Tertiary tectonic evolution of the Pannonian Basin system and neighbouring orogens: a new synthesis of palaeostress data

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Abstract: Compilation of a microtectonic observation data base for most of the data measured in the Pannonian Basin and surrounding orogens permits a detailed reconstruction of the Tertiary stress field evolution. Combination of tectonic observations, borehole, gravity and seismic data, palaeogeographic and stratigraphic information led to an understanding of fault kinematics and description of the structural evolution in seven major tectonic episodes.

The first two episodes depict the kinematics of the two major separated blocks, the Eastern Alpine–Western Carpathian–Northern Pannonian (Alcapa) and the Southern Pannonian–Eastern Carpathian (Tisza–Dacia) microplates. A Mid-Eocene to Early Oligocene N–S compression led to contractional basin formation both in the foreland (Western Carpathians) and hinterland (Hungarian Palaeogene basins) of the orogenic wedge. Due to oblique convergence, the Palaeogene basins are generally asymmetric and often dissected by dextral tear faults.

Northward advance of the Adriatic promontory initiated the separation of the Alcapa from the Southern Alps and its eastward extrusion. This process probably started during latest Oligocene and reached its climax during the Early Miocene. The main displacement was accommodated by dextral slip along the Periadriatic and Mid-Hungarian shear zones and during and after this tectonic episode Alcapa suffered 50° CCW rotation. At about the same time period the Tisza–Dacia block also experienced rotation of 60–80°, but clockwise. These opposite rotations resulted in the marked actual deviation of earlier compression axes, which are now N or NW in the Eastern Alps, WNW-ESE in the Western Carpathian–Pannonian domain and NE–SW in the Tisza-Dacia domains. Termination of rotations can be considered as the time for final amalgamation of the two separate blocks and the beginning of extensional tectonics in a single Pannonian unit.

The Pannonian Basin system was born by rifting of back-arc style during the late Early and Mid-Miocene time. Extension was controlled by the retreat and roll-back of the subducted lithospheric slab along the Carpathian arc. Two corners, the Bohemian and Moesian promontories formed gates towards this free space. At both the northern and southern corners, broad shear zones developed. The initial NE-directed tension was gradually replaced by a later E- to SE-directed tension as a consequence of the progressive termination of subduction roll-back along the arc from the Western Carpathians towards the Southern Carpathians. There is growing evidence that an E–W-oriented short compressional event occurred during the earliest Late Miocene but during the most of the Late Miocene extension was renewed. Starting from the latest Miocene roll-back terminated everywhere and a compressional stress field has propagated from the Southern Alps gradually into the Pannonian Basin, and resulted in Pliocene (?) through Quaternary tectonic inversion of the whole basin system.

During the past 15 years, the understanding of the tectonic evolution of the Pannonian Basin has increased considerably. This improvement had several sources. One important contribution has come from the increased precision in biostratigraphy (e.g. Báldi 1986), and radiometric dating of magmatic rocks (Pécskay *et al.* 1995). Another factor was the publication of several seismic sections from the basin system including seismic stratigraphic results (Kilényi *et al.* 1991; Lőrincz & Szabó, 1993; Pogácsás *et al.* 1988, 1994*a*,*b*; Rumpler & Horváth 1988; Tari *et al.*

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1992; Horváth *et al.* 1995). The third source of information has been the systematic structural investigation, i.e. a great number of tectonic measurements and related palaeostress calculations have been carried out in areas where brittle rocks crop out. The main goal of the present paper is to review and interpret this palaeostress data set.

Tectonic measurements started in the frame of a Hungarian–French co-operation. The first results seemed to fit the general extensional–transtensional character of the Pannonian Basin (Bergerat *et al.* 1983, 1984). The following research gradually involved different authors in other areas of the basin system and its surroundings (Poljak 1984; Bergerat & Csontos 1988; Nemčok *et al.* 1989; Fodor *et al.* 1990) and permitted the first basin-scale synthesis (Csontos *et al.* 1991). By the beginning of the 1990s numerous groups started their researches in the Western Carpathians, Southern and Eastern Alps. Their results describe the local structural geology and often modify ideas on the general structural evolution of the whole area (Kováč *et al.* 1989; Marko *et al.* 1990, 1991; Nemčok & Lexa, 1990; Decker *et al.* 1993, 1994; Linzer *et al.* 1995; Nemes *et al.* 1995*a*,*b*; Ratschbacher *et al.* 1993*a*,*b*; Kováč & Hók, 1993; Vass *et al.* 1993, Györfi *et al.* this volume). Recent attempts have been also made for regional synthesis using stress data (Nemčok 1993; Decker *et al.* 1993; Linzer *et al.* 1995; Marko *et al.* 1995).

This paper reviews the results of tectonic measurements carried out in the Pannonian Basin system and neighbouring parts of the Alpine–Carpathian–Dinaric system. Interpretation of measurements has been improved by the analysis of map-scale structures, stratigraphic research, borehole data, seismic sections and gravity maps. Incorporation of palaeomagnetic data in tectonic reconstruction is also inevitable because they show large and complex during the Tertiary (Márton 1987; Márton & Mauritsch 1990; Márton & Márton 1996). Structural models based on map analysis or analogue modelling (Balla 1984; Ratschbacher *et al.* 1991*a*,*b*) also contributed to our geodynamic models. Using our palaeostress database and the above mentioned

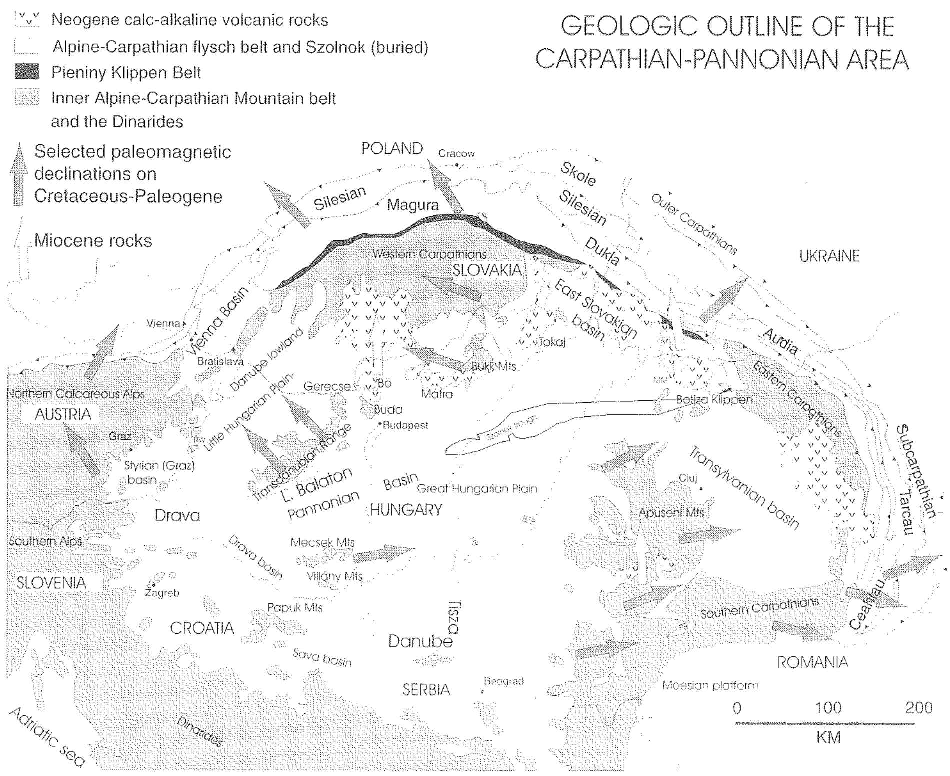


Fig. 1. Major units and pre-Neogene outcrops of the East Alpine–Pannonian–Carpathian region. Palaeomagnetic data after Márton & Mauritsch (1990); Márton & Márton (1996); Krs *et al.* (1982, 1991); Bazhenov & Burtman (1980); Bazhenov *et al.* (1993); Pătraşcu *et al.* (1992, 1994).

other structural data, we give a detailed description of the Tertiary stress field and structural evolution than we attempt to arrive at a geodynamic model which connects the stress field evolution and the kinematics of the area.

Geological setting

The Carpathian mountains and the Dinarides encircle an inner lowland. This Intra-Carpathian Basin is not uniform, but is subdivided by low-altitude inselbergs, e.g. the Transdanubian Range (TR), Bükk, Mecsek, Villány, Papuk and Apuseni mountains (Fig. 1). The major subbasins are: the Graz Basin, the Little Hungarian Plain–Danube Lowland, the Great Hungarian Plain, which form together the Pannonian Basin. The Transylvanian, Vienna and East Slovakian Basins are situated close or above the contact of the inner Eoalpine nappes and the Outer Carpathian Flysch Belt (Fig. 1).

The basins were formed by stretching of continental lithosphere during the Neogene, synchronously with the compressional deformation of the Outer Carpathian flysch nappes (Horváth & Royden 1981; Horváth & Rumpler 1984). The gross geological structures of the Carpathians and the Dinarides are similar. Both are formed by nappes verging outward from the Pannonian centre. The external units are composed of a Late Cretaceous–Early Neogene flysch wedge, deformed during the Tertiary and thrust on their foredeep and foreland (Fig. 1). The internal units are made up of less continuous exposures of Mesozoic rocks and their crystalline basement, deformed during different periods of the Jurassic and Cretaceous. A narrow, strongly deformed zone of Mesozoic rocks, the Pieniny Klippen Belt is situated between the Inner West Carpathians and the Flysch belt (Birkenmajer 1985). The Neogene volcanic belt follow the outlines of the chains on their internal side. Large amounts of volcanic rocks are buried beneath sediments of the Pannonian Basin. Modern seismic and borehole investigations revealed that the substrata of the Neogene Basins is composed of Alpine nappes (Papp 1990; Horváth 1993; Tari 1994), which are the continuations of outcropping units around the basin system.

Based on palaeobiogeographical contrast in the northern and southern inselbergs (TR and Mecsek-Villány Mts respectively) Géczy (1973, 1984) recognized a major faunal discrepancy: the TR showed an affinity to a Tethyan palaeobiogeographical province, while the Mecsek and Villány Mts to a European palaeobiogeographical province during the Early Jurassic. This interpretation is corroborated by a number of subsequent faunal studies (Vörös 1993). This implies that the substratum of the Pannonian Basin is not uniform, but is composed of at two major blocks of different original paleogeographic position (Fig. 2).

Palaeomagnetic measurements also support this tectonic subdivision. The declinations of the TR, Bükk Mts, together with the Inner West Carpathians show Tertiary counterclockwise rotation (Márton 1987; Márton & Márton 1996). In contrast, data from the Mecsek, Villány and Apuseni Mts, together with the East and South Carpathians show Tertiary clockwise rotation (Márton & Márton 1978; Balla 1987; Pătraşcu *et al.* 1992, 1994; Bazhenov *et al.* 1993; Fig. 1). Because the amount of rotation is quite uniform within the two units, it is probable that the two major blocks moved as separate microplates during the Tertiary (Balla 1984, 1988; Pătraşcu *et al.* 1994).

These continental microplates are called the Alcapa and Tisza–Dacia units (Balla 1984; Csontos *et al.* 1992; Vörös & Csontos 1992; Fig. 2). They had significantly different Mesozoic position and were juxtaposed in the late Tertiary. The boundary between the two units is a tectonic zone, the Mid-Hungarian fault system, which is buried for most of its length below the Neogene volcanic and sedimentary rocks of the Pannonian Basin.

The Alcapa block (Fig. 2) is located north of the Mid-Hungarian fault system. Its original northern limit is the Pieniny Klippenbelt (Fig. 1). This northern limit gradually shifted to more external nappe boundaries, as during evolution of the Alcapa, more and more Outer Carpathian flysch nappes were accreted to the internal Mesozoic unit. The western boundary of the Alcapa can be found in the Eastern Alps along the low-angle normal fault at the eastern limit of the Tauern window.

Rotations derived from palaeomagnetic data

In structural analysis and stress field reconstruction in particular, the amount and time of rotations of different units are crucial. We briefly discuss the most important palaeomagnetic data, because description of stress field history is not possible without this information.

Senonian rocks within the Northern Calcareous Alps show slight clockwise rotation (Mauritsch & Becke 1987; Mauritsch & Márton 1995). The rest of the Alcapa exhibit larger clockwise rotations (Márton 1993). In the northern Pannonian Basin and the Western Carpathians

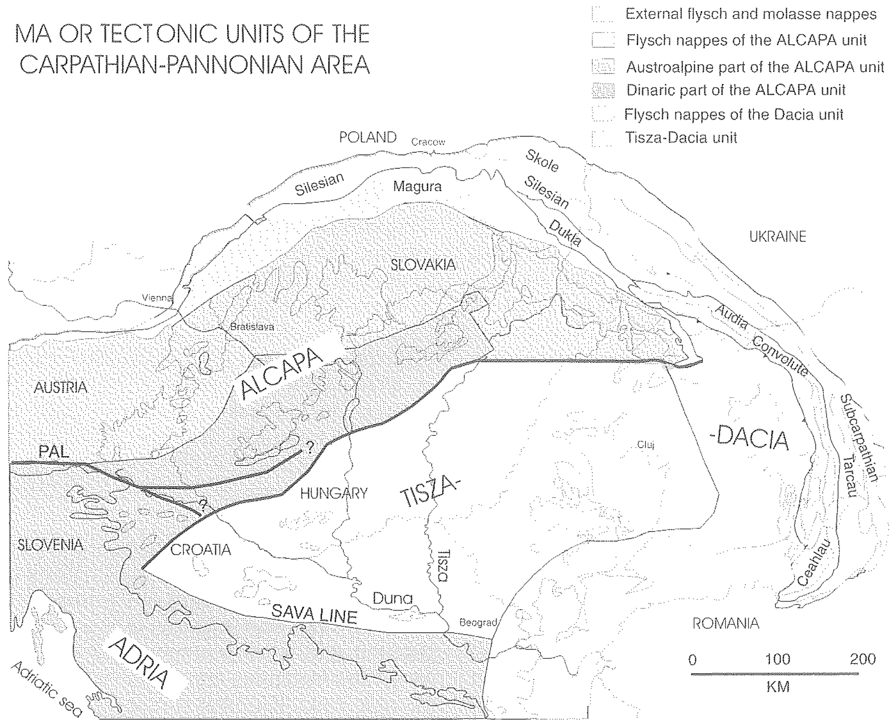


Fig. 2. Major tectonic units of the study area. This division is mainly valid for the late Palaeogene–Early Miocene period.

rotation attained 80° CCW (Márton & Márton 1996; Túnyi & Márton 1996; Márton *et al.* 1996). This happened in two phases: an early 45–50° CCW and a later 30–35° CCW rotation can be distinguished. The age of the first rotation is well-constrained, having occurred in the late Ottnangian–early Karpatian (18–17 Ma; Márton & Márton 1996). The second phase ended before the Mid-Miocene because Badenian and younger rocks did not show anomalous palaeodeclination (Márton & Márton 1989). The only exception occurs in the northeastern part, near the East Slovakian Basin, where 30° CCW rotation occurred as late as the end of the Sarmatian (Márton & Pécskay 1995; Orlicky 1996). The northern Carpathian flysch belt also shows 45–60° CCW rotation (Krs *et al.* 1982, 1991) but this is less constrained in time.

