

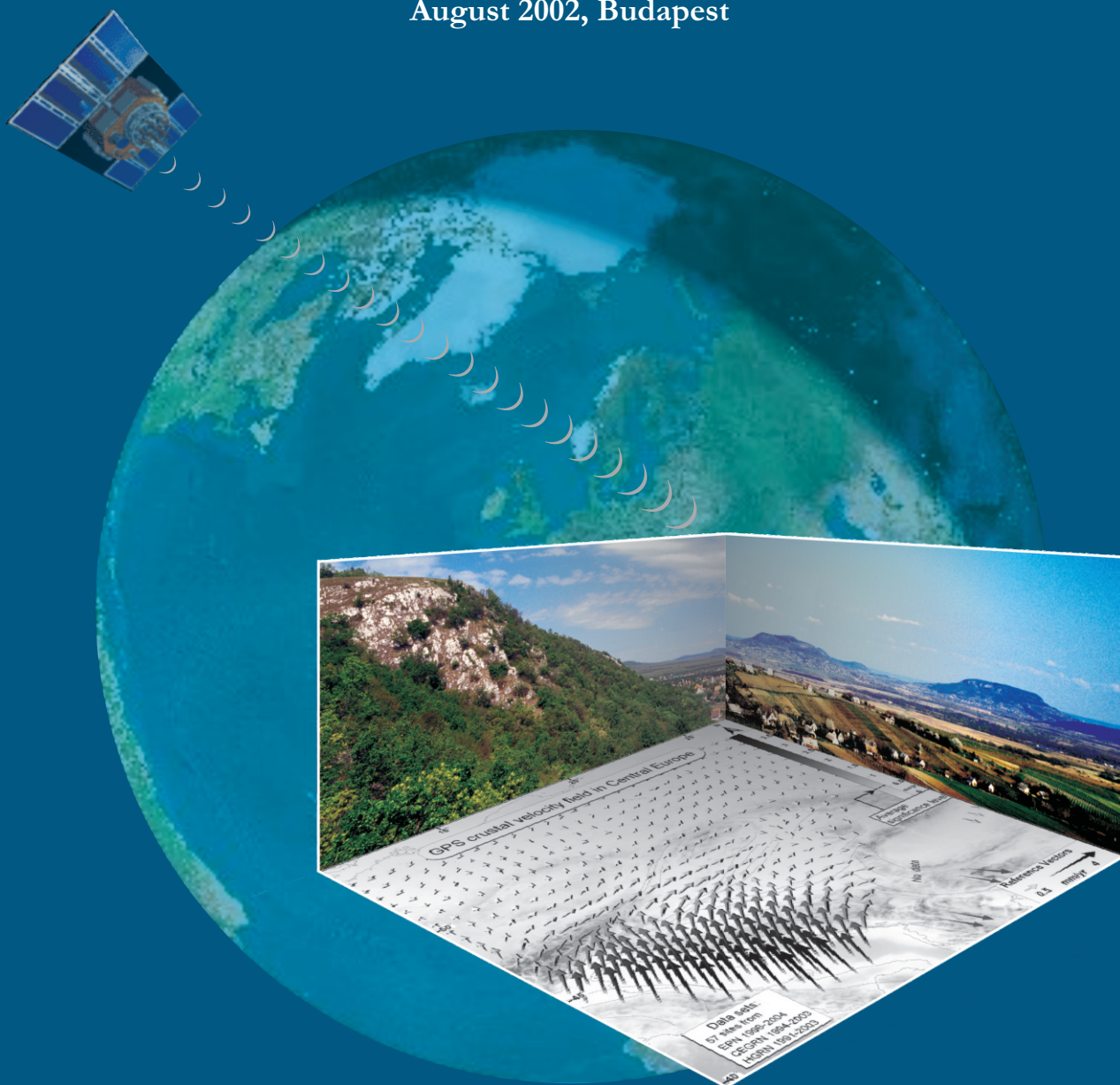
Occasional Papers of the Geological Institute of Hungary, volume 204

# Proceedings

of the workshop on

*“Application of GPS in plate tectonics, in research on fossil energy resources and in earthquake hazard assessment”*

Geological Institute of Hungary  
August 2002, Budapest



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Editors:

László FODOR and Károly BREZSNYÁNSZKY

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*Cover pictures: GPS-based velocity field in the Pannonian Basin (horizontal, see Grenczy this volume); late Miocene fault scarp of the Eastern Vértes Fault Zone, (photo G. Csillag, left); panoramic view of the Tapolca Basin with remnants of Pliocene volcanic edifices (photo G. Csillag, right)*

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## Preface of the Editors

In 1989 the U.S. and Hungarian governments signed an agreement on scientific and technological cooperation, in which the two countries agreed to contribute equal amounts of financial support annually to establish a joint fund. The bilateral agreement stipulates that the purpose of the U.S.–Hungarian Science and Technology Joint Fund is to encourage and support a wide range of scientific and technological cooperation between the two countries, based on the principles of equality, reciprocity, and mutual benefit.

Within the scope of the bilateral program, U.S. and Hungarian scientist could receive support from the Fund for (1) cooperative research projects, (2) bilateral scientific symposia (workshop) and (3) project development visits. Researches from U.S. and Hungarian government agencies, scientific institutes and societies, universities, and other national research and development centers were eligible to apply for Joint Fund support. Applications to the Joint Fund must propose collaborative U.S.–Hungarian scientific research and/or technological development activity, involving researches from both countries. All projects must contribute to the advancement of mutually beneficial scientific knowledge. Research results must be publishable in open source journals. Topics of the collaboration are of wide range, including basic sciences, environmental protection, biomedical sciences and health, agriculture, engineering research, energy, natural resources, including earth sciences, transportation, science and technology policy and management, and other fields.

The Joint Fund Secretariat logged in proposals submitted to the Joint Fund. The Secretariat examined each proposal to determine whether they meet all specified criteria and conditions, including the required number of copies, and decided which U.S. and Hungarian technical agencies are most appropriate for conducting the review.

U.S. and Hungarian agencies among them the U.S. and the Hungarian Geological Surveys concurrently subject the proposals to their own clearly defined peer review procedures. Reviewers' evaluations were based on the following common criteria: (1) intrinsic scientific or technical merit; (2) previous research performance and competence; (3) significance of the research for international cooperation; (4) reasonability of budget; and (5) proposal relevance within national priorities.

Following U.S. and Hungarian technical agency consultations, grant proposals reached the Joint Board, which, based on the results of the reviews, may approve, decline a funding request or postpone a decision. Following the approval of an award, a grant letter confirming the terms of the grant was sent to the principal investigator.

During the decade of Joint Fund activity several earth sciences projects have been carried out with significant scientific results. The present workshop was one of the latest winner of Joint Fund support, with the project leadership of Dr. Károly Brezsnýánszky from the Hungarian side and Prof. John Frederick Dewey from the U.S. side.

The topic of the workshop "Application of GPS in plate tectonics, in research on fossil energy resources and in earthquake hazard assessment" is important both scientific and economic point of view.

The workshop and the related field trip to the Transdanubian Range, with participation of U.S. and Hungarian scientists, was organized by the Geological Institute of Hungary and was held in Budapest between 24<sup>th</sup> August to 1<sup>st</sup> September, 2002.

The results of the workshop can be summarized in the following.

Hungarian scientists and the general public learnt of the workshop through notices describing it as a discussion about recent development of space geodesy, tectonics and structural geology. The presentations clearly demonstrated that these

rapidly growing fields are able to describe and synthesize a number of earth science problems, some of which have large societal impact.

The results of the workshop included the development of new and important research relationships and co-operation, and potential for a number of joint research projects, between Hungarian and U.S. geologists in the Balaton and Transdanubian Highlands as well as in more regional tectonic studies.

New ideas and results on the problems of neotectonics of the Pannonian Basin, and past plate tectonic reconstructions were discussed during the meeting and field trip. A general agreement was reached that the present-day plate motions are markedly different from older (Miocene) deformations. The discussions strengthen the earlier suggestions that the current style of neotectonic deformation may result in increased seismic risk. However the workshop pointed out that more precise data was still needed for thorough evaluation of any natural hazards related to tectonic movements. For example, the precise time of the onset of neotectonic phase has serious implications for the rate of subsequent motion, and thus on present-day hazard estimation.

The workshop demonstrated clearly that modern geodetic techniques have major impacts on neotectonic research. Ongoing Hungarian research projects clearly demonstrate 1.3 mm/year bulk shortening between the Lake Balaton and Budapest and considerable differential motions of Transdanubia with respect to other areas, like the Great Hungarian Plain. Similarly, preliminary GPS measurements on the terraces of the river Danube confirm earlier ideas of large-scale neotectonic deformation related to uplift of the Transdanubian Range.

Geology is a field-based science. The field trip demonstrated that new field observations, supplemented with modern measuring techniques are an essential part of geological research. Integrating data from geodesy, geophysics, geomorphology, volcanology, hydrology were all needed to understand the geology of the field area. This integrated method is the only way to answer some of the problems of Transdanubia, for example the age and origin of uplift, formation of enigmatic Transdanubian valley system (including the Danube gorge), origin of the morphological depression of the Lake Balaton, etc. During the field trip this integrated approach combined with different interpretations by different geologists offered some new hypothesis, such as a compressional origin of Lake Balaton depression, combined eolian-fluvial origin of valleys, that should be tested by future research. The American scientists were particularly impressed by the reconstruction of Pliocene volcanoes of the Balaton Highland. This reconstruction offers a powerful tool to estimate pre-volcanic early Pliocene and post-volcanic Quaternary denudation, related to the uplift.

All these scientific debates are related to general concepts for future research on fossil energy resources. Neotectonic deformation capable of creating large-scale traps for hydrocarbons (like in the southern part of the Pannonian Basin) but could also contribute to destruction of already existing traps and seals by means of recent deformation.

It is also clear that neotectonics exerts a first-order control on fluid flow via the past and active fault pattern of the area, and convincing evidences of this control is found in the Transdanubian Range. Associated regional uplift of Transdanubia could play major role in establishing the (hydrothermally influenced) fluid flow.

This volume contains short papers of the participants and an excursion guidebook of the field trip to the Transdanubian Range, particularly to the Balaton Highland area.

John F. Dewey in his paper outlines the general rule for transpressional and transtensional deformation, which can be used as theoretical background for studies of deformation of the Transdanubian Range during both older (Cretaceous, Miocene) and neotectonic phases.

John C. Weber describes a case study of application of the GPS technique in the actively deforming area of the Caribbeans, namely in Trinidad Island and its surroundings. They show that the major displacement zone occurs in the middle of the island and may trigger surface deformation as well. The precise GPS measurements could modify views of the neotectonic fault geometry and the earlier plate tectonic models of a broader area.

Gyula Grencsics summarizes the results of the Global Positioning System technique in central Europe obtained through coordinated studies from 1991 onward. He clearly demonstrates that present-day velocity field of the Pannonian basin is highly variable, both in direction and in amount. The data also suggest that the driving force of neotectonic deformation is the northward movement of the Adriatic plate, which was absorbed mainly in the Dinarides and Southern Alps, but also in the Pannonian region.

The direct effect of neotectonic plate convergence was detailed in Fodor et al. paper. In addition to briefly describe neotectonic structures in western and central Pannonian basin, the authors suggest a simplified model for neotectonic structural evolution of the Pannonian area. The model involves gradual temporal propagation of inversion structures from SW to NE, from the direction of the Adriatic plate toward the basin interior. If valid, this model may change considerations about temporal evolution of active deformation and long-term seismic hazard evaluation.

Nicholas Pinter undersigns the importance of tectonic geomorphology as a sensitive and newly growing technique in searching for young Quaternary structural elements. He also presents the preliminary result obtained by this method in southwestern Hungary, in the Zala Hills. Geomorphic indices suggest active surface deformation of the area, corroborating earlier assumptions. Further application of similar studies may have importance for active deformation, seismic risk evaluation within the poorly outcropping Pannonian Basin.

The paper of Emőke Jocha-Edelényi exploits a very important applied aspect of recent and past deformation phases. The structural geological control upon the fluid flow system of the Transdanubian Range is of primary importance, because human impact modified seriously natural circumstances. The determination of vulnerability, preservation and maintenance of karst water reservoir system is only possible when taking into account combined geological, structural and hydrogeological data sources.

Finally the excursion guidebook (Fodor et al.) describes general structural, volcanological, geomorphological and stratigraphical features of some parts of the Transdanubian Range, classical research area of the Hungarian geological research. In addition to brief description of some published and stratigraphically important sites, the guidebook concentrates on the results of new field works and on new interpretations of the structure of the Transdanubian Range. New mapping in the Vértes Hills clearly underlies the role of late Miocene faulting, which lasted in the early Pliocene as well. The resulted fault pattern might have guided inversional neotectonic structures, as revealed by inverted normal faults of the area. The determination of precise geometry of the Cretaceous imbricate structure of the Balaton Highland may also have neotectonic importance. Gently dipping thrust planes could be reactivated during Miocene transpression, and, as a newly formulated idea, can play role in neotectonic folding as well.

Neotectonic deformation played considerable, although not exclusive role in Pliocene–Quaternary landscape evolution of the Transdanubian Range. Understanding of this tectonic control and separation from the role of exogenic forces, like wind and rivers are important for societal application of structural geological results. Volcanological research can contribute considerably in solving of these neotectonic and geomorphic questions. Precise reconstruction of volcanic landforms may help to understand landscape evolution as well as the timing and amount of post-volcanic erosion.

Dr. Károly Brezsnýánszky  
Director of MÁFI  
Hungarian Project Leader

Dr. László Fodor  
Editor

# Transtension in the Coso region of the central Basin and Range

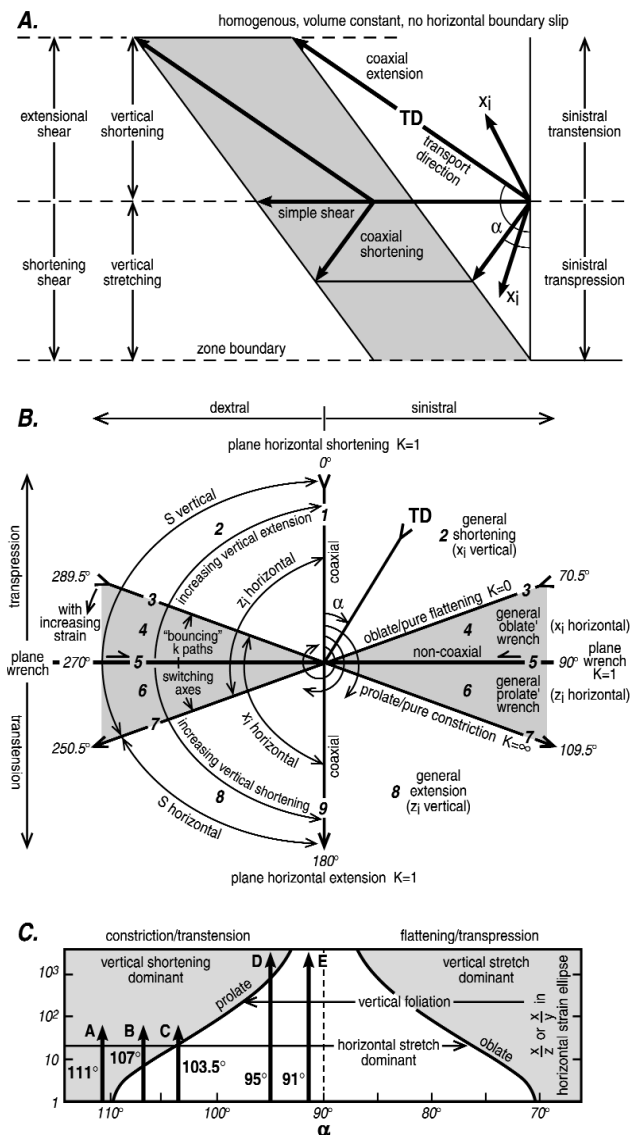
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## Background and problems: kinematic theory of transtension

SANDERSON and MARCHINI (1984), following HARLAND (1971), showed, in their paper on transpression, that most deformation zones involve the oblique relative motion of their boundary walls and, therefore, that the resulting strains combine coaxial and noncoaxial components (Figure 1A). MCCOSS (1986), FOSSEN and TIKOFF (1997, 1998), TIKOFF and FOSSEN (1993, 1995, 1999), FOSSEN et al. (1994), TIKOFF and TEYSSIER (1994), TEYSSIER et al. (1995), DUTTON (1997), TIKOFF and GREENE (1997), TEYSSIER and TIKOFF (1999), KRABBENDAM and DEWEY (1998), DEWEY et al. (1998), and DEWEY (2002) expanded these concepts to, and developed models for, zones of general transpression and transtension. Transtension (Figure 1A) is oblique divergence between bounding plates or blocks, which combines a coaxial orthogonal extension with a deformation zone boundary-parallel, noncoaxial, component; this generates a bulk constrictional strain. The coaxial component determines the rate and amount of crustal/lithospheric thinning and part of the horizontal extension. The non-coaxial component controls the vorticity, the horizontal shortening, and part of the horizontal extension. The instantaneous stretching direction ( $X_i$ ) bisects the angle ( $\alpha$ ) between the direction of divergence (transport direction, TD) and the zone boundary orthogonal. In the brittle regime, normal fault arrays accommodate vertical shortening and horizontal extension whereas simultaneous wrench fault arrays allow horizontal shortening and

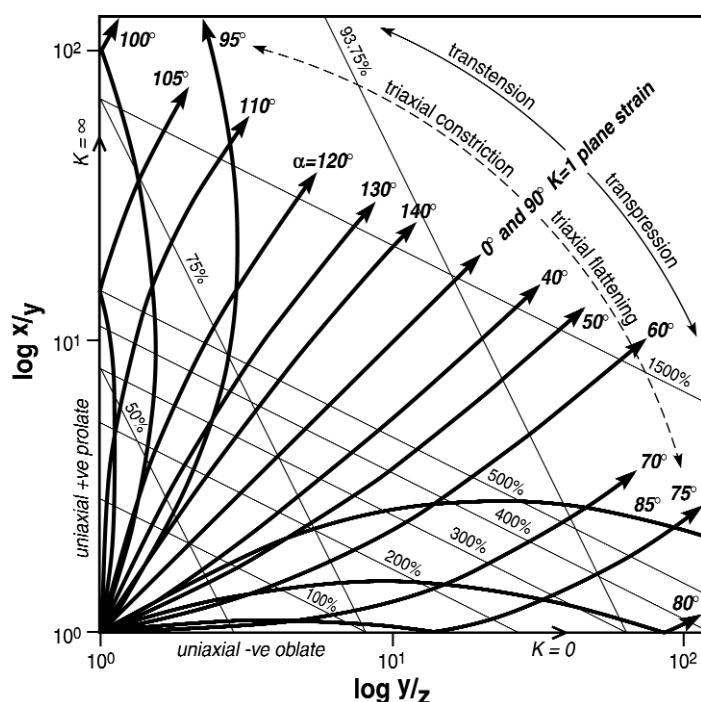
**Figure 1.** A – Transtension and transpression in the horizontal plane.  $X_i$  is the instantaneous stretching direction. B – Fields of transtension and transpression and their fabrics.  $270^\circ > \alpha > 90^\circ$  in transtension. The prolate and oblate lines are at  $19.5^\circ$  to the zone boundary at infinitesimal strain and move towards the zone boundary as strain increases. C – Deformation fields for  $65^\circ < \alpha < 115^\circ$  showing five deformation paths in transtension.



extension. All faults, except those that are vertical and parallel with the zone boundary or TD, rotate either with or against vorticity. Block rotation controls fault slip-direction and sense. This causes serious compatibility problems because blocks of varying size and shape bounded by normal and wrench faults rotate at different rates about vertical and horizontal axes. Where TD is at greater than  $19.5^\circ$  to the zone boundary, the coaxial component, vertical shortening, normal faults, and horizontal foliation dominate. At angles less than  $19.5^\circ$ , the non-coaxial component, horizontal shortening, wrench faults, and vertical foliation dominate at small strains and for TD close to the zone boundary.

Figure 1B indicates the TD and  $\alpha$  convention and the implications at infinitesimal strain;  $90^\circ < \alpha < 270^\circ$  indicates transtension. Figure 1C illustrates the implications of progressive finite strain in transtension and transpression for constant  $\alpha$  (i.e. constant TD) paths (after McCoss, 1986; TEYSSIER et al. 1999). In transtension, for  $109.5^\circ < \alpha < 250.5^\circ$ , progressive constrictional strain yields a dominant vertical (Z) and a lesser horizontal (Y) shortening. For  $270^\circ > \alpha > 250.5^\circ$  and  $109.5^\circ > \alpha > 90^\circ$ , Z is horizontal and Y vertical. The curved line in the transtensional field in Figure 1C is the line of pure prolate constriction ( $K = 8$ ) through which  $\alpha$ -constant strain paths pass, for  $\alpha < 109.5^\circ$ , from a field of dominant horizontal (Y vertical) to dominant vertical (Z vertical) shortening (TEYSSIER and TIKOFF, 1999). As  $\alpha$  lessens towards  $90^\circ$ , the transient prolate strain line is transected at higher strains. For  $\alpha = 95^\circ$  (path D), the horizontal stretch has to be 1000 to intersect the prolate line. Therefore, for realistic stretches up to about 10 (path B), for example  $\alpha = 103.5^\circ$  (path C), transtension generates Z-horizontal strains, which occur either at low strain or at very small, noncoaxially-dominated  $\alpha$  values (path E).

The logarithmic FLINN (1962) plot (Figure 2), allows the representation of any strained state by the  $K$  value ( $K = a-1/b-1$ , where  $a = X/Y$ ,  $b = Y/Z$ , and  $X > Y > Z$  are the axes of the finite deformation ellipsoid). At  $K = 1$  (plane or biaxial strain), coaxial (irrotational) and noncoaxial (rotational) strains are, respectively, orthorhombic (at  $\alpha = 0^\circ, 180^\circ$ ) and monoclinic (at  $\alpha = 90^\circ, 270^\circ$ ), and are linear. At  $K = \infty$ , strains are prolate (uniaxial-positive) and, from  $K = \infty$  to  $K = 1$ , are triaxial in the constrictional field. Volume-constant transtension with no zone length change generates  $K > 1$  constrictional strains. All  $\alpha$  values for TD neither parallel with nor orthogonal to the zone boundary generate non-linear (varying  $K$ ) deformation paths, which, for  $90^\circ < \alpha < 109.5^\circ$  and from  $250.5^\circ < \alpha < 270^\circ$ , “bounce” off the ordinate if they pass through the prolate line, meaning that the deformation ellipsoid passes instantaneously through a pure prolate form at which point the Y and Z axes switch.



**Figure 2.** Logarithmic Flinn Plot for constant  $\alpha$  deformation paths (heavy lines) in constriction/transtension and flattening/transpression; thin lines indicate percent elongation or shortening

with the coaxial zone is parallel with TD and does not rotate; that with the non-coaxial zone is normal to TD and rotates counterclockwise. Vertical shortening is greater in the coaxial zone and the normal fault arrays change trend across the boundary. The rotating boundary with the non-coaxial zone is a vertical transition from normal/wrench to wrench faulting. Elevation changes across the transitions, dropping from the non-coaxial zone and again into the coaxial zone. Figure 3J portrays a similar kinematic picture when  $90.5^\circ < \alpha < 109.5^\circ$  at modest strain. Here, the transtensional and noncoaxial zones have similar normal/wrench-fault patterns, but the transition from wrench/normal faults to normal faults is at the non-rotating transtension/coaxial transition. Again, elevation changes across the transitions but the change at the coaxial boundary is larger because of the greater constriction in the transtensional zone. These geometries will be modeled in the proposed research.

In plan view of homogeneous instantaneous dextral transtension at  $\alpha = 210^\circ$  (Figure 4A),  $Y_i$  is horizontal and normal to  $X_i$ , while  $Z_i$  is vertical. One line of no infinitesimal strain (LNIS) is parallel with the zone boundary, the other is normal to TD. Folds of horizontal surfaces would be expected to form with hinges parallel with  $X_i$  and extensional features such as

transpression. Figure 3 illustrates, in plan view, some geometric and kinematic factors that contribute to transtensional zones (DEWEY 2002). Figure 3B illustrates transtension when  $109.5^\circ < \alpha < 180^\circ$  and Figure 3C when  $90^\circ < \alpha < 109.5^\circ$  with their associated structures. Figures 3E, F, and G show transtensional systems in which zone boundary walls are non-parallel and with a constant slip vector such that the strain rate increases as the zone narrows and, consequently, vertical shortening increases with a consequent decrease in elevation. Transtension cannot be viewed independently of the way in which the zone ends. Figure 3I is a transtension zone where  $109.5^\circ < \alpha < 180^\circ$  and the zone is terminated by purely non-coaxial and purely coaxial zones. The boundary

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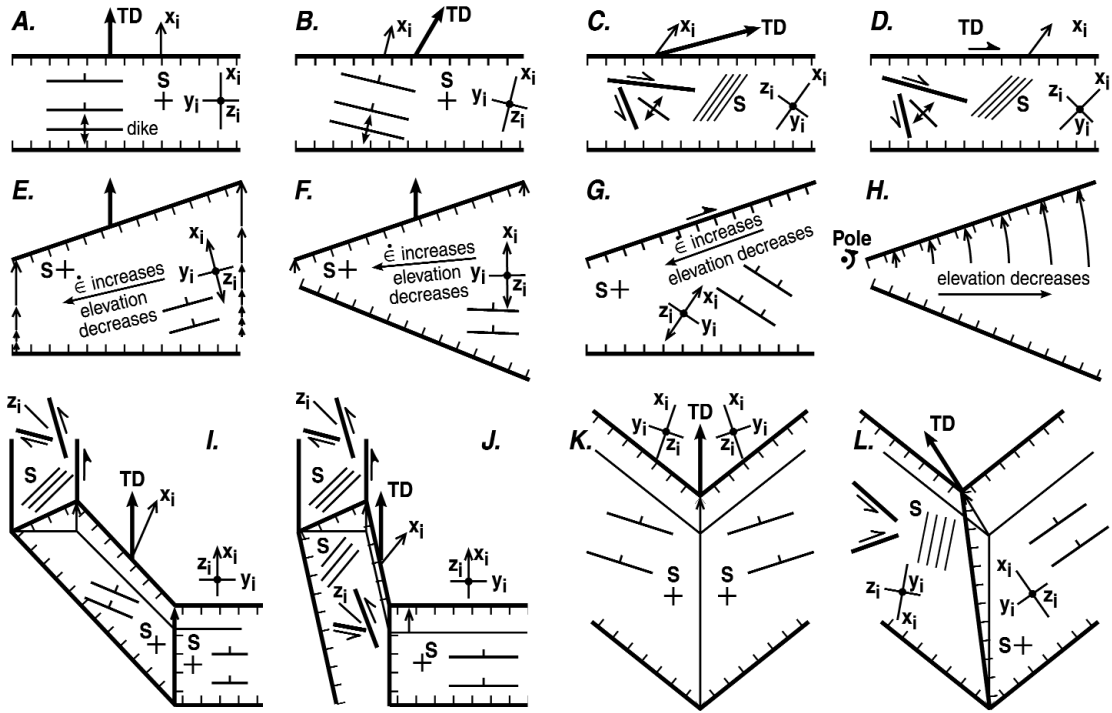


Figure 3. Twelve kinematic configurations for transtensional geometries. In I through L, thin lines are initial zone boundary positions linked by arrows to current (bold lines) zone boundary positions. Lines with single tick mark are normal faults

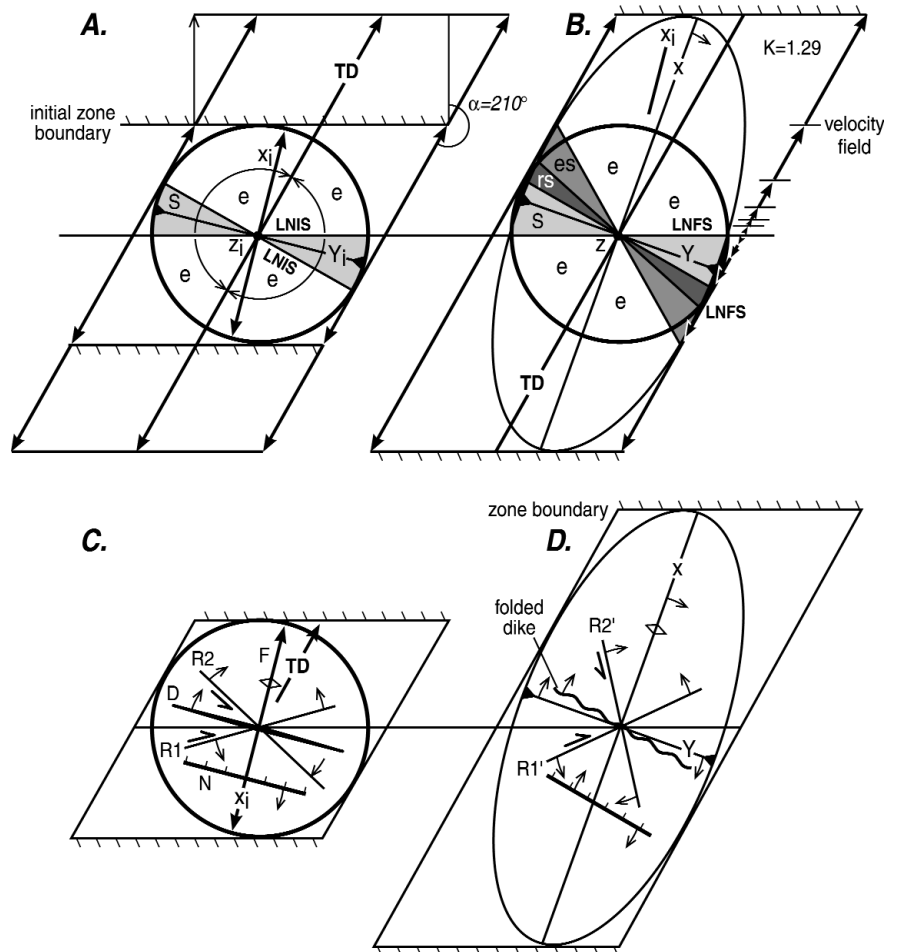


Figure 4. Transtensional strain in plan view. A – Instantaneous strain showing the two lines of no infinitesimal strain (LNIS), the fields of shortening (s) and elongation (e), and the incremental strain axes (principal shortening direction  $Z_i$  vertical, secondary shortening direction  $Y_i$  horizontal, and extension direction  $X_i$  horizontal). B – Finite strain at 100% stretch showing finite fields of shortening (s), reduced shortening (rs), lines now elongating but shorter than original length (es), and elongation (e). C – Orientation of structures at infinitesimal/very small strain; F = fold hinge, D = dike, N = normal fault, R1/R2 = wrench faults. Structures in the obtuse angle between TD and the zone boundary rotate clockwise with vorticity, those in the acute angle rotate counterclockwise against vorticity. D – Orientation of structures at 100% extension assuming purely internal rotation

tension gashes and dikes in the  $Y_i/Z_i$  plane with normal faults striking parallel with  $Y_i$  (Figure 4B). Conjugate extensional faults with dips of about  $60^\circ$  should intersect in  $Y_i$ . Shortening in  $Y_i$  could yield a second conjugate set of Riedel (R1) and anti-Riedel (R2) faults (Figure 4C). The LNIS parallel with the zone boundary is stationary with respect to material points; lines parallel with it neither rotate nor change in length. The LNIS normal to TD is a line of maximum shear-strain rate and zero longitudinal strain but, in contrast, is a line through which material points move. Lines rotate towards directions of zero angular velocity at rates inversely proportional to the shortening/elongation rates. Lines in the obtuse angles between the zone boundary LNIS and TD rotate clockwise with the kinematic vorticity ( $W_k = \cos \beta$  where  $\beta$  is the minimum angle between directions of zero angular velocity; Tikoff and Fossen, 1995); in transtension,  $0 < W_k < 1$ . Lines in the acute sections between TD and the zone boundary rotate counterclockwise against vorticity. In the acute angles between the two LNIS, lines shorten; in the obtuse angles, they lengthen. After 100% extension (Figures 4B, D), the finite strain axes X and Y have internally-rotated clockwise with respect to the zone boundary. Similarly, fold hinges, anti-Riedels (R2), tension gashes, dikes and normal faults have rotated clockwise, whereas the Riedels (R1) have rotated counterclockwise, towards TD. Between the orthogonal to TD and the clockwise-rotating line of no finite strain (LNFS), along which lines are their original length, a field of reduced shortening (rs) is one in which lines, although elongating, are shorter than their original length. The field of extension (e) is one in which all lines are longer than their original length. Dikes and normal faults are folded with vertical and steeply-dipping hinges in the ZX plane and horizontal surfaces are folded with hinges parallel with X whereas the Riedels (R1) and anti-Riedels (R2) are extended (Figure 4D). The LNIS and LNFS are, respectively, the intersections of the horizontal plane with surfaces of no infinitesimal and no finite strain; these are the circular sections of plane strain, the cones of uniaxial strain and the more complicated surfaces of triaxial strains. Normal faults intersecting in Y have flattened in dip and their strike has rotated clockwise with vorticity. Very likely, only one member of the conjugate set will develop fully to allow block rotation in a fault array and to avoid the conjugate intersection problem. The fault slip direction plunge (slickensides) is oblique and changes, combining simultaneous block rotations around horizontal and vertical axes. The relationship between rotating normal faults and dikes is likely to be complicated from late dikes cutting normal fault arrays to early dikes segmented and rotated in normal fault blocks.

Figure 5 shows progressive dextral transtensional finite strain at  $\alpha = 253^\circ$  to illustrate structures and fabrics in three different crustal levels and rheologies, B in the brittle upper crust, D at a deep crustal level where homogeneous ductile strains

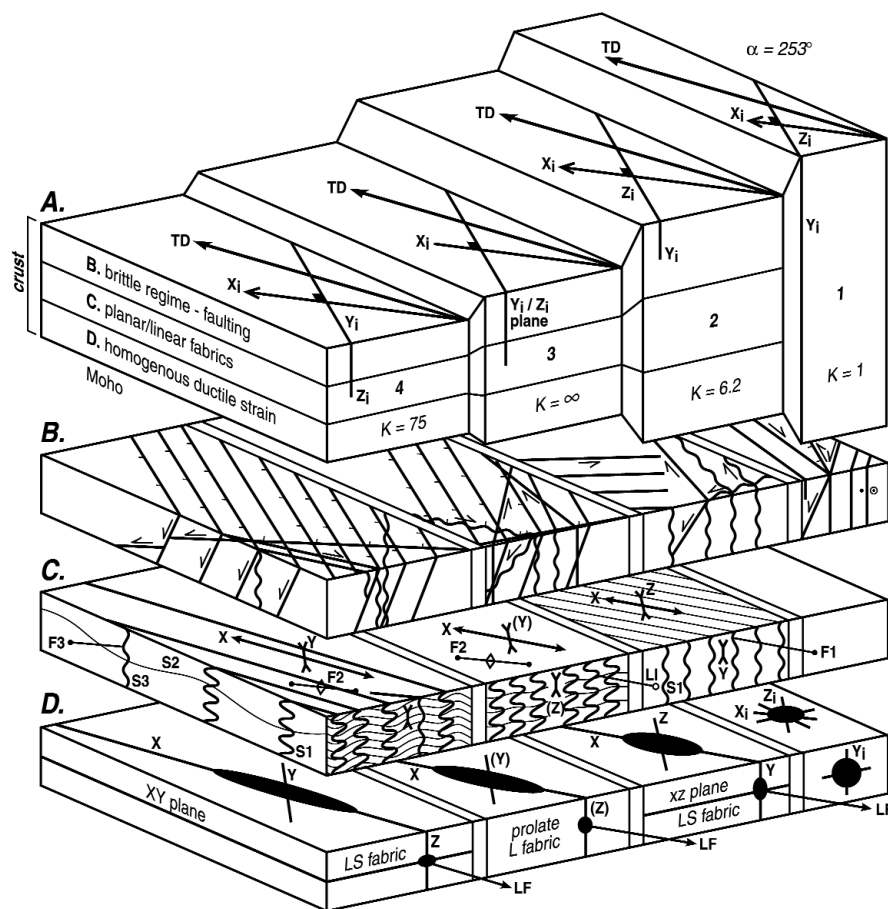


Figure 5. Structures and fabrics of progressive (1, 2, 3, 4) transtension for  $\alpha = 253^\circ$  at three structural levels

are reflected in gneissic fabrics, and C where foliations, lineations, shear bands and kink bands are developed. Zone 1 represents rocks at infinitesimal strain and zones 2, 3, and 4 increasing transtensional strain to illustrate both a time and space sequence. The strain path passes from a zone (2) of vertical Y through pure prolate constriction (3) to a zone (4) of horizontal Y. C2 has a vertical foliation (S1) and a horizontal stretch in X accommodating, respectively, shortening and elongation in the horizontal plane. The importance of this is that polyphase deformation can be generated by progressive finite strain in a constant incremental strain field at a diminishing horizontal strain rate as the zone widens. In the brittle layer, conjugate vertical wrench faults intersecting in  $Y_i$  develop in zone 1 to accommodate the dominant horizontal shortening, but simultaneous normal faults striking parallel with  $Z_i$  are necessary to accommodate vertical shortening in  $Y_i$ . With increasing strain (B2), the wrench faults rotate towards

TD around a vertical axis ( $Y_i$ ), and the normal faults rotate around both  $Y_i$  and a horizontal axis (strike of the fault surface), yielding a constantly-changing oblique-slip direction on the fault surface. Constriction could buckle both wrench and normal faults. As fault blocks rotate, the faults move into less-favorable orientations for slip as the shear/normal stress ratio diminishes and, hence, new fault arrays are likely to form. Therefore, it is probable that, with progressive bulk deformation, there will be an increasing number of faults and fault blocks, some active and some inactive, with a diminution in fault block size.

