending lake was controlled by surface down-warping (Fodor et al., 2005). A fold-controlled rather than faulted origin of the lake would be more in agreement with general neotectonic structural style projecting data both from south-western (this study) and south-eastern surroundings of the lake (Sacchi et al., 1999; Csontos et al., 2005—this volume).

Neotectonic folds and faults are perpendicular or highly oblique to “meridional” valleys in the Zala and Gödöllő and Vértes Hills (Fig. 13). In addition, the radial “meridional” valley system cannot be reconciled with the syn-rift fault pattern. While “meridional” valleys are straight, syn-rift faults are curved to corrugated, and show several oversteps, similarly to those observed in modern rift systems (Patton et al., 1994). Particularly in the south-western side of Lake Balaton

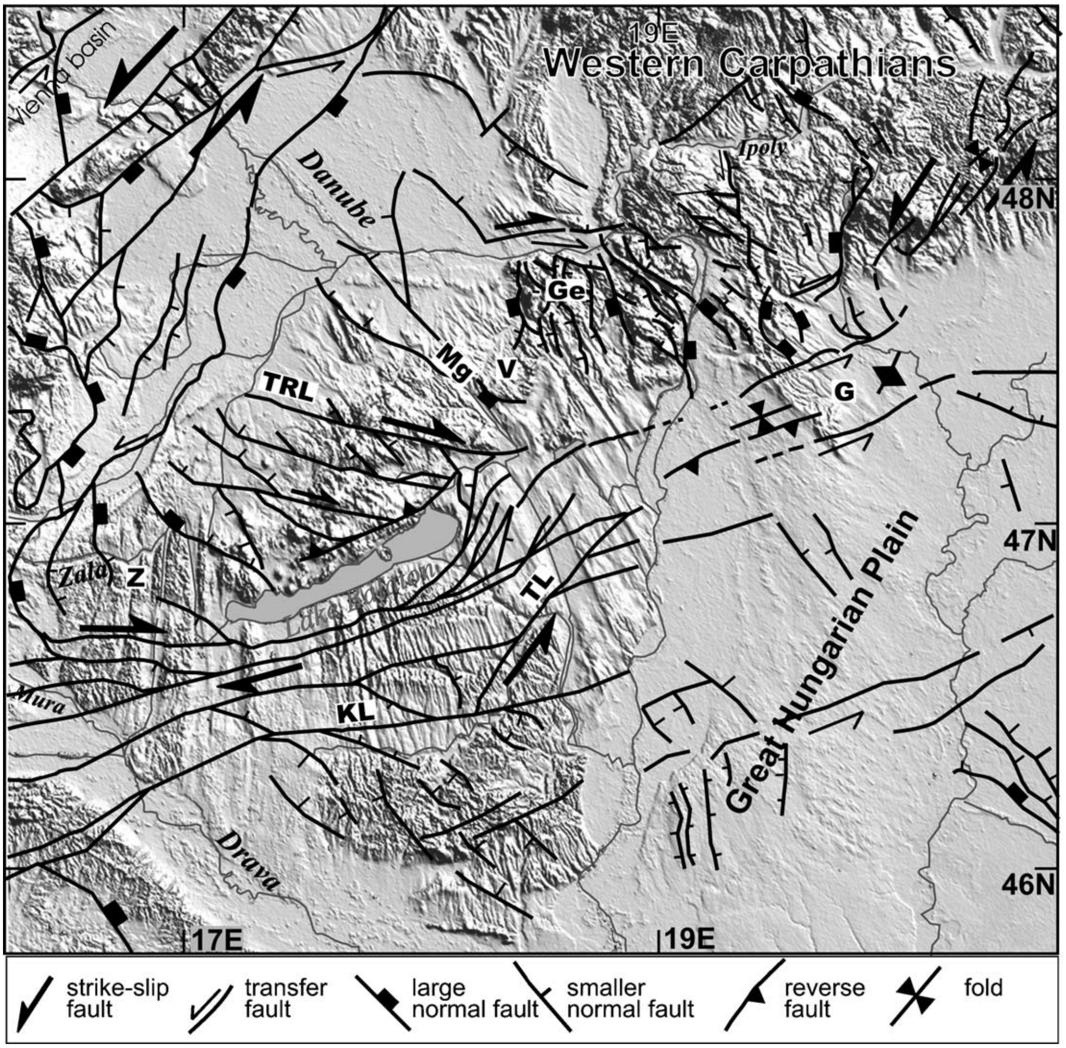

Fig. 13. Syn-rift fault pattern superposed on digital elevation model of the western and central Pannonian Basin. Fault pattern is after Fodor et al. (1999), large normal faults have more than 1 km separation. Note good fit of faults and topography in the Mid-Hungarian Shear Zone (MHZ) including the Kapos (KL) and Tamási Lines (TL) and along some exhumed syn-rift faults in the Gerecse Hills (Ge). However, meridional valleys often lack fault control. Z, V, G: Zala, Vértes and Gödöllő Hills, respectively (see study areas on Fig. 1); Mg: Mór graben; TRL: Telegdi Roth line.

the radial–linear valley system has little in common with the geometry of Miocene normal faults. Accordingly we suggest that the meridional valleys have no noticeable control from neither neotectonic nor syn-rift faults or folds although our analyses cannot exclude that small-scale faulting or joint systems could play a limited role in localisation of short segments of drainage. Denudation processes mainly influence the shape of valleys and lakes. Concerning the mechanism of denudation, we reinforce earlier conclusions of Lóczy (1916), Cholnoky (1920) and many others in favour of a dominantly eolian (deflation) origin of the Transdanubian valley system with minor role of fluvial incision as well. We also suggest that the formation of a number of lakes found in 'longitudinal depressions' was also connected to surface warping and deflation effect of catabatic wind (Fodor et al., 2003).