Senonian rocks in the southern part of the Eastern Alps show moderate (30–60°) counterclockwise rotation. A similar 30° of CCW rotation was demonstrated close to this site in northern Slovenia on Badenian rocks (Márton & Jelen in preparation). All these data suggest that the Alcapa terrane rotated consistently in counterclockwise sense, but by different amounts, which indicate internal deformation of Early to Mid-Miocene age.

The Tisza–Dacia unit underwent opposite rotation with respect to the northern unit. Data from southern Hungary (Mecsek) and the Apuseni Mts, Southern Carpathians derived from Palaeogene and earliest Miocene rocks show clockwise rotation of about 90° (Márton & Márton 1978; Surmont *et al.* 1990; Pătraşcu *et al.* 1992, 1994). Because volcanics of 15 Ma and younger age do not show any important rotation, it must have happened during the early Miocene. In conclusion, rotation of the Alcapa and Tisza–Dacia units with different sense occurred at a similar time, between 19 and 15 Ma.

Stress field derived from tectonic observations

We carried out systematic tectonic studies at the margins of the Pannonian Basin system, in Hungary, Austria, Slovakia, Czech Republic, Romania and Slovenia. The field observations consisted of measurements of brittle structures, mainly faults, but joints and folds were also taken into account. Sense of the fault slip was determined using kinematic indicators occurring on and along the fault planes (Angelier 1979*a*; Hancock 1985; Petit 1987).

A theoretically well-known equation defines the relationship between the different type of faults and the stress field (Anderson 1951; Bott 1959). The three orthogonal principal stress axes: the maximum (σ_1), the medium (σ_2) and the minimam axes (σ_3) can be derived from the observed brittle failure pattern. Concerning their mutual position with respect to horizontal, three main type of stress state can be determined in a given point: compressional, strike-slip and tensional stress states if σ_3, σ_2 or σ_1 are vertical, respectively. The method of palaeostress calculation from fault slip (striae) observation was elaborated and extensively discussed by Angelier (1979*b*, 1984, 1990).

At several outcrops faults were not formed under one stress field. In such cases two types of separation into homogeneous subsets were carried out. First, the software of Angelier & Manoussis (1980) allows an automatic phase separation. Parallel to this, faults were separated manually, taking into account Anderson's simple geometric assumptions. Then the manual grouping of faults was tested by computer calculations. In most cases, the 'automatic' and 'manual' phase separation yielded similar or identical stress axes.

A palaeostress data base was set up from the calculated stress data. This contains the site name, coordinates, age of deformed rock, the direction of the main stress axes, the type of calculation method and the number of observations. To cover the whole study area, we completed the data base for the surrounding mountain ranges from publications (see Appendix for sources of the stress data). Stress calculation methods used by different authors were, unfortunately not the same. In some cases the stress axes were only estimated by graphical methods of Alexandrowski (1985) or Angelier & Mechler (1979). The majority of authors used computer calculation of the stress axes, however they applied different softwares (e.g. Nemčok 1993; Ratschbacher *et al.* 1994). This difference can hardly be avoided and some discrepancy in stress data can be attributed to this fact.

In our analysis special attention was given for the timing of each determined stress state. This timing was based on observations at several levels. First, the age of the brittle deformation was interpreted from the outcrop itself, using synsedimentary, syndiagenetic structures, the age of the deformed and the overlying undeformed rocks, relative chronology between structures belonging to separate phases. Then outcrop-scale structures and the calculated local stress field was compared to map-scale structures, to the occurrence of thick sedimentary or volcanic sequences. For example, structures often pre- or post-date the tilting (or folding) of beds, the age of which can be determined from local data. All these data gave lower and upper time constraints for the deformation and palaeostress field for a smaller area. The basin-wide synthesis is composed from such constrained age determinations.

Figure 3 shows an example of this work. The presented Alsótold section is composed of tilted layers of Karpatian to Badenian in the footwall, and Pannonian in the hanging wall of an important, map-scale fault (Noszky 1940). Part of the small faults within the footwall block is tilted, they are characterized by NE–SW tension (Fig. 3b, lower left corner). Hence, this stress field is older than the tilting. Conjugate normal and strike-slip faults were formed by ESE–WNW tension, perpendicular to the axis of the tilt. Part of them are symmetrical to bedding thus they are contemporaneous with the tilting, while others (mainly in the Pannonian) are post-tilting. The deformed Mid-Miocene to Pannonian rocks suggest continuous Sarmatian to Pannonian age for the tilting within the same ESE tension. WNW–ESE compression is indicated by few data; their relation to tilting is not clear, but they do not seem to suffer the full amount of tilt. Finally, NNW–ESE compression postdates the tilting, thus represents a Pontian to Quaternary episode. Seismic data from the nearby Zagyva trough impose a younger lower time constraint for this deformation, around the beginning of the Pliocene.

Different durations were attributed to the separated tectonic phases (palaeostress fields) in different sub-areas of the basin system and surrounding mountain chain. In fact, these variations partly reflect the dating possibility within the sub-areas, but also reflect real inhomogeneity of the stress field. Except for clear cases (e.g. improvement in dating after publication of the data), the inferred timing of the original authors was kept.

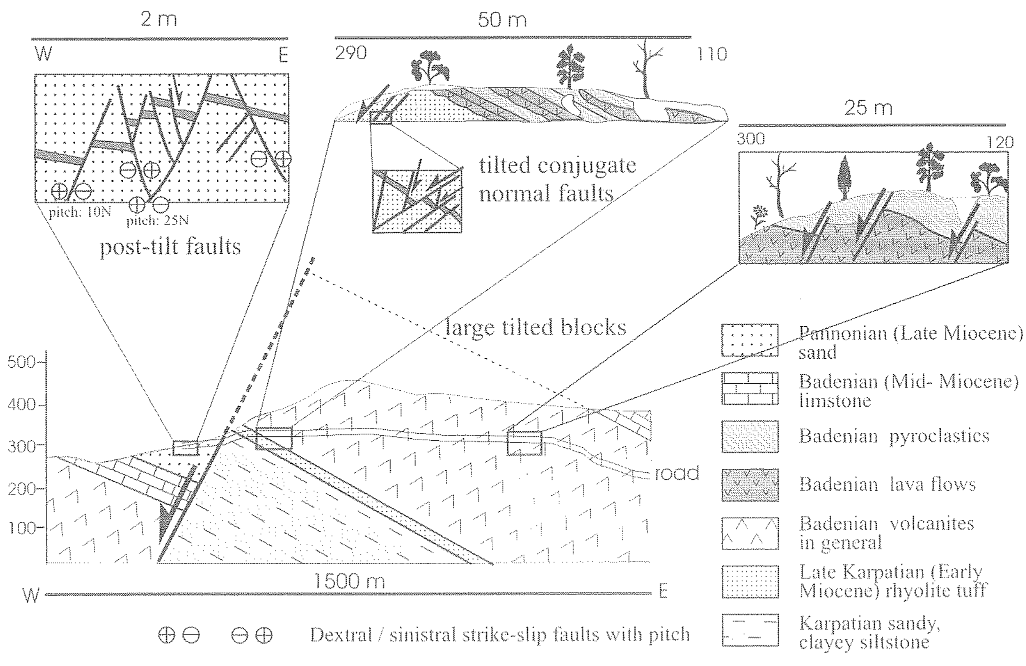
Fault slip data and stress field calculations were used for kinematic analysis of Tertiary structures known from earlier works or established during this study. At first, we made attempts to directly determine the kinematics of some map-scale faults by field observations. Good outcrop-scale kinematic observations were often extrapolated to nearby faults where direct determination of their nature was not possible. Furthermore, the knowledge of the stress field permitted an estimation of kinematics of faults below the thick Quaternary or Neogene sedimentary cover in the deeper parts of the Pannonian Basin. In such areas, we used also published tectonic data (i.e. structural maps, description of folds and faults without stress calculations). In addition, seismic reflection sections, borehole data, gravity map, information on offset paleogeographic boundaries, depositional environments and geometry of the basin fill were also taken into consideration.

The results are presented in a series of maps, together with the stress field data. Reflecting our present knowledge, the structural maps show different levels of accuracy and details. Because of the different early Tertiary tectonic history, we discuss the observed stress field and fault pattern separately for the Alcapa and Tisza–Dacia units for the Palaeogene to the Mid-Miocene. Late Early Miocene was the time of major and opposite sense rotations of these units, and from latest Early Miocene one can consider a more or less uniform Carpathian–Pannonian domain.

Pre-rift events (Eocene to Early Miocene)

Alcapa: repeated dextral transpression – Eocene to early Ottnangian

Eocene to early Oligocene transpression (Figs 4a, 5a) The oldest Tertiary stress field was determined in the Alcapa terrane. The orientation of the stress axes are similar in the different subareas. σ1 was horizontal and oriented to NW–SE in the Eastern Alps (Nemes *et al.* 1995*b*); (Decker *et al.* 1993, 1994; Linzer *et al.* 1995). In the Transdanubian Range σ1 varies slightly from WNW–ESE to NW–SE (Fodor *et al.* 1992; Bada *et al.* 1996) (Fig. 5a). This direction was also determined in the westernmost Carpathians (Marko *et al.* 1990, 1995), and similarly oriented σ_{Hmax} (with vertical σ_1) is postulated in the Central Carpathian Paleogene Basin of the Western Carpathians (Jablonský *et al.* 1994). In the easternmost Alps the characteristic structures are ENE to NE trending thrust or fold axes. NW trending dextral faults could reflect oblique thrusting (Linzer *et al.* 1995). Such contractional deformation started in the Cretaceous, but the deformed Senonian to Lower Eocene sediments demonstrate the continuation during the early Palaeogene.



(a)

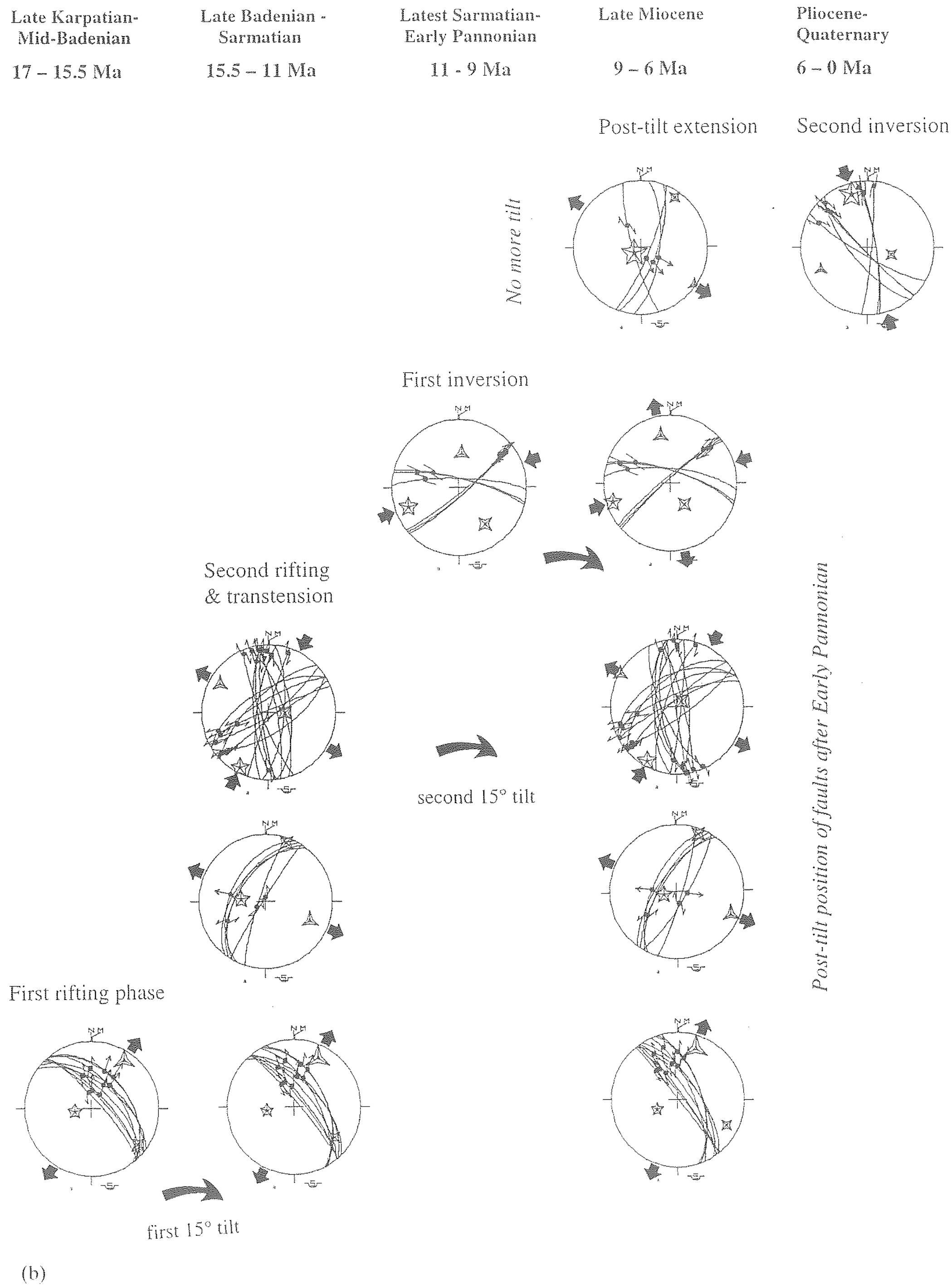


Fig. 3. Example of separation of tectonic phases (Alsótold, Cserhát hills, NE Hungary). (**a**) Geological cross-section, partly after the map of Noszky (1940). Note the eastward tilt of the sequence. (**b**) Stereographic projections of measured faults and calculated stress axes. Pre-tilt (original) and post-tilt (recent) positions are equally shown. Dots represent slickenside lineations on fault planes; double arrows: strike-slip; centrifugal arrows: normal faults (motion of hanging wall); σ_1, σ_2, σ_3 stress axes: stars with 5, 4, 3 branches.