### Fault block rotation in transtension

There are some difficult kinematic problems associated with fault block rotation in transtensional zones (DEWEY 2002). Faults form, generally, in domains or arrays dominated by one of the conjugate orientations, whether on surfaces of maximum shear stress (governed by internal friction) or on favorably-oriented surfaces of weakness (governed by sliding friction; PRICE 1966). This allows large bulk strains by fault block rotation opposite to fault slip and obviates the conjugate intersection problem. Transtensional constriction favors a wrench/normal fault combination. Consider domains of conjugate wrench faults in transtension (Figure 6), forming on surfaces of maximum shear stress, and rotating in opposite senses to orientations at which the shear/normal stress ratio is at a minimum for slip to occur. Now, fault slip and block rotation cease (Figure 6B); new faults must form if bulk deformation continues. If the blocks behave completely rigidly, gaps develop at domain boundaries, perhaps expressed as small extensional/rotational sedimentary basins. However, there are three major compatibility problems. First, the transtensional zone shortens parallel with its boundary (Sh1 and Sh2). Secondly, the zone widens by different amounts for each domain (33% versus 278%). Thirdly, wrenching cannot account for the vertical shortening, which needs simultaneous arrays of normal faults striking, initially, parallel with  $Y_i$  (Figure 6C). If the normal fault blocks remain rigid as they rotate simultaneously around horizontal and vertical axes, the zone lengthens then shortens as the LNIS is passed. Perhaps buckling could account for the vertical shortening of the wrench faults and the horizontal shortening of the normal faults. Alternatively, if the origin of the conjugate wrench faults is determined by incremental shortening in  $Y_i$ , then rotated as passive markers to maintain complete compatibility of zone length and width (Figures 6A, points 1, 2, 3, 4), the blocks change shape. Only if the domains rotate to orientations of minimum shear/normal stress and then cease slipping do the domains generate zone width compatibility problems. Passive, compatibility-driven rotation involves substantial internal deformation of the blocks, certainly sufficiently large to be expressed in small-scale brittle rock fabrics.

A possible approach to this problem is illustrated in Figure 6C where synchronous wrench and normal fault systems take up the lesser horizontal and greater vertical shortening respectively (DEWEY 2002). The incompatibility of the difference in zone widening contribution between the R1 (167%) and R2 (33%) sets is because rotation of each is to the minimum shear/normal stress ratio. If zone width compatibility is maintained, either R1 does not reach, or R2 rotates beyond, the stress limit orientation. A logical solution to block rotation in the brittle transtensional regime is that, although the bulk instantaneous strain/stress field determines the wrench/normal fault regime at infinitesimal or very small strains, compatibility dictates their rotation, buckling and intersection relation-

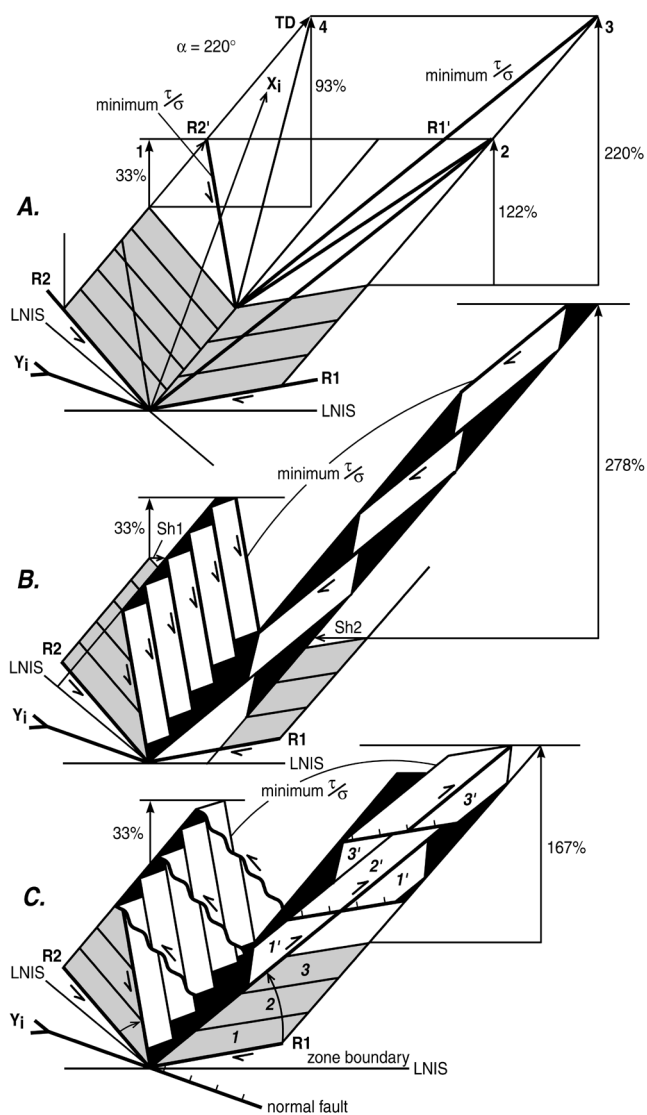


Figure 6. Rotation of and relationships between wrench faults and normal faults in transtension



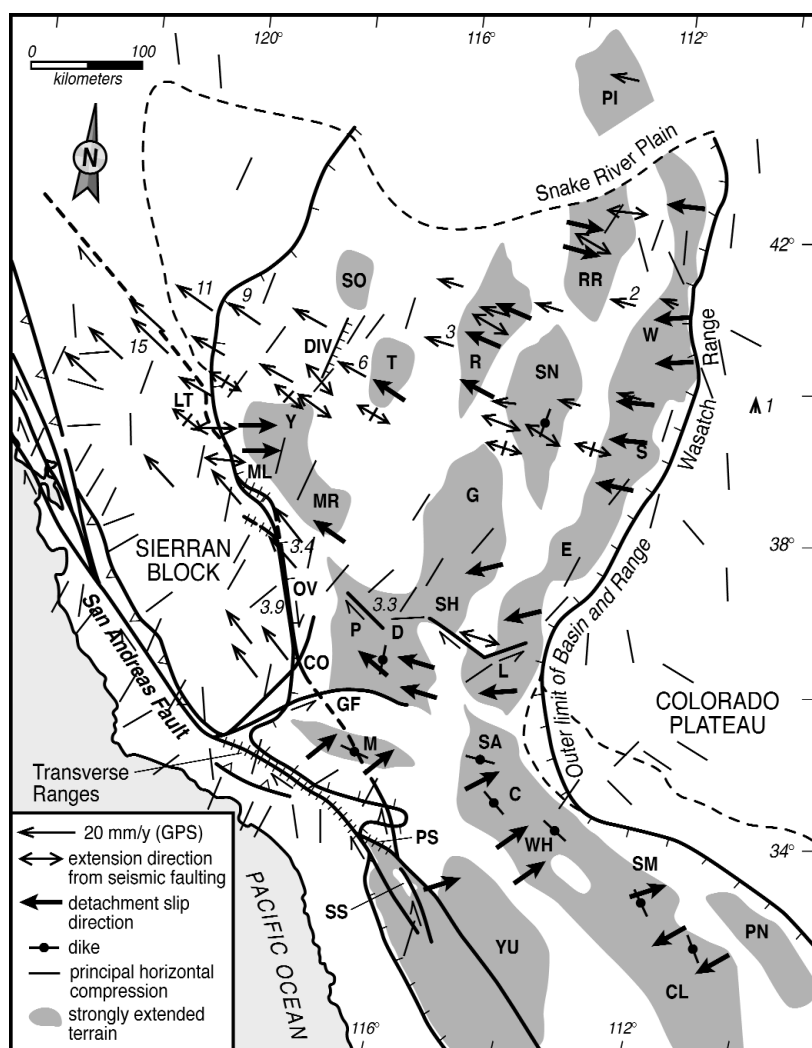
ships, where elongation of the wrench fault blocks is effected by the normal faults and the small, early, shortening of the normal faults by buckling (DEWEY 2002). Some consequences of this are that slip directions and senses and stress patterns are dictated by block rotation and that, as faults lock at unfavorable intersections and orientations, new fault systems increase fracture penetration to decrease block size. Outcrop and hand-specimen-scale tensile and shear fractures and some ore bodies are, probably, a response to these complexities.

### Basin and Range province

The northern segment of the Basin and Range province (Figure 7), from the Snake River plain to the Tahoe/Walker Lane/Las Vegas seismic (dextral shear) zone, is about 600 km wide between Winnemucca and the Wasatch Front and is characterized by roughly boundary-orthogonal extension of about 100% [from GPS-derived velocity fields (OLDOW et al. 2001; THATCHER et al. 1999; WERNICKE et al. 2000), seismic faulting, dikes, and the horizontal orthogonal to the principal horizontal compression (ZOBACK and ZOBACK 1989)]. Earlier detachment slip directions are, similarly, roughly boundary-orthogonal; the segment has undergone extensional, bulk, roughly plane strain. The GPS-derived displacement (THATCHER et al. 1999; WERNICKE et al. 2000) between its boundaries gives an average horizontal extensional strain rate of  $0.58 \times 10^{-15} \text{ s}^{-1}$ . If, however, elevation is a guide to extensional crustal thinning, concentrating extension in the Lahontan and Bonneville basins gives a horizontal strain rate of  $0.87 \times 10^{-15} \text{ s}^{-1}$ , which would account for the jump in GPS velocity across the Dixie Valley Fault Zone (THATCHER et al. 1999; WERNICKE et al. 2000) and the elevation changes. Concentrating extension west of the Dixie Valley Zone gives a rate of  $2.0 \times 10^{-15} \text{ s}^{-1}$ .

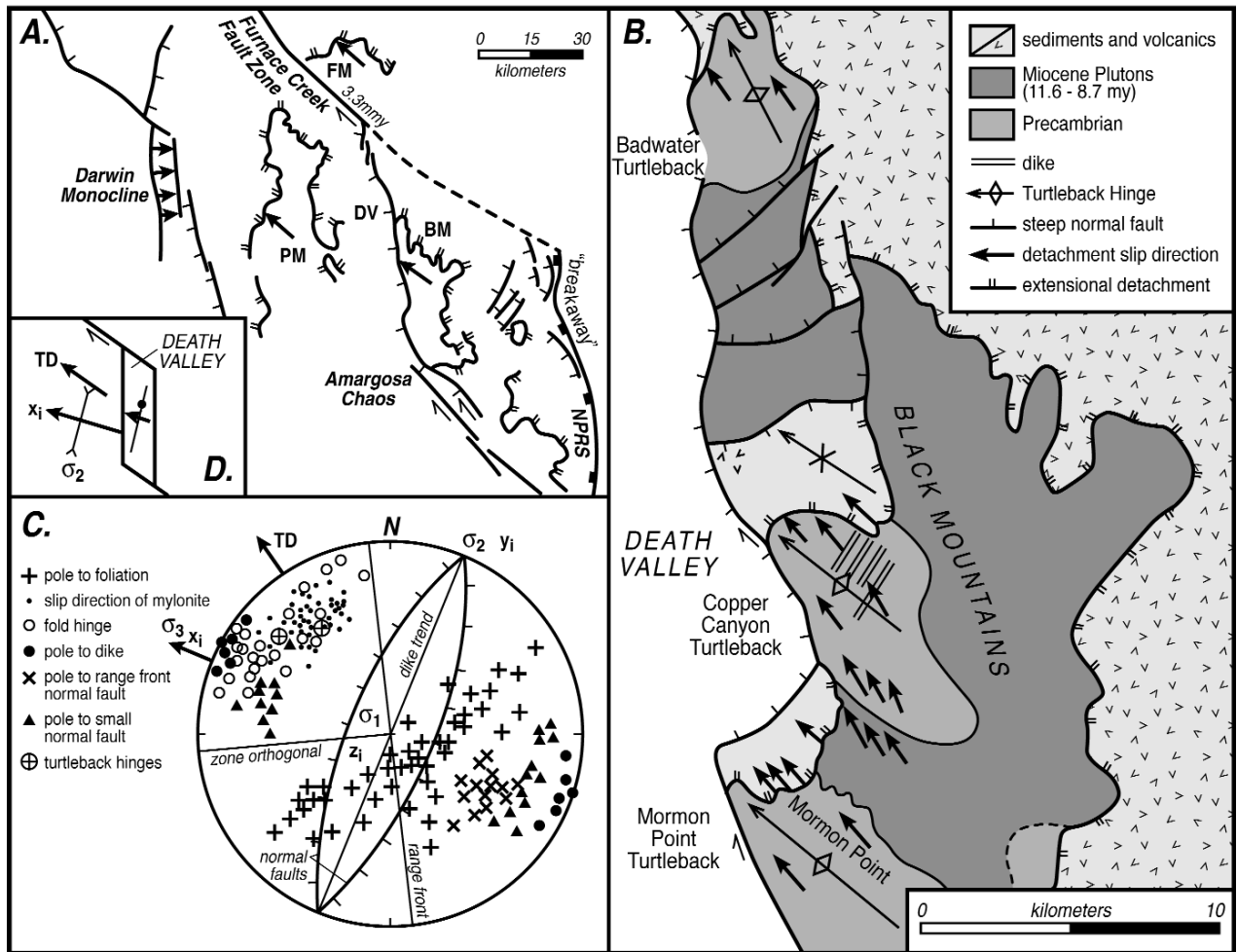
The central segment of the Basin and Range, between the Sierra Nevada block and the Colorado plateau, and the Lake Tahoe-Las Vegas seismic zone and the Garlock fault, is about 300 km wide (Figure 7). Within the Walker Lane, GPS data reveal dextral shear and clockwise rotation (OLDOW et al. 2001). The relative motion between the Sierra Nevada and the Colorado

Plateau is at about  $23^\circ$  to the zone boundaries indicating dextral transtension. The horizontal extensional strain rate is  $0.9 \times 10^{-15} \text{ s}^{-1}$ , if the whole zone is extending, to  $3.8 \times 10^{-15} \text{ s}^{-1}$  if only the more active seismic zone between Death Valley and the Sierras is extending. The higher extensional strain rate of the central segment probably accounts for its lower elevation with part of Death Valley, the Badwater Basin, below sea level. A transtensional origin for the central segment is supported by the geology of Death Valley (Figure 8). BURCHFIELD and STEWART (1966) first suggested that Death Valley is a transtensional pull-apart terminated by the dextral Furnace Creek and Amargosa fault zones. The active Furnace Creek fault zone continues south-eastwards as an inactive structure with di-



**Figure 7.** The Basin and Range Province. Data mainly from ARGUS and GORDON (2001) DIXON et al. (1995, 2000), THATCHER et al. (1999), WERNICKE (1992), WERNICKE et al. (2000), ZOBACK and ZOBACK (1989).

Rates from GPS in mm/yr. Highly-extended terrains from WERNICKE (1992). C = Chemehuevi, CL = Catalina, CO = Coso/China Lake, D = Death Valley, DIV = Dixie Valley/Fairweather, E = Escalante, G = Grant, GF = Garlock fault, L = Lake Mead, LT = Lake Tahoe, M = Mohave, ML = Mammoth Lake, MR = Mineral Ridge, OV = Owens Valley, P = Panamint, PI = Pioneer, PN = Pinalena, PS = Palm Springs, R = Ruby, RR = Raft River, S = Sevier, SA = Sacramento, SH = Sheep, SM = South Mountains, SN = Snake, SO = Sonoma, SS = Salton Sea, T = Toiyabe, W = Wasatch, WH = Whipple, Y = Yerrington, YU = Yuma

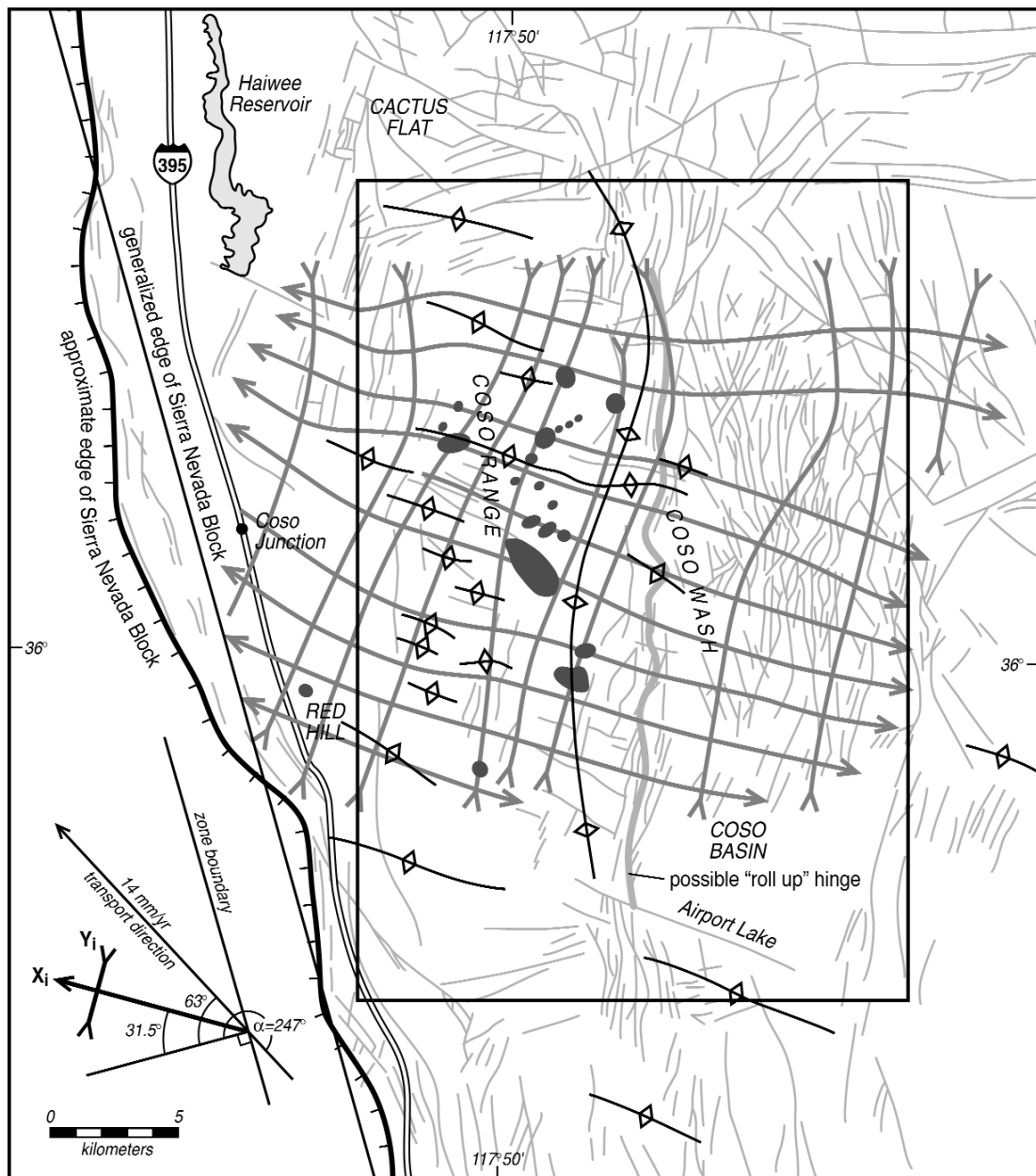


**Figure 8.** Death Valley. A – Regional map: BM = Black Mountains, DV = Death Valley, FM = Funeral Mountains, NPRS = Nopah/Resting Springs Range, PM = Panamint Mountains. B – Geologic map of the Black Mountains. C – Lower hemisphere Wulff net projection of structural data from the Black Mountains. D – Dextral transtensional model for Death Valley. Data from HOLM (1995), HOLM et al. (1994), HOLM and WERNICKE (1989), WERNICKE and SNOW (1998), KEENER et al. (1993) and the writer's observations

minishing offset (SNOW and WERNICKE 1989) to terminate at the northern end of the Nopah/Resting Springs breakaway zone. The Black Mountains detachment has three northwest-plunging antiformal “turtle-backs” of Precambrian basement whose surfaces are the mylonitic carapace (MILLER 1991) to the detachment (Figure 8B). The mylonites are deformed by small folds whose hinges are sub-parallel with the slip direction, as determined by slickensides, and with the pole to the mylonite foliation girdle (Figure 8C), probably defining TD, roughly parallel with the Furnace Creek fault zone. Poles to dikes and the horizontal trace of the orthogonal plane to small normal faults are taken to define the instantaneous stretching direction  $X_i$ . Hence, a dextral transtensional system in Death Valley (Figure 8D) can be defined involving a zone boundary striking just west of north with TD and  $X_i$  trending at  $327^\circ$  and  $293^\circ$  respectively. McKee (1968) derived a long term slip rate of 9.3 mm/yr for the active Furnace Creek fault zone, accounting for over half of the long term relative motion between the Sierra Nevada and the Colorado Plateau, which suggests that the higher strain rate of  $3.8 \times 10^{-15} \text{ s}^{-1}$  is the more likely. Bulk and small-scale structures suggest a horizontal stretch in X of 150%, a vertical shortening in Z of 47%, and a horizontal shortening in Y of 25% giving a bulk  $K = 5.62$ . Progressive constriction is suggested by the folding of the older extensional detachment versus the relative planarity of the younger range-front faults. The pattern of ranges in the Basin and Range is instructive. In the northern segment, ranges strike, fairly uniformly, NNE, consistent with roughly coaxial WNW extension. Counterclockwise vertical-axis rotations (HUDSON and GEISSMAN 1987; LI et al. 1990) appear to be related to small transfer zones. In the central segment, not only is elevation lower but range orientation is very variable and range length is much shorter. Local clockwise and counterclockwise vertical-axis rotations are evident, both here and in the Walker Lane, from paleomagnetic (PETRONIS et al. 2002), GPS (OLDOW et al. 2001), and structural/stratigraphic (SNOW and PRAVE 1994) data but the shapes of rotated blocks and their marginal compatibility relationships are unclear. These rotations are consistent with dextral transtension in the central segment, where synvorticity clockwise rotation would be expected to be dominant over countervorticity, counterclockwise rotation.

### The Coso region

The Coso area has been well mapped in outline (DUFFIELD and BACON 1981). The volcanically and seismically active Coso/China Lake area (CCL) (Figure 9) appears to be the site of a newly-developing/nascent transtensional system as deformation moves westwards. The area occupies a critical tectonic position and role in being the principal locus of transtensional strain that takes up a large part of the 14 mm/yr relative motion between the Sierra Nevada and the Colorado Plateau (ARGUS and GORDON 2001; DIXON et al. 1995) at a transtensional angle  $\alpha$  of about  $247^\circ$  (Figure 9). GPS data indicates a separation rate of the Sierra Nevada block from the Panamint range of about 12 mm/yr. Although there is some seismicity across the whole of the central Basin and Range, intense seismicity and Pleistocene/Holocene bimodal volcanic activity is concentrated in the CCL between the eastern margin of the Sierra Nevada Block and the Argus Range; therefore, the CCL can be considered as a nascent transtensional system between the Sierra Nevada and the Argus Range boundaries trending about  $343^\circ$  (Figure 9). The CCL is part of a seismically and volcanically active dextral transtensional megashear that links the Landers fault, via the Blackwater fault, with the Owens Valley, Mammoth and Tahoe and is, at first sight, a left-stepping transpressional connection from the Airport Lake fault zone to the Owens valley. However, the NNE shortening attributed to trans-



**Figure 9.** Outline structural map of the Coso/China Lake area from DUFFIELD and BACON (1981) and satellite imagery. Grid-instantaneous extension and shortening directions from seismic data (J. UNRUH verb. comm.), Black Pleistocene/Holocene volcanic centers, thin black lines: fault traces.

pression is generated probably by the NNE horizontal shortening of regional transtension. If transtensional strain is concentrated between the Sierra Nevada and the Argus Range at a displacement rate of 12 mm/yr, the horizontal extensional strain rate is about  $10^{-14}\text{s}^{-1}$ . This is one of the highest regional rates known to the writer and renders the CCL an unusual and exciting place in which to study transtension because, although block rotation has been occurring for only three million years, rotation rates could be as high as  $10^\circ/\text{My}$ .

The fault pattern in the CCL suggests co-eval normal faulting that effects WNW extension and vertical shortening, and conjugate wrench faulting that effects WNW extension and NNE shortening, a pattern deduced by Dr. Jeff Unruh (verb. comm.) from seismic data (Figure 9) and also seen in the Tahoe region (Dr. Richard Shweickert, verb. comm.). There is a striking correlation between the WNW extension/NNE shortening directions deduced from seismicity and the instantaneous stretching (Xi)/shortening (Yi) directions deduced from the  $343^\circ$ -trending transtensional zone boundaries and the  $316^\circ$ -trending relative motion of the Sierra Nevada block (Figure 9). The NNE shortening is also taken up by WNW folds that form basement highs (Figure 9) and partly by NNW dextral and NE sinistral wrench faults. Many of the CCL faults are non-planar and are probably folded. The normal faults and the NE wrench faults “should be” rotating clockwise whereas the NNW wrench faults “should be” rotating counterclockwise; there is as yet insufficient published paleomagnetic data to address this problem. Chris Pluhar of UC Santa Cruz has discovered  $18^\circ$  clockwise rotations in Wild Horse mesa east of Coso Wash (Dr. Frank Monastero, verb. comm.) but the shape and bounding faults of the rotating blocks are unclear. The earliest (3ma) lavas are pervasively faulted whereas the youngest lavas are cut by few faults; this may prove to be important in determining strain rates with time.

Future research in the transtensional zone that bounds the east side of the Sierra Nevada for five hundred kilometres between Lake Tahoe and China Lake (Figure 7) should address the following questions relating to how normal and wrench faults intersect and rotate in transtension and how this effects brittle strain, fluid flow, and the localization of igneous conduits. These questions are appropriate, also, for any transtensional zone.

1. To develop a coherent kinematic theory of transtension in the upper brittle crust that can explain structural and palaeomagnetic data in transtensional zones.
2. To use the CCL as a laboratory to iteratively build and test transtensional models.
3. To study how brittle blocks of varying shape and size bounded by normal and wrench faults deform especially how compatibility problems at rotating block margins and fault intersections are solved. The shapes of blocks and the relative ages of their bounding faults should be mapped and analysed to determine whether they operated in synchronous domainal arrays of normal and wrench faults, in more complicated systems of intersecting normal and wrench faults, or in polyphase alternating normal and wrench systems. The shapes and slip kinematics of faults should be analyzed to determine whether they are folded and, if so, are older less planar than younger faults, and can shortening strains be deduced from them?
4. To determine how the expected NNE shortening is accommodated, whether by wrench faulting alone or the buckling of normal faults, or a combination of both and/or other mechanisms.
5. To determine the role of block intersection compatibility problems in localizing pathways of fluid flow in the Coso geothermal field, especially volcanic conduits.
6. Geologic criteria such as tilted young sediments and volcanics, and paleomagnetic data should be used to assess the amounts and timings of fault block rotations around vertical and horizontal axes. There is a substantial amount of paleomagnetic data being acquired by Chris Pluhar and it is not proposed to duplicate this effort.
7. Particular attention will be paid to fine-scale structural analysis at fault intersections, bends and terminations to determine strains resulting from compatibility problems arising from rotations of normal and wrench fault bounded blocks with complex shapes and different rotation senses and rates.
8. Fracture systems should be analyzed to detect possible volume changes at fault intersections and the role of volume increases and compatibility gaps in localizing fluid flow, mineralization, and volcanic conduits.
9. The intense and pervasive seismicity of the CCL should be analyzed to determine fault slip senses, and to derive a velocity field from moment tensor sums. An instantaneous strain field for the region (Figure 9) from seismic data supports a constrictional transtensional strain field with NNE and vertical shortening, and WNW extension.
10. Geophysical data, including magnetic, gravity, heatflow, and seismic reflection, should be incorporated into a synthesis of CCL transtension. The lithosphere appear to be very thin and weak under CCL (SMITH et al. 2002). There is a very low velocity zone at four to five kilometers beneath CCL, probably a magma chamber.
11. Work should continue on building theoretical kinematic models for transtension in the brittle field involving the simultaneous rotation of normal and wrench faults and the solution to compatibility problems at block margins. In addition to the usual regional monoclinic symmetries of transtension, more local triclinic symmetries involving an added noncoaxial component generated by extensional simple shear in a vertical plane normal to the zone boundaries should be modeled. Also, transtension involving non-parallel boundaries (Figure 3E–H) should be modeled. Preliminary work on triclinic symmetries and non-parallel boundaries indicates that these introduce further layers of complexity to transtension, particularly in the brittle field.



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# Neotectonics in the Trinidad and Tobago, West Indies segment of the Caribbean-South American plate boundary

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## Introduction

The twin-island nation of Trinidad and Tobago offers a rich source of neotectonic signals, which, over the past decade, I have been working (collaboratively) to quantify, understand, and relate to regional tectonic processes (Figure 1). The signals include horizontal motions and strain related to active dextral wrenching in the Caribbean-South American plate boundary zone, subsidence and uplift at steps and bends the active strike-slip fault system, and sinking of the obducted oceanic arc-forearc lithosphere of Tobago. This paper provides a summary of my work to date in the region, it includes discussions on our use of far- and near-field geodesy to study the horizontal motions, palaeoseismology to search for fossil earthquakes on the Central Range Fault, the principal active strike-slip fault in Trinidad, and geomorphology to study geologic-time-scale records of vertical motions. Because some of the geology developed during an earlier phase of oblique convergence affects the neotectonics, I also include a brief discussion of these features and events. The techniques discussed here, particularly those that exploit neotectonic signals accumulated over long times, may be applicable to studying the slower, but possibly significant, intraplate neotectonics of Hungary.

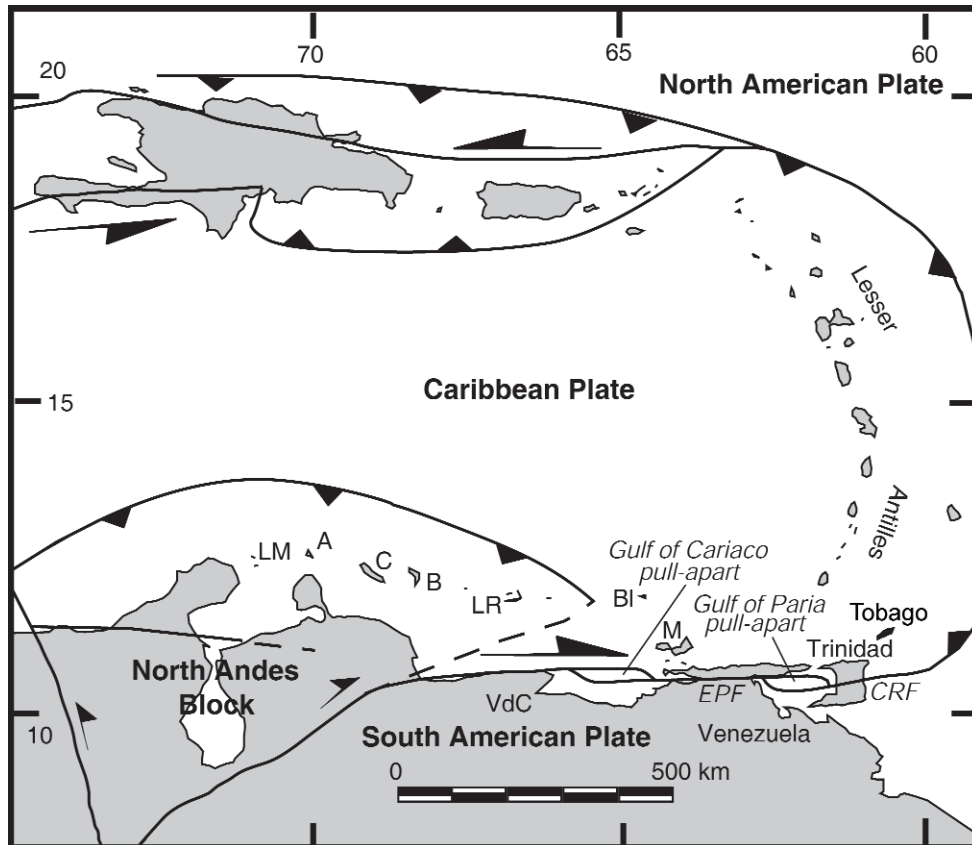
## Regional background

### *Caribbean-South American plate tectonic setting*

MOLNAR and SYKES (1969) first suggested that the lithosphere in the Caribbean region moves as a single, rigid plate. However, with only geologic methods available until the 1980s and 1990s, the precise direction and rate of the current motion of the Caribbean plate relative to its neighbours remained highly debated, e.g., compare the interpretations of SPEED (1985), ROBERTSON and BURKE (1989), SPEED et al. (1991), and DENG and SYKES (1995). More recently, GPS (Global Positioning System) data became available, which now clearly demonstrates that the northern and southern boundaries between the Caribbean and neighbouring North and South America plates are currently east-west trending, wrench-dominated (strike-slip) boundaries developed in continental and previously accreted oceanic and arc lithospheres (DIXON et al. 1998; WEBER et al. 2001a; PEREZ et al. 2001; MANN et al. 2002) (Figure 1).

In WEBER et al. (2001a), we used GPS data from eight Caribbean plate-interior and plate-edge sites and demonstrated that the Caribbean plate is rigid to at least  $\pm 1.5$  mm/yr, the approximate noise level in the GPS velocity data. WEBER et al.'s (2001a) pole of rotation derived from these recent GPS measurements indicates that, relative to the South American plate, the Caribbean plate currently moves ~eastward at  $20 \pm 2$  mm/yr. The GPS results of PEREZ et al. (2001), although focused on more local motions in Venezuela, confirmed this plate-wide result.

The presence of pervasive northeast-trending Neogene contractional structures (PEREZ and AGGARWAL 1981; SCHUBERT 1981; SPEED 1985; VIERBACHEN 1984) developed in well-dated Neogene foreland basin deposits, which get younger to the east (SPEED 1985), led many workers to infer that active contraction is occurring in the plate boundary. It is now generally agreed upon that these structures reflect a pre-mid-Miocene phase of Caribbean-South American



**Figure 1.** Regional structural geology and plate tectonic setting of Trinidad and Tobago study area. The Caribbean plate currently moves ~20 mm/yr eastward relative to South America (WEBER et al. 2001a) along the right-stepping El Pilar (EPF) and Central Range (CRF) strike-slip faults and across associated Gulf of Paria and Gulf of Cariaco pull-aparts. The Lesser Antilles is a modern arc developing as the Caribbean plate overrides and subducts Atlantic lithosphere of the North and South American plates. Older oceanic fore-arc and arc lithosphere that was obducted over and accreted to South American continent crops out mainly as a series of submerged islands including: Tobago, Margarita (M), Blanquilla (BI), Los Roques (LR), Bonaire (B), Curacao (C), Aruba (A), and Los Monjes (LM); the Villa de Cura nappe (VdC) is its largest “trapped” onland exposure

oblique contraction and that most are “fossil” transpressional structures. PINDELL et al. (1998) used regional geologic data, in part from Venezuela and Trinidad, to construct a semi-quantitative record of Caribbean-South American relative plate motion back through the entire Cenozoic. According to the PINDELL et al. (1998) geologic models and plate motion reconstructions, the margin accommodated oblique convergence from 59 Ma to 12 Ma, then began experiencing pure wrenching at ~10 Ma. According to the GPS results of WEBER et al. (2001a), the dextral wrenching phase continues today. Active contractile structures are developing only along restraining bends and strike-slip faults that are oblique to local plate motion azimuths (WEBER et al. 2001a); active extensional structures are developing along right-steps and pull-apart basins (SCHUBERT 1985; BABB and MANN 2000; FLINCH et al. 2000) (Figure 1).

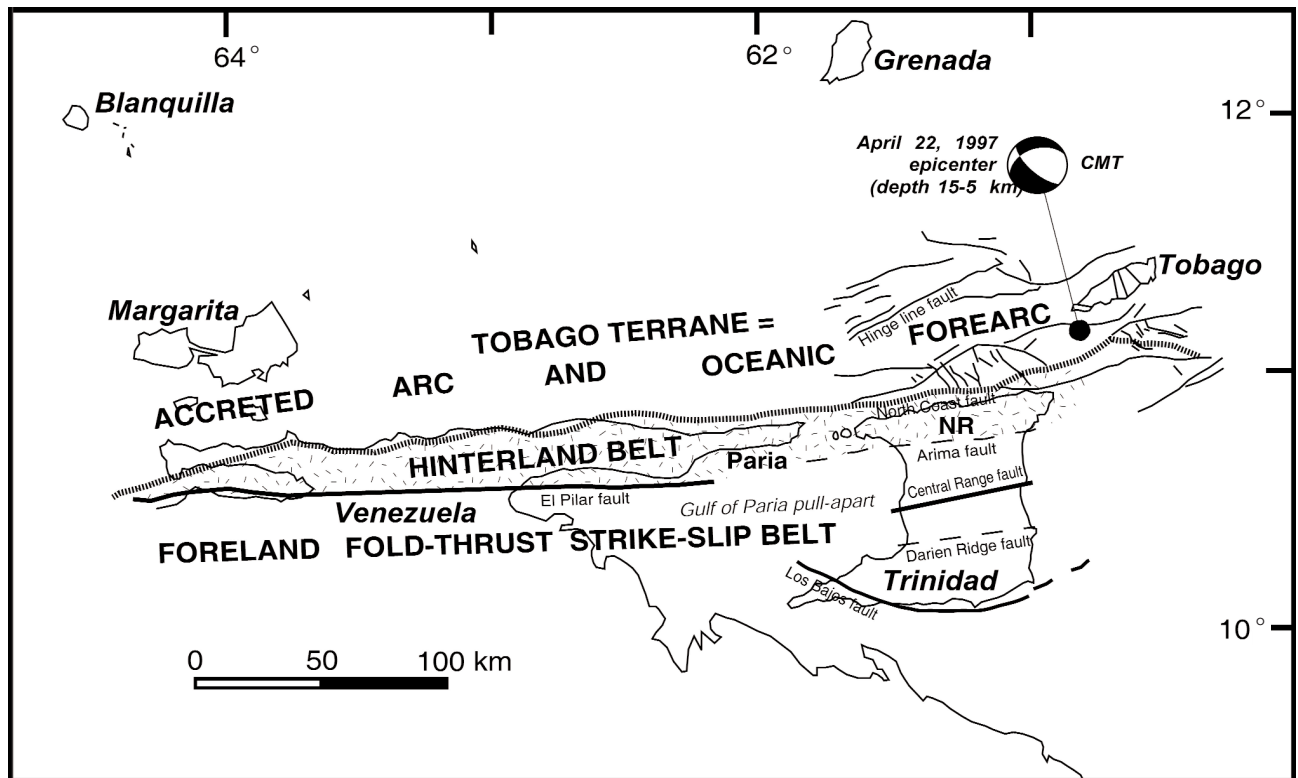
#### *Lithotectonic belts in the plate boundary*

As background for discussing the neotectonics in Trinidad and Tobago, it is necessary to first discuss the geology in the plate boundary zone. The three major ~east-west-trending lithotectonic belts are exposed, and one additional lithotectonic unit is inferred in the subsurface. The belts discussed are: 1) a foreland fold-thrust and strike-slip belt, 2) a hinterland metamorphic belt, 3) the Tobago terrane, and 4) a subsurface mantle shear zone (Figure 2).