Relatively good spatial correlation exists between steep slopes of the Transdanubian Range and WNW striking strike-slip and (N)NW trending normal faults (e.g., Telegdi Roth Line, and Gerecse Hills, respectively, Fig. 13). However, slope-controlling faults are older (syn- to post-rift) structures that were exhumed relatively recently during regional uplift. Undoubted evidence for Plio–Quaternary reactivation will need further arguments including the study of geomorphic surfaces.

5.3. Timing of deformation

Time constraints of neotectonic deformation have spatial variations in the Pannonian Basin. Despite the preliminary character of such data, we formulate the following working hypothesis. In the south-western Pannonian Basin folding and strike-slip faulting could locally be initiated within the Late Miocene (Mecsek–Villány Mts., Csontos et al., 2002). Regional contraction started at the Miocene–Pliocene boundary in southern Transdanubia suggested by the onlap geometry of Pliocene over deformed Late Miocene (Sacchi et al., 1999; Sacchi and Horváth, 2002). In Slovenia, Early Pliocene folding was post-dated by erosion, subsequent transtensional basin formation and regional vertical-axis rotation (Vrabec, 1999; Márton et al., 2002). On the other hand, continuous Pliocene to Quaternary folding was demonstrated near the Slavonian Mts. (Tomljenović and Csontos, 2001). and in the Zala Hills. In this latter area the young (Quaternary?) folding is reflected by the deformation of an earlier denudation surface and is also indicated by the surface morphology, deflection of drainage, occurrence of wind gaps.

The initial folding event could induce an early uplift of southern Transdanubian Range. Volcanologic reconstructions of basalt occurrences and geomorphologic analyses revealed important denudation taking place both before and after volcanism (Németh and Martin, 1999; Csillag, 2004). Magmatic rocks occur upon pre-Miocene rocks where Late Miocene was eroded (maps of Lóczy, 1916; Budai et al., 1999). The pre-basalt denudation has occurred in the Early Pliocene, before the major phase of volcanism of 2–4.5 Ma (Balogh et al., 1983) and marks the first sign of uplift and thus the beginning of the inversion.