In the Buda and Gerecse Hills the best documented map-scale structures are E–W-trending dextral strike-slip faults, the Nagykovácsi and Budaörs zones (Maros 1988; Balla & Dudko 1989), whose separation was partly accommodated by the blind Buda imbricate stack (Fodor *et al.* 1994) and by reverse faults and folds east of the Danube (Csontos & Nagymarosy 1998). Some en echelon normal faults are also related to strike-slip faults. On the basis of fault slip data, dextral reactivation was postulated just north of the Gerecse, along the Hurbanovo–Diósjenő line.

In these areas the original palaeogeography, structural pattern was reconstructed on the basis of sedimentary dykes, synsedimentary faults and tilting, different sediment thickness and facies, fault-related talus cones, bioperforated fault planes, soft-sediment deformations, different types of gravity flows (Fáy-Tátrai 1984; Fodor *et al.* 1992, 1994; Bada *et al.* 1996; Sztanó & Fodor 1997). Some new observations suggest that faulting can slightly predate Mid-Eocene sedimentation which started on an already dissected topography (Kercsmár 1996). The observations were extrapolated to the central and southwestern Transdanubian Range where measurements are scarce but WNW–ESE compression was determined in some Senonian sites (Fig. 5a). The NE-trending occurrence of Eocene depressions were probably bordered by reverse faults and segmented by E–W dextral faults (Fodor *et al.* 1992). Activity of NE-trending (reverse) faults might be indicated by northwestward thickening Eocene sediment wedges (Mindszenty *et al.* 1988).

Dextral transpression occurred also in the westernmost Carpathians (Plašienka 1991; Marko *et al.* 1990). Geological maps and some scattered microtectonic observations suggest a similar scenario in the Súl'ov Palaeogene Basin (Salaj 1995; Marko & Fodor unpublished data 1991) while the Central Carpathian Palaeogene Basins were probably characterized by WNW-trending normal faults (Kováč, Marko, Fodor unpublished data 1994). This phase of transpression started before the Eocene (Plašienka 1991) and the main phase occurred during the Oligocene (Marko *et al.* 1990).

Latest Oligocene to Early Miocene escape tectonics (Figs 4b, 5b). The Early Miocene (late Egerian–early Ottnangian) stress axes did not deviate significantly from the Eocene ones in the main part of the Western Carpathians and in the Pannonian region (Figure 5b). NW–SE or WNW–ESE σ_1 and perpendicular σ_3 reactivated E–W (ENE–WSW) trending dextral, NNW–SSE- to NNE–SSW-trending sinistral and reverse faults. The separation of this event from the previous one is possible because of good synsedimentary evidence for the older event or, locally by the existence of marked erosion before Eggenburgian times. In the Eastern Alps a new, strike-slip-type stress field was established, whose σ_1 was oriented around N–S (locally slightly deviating to NNW–SSE or NNE–SSW). Corresponding to this, conjugate set of strike-slip faults developed (Decker *et al.* 1993). The main structures are NE- to ENE-trending sinistral fault zones within the Northern Calcareous Alps and northward thrusting in the marginal Flysch and molasse belts. (Fig. 5b). However, the time constraint is relatively poor concerning the beginning of this stress field (Decker *et al.* 1993; Linzer *et al.* 1995). The eastern margin of Inner Western Carpathians was also characterized by a different σ_1 direction, N–S to NNE–SSW (Fig. 5b) (Nemčok *et al.* 1993). This stress field might induce dextral transpression along the eastern Pieniny Klippen Belt (Ratschbacher *et al.* 1993*a*) and thrusting within the Carpathian flysch belt. However, it is difficult to date the stress data, and the phase could have started later (during the Mid-Miocene).

The main deformation of the whole area is due to the dextral (re)activation of the Periadriatic line. Its general dextral character is well documented in the Western Alps (Laubscher 1988; Schmid *et al.* 1989). Recently, Nemes *et al.* (1995*a*) and Fodor *et al.* (1998) offered further structural evidence for dextral shear at the eastern end of the fault system in Slovenia.

To follow the continuation of the Periadriatic line into the Pannonian area, one has to reconstruct the large scale CCW rotation of the Alcapa domain. Before Ottnangian time the southern border of the Alcapa, the Mid-Hungarian shear zone was oriented NW–SE, thus forming a straight eastward continuation of the Periadriatic fault system. The kinematics and amount of dextral separation and timing of the motion of this formerly unique fault system was extensively discussed by Balla (1988), Balla & Dudko (1989), Csontos *et al.* (1992), Tari (1994) and Tari *et al.* (1995). The key point in their reconstruction is the similarity between the Dinaric and Bükk-type late Palaeozoic, the South Alpine and Transdanubian Mesozoic, the Hungarian and Slovenian–Italian Palaeogene developments. The similarity of these presently distant localities were already discussed by Lóczy (1913) and more recently by Premru (1981), Kázmér & Kovács (1985), Ebner *et al.* (1991), Schmidt *et al.* (1991), Csontos *et al.*

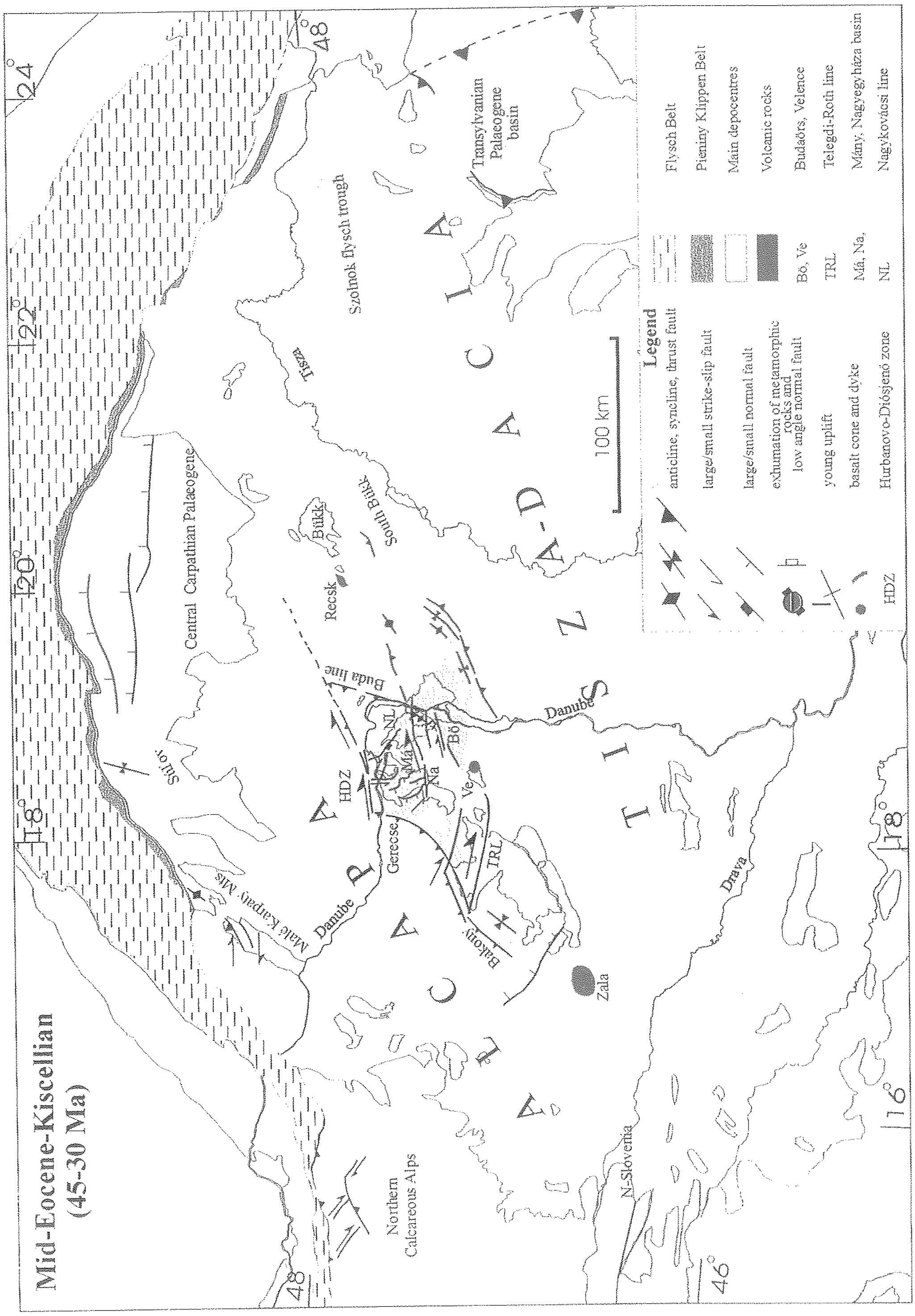

Mid-Eocene-Kiscellian
(45-30 Ma)
Northern Calcareous Alps
N-Slovenia
Malé Karpaty Mts
Danube
Bakony
Zala
Gerecse
TRL
HDZ
Ve
Na
Má
Bö
NL
Buda line
Danube
Drava
Sitno
Central Carpathian Palaeogene
Recsk
Bükk
South Bükk
Tisza
Szolnok flysch trough
Transylvanian Palaeogene basin
ALCAPA
TISZA-DACIA
100 km
Legend
anticline, syncline, thrust fault
large/small strike-slip fault
large/small normal fault
exhumation of metamorphic rocks and low angle normal fault
young uplift
basalt cone and dyke
HDZ Hurbanovo-Diósjenő zone
Flysch Belt
Pieniny Klippen Belt
Main depocentres
Volcanic rocks
Bö, Ve Budaörs, Velence
TRL Telegdi-Roth line
Má, Na, Mány, Nagyegyháza basin
NL Nagykovácsi line
16°
18°
20°
22°
24°
46°
48°

Fig 4a

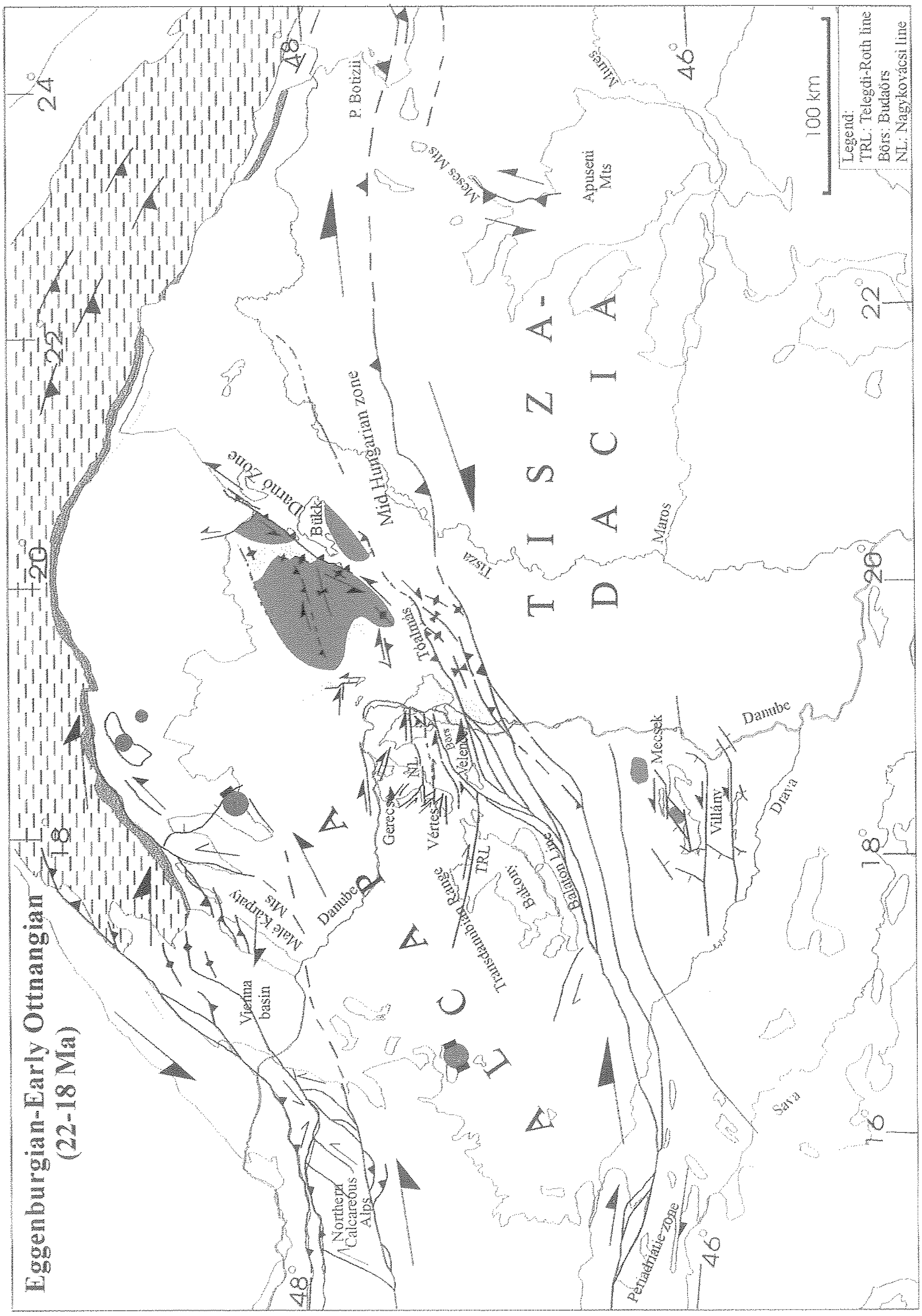

Eggenburgian-Early Ottnangian
(22-18 Ma)
Legend:
TRL: Telegdi-Roth line
Börs: Budaörs
NL: Nagykovácsi line
100 km
Northern Calcareous Alps
Vienna basin
Male Karpaty Mts
Danube
Gerecse
Vértes
Transdanubian Range
Bakony
TRL
NL
Börs
Velence
Balaton Line
Tóalmás
Bükk
Darnó Zone
Mid Hungarian zone
Tisza
Maros
Mecsek
Villány
Drava
Sava
Periadriatic zone
Apuseni Mts
Meses Mts
P. Botizii
Mures
A L C A P A
T I S Z A -
D A C I A
48°
46°
16°
18°
20°
22°
24°