#### *Fold-thrust and strike-slip belt*

Deformed upper crustal rocks of the fold-thrust and strike-slip belt are exposed in the Serrania del Interior in Venezuela, as well as in central and southern Trinidad. Exposures in Trinidad include the active Central Range transpressional belt (see below), the central Trinidad Nariva shale belt, the Southern Range, and the Southern Basin (Figures 2,





**Figure 2.** Principal active and fossil faults, and tectonic and geologic elements around Trinidad and Tobago. Principal active strike-slip faults in eastern Caribbean-South American plate boundary zone are: El Pilar, Central Range, Los Bajos, and Darien Ridge (?) faults, and those in Gulf of Paria pull-apart – FLINCH et al. (2000) gives the full details there. Faults in Trinidad-Tobago offshore and onland Tobago are taken from SNOKE et al. (2001). Stippled pattern marks mountainous Northern Range-Paria coastal hinterland metamorphic belt. The accreted arc-oceanic forearc Tobago terrane – continental South America boundary, after SPEED and SMITH-HOROWITZ (1998) and SNOKE et al. (2001), was reactivated during the April 22, 1997 Tobago earthquake, for which CMT focal mechanism, and CMT and NEIC epicenter and focal depth range are shown

3). The fold-thrust and strike-slip belt includes strongly shortened and sheared (i.e., transpressed and wrenched) Mesozoic north-facing South American passive margin deposits, and Cenozoic foreland basin deposits. Structures include northeast-trending, upright folds and thrusts that displace these deposits southeastward over continental South America (e.g. MASCLE et al. 1979; KUGLER 1961). The contractile structures are either truncated by or merge into the active dextral strike-slip faults.

Based on the stratigraphic ages available from the deformed foreland basin deposits (SPEED 1985; PINDELL et al. 1998; ALGAR and PINDELL 1993), deformation in this belt is entirely of Neogene age, and the locus of transpression related to oblique collision gets younger from east to west. Rocks as young as Pleistocene are folded and faulted in the Southern Basin of Trinidad (KUGLER 1961). Active folding and thrusting occurs along the N68°E active Central Range strike-slip fault in Trinidad, which is highly oblique to the current plate motion, and active strike-slip faulting probably also occurs in southern Trinidad (i.e., on the Los Bajos fault) and along the south coast (see below) (Figures 1, 2).

### *Hinterland belt*

The rocks that lie north of the fold-thrust and strike-slip belt make up the internal or hinterland part of the Caribbean-South American orogen, which is expressed topographically as a linear belt of coastal mountain ranges. AVÉ LALLEMANT (1997) synthesized the geologic history of this belt. In the Araya Peninsula, in central Venezuela, oceanic and subduction-related terranes containing high-pressure mineral assemblages are exposed. Greenschist- and subgreenschist-grade lateral equivalents of the Mesozoic South America passive margin deposits discussed above are present in the Paria Peninsula in eastern Venezuela and in the Northern Range of Trinidad (FREY et al. 1988, ALGAR and PINDELL 1993, WEBER et al. 2001b) (Fig. 2). FOLAND et al. (1992) and FOLAND and SPEED (1992) reported  $^{40}\text{Ar}/^{39}\text{Ar}$  age spectra from metamorphic white mica from the Northern Range schists as having Neogene (~25 Ma) ages; these pre-date fission-track cooling ages (ALGAR 1993; ALGAR et al. 1998; WEBER et al. 2001b) and probably date the age of fabric development. Zircon fission-track data indicate that the rocks in the western and central Northern Range cooled through ~230–330 °C at ~12 Ma (ALGAR 1993; ALGAR et al. 1998; WEBER et al. 2001b). Zircon fission tracks are unreset in the eastern Northern Range; a single reset apatite fission

track age there indicates that these rocks cooled through  $\sim 110^\circ\text{C}$  at 22 Ma (WEBER et al. 2001b). CRUZ et al. (in review) presents new fission-track ages from correlative hinterland rocks in the Paria Peninsula, Venezuela.

Clearly the age of the hinterland metamorphic fabrics, as well as that of the exhumation of these rocks, is related to pre-mid-Miocene transpression (e.g., TEYSSIER et al. 2002). However, the high coastal topography, and young fission-track ages in Paria (apatite fission-track ages:  $= 5.2 \pm 1.6$  Ma; zircon fission-track ages:  $= 4.7 \pm 1.8$  Ma; CRUZ et al. in review) suggest that uplift and erosional exhumation, probably driven by isostasy acting on a deep crustal root that developed during pre-mid-Miocene transpression, continued until recently.

### *Tobago terrane*

The Cretaceous plutonic, volcanic, and metamorphic oceanic arc-forearc rocks exposed in Tobago are distinctly different from the South American passive margin deposits and foreland basin deposits exposed in Trinidad (see e.g., SNOKE et al. 2001). SPEED (1985) was the first to reconcile this difference in the context of the terrane concept; he also mapped the broader extent of the oceanic arc-forearc rocks and called them the Tobago terrane after their exposures in Tobago (Figure 2). The geologic map, cross-sections, and report of SNOKE et al. (2001) provide a summary of Tobago's geologic features and a discussion of the history of ideas related to their development. There is general agreement that Tobago's oceanic arc-forearc lithosphere was obducted over or wedged into the hinterland metasedimentary rocks during pre-mid-Miocene oblique convergence (RUSSO and SPEED 1990; SPEED and HOROWITZ-SMITH 1988; SNOKE et al. 2001). This obduction event has had significant, long-lasting effects. The Tobago terrane-South American continental boundary has recently been reactivated and inverted, and is an active dextral-normal fault along which sinking of the dense, gravitationally unstable Tobago oceanic arc-forearc lithosphere is occurring (Figure 2). A long chain of geologically and geomorphically similar, accreted and sunken oceanic arc-forearc islands continues along strike from Tobago to the west (e.g. Aruba, Bonaire, Curacao), reflecting tectonic sinking across a much broader region (Figure 1).

### *Mantle shear zone*

A shear-wave splitting experiment revealed that a large ( $\sim 2$  second delay time) seismic anisotropy exists in the mantle beneath Trinidad and northeastern Venezuela (RUSSO et al. 1997). This highly anisotropic mantle has a fast polarization direction that is east-west, parallel to the trend of lithotectonic belts and major strike-slip faults in the plate boundary. RUSSO et al. (1997) interpret these data to indicate the presence of a broad east-west trending mantle shear zone underlying much of the plate boundary zone. The inferred shear zone probably has a minimum width of about 100 km, and the best estimate of its vertical dimension is  $\sim 200$  km, suggesting that it extends beyond the base of the lithosphere into the asthenosphere (RUSSO et al. 1997).

## **Trinidad and Tobago's neotectonics**

### *Horizontal neotectonics*

#### *Far-field Geodesy*

The plate-wide GPS study of WEBER et al. (2001a) first provided us with some important new geodetic results to address some of the open local neotectonic questions in the plate boundary. Along the El Pilar fault, the principal active transform fault in central Venezuela, our GPS-derived Caribbean-South American pole of rotation predicts  $090^\circ \pm 3^\circ$  directed motion, which is parallel to the strike of this seismically active fault (Figure 1). El Pilar slip vectors from large earthquakes support this  $090^\circ \pm 3^\circ$  prediction (see WEBER et al. 2001a, Figure 2). Unfortunately, none of the possible active faults in Trinidad are marked by current seismic activity. Regional geologic syntheses (ALGAR and PINDELL 1993; FLINCH et al. 2000; BABB and MANN 2000), suggested that the El Pilar Fault might right-step into central Trinidad, and the step could form a pull-apart basin in the Gulf of Paria (Figures 1, 2). The WEBER et al. (2001a) GPS-derived Caribbean-South American pole of rotation predicts motion toward  $085^\circ \pm 3^\circ$  in central Trinidad, which is highly oblique to the aseismic,  $068^\circ$ -trending Central Range Fault, the active strike-slip fault there (see below).

#### *Near-field geodesy*

In Trinidad, we were first able to measure near-field motions by combining high-quality angle and baseline observations from triangulation data collected  $\sim$ a century ago (1901–1903) by the British Ordnance Survey with 1994–1995 GPS data that we collected at 27 common sites (WEBER et al. 1999; SALEH et al. 2004; SALEH et al. in preparation). This work gave the first geodetic indication that a significant portion of the total 20 mm/yr Caribbean-South American of dextral

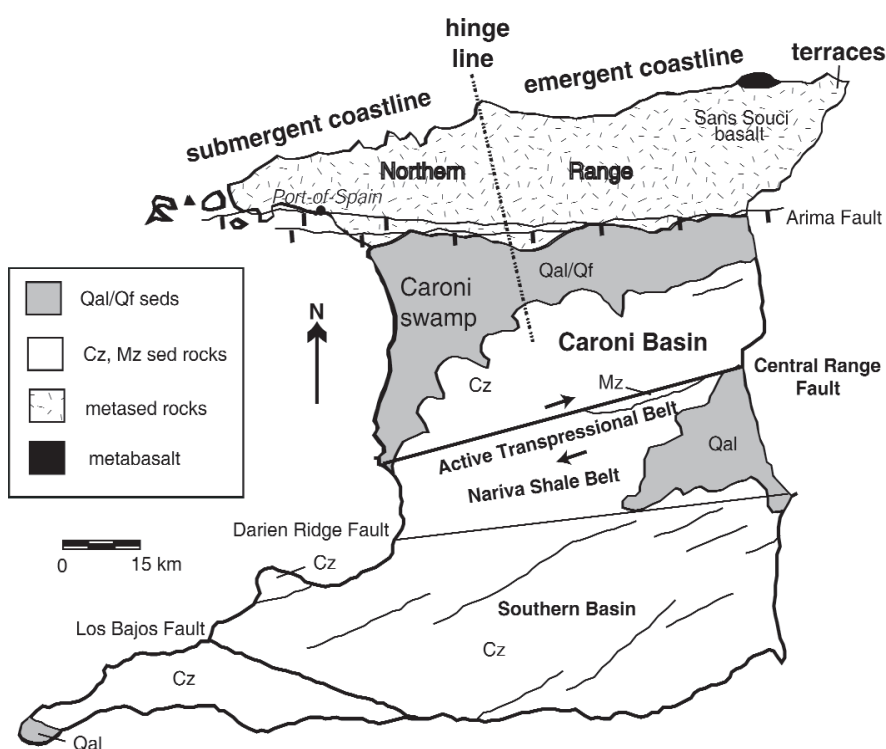
shear is taken up on-land in Trinidad, and demonstrated that over the past century geodetic strain accumulated mainly on the Central Range Fault, a previously unrecognized active fault. These geodetic data are robust and modeling them clearly shows that the Central Range Fault is the locus of maximum shearing in Trinidad, but it is more difficult to resolve the level of motion on the secondary faults using these data. Many earlier workers simply assumed that Venezuela's El Pilar Fault continues on-strike to the east into Trinidad, and that most plate motion was taken up there.

When analyzed using elastic dislocation models (SAVAGE and BURFORD 1970), the triangulation-to-GPS data gave a  $12 \pm 3$  mm/yr Central Range fault-parallel slip-rate. This value compares well with the  $14 \pm 3$  mm/yr of total dextral eastward shear we derived from repeat (1994–1999) GPS measurements at two sites spanning Trinidad (see WEBER et al. 2001a for complete details). In addition, Balkaransingh (1999) compared 96 1901–1903 triangulation-derived positions to 1963 triangulation-trilateration positions. Although her 1901–1963 displacements estimated are noisier than our 1901–1995 displacements, they also gave a comparable  $13 \pm 3$  mm/yr (dextral), best-fit, fault-parallel, elastic Central Range Fault slip rate.

### Central Range Fault geomorphology and palaeoseismology

Using aerial photographs and a high-resolution digital elevation model, we next mapped and studied the major topographic lineament coinciding with the high-strain Central Range Fault zone first identified geodetically. The N68°E-trending Central Range Fault lineament shows many features diagnostic of an active strike-slip fault, including, most importantly, dextrally offset streams (PRENTICE et al. 2001; CROSBY et al. 2001, 2003). The Central Range Fault is highly oblique to Caribbean-South American plate motion, which is directed  $N86^\circ \pm 2^\circ E$  in central Trinidad (WEBER et al. 2001a), and we infer that transpression is the style of deformation along it (Figure 3). The active tectonics of the Central Range transpressional system is quite different from that of the El Pilar transform fault in Venezuela, where, using plate-wide GPS data, we inferred that pure wrenching is occurring (WEBER et al. 2001a). Our calculated 1901–1995 Central Range Fault fault-normal shortening rate is  $1.2 \pm 1.2$  mm/yr, i.e., we cannot yet resolve this component of slip geodetically. However, longer time-scale geomorphic and geologic features, e.g. the folds and thrusts KUGLER (1961) mapped in the Central Range, and the hilly Central Range topography that is being held up despite the highly erodable, mudstone-rich section exposed there, suggest the accumulation of a few mm/yr of fault-normal shortening. Topography is probably growing in response to crustal thickening and incipient root development. In summary, the Central Range transpressional belt contains the youngest, most actively developing mountains in the Trinidad–Tobago segment of the plate boundary.

Locked faults store motion and then periodically release it during large earthquakes, whereas creeping faults do not. An extremely important open question raised by our near-field geodetic research, with significant implications for the seismic risk of Trinidad, is whether the Central Range Fault is locked or creeping. Elastic modeling the 1901–1995 geodetic data does not help us answer this question. But, a number of additional lines of evidence suggest that creep may be unlikely. First, creeping strike-slip faults are, in general, rare. A characteristic feature of the few known creeping faults in California (i.e. the Hayward and Calaveras faults; e.g. LIENKAEMPER et al. 2001) is that they have very frequent small earthquakes and are thus easy to identify seismically. According to maps made using data from the local Trinidad seismic network, which has been operating since 1953 (e.g. LATCHMAN 1997: Figures 5, 7), there is no cluster-



**Figure 3.** Geologic map of Trinidad, after KUGLER (1961) showing features discussed in text. Q = Quaternary; Cz = Cenozoic; Mz = Mesozoic; Note: Qf (= Quaternary fan deposits) are buried in the west by Qal (= Quaternary alluvium) of the Caroni swamp, and in the east are exposed and highly dissected by modern streams

ing of small magnitude events on the Central Range Fault. We do not know the performance characteristics of this network, so it is possible that microseismicity related to creep could go undetected, or that a clustering pattern could result from better event locations. However, maps made from the local Venezuelan (FUNVISIS) catalogue, with events dating back to 1910, also show no clustering of earthquakes on the Central Range Fault. In addition, after several days of direct searching in the field for offset cultural features (e.g. offset roads, dam that etc.) mark might creep along our mapped Central Range fault trace, I found none. Finally, Trinidad's long historic earthquake record is complete to ~1800 (ROBSON 1964), and although the possibility of several historic earthquakes along the Central Range Fault can not be ruled out, they are not unambiguously required by this record. Thus, we suspect that the Central Range Fault may be locked rather than creeping, and could constitute a major seismic risk for Trinidad. Far-field motion of 12 mm/yr, having possibly accumulated over the past several centuries, suggests that several metres or more of motion could be stored up along this fault.

We next began a programme of palaeoseismology research aimed at establishing whether or not any large prehistoric ("fossil") earthquakes have occurred on the Central Range Fault. We reasoned that if we could find evidence for fossil earthquakes "frozen" in the Holocene sediments that were deposited along the mapped fault trace, then the fault was locked and stored and released significant elastic motion in the recent past, and could be locked today. Our work to-date demonstrates clearly that the Central Range Fault has been active in the mid-Holocene or later (< 5,000 yrs BP) (PRENTICE et al. 2001; CROSBY et al. 2001, 2003), which qualifies it as an active fault by palaeoseismological standards. As of this writing we are still waiting for  $^{14}\text{C}$  ages from the past field season to more fully answer the locked versus creeping question.

Additional geomorphic and palaeoseismic data in Trinidad suggest that a few mm/yr of dextral-slip is taken up across the Los Bajos Fault (CROSBY et al. 2001; 2003), which links to faults along the south coast and in the southern offshore, i.e. in the Columbus channel (Figure 2). We have not yet been able to directly measure Los Bajos slip rates geodetically. Nonetheless, subtracting our 12 mm/yr Central Range Fault slip rate from the 20 mm/yr plate motion total, derived in part using GPS data from a site (#0069) in northern Trinidad that, within uncertainty, moves at the full (20 mm/yr) Caribbean–South American plate motion rate (WEBER et al. 2001a), suggests that some of the missing ~6 mm/yr of motion may be taken up in southern Trinidad.

### *Vertical neotectonics*

#### Sinking into the Gulf of Paria

Vertical neotectonic motions are difficult to measure geodetically. But, geology and geomorphology can potentially capture a longer time-integrated record of such signals. I next describe features in northern Trinidad that I believe record to a geomorphic signal of long-term subsidence into the Gulf of Paria pull-apart basin. In the eastern Northern Range the entire coastline is sharp and emergent with young (<1 Ma) subaerial exposures of marine terraces at elevations of ~15 metres above sea level (Figure 3). After crossing a "hinge-line" in the west-central Northern Range, the coastline becomes highly scalloped, drowned, and submergent (Figure 3).

The morphology of the young alluvial fan system along the southern margin of the Northern Range records the same vertical signal (Figure 3). In eastern Trinidad (e.g. around the village of Valencia), the range-front fans are elevated relative to local basse-level (i.e., they have probably been uplifted) and highly dissected by modern streams. In western Trinidad, the correlative range-front fan deposits are buried in the subsurface of the Caroni Basin beneath the Caroni swamp; we know that the fan gravels are present there because they have been extensively drilled an important source of ground water (e.g. in the villages of el Socorro, Curepe).

This geomorphic pattern is clear and consistent and I believe indicates that north-western Trinidad is sinking into the Gulf of Paria pull-apart basin via a see-saw tectonic mechanism, with a fulcrum at the "hinge line" discussed above (Figure 3). In addition, I observed a mirror image of the Trinidad pattern during a visit to the Paria Peninsula in Venezuela in January, 2004. Thus, the regional geomorphology "feels" and records active deformation in the Gulf of Paria pull-apart basin. Subsidence into this basin may have been a major cause of long-term vertical tectonic motions in Trinidad over the past few million years, the approximate duration of the pull-apart. The east-west oriented cross-sections of FLINCH et al. (2000) show that about 50 km of total stretching has taken up across the Gulf of Paria. Divided by the full plate-motion-rate we observe today (20 mm/yr) gives a probable age for the Gulf of Paria pull-apart of a few m.y. The current stretched width of the Gulf of Paria is about 125 km. The stretching factor  $\alpha = w_b/w_o$  ( $w_b$  = new, deformed, stretched length of crust,  $w_o$  = original length of crust) is therefore ~1.6 for the Gulf of Paria.

Regional subsidence into pull-aparts occurs in response to stretching and thinning of the lithosphere. Simple theory based on conservation of mass and isostasy makes quantitative predictions regarding the magnitude of subsidence in the center of such a basin (TURCOTTE and SCHUBERT 2002: p. 75–76), ~7.5 km of total subsidence, tapering off toward the basin edges.



Crude estimates can be made for possible a long-term subsidence rate and total subsidence in Trinidad from the currently available geomorphic data. The subaerial marine terraces in north-eastern Trinidad formed at wave base that are now at 15 metres above sea level do not have well-constrained ages (Figure 3). They consist of young, unconsolidated quartz-rich gravels overlying tilted Northern Range metamorphic hinterland rocks that KUGLER (1961) mapped as Pleistocene in age, so a possible age range between ~1 Ma and 10,000 ka seem reasonable. Doing the division, one gets a reasonable range of uplift rates, 0.02–1.5 mm/yr. Assuming that sinking in the west is symmetric with uplift in the east about the “hinge-line” discussed above, long-term subsidence rates about 30–40 km west of the “hinge-line” in at the western end of the Northern Range must then range between about 0.02–1.5 mm/yr. This translates into a possible range for total subsidence of 0.04–3 km over the past few m.y. A value of about 1 km seems most in line with the observations and the 7.5 km of total basin-center subsidence derived via theoretical considerations above.

### Sinking of the Tobago terrane

On April 22, 1997 the largest (M 6.6) recorded earthquake in the Trinidad–Tobago segment of the Caribbean–South American plate boundary zone occurred near the boundary between the obducted oceanic arc–forearc of the Tobago terrane and continental South America (Figure 2). Focal mechanisms, locations, and epicentral depths are available for this event from the NEIC (U.S.G.S. National Earthquake Information Center) and the CMT (Harvard Centroid Moment Tensor) catalogues (Figure 2). During the event a WSW striking (250° azimuth), shallowly dipping, dextral-normal fault ruptured a shallow (= 5–15 km deep) fault patch ~10 km south of Tobago. Given the current ~E–W dextral shearing in the plate boundary, e.g. shown by the recent GPS data, the 1997 Tobago event is anomalous; it did not occur on an ~E–W striking, subvertical, dextral strike-slip fault, nor on a subsidiary fault related to dextral shearing. The event is of further interest because it ruptured a fault plane with an extremely low (28°) primary dip angle that accommodated a relatively large component of normal slip. I next discuss work in preparation related to the tectonic significance of this earthquake (WEBER et al. in preparation).

We use GPS data before and after the April, 1997 earthquake at two sites in Tobago to determine the coseismic offsets that occurred during this event; we then use the offsets in elastic dislocation models and invert for best-fit fault plane and fault slip parameters (WEBER et al. in preparation). The GPS-derived offsets are largest in southern Tobago, closest to the earthquake epicenter, and taper off to the north, and both GPS sites moved mainly to the north during this WSW-striking dextral-normal faulting event.

To model the fault plane and fault slip parameters, we used an elastic dislocation model of POLLITZ (1997), and the aftershock distribution (LATCHMAN et al. in preparation) to fix the fault dimensions (30 km along strike and 25 km down dip). We also fix the fault strike to that from the CMT solution, 250°, which is in good agreement with the aftershock pattern (LATCHMAN et al. in preparation). We obtained a best-fit model with a fault dip of 28° NW, midway between the reported CMT value of 41° NW and NEIC value of 20° NW, slip rake of –142°, and a 100 cm average fault slip. We get a very close model match to our GPS-derived and empirically calculated displacements. A lower, 12° NW, fault dip angle was estimated earlier by SPEED and HOROWITZ-SMITH (1998) for the terrane boundary from an uncorrected seismic (time) section. Our 28° NW dip estimate is robust and is probably more accurate than either the earlier geologic or the seismologic estimates. Our 100 cm of slip resolves into 79 cm of normal dip-slip and 62 cm of dextral strike-slip.

The 250° strike and normal dip-slip, and the relatively large (1.3:1.0) normal dip-slip to dextral strike-slip ratio, for the April 22, 1997 event cannot be directly related to ~E–W dextral shear straining in the plate boundary. Recall that oceanic arc-forearc lithosphere of the Tobago terrane was obducted over or wedged into the metasedimentary rocks of the Northern Range along the Tobago terrane–South American continent boundary during pre-mid-Miocene oblique convergence (RUSSO and SPEED 1990; SPEED and HOROWITZ-SMITH 1989) (Figure 2). The 250° strike of the fault patch ruptured during the Tobago event matches that of the mapped terrane boundary (SPEED and HOROWITZ-SMITH 1998; SNOKE et al. 2001). Given our 28° NW dip estimate, the epicenter and fault plane for this event can be brought into coincidence with the subsurface projection of the mapped Tobago terrane – South American continent boundary if we assign the earthquake a ~5 km epicentral depth, which is within the permissible CMT and NEIC 5–15 km range (Figure 2).

We conclude that the April 22nd event reactivated and inverted the Tobago terrane boundary, which initially had a low dip angle and opposite (i.e., thrust) dip-slip sense of motion. We interpret that gravitational forces drove the large component of normal dip-slip, which resulted in a rapid readjustment of the unstably arranged dense oceanic arc-forearc over continental South America lithospheres. During the event, the southern, oceanic edge of the Tobago terrane largely “sunk” into a gravitationally more stable position. Similar earthquakes acting over the past few m.y. can explain the peculiar sunken geomorphic expression of the entire accreted oceanic arc-forearc terrane (i.e. the Aruba–Bonaire–Curacao island chain) in the plate boundary zone (Figure 1).

## Summary

Despite being set in an apparently simple dextral transform plate boundary, e.g. as shown by the plate-wide GPS results of WEBER et al. (2001a), the active tectonics in Trinidad and Tobago is quite complex. The principal active strike-slip fault in Trinidad, the Central Range Fault, is highly oblique to plate motion, and owing to the obliquity, a transpressional belt is developing in central Trinidad. Based on geometric estimates for the age of the Gulf of Paria pull-apart, which links the Central Range Fault to the active El Pilar transform fault in Venezuela, the inferred age of the transpressional belt is just a ~few Ma. Some contractile structures in the foreland fold-thrust strike-slip belt are probably fossil structures related to the pre-mid-Miocene oblique convergent phase of plate motion. An inferred fossil crustal root developed during earlier oblique convergence probably holds up the coastal mountains of the hinterland belt, but, the geomorphology of northwestern Trinidad also records a km or so of total vertical subsidence into the Gulf of Paria pull-apart basin, indicating that significant parts of the coastal mountain belt are actively sinking. The dense oceanic arc-forearc terrane exposed in Tobago was obducted over South America during oblique convergence; the Tobago terrane boundary has been reactivated and inverted, and today Tobago sinks along it. The techniques discussed here that exploit neotectonic signals accumulated over long times may be applicable to studying the slower, but possibly significant, intraplate neotectonics of Hungary.

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## Crustal motions from space geodesy: a review from EPN, CEGRN, and HGRN data

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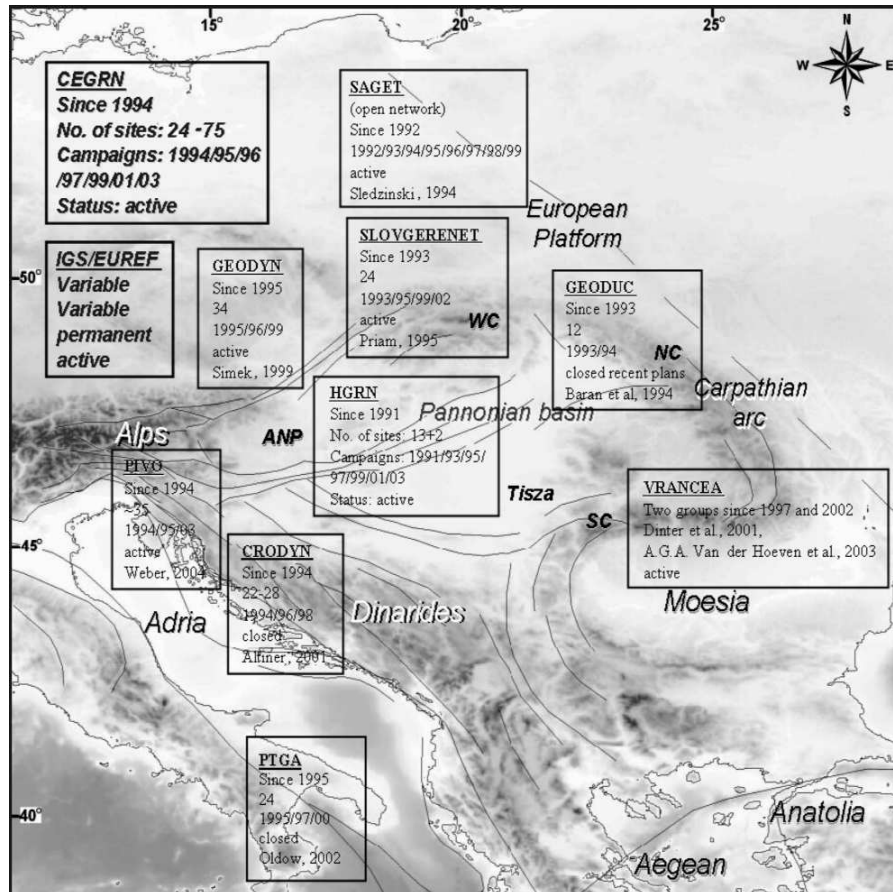
### Abstract

From the Adria microplate, the main engine of the Central European tectonics, in the southwest to the rigid, aseismic European Platform in the north-east one regional and several smaller scale geodynamic networks have been measured since the early 1990s using the Global Positioning System (GPS) space geodetic technique. A continental permanently observing GPS network has also been providing data since the mid 1990s, and there are several stations of the global network in the region as well. After almost a decade, some of these networks have already produced intraplate velocity solutions. In general, the site density of these networks is in inverse proportion to the covered area, thus an adequately dense velocity data at a regional level, required by the complex crustal structure, and highly interconnected tectonic processes, can only be achieved by the integration of the results of these networks. Our goal is an integrated dense intraplate crustal velocity field for the better understanding the present tectonics of the region.

The first national GPS network in Central Europe was the Austrian AGREF (Figure 1); however, it has not yet been completely re-measured again. The Hungarian GPS geodynamic network, the HGRN was established between 1990 and 1991 with 13 sites on outcropping solid bedrock and measured in every second year with 3 times 24-hour campaigns since 1991 (FEJES et al. 1993). The Polish SAGET is an open network since 1992; it includes the data of all simultaneously measured European GPS sites (SLEDZINSKI 1991). The SLOVGERENET (PRIAM 1995) and the Ukrainian GEODUC (BARAN et al. 1994) were established in 1993 but while the SLOVGERENET is still active, the GEODUC has not been resurveyed since 1994. The CRODYN network was operated by a German–Croatian cooperation investigating the eastern boundary of the Adria microplate but after three observational campaigns the project ended (ALTINER 2001). The Bohemian Massif is represented by the Czech GEODYN (SIMEK 1999) network since 1995. There are a couple of more or less independent GPS networks covering the Vrancea seismic zone and its surrounding areas since 1997 and 2002 (DINTER et al. 2001; A.G.A. VAN DER HOEVEN et al. 2003). In the Sudeten (SCHENK et al. 2003) and the Tatra Mountains (CZARNECKI et al. 2003) two local scale networks has been operating and already provided results. At the northern tip of Adria having a special importance, the PIVO experiment (WEBER et al. 2004) and the re-measurement of the Slovene GPS network after 1994–1995 are being carried out. The regional scale Central European GPS Geodynamic Reference Network (CEGRN) is the product of the joint effort of all Central European countries (PESEC 2003; FEJES 1993). The network covers the whole region with homogeneous site density providing high quality GPS data (BECKER et al. 2001; GRENERCZY 2002; HEFTY 2001). The CEGRN was established in 1994 and are measured seven times so far between 1994 and 2003 with 5 times 24-hour simultaneous observations. There are also several IGS (MUELLER and BEUTLER 1992) and European Permanent GPS Network EPN/EUREF (BRUYNINX et al. 1996) permanent GPS sites although their distribution is not appropriate and their number is small, they have much more reliable continuous data sets that enable easier detection of coordinate outliers, offsets and periodicities.

In order to obtain a dense intraplate velocity map for Central Europe, combination of velocity solutions of as small as national scale networks would be necessary. To integrate these to a combined map a common reference is required. All solutions therefore are relative to the European Platform and are in the International Terrestrial Reference Frame 2000. From GPS station velocities a continuous velocity field was calculated by interpolating the data on a grid covering the region of Central Europe by different interpolation methods to reveal the general, large-scale crustal motion pattern from the plate boundary to the Eurasian plate interior.





**Figure 1.** Summary of the GPS geodynamic networks from Adria to the Eurasian plate interior after GRENERCZY and KENYERES (2004). Inserted tables show the first measurements, number of sites, campaigns, and current status. ANP: Alpine – North Pannonian unit, WC, NC, SC; Western, Northern, and Southern Carpathians

The obtained GPS intraplate velocity field (Figure 2) constrains the motion of the Adria microplate and reveals that at the southern part of the microplate, GPS sites have 3.5–5 mm/yr north–north-eastward oriented velocity that decreases to 3–4 mm/yr in the central part. In the northern part it further decreases to 2.5–3.5 mm/yr northward motions without major difference between the eastern and western shores of the Adria. The data show remarkable uniform counter-clockwise block rotation for the Adria, but still do not allow proving that it is one rigid block since the sites are in or very close to actively deforming zones affecting their velocities thus they would not represent one rigid Adria block. At the northern tip of the Adria block the velocities suggest a northward intruding and east-west contracting Adria wedge with an average 2.5 mm/yr northward, convergent component with the Alps. According to our data the Alps is presently absorbing all of this 2.5 mm/yr crustal movement causing 30 ppb/yr contraction rate over the Southern Alps since no northward movement can be traced further to the north. At the northeastern boundary of the Adria block, the Dinarides also takes up from 3 to 4 mm/yr motion that is about 1 mm/yr less than the Adria convergence rate. The entire Pannonian Basin is also in a compressional stage and the observed present contraction rate over the entire basin is around 4 ppb/yr although the absorbed crustal movement is around 1–1.5 mm/yr. The contraction is not uniform, there is a significant strain partitioning since much of the deformation is mainly taking place in the western and central part, meanwhile the eastern part and the Northern Carpathians may only absorb a few tens of a millimetre per year crustal movement. The data also proves that the Alpine – North Pannonian unit is presently squeezed out from between the stable Bohemian Massif and the northward intruding Adria block with 1–1.5 mm/yr velocity relative to the stable and rigid northern domains towards the Pannonian Basin that provides the weakest boundary condition. The majority of this lateral extrusional movement is taking place along a southern dextral transpressional fault system including the Fella–Sava line and the Periadriatic lineament and a sinistral fault system at its northern boundary including the Mur–Mürz line. In the central section of the Pannonian Basin, this eastward movement seems to slow down and stop, and the kinetic energy of this crustal segment is absorbed probably causing the observed increased seismicity there. At present the northern domains like the Bohemian Massif, East European Precambrian Platform, and German–Polish Depression show no crustal motions above the 0.5 mm/yr level. Further to the north of the Carpathians and the Alps the tectonic domains can be considered to be part of the European Platform, part of the rigid Eurasian plate interior.

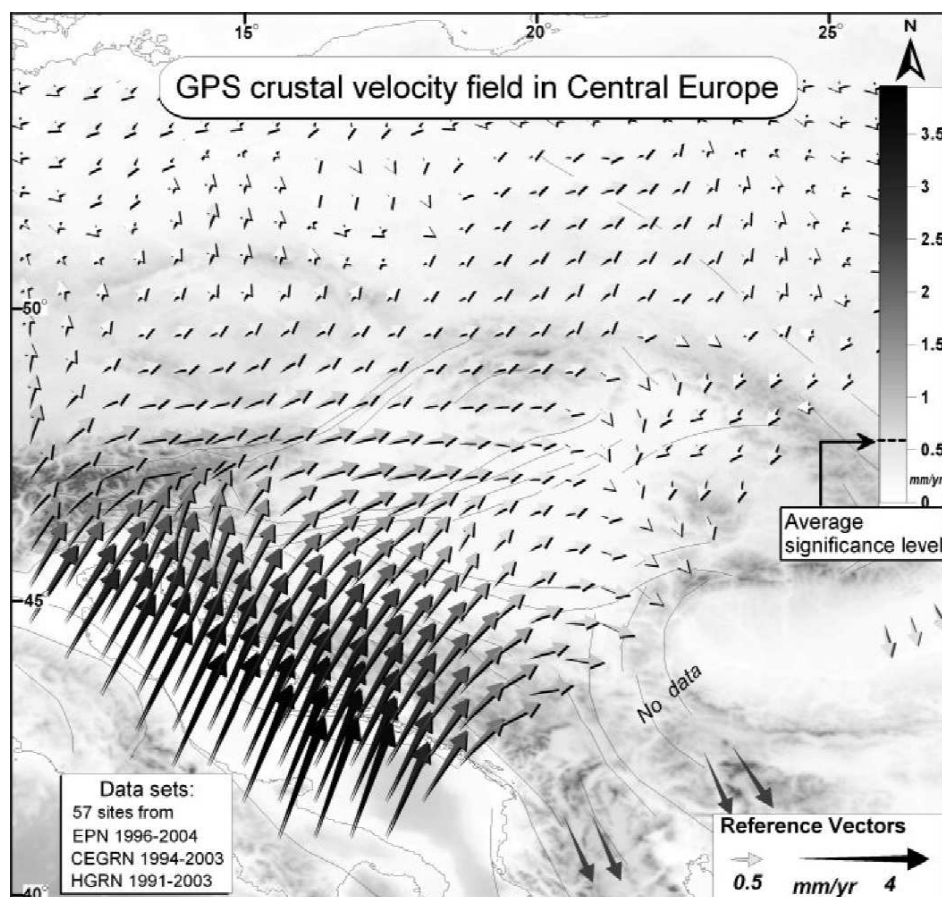


Figure 2. Interpolated GPS velocity field after GRENERCZY and KENYERES (2004). The velocities are in ITRF2000 referenced to stable Eurasia that is defined by sites situated on the European Platform. For geographic names see Figure 1

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## New data on neotectonic structures and morphotectonics of the western and central Pannonian Basin

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### Abstract

We carried out a complex neotectonic-geomorphologic research in several selected areas of the western and central Pannonian Basin, in order to understand the connection between upper crustal neotectonic deformation and landscape evolution. The methods incorporate surface geology, geomorphology, digital terrain models, and subsurface data sets, like boreholes and seismic reflection profiles. Observations suggest that most of the neotectonic structures are related to inverse reactivation of earlier faults formed mainly in the Miocene syn-rift and post-rift phases. The typical structures are folds, mainly blind reverse faults and transpressional strike-slip faults, although normal or oblique-normal faults can also be present. These structures exerted considerable control on landforms and drainage pattern while inducing surface upwarps and the development of drainage divide or river deflections. On the other hand, our analysis disproves major neotectonic origin of some landforms, particularly the fan-shaped 'meridional' valley system of the western Pannonian Basin. The major neotectonic strike-slip faults are trending east-northeast, and have dextral to sinistral kinematics in the south-western and central-eastern part of the studied area, respectively. This along-strike changing kinematics within the same zone is in agreement with fan-shaped stress trajectories, and recent motion of crustal blocks suggested by GPS data.