Both in the Vértes Hills and northern Gödöllő Hills extensional or transtensional normal faulting seems to continue after the Miocene, during the final phase of fluvial post-rift sedimentation and early episodes of denudation. In the Gödöllő Hills the transtensional deformation lasted up to middle Pliocene times. This timing is in contrast with simultaneous compressional or transpressional deformation of the southern Pannonian Basin. Our observations reinforce earlier suggestion of Tari (1994) that the compressional/transpressional deformation gradually progressed from the southwestern to central part of the Pannonian Basin during the Pliocene. By the end of the Pliocene inversion of extensional or transtensional faults and onset of surface uplift have started in most part of the Pannonian Basin. These processes led to the establishment of the present-day structural pattern. This gradual propagation of inversion suggests that the neotectonic deformations were progressively dominated by the effect of the motion of the Adriatic plate parallel with the decreasing influence of the Carpathian subduction zones.

5.4. Regional geodynamic considerations

Finally we tentatively extend our detailed observations into a wide Alpine–Carpathian–Pannonian–Dinaric framework. We discuss deformation and motion of three major plate tectonic units through their velocity field derived from GPS data, and the observed neotectonic structures. Temporary names for these units were adopted because of the still weakly defined unit boundaries and also to avoid confusion when ‘recycling’ names valid for Neogene tectonic units.

Northward motion and counterclockwise rotation of the Adriatic microplate can induce most of the intra-Carpathian deformation (Anderson and Jackson, 1987; Bada et al., 1999, 2001; Oldow et al., 2002; Weber et al., 2004). Present-day shortening of 2–4 mm/y between the Bohemian Massif and Adria is now largely absorbed within the Alps and the northern Dinarides (Grenerczy and Kenyeres, 2004). Late Miocene to recent folding and thrusting were documented in the Southern Alps (Castellarin and Cantelli, 2000), Sava fold belts (Placer, 1999; Vrabec and Fodor, 2005) and Croatia (Tomljenović and Csontos, 2001) and suggest persistent structural pattern for the last 6 Ma (Fig. 14). Because of the obliquity of convergence and the rotation of Adria, ~N–S contraction was partly accommodated by strike-slip faulting, resulted in transpression in Slovenia and Slavonia. The only exception is central Slovenia where Quaternary deformation is reflected by the opening of several transtensional basins (Vrabec and Fodor, 2005).

Within the Carpatho–Pannonian domain the 'western unit', containing the easternmost Alps, southwestern Carpathians and western Pannonian Basin, is still moving eastward with a differential motion of ~1.3 mm/y with respect to stable Europe (Grenerczy et al., 2000). The unit is clearly bounded by the Mur–Mürz–Žilina Line to the north, which shows strong seismicity and several active faults near the Vienna Basin (Hinsch and Decker, 2003; Decker et al., 2005). The eastward motion of the 'western unit' is accommodated in a broad zone stretching from the Lake Balaton to the Danube and possibly into the central Western Carpathians (Fig. 14). Shortening could induce regional surface uplift and enhance the incision of the Danube River across the northern Transdanubian Range (DB) (Ruszkiczay-Rüdiger et al., 2005—this volume b). Further to the east, the 'north-eastern unit' (including the NE