Fig 4b

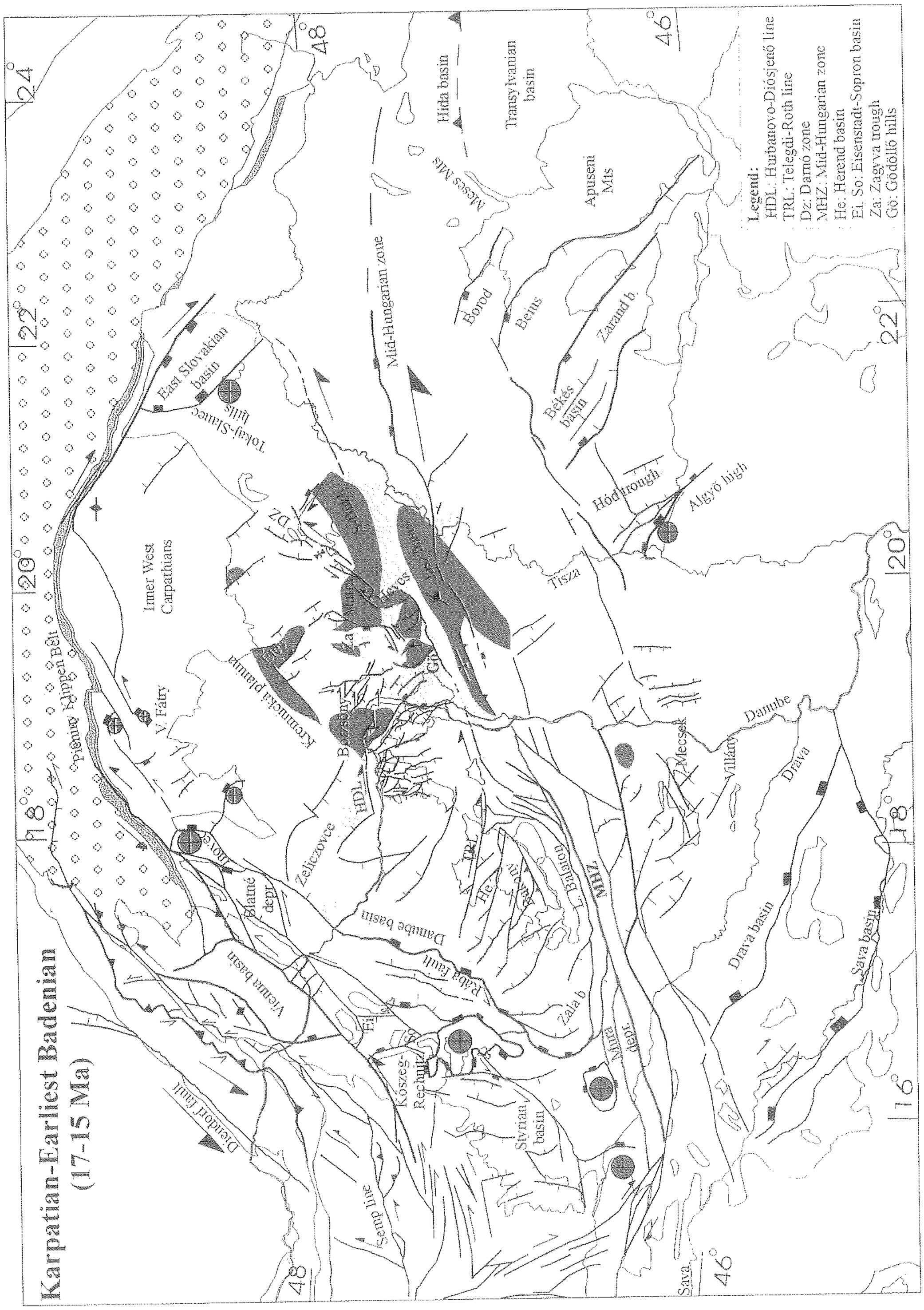

Karpatian-Earliest Badenian
(17-15 Ma)
Legend:
HDL: Hurbanovo-Diósjenő line
TRL: Telegdi-Roth line
Dz: Darnó zone
MHZ: Mid-Hungarian zone
He: Herend basin
Ei, So: Eisenstadt-Sopron basin
Za: Zagyva trough
Gö: Gödöllő hills
Pieniny Klippen Belt
Inner West Carpathians
East Slovakian basin
Tokaj-Slanec hills
Mid-Hungarian zone
Transylvanian basin
Hida basin
Meses Mts
Apuseni Mts
Borod
Beius
Zarand b.
Békés basin
Hód trough
Algyő high
Tisza
Danube
Drava
Drava basin
Sava basin
Sava
Mecsek
Villány
Vienna basin
Danube basin
Rába fault
Zala b
Mura depr.
Styrian basin
Kőszeg-Rechnitz
Semp line
Diendorf fault
Blatné depr.
Zeliezovce
Kremnica planina
V. Fátry
L. Balaton
MHZ
TRL
HDL
Dz
Börzsöny
Mátra
Bükk
Cserhát
48
46
24
22
20
18
16

Fig 4c

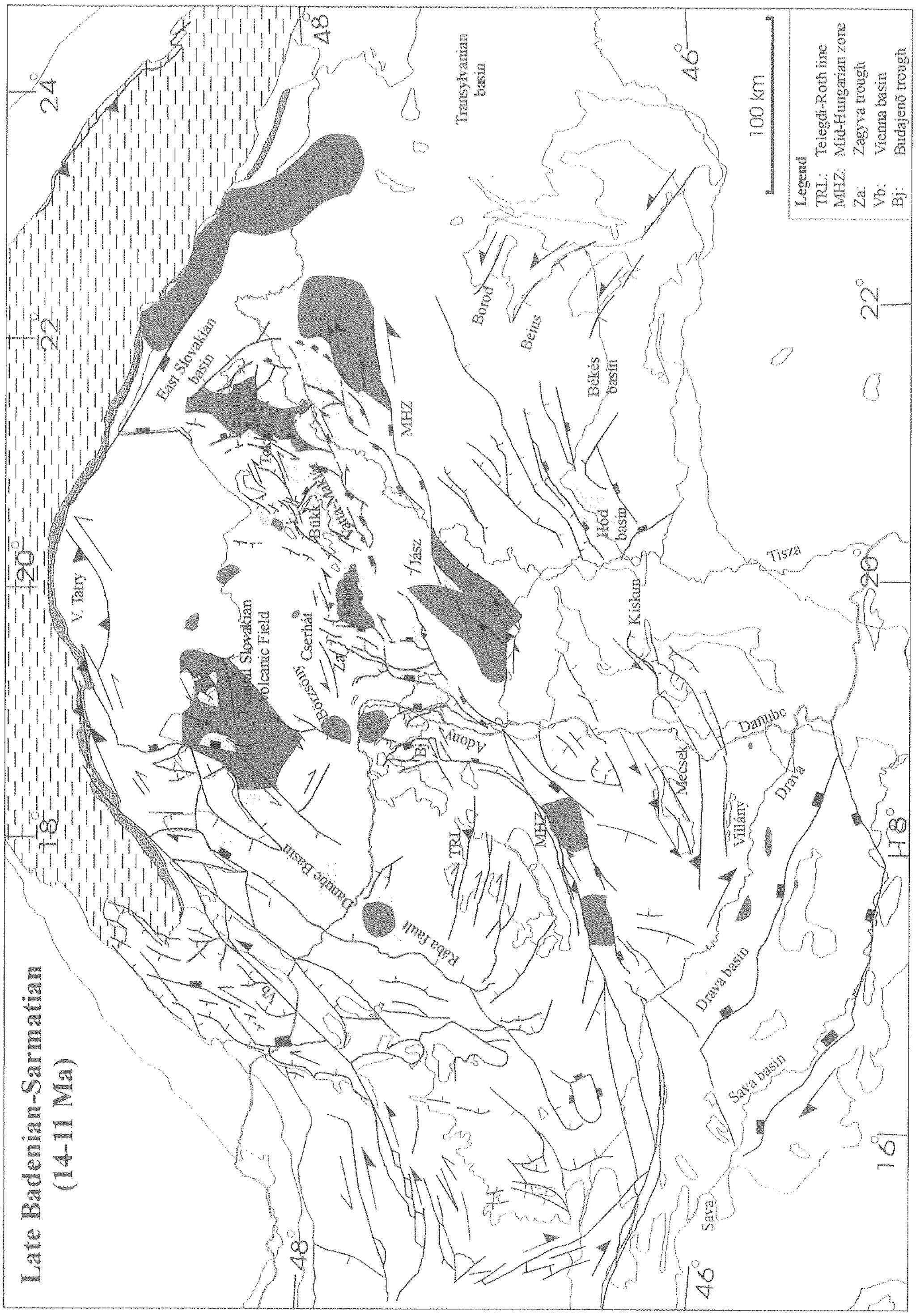

Late Badenian-Sarmatian
(14-11 Ma)
Legend
TRL: Telegdi-Roth line
MHZ: Mid-Hungarian zone
Za: Zagyva trough
Vb: Vienna basin
Bj: Budajenő trough
100 km
Transylvanian basin
East Slovakian basin
Central Slovakian Volcanic Field
V. Tatry
Danube Basin
Rába fault
Cserhát
Börzsöny
Mátra
Bükk
Tokaj
Jász
MHZ
TRL
Bj
Adony
Kiskun
Hód basin
Békés basin
Borod
Beius
Mecsek
Villány
Drava
Drava basin
Sava basin
Sava
Danube
Tisza
Vb
Za
48°
46°
16°
18°
20°
22°
24°