### Introduction, methodology

Neotectonic phase in the Pannonian Basin comprises deformation of the last 5–6 million years (TARI 1994; HORVÁTH 1995). The deformation was, in general, characterized by compressional or transpressional stress, similar to present-day conditions (BADA et al. 1998, 1999; GERNER et al. 1999). This last tectonic phase largely reactivated somewhat earlier structures, originated in the syn-rift (18–13 Ma) and post-rift (13–6 Ma) events of the basin evolution. The upbuilding of neotectonic intra-plate stresses induced a general surface uplift in the western Pannonian Basin, while inversion of fault slip on earlier normal faults enhanced it locally (RÓNAI 1974; HORVÁTH and CLOETINGH 1996; CLOETINGH et al. 1999).

Due to vertical motions, former lacustrine and fluvial sedimentation basically terminated and neotectonic deformation occurred on dry land. Thus understanding the link between crustal deformation and surface processes (landscape evolution) is essential in neotectonic analysis of the Pannonian Basin. Another consequence is that no or very limited syn-tectonic sedimentation occurred; Pliocene is generally missing and Quaternary is mainly represented by loess, fluvial or slope sediments having poor outcrop conditions.

Taking into account these features, we carried out an investigation of map-scale neotectonic structural elements observed both in the late Miocene post-rift sediments and/or on the surface. The applied methodology included the comparison of outcrop data, digital elevation models, topographic maps, and aerial photos from the surface and industrial seismic profiles, other geophysical data, and boreholes from the subsurface. The idea behind this comparative methodology is that surface neotectonic indices should have corresponding structures below the surface. In other words, we looked neotectonically deformed landscape from below the surface and tried to determine those sub-surface structures (faults and folds), which could be responsible for surface deformation. We focused on map-scale structures, which are within the resolution of applied subsurface geophysical methods and borehole data.



Our ‘precursor’ was STRAUZ (1942, p. 52) who concluded that “...structural geometry of southern Transdanubia, covered by young sediments, can not really be reconstructed on the basis of surface geology but using geophysical observations related to formations below the young surface layers”. We followed the successful method of HORVÁTH and RUMPLER (1984), TARI (1994), HORVÁTH (1995) and TÓTH et al. (1997) who demonstrated subsurface neotectonic structures using seismic reflection profiles. During our research, we intended to extend this line of promising and pioneering observation with a (1) broader surface and subsurface database, (2) detailed study of three selected areas, (3) describing and understanding of the origin of main landforms, and (4) deciphering the role of neotectonic deformation in landscape evolution.

The selected areas, the Zala, Vértes and Gödöllő Hills are located in the western and central part of the Pannonian Basin, close to the axis of the uplifted Transdanubian Range (TR), and have particularities in research methodology and neotectonic structures. In the Zala Hills (south-western Hungary) the surface effect of well-imaged neotectonic structures (PÁVAI VAJNA 1925; HORVÁTH and RUMPLER 1984) was the main focus of research. The change from post-rift to neotectonic phases was detected in the Vértes Hills in the central TR; this case study shows the combination of surface data and boreholes (CSILLAG et al. 2002). Extensive loess cover over post-rift suite and the existence of geomorphic indices for young deformation represent the challenges in the Gödöllő Hills (FODOR et al. 2001); the geomorphic indices were detected by subsurface data sets.

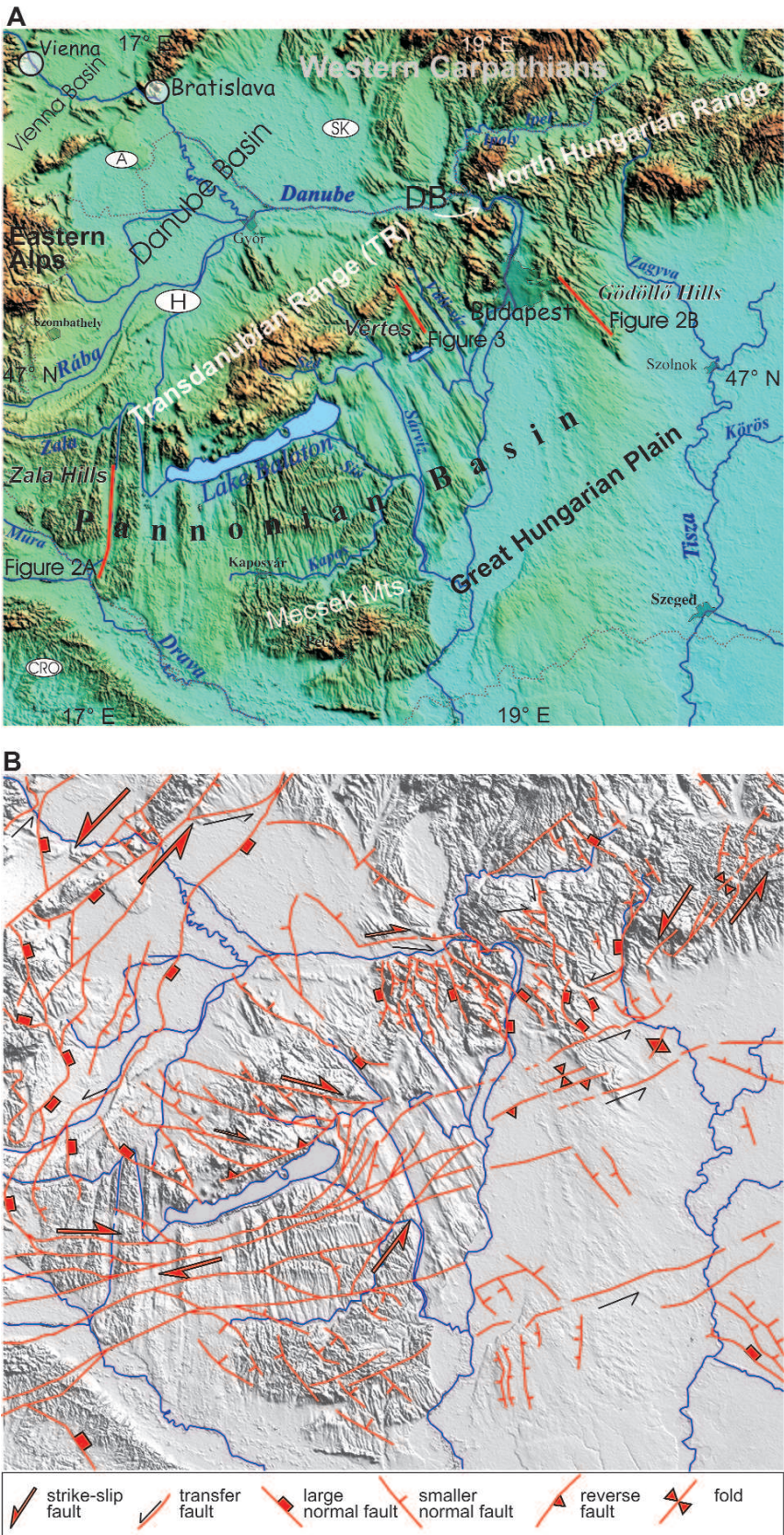
### **Fault reactivation and the Transdanubian valley system**

Probably the most characteristic landforms of the western Pannonian Basin (a region called Transdanubia), are the N–S to NW–SE trending ‘meridional valleys’, which were frequently supposed to be tectonically controlled (ÁDÁM et al. 1969, BREZSNYÁNSZKY and SÍKHEGYI 1987, GÁBRIS 1987, GERNER 1992; JÁMBOR et al. 1993; SÍKHEGYI 2002; see also the review of GERNER 1994). To check the role of structures to these landforms, we compared syn-rift fault pattern to the valley system (Figures 1A, B) because Miocene faults were likely be reactivated during the neotectonic phase. This comparison, and some seismic sections across the valleys suggest that the radial-linear ‘meridional valleys’ are generally not controlled by neotectonically reactivated Miocene faults with displacement over 20 metres. Small-displacement faults or joints could play a limited role in localisation of short valley segments. Such possible tectonic control was suggested, for example, by recent study of MAGYARI et al. (2004) south of the Lake Balaton. However, fracture-length–displacement scaling laws (WATTERSON and WALSH 1988) suggest that these small-displacement fractures can hardly control entire long valleys. Instead of tectonics, we reinforce earlier conclusions of LÓCZY (1913), CHOLNOKY (1918) and many others in favour for dominant aeolian (deflation) origin for the meridional valley system (FODOR et al. 2003), not denying the important role of fluvial incision (SOMOGYI 1961; ERDÉLYI 1962; PÉCSI 1986). Our numerous arguments include, among others, the wide distribution of wind-polished pebbles (JÁMBOR 1992, 2002), wind-polished rock surfaces (FODOR et al. 2003), landforms related to deflation (CHOLNOKY 1918, 1936), most of them found in wind channels.

### **Neotectonic folding in the Zala Hills**

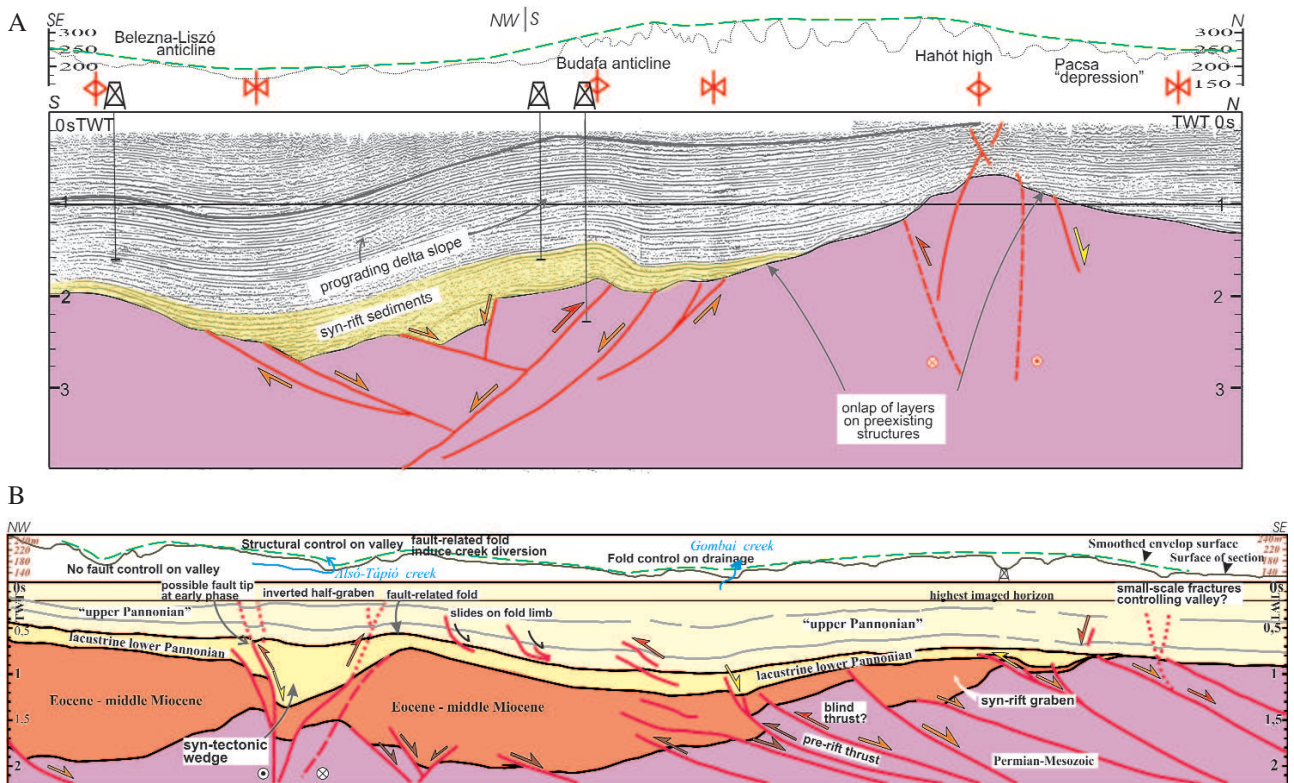
The Zala Hills of south-western Hungary have a more dissected topography than areas around its northern, eastern and southern surrounding. Geomorphic indices also suggest actively deforming landforms, like asymmetric valleys (KOVÁCS 1999; PINTER this volume). The long and wide meridional and smaller side valleys incise a thick post-rift Pannonian sequence. E–W trending folds were recognised long-time ago both in the Zala Hills and their eastern Slovenian neighbourhood (PÁVAI VAJNA 1925; HORVÁTH and RUMPLER 1984). We compared structures imaged by seismic reflection profiles to surface morphological elements (highs and lows within the intra-valley ridges) and drainage divides. Our observation is that anticline hinges quite closely correspond to E–W trending morphological highs and drainage divides, while synclinal hinges broadly match with topographic lows, often marked with major drainage elements (Figures 2a, 4). We also suggest that a folded surface, which envelopes the undulating topography may correspond to a former alluvial fan. This surface has locally preserved pebble lag with poorly constrained Pliocene or early Quaternary age (STRAUSZ 1949). In addition to recognition of the two major anticlines seen on Figure 2A (Hahót and Budafa folds), STRAUZ (1943) observed that the pebble lag is missing near the hinges of the folds; he suggested syn-sedimentary folding and non-deposition of gravel.

We suggest that landscape evolution of SW Hungary was primarily determined by upper crustal folding, due to reactivation of former syn-rift structures, mainly normal and few strike-slip zones. Folding started in the latest Miocene or in the earliest Pliocene after the cessation of post-rift fluvial sedimentation. Folded post-rift sediments were denudated before the formation of alluvial fan(s) of the Dráva and/or Rába rivers. Deformation continued later and folded the alluvial fan (palaeosurface) during the late Pliocene and/or Quaternary. Due to denudation by wind and rivers the once con-



**Figure 1.** A) Digital elevation model of the western and central part of the Pannonian Basin (Red lines mark the sections of Figure 2A, B and 3. Digital Elevation Model is after TIMAR et al. (2003) B) Syn-rift fault pattern (FODOR et al. 1999) superposed on digital elevation model of the same area. Note discrepancy between most of the faults and landforms, particularly the radial 'meridional valleys'.





**Figure 2.** Representative seismic reflection profiles in the Zala and Gödöllő Hills A) N-S seismic reflection profile made in the Zala Hills, Principális valley, showing major folds, which affected the whole post-rift sequence (partly after HORVÁTH and RUMPLER 1984; BADA et al. 2004). The topographical section 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 B) NW-SE trending profile showing the Tóalmás zone and the folded structure of the southern Gödöllő Hills (after FODOR et al. 2001)

tinuous folded gravel carpet was destroyed and a strongly dissected topography developed. However, the recent folding is still reflected in undulations of the smoothed envelope surface of the topography. Surface upwarplings above folds predicted the location of drainage divides on the inter-valley ridges and sometimes within main meridional valleys. Particularly, deflection of the Lower Válicka was induced by formation of the Budafa and Lovászi anticlines. This suggestion is in agreement with PINTER (this volume). However wind deflation and fluvial erosion in some meridional valleys kept up with gentle upwarping and erased the effect of folding on valley bottoms.

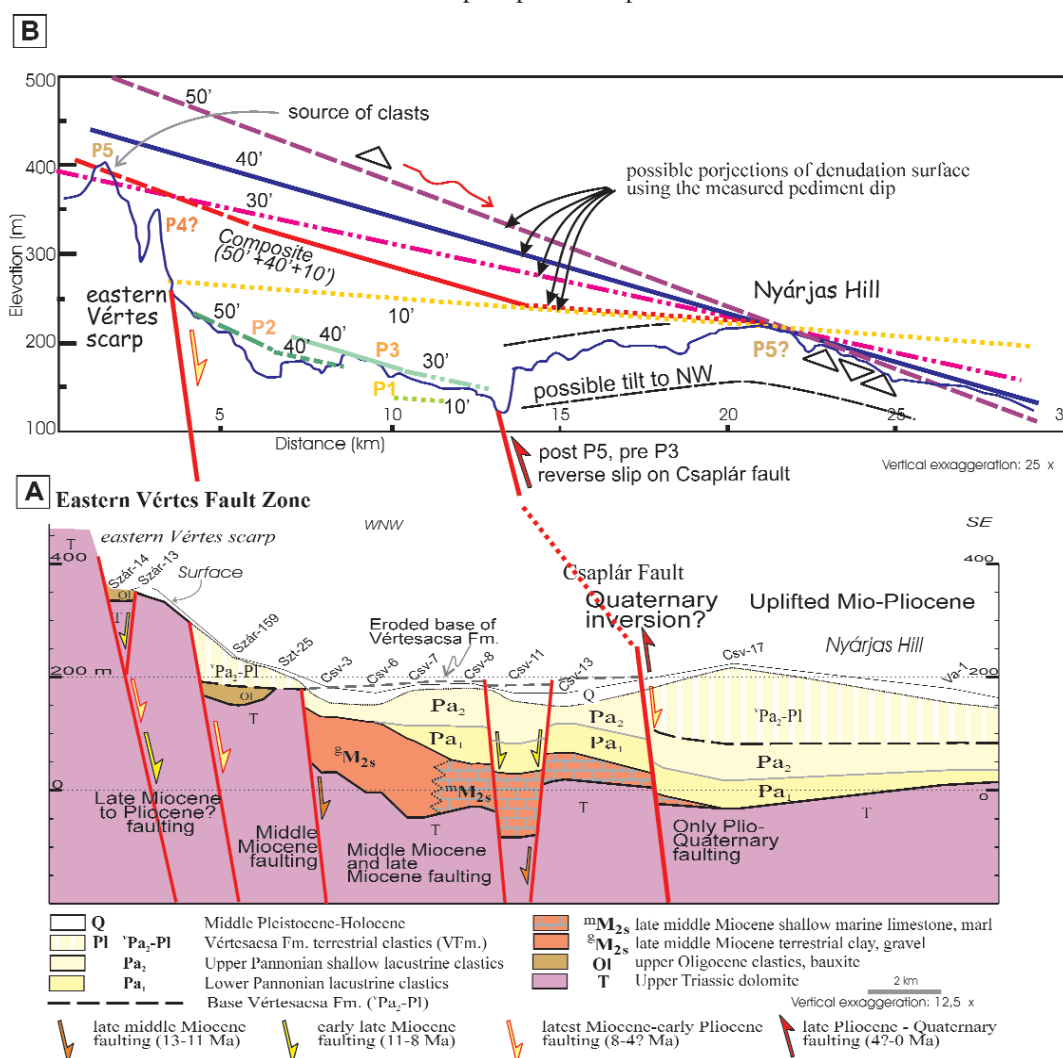
### Neotectonic transpression in the Gödöllő Hills

The Gödöllő Hills are composed of more-than-1 km thick post-rift sediments package of late Miocene to early Pliocene age and middle to late Pleistocene loess, sand and slope sediments (SZENTES 1943; UHRIN 2004). In the northern Gödöllő Hills, some NW-oriented syn-rift normal faults were slightly reactivated partly or completely after the deposition of the post-rift suite (the exact age of this slight faulting is not well constrained). Few of these faults were later inverted as reverse (or reverse-dextral) faults (FODOR et al. 2001, 2003). The reverse displacement was distributed to small fault branches and induced slight folding of the post-rift sequence.

The central Gödöllő Hills are cut by an ENE trending sinistral transpressive zone, the Tóalmás zone (Figures 2B, 4). Neotectonic transpression inverted early post-rift (early late Miocene) transtensional syn-sedimentary grabens, which themselves were superposed onto a Miocene (pre- to syn-rift) strike-slip zone (TARI et al. 1992; CSONTOS and NAGYMAROSY 1998). Slip reversal from normal-sinistral to reverse-sinistral was connected to or completely accommodated by folding in the upper post-rift suite. The sinistral strike-slip zone is connected to en echelon E-W trending folds of the southern Gödöllő Hills. A syncline hinge can be detected between Uri and Pánd, while an anticline is located more to the south, between Pánd and Ceglédbercel. The deformation pattern and kinematics are in agreement with recent stress field, characterized by NE-SW compression and NW-SE tension (TÓTH et al. 2002).

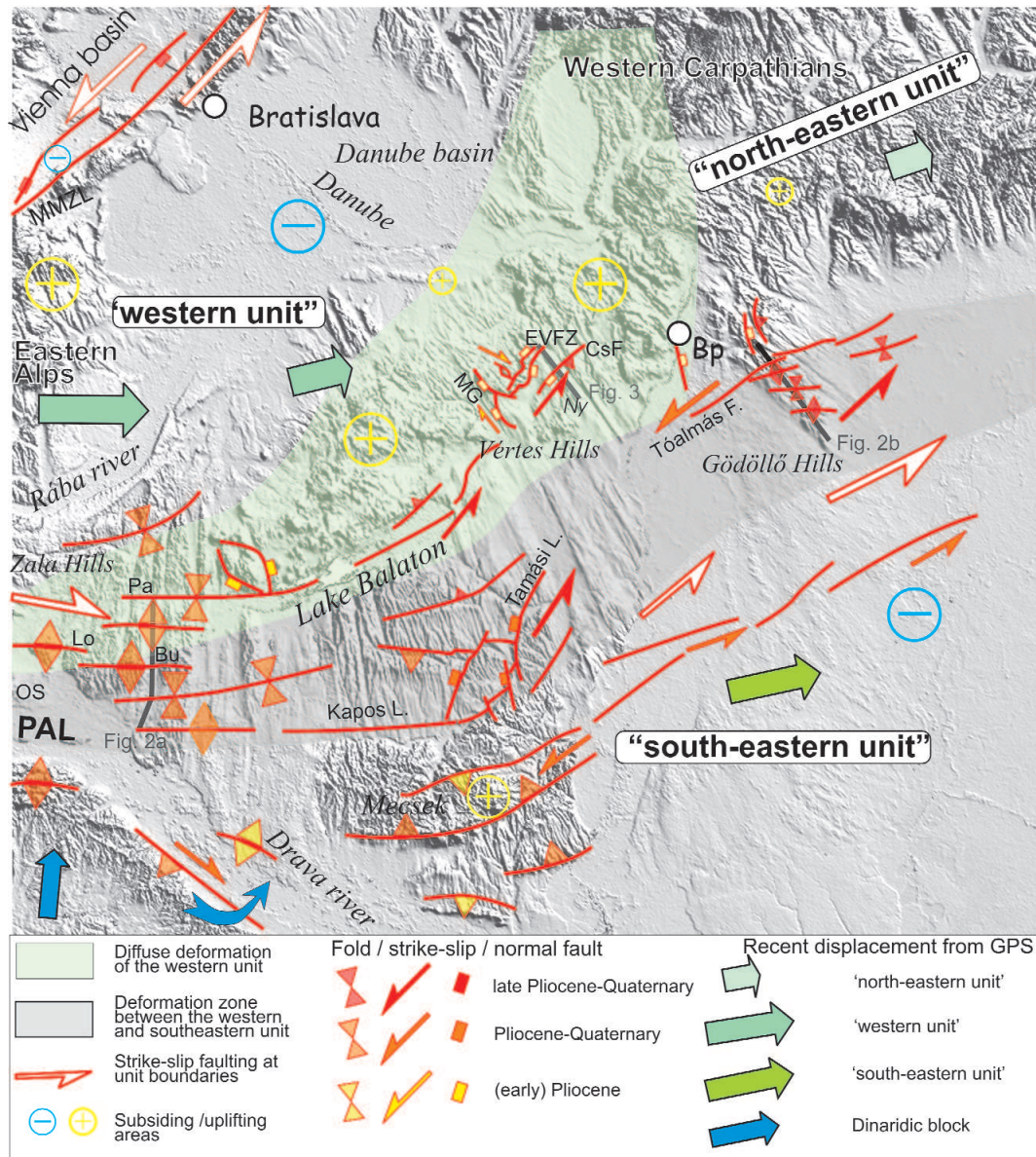
The growing anticlines can induce diversion of creeks or the formation of new drainage divides. The most prominent drainage deflections are on the Alsó-Tápió and Kókai creeks, which are partly flowing parallel to branches of the

Neotectonic structures of the Gödöllő Hills always post-date the youngest imaged post-rift layers. Deformation also affected denudation surfaces and the drainage system, which started to develop (and having been deformed) after post-rift sedimentation, from the middle Pliocene onward. Thus the observed transpressional deformation can be middle Pliocene to Quaternary. Earthquake distribution (Tóth et al. 2002) is interpreted to support the historical activity of the Tóalmás zone and particularly the southern folds, thus they are considered as recently active.



**Figure 3.** A) Geological cross section of the south-eastern foreland of the Vértés Hills parallel to Plio-Quaternary sediment transport direction, based on boreholes; after CSILLAG et al. (2002), modified. Note different age of fault activity indicated by different arrows. Vertical exaggeration is 12.5 times B) Reconstruction of supposed denudation surface upon which dolomite debris was transported from the Vértés scarp to the Nyárjas Hill. Lines indicate possible projections using different dip angle measured on surfaces P1–P3. (Green colours on the figure). Triangles indicate Triassic dolomite clasts derived from the eastern Vértés scarp. Vertical exaggeration is 25 times





**Figure 4.** Simplified map of neotectonic structures of the western Pannonian Basin superposed on digital elevation model. Note that colour code for structures indicates two major events of deformation, while some structures were active through the whole Pliocene–Quaternary period. Partly after FODOR et al. 2003, see text for further data sources. Large arrows schematically show the recent motion of units with respect to fixed Bohemian Massif (GRENERCZY et al. 2002, GRENERCZY this volume). Bu: Budafa anticline; CsF: Csaplár fault; EVFZ: Eastern Vértess Fault Zone; Lo: Lovászi anticline; MG: Mór graben; OS: Ormos–Selnica anticline; Pa: Pacsa syncline; Ny: Nyárjas Hill; PAL: Periadriatic Line

(EVFZ), which has the Vértess Formation in its hanging wall (Figures 3, 4). Some other fault branches are located toward the Vértess foreland and are sealed by late Pleistocene loess, other Quaternary sediments or Quaternary denudation surfaces (pediments). Within the Vértess foreland the Csaplár fault runs along the northwestern slope of the Nyárjas Hill, which is standing out from lower hills of the foreland. The fault crosscuts Pannonian formations and has the Vértess Fm in its south-eastern hanging wall (Figure 3). Other faults in the Vértess area include the eastern boundary fault of the Gánt depression, NW-trending boundary fault of the Mór graben and faults of the Kápolnapuszta pull-apart basin (FODOR et al. 2004; FODOR et al. this volume). Fault slip data indicate normal slip along the EVFZ and normal-sinistral slip along the Csaplár fault, and dextral-normal along NW-trending faults. The stress field was tensional-transensional, having broadly E–W oriented horizontal minimal stress axis.

The Eastern Vértess Fault Zone initiated in the Sarmatian (11–13 Ma) as indicated by abrupt thickening of sedimentary wedge across some fault branches (CSILLAG et al. 2002). During the late Miocene lacustrine transgression the eastern Vértess fault scarp evolved into a fault-controlled, abrasion-dominated cliff indicated by syn-sedimentary dykes and by conglomer-

ates and breccias draping the fault scarp (CSILLAG et al. 2002). Some fault branches and the Csaplár fault displace the Vértesacska Fm (Figure 3). Unfortunately, the age of this formation, critical for timing of faulting, is poorly constrained as latest Miocene or, alternatively, latest Miocene – early (?) Pliocene. These data suggest a more or less continuous, tensional–transtensional deformation from the Sarmatian to the end of the Miocene or, alternatively, up to middle (?) Pliocene.

Geomorphic evolution may suggest a younger deformation in the Vértes foreland. Near the Nyárjas Hill a drainage anomaly can be observed. While most of the linear drainage flows south-eastward, some creeks near the Nyárjas Hill turn north-westward. On the other hand, dolomite clasts were found on top of the Nyárjas Hill derived from the eastern Vértes scarp (Figure 3). These clasts were likely transported along a denudation surface, similar to those found in the Vértes foreland or at the top of the Vértes.

We reconstructed the possible geometry of the denudation surface descending from the Vértes scarp to the Nyárjas Hill applying measured dips of pediment surfaces of the foreland found at lower topographic levels (Figure 3). We concluded that in almost all cases, the denudation surfaces project above the highest surface of the Vértes Hills (CSILLAG et al. 2002). Thus we assume that the Nyárjas Hill was uplifted after the transport and deposition of dolomite debris. The uplift could be realised by inverse reactivation of the Csaplár fault. The uplift can reverse flow direction on the north-western side of the Nyárjas Hill and explain drainage anomalies. Reverse (or sinistral–reverse) slip of the Csaplár fault could happen in the late Pliocene or Quaternary and marks the change of faulting style from tensional to compressional or transpressional. We correlate this change with the onset of “inversion phase” at this part of the Pannonian Basin.

### Integration of new neotectonic structures into a simple model

Compilation of our observations and existing neotectonic data permit to draw a simplified neotectonic map of Transdanubia (Figure 4). This brief discussion does not include all available neotectonic data but try to integrate some new findings into a coherent picture, mainly focused on the TR. In addition, neotectonic structures can be compared to recent data derived from GPS measurements (GRENERCZY et al. 2002; and this volume).

Neotectonic folding of the Zala Hills represent the continuation of folds and transpressive structures in northern and central Slovenia (PÁVAI VAJNA 1925; MÁRTON et al. 2002; VRABEC and FODOR 2004). The syncline south from the Budafa anticline can continue south from the Lake Balaton as suggested by studies of SACCHI et al. (1999), BADA et al. (in press) and CSONTOS et al. (in press). Outcrop-scale structures south of the lake may also be in agreement with compressional deformation (MAGYARI et al. 2004). The ‘Pacsa syncline’ north of the Budafa–Lovászi structure may continue in the depression of the Lake Balaton (FODOR et al. this volume), although the size and existence of folding is questionable. North-eastward from the central part of the lake, NE trending fault zones were demonstrated by seismic sections (SACCHI et al. 1999; VIDA et al. 2001). Small-scale faults and possible liquefaction features near the Várpalota Basin can be associated with these faults (KÁZMÉR et al. 2005). The sinistral–reverse Csaplár fault can represent a further continuation of this sinistral fault array. Finally, sinistral transpression was observed in the central Gödöllő Hills, along the Tóalmás zone, where en echelon folds were associated to faults. Considerable surface uplift, related incision of the Danube (PÉCSI 1959; RUSZKICZAY et al. in press a, b), and strong earthquake activity between Mór and Komárom (TÓTH et al. 2002) can also be associated to other compressive or transpressive Transdanubian structures. Although the exact kinematics of these structures is not clear, the existence of a diffuse zone of neotectonic deformation is a necessity between the eastward moving ‘western unit’ and the ‘stable’ ‘northeastern unit’ (Figure 4).

All these deformation seem to accommodate eastward motion of the westernmost Pannonian Basin and the easternmost Alps. This unit, called ‘western unit’ in FODOR et al. (2003), is moving 1.3 mm/y with respect to the Bohemian massif (GRENERCZY et al. 2002; GRENERCZY this volume). This motion was accommodated in a broad zone, probably incorporating most of the whole Transdanubian Range and south from the Lake Balaton up to the Kapos river.

New GPS data suggest that the ‘south-eastern unit’, south of the Kapos river is also moving eastward (GRENERCZY this volume). The rate of motion within the unit does not seem to change so dramatically than within the ‘western unit’. Thus, an increasing left-lateral displacement can be observed along the boundary zone of the two eastward-moving units, a diffuse deformation belt between the Lake Balaton and Kapos river. The sinistral ‘tendency’ of the faults was observed from the northern Lake Balaton along the Vértes Hills up to the Gödöllő Hills (Figure 4). Sinistral neotectonic slip along the Kapos Line was suggested, among others, by SÍKHEGYI (2002). The western wedge-shaped part of the ‘south-eastern unit’ suffered serious compression–transpression, documented by WEIN (1967), CSONTOS et al. (2002).

Our observations also suggest that inversion (neotectonic phase) was not established at the same time in the whole Pannonian Basin. Folding in the southwest and south started in latest Miocene or earliest Pliocene, while tensional or transtensional deformation persisted in the Vértes and Gödöllő Hills up to middle or late Pliocene (Figure 4). This suggest north-eastward ‘progradation’ of inversion structures toward the central and eastern Pannonian Basin, a conclusion put forward already by TARI (1994) and HORVÁTH (1995).



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# Applications of tectonic geomorphology for deciphering active deformation in the Pannonian Basin, Hungary

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## Abstract

Tectonic geomorphology has proven to be a useful tool for identifying and quantifying active and geologically recent tectonic deformation. Some of these techniques have been applied within the Pannonian Basin, but the region presents some particular challenges to deciphering the pattern of Quaternary deformation. In the Pannonian Basin, Neogene basin-forming extension has been supplanted by transpression. Although the magnitude of shortening appears to be minor relative to total extension, compressional structures appear to be the primary influence on both petroleum accumulation and topographic relief within the basin. In many geological settings around the world, the field of tectonic geomorphology has provided tools for evaluating the nature, pattern, and rates of active deformation. In western Hungary, we have completed morphometric analyses to assess the likelihood of active motion on petroleum-producing anticlinal structures in the region. South of the Zala River, fluvial terraces were mapped and measured over the crests of these structures. The present-day terrace geometry is consistent with anticlinal warping during the Quaternary, and SL indices and Transverse Topographic Symmetry Factors were calculated for the major drainages in the region. These morphometric tests suggest broad, slow uplift and westward propagation of the Budafa-Lovási anticlines as well as possible incipient motion across two parallel structures.

The incentive for future geomorphic work remains strong. Geomorphic mapping of recent deformation could help answer questions regarding thermal history, migration, and overpressuring in Hungarian oil fields, and the timing and rates of deformation in the Pannonian Basin are important for assessing seismic hazard and nuclear safety in the region.

## The Pannonian Basin

The Pannonian Basin (Figure 1) is a back-arc basin filled with up to 7 km of sediment, formed during late early to mid-Miocene extension (ROYDEN and HORVÁTH 1988). The Pliocene and Quaternary represent a period of resurgent tectonics dominated by regional transpression (HORVÁTH and CLOETINGH 1996). The present-day pattern of stress in the basin shows maximum horizontal stress that ranges from north-south in the western Pannonian Basin to east-west in the east (GERNER et al. 1999). Tectonic activity has manifested itself by historical seismicity (Figure 2), including earthquakes near Komárom (1763, 1783, 1806, 1851), Mór (1810), Jászberény (1868), Kecskemét (1908, 1911), Eger, (1925), Dunaharaszti (1956) and Berhida (1985) (STEGENA and SZEIDOVITZ 1991; ZSÍROS 2000, TÓTH et al. 2002).

Recent determinations of earthquake focal mechanisms in the basin (GERNER et al. 1999; TÓTH et al. 2002) suggest that most of the recent seismicity has occurred on reverse and strike-slip faults. The NE-SW-trending Mid-Hungarian shear zone played an important role in Miocene tectonics and now subdivides the basin into two distinct tectono-stratigraphic units. Shallow seismic-reflection profiling on the Danube and on Lake Balaton demonstrated compressional deformation associated with the fault zone (TÓTH et al. 1997; SACCHI et al. 1999, VIDA et al. 2001). Two major branches of Mid-Hungarian shear zone, the Kapos and Balaton Lines, converge and continue into Croatia and Slovenia, where Neogene listric faults have been reactivated as Pliocene to Quaternary reverse faults (RUMPLER and HORVÁTH 1988; HORVÁTH 1995; TOMJENOVIC and CSONTOS 2001; MÁRTA et al. 2002). Surface topography and fluvial incision over these structures suggest recent activity in the form of fault-related anticline growth. Mesozoic rocks are often thrust on top of uppermost Miocene sediments, suggesting major inversion of basin-margin faults of the southern Pannonian Basin (WEIN 1967; PRELOGOVIC et al. 1995).