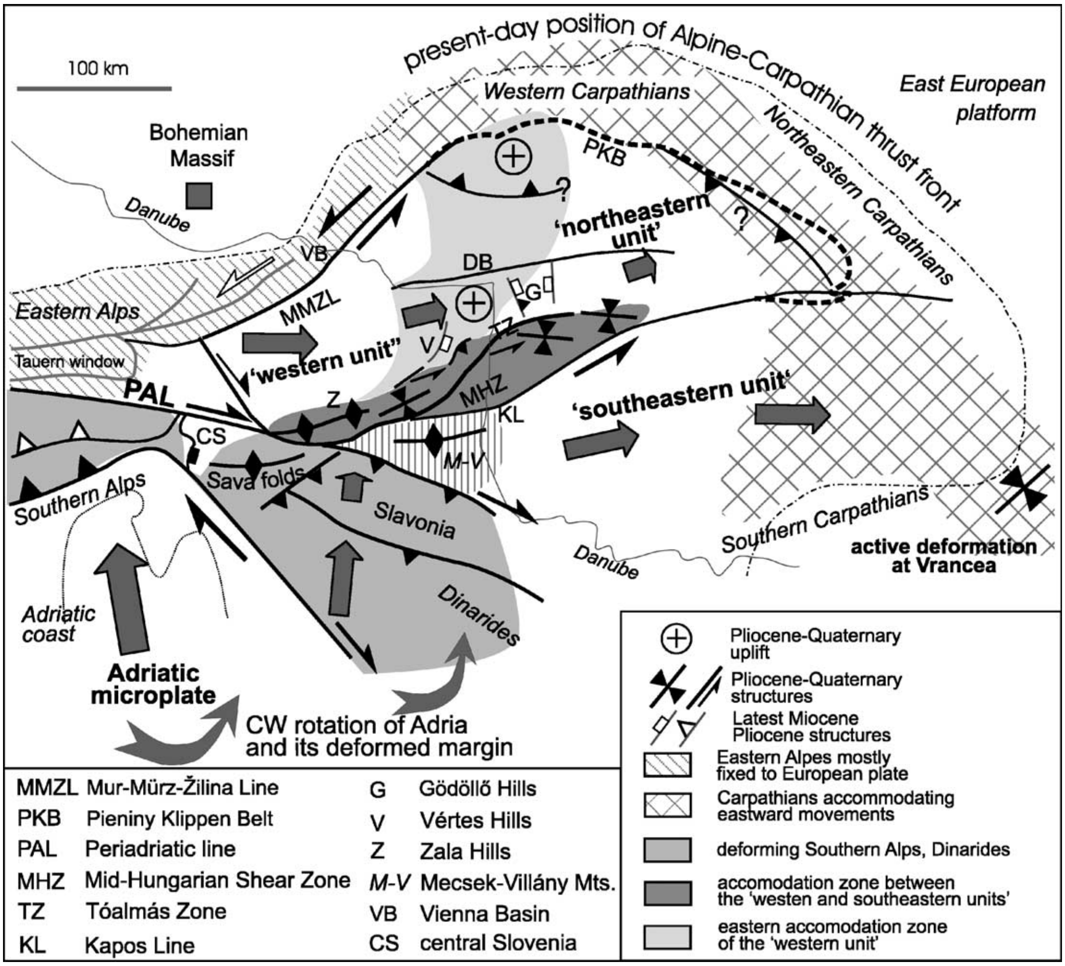

Fig. 14. Structural sketch of the main neotectonic blocks and deformation elements in the western and eastern part of the Pannonian basin. Large arrows schematically show the recent motion of units with respect to fixed Bohemian Massive (Grenerczy et al., 2000; Grenerczy and Kenyeres, 2004). DB: Danube Bend area.

part of the Pannonian Basin and the NE Carpathians) shows one magnitude smaller motion (0.3 mm/y). Its final docking may induce reverse faulting and earthquakes along its north-eastern edge within the Carpathian arc. The 'south-eastern unit' comprising the SE Pannonian Basin and parts of the Southern and Eastern Carpathians is currently moving eastward (Grenerczy and Kenyeres, 2004). This motion can be connected to the active deformation in the Vrancea zone near the south-eastern corner of the Carpathians (Maţenco and Bertotti, 2000). The western wedge-shaped part of this unit suffered transpressional deformation (Csontos et al., 2002) due to the push of the Dinarides from the south.