Fig 4d

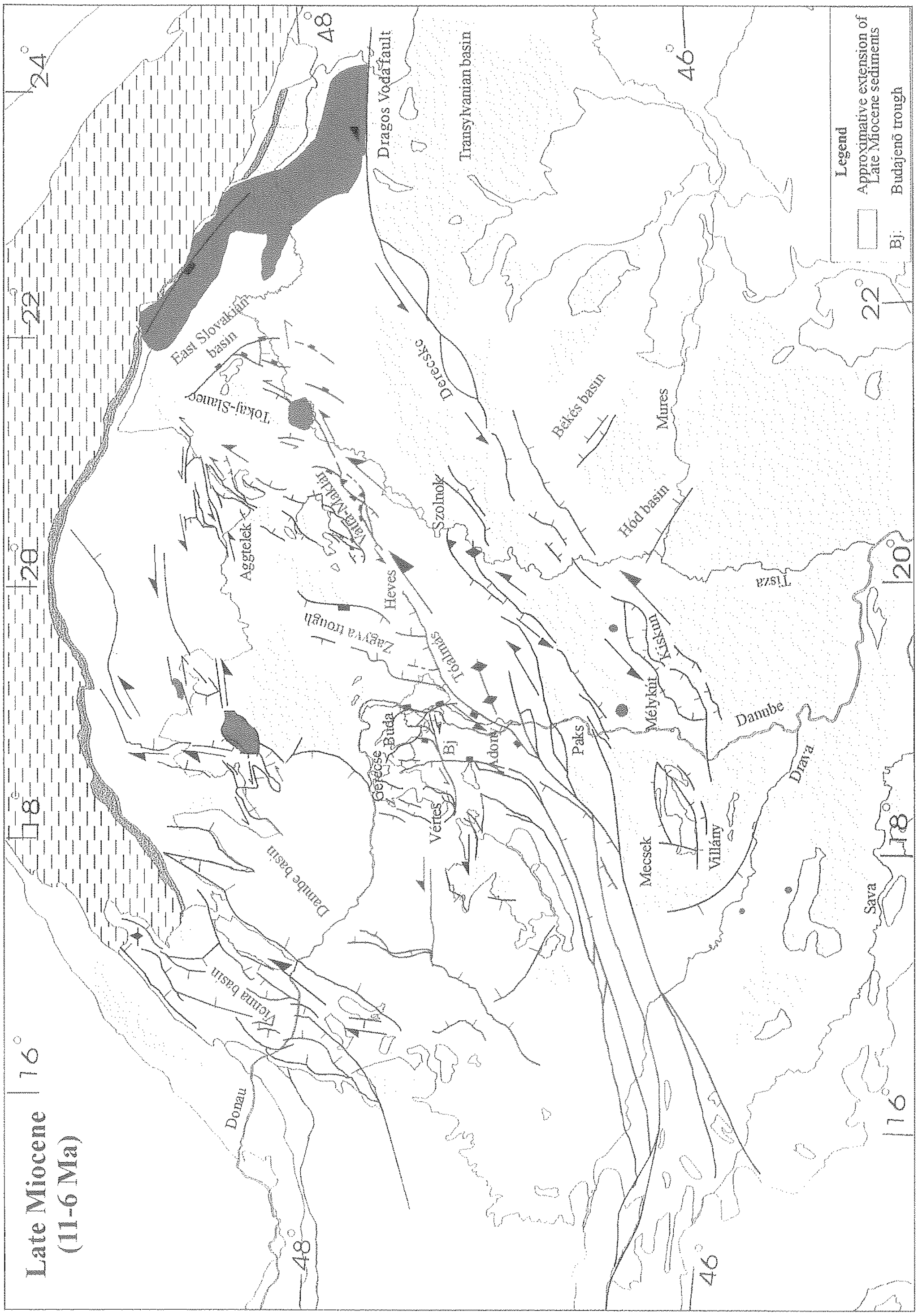
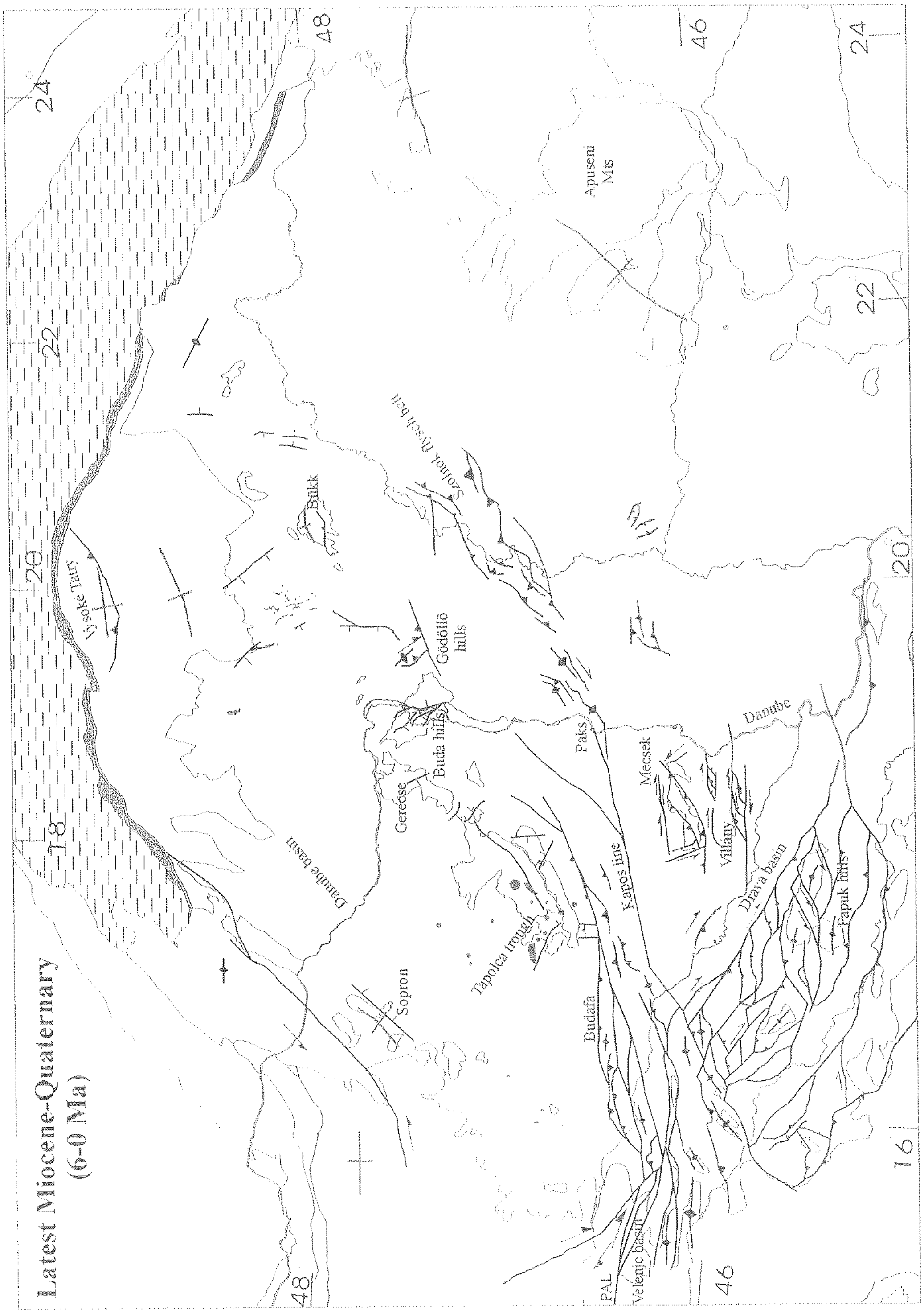

Late Miocene
(11-6 Ma)
Legend
Approximative extension of Late Miocene sediments
Bj: Budajenő trough
Dragos Vodă fault
Transylvanian basin
East Slovakian basin
Tokaj-Slanec
Derecske
Békés basin
Mures
Szolnok
Aggtelek
Hód basin
Heves
Zagyva trough
Tisza
Kiskun
Mélykút
Danube
Paks
Adony
Buda
Bj
Gerecse
Vértes
Mecsek
Villány
Drava
Sava
Danube basin
Vienna basin
Donau
16°
18°
20°
22°
24°
46°
48°

Fig 4e



Latest Miocene-Quaternary
(6-0 Ma)
Vysoke Tatry
Bükk
Szolnok flysch belt
Gödöllő
hills
Buda hills
Gerecse
Danube basin
Sopron
Tapolca trough
Budafa
Kapos line
Paks
Mecsek
Villány
Drava basin
Papuk hills
Danube
Apuseni
Mts
PAL
Velenje basin
48
24
22
20
18
16
46

Fig 4f

(1992) and Haas *et al.* (1995). The former unity of the Slovenian and the North Hungarian Palaeogene Basins was inferred by Báldi (1986) and Nagymarosy (1990).

The strong deformation by subsequent Miocene phases overprints the Early Miocene internal structure of the Mid-Hungarian shear zone. The stress data are, however, in accordance with dextral slip along this fault system. Some other kinematic data are known along the northern border of the shear zone. South of the Velence hills dextral faults were mapped in the subsurface which displaced Eocene rocks (Dudko *et al.* 1989). These structures are supposed to continue in the Tóalmás zone (Csontos & Nagymarosy 1998) which determines the location of an Egerian–Eggenburgian delta (Sztanó 1994). South of the Lake Balaton exotic duplexes of the Palaeogene Basin were found (see on Fig. 4a, after Sztrákos 1975; Balla *et al.* 1987; Nagymarosy 1990; Kőrössy 1990). These duplexes were formed during the separation of the formerly unique Hungarian–Slovenian Palaeogene Basin. Both surface observations (in Slovenia) and borehole data (in Hungary) suggest that the Slovenian and Hungarian Basin fragments were separated before the Karpatian, due to Early Miocene dextral slip (Balla & Dudko 1989; Fodor *et al.* 1998).

Within the inner part of the Transdanubian Range the activity of E–W trending dextral faults was demonstrated. In the Buda and Gerecse hills dextral slip reactivated Eocene faults (Fodor *et al.* 1994; Bada *et al.* 1996). Such faults separate areas with thick and thin upper Oligocene series, suggesting younger, Early Miocene displacement (Nagykovácsi line, Balla & Dudko 1989). Upper Oligocene sandstones were dragged into tectonic lenses within fault zones during Early Miocene slip (Gyalog 1992). Normal faults belonging to later rifting phase displace some dextral strike-slip faults, thus their activity ended by the Karpatian.

In the Western Carpathians NW–SE compression influenced the Eggenburgian sedimentation (Kováč *et al.* 1989). In fact, the main boundary faults of the depressions seem to reactivate Oligocene transpressional structures, but a marked discordance indicates Early Miocene reactivation, too (Kováč *et al.* 1990).

In contrast to dextral transpression, different kinematics occurred along the Darnó-zone. There anticlines and reverse faults were mapped in Egerian to Eggenburgian rocks and demonstrated on seismic lines (Fig. 4b), (Telegdi-Roth 1951; Sztanó & Tari 1993). In addition to shortening, the en echelon geometry and oblique orientation of structures to the compression may suggest slight sinistral component of slip. The deformation is well-dated along the zone. The first step might be indicated by the erosion of the thick Kiscellian series in the hanging wall of the reverse faults before the beginning of Eggenburgian sedimentation, during late Egerian (Báldi 1986; Nagymarosy 1990). On the downthrown side the depositional geometry of the Eggenburgian sediments suggest that the zone represented the easternmost boundary of an embayment (Sztanó 1994). The tectonically controlled shore is marked by talus cones, fan deltas, liquefactions and gradual displacement of the source area of the fan deltas (Sztanó & Józsa 1996). In the northern part, similar mismatch between the source area and pebble material of deltas was demonstrated (Szentpétery 1988). Slices of Eggenburgian rocks incorporated into faults suggest post-sedimentary continuation of the motion. However, the bulk of the displacement predates the first rhyolite tuff horizon (19 Ma; Radócz 1966) and the Ottnangian coal series seals the compressional structures.

The Tisza–Dacia unit: Eocene through Early Miocene compression

The earliest Tertiary deformations in the Tisza–Dacia block were found in northern Transylvania. Here Eocene to Early Miocene rocks are exposed along the margins of Mesozoic to crystalline outcrops. These sediments are considerably deformed at the marginal parts, while they are flatlying towards the centre of the basin. The deformed outcrops show polyphase tectonism, first characterized by an ENE–WSW to NE–SW compression, followed by a NNW–SSE to NW–SE compression (Fig. 5). Thrust faults are dominant, but strike-slip faults are also abundant. In some outcrops folds are also seen. Eocene to Early Oligocene rocks are affected by both phases, while Early Miocene rocks record only the second phase. Local map-scale structures range from folds to thrust faults. Two

Fig. 4. Structural evolution and basin formation in the Pannonian Basin and adjacent mountain ranges through six episodes. Structural geometry is more detailed in the Pannonian part and schematic within the Alpine and Carpathian ranges. Main sedimentary and volcanic areas are stippled. References for structural geometry are discussed in the text.

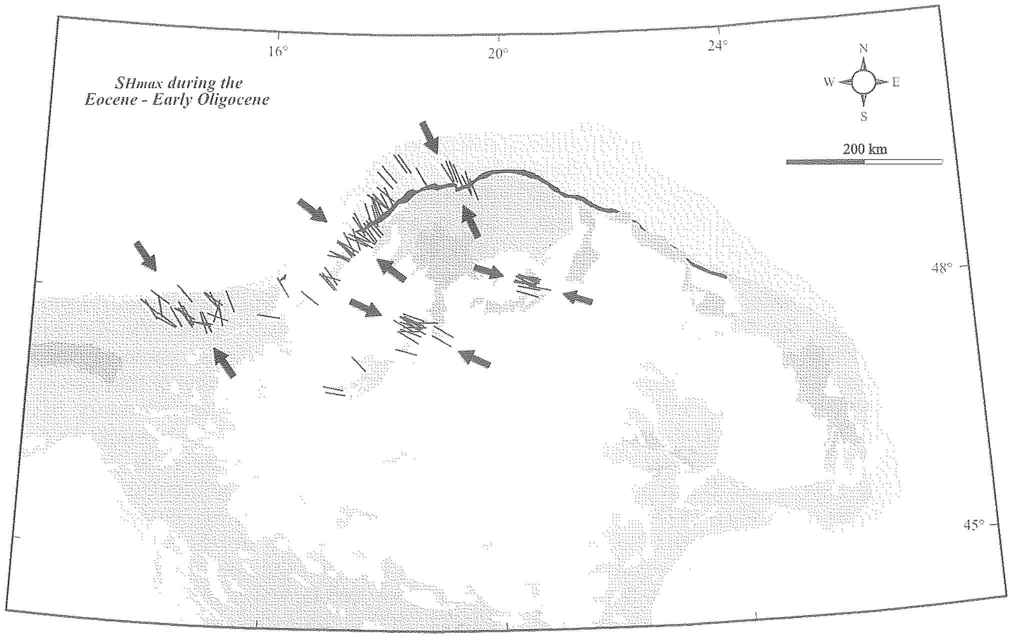

16°
20°
24°
SHmax during the
Eocene - Early Oligocene
N
W
E
S
200 km
48°
45°

Fig. 5(a)

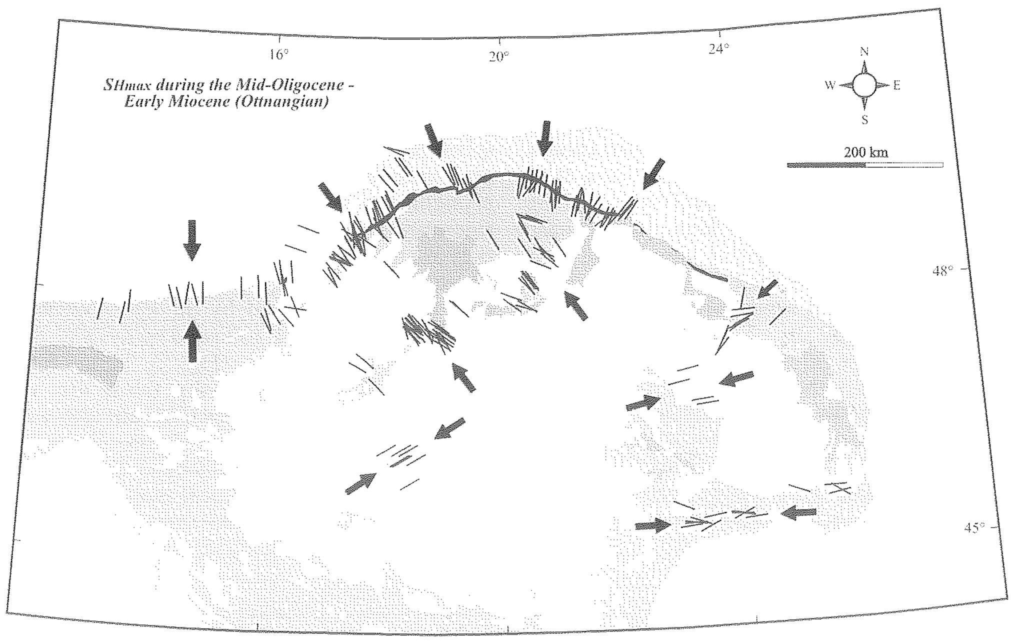

16°
20°
24°
SHmax during the Mid-Oligocene -
Early Miocene (Ottnangian)
N
W
E
S
200 km
48°
45°