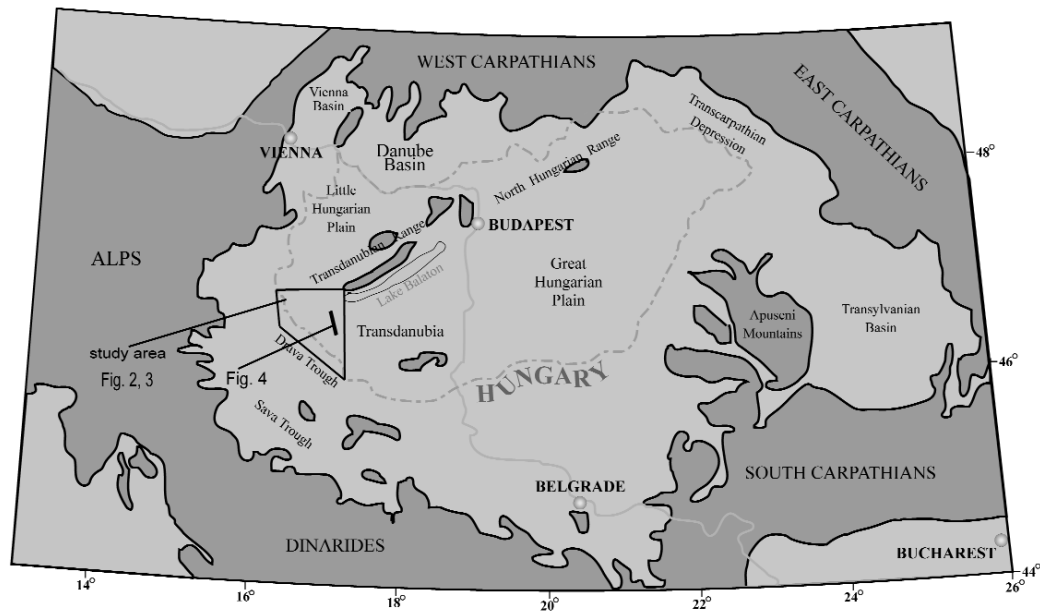


Figure 1. Features of the Pannonian Basin and surrounding regions

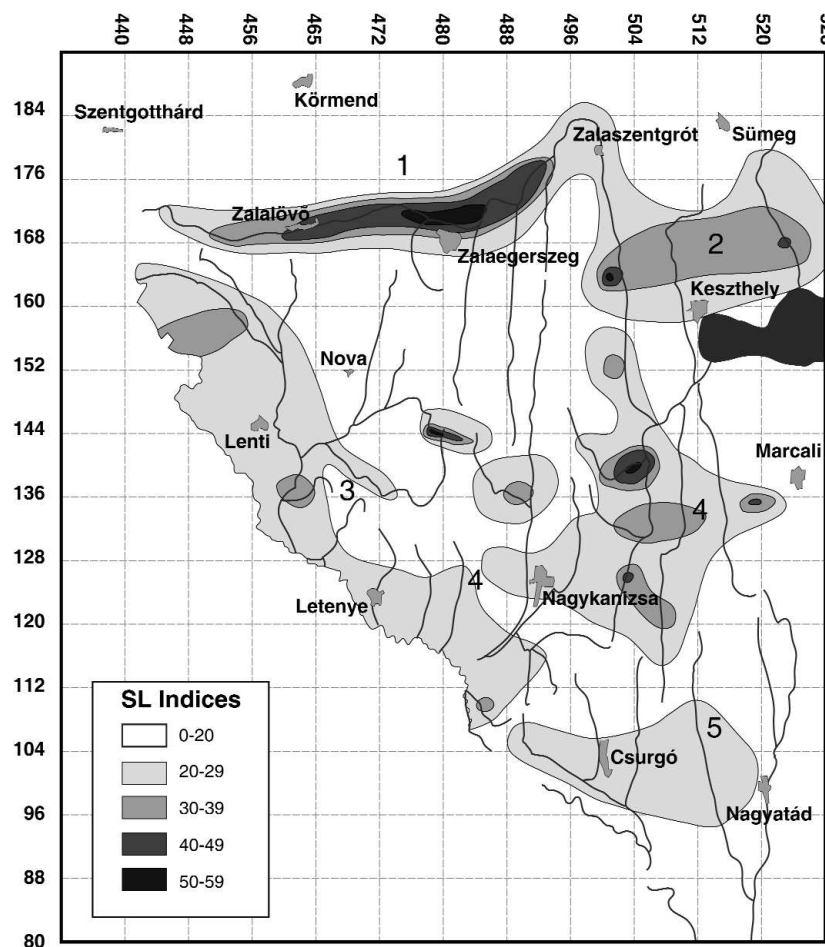


Figure 2. SL Index values across the western Hungary study area. Figure shows smoothed and contoured SL values. High SL values indicate deviations in cross-sectional stream profiles from their idealized, equilibrium form, which may be indicative of active and/or recent tectonic activity in an area (Map created by A. KOVACS)

It has been suggested that the overpressured hydrocarbon fields of the Pannonian Basin can be best understood by supposing several hundred metres of erosion from atop the structural highs during the Quaternary (HORVÁTH 1995; HORVÁTH and CLOETINGH 1996). Various data suggest a Quaternary shift from subsidence to uplift (LEÉL-ÖSSY 1997; ANGELUS et al. 1997; SÍKHEGYI 2002). Preliminary estimates of uplift and subsidence in the Pannonian region have been made based on sediment thickness in basins and elevations of river terraces (RÓNAI 1974, RUSZKICZAY et al. in press) and geodetic surveys (JOÓ 1992). Although the upper portions of seismic sections are often cut, the uppermost (upper Miocene) reflections are always tilted and eroded approaching the basin-margin ranges and intra-basinal hills. Estimates of basin-margin uplift suggest 600 to 800 m of Quaternary motion (HORVÁTH 1995; VAKARCS et al. 1994).

### **Basin inversion**

Although there remains some disagreement about its definition, most researchers take “basin inversion” to mean the reversal of movement on basin-forming extensional faults during subsequent shortening or transpression (WILLIAMS et al. 1989). The degree of inversion can range from minor to complete overprinting of the original basin, as seen in the Alps (BUTLER 1995) and the Pyrenees (HAYWARD and GRAHAM 1989). In the latter case, the pre-existing extensional fabric can dominate long after compression is pervasive. Conversely, even relatively minor shortening can result in compressional structures that dominate the near-surface geology and are crucial to petroleum prospectivity (MORLEY 1995).

Space problems during inversion can cause basin fill to be expelled laterally, through backthrusting, and/or upward (HAYWARD and GRAHAM 1989). Upward expulsion in the form of shallow anticlines is an important mechanism for creating hydrocarbon traps, and mildly inverted basins are associated with the concentration of petroleum into large fields (MACGREGOR 1995). In addition, large-scale vertical motions may affect thermal history and maturation processes (CHRISTIAN 1996). As one author put it, “In many cases, inversion tectonics is the controlling factor on hydrocarbon prospectivity” (BUCHANAN and BUCHANAN 1995).

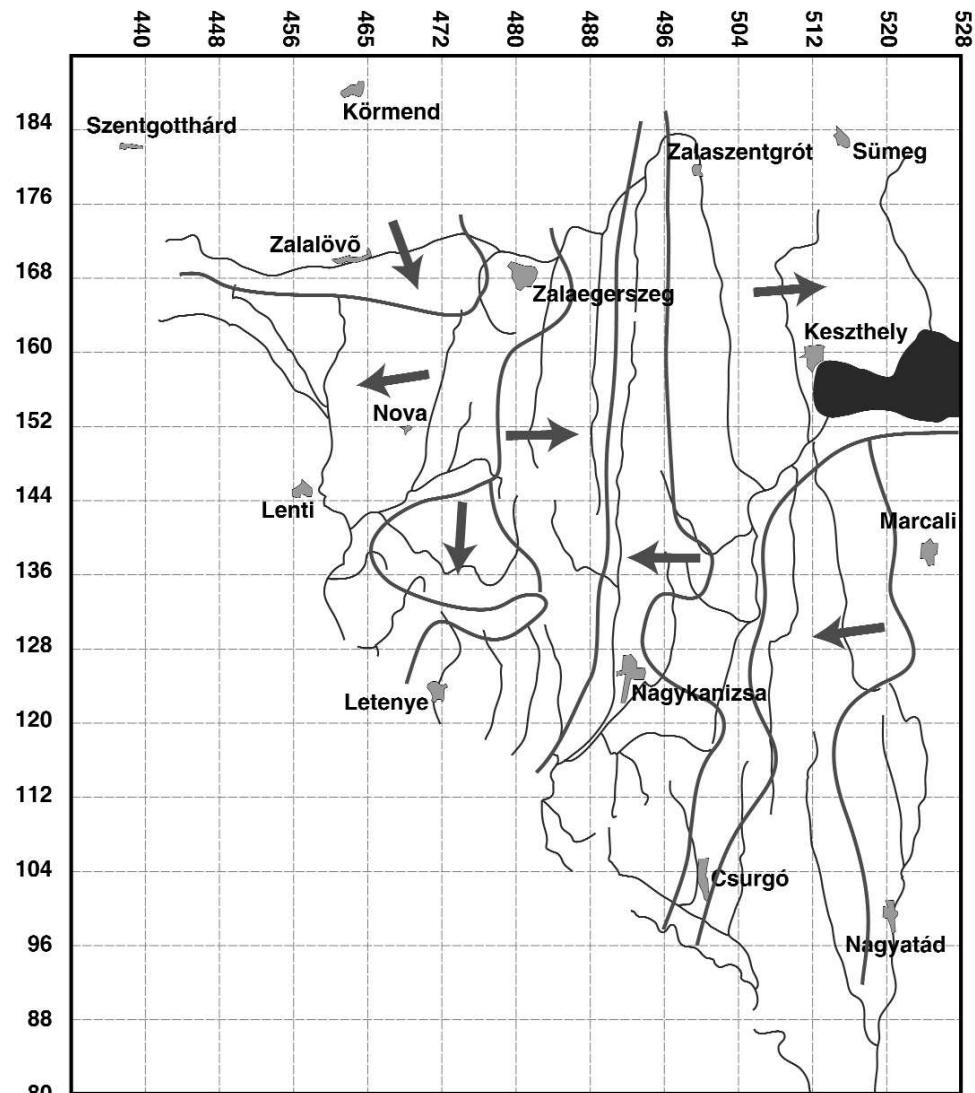
### **Geomorphic tools**

Tectonic geomorphology is a relatively new, interdisciplinary field at the boundary between structural geology/tectonics and surface processes. The most common goal of tectonic geomorphology research is to use Quaternary landforms and stratigraphy to infer the nature, patterns, rates, and history of near-surface tectonic processes. In many cases, older geological markers indicate deformation different from, in some cases even opposite, the active processes (for example, inversion). Tectonic geomorphology provides a whole kit of tools for deciphering the most recent activity on live structures (KELLER and PINTER 2002; PINTER 1996).

Many of the tools applicable to the study of basin inversion in general, and the Pannonian Basin specifically, can be illustrated by briefly discussing my on-going project in California. The Northern Channel Islands are the southernmost anticlinal range of the Western Transverse Ranges, and are inferred to be growing in response to slip on an underlying thrust ramp (SHAW and SUPPE 1994). Uplifted coastal terraces are preserved around the perimeters of the islands, and my students and I are precisely measuring these terraces at numerous sites using combinations of laser surveying, kinematic GPS positioning, and numerical analysis of DEMs (terrace-recognition algorithms and krigging). Combined with interpretation of public and industry seismic sections — 4 — from the surrounding shelves, this data has allowed us to determine the following: 1) patterns and rates of deformation, 2) geometry of the underlying fault, and 3) the changing history of slip during the Quaternary. Without going into all the details here, surface deformation is consistent with a listric fault geometry at depth, and slip on a listric thrust can be uniquely derived from a surface-deformation field using results of PINTER et al. (2001, 2003) and SEEGER & SORLIEN (2000). The Northern Channel Islands share many inversion-related characteristics with the Pannonian Basin: 1) extensional fabric dating to the late Tertiary, 2) modern transpressional stress, 3) anticlinal uplift over listric-shaped faults, and 4) terraces (coastal in California, fluvial in Hungary) that record late Quaternary surface deformation.

### **Morphometric analyses in western Hungary**

Work in western Hungary focused on two potentially active structures, the Budafa and Lovászi anticlines. Morphometric analyses included calculation of regional geomorphic indices using 1:200,000-scale topographic maps, and calculation of more focused geomorphic indices a scale of 1:50,000 over an area bordered by the Kis-Balaton valley to the east, Sand to Semjénháza to the south, and Nova to Söjtör to Felsőrajk to the north.



**Figure 3.** Transverse Topographic Symmetry Factor (T) values across the western Hungary study area. Figure shows domains of similar asymmetry. Transverse Topographic Symmetry is a vector value, including both a direction and a magnitude at all locations, but only the vector directions are summarized here. Like all geomorphic indices, Transverse Topographic Symmetry is just a reconnaissance tool for suggesting areas worthy or more focused research. Here, the N-S orientation of T domains is most likely an artifact of the dominant N-S drainage pattern; the more suggestive signature, at least to this author's eye, are the subtle deviations from the N-S pattern, for example along a line just north of Lenti-to-Nagykanizsa and possibly another trend from Zalalövő to Zalaegerszeg to Zalaszentgrót (Map created by A. Kovács)

Geomorphic indices calculated included the Stream Length – Gradient Index (“SL Index”) and the Transverse Topographic Symmetry Factor (TTSF, or T). Both indices were calculated at both the 1:50,000 and 1:200,000 scale. SL Index is calculated as:  $SL = [\Delta H / \Delta L] L$ , where  $\Delta H / \Delta L$  is the dimensionless gradient of a channel reach and  $L$  is the distance from the reach centroid to the channel head. SL Indices can detect areas of anomalous uplift within a landscape. The Transverse Topographic Symmetry Factor is calculated as:  $T = D_a / D_b$ , where  $D_a$  is the distance from the channel to the basin midline and  $D_b$  is the distance from the lateral basin margin to the basin midline. TTSF can detect areas of lateral tilting. For details of both indices, see KELLER and PINTER (2002).

A map of SL indices was created using values for all segments on stream channels crossing the 20 m contours on the 1:200,000 topographic maps (Figure2). The SL indices varied between 5 and 60, and five SL categories were distinguished: 0–20, 20–30, 30–40, 40–50, and 50–60. On the 50,000-scale maps in the focused study area, SL indices were calculated on stream segments between the 10 m contour lines. Regionally, the highest values were in the area of Galambok. The Neszele–Pózva, Zalacsány, and Mihály areas also contained high values, forming rough SSW–NNE-oriented zones. The Principális valley to the north of Magyarszerdahely contained the lowest SL values in the study area.



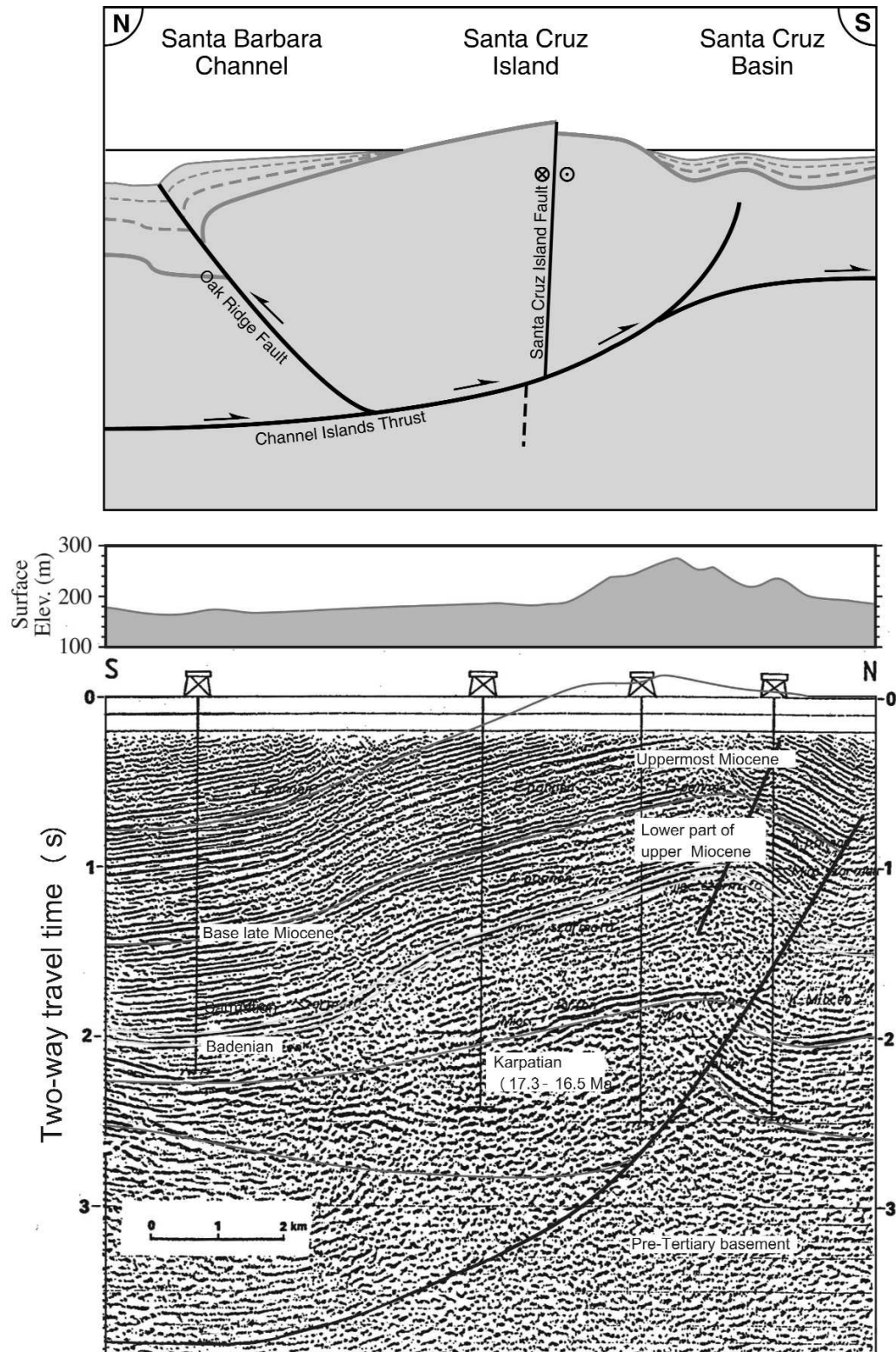


Figure 4. The Budafa field, Hungary's oldest, is located southwest of Budapest. The Miocene listric normal fault has been reactivated as a thrust, over which the Quaternary sediments and the surface topography are deformed (after HORVÁTH 1995)

Keeping in mind that SL and other geomorphic indices are just reconnaissance tools, the results of the mapping above suggest that several areas and trends may merit examination in greater detail, including: the Zalaszentgrót–Zalalövő zone along the Zala River, the Zalacsány-to-Tapolca trend (2), the Garabonc–Magyarszerdahely–Lovászi area (3), the (4) Marcali–Galambok–Nagykanizsa–Letenye area, and the Zenta–Csurgó–Zákány region (5).



Maps of Transverse Topographic Symmetry Factors was created in the regional (1:200,000) and focused (1:50,000) study areas defined above. T values varied from 0 to 0.9, with high values indicating areas of potential tilting influence on drainage pattern. Like all geomorphic indices, single values of T have no significance, but rather we look to broad areas with similar values to indicate a possible tectonic signature on the drainage pattern. In the case of Transverse Topographic Symmetry Factor, the index includes both a numerical value (asymmetry of the valley) as well as a vector orientation (direction of asymmetry). The map presented here (Figure 3) shows an interpretation of the raw T magnitudes and directions, in which areas of similar values have been delineated. The precise locations of the domain boundaries on this figure are approximate, but the domains should be seen as reconnaissance tool for suggesting areas of possible differential neotectonic activity. Transverse Topographic Symmetry Factor and the SL Index are potentially sensitive to different types of activity — uplift and subsidence of discrete blocks versus broad tilting — but the calculation of both sets of indices in western Hungary suggests some similar patterns. In particular, the Garabonc–Magyarszerdahely–Lovászi and Miháld–Zalalövő–Zalaegerszeg–Murakeresztúr zones show anomalous values using both techniques. Such coincident results from multiple indices often represents the strongest evidence that a real tectonic signature may be present.

### Significance

Geomorphic indices seen to be in agreement with suggestions of HORVÁTH (1995) and FODOR et al. (2003, 2004) that southwestern Hungary is characterised by Quaternary folds. Tectonic geomorphology research has the potential to yield results significant to the geology of the Pannonian Basin, to petroleum geology in general, and to a broader range of applications in Hungary. The principal local goal is to test the hypothesis that uplift and petroleum migration on the Lovászi, Budafa, and other structures is a recent phenomenon, which explain the anomalous maturity in these fields (Figure 4). The broader goal of this research is to develop reconnaissance tools for identifying active basin inversion and more local tools for determining the rates and timing of fault slip and fold growth. Finally, it should be mentioned that studying active deformation in the Pannonian Basin has a broad range of applications. In particular, the seismic hazard in Hungary remains largely unknown. Evidence of patterns and rates of slip on Quaternary-active faults would have significant implications for the Paks nuclear power plant south of Budapest, the Gabčíkovo hydroelectric dam in the Little Hungarian Plain, and the high-level nuclear waste site now planned for Mecsek Mts (a probable Quaternary inversion structure).

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# Karsthydrogeology of the Transdanubian Range, Hungary: Geological constrains and human impact on a unique karst reservoir

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Research of the karstwater-bearing formations is a practical and a very interesting theoretical question for the Hungarian hydrogeologists, because more than 90% of drinking water is from subsurface water, and one third is from karstic water aquifers. Furthermore, this karst water is the source of the world-famous thermal springs, which are situated at the margins of the karst mountains (Budapest, Hévíz, Harkány, Miskolctapolca, Eger). During the last years of our work we focused on the Transdanubian Range, because the water stored in the karstified rocks is one of the 40 main drinking-water reservoirs of Hungary (Figure 1).

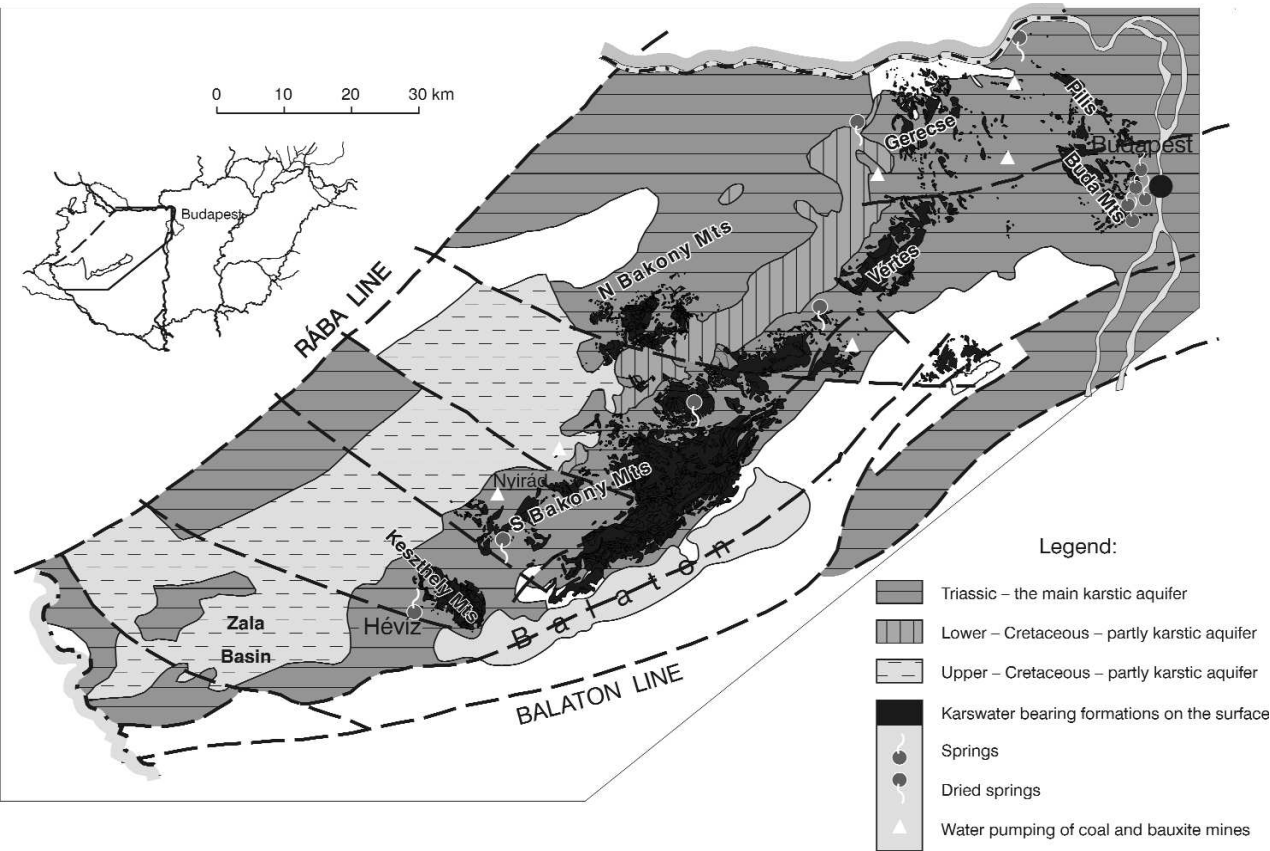


Fig. 1. Karstwater bearing formation of the Transdanubian Range

Due to the withdrawal of the coal and bauxite mining, the original karst water level of this region dropped by 100–180 m in mining areas, and by an average of 30 m throughout the entire region in the period between 1960 and 1990. The decline of water level caused numerous environmental problems. The water level has decreased in the wells, the boggy areas have dried up, and several springs have run dry, among them the famous Tata springs. Water inflow of the world-famous Hévíz thermal lake decreased from 600 l/min to 300 l/min, as well as, the Budapest thermal springs experienced some loss of water.

In 1990 the water pumping decreased, due to environmental and financial reasons, and the rehabilitation began. The refilling process offered a unique opportunity to study the geological factors that determine water flow between the recharge and discharge areas.

From hydrogeological point of view all geological features of the karstwater aquifers are essential as

— the spatial position of the permeable and impermeable rocks and their relationship with each other and with the research areas,

— as well as the fissure and cave system in the permeable and the fissure system in the impermeable formations, which were developed partly by the original characteristic of development, and partly by the subsequent karstification and tectonic processes.

The geological elements were determined by the original sedimentation, and post-depositional tectonic events.

### **Evolution of the Transdanubian Range from hydrogeological point of view**

#### *Geology*

The main karstwater aquifer of the Transdanubian Range (TR) is approximately 2.500–3.000 m thick, and it mainly consists of Triassic, in the first place Upper Triassic carbonate rocks. In a significant part of the Bakony Mountains the Triassic is covered by Jurassic rocks, mostly carbonates, which practically settled with continuous sedimentation. However, the thickness of Jurassic is less than 200 metres.

Different types of Triassic rocks were formed in varying paleo-environment. The most characteristic facies is the pure carbonate rocks (Dachstein Limestone, Hauptdolomit), other carbonates contains fine clastic components in variable amount. In hydrological point of view the marl formations (Veszprém Marl and the Kössen Formation) are non-permeable, and they dissect the thick pure carbonate formations. Obviously, the almost pure carbonate formations are not homogeneous as well, their permeability vary. They have a variable primary porosity, determined by the sedimentation and lithification, and the secondary porosity is variable as well. The secondary porosity formed after lithification and it is determined by tectonic events and the karstification processes.

The first known sedimentary hiatus of the main karstwater aquifers was during the earliest Jurassic. This was the first, though limited possibility of karstification of the uppermost Triassic Dachstein Limestone. During the Jurassic there was no long terrestrial periods with the possibility of significant karstification.

The Austrian and pre-Gosau tectonic phases, before the Late Cretaceous were important in the evolution of the main karst aquifer and the evolution of the whole Transdanubian Range. Due to the compressional events a major synclinal structure was formed, and the Transdanubian Range became moderately folded and imbricated (TARI 1994, 1995). At that time, in the central part of the northeast–southwest trending syncline Early and early Late Cretaceous formations were on the surface. These sequences consist of mostly aleurolite, argilleous and marl sediments, with some carbonate intercalations of significant thickness. Northwards and southwards, in a 15–25 km wide zone, gradually older formations were exposed on the surface, due to the synclinal structure. The tropical climate in that time created extremely favourable conditions for the karstification of the carbonate rocks, which had been on the surface for long period.

The pre-Senonian surface is overlain by Upper Cretaceous sedimentary rocks, among others with the karst-water-bearing Ugod Limestone Formation. This biogenic limestone is well karstified, due to the hiatus between the Late Cretaceous and the middle Eocene sedimentation.

According to our present knowledge, the Transdanubian Range are of allochthonous structural position, and two important tectonic lines, the northwestern Rába and the southern Balaton Line form their limits. The original location of the Transdanubian Range was north of the Southern Alps (KÁZMÉR and KOVÁCS 1985). It was pushed out along the above-mentioned transcurrent lines from its original formation area (approx. from actual Graz zone) during the Cenozoic, mainly in late Oligocene and in early Miocene, as the result of the convergence between the African and European plates.

The formation of numerous tectonic lines, which plays an important role in the subsurface water flow is connected to this “escape” event. The tectonic movements began in late Oligocene and reached maximum intensity in the Miocene though the syncline structure were preserved. Of course the TR did not behave as one rigid block, but was broken into coulisse-like blocks bounded by WNW–ESE tectonic lines (horizontal strike-slip faults) (MÉSZÁROS 1983). The role of



the horizontal is displacement zones are multiple. These strike-slip faults influence water movement, they have good permeability along the tectonic line, while they are impermeable for the perpendicular fluid flow. They play an important role in the formation of basins filled with thick Miocene sediments. They cut through the original Mesozoic facies juxtaposing blocks with different hydrogeology, transmissibilities and geological structure. As a consequence, these tectonic lines determine the flow of subsurface water.

From the younger sediments the carbonate and coarse detrital Badenian and Sarmatian (Middle Miocene) deposits and the Pannonian (Late Miocene) clastic formations are of main importance. The Pannonian and younger tectonic movements have important hydrogeological role, too. These events influenced the present morphology of the area (FODOR et al. this volume), mainly by the regional uplift of the TR due to compressional forces (HORVÁTH 1995). The elevated structural position of the TR determines the recharge and discharge areas as well as the surface and subsurface water divides.

### Lake Hévíz

On the south-western side of the syncline structure of the Transdanubian Range the most important natural discharge of karstic water is the Hévíz Lake. The springs feeding the Lake originate in the spring cave (discovered by divers in 1975), that formed in Pannonian sandstone. On the east side of the cave the spring is 17.2 °C, cold, on the west side of the cave the spring is 39.6 °C warm. The mixture at the mouth of the cave is 38.8 °C warm. Isotopic studies showed that the warm spring has water of sp a few ten thousands years old, while the cold water is a few thousands years old only. These facts unambiguously show that there are two different flows of water.

The Hévíz Lake is situated in a fault zone, which creates steps in the upper surface of the Triassic rocks (Figure 2). This fault system has predominantly dip-slip movements during the Pannonian. This movement reactivated the Oligocene– Miocene slip-strike fault, which borders the Keszthely Mountains on the west. The Keszthely Mountains is an uplifted block of Triassic formations. Pannonian clastics surround this footwall karst, and are characterized with decreasing grain size with increasing distance from Triassic outcrops.

The recharge area of the cold water flow is the nearby Keszthely Mountains, where the well karstified formation are on the surface (Figure 3). The water moves in the karst westward along a short flowpath. Due to the side low transmissibility strike-slip zone of the western Keszthely Mts situated perpendicular to the flowpath, the water moves upward and feeds the cold side of the spring. The coarse Pannonian clastic sediments, settled directly on the Triassic karst, gave favorable condition to the formation of the spring cave of the Hévíz.

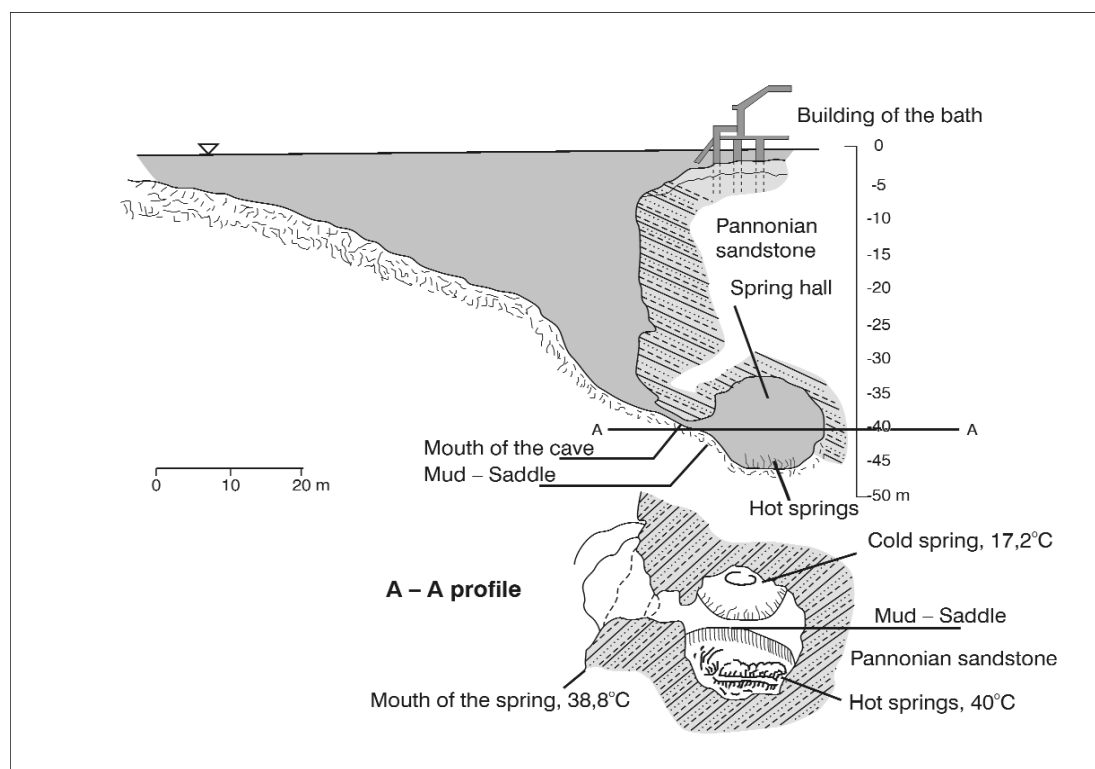


Figure 2. The Hévíz spring lake

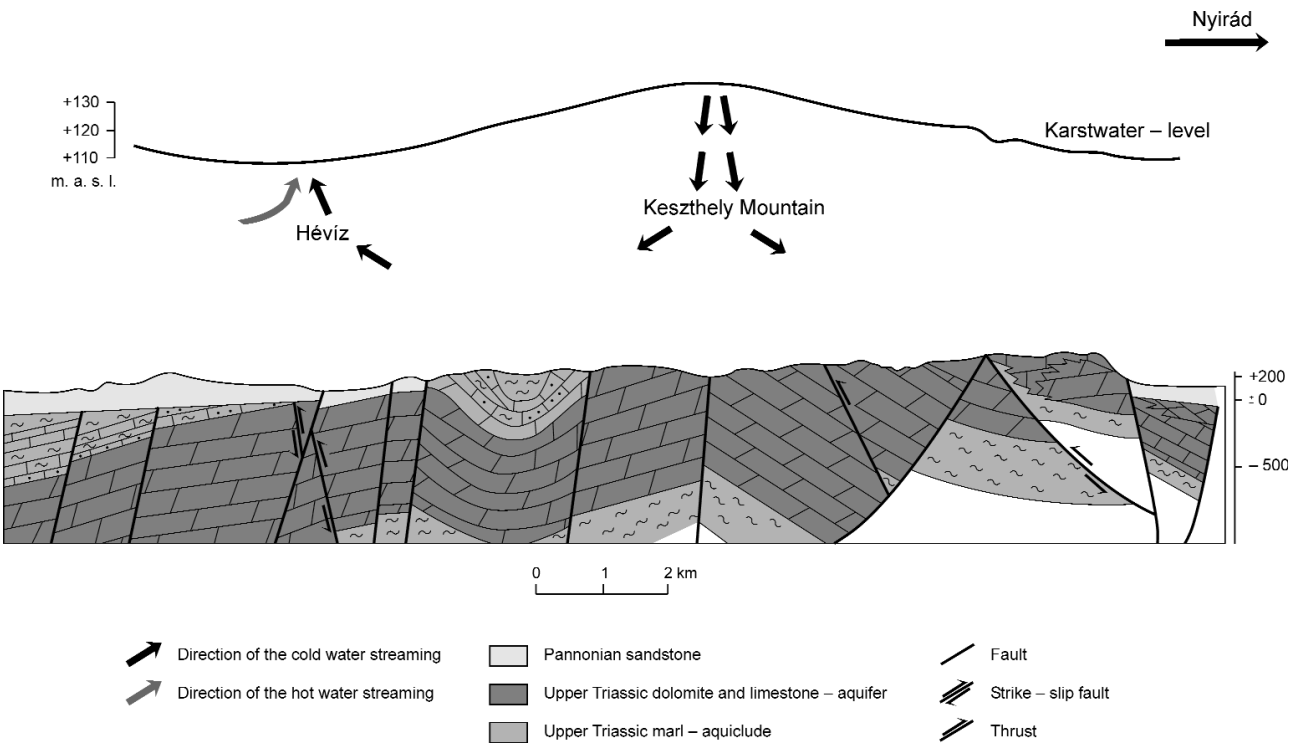


Figure 3. The Hévíz spring-lake profile between the lake Hévíz and the Keszthely Mountains, showing geological cross section (after BUDAI et al. 1999) and the supposed fluid motions

*The warm side:* Further northwest flow path and the geology is more complex, connected structurally to karstic formations, which are on the surface in the Transdanubian Range, NW or N from the Keszthely Hills. The precipitation, recharges in the mountaneous areas first to southwest in the depth (Figure 4). Then it reaches the fault zone juxtaposing carbonate rocks and impermeable Triassic and Upper Cretaceous formations. This structure —the Nagylengyel Zone — is about 10 km wide trending W–SW from Hévíz Lake. The fault zone directs the flow back (SE) towards Hévíz Lake. As a consequence of this long flow path the karst water warms up, and moves up to the spring cave on the west side of the fault fault.

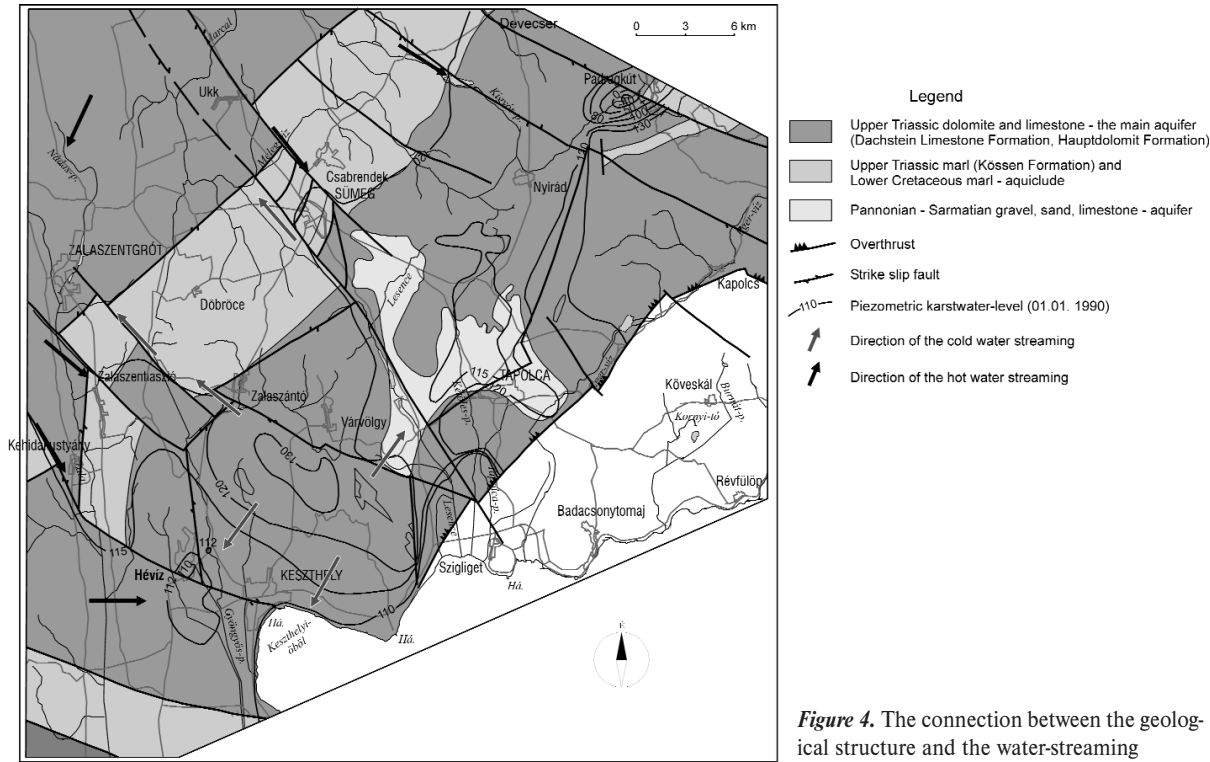


Figure 4. The connection between the geological structure and the water-streaming

Water withdrawal at Nyírád drained the warm water flow with 300 m<sup>3</sup>/day. The withdrawal significantly decreased since 1990, today is 25 m<sup>3</sup>/day. As a consequence, in the area the piezometric water level increased, and the yield of Hévíz Lake increased, today it is 435 m<sup>3</sup>/day.