The motion between the 'western' and 'south-eastern units' is accommodated in a wide zone, which is close but not identical to the Miocene Mid-Hungarian Shear Zone (MHZ) (Fig. 14). Because the eastward motion of the 'western unit' is currently accommodated before the Danube River, and the 'eastern unit' does not show such velocity gradient, the kinematics of their broad accommodation zone changes along strike. The zone has dextral–transpressional character at its southwestern part in Slovenia as demonstrated by structural analysis (Fodor et al., 1998; Vrabec, 1999) and GPS data (Weber et al., 2004). Slightly to the east, pure shortening induced folding in the Zala Hills. An increasing sinistral slip component of individual faults occurs from the Lake Balaton towards the east (Bada et al., 2003). Post–Miocene sinistral faults were identified at the eastern Lake Balaton, (Vida et al., 2001; Csontos et al., 2005—this volume), in the SE foreland of the Vértes Hills. Further to the east, the Tóalmás zone and related folds in the southern Gödöllő Hills and Jászság depression are located where the 'south-eastern unit' passed the blocked 'western unit' and pressed against the relatively stable 'north-eastern unit'. More to the east, the 'south-eastern unit' has clear sinistral motion with respect to the stable 'north-eastern unit' (Fig. 14). This motion is reflected by a number of sinistral transtensional or transpressional shear zones starting with the Kapos Line and crossing the Great Hungarian Plain (Pogácsás et al., 1989; Detzky-Lőrincz et al., 2002; Síkhegyi, 2002; Bada et al., 2005). Tóth et al. (1997), Timár (2003) and Windhoffer et al. (2005—this volume) clearly demonstrated persistent Quaternary activity along these fault zones.

6. Conclusions

Neotectonic deformation in the Pannonian Basin took place in terrestrial environment and was partly preserved in deformed landforms. The presented case studies emphasise that the exact demonstration of neotectonic structural elements requires the use of complex methodology including geomorphology, geology, and geophysics. Seismic reflection profiles are of utmost importance to image subsurface structures. Established subsurface structures have to be crosschecked by surface geologic or geomorphic indices to validate neotectonic age of deformation. The value of small-scale surface structural data is somewhat limited because of the scarcity of outcrops and dubious origin of some structures. Neotectonic studies should involve the analysis of geomorphic processes, like fluvial incision, eolian erosion, and slope sedimentation. Separation of exogene forces and deformation in landscape evolution is essential for neotectonic studies.

Our study suggests that folding is the characteristic neotectonic deformation style in the western–central part of the Pannonian Basin. Folding might have several episodes punctuated by denudation and by formation of sub-planar landforms, which were in turn deformed by subsequent folding phases. Surface upwarpings kept substantial control on the drainage network and induced the development of drainage divides and river deflections. Folds are mainly connected to the reactivation of Miocene normal or strike-slip faults. They represent fault-related folds connected to reverse slip on former fault planes, which propagate upward into thick, mostly unfaulted post-rift suite. On the other hand, neotectonic faults with more than 15–20 m vertical separation are scarce in the Upper Miocene post-rift sequences both on the surface and on seismic sections. This observation is in line with the fact that surface rupture across loose Plio–Quaternary sediments was barely documented.

Contractional neotectonic deformation and related uplift could start as early as the latest Miocene or earliest Pliocene in the southern part of the Pannonian Basin. However, our study gives indications that normal or transtensional faulting was maintained in the central Pannonian Basin and change in fault kinematics could occur during the Late Pliocene but not earlier. Gradual expansion of compression-dominated areas indicates increasing influence of the Adriatic plate motion as driving force for neotectonic deformation.

Based on documented structural elements, historical earthquake distribution and GPS-calibrated velocities, a three-plate structural–kinematic model is suggested. The south-eastern Pannonian–Carpathian block is moving eastward with respect to the north-eastern block while the western Pannonian–Alpine block is gradually pushed against the north-eastern block. This model may account for most of the observed neotectonic structures and related tectonically controlled landforms.

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