Fig. 5(b)

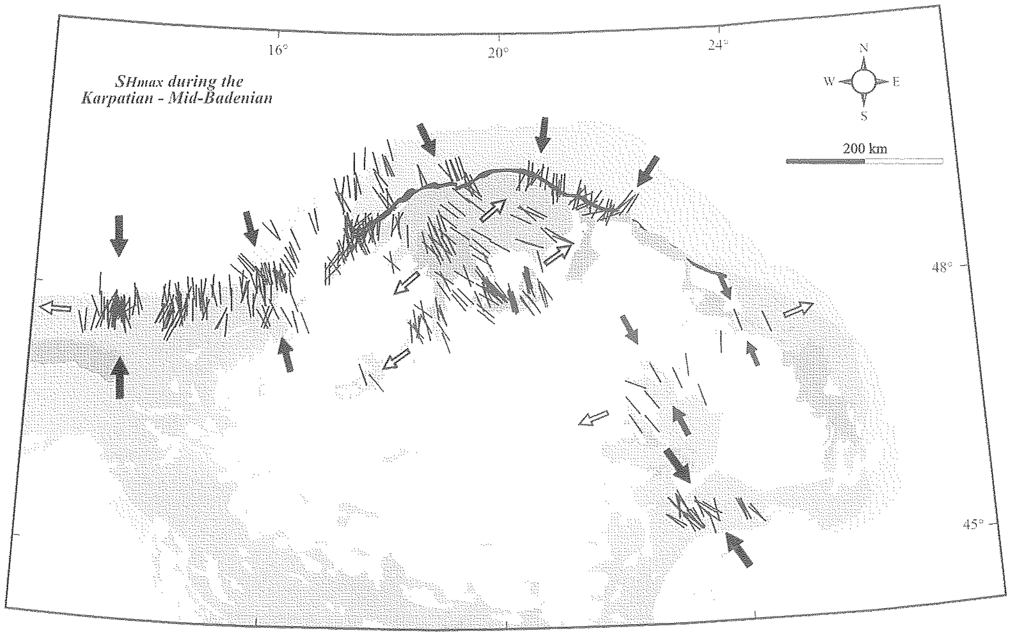

SHmax during the
Karpatian - Mid-Badenian
16°
20°
24°
N
W
E
S
200 km
48°
45°

Fig. 5(c)

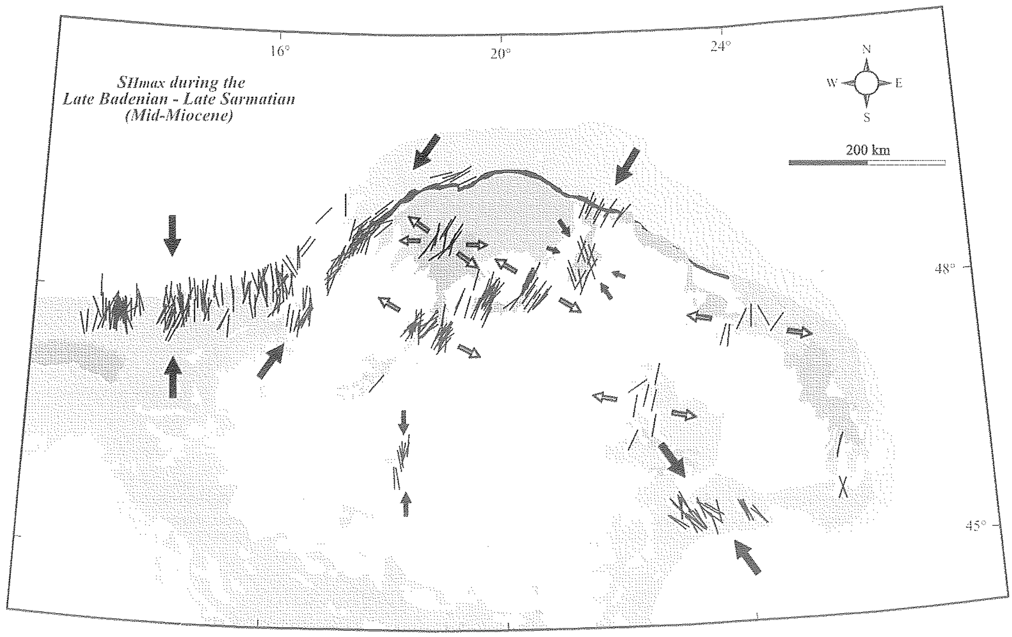

SHmax during the
Late Badenian - Late Sarmatian
(Mid-Miocene)
16°
20°
24°
N
W
E
S
200 km
48°
45°

Fig. 5(d)

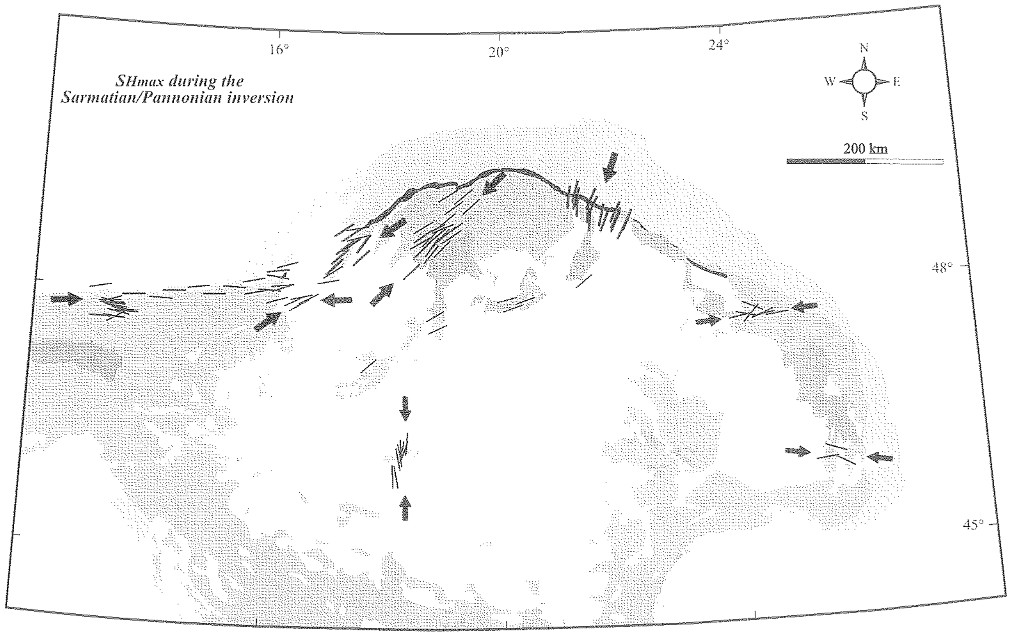

SHmax during the
Sarmatian/Pannonian inversion
16°
20°
24°
N
W
E
S
200 km
48°
45°

Fig. 5(e)

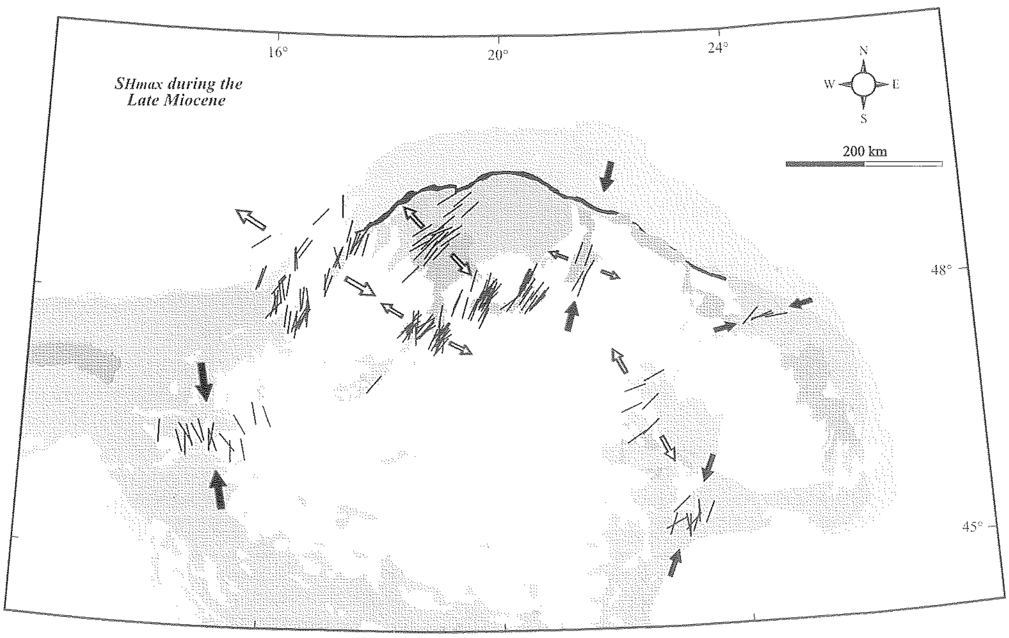

SHmax during the
Late Miocene
16°
20°
24°
N
W
E
S
200 km
48°
45°

Fig. 5(f)

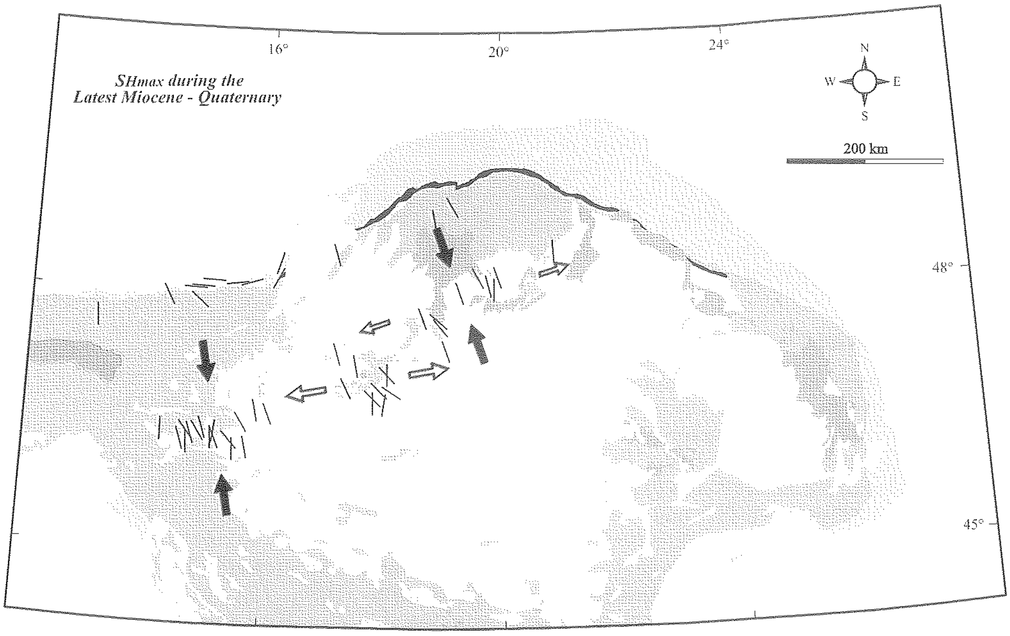


Fig. 5(g)

Fig. 5. Stress data of the Pannonian–Alpine–Carpathian region. Only the maximal horizontal stress axes are represented. Large arrows indicate the general trend of stress type (compressional, strike slip, tensional). Geological outlines after Mahel *et al.* (1984). Sources: Decker *et al.* (1993); Fodor (1995); Nemčok (1993); Nemčok *et al.* (1993); Nemčok & Lexa (1990); Ratschbacher *et al.* (1993*a*, *b*); Fodor *et al.* (1992, 1994); Vass *et al.* (1993); Csontos & Bergerat (1992); Györfi *et al.* (this volume); and own unpublished data.

major thrusts are known: the Meses thrust putting Late Cretaceous granites on top of overturned Eocene, deposits and the Botiza area, where an imbricate sheet of Mesozoic to Eocene rocks is thrust upon Late Oligocene–Early Miocene rocks. The Meses thrust is an oblique transpressional zone related to the first stress-field, while the Botiza imbricate stack was created by the second deformation (Fig. 4a, b).

Timing of the deformation is partly ascertained using superposition, partly from rocks not showing earlier fault sets. Dating was complemented by considering the sedimentological features of the Transylvanian and Maramures basins (Györfi *et al.* this volume). We inferred that the NW–SE-oriented Transylvanian Palaeogene Basin was created by compression. One potential thrust responsible for the flexure of the basin is found at the Meses Mts, while other supposed faults remain hidden beneath the thick Mid- to Late Miocene fill of the Transylvanian Basin. An Early Oligocene age was suggested for this deformation. The E–W-oriented Late Oligocene–Ottnangian turbiditic Borsa and Early Miocene Ottnangian–Karpatian coarse clastic Hida Basins were supposedly created by the Botiza thrusts and the second, NNW–SSE compressional phase. The age of this deformation should be therefore Latest Oligocene to Early Miocene.

Rifting of the Pannonian Basin: latest Early to Mid-Miocene (Figs 4c, d, 5c, d)

Easternmost Alps

Late Early and Mid-Miocene times represent the major rifting period in the Pannonian region (Horváth & Royden 1981; Tari 1994). The onset of rifting corresponds to a pronounced change in stress field and tectonic evolution was complex in different sub-basins. In contrast, the easternmost Alps experienced a relatively simple evolution: the same stress field established for the latest Oligocene and earliest Miocene seems to continue during the rifting of the Pannonian Basin. In the Eastern Alps σ_1 was around N–S and σ_3 was typically horizontal (Figs 4c, d, 5c, d). However, the earlier frontal thrusting onto the European foreland ceased and the new structures are conjugate strike-slip faults and related transtensional and transpressional features (Ratschbacher *et al.* 1989, 1991*a*, *b*; Decker *et al.* 1994; Linzer *et al.* 1995; Nemes *et al.* 1995*a, b*). The main subsidence is Karpatian–early Badenian within the large Styrian (Ebner & Sachsenhofer 1995) and other small pull-apart basins (Ratschbacher *et al.* 1991*a*).