### Zala Basin

The Nagylengyel Zone forms the border between the Transdanubian Range and the Zala area. At the Zala area, several thousand-metre thick Pannonian sediments cover the outstanding karstwater-bearing formations, which are on the surface in the mountain range. Here, the synclinal structure continues, though it was strongly fractured in several phases (Figure 4.).

Water quality of Zala region is significantly different than the karst water quality of the mountainous area (Figure 5). Total dissolved solids of Zala region is significantly higher than in the mountainous area. In the Zala region the significant fluid reservoir is the Upper Cretaceous Ugod Limestone that contains great amount of hydrocarbons. The hydrocarbon occur at the southwestern side of the Nagylengyel Fault Zone alone (Figure 6).

### Hydrogeological Units of the southwestern Transdanubian Range

Due to the geological structure the Transdanubian Range divides into two great hydrodynamic units, which are different by water-composition and by the occurrence of hydrocarbons, on the SW side of the TR.

The border of these hydrodynamic units formed by a NW–SE trending, 8 to 10 km wide fault zone. This strike-slip fault is excellent water conduit in the direction of the strike, however it has low water conductivity perpendicular to the strike.

The hydrodynamic unit of Transdanubian Range is situated northeast from this zone. Water infiltrates in the elevated parts of the mountain, and

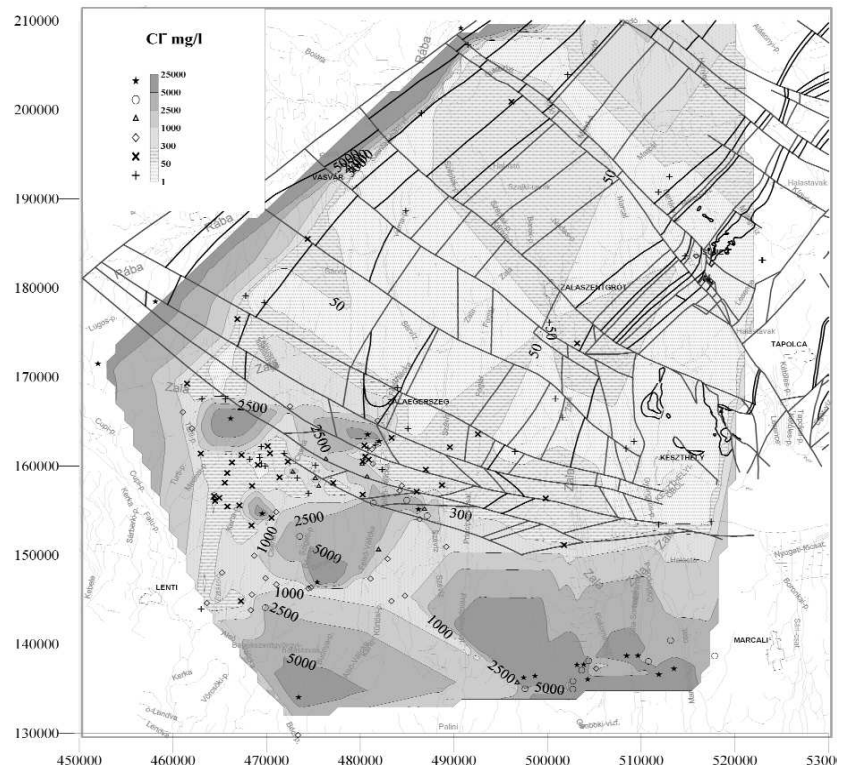


Figure 5. Cl<sup>-</sup> content of the karstwater (after I. HORVÁTH)



Figure 6. The pre-Senonian surface in the Zala region



then it moves toward the natural discharge areas at the margins through the karst formations. The karstwater moves in SW–W direction, when it reaches the impermeable fault zone it turns toward E–SE and it reaches the surface through the springs of Lake Hévíz.

The Zala region situated SW from the Nagylengyel fault zone and forms a separate hydrodynamic unit from the Transdanubian Range. The direction of fluid flow is from NW to SE or from W to E in the deep-seated karst aquifer. The northeastern border of the fluid movement is determined by the impermeable fault zone. Part of the Zala basin is situated in Slovenia and bordered by high mountains. These mountains are the recharge areas of the Zala region, and the difference in elevation provided the hydraulic pressure that drives the water and hydrocarbon flows.

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# Tectonic development, morphotectonics and volcanism of the Transdanubian Range: a field guide

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**Key words:** neotectonics denudation, faults, GPS, phreatomagmatic, basalt, maar, tuff ring, scoria, peperite, Neogene

## Abstract

Between 24th August and 1st September, 2002 the Geological Institute of Hungary held an International Workshop on “Application of GPS in plate tectonics, in research on fossil energy resources and in earthquake hazard assessment”. Following the plenary sessions a short field trip ran to western Hungary with an aim to present field sites where the interplay of neotectonics, geomorphic evolution, various denudation processes, sedimentary systems and interaction of volcanism and sedimentation in a fluvio-lacustrine environment can be studied.

The Balaton Highland represents a classical area where the Middle Triassic asymmetric rifting of the Neo-Tethian carbonate platform was documented. Classical and recent studies described the major elements of the Cretaceous shortening phase. The southeast-vergent thrusts can be connected to the formation of the Transdanubian syncline structure, and probably also to nappe emplacement of the whole unit. Miocene and eventually Plio-Quaternary reactivation of some of the older thrusts could represent a working hypothesis for future neotectonic studies, dealing with the neotectonic uplift of the Balaton Highland and formation of the Lake Balaton depression.

Some publications and recent mapping work in the Transdanubian Range revealed the deformation of the late Miocene (Pannonian to Pontian) sequences. This Pliocene or Quaternary deformation shows varying kinematics, from blind reverse faulting and folding (Zala Hills) to normal faulting (Vértes Hills) through transpressional or transtensional fault zones (Lake Balaton, Vértes Hills). The timing and connection of this deformation pattern to measurements of recent stress field and recent crustal motion (GPS) represents the future challenge of neotectonic research.

Western Hungary is also a region, where Neogene small-volume alkali basaltic intracontinental volcanism occurred. The systematic study of the strongly eroded volcanic edifices, often degraded to their volcanic root zones, gives a good opportunity to estimate the level of erosion over a large area. Recent researches demonstrated the presence of extensive Pannonian sedimentary cover in syn-volcanic time. The distribution of volcanic vent remnants, their asymmetric shape and their level of erosion are in good concert with other geological field data supporting a 100 m-scale erosion of Pannonian sedimentary cover and the presence of hydrogeologically active zones such as valley systems, swamps, wetlands in the Pliocene. Our recent analyses reinforce earlier assumptions that wind erosion played important role in formation of the Transdanubian valley system. Evidences include the occurrence of ventifacts, wind-polished rock surfaces at different topographic positions, the presence of deflated large-scale landforms (yardangs), the lateral transition of wind-blown sand and loess, and the relative scarcity of fluvial sediments. The large shallow lacustrine system and its late Pleistocene–Holocene sediments give opportunity to connect climatic and landscape evolution. Our present results suggest that only a combined tectonic, volcanologic, geochronologic, sedimentologic approach can help to unravel Plio-Quaternary deformation and related landscape evolution.

## Introduction

With the support of the Hungarian Science and Technology Joint Fund, an international Workshop on “Application of GPS in plate tectonics, in research on fossil energy resources and in earthquake hazard assessment” was held between 24th August and 1st September, 2002 at the Geological Institute of Hungary (MÁFI). A field trip followed the workshop to the Transdanubian Range (TR). The goal of the field trip was to initiate discussion about the tectonic aspect of the TR,

the best-known part of Hungary from general geological point of view. This region is excellent to study neotectonic deformations, Pliocene–Quaternary basaltic intraplate volcanism and to unfold the connection of volcanism, deformation and landscape evolution in a tectonically active area.

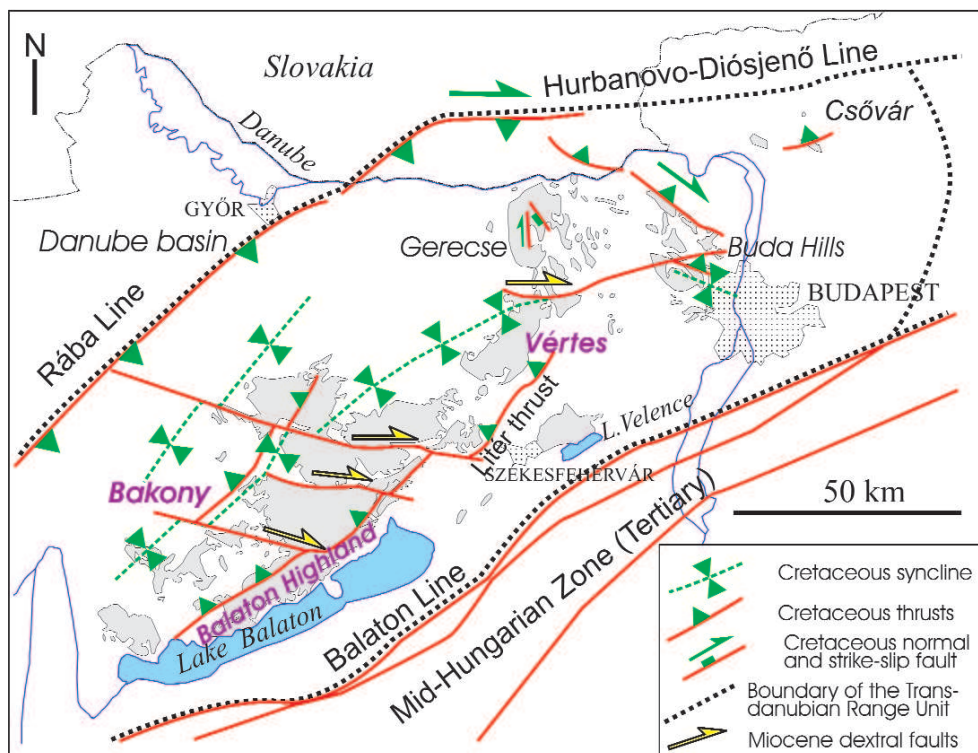
The excursion was organised in two parts. Pre-workshop one-day field trip gave introduction to Tertiary deformations of northern Transdanubian Range and dealt with neotectonic significance of the Danube gorge area. Post-conference three-day field trip was dedicated to the central and southern Transdanubian Range, namely to the Vértes Hills and Balaton Highland areas where recent geological mapping (BUDAI et al. 1999) and specific researches (CSILLAG et al. 2001, 2002a, b; BUDAI and CSILLAG 2000; NÉMETH et al. 1999; CSERNY and NAGY-BODOR 2000; FODOR et al. 2002) revealed new data and interpretations (Figures 1–3.). The present work gives a general geological overview of the TR, and describes the stops of the field trip. As a field guide, this work contains already published data, new observations not published elsewhere and also some models, which still wait for further studies.

### Summary of structural evolution of the Transdanubian Range

The Transdanubian Range (TR) and the Balaton Highland have a complex deformation history although detailed modern structural documentation and analysis started relatively recently (MÉSZÁROS 1982, 1983; BALLA and DUDKO 1989; DUDKO 1991; DUDKO et al. 1992; FODOR et al. 1992; KISS et al. 2001). This summary gives a very short review of the major tectonic events not intending to describe the total structural complexity.

The trace of Middle Triassic rifting was mainly discovered on the basis of differentiated sedimentary sequences (BUDAI and VÖRÖS 1992). Late Triassic climatic changes superposed on faulting resulted in formation of good hydrocarbon source rocks (HAAS 1993). Renewed disintegration of the carbonate platform and associated subsidence in the Early and Middle Jurassic led to variable, although thin deep-water sequences and to the occurrence of sedimentary dykes and faults (GALÁČZ 1988; FODOR and LANTOS 1998; VÖRÖS and GALÁČZ 1998). The most important pre-Tertiary tectonic event is marked by the folding and thrusting of the TR in the Cretaceous, mainly in the early to middle Albian (started probably in the Aptian), which resulted in the so-called “synclinal structure” of the TR (LÓCZY 1925; TARI 1994, 1995; FODOR 1997, 1998; KISS and FODOR 2003 — Figure 1).

In a regional context the shortening was associated with nappe thrusting of the TR onto different Austroalpine units (HORVÁTH 1993; TARI 1994; FODOR and KOROKNAI 2000; CSONTOS and VÖRÖS 2004). All these deformations led to



**Figure 1.** Simplified structural sketch of the Transdanubian Range (TR) showing mainly the Cretaceous structures (after TARI 1994, FODOR 1998; KISS et al. 2001, KISS and FODOR 2003, modified). Pre-Tertiary outcrops are in grey



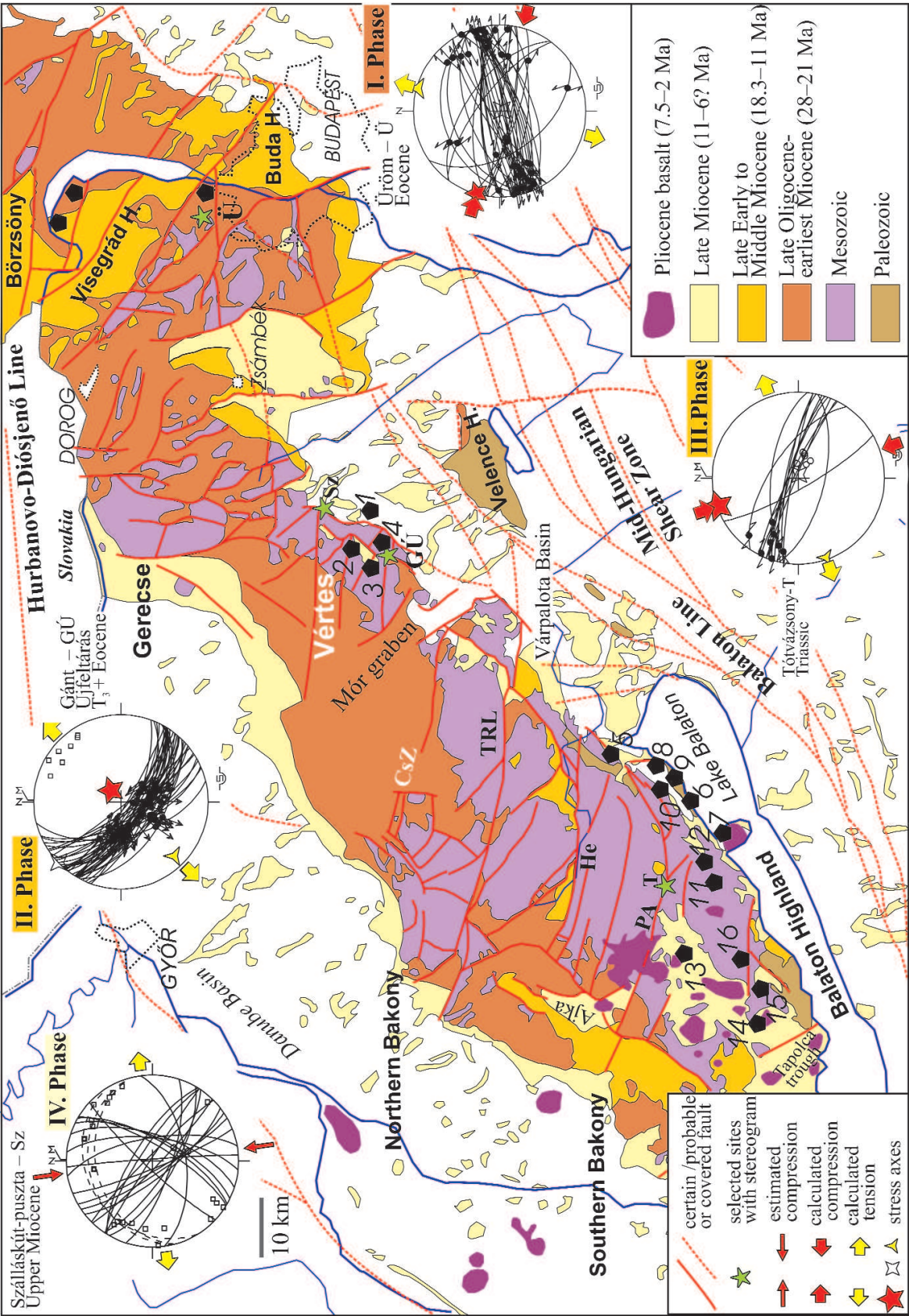
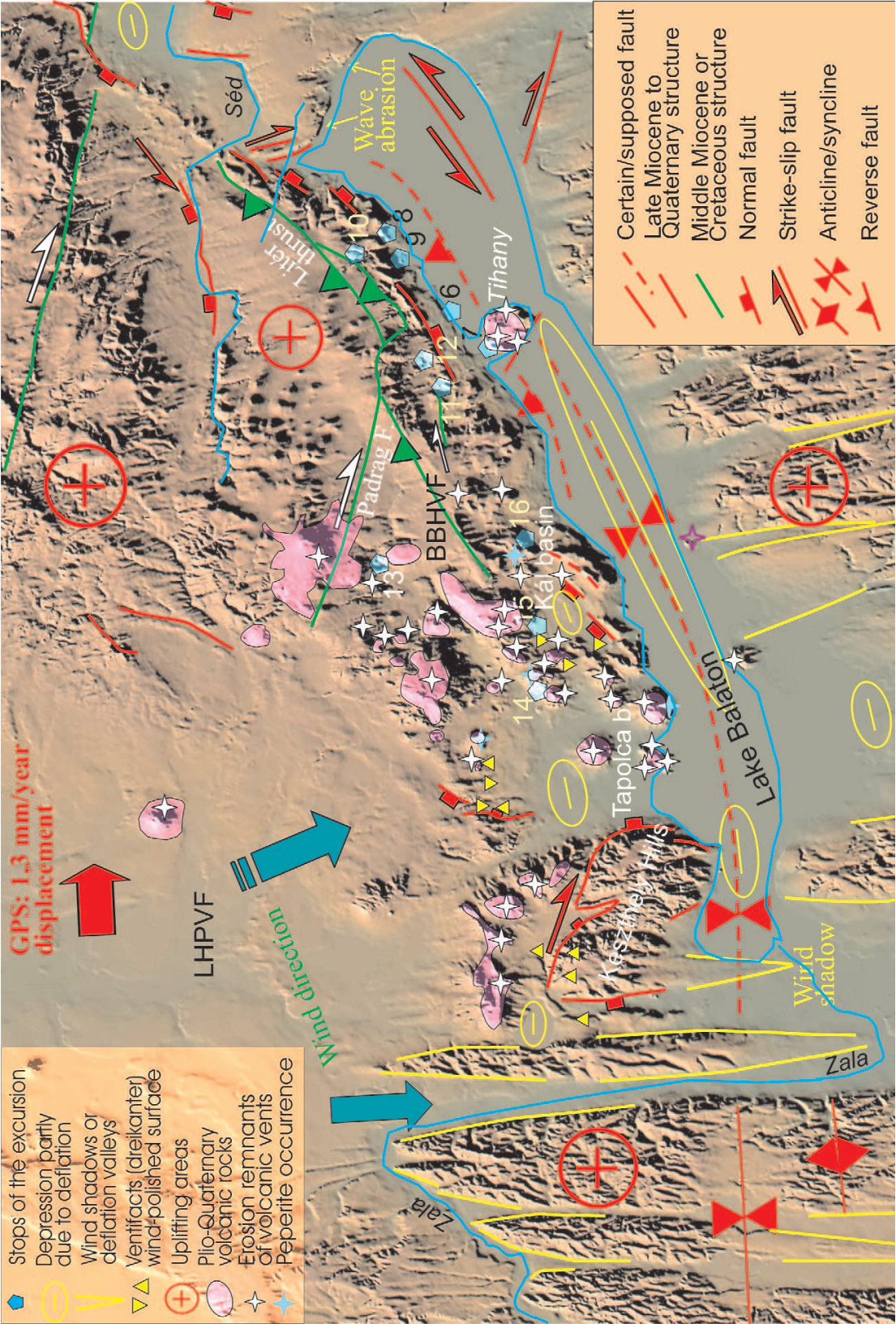


Figure 2. Geological-structural map of the Transdanubian Range (modified from MÁRTON and FODOR, 2003). Note Tertiary fault pattern and excursion stops (pentagons). Stereograms show examples for fault pattern of the 4 major Tertiary deformation phases (lower hemisphere projections, Schmidt net); their locations are marked with green stars. TRL, He, PA, CsZ: TELEGDI ROTH Line, Herend and Padrag faults, Csesznek Zone, respectively, Ajka: Ajka graben





**Figure 3.** Very simplified working model for neotectonic features of the Lake Balaton, Balaton Highland, Tapolca and Káli basins. Note that late Miocene-Quaternary structures may belong to several successive phases with different ages. The sketch does not show all structures but mainly concentrates on those mentioned in the text. Large green arrows indicate main wind directions. Ventifact locations, wind directions are from JÁMBOR (1992, 2002), FODOR et al. (2003). Deflated depressions are from CHOLNOKY (1918) and our own suggestions. Faults from BUDAI et al. (1999), SÁCCHI et al. (1999), VIDA et al. (2001), LOPEZ CARDOZO et al. (2002), BADA et al. (2003), FODOR et al. (2003) and suggestions from the excursion. LHPVF – Little Hungarian Plain Volcanic Field, BBHVF – Bakony-Balaton Highland Volcanic Field. The Digital Elevation Model was downloaded from <http://lazarus.elte.hu/hun/maps/shading/balaton.jpg>.



NE–SW trending strike of the area. It is also important that this deformation event might have contributed to maturation of the hydrocarbon source rocks.

Following minor deformations associated with Albian–Cenomanian and Senonian basin formation, Palaeocene and early Eocene was characterized by stability and denudation (etchplain formation) (KAISER 1997). The birth and subsequent evolution of the middle to late Eocene basins was determined by WNW–ESE to NW–SE oriented compression and perpendicular tension (FODOR et al. 1992 — Figure 2). Similar structures reoccurred in the late early Miocene. The major Neogene rifting phase of the Pannonian basin reactivated most of the inherited structures trending between NW–SE and NE–SW (FODOR et al. 1999). Particularity of the TR is the presence of WNW–ESE trending dextral faults (MÉSZÁROS 1983). These are partly syn-rift faults but also mark a post-rift transpressional event at the end of the middle Miocene approximately at 13–11 Ma (KÓKAY 1976, 1996; MÉSZÁROS 1983; TARI 1991; KISS et al. 2001 — Figure 2). Other interesting features are the N–S to NE–SW oriented normal faults, which occurred during the late Miocene post-rift phase (Figure 2). All these Miocene faults play a major role in the present morphology of the TR while forming steep slopes between footwall Mesozoic carbonates and soft hanging-wall Tertiary clastics.

Starting in the latest Miocene, a considerable amount of basaltic magma erupted in the TR and Balaton Highland area in the Pliocene (BUDAI et al. 1999 — Figures 3, 4). The tectonic role and style of this magmatism is still debated (SZABÓ et al. 1992; PÉCSKAY et al. 1995; HARANGI 2001). A possible interpretation could be that magmatism is somehow connected with the latest stage of post-rift faulting (FODOR et al. 1999). On the other hand, basalt volcanoes occurred after the cessation of post-rift sedimentation and preserve incipient denudation surfaces. In that way, they could belong already to the inversion phase of the Pannonian basin, which was, in general, associated with uplift and denudation (HORVÁTH 1995).

The origin, timing of this neotectonic uplift and related structures are also enigmatic. The present-day stress field, and the focal mechanism of earthquakes are distorted around the TR (GERNER et al. 1999, BADA et al. 1999). Because the maximal horizontal compression is only poorly constrained between NNW to ENE, exact structural characterization, and estimation of societal risk of the present-day tectonics is not yet possible.

One of the most coherent neotectonic data sets comes from the GPS measurements. These suggest that the easternmost Alps and western Hungary move 1.3 mm/year toward east-northeast, while the motion completely disappear toward northeastern Hungary (GRENERCZY et al. 2000). It means that the displacement was accommodated within the TR, the supposed shortening is mainly concentrating between the Lake Balaton and the Danube gorge area, north of Budapest. However, the geometry and kinematics of discrete structures related to the deformation still remain obscure. It is still not known how much of earlier structures, including Cretaceous thrust planes, Miocene strike-slip or normal faults were re-

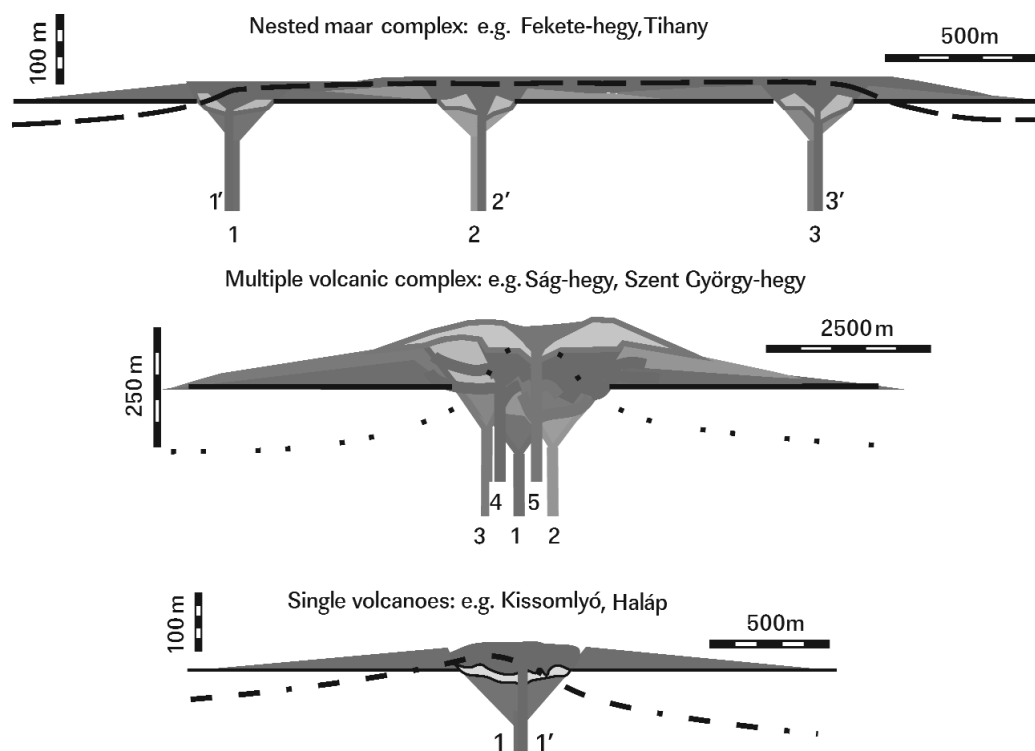


Figure 4. Volcano types of Late Miocene to Pliocene vents in Western Hungary, based on identified volcanoclastic lithofacies distributions of erosion remnants of small volume mafic volcanoes. Numbers represent time sequence of eruptive events

activated. This is the reason that structural analysis, modern measurement techniques (e.g. detailed GPS measurements) would be required to understand neotectonics. Another question is the morphological response to this recent deformation. The general uplift of Transdanubia, deformed terraces of the Danube (PÉCSI 1959), formation of some NE–SW trending depressions, like the Lake Balaton (Figure 3) might be connected to shortening although the exact way of deformation has not unraveled yet.

The questions of inversion, uplift, Plio-Quaternary structures are not merely scientific ones but have serious societal issues. In addition to earthquake hazard, the occurrences of hydrocarbons were ultimately related to these young tectonic events. Part of the hydrocarbon reserves of Transdanubia is stored in inversion structures (PÁVAI VAJNA 1919; DANK 1962; HORVÁTH and RUMPLER 1984). On the other hand, neotectonic uplift could completely change fluid flow in the basin, influence sealing capacity, thus having major implications in research for remaining hydrocarbon traps.

### **Miocene–Pliocene volcanism in the western Pannonian Basin and its relationship to the geomorphological development of the region**

Late Miocene to Pliocene eroded maar complexes, tuff rings as well as strongly modified edifices of scoria cones are preserved in the Pannonian Basin (JUGOVICS 1915, 1916, 1968, 1969, 1971; NÉMETH and MARTIN 1999a,b, 2004a). They belong to the Bakony–Balaton Highland (BBHVF) and the Little Hungarian Plain Volcanic Field (LHPVF) (NÉMETH and MARTIN 1999a — Fig. 5). The bulk of the volcanic rocks of both volcanic fields are considered to have developed subaerially (LÓCZY 1913, 1920; JUGOVICS 1969, 1971; JÁMBOR et al. 1981; BORSY et al. 1986; NÉMETH and MARTIN 1999a). They are commonly erosional remnants of nested maar complexes with effusive intra-crater lava units, well-localized multivert complexes and/or single phreatomagmatic volcanic vents (Figure 4). Base surge and fallout deposits were formed around maars and tuff rings by phreatomagmatic explosions, caused by interactions between 1) water-saturated sediments, 2) surface water (ephemeral lakes, swamps) and/or 3) fracture-stored ground water (karst) and alkali basalt magma (NÉMETH and CSILLAG in: BUDAI et al. 1999; NÉMETH and MARTIN 1999a; MARTIN et al. 2002). Particularly well exposed is a great variety of peperites, (SKILLING et al. 2002) which commonly constitute the root of the volcanic vent (MARTIN and NÉMETH 2000, 2002a, b, c) as a result of non-explosive to very mildly explosive interaction of magma with wet, unconsolidated sediments. There is often a detectable slight time delay between the formation of initially phreatomagmatic landforms and lava effusion on the basis of volcanic textures and interrelations between coherent lavas and volcanoclastic rock units (MARTIN and NÉMETH 2002a, b, c). The composition of the initial volcanic products according to major element geochemistry of volcanic glass shards from pyroclastic rocks is commonly more evolved (tephrite, phono-tephrite) in contrast to more basic composition of subsequent coherent lavas (MARTIN et al. 2003; NÉMETH et al. 2003a). The time difference between phreatomagmatic explosive and effusive eruptions and the bimodality of melt involved in these eruptions may suggest a dual melt involvement in volcanic eruptions in western Hungary (MARTIN et al. 2003; NÉMETH et al. 2003a).

The amount of erosion of the BBHVF has been calculated using different geometrical models of former volcanic landforms (Figure 4) based on the identified volcanic lithofacies-distribution of these volcanic erosional remnants (NÉMETH and MARTIN 1999b; NÉMETH et al. 2001, 2003b, c). Different erosion rate calculations all together gave a relatively uniform result of a few hundred of metres of erosion (NÉMETH et al. 2003b), which numbers are in good concert with other numbers derived from general stratigraphical evidences as well as field relationships of rock units. Newly identified diatreme-filling rocks (NÉMETH et al. 2003c) with sedimentary grains that derived from already eroded Pannonian sedimentary units, are the evidence of a Pannonian sedimentary cover during syn-volcanic time. A phreatomagmatic origin of most of the studied volcanoes in western Hungary is based on the presence of chilled juvenile volcanic glass shards in the pyroclastic units (HEIKEN 1972, 1974; HEIKEN and WOHLTZ 1986). The widespread phreatomagmatism identified in the region, the characteristic north-south lineament of the remnants of the volcanic edifices (JUGOVICS 1937, 1968, 1969; NÉMETH and CSILLAG in: BUDAI et al. 1999) and the presence of other elongated pyroclastic rock units deposited from horizontally moving pyroclastic density currents (NÉMETH and MARTIN 1999c) suggest that these volcanoes erupted in hydrogeologically active zones (e.g. valleys or lowlands — NÉMETH and MARTIN 1999a; MARTIN et al. 2003). The calculated amount of 100–300 m erosion since volcanism ended should be interpreted as a minimum value for the erosion of pre-volcanic, predominantly Pannonian sedimentary units (NÉMETH et al. 2003b). It is also inferred that during volcanism most of the area in Transdanubia still had Pannonian sedimentary cover, excluding few elevated blocks.

### **Evolution of Lake Balaton**

A critical review of works of past researcher (LÓCZY 1913; CHOLNOKY 1918; RÓNAI 1969; MAROSI and SZILÁRD 1981; ERDÉLYI 1983; ZÓLYOMI 1987) and new results of the integrated research of the lacustrine sediments (CSERNY 1993; CSERNY and NAGY-BODOR 2000; TULLNER and CSERNY 2003) allowed us to enhance the accuracy of the geological evolution history of Lake Balaton and its surroundings through the Quaternary (Figure 5).

During the main part of Pleistocene differential uplift of the lake's surroundings continued along pre-existing faults and was accompanied with intense denudation (erosion, deflation). Tectonically preformed depressions and valleys together with deflation depressions formed simultaneously on the surface of blocks uplifting at different speed relative one to another. Denudation could read 150–200 m relative to the original Pannonian sediment surface. Considerable masses of sediments were deposited by seasonal streams active occasionally for more prolonged periods flowing from the valleys of the Balaton Highland. These sediments are exposed along the southern shore. Climatic conditions were presumably rather extreme during the whole Pleistocene. This is confirmed by the coexistence of different formations in the southern shore: such as eluvial red clay indicates warm, fluvial gravel and sand support wet, whereas loess argues for dry and cool climate, respectively. A wide valley with several terraces incised in the solid (Miocene) rock of the lake basement at Zánka and its cross-bedded sediment fill observed on seismo-acoustic profiles refer to a river of high energy. Apart from intense denudated processes neotectonic movements were also characteristic in the area. It can be suggested that as a result of differentiated structural movements and intense denudation the blocks south of the present lake basin were slightly uplifted by the end of Middle Pleistocene (ERDÉLYI 1962). The same process could form depressions in the present site of the lake. The combination of exogenous and endogenous processes changed the drainage pattern by the Würmian period. At the same time, slow subsidence of side basins and denudation of formerly accumulated sediments went on during the long and dry Würmian accompanied by the formation of loess series on the southern shore.

Towards the end of Pleistocene, approximately 15,000–17,000 years BP several, shallow ponds with clean and cold water formed in the site of Lake Balaton (TULLNER and CSERNY 2003, Figure 5). Inundation followed progressively from west to east. Warming climate and increasing precipitation brought about the rise of water level. Moreover, abrasion progressively destroyed the dams separating the ponds and a uniform lake was formed. Later, the water depth changed frequently as a function of temperature, precipitation (and sometimes probably human intervention). Prolonged high water level is proven by terraces in the surroundings of the lake extending between 104.1 m and 112.5 m above Adriatic sea level. Disregarding the artificial interventions before 1863 — if there were any — we can say that under natural conditions the lake level changed during the Holocene period between 103.0 and 108.0 m above Adriatic sea level (TULLNER and CSERNY 2003).

### Short description of excursion stops

#### *Stops in the Vértes Hills*

The eastern margin of the Vértes Hills represents one of the most spectacular morphological scarps of the TR. The scarp is consisted of N–S, NNE–SSW and NW–SE to WNW–ESE trending segments. All morphological scarps correspond to Miocene or younger faults (CSILLAG et al. 2002a; FODOR et al. 2002 — Figures 6, 7). The kinematics can be estimated using some outcrop-scale fault-slip data suggesting E–W tension and locally N–S compression (normal and strike-slip type deformation, respectively).