Pannonian Basin–Carpathians

East of the Vienna and Styrian Basins, the Pannonian Basin system and the internal part of the Carpathian orogen were characterized by a tensional, locally transtensional stress field. The direction of the minimal stress axis (σ_3) varied in time and space between NE–SW and NW–SE (or even NNW–SSE). These varying directions can be generally separated into two distinct events. NE–SW- to ENE–WSW-tension-generated structures, which are systematically older than those characterized by E–W to SE–NW tension. Recent research by Fodor *et al.* (in prep.) in North Hungary demonstrates that the change between the two stress fields did not occur at the same time in different basins, thus the stress field remained inhomogeneous during the Mid-Miocene. First, we describe these two stress fields and then discuss their timing within the sub-areas.

NE–SW to ENE–WSW tension: stress field and fault characteristics (Figs 4c, d, 5c, d). The first phase of rifting was characterized by NE–SW to ENE–WSW tension or strike-slip type stress field with σ_{Hmax} NNW–SSE to NW–SE throughout the whole Pannonian area (Fig. 5c). The main structures are normal and normal-oblique faults with orientations varying from NW–SE to NNE–SSW, but the main direction changes from one sub-basin to another. E–W- to WNW–ESE-trending dextral and N–S- to NE–SW-trending sinistral faults are also common.

In the area of the Danube Basin and its southeastern rim, the normal faults are trending NW–SE or NNE–SSW. The larger ones are low-angle, determined by seismic sections (Tari *et al.* 1992). Fission track ages from the Rechnitz and Inovec hills demonstrate that exhumation of the footwall and low-angle normal faulting were coeval (Dunkl & Demény 1997; Kováč *et al.* 1994). Transition from ductile to brittle deformation around 20 Ma (Ratschbacher *et al.* 1990) also corroborates this relationship. These data argue for the formation of the Rechnitz metamorphic core complexes due to simple shear extension (Tari *et al.* 1992; Györfi 1992; Tari 1996). Although the connection of the exhumation of metamorphic rocks and low-angle normal faulting was not clearly demonstrated at other locations of the study area, we tentatively suggest the existence of several similar core complexes (Fig. 4c).

High-angle normal faults often border a series of tilted blocks or synsedimentary half grabens, mainly trending NW–SE. Such blocks constitute the Zala Basin, the southern and central Bakony, the main part of the Gerecse and Buda hills, the subsurface Gödöllő area, the Etes graben and a number of depressions within the Western Carpathians (Fig. 4c). NW-trending border faults limit the depressions between the Bükk hills and the Tokaj hills. The Eastern Slovakian Basin itself may also be such a large graben, although moderately dipping boundary faults might have dextral component too (Tomek & Thon 1988; Kováč *et al.* 1995). High-angle normal faults probably flatten to low-angle detachment faults, like near the Rechnitz windows (Györfi 1992; Tari 1996).

The system of large, NW–SE-oriented normal faults and rotated half-grabens continued south of the Mid-Hungarian zone, both in the buried areas and in the exposed ranges. Larger and very deep basins are found in SE Hungary, in the Danube–Tisza interfluve and immediately east of the Tisza river (Posgay *et al.* 1995). Here the Hód and Békés Basins (Fig. 4c) reach at present 7 km depth (however large part of the subsidence occurred during the post-rift phase). Basins of similar orientation and bordered by NW–SE normal faults flank the western Apuseni Mts (Györfi 1993). These basins are bordered by NE-dipping low-angle normal faults (Györfi & Csontos 1994), the footwalls of which were uplifted synchronously (fission track data of Dunkl pers. comm. 1993; Posgay *et al.* 1995). In the southwestern Pannonian Basin, the Drava and Sava grabens can also belong to the system of tilted blocks (Prelogović *et al.* 1995; Tari & Pamić 1998).

Border faults often have strike-slip or oblique-slip thus the depressions have transtensional character (Fig. 4c). One of the best documented example is the Vienna Basin where fast subsidence occurred between sinistral and normal border faults of a rhombohedral pull-apart depression (Royden 1985; Fodor 1995). The sinistral nature of the southeastern border faults is also indicated by displaced talus cones (Vass *et al.* 1988*b*). In the Southern Carpathians, near the margin of the Tisza–Dacia block E–W trending dextral shear zone developed (Ratschbacher *et al.* 1993*b*; Rabagia & Fülöp 1994).

Within the Pannonian Basin, some other fault zones had strike-slip or oblique-slip character. A set of en echelon half grabens are situated within the sinistral Darnó shear zone of North Hungary (Bergerat *et al.* 1984; Márton & Fodor 1995). The Bakony hills are dissected by WNW–ESE-trending dextral faults (Mészáros & Tóth 1981; Mészáros 1982) which border small Miocene Basins near Herend and Várpalota (Kókay, 1976; Balla & Dudko 1989). The western segment of the Hurbanovo zone could also have dextral character, suggested by nearby map-scale faults (Bence *et al.* 1991) and fast subsidence in the Zeliezovce sub-basin (Lankreijer *et al.* 1995).

E–W to SE–NW tension (Figs 4d, 5d). The second phase of rifting is mainly characterized by tensional or strike-slip type stress field. The direction σ_3 is variable, from E–W to SE–NW (Fig. 5d). This difference is interpreted as variation within a slightly inhomogeneous stress field. This stress field created new faults but often reactivated with different kinematics the faults of the first rifting phase. The NW–SE-trending faults became dextral strike-slips, the NNE–SSW-trending strike-slips changed to normal or normal-oblique faults. One of the best documented examples is the Vienna Basin which lost its pull-apart character when the stress field and, thus, the strike-slip kinematics of the border faults changed (Fig. 4c,d; Fodor 1995). A similar phenomenon was demonstrated in basins of the westernmost Carpathians (northern branches of the Danube Basin) which were essentially bordered by N–S to NE–SW-striking normal faults (Kováč *et al.* 1990; Kováč & Baráth 1995).

Faults of the second phase are clearly superimposed on the first generation from the Gerecse to Bükk hills (Fodor *et al.* in prep). The separation along the NW–SE-trending older faults decreased, they are reactivated as dextral oblique faults or are cut by the new, N–S- to NNE–SSW-trending fault sets (Fig. 4d). The slip along the border faults resulted in a second tilt episode in ESE or WNW directions.

Along the Mid-Hungarian shear zone depressions are mainly trending ENE–WSW. Seismic sections show a number of criteria which is typical for strike-slip fault. Tari (1988) used such criteria and demonstrated sinistral strike-slip faults in the northern Vatta-Maklár trough which is in good agreement with surface measurements (Csontos 1988; Márton & Fodor 1995).

The system of ENE-trending sinistral strike-slip faults and rhomb-shaped depressions frequently occur in the southern Pannonian Basin. Their pull-apart origin was already suggested by earlier publications (Rumpler & Horváth 1988).

In the western side of the Apuseni Mts the NW–SE extensional stress-field did not create new major normal faults but the older NW–SE faults were reactivated as strike-slip faults. Microtectonic data are recorded in Badenian sediments (Györfi 1993).

In the Mecsek hills a strike-slip type stress field was recorded in Early Miocene rocks. It is characterized by N–S σ_1 and perpendicular σ_3 directions. Major structures are NE–SW sinistral and NNW–SSE dextral faults. Sinistral faults are connected to E–W en echelon folds (Benkovics 1997).

Timing and tentative separation of rifting phases

In this section we analyse the time constraints for the onset of the first rifting episode and its change to the second tensional stress field. This overview is presented from west to east, starting from Vienna Basin. During the Karpatian and Early Badenian continued thrusting in the northern Outer Carpathians permitted the sinistral opening and displacement transfer from its southeastern to the northwestern boundary fault. However, from the middle Badenian onward, thrusting ceased in the western sector of the Western Carpathians (Jiříček, 1979), thus sinistral shift was maintained only along the SE boundary fault. The change in stress field occurred contemporaneously with this migration of thrust activity, after Early Badenian (Fodor 1995; Marko *et al.* 1995).

The Ottnangian initiation of sedimentation around the Rechnitz windows is in good agreement with the time of its exhumation (20–18 Ma). The formation of the Styrian and Sopron basins can thus be connected to the first phase of rifting (Tari 1994). In fact, at the western edge of the Danube Basin, only the Ottnangian sediments seem to be affected by ENE–WSW tension, the lower Badenian rocks show only the effect of the younger stress fields (Fodor 1995). A pronounced unconformity occurs within the Badenian series of the Danube and Zala Basins, on the hanging wall of the large low-angle detachment fault above the Rechnitz windows (Horváth *et al.* 1995). This unconformity can be correlated to the change in the stress field during the middle Badenian.

Stress field change is well dated in the central and northeastern Pannonian Basin and Western Carpathians. The oldest, synsedimentary activity is indicated by different thickness and facies development within the Darnó zone and the Etes trough, starting already in the Ottnangian (Hámor 1985; Márton & Fodor 1995). Differential subsidence continued through Karpatian and earliest Badenian. Early Badenian initiation of sedimentation is somewhat younger in the Zeliezovce depression of the Danube Basin (Kováč *et al.* 1993; Lankreijer *et al.* 1995). Outside these zones, the initiation of the NW–SE grabens are less well constrained, because they are superimposed on Egerian to Eggenburgian rocks and Ottnangian to Karpatian strata are missing.

The earliest indication of change in the stress field occurs in the subsurface southern Gerecse and Buda hills (Fig. 4c, d). Here Karpatian(?)–Lower Badenian clastics are situated in grabens which are discordantly covered by Middle to Upper Badenian sediments (Kókay 1989). These latter sediments were already deposited in new N–S-trending grabens (Kókay 1990). Thus the rotation of tension from NE to E had already occurred before or within the Middle Badenian (15–14 Ma).

Synsedimetary tectonism of the neighbouring Budajenő trough and south Buda hills give also good time constraint for the onset of the second tension. Both the Upper Badenian and Sarmatian limestones are dissected by synsedimentary dykes and microfaults (Bergerat *et al.* 1983, Bada *et al.* 1996). Sediment transport direction points in each case to the direction of gravity lows, from the master faults or tilted block edge toward basin centres. This suggests that the shape of the depressions were formed by late Badenian tectonism and kept during the Sarmatian. Considerable thickness variation of near-shore limestone versus basin marl indicate amplification of master faults during Sarmatian sedimentation.

Eruptions of voluminous stratovolcanoes of the Börzsöny, Mátra hills and the Kremnicka Planina highland started during Early Badenian (Konečny & Lexa 1984; Vass *et al.* 1979) and can be connected to the first rifting episode. Some sites of these volcanoes are still deformed by the earlier NE–SW tension (Bence *et al.* 1991; Kováč & Hók 1993; Vass *et al.* 1993) while other sites bear the trace of only the younger deformation. The 15 Ma Karancs subvolcanic andesite body (Orlicky *et al.* 1995) is deformed by NE–SW tension. Intrusion of the NW–SE-trending dykes of the Cserhát (14 Ma, Pécskay *et al.* 1995) may have also been facilitated by NE–SW tension. However, they are dragged by E–W sinistral faults in a ductile manner immediately after intrusion, before complete cooling (Balla 1989). Looking to the coeval sediments, the latest Early Badenian limestone ('lower Leithakalk') has the trace of both stress field, while Upper Badenian to Sarmatian limestones were already affected only by the second tension.

Seismic sections (Benkovics 1991; Tari *et al.* 1992) demonstrate that the Zagyva and Tura master faults were active during the late Badenian–Sarmatian sedimentation (Fig. 4d). Borehole data and surface thickness and facies (Hámor 1985) suggest that late Badenian to Sarmatian sediment wedges are progressively thickening toward the master faults while their facies is generally changing from carbonate rocks on the edge of the tilted blocks to (fine) clastics basinward. All these data suggest that between the Gerecse hills and Darnó zone the change in stress field can be placed at the beginning of or within Late Badenian (14–13 Ma).

In the central Western Carpathians NE–SW tension was replaced by E–W tension during the Late Badenian and Early Sarmatian. This tension changed again to NW–SE tension at the end or after the Sarmatian (Nemčok & Lexa 1990; Hók *et al.* 1995). The first change of tensional direction is coeval with similar changes just south of this region, in North Hungary.

The timing markedly changes in the Tokaj–Slanec hills. Here the stress evolution has three stages. The maximal horizontal stress axis of the first event was oriented WNW–ESE, then NNW–SSE, and finally NNE–SSW. The first stress field was strike-slip type, while the two younger were tensional ones. The lower part of the volcanic suite (upper Badenian to lower Sarmatian) was deformed by three different stress fields, while the upper volcanic suite (upper Sarmatian–lowermost Pannonian) was deformed only by the youngest tension. Consequently, a change in stress field occurred during the first half of the Sarmatian, and another one only at the end of this period. Similar statements can be made using dykes: the older ones are trending NW–SE, while progressively younger ones changed in direction to N–S then to NE–SW (Gyarmaty *et al.* 1976; Molnár 1994). One uppermost Sarmatian site shows synsedimentary faults trending NE, fitting to a latest Sarmatian onset of the youngest, NE–SW tension.

Palaeomagnetic data suggest 30–35° CCW rotation which occurred within the Sarmatian (Márton & Pécskay 1995; Orlicky 1996). We tentatively correlate this rotation with the older change of stress field. This change can be only apparent, because by eliminating the rotation, the measured WNW–ESE maximal stress axes turned back to original NNW–SSE. Then the same NNW–SSE compression and perpendicular tension continued, without rotation. The best structural expression of the stress field is the fast subsidence along NW–SE faults in the neighbouring East Slovakian Basin (Fig. 4d), (Vass *et al.* 1988*a*; Tomek & Thon 1988; Rudinec 1989; Kováč *et al.* 1995).