Some of the fault segments could be initiated in the late middle Miocene (Sarmatian, 11–13 Ma), as reconstructed on the basis of palaeogeographic maps and borehole data. Sedimentary dykes demonstrate ongoing late Miocene fault activity (CSILLAG et al. 2002a). On the other hand, mapping of the Geological Institute (Zsolt Peregi, Gábor Csillag, László Fodor) revealed the presence of shoreface gravels and breccias on the slopes (Figure 8). While the gravels, conglomerates occur at the mid-slope, breccias occupy the higher part of the slope. Their transition is gradual. This geometry sug-

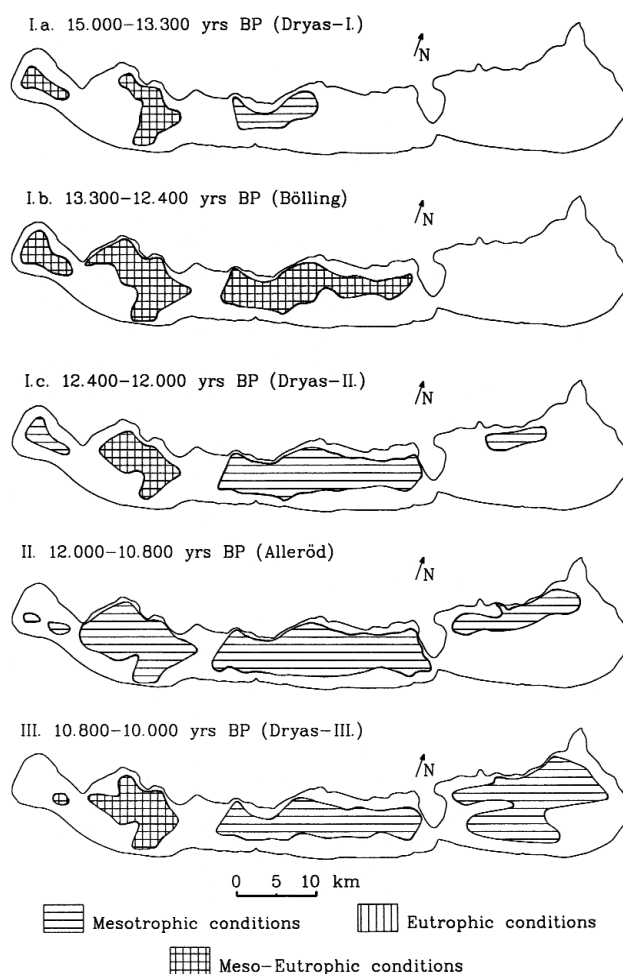
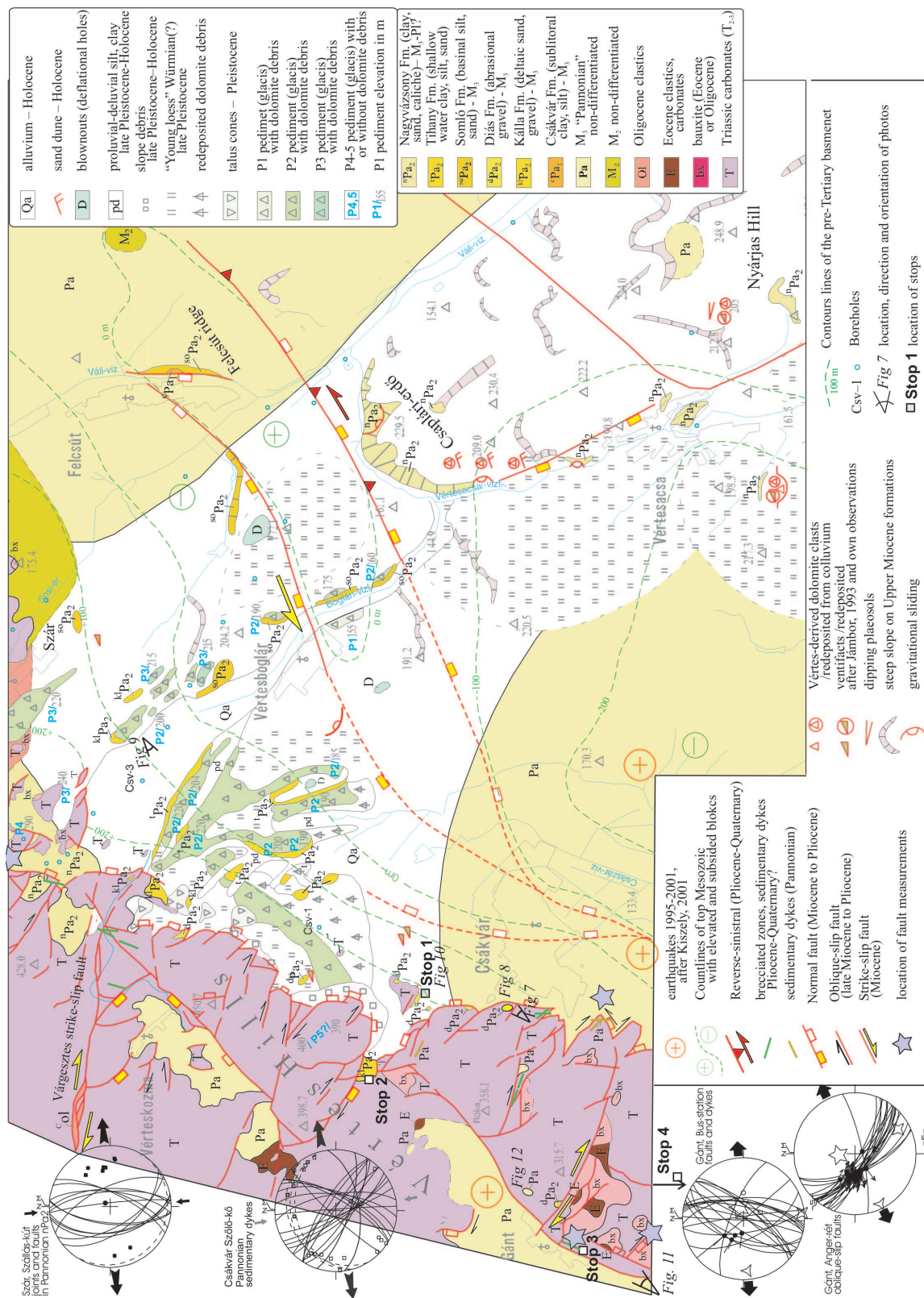


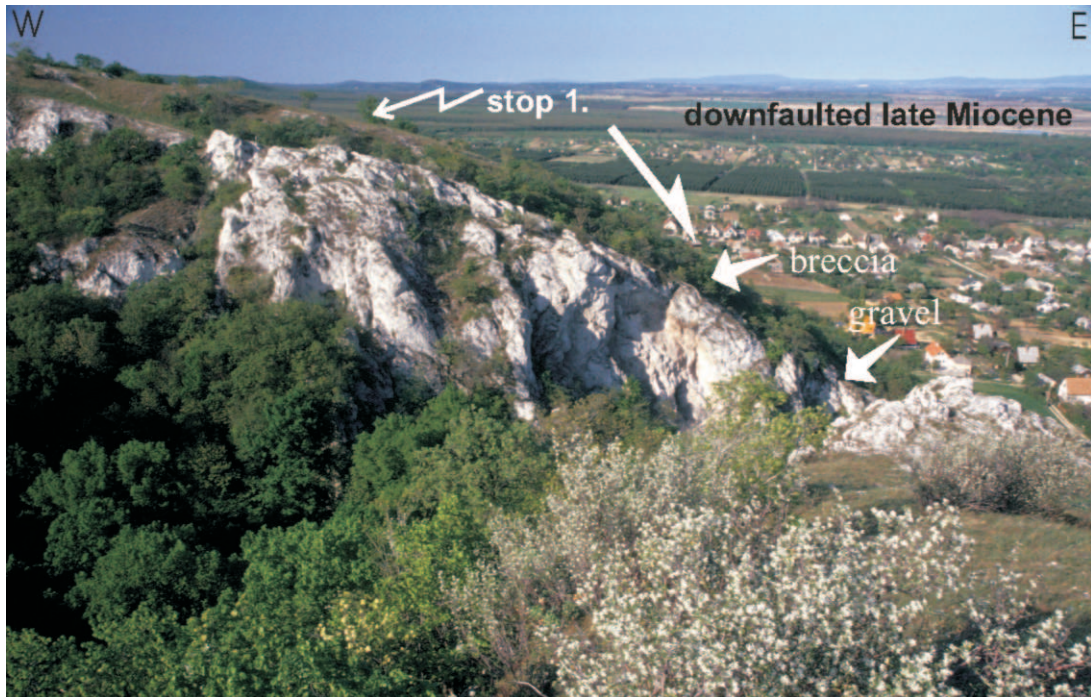
Figure 5. Evolution of the Lake Balaton (after CSERNY and NAGY-BODOR 2000; TULLNER and CSERNY 2003)





**Figure 6.** Simplified structural-geomorphological map of the eastern Vértess Hills and its foreland (after CSILLAG et al. 2001, 2002a; FODOR et al. 2002). The Várgesztes strike-slip fault is after GYALOG (1992), modified. Note northwestern flow direction of the Vértessacs-víz, probably due to uplift of the Nyárjas Hill. This uplift might have separated the Vértess-derived dolomite clasts of the Nyárjas from the source area (NE from Vértesszoma). The eastern fault system of the Vértess affected the hanging wall late Miocene sediments. Earthquakes after KISZELYI (2001). Quaternary formations and landforms are only marked in the central part of the foreland.





**Figure 7.** View to the eastern morphological scarp of the Vértes Hills at Csákvár. The scarps are due to late Miocene to Quaternary(?) faulting with mainly normal kinematics. Note position of late Miocene scarp breccia (Figure 6.) and abrasional gravel (conglomerate). Stop 1. is located in the background, behind the visible scarp

gests that the late Miocene Lake Pannon reached a region, which was characterised by tectonically controlled palaeoslopes, where rocky shores developed (CSILLAG et al. 2002a).

The upper Miocene sequence starts with gravel covered by silt, clay, marl, which change to more sandy sediments. The youngest sediments consist of clays, silts with caliche horizons, sand, and breccias at the marginal positions (TÓTH 1971; CSILLAG et al. 2002b). The age of this formation can be of latest Miocene or early Pliocene.

Mapping clearly demonstrated that the late Miocene sediments were equally affected by faulting. The maximal vertical displacement could be around 400 m. The age of deformation could be Pliocene and/or Quaternary. Small magnitude historical and instrumentally registered earthquakes indicate ongoing tectonic deformation (KISZELY 2001). However, their connections to mapped faults have not been determined yet, thus a definite statement about Quaternary activity of Miocene boundary faults cannot be made.

Southeast of the major fault scarps, mapping and geomorphological observations revealed the Plio-Quaternary landscape evolution (CSILLAG et al. 2002a). Gently SE dipping, extensive pediment surfaces erode the late Miocene to Pliocene sequence (Figures 6, 9). These are covered by dolomite colluvium, originated from the scarps by surface wash. More linear fluvial erosion resulted in a system of parallel incised valleys generally flowing SE. The activity of wind is marked by numerous ventifacts, wind holes, wind-blown sand and loess indicating that deflation could take role in formation of parallel valley system.

Quaternary fault activity can be suspected on the base of deformed landscape and diverted drainage in the south-eastern foreland of the Vértes Hills. The oldest and highest pediment surface (in the Nyárjas Hill, Csaplári-erdő area) could be uplifted and tilted, resulting in unusually high position of scarp-derived colluvial material ca. 15 km away from the scarp



**Figure 8.** Late Miocene breccia gradually passing downward to shoreface gravel. Note large dolomite clast in breccia matrix. West of Csákvár, for location see Figures 6, 7. Note slightly rounded dolomite clast below hammer





**Figure 9.** Pediment surface eroding late Miocene sediments in the eastern foreland of the Vértes (after CSILLAG et al. 2002a). The photo was taken from the same pediment as found in the center of the photo. The pediment is covered by ~1 m thick dolomite colluvium. Background shows the eastern fault scarp of the Vértes Hills

(CSILLAG et al. 2002a — Figure 6). Probably due to this uplift, an intra-valley drainage divide was formed in a southeast trending valley west from the Nyárjas Hill (Vértesacsai-víz). The creek is now flowing to northwest instead of general southeastward flow direction. The uplift of the Nyárjas block could probably be related to sinistral-transpressional reactivation of a NE–SW trending fault running at the northern tip of the Csaplári-erdő.

#### Stop 1. Csákvár

The first outcrop is located 1.5 km northwest of Csákvár along the road to Oroszlány, on the southern slope of the Bagó-kő, 150 m from the road. The outcrop is located on the hanging-wall of the main range-bounding fault system. The exposed sequence possibly belongs to the latest Miocene – early Pliocene sedimentary unit although its Quaternary classification is also possible. The moderately dipping poorly cemented breccia shows divers signs of early diagenesis and/or Quaternary palaeosol formation (Figure 10). The dip (~10–15°) could be considered as original. Updip the beds could be projected to the top of the Bagó-kő (~200 m). This hill consists of Triassic dolomite, the possible source rock of the clasts of the Plio-Quaternary beds. The valley between the outcrop and the Bagó-kő was cut by a more recent denudation event.

The small hillock offers a good view of the eastern scarp of the Vértes Hills in the vicinity of Csákvár (Figures 6, 7). The change in dip degree of the scarp corresponds broadly to change from terrestrial to lake sediments (breccia to abrasional gravel) formed on tectonically controlled palaeoslope (Figures 7, 8).



**Figure 10.** Pliocene or Quaternary talus breccia on the southern slope of the Bagó-kő, northwest from Csákvár. Note irregular upper surface of the Triassic dolomite and the cemented breccia layer above loose angular clasts. View is toward SSE, toward the main scarp



### Stop 2. Csákvár, road cut

The second outcrop is located along the road Csákvár–Oroszlány, ~500 m south of the junction to Gánt and was mapped by Zsolt Peregi (Figure 6). It shows the alternation of a clay and breccia of late Miocene to Pliocene age. The occurrence of beds with angular dolomite clasts could indicate tectonic activity during the deposition although a simple sedimentological explanation (fan-delta) is also possible. The outcrop is close to a WNW–ESE trending scarp, which corresponds to a dextral-normal fault with ~150 m of vertical displacement. The displacement might have occurred during the latest Miocene or in the Plio-Quaternary, in connection with the motions along the major range-bounding fault system to the east.

### Stop 3. Gánt, bus stop at the southern end of the village

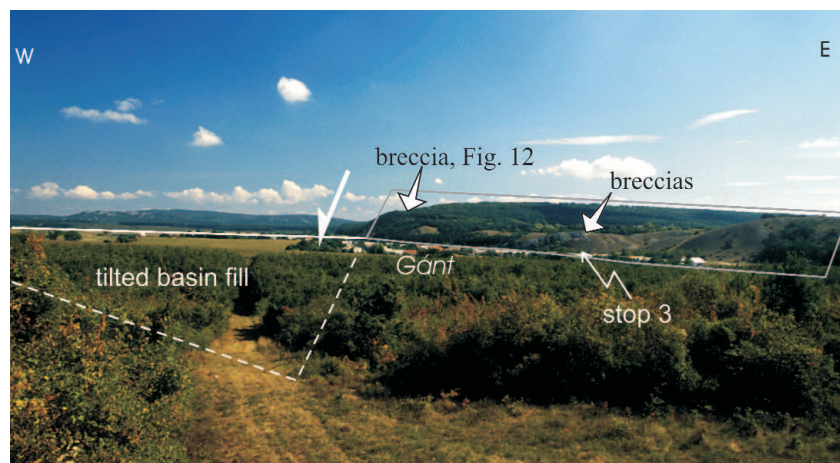
The field trip continues along the NNE–SSW trending normal fault, which bounds the Gánt half-graben to the east. The gently dipping slopes under the meadows and houses of Gánt correspond to east-tilted upper Miocene (Pannonian) strata penetrated by boreholes (Figures 6, 11) (FODOR et al. 2002). At the graben margin, along the boundary fault, peculiar dolomite breccias can be studied near the southernmost bus stop of the village (CSILLAG, unpublished mapping result). These breccias either occur as sub-vertical bodies (“dykes”), or cover the surface of the dolomite (Figure 12). They may represent fault breccias or were deposited in terrestrial settings on moderate palaeoslopes. In any case, these breccias seem to be connected to the development of the Gánt half-graben. Their latest Miocene and/or Pliocene age and exact origin needs further research. Few fault slip data and other brittle structures suggest E–W tension, which affected the host dolomite (Figure 6).

### Stop 4. Gánt, Bagoly Hill

The classical Gánt Bagoly Hill bauxite quarry is found between the village and the Csákvár–Csákkerény road and shelters a mining museum. This outcrop offers excellent opportunity to study the Eocene and Miocene fault pattern and to see in detail features occurring on fault planes.

Bauxite was accumulated in shallow, relatively wide dolines having been formed on Upper Triassic dolomite. Pelitomorphic and pizolitic bauxite layers alternate while cross bedding and gradation are also characteristic (MINDSZENTY et al. 1989, 1997). The bauxite is covered by brackish then marine shallow water sequence of clay, silt, marl and limestone (BIGNOT et al., 1985, MINDSZENTY et al. 1989).

On the northern quarry wall, below the mining museum, an E–W trending fault zone represents the northern margin of the pit. The fault zone is composed of en echelon segments, which are connected to relay ramps (Fig. 13). Kinematics of the faults is normal-dextral in each case (ALMÁSI 1993; ALMÁSI and FODOR 1995; MINDSZENTY et al. 1997). Small connecting fault splays between major fault segments have gentle dip, and show dip slip. They probably cut across the already strongly fractured relay ramps (ALMÁSI and FODOR 1995).



**Figure 11.** Panoramic view of the Gánt depression, looking from south, staying south of Gánt. Note the eastward dipping half graben, which is bound by normal faults running just east from Gánt. The late Miocene to Pliocene(?) basin fill argues for Pliocene to Quaternary faulting. Breccia bodies and dykes(?) follow – and are related(?) to – the fault scarp



**Figure 12.** Breccia body covering the Triassic dolomite on the western slope of the Gánt Hill, facing the Gánt half-graben. The breccia could be formed in connection with boundary faulting and/or erosion along the fault scarp. The age of the sediment could be late Miocene to Quaternary



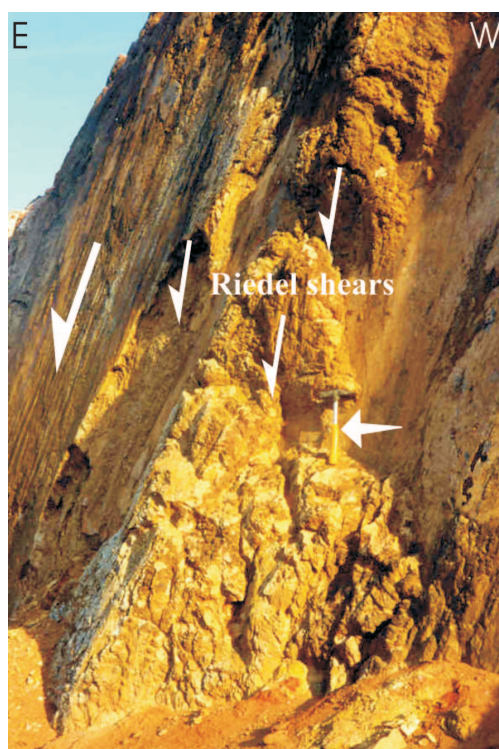


**Figure 13.** En echelon normal-dextral faults at the northern boundary of Gánt, Bagoly Hill bauxite pit, looking to the north. The en echelon faults are connected by strongly fractured relay ramps (after ALMÁSI 1993; ALMÁSI and FODOR 1995)

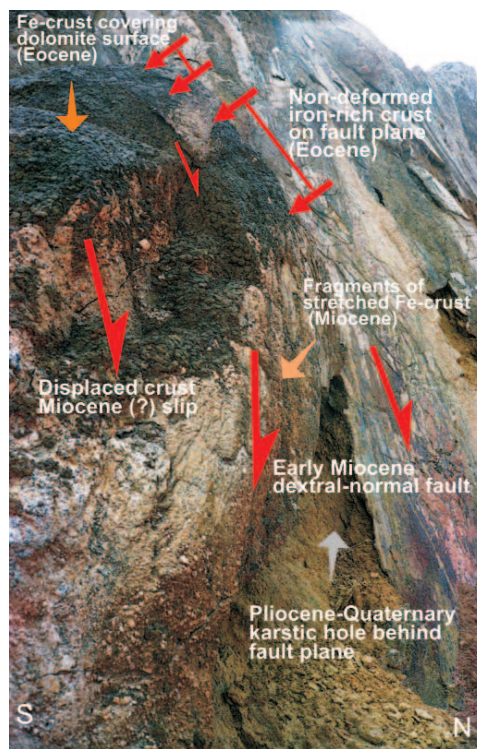
have existed already before the encrustation (Fig. 15). Consequently, fault activity started before the marine Eocene transgression, during or close to the deposition of the bauxite (MINDSZENTY and FODOR 2002). The earliest fault motions could also induce the redeposition of a former palaeosol, which once covered the whole denudated area. The

A parallel fault occurs at the southern half of the pit (Figure 14). Here the fault plane is continuous but curved. Pitches of oblique slickensides change gradually corresponding to changing fault strike. Riedel shears, dragged clasts, large grooves with clasts at their end can be seen along the main fault plane (Figure 14). All features suggest dextral-normal slip, probably during the early Miocene.

Another particularity of the bauxite quarry is the deformation of the Eocene bauxite and related ferrigenous crust. The ferrigenous crust between the bauxite and overlying dolomite was formed during the initial Eocene marine transgression, just after the deposition of the bauxite (GERMÁN-HEINS 1994). This crust covers the higher parts of the southern major fault plane, which should



**Figure 14.** Secondary features along the southern boundary fault of the Bagoly Hill mine. Note Riedel shears, displaced fragments of the Fe-crust, etc. (MINDSZENTY and FODOR, 2002). Hammer for scale



**Figure 15.** The southern main fault of the Bagoly Hill mine. The fault displaces Eocene cover sequence (hanging wall) against Eocene bauxite and Triassic dolomite (footwall). The upper part of the fault plane is covered by ferrigenous crust, which are related to the Eocene marine transgression (GERMÁN-HEINS, 1994). This crust post-dates the first motion of the fault, probably coeval with the bauxite formation (MINDSZENTY and FODOR, 2002)



resulting bauxite shows sedimentological features related to deposition on alluvial fans by different gravity flows (MINDSZENTY et al. 1989). Gently dipping, glittering ferrigenous sliding planes were probably originated from gravity sliding just after deposition and, together with slump folds, indicate unstable tectonic conditions. Small synsedimentary faults, pelitomorph bauxite dykes are the direct evidence of faulting (MINDSZENTY et al. 1989).

Although the faults of the Bagoly Hill quarry seem to have Eocene to Miocene age, other faults of the Gánt area might have Plio-Quaternary activity. Particularly, the NW–SE trending boundary fault system of the Haraszts and Újfeltárás displaces the NNE-trending boundary fault of the Gánt half-graben (Figure 6). Thus, this fault branches have at least early Pliocene age.

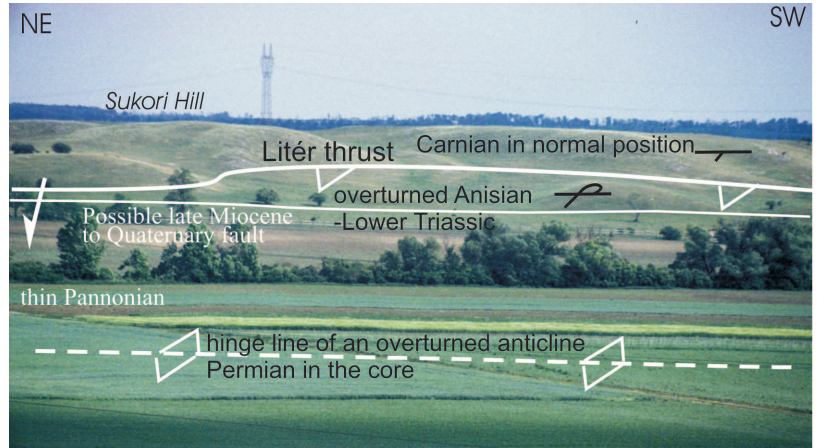


Figure 16. Panoramic view to the Sukori Hill. In the foreground, the morphological scarp corresponds to young Miocene normal fault (?), while the Litér thrust occur on the hill

### Stops in the Balaton Highland

#### Stop 5. Sólly, southern slope of the Ór Hill – view toward the Litér thrust

The viewpoint is located east of Sólly, 200 m east from the railway station, near the protected key section of the Lower Triassic Csopak Marl Formation. The point offers a panoramic view on the Litér thrust, a major fault of the Balaton Highland area (definition by BÖCKH 1872, 1874). This NE–SW trending thrust occurs on the southeastern limb of the major syncline of the TR (TARI 1994, BUDAI et al. 1997, BUDAI et al. 1999). The largest throw is between Silurian in the hanging wall and Upper Triassic in the footwall.

Looking from the viewpoint, Upper Permian to Lower Triassic rocks occur below thin Pannonian under the agricultural fields at the foreground. We suggest that this area represents the hanging wall of late Miocene to Quaternary(?) normal fault located close to the north-western slope of the Sukori Hill (Figure 16). The pre-Pannonian rocks form the core of an overturned anticline (Figure 16). In the background, on the northwestern slope of the Sukori Hill, overturned uppermost Lower Triassic marls and the Lower Anisian Aszófő Dolomite and Iszkahegy Limestone thrust over normal succession of Upper Triassic (lower Carnian) dolomites (Budaörs Fm) (Figure 16). Geophysical data indicate that the overturned beds form a thin flake over the normally dipping younger Triassic formations (Figure 17). Our interpreted section suggests that the thin overturned flake is the continuation of the overturned anticline (Figure 17), which before normal faulting, was at the same topographic level as the Sukori Hill. The formation of the overturned anticline is connected to the reverse displacement of the Litér thrust. However, the exact fault geometry is not very clear at this location. It is possible that the overturned anticline is detached from the Silurian-Devonian base along a late, sub-horizontal detachment (white arrows on Figure 17) or follow a moderately dipping thrust ramp, located just below the slope of the Sukori Hill (black arrows on Figure 17).

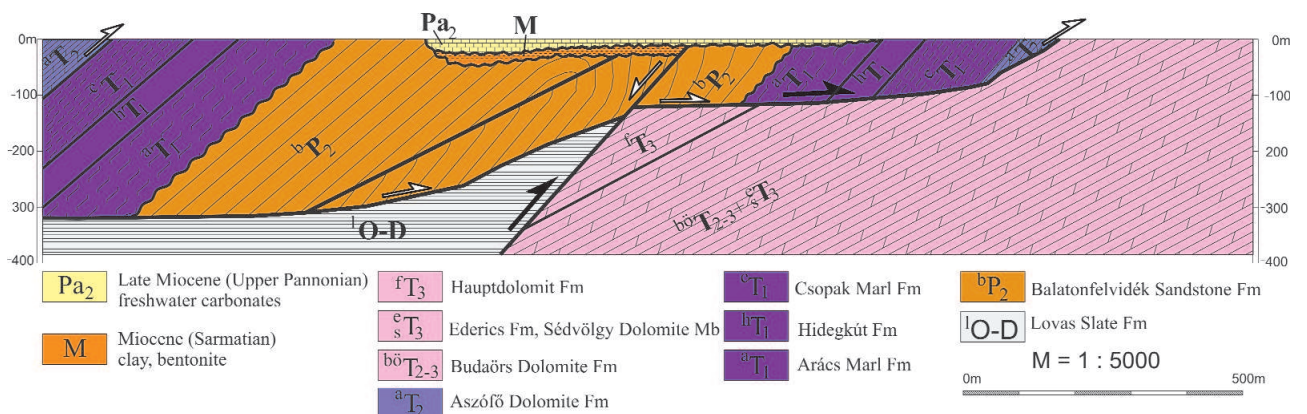


Figure 17. Cross section at Sólly, across the Litér thrust, after BUDAI et al. (1997), modified. Black and white arrows show the suggested possible position of the Litér thrust

### Stop 6. Balatonfüred, beach Limnogeological research of Lake Balaton

One of the objectives of limnogeological research of Lake Balaton was to investigate the effectiveness of thin-layer dredging for the removal of the silt blanket because of environmental purposes, as well as the underwater movement of sediments (creeping, agitation and re-suspension). Apart from the already well-established laboratory methods (sedimentology, soil mechanics, geochemistry) we studied the distribution of radioactive isotopes (caesium, lead, potassium, etc.) in the upper section of the calcareous silt (CSERNY et al. 1995) in order to determine the intensity of silting up and redistribution of sediments. During this investigation we revealed among others  $^{134}\text{Cs}$  and  $^{137}\text{Cs}$  isotope contamination in the 0,5 m uppermost part of the sediments that can be traced only from 1951 onward since atmospheric nuclear tests took place. Two distinct maximums can usually be observed in the layers, namely the maximums of the year preceding the 1964 Nuclear Test Ban Treaty and the 1986 Chernobyl nuclear accident. They facilitate to assess how fast the silting up proceeds. Their absence indicates underwater erosion, whereas balanced averaged values refer to underwater agitation and subsequent accumulation. Under calm hydrological circumstances the speed of mud accumulation amounts to 14 mm/year during the last 40 years in the central part of the Keszthely embayment. At the same time, the average silting up of the major part of the lake is 5 mm/year (these values are related to loose, unconsolidated sediments of high water content).

Another research theme focused on environmental-geochemical investigations of the Zala river–Kis-Balaton–Keszthely embayment and aimed at assessing the filtering effect of the Kis-Balaton. In order to answer this problem we studied the trace element and heavy metal content of mud, suspended sediments and water taken from the related hydrological system. Apart from the parameters determined in the field (pH, Eh, alkalinity, KOI) we determined the concentration of major elements (Na, K, Ca, Mg, Fe,  $\text{PO}_4\text{-P}$ ,  $\text{NO}_3\text{-N}$ , Cl), as well as 16 trace elements (V, Cr, Mn, C, Ni, Cu, Zn, As, Rb, Sr, Mo, Cd, Sb, Cs, Ba, Pb) and studied the interaction between water and sediment. We arrived at the conclusion that the water system was not contaminated by metals, the measured concentrations were in the same order as in the oceans, except for lead, zinc, cadmium and copper. The latter anomalies are due to atmospheric contamination. The amount of the measured elements is the result of quite complex — biological, hydrological and chemical — processes (ELBAZ-POULICHET et al. 1997a, b). As a result of these complex processes, the water–suspended sediment system of the Kis-Balaton has significant filtering effect.

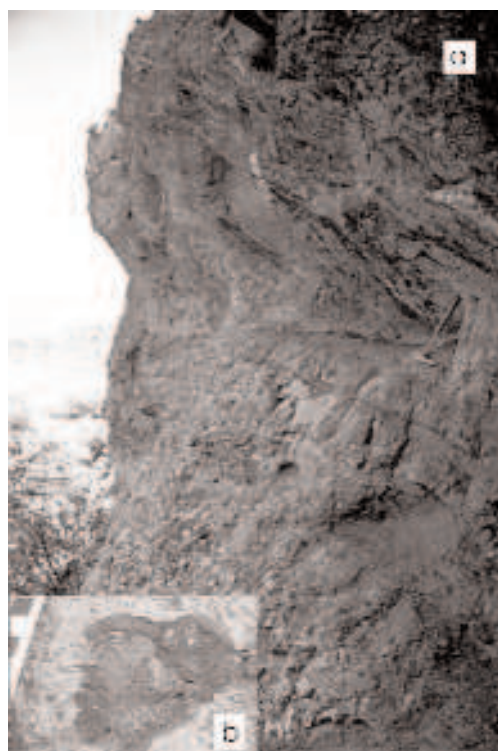
A third study is focused on investigating the relationship between the amount of phosphorous in lacustrine sediments and the eutrophic state of water. The existing relationship between the amount and nature of phosphorous in lake sediments, as well as proliferation of algae and the eutrophic state of the lake are already well known. The summarised phosphorous section of drilling profiles in the four sub-basins of the Lake Balaton shows that the total phosphorous content values similar to the actual situation existed already in the sediments deposited both before and after the lake's formation. Since phosphorous is not mobile in the sediments, the approximately 600 g/l average value in certain layers is due to its syn-genetic origin. In consequence, it can be suggested that phosphorous was transported from the geological formations of background areas. Enrichment observed in some parts of the profile can be explained by diminishing water volume (due to small precipitation, high temperature and intense evaporation) suggesting simultaneous proliferation of vegetation (CSERNY and NAGY-BODOR 2000).

Participants of the field trip could examine samples taken from the Keszthely embayment where dredging of the upper part of the mud has taken place since 1992. The upper 20–30 cm of the sedimentary pile is a darker mud containing a great amount of phosphorous and spores, which induced an algal bloom. That is why the major task of water management activities in the Lake Balaton area is to reduce the amount of phosphorous in the water and the uppermost sediments. One of the solutions is the reestablishment of the original conditions in Kis-Balaton (swamps on the major tributary of the Lake, the Zala river) because the swamps function as natural filters of phosphorous. The other solution is the dredging of the uppermost section of the lake sediments, carrying a great amount of phosphorous and algal spores.

### Stop 7. Tihany, Kiserdő-tető

At the Kiserdő-tető in a continuous succession of massive-to-weakly bedded tuff breccia and lapilli tuff (diatreme-facies) crops out and is overlain by alternating coarse-fine-coarse lapilli tuff units (Figure 18). The pyroclastic rocks of this later succession has volcanoclastic texture characteristic for reworking, presumably in a water-laid environment (NÉMETH 2001, NÉMETH et al. 2001b). The basal volcanoclastic units are chaotic, ill-sorted, matrix supported tuff breccias and lapilli tuff, rich in accidental lithic clasts, cauliflower bombs and larger chunks of irregularly ordered lapilli tuffs inferred to be recycled clasts from earlier eruptions. The textural characteristics of this unit are interpreted to be the result of magma and water or water-rich slurry interaction leading to phreatomagmatic explosive eruption. The basal unit is overlain by a reworked, bedded, alternating unit, preserved in the highest topographic position among all the volcanoclastic successions in Tihany (Figure 19).

Reworked volcanoclastic beds of the Kiserdő-tető, interpreted as foresets of delta fronts built into maar lake(s) of Tihany, are generally 5–25 cm thick, and dip steeply (20–30°) towards the former crater position (Figure 18), forming a concave geometry controlled by the steep inner morphology of the former maar basin (NÉMETH 2001). The total thickness of the



**Figure 18.** Contact between the lower diatreme facies pyroclastic rocks (lower unit) and the alternating coarse-fine pyroclastic beds of the maar lake turbidite and debris flow deposits (after BUDAI et al. 2002)

foreset sequences is at least 40 m. The individual beds usually have irregular tops and bases. Coarse-grained beds are usually laterally discontinuous and they grade into fine-grained cross beds. Scoriaceous fragment-rich scour-fill structures are common. Locally, carbonate (calcite) enrichment is also common, in several cases forming faint bedding. The large scoriaceous clasts in several places contain a large amount of probably secondary carbonates (calcite) and the clasts are carbonate-cemented (NÉMETH 2001). The common carbonate cementation may have been the result of high carbonate content of the maar-lake. The carbonate content of these beds could also be explained by substantial hot spring water entering into the maar-lakes.

Coarse-grained beds show characteristic inverse-to-normal grading (NÉMETH 2001). Between the coarse-grained foreset beds there are a few fine-grained primary volcanoclastic beds, usually with fallout characters (Figure 19). They usually contain numerous accretionary lapilli, with a maximum diameter of 1 cm. These interbedded primary volcanoclastic beds suggest ongoing volcanic eruptions adjacent to the earlier-developed maar basins coinciding with clastic sedimentation (NÉMETH 2001).

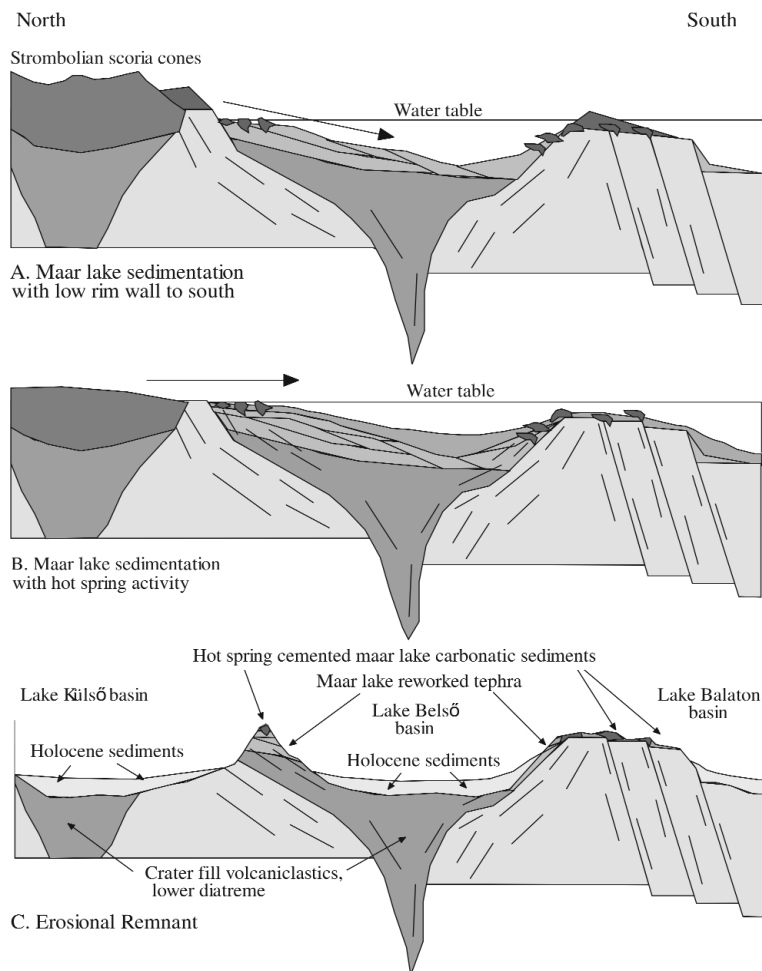
#### Stop 8. Káptalanfüred, Köcsi-tó quarry

The Alpine sedimentary cycle of the TR started with the deposition of the Upper Permian *Balatonfelvidék Sandstone Formation*. It has been deposited in several cycles; 1st: fluvial, 2nd: continental sabhka facies. Due to the synclinal structure of the Bakony Mts the sandstone can be found in subsurface in the Northern Bakony, too (MAJOROS 1980).

The Köcsi-tó quarry exposes sandstone, conglomerate and pebbly sandstone (Fig. 20). The lowermost conglomerate layers are covered by a pebbly sandstone unit with large-scale trough cross bedding, indicating transport from the NE (BUDAI et al. 1999, 2002). Pebbles are quartzites, red dacites or foliated sandstones cemented by siliceous-kaolinitic matrix. The sediments were deposited in fluvial environment.

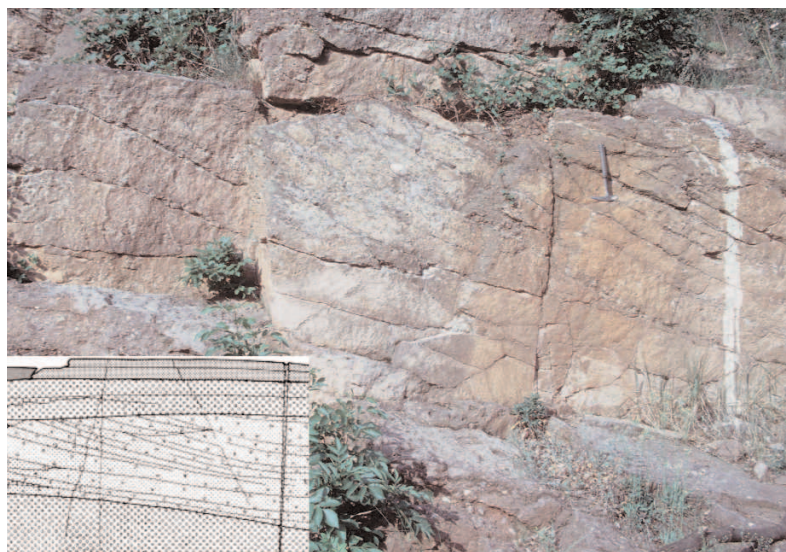
#### Stop 9. Balatonarács, railway cut

The classical outcrop (LÓCZY 1913, HAAS 1985, HAAS et al. 1988, BUDAI et al. 1999, 2002) shows the boundary between the Upper Permian Balatonfelvidék Sandstone Formation and the Lower Triassic light gray dolomite (Figure 21). There could be a small hiatus near the Permian-Triassic boundary, but it has not been definitely proven yet.



**Figure 19.** Evolutionary model and depositional system for the Tihany Maar Volcanic Complex (after NÉMETH et al. 1999)

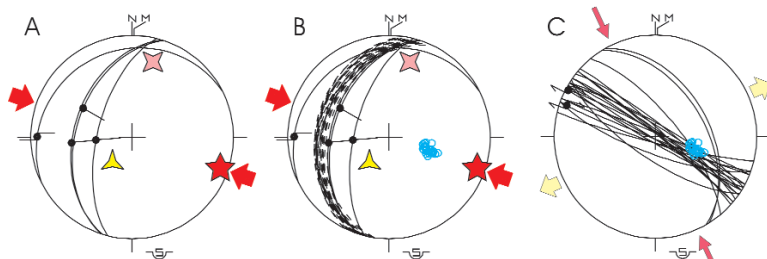




**Figure 20.** Cross-bedded sandstone and conglomerate layers of the Balatonfelvidék Sandstone Formation in the Köcsi-tó quarry, Balatonalmádi-Káptalanfüred after BUDAI et al. (2002). Inset shows interpreted drawing of the whole outcrop



**Figure 21.** View to the railway cut of Balatonarács. The boundary surface between the red Permian sandstone-siltstone and the light grey Lower Triassic Nádaskút Dolomite is poorly visible. Note small reverse fault somewhat steeper than bedding in Permian beds, at the middle part of the photo



**Figure 22.** Stereograms from Balatonarács railway cut, presenting unpublished data of A. DUDKO and L. FODOR. A) Shows small reverse faults formed due to WNW-ESE compression (large arrows); B) Reverse faults are parallel or slightly steeper than bedding (dashed lines); C) Faults formed after the tilting, probably during Miocene; stress axes are only estimated. Schmidt net, lower hemisphere projection. Dashed curves are projections of bedding planes, circles for poles of bedding (after FODOR 1997)

The upper part of the Permian Balatonfelvidék Formation is built up of red sandstone and siltstone of fluvatile origin. In the outcrop, the sandstone beds are overlain by alternating sand- and siltstones, characterized by parallel and low angle cross lamination with ripple marks and bioturbation. The overlying Lower Triassic transgressive sequence begins with the Nádaskút Dolomite underlying the Arács Marl Formation (Figure 21). The depositional environment of these beds was probably a shallow subtidal lagoon of low energy. Triassic sediments were deposited on a wide ramp, preceding the Anisian rifting of the area.

Within the Permian siltstone, reverse faults could be detected (observation of Dudko A. and Fodor L.). They were formed by WNW–ESE compression (Figures 21, 22). Faults are parallel to the bedding planes or slightly steeper, and could form at sub-horizontal or gently dipping bed position. Formation of faults could represent an early stage of deformation, which can be related to the formation of the TR syncline and the Litér thrust. Some sub-vertical faults could be Miocene in age.

#### Stop 10.

##### Felsőörs, Malom-völgy

This well-protected outcrop is a geological conservation area of the Balaton Highland National Park, found in the northern part of Felsőörs village. This section is one of the most spectacular outcrops of the Middle Triassic in the Balaton Highland.

The lowermost part of the section is built up of bedded *Megyehgy Dolomite* which is the youngest formation of the Lower Anisian carbonate ramp. The transition with the overlying pelagic basinal sequence is characterized by bituminous marly dolomite. The lower part of the *Felsőörs Limestone* is built up of nodular cherty limestone, overlain by marly flaser-bedded brachiopod- and crinoid-bearing limestone. The upper part of the formation is characterized by even-bedded dark-gray limestone with rich Upper Anisian ammonite-assemblage. The overlying greenish volcanite with limestone intercalations belongs to the *Vászoly Formation*. The Anisian/Ladinian boundary can be fixed within this formation, at the base of the Reitziites reitzi Zone (VÖRÖS 1993, 1998) (Figure 23). New U/Pb ages suggest 241.5 Ma for the base of Ladinian (PÁLFY et al. 2003). Cherty nodular limestone above the



volcanic sequence belongs to the *Buchenstein Formation*.

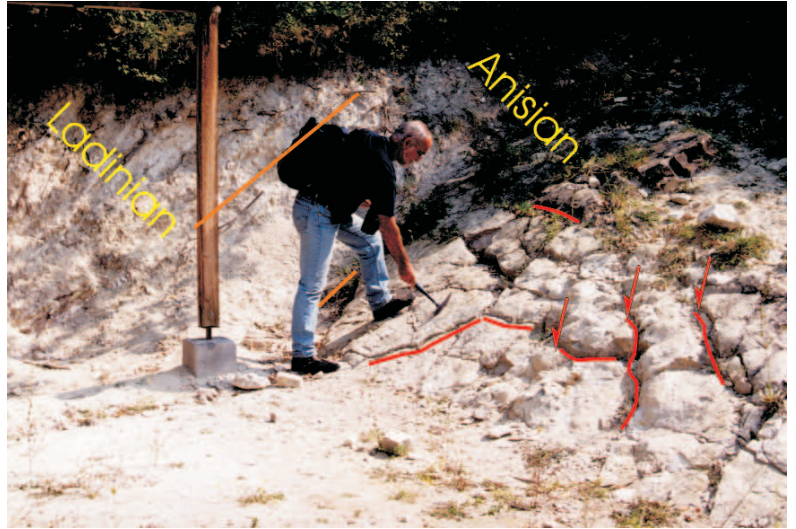
The late Anisian–Ladinian basins probably represent half grabens (BUDAI and VÖRÖS 1992) related to the rifting of this part of the Tethys Ocean at the beginning of the Pelsonian. Small-scale faulting was observed in the section during the excursion (Figure 23). The faults are smooth, curved, and do not seem to cut overlying layers. Altogether, the structures seem to have affected soft layers before complete diagenesis and could be connected to syn-sedimentary tectonic activity.

#### Stop 11.

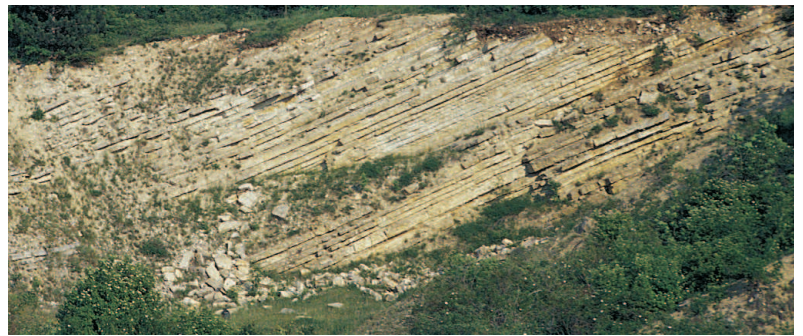
##### Pécsely, Meggy-hegy

The outcrop is located south of Pécsely village, east of the main road, on a wide ridge. It contains the well-bedded, lower Carnian *Füred Limestone*, a formation consisted of light grey, nodular cherty limestone beds and thin intercalated marl layers. This cyclic basin succession showing a shallowing upward trend represents the progradation of the neighbouring carbonate platforms during the earliest Carnian (BUDAI and HAAS 1997).

The eastern quarry wall exposes moderately dipping limestone beds (Figure 24). They represent the northern limb of an anticline, trending E–W. Some dextral slickensides occur on fault planes, which are almost perpendicular to beds and fold axis (Figure 25). The slickensides are parallel to the fault plane–bedding plane intersection line. These small structures either were formed at horizontal bed position before folding or reflect guided slip along bedding plane during the folding (tilting). In the former case, the NNE–SSW compression characterizing the dextral faults could pre-date the folding and may associate to some NW–SE trending ‘cross-structures’ present in the Balaton Highland (DUDKO in BUDAI et al. 1999). The formation of the anticline itself could be connected to the same NNE–SSW compression. Alternatively, the folds just represent dragging along E–W trending dextral faults frequently cross-cutting the quarry. The outcrop-scale dextral faults are part of a longer zone, which can be followed for long distances (BUDAI et al. 1999, BUDAI et al. 2002). Conjugate strike-slip faults suggest NW–SE compression for the faulting (Figure 26). In that scenario, the observed folds may represent structures parallel with the shear zone; this geometry



**Figure 23.** Boundary layers of the Anisian and Ladinian stages at the Felsőörs section (sensu VÖRÖS 1993). Wavy or irregular fractures (red lines and arrows) cutting bedding planes could be connected to syn-sedimentary or syn-diagenetic deformation (director K. Brezsnýánszky for scale)

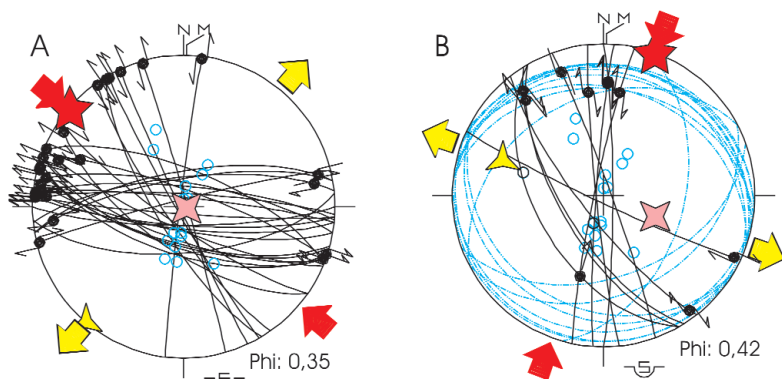


**Figure 24.** North-dipping beds of Füred Limestone Formation in the Meggy Hill, Pécsely. Note marl intercalations, which contributed to bedding-parallel flexural slip during folding



**Figure 25.** Dextral fault plane with slickensides parallel to bedding. These features represent either pre-tilt faults or transfer fractures between bedding-parallel slip planes, formed during folding (FODOR, 1997)





**Figure 26.** Small-scale brittle structures in the Meggy-hegy quarry, unpublished data of F. BERGERAT, A. DUDKO and L. FODOR. A) Conjugate dextral and sinistral faults suggest NW-SE compression, related to Cretaceous(?) or Miocene(?) deformation phase. Faults seem to post-date folding with E-W axis, which are depicted by poles of bedding. B) N-S dextral faults could suggest NNE-SSW compression which also characterise the folds themselves. If dextral faults are transfer fractures, stress axes could not be inferred Phi ( $\Phi$ ) indicates ratio of stress axes:  $\sigma_2 - \sigma_3 / \sigma_1 - \sigma_3$

mező, a relatively flat area, which is underlain by Upper Triassic Hauptdolomit Formation. The dolomite is very thin here, proved by geophysical soundings (HOFFER and SZILÁGYI 1990). This thin flake was thrust south-eastward on Middle Triassic (BUDAI and CSILLAG 2000) thus giving a young-on-older thrust geometry (Figure 28). This relationship clearly demonstrates that the Nagymező flake was emplaced after initial deformation events. It is suggested that its basal thrust plane cuts the Litér thrust, situated at the northwestern side of the Hauptdolomit flake.

The late young-on-older thrust may have importance for Miocene (and younger?) tectonic deformation phases. The imbricate slice is limited on the southwest by a lateral ramp, which can be interpreted as a transfer strike-slip fault (Figure 28). This fault segment can be connected to the Padrag dextral strike-slip fault further to the west (MÉSZÁROS 1982). The connection is made across the Litér thrust, which shows a 2 km bend or displacement at the junction with the Nagymező transfer fault. The Padrag fault has the same 2 km separation on Eocene markers as had the lateral ramp in the Nagymező. It seems to be geometrically logical that the Padrag fault was mainly accommodated by the drag of the older Litér thrust and by the development of the Nagymező flake. This idea is similar to what MÉSZÁROS (1982), TARI (1991) and DUDKO (1991) already suggested. The displacement could have Miocene age, while displacing Palaeogene beds and influencing the distribution of middle Miocene formations (MÉSZÁROS 1983; TARI 1991). The frontal reverse fault of the Nagymező flake seems to be sealed by late Miocene shoreface sand and gravel, so the activity predates late Miocene. Projecting the age from other parallel dextral faults of the TR (MÉSZÁROS 1983; KÓKAY 1976), a middle Miocene (Sarmatian?) age can be postulated for the strike-slip and connected thrust faulting.

This viewpoint represents a good occasion to discuss the origin of the Lake Balaton from structural point of view. Traditionally, Quaternary longitudinal normal faults were connected to the formation of the depression of the lake. In fact, normal faults seem to be connected to Pannonian (late Miocene) sediments near the north-eastern corner of the lake (Figure 3). However, the age of these faults is somewhat uncertain; they could be syn-sedimentary with respect to sediments or even slightly pre-date the abrasional Pannonian strata. Slight post-Miocene reactivation is also postulated.

may explain the oblique position of folds with respect to the compression.

#### Stop 12.

View to the Tihany peninsula, Lake Balaton and Nagymező from György Hill, Balatonfüred

The panoramic view toward Tihany peninsula depicts a sub-horizontal morphological surface cutting the older volcanic edifices (Figure 27). This erosional surface might correspond to a Pliocene(?) pediment which descends from the Balaton Highland southward to the Somogy Hills. It is not yet clear whether this surface was tilted, (folded) during Plio-Quaternary tectonic activity or just dissected by divers young denudation processes.

The view eastward shows the Nagy-



**Figure 27.** Panoramic view to the Tihany peninsula from Balatonfüred, György Hill. Note sub-horizontal denudation surface eroding the late Miocene (~7,5 Ma) volcanic rocks

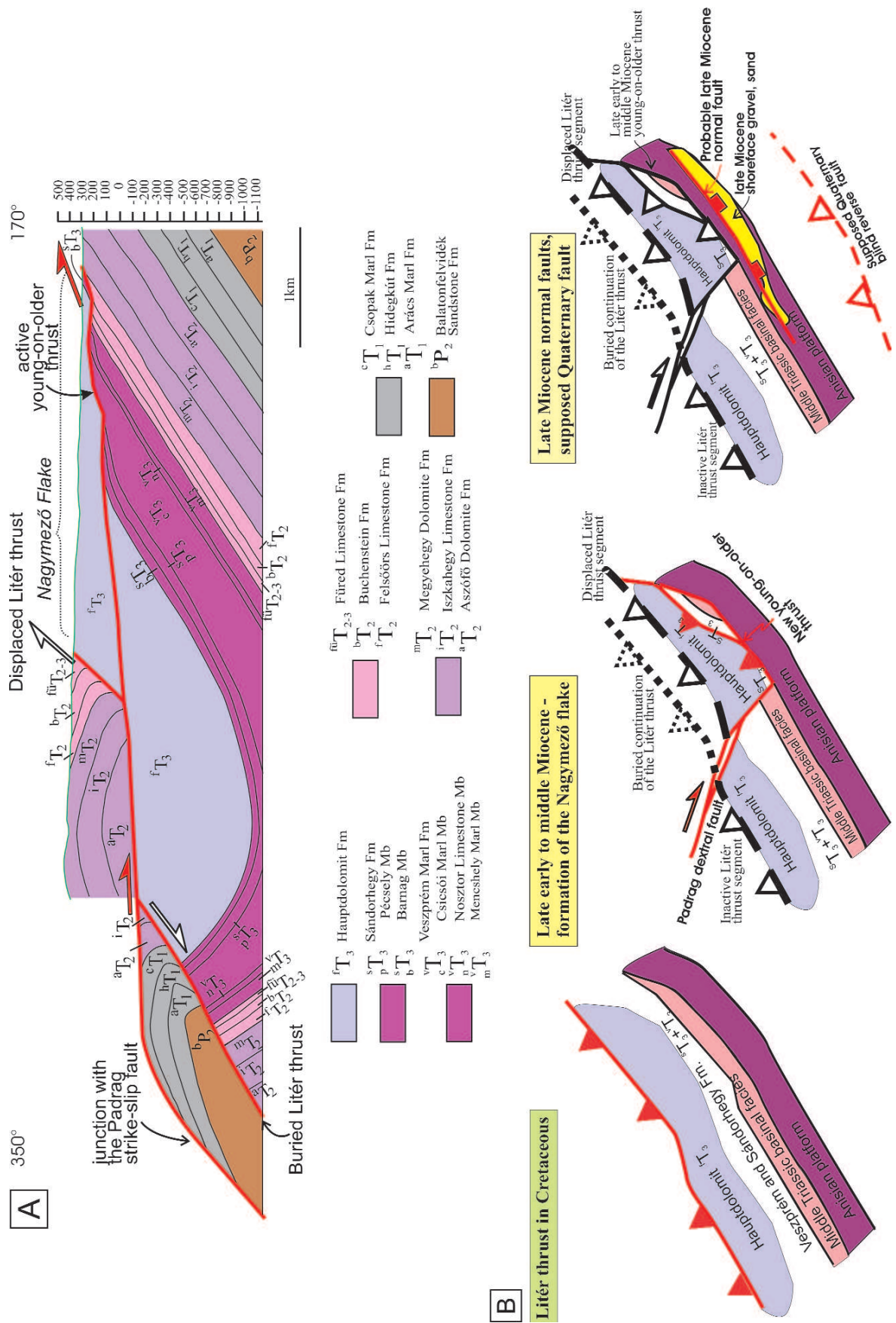


Figure 28. A) Cross section through the Nagymező flake, after the measurements of HOFFER and SZILÁGYI 1990, BUDAI et al. 1999, modified. B) schematic evolution of the Litér thrust, Padrag fault and the Nagymező flake (partly after MÉSZÁROS 1982, TARI 1991, BUDAI et al., 1999). Red lines indicate active structures while black ones correspond to inactive ones



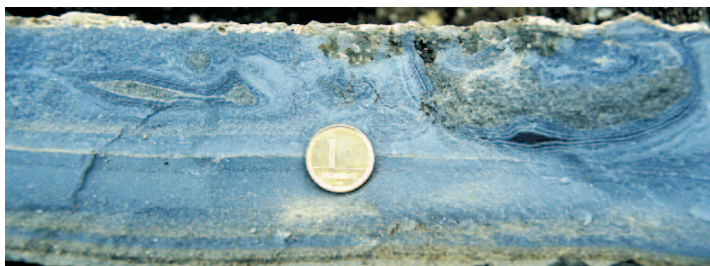
On the other hand, participants of the excursion formulate another hypothesis, which use geometrical similarities to neighbouring areas, and assume the reactivation of older structures of the Balaton Highland. For example, the young-on-older thrust of the Nagymező, or similar features, now still in blind position, could influence the formation of the depression of the Lake Balaton during the Quaternary, and also could lead to uplift of the Highland areas (Figure 28). In that case, the structural depression of the lake would represent a small amplitude fold (syncline), in other word compressional or transpressional feature (Figure 3).

This idea would be in agreement with some neotectonic data. South from the lake ERDÉLYI (1962) postulated NE–SW trending anticline and parallel, NW verging steep reverse faults based on some scattered observations on small-scale faulting in Plio-Quaternary sediments. SACCHI et al. (1999) demonstrated compressional structures below and south of the Lake Balaton, in late Miocene sediments. Studies of seismic sections south from the lake generally reveal the existence of folds or transpressional structures (SÍKHEGYI 2002, BADA et al. 2003, CSONTOS et al. 2003). The present-day stress field and earthquake focal solutions (BADA et al. 1999) seem to be perpendicular or oblique to the long axis of the lake, thus suggesting other than normal type of faulting (BADA et al. 2003). However, it is to note that the stress field seems to be complicated near the lake. Shallow seismic survey near Siófok (VIDA et al. 2001) and field observation near Balatonfűzfő (LOPEZ CARDOSO et al. 2002) demonstrate NE-trending left lateral and NNE–SSW trending normal faults having affected the late Miocene sediments. The Pannonian outcrop at Papvásár-hegy, Balatonfűzfő was deformed by a conjugate set of strike-slip faults, characterized by N–S compression and perpendicular tension (data of BADA, KOROKNAI, FODOR, in BALLA and DUDKO 1996). All these data suggest that either the Pliocene–Quaternary deformation history was multiphase or the coeval deformation and stress field was complex near the Lake Balaton. The former solution was suggested by MAGYARI et al (2003) in the Somogy Hills.

In any structural solution, it is clear that the amplitude of the tectonic motion could be smaller than the topographic difference between the bottom of the lake and the peaks surrounding it. Thus, the morphological depression of the lake was partly created by denudation processes, namely fluvial erosion, deflation and abrasion by the present and ancient lake waters.

### Stop 13. Pula, alginite quarry

The partially active alginite quarry is located NW of Pula village. The Pula Alginite Formation was deposited inside a former tuff ring (JÁMBOR and SOLTÍ 1975, HAJÓS 1976, SOLTÍ 1981, VOS et al. 2000, NÉMETH et al. 2002). The ring is now mainly covered by Quaternary sediments (BUDAI et al. 1999). The Pula maar is interpreted to be a Pliocene eroded, phreatomagmatic volcano on the basis of the presence of chilled volcanic glass and accidental lithic lapilli-rich volcaniclastic rock units in association with the volcanic depression (NÉMETH et al. 2002). The volcano is part of the Mio/Pliocene



**Figure 29.** Features demonstrating strong syn-sedimentary deformation and shaking in the Pula maar sequence. Loading structures in fine-grained volcaniclastic turbidites from Pula. Coin for scale is 1 cm across



**Figure 30.** Small-scale syn-sedimentary faults in the Pula alginite mine

Bakony – Balaton Highland Volcanic Field. The remnant of the maar consists of a (1) distinct depression with a thick alginite, lacustrine laminite infill interbedded with coarse grained lapilli tuffs (Figure 29) (2) a narrow belt of a primary pyroclastic unit (tuff ring) in the marginal zone of the depression (erosion remnant of the tuff ring) and, (3) a re-worked coarse-grained volcaniclastic unit in the marginal zone. The lower part of the central alginite-rich unit consists of greenish-white diatomite, while the upper part is grey, bedded and laminated (Figure 30) (JÁMBOR and SOLTÍ 1975, HAJÓS 1976, PÁPAY 2001). The crater is a north–south elongated depression, currently forming a ~50 m deep basin.

The rocks have been grouped into four major lithofacies on the basis of their bedding, sorting, grading and compositional characteristics. The central part of the depression is filled with finely bedded, laminated, normally graded, fine-grained volcanic silt and sandstone with angular quartz and minor non-to-weakly vesicular, non-abraded tephrite to phonotephrite glass shards (facies 1 — central laminated) all indicative for turbidite sedimentation (SCHARF et al. 2001).

Thicker bedded, coarse-grained lapilli tuff beds are predominantly inverse-graded indicative of grain flow deposition (facies 2 — central juvenile-rich fa-

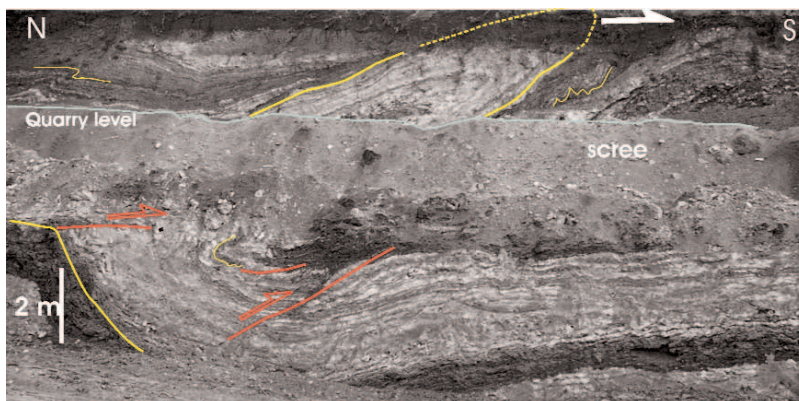
cies). Tephrite/phonotephrite glass shards are weakly vesicular, microlite poor and blocky suggesting a phreatomagmatic origin (HEIKEN and WOHLTZ 1986). The coarse-grained beds often truncate underlying laminae causing dewatering structures, soft sediment deformation including faulting and development of dish structures (Figure 29). All these features suggest active syn-sedimentary slumping and shaking, and are interpreted as debris flow and/or turbidity current emplacement from the crater rim accompanied by palaeo-earthquakes e.g. (SMITH 1986).

The marginal area of the depression is formed by a narrow belt of cross, parallel- and dune-bedded phreatomagmatic lapilli tuff and tuff beds often with rim-type accretionary lapilli (SCHUMACHER and SCHMINCKE 1991) and large proportion of accidental lithic clasts (facies 3 — tuff ring facies; ~30 m thick), that is interpreted to be primary in origin (NÉMETH et al. 2002).

The fourth facies (facies 4) dips at 20–30° towards the centre of the volcanic depression and its sedimentary features suggest a reworked origin by debris flows, which are generated on the inner wall of the crater (SMITH 1986, WHITE 1992, NÉMETH 2001). The common presence of abraded pyroclastic fragments in these beds suggests that already diagenised pyroclastic rocks existed prior to their disruption (FISHER and SCHMINCKE 1984). The large number of lava clasts and their diverse shape indicates that older lava units were disrupted by the volcanic eruption of Pula and subsequently reworked in form of debris flows that developed on the collapsing inner wall of a phreatomagmatic volcano (NÉMETH et al. 2002).

The laminated, poorly consolidated alginite is horizontally bedded. Along the crater rim syn-sedimentary faults could be activated and induced sedimentary deformations, like sliding, small slumping, brecciation, redeposition (Fig. 30) (NÉMETH et al. 2002) similarly to other maar type volcanic basins such as Meerfelder (Eifel, Germany) (DROHMANN and NEGENDANK 1993, BRAUER et al. 2000) and Eckfeld (Eifel, Germany) (FISHER et al. 2000).

In a newly exposed part of the outcrop, some dm-to-m-scale folds were visible in the upper part of the laminated maar sediments and the Quaternary cover (Figure 31). Folds were locally overturned and verging southward parallel to the present-day low-angle slope direction. During the field-excursion it was evaluated if the overlying Quaternary colluvium was affected by folding or it just appeared in erosional channels. The origin of folding was also considered from different point of view. It is possible that folds were induced by 1) syn-volcanic earthquakes affected non-lithified sediments of the maar and/or 2) they could have been formed in the Quaternary under periglacial climate as a result of alternation of freezing-and-melting process and soil creep. For conclusive interpretation further study of the phenomenon would be desirable to unravel Plio-Quaternary tectonism of the area.



**Figure 31.** Folds in the upper level of the Pula alginite mine. Note that fold geometry changes along fold axial plane; it is slightly asymmetric fold in lower dark argillitic layers but overturned in the upper part. This could suggest strong southward (right) drag affected near the top of the section (arrow)

#### Stop 14. Hajagos Hill

Few tens of metres thick volcanoclastic rocks underlie capping basanite lava layers at Hajagos Hill. This volcanoclastic sequence dips gently towards the center of the Hajagos Hill. The volcanoclastic beds consist of uniform coarse-grained and fine-grained bed couplets (MARTIN and NÉMETH 2000). The slightly micro-vesicular, predominately tachylite lapilli are semi-rounded but small grains of angular, blocky slightly oriented textured sideromelane lapilli are also present. Coarse-grained lapilli tuff beds display normal grading. Lapilli are dominantly semi-rounded and display rims comprised of altered glassy material. The lapilli are interpreted as clasts derived from a pre-existing cemented deposit, with the rims around larger volcanic clasts representing remnants of the former matrix of a lapilli tuff. Therefore the lapilli tuff beds inferred to be reworked volcanoclastic rocks deposited in a volcanic depression such as a maar or crater lake of tuff ring (MARTIN and NÉMETH 2000). The well-localized, semi-circular distribution of a negative Bouguer-anomaly, positive magnetic anomaly, the semicircular gently inward dipping volcanoclastic sequences all indicate that Hajagos Hill is an erosional remnant of a maar/tuff ring (MARTIN and NÉMETH 2000). The erosional remnant today forms a dish-like structure filled with basanite lava flows with up to 3 well-distinguished lava units each having an individual thickness of up to 10-metres. The pyroclastic rocks are rich in chilled juvenile, weakly to moderately vesicular basanitoid lapilli (MARTIN and NÉMETH 2000).

The volcanic depression was partly filled with reworked volcanoclasts deposited in a crater lake. The vent zone of the Hajagos Hill volcano was occupied by volcanoclastic slurry and large (metre-scale) chunks of irregular shape siliclastic sediments derived from rocks immediately underlying the volcanic sequences (MARTIN and NÉMETH 2000).





**Figure 32.** Volcanic features of the Hajagos Hill. Marginal zone of a basanitoid intrusive body at the crater zone of the Hajagos Hill. Note the onion-like jointing pattern of the coherent flow body as well as the yellowish, light colour siliclastic matrix of the peperite surrounding the intrusive body. Trees are about 3 m high



**Figure 33.** Blocky peperite from the marginal zone of a basanite intrusive body of Hajagos Hill. Note the jigsaw fit structure of fragmented basanite clasts in the fine siliclastic matrix. Note hammer for scale

quartz sand/silt (MARTIN and NÉMETH 2000). Pyroclastic deposit hosted peperite is present over areas up to several metres across and is characterized by complex relationships between intrusive basalt and tuff or lapilli tuff. The volcanoclastic host most closely resembles the maar-forming volcanoclastic and/or vent-filling pyroclastic (diatrema-filling) deposits according to its composition and texture (MARTIN and NÉMETH 2000). In the vicinity of the magmatic bodies, the sedimentary structures are destroyed but further away from the contact they are well preserved (MARTIN and NÉMETH 2000). The liquefaction of fine matrix of the volcanoclastic host seems to be effective as it is recorded by oriented minerals, glass shards and quartz grains preserved along larger magmatic clasts (MARTIN and NÉMETH 2000). The pillow-shape blobs are typically about 20 cm long but in places are associated with larger, ellipsoidal or tongue-like bodies up to 1 m. The margins of the bodies are either smooth or highly irregular and crenulate on a fine scale (MARTIN and NÉMETH 2000, 2002c). Other type of globular peperite has been identified at Hajagos Hill that was developed at the base of lava flows in-filled their craters (MARTIN and NÉMETH 2000) as results of lava entering to a water-rich environment (JONES and NELSON 1970). The peperite at Hajagos Hill may have formed by a combination of at least two processes. Basanite magma may have flowed over a swampy area or shallow ponds, where a large amount of steam locally formed the "mega bubbles" in the lava flow (MARTIN and NÉMETH 2000). In this case the lava flows are interpreted as having been emplaced into a surrounding swampy area after overflowing the former tuff ring rim. Alternatively, the peperite formed subaqueously, while lava flows erupted into the crater lake (MARTIN and NÉMETH 2000, 2002a, 2004b).

Dykes invaded subsequently the vent-filling mixture of sediments forming both globular and blocky peperite (Figure 32) and proving the host sediment's wet and unconsolidated state (MARTIN and NÉMETH 2000). Peperites (WHITE et al. 2000) have been divided into two end-member textural types, which correspond to blocky and globular (BUSBY-SPERA and WHITE 1987). Globular peperite generally tends to develop when magma intrudes into fine-grained host sediments such as silt or mud (BUSBY-SPERA and WHITE 1987). In contrast blocky peperite forms more commonly in coarser grained hosts such as coarse sand or conglomerate (BUSBY-SPERA and WHITE 1987). However, evidences from the western Pannonian Basin seem to demonstrate that no direct link may exist between host sediment grain size and types of peperite developing along dyke and/or lava flow margins associated with sub-surface settings of phreatomagmatic volcanoes (MARTIN and NÉMETH 2000, 2002a, b, c, 2004b). At Hajagos Hill both blocky and globular peperite coexist in relationship with a seemingly similar host sediment. The presence of blocky peperite in a coarse grained, volcanoclastic host sediment in the same localities highlights the complexity and the more diverse controlling parameters forming different types of peperite. Blocky peperite at Hajagos Hill is breccia-like and is always related to columnar jointed basanite (Figure 33), with fan-like columns or developed along the margin of fluidally shaped magmatic bodies (MARTIN and NÉMETH 2000, 2004b). The basanite dykes are interpreted as feeder conduits to a lava lake, which filled the crater. The feeder dykes propagated through the phreatomagmatic deposits generating peperite by interaction with 1) muddy, water-saturated siliclastic and 2) volcanoclastic sediments.

Fluidal peperite is common in Hajagos Hill where the host is either 1) lapilli tuff/tuff and 2)



### Stop 15. Kál Basin, Szentbékállá

The outcrop represents a nature-protected area west of the village of Szentbékállá. The cliffs consist of silicified upper Miocene (Pannonian) sandstone and pebbly sandstone (BUDAI et al. 1999, 2002). The deposition probably occurred in an embayment of the Lake Pannon, close to the shore (BABINSZKI et al. 2003). Small and large scale cross bedding indicate agitated water (currents?). The origin of silicification is under debate, it can probably be related to ground water silification, after the model of THIRY and BERTRAND-AYRAULT (1988), THIRY and MARÉCHAL (2001). The deposition of late Miocene sand and silt preceded the Pliocene basalt volcanism.

The other interesting aspect of this outcrop is the numerous wind-polished surfaces on the Pannonian sandstone blocks (BUDAI et al. 2002, JORDÁN et al. 2003, FODOR et al., 2003). These surfaces occur mainly on the top part of the blocks, but some joint surfaces are also polished (Figure 34). Individual pebbles were frequently half-eroded. These wind-polished surfaces reflect strong wind deflation period(s) derived already on the basis of landforms (CHOLNOKY 1918) and numerous ventifacts (dreikanter) found in the whole TR (Figure 3) (JÁMBOR 1992, 2002). Wind direction at Szentbékállá corresponds to NW–SE (FODOR et al., 2003), which also fits well with the regional wind direction suggested by JÁMBOR (2002).



Figure 34. Wind-polished surfaces of the Pannonian sandstone at Salföld, in similar situation like at Szentbékállá. Paleo-wind direction from NW to SE, away from viewer

### Stop 16. Hegyestű

The protected outcrop is found in an abandoned quarry between the village of Zánka and Monoszló, along the marked tourist path on the top of Hegyestű Hill. It shows a nearly 3 dimensional view of a Pliocene volcanic neck with hexagonal columns, whose textures are in relationship with the cooling rate of the igneous body (BUDKEWITSCH and ROBIN 1994). At Hegyestű, no phreatomagmatic lapilli tuff or tuff have been found yet. In contrast, 2.5 km south of Hegyestű, volcanic glass shard-rich lapilli tuff crops out in a small semicircular region (NÉMETH et al. 2003c). This rock has been interpreted as a diatreme-filling pyroclastic rock facies. (NÉMETH et al. 2003c). The common presence of sedimentary grains derived from the late Miocene lacustrine formations suggests that these sedimentary units were still intact in the Kál Basin region in syn-volcanic time which information bears great significance in understanding the landscape evolution of this region in the last few millions of years (NÉMETH et al. 2003c).

Late Miocene deformation induced the formation of the Kál Basin (DUDKO in BUDAI et al. 1999). The late Miocene shore-face sand and gravel is detectable along the rims of the present-day morphological depression of the Kál Basin (BUDAI et al. 1999). These marginal facies follow basin-bounding syn-sedimentary faults (DUDKO 1991, DUDKO et al. 1992, BUDAI et al. 1999, JORDÁN et al. 2003). On the other hand, the Pannonian basin fill was affected by renewed faulting of Pliocene or Quaternary age (JORDÁN et al. 2003).

In addition to prove late Miocene and questionable Plio-Quaternary tectonics, Quaternary denudation processes were important in the formation of the depression. The wind erosion inferred to play an important role in this denudation process (FODOR et al. 2003) on the basis of existence of 1) number of sites of Pannonian sandstone containing wind-polished surfaces (see Szentbékállá, stop 15); 2) the present-day depression has practically no outflow drainage or central valley (Figure 3). Fluvial erosion does not seem to have affected significantly the depression during the late Pleistocene–Holocene. However, at least ~100 m Pannonian sediments were eroded from the basin if taking account the highest position of Pannonian sediment on the margin, and the depth of the depression. The erosion of the softer Pannonian sediments was almost complete, while Permian–Triassic rocks are cropping out at the rims and in the deepest topographic position. Only the younger basalt cover, phreatomagmatic diatreme deposits or silicification of the lacustrine rock units preserved remnants of the Pannonian sediments at specific locations. It is a reasonable interpretation in the present stage of research that denudation of the Kál Basin was driven by deflation rather than fluvial processes. This suggestion is similar to interpretations of earlier works (CHOLNOKY 1918, JÁMBOR 1992, 2002).

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Panoramic view from the wind-polished sandstone blocks (late Miocene depositional age) near Szentbékállá

Presentation at Felsőörs, near the protected section of middle Triassic formations by Tamás Budai, János Haas (standing in the middle)



Participants of the excursion in the quarry of basalt columns at Hegyestű, Zánka.

Tibor Cserny, Gábor Csillag, Giovanni Sella, Károly Brezsnýánszky, John Dewey, László Fodor, John Weber, István Bíró, Dóra Halász, Károly Németh (standing row), Ada Kiss, Nicholas Pinter, Maria Mange, Suzanne Weber (front row)

Testing and tasting of the connection of wine and geology in the Balaton Highland, at the cellar of Tibor Cserny.

Károly Németh, Giovanni Sella, Zsófia Ruszkiczay-Rüdiger, Ada Kiss, Károly Brezsnýánszky, László Fodor, John Dewey, Dóra Halász, Anikó Cserny (standing row), István Bíró, John Weber, Suzanne Weber, Nicholas Pinter, Maria Mange (front row)



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