All data suggest that the first rifting phase terminated gradually from the western to the northeastern Pannonian Basin. The transition occurred at the beginning of or during the Mid-Badenian west of the Danube, during the Late Badenian east of the Danube, around the Zagyva graben, and only in late Sarmatian in the Tokaj–Slanec hills and East Slovakian Basin.

Within the Tisza–Dacia unit no such detailed chronology can be given. In seismic sections of the Great Hungarian Plain the change from NE–SW tension to NW–SE tension can be put somewhere in the Badenian. Syn-rift deposits in NW–SE-oriented grabens are pre-Mid-Badenian in age, while syn-rift deposits of NE–SW tilted grabens and pull-apart sub-basins are Badenian (Györfi 1993; Györfi & Csontos 1994). In the western flanks of the Apuseni Mts the change in the stress-field orientations occurs in the Mid-Badenian. Older deposits are affected by both tensional phases, while Mid- to Late Badenian and Sarmatian sediments are affected by NW–SE and NNW–SSE tension (Györfi 1993).

Early stage basin inversion (earliest Late Miocene) (Fig. 5e)

ENE–WSW or E–W compression at about 10 Ma was determined in several sites of the Eastern Alps (Decker *et al.* 1993; Linzer *et al.* 1995; Peresson & Decker 1997). The stress field was clearly demonstrated and separated from other phases in the eastern border of the Vienna Basin (Fodor *et al.* 1990).

Some conjugate strike-slip faults in Hungary are also characterized by NE–SW to ENE–WSW oriented σ_1 (Fig. 5e). The demonstration is relatively clear in sites where both the ESE–WNW tension and ENE–WSW compression are present (Fig. 3.). Where only NE–SW oriented σ_1 was calculated, the faults might be attributed to this phase but they may also be considered as local deviation of the Mid-Miocene ESE–WNW tension. In this case, the occurrence of such stress data would be attributed to variation in magnitudes of the stress axes (change from tensional to strike-slip or even compressional stress field). The main microtectonic features are the inversion of sense of NNE–SSW-trending strike-slip faults, from sinistral to dextral. However, this inversion was never large enough to inverse offsets on map-scale. Steep reverse faults are sometimes associated to these strike-slip faults.

We ranged such determinations into an individual inversion phase (with ESE–WNW σ_1) only when additional time constraints, mainly relative timing with respect to other stress fields were present. Such examples from South Bükk occurred obviously in tilted position, which suggest a pre-Pannonian age. In the Cserhát this phase occurs after the bulk of tilting (Middle Miocene, Fig. 3.). This inversion phase is relatively clearly visible on seismic sections in the southern Pannonian Basin and south of Budapest, where inversion of the former basins occurs (Horváth 1995). This inversion is coupled by the erosion of the Sarmatian deposits and non-deposition of the Early Pannonian on the emerged blocks.

In the Eastern Alps this phase was determined only in Mesozoic and Senonian rocks thus the time constraint is not really good (Decker *et al.* 1993; Peresson & Decker 1997). In the eastern Vienna Basin and northern Hungary, Badenian and occasionally Sarmatian sediments were also deformed (Fodor *et al.* 1990). The relative chronology of this stress field with respect to mid-Miocene stress field and tilting would suggest post-Mid-Miocene age in the central Pannonian area. The general lack of this deformation in Upper Miocene rocks and some relative chronological criteria with respect to Late Miocene deformation, bracket this phase around the turn of Mid- and Late Miocene, at latest Sarmatian or earliest Pannonian. However, some sites might allow other timing as well, e.g. a site in the eastern Vienna Basin would suggest considerably younger, Pontian age (Sauer *et al.* 1992; Peresson & Decker 1997).

In the eastern Pannonian Basin and western Transylvanian Basin the orientation of σ_1 did not change during the late Mid-Miocene–Late Miocene, thus this event does not appear as a separate phase. Also, in the central Western Carpathians the NE–SW maximal horizontal stress axes (strike-slip type stress) were not changed through the Middle–Late Miocene boundary.

Renewed extension and strike-slip faulting (Late Miocene) (Figs 4e, 5f)

After the main Mid-Miocene extension, the Pannonian Basin system was influenced by thermal subsidence (Horváth & Royden 1981; Royden *et al.* 1982). However, E–W to SE–NW tension was determined in Sarmatian and Pannonian sedimentary rocks around several sub-basins in the western and central Pannonian Basin and Western Carpathians (Fig. 5f). This stress field is similar to that of the Mid-Miocene and demonstrates the renewal of tectonic activity after the main phase of rifting. Master faults of the western and central Pannonian Basins were reactivated mainly as normal faults with oblique component. Some seismic sections, Tari *et al.* (1992) suggest that faults frequently do not cut through the lower Pannonian sequence as a single plane, but the deformation was distributed within and sealed by the soft young sediments. In some other places growth in Late Miocene deposits along listric normal faults is observed. Faulting also occurred along newly formed depressions, where no previous sedimentation took place (depressions of eastern Vértes, western Gerecse, the Aggtelek hills; Grill *et al.* 1984).

Some NE- to ENE-trending sinistral faults were also renewed. Such reactivation is indicated by important sediment thickness and en echelon geometry of depressions in the Vienna Basin (Wessely 1988; Hamilton *et al.* 1990; Fodor 1995). Strike-slip faulting of the southern boundary zone of the North Pannonian unit was also renewed and was associated with transpressional accommodation structures (Fig. 4e). The timing of motion is relatively precise here: Pannonian (Tóalmás: Csontos & Nagymarosy 1998), Pannonian to Pontian (Vatta-Maklár: Tari 1988; Csontos *et al.* 1991), late Miocene (Szolnok: Lőrincz & Szabó 1993). Further southwards, a sinistral transtensional shear zone was demonstrated from Paks to Szolnok (Pogácsás *et al.* 1989). Locally deep pull-apart basins were described in the Derecske region (Rumpler & Horváth 1988), in the southern part of the Danube–Tisza interfluve. All these basins were created by ENE–WSW-oriented left-lateral shear zones. In North Transylvania an important E–W-oriented left-lateral fault can be observed (Dicea *et al.* 1980) with Late Miocene activity (Györfi *et al.* this volume). Sarmatian to Pannonian volcanic rocks are deformed and partly intrude the fault. Due to later deformation, Late Miocene structures are poorly identified in southern Transdanubia. The important thickness variation of Pannonian sediments suggests continuing tectonic subsidence in the Drava and Sava grabens.

Late-stage basin inversion (Pliocene through Quaternary) (Figs 4f, 5g)

The direction of the maximal stress axes are varying from NW–SE to N–S (Fig. 5g). The stress field is strike-slip type or compressional at the Alpine–Dinaric junction and in the northern Pannonian Basin and mainly tensional in the eastern-central part. This stress field is similar to the recent one (Gerner 1992; Gerner *et al.* this volume).

The transpressional character of the stress field and structures of this phase are well expressed in the southern Pannonian Basin, from NE Slovenia to the Mecsek hills. The WNW (NW) trending dextral and ENE (NE) trending sinistral faults form obtuse angle. Strike slip was associated with open to tight folding, reverse faulting and with steepening of beds, all structures striking parallel to, or situated en echelon with respect to strike-slip faults. This geometry suggests that the strain was probably partitioned between strike-slip and reverse faults or folds, which is typical for transpressional deformation.

Dextral transpression was documented along the Periadriatic zone and connected to the Sava folds region (Polinski & Eisbacher 1992; Poljak, 1984; Premru 1976; Fodor *et al.* 1998). Similar transpressional tectonics was recently documented in the Papuk Mts by Jamičić (1995) and in other parts of northern Croatia (Prelogović *et al.* 1995). Here the reverse slip probably reactivated the earlier normal boundary faults of the Drava and Sava grabens (Tari & Pamić 1998). Along the parallel WNW trending Babócsa high (Kőrössy 1989) partitioning of NNE-vergent reverse and strike-parallel dextral slip is postulated. In the region, where the Periadriatic line changes their strike to E–W, folding and reverse reactivation of earlier normal faults was described (Rumpler & Horváth 1988; Pogácsás *et al.* 1994).

Between Zagreb and the Lake Balaton Middle to Upper Miocene sediments are folded with ENE–WSW-oriented fold axes. Folding affected Pannonian to early Pontian sediments below the Holocene mud of the Lake Balaton (Sacchi *et al.* this volume).

Strike-slip and reverse faults are well documented on both the northern and southern boundary of the Mecsek hills (Vadász 1935; Wein 1967; Csontos & Bergerat 1992; Tari *et al.* 1992). Accordingly, the whole Mecsek Hills seem to represent one transpressional, inverted pop-up structure (Csontos & Bergerat 1992; Tari 1992). Seismic reflection sections demonstrate that these fault zones are continuing north-eastwards up to the eastern part of the Pannonian Basin (Pogácsás *et al.* 1989). Structures become more transtensional in the northern Pannonian region. Here E–W to ENE–WSW tension was determined in Pliocene basalts and Pannonian sediments (Bergerat & Csontos 1987; Dudko *et* *al.* 1992; Bada *et al.* 1993; Vass *et al.* 1993). The corresponding σ_1 was sometimes horizontal but mainly vertical. There are few strike-slip faults in the central, northern Pannonian Basin and Western Carpathians which are clearly documented during this stage. One example is the southern boundary zone of the Vienna Basin where transtensional and transpressional segments are connected (Marko *et al.* 1991; Hubatka & Pospišil 1990). In its continuation south-vergent reverse faults were postulated within the Western Carpathians, south of the Tatra Mts (Tomek 1988; Tomek *et al.* 1989).

Some observations on synsedimentary tectonics suggest that the transpression and inversion may have (at least locally) started in the Pannonian (Wein 1967). However, the deformed nature of Pannonian and Pontian sediments support the amplification of deformation toward the end of the Late Miocene and the beginning of the Pliocene.

Along the Periadriatic zone and within the Drava graben borehole and seismic data show that Pliocene strata seal the bulk of the deformation (Vrabec 1994; Kőrössy 1989, 1990). However, the thick, slightly deformed basin fill of the small Velenje Basin in Slovenia demonstrates the continuation during and after the Pliocene (Brezigar *et al.* 1987). The eventual connection between the localization of basalt volcanoes, volcanic lines and faulting would suggest Pliocene to early Quaternary age of tectonism. Where Pliocene sediments are lacking, the upper time constraint of the deformation is not good. However, the general lack of Pliocene sediments, the strongly eroded character of Miocene rocks, the close connection of subsurface structure with recent morphology, and earthquakes would suggest that inversion deformation has continued through the Pliocene and Quaternary. The most recent period of the structural history is discussed by Gerner *et al.* (this volume).

Discussion and geodynamic model

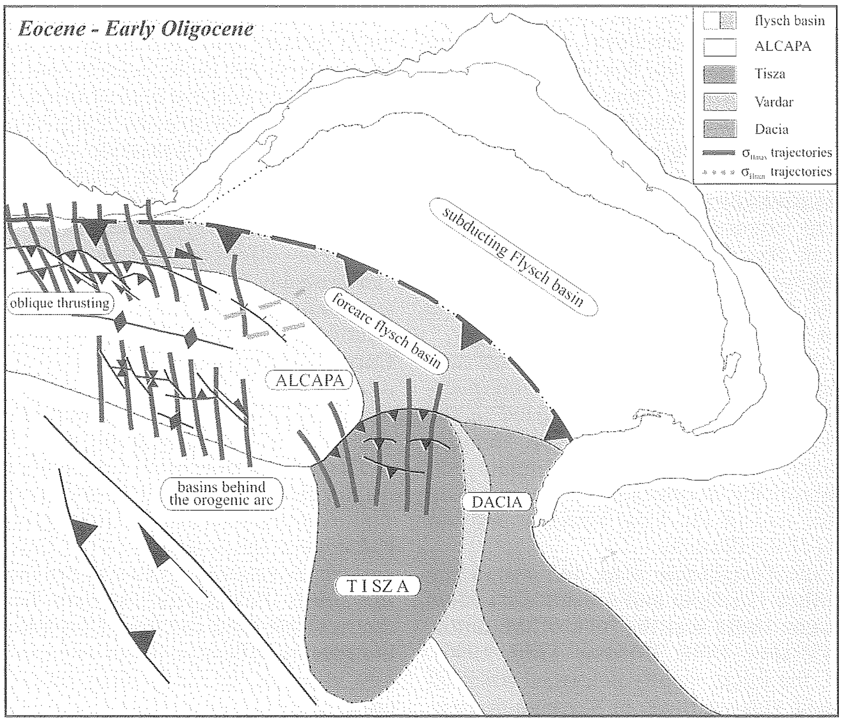
The middle Eocene to early Oligocene stress field reflects the Mesoalpine convergence of the European and Adriatic plates and intervening crustal blocks. This convergence resulted in the dextrally oblique nappe stacking of the Northern Calcareous Alps (Linzer *et al.* 1995). From the Late Eocene, this nappe stack and the Inner Western Carpathian nappes were thrust onto the Alpine and Carpathian Flysch Belt. In the hinterland of the thrust and fold belt, the Hungarian Palaeogene Basin system was formed possibly as a retroarc flexural basin (Tari *et al.* 1993). Due to the oblique convergence, this basin system was also dissected into isolated depressions by E–W dextral tear faults and associated en echelon normal faults (Fig. 6a).

Taking into account the Early Miocene rotations, the original orientation of σ_1 was N–S in the Pannonian–Western Carpathian part and was different from the Alpine NW–SE direction (Fig. 6a). This reconstruction indicates a slightly divergent pattern of the stress trajectories which can probably be explained by different boundary conditions along the orogenic area: with a relatively free interface east from the Western Carpathian unit. Reconstruction of the rotation of the Tisza–Dacia block also yields an original N–S compression, parallel to that of the eastern Alcapa σ_1 directions (Fig. 6a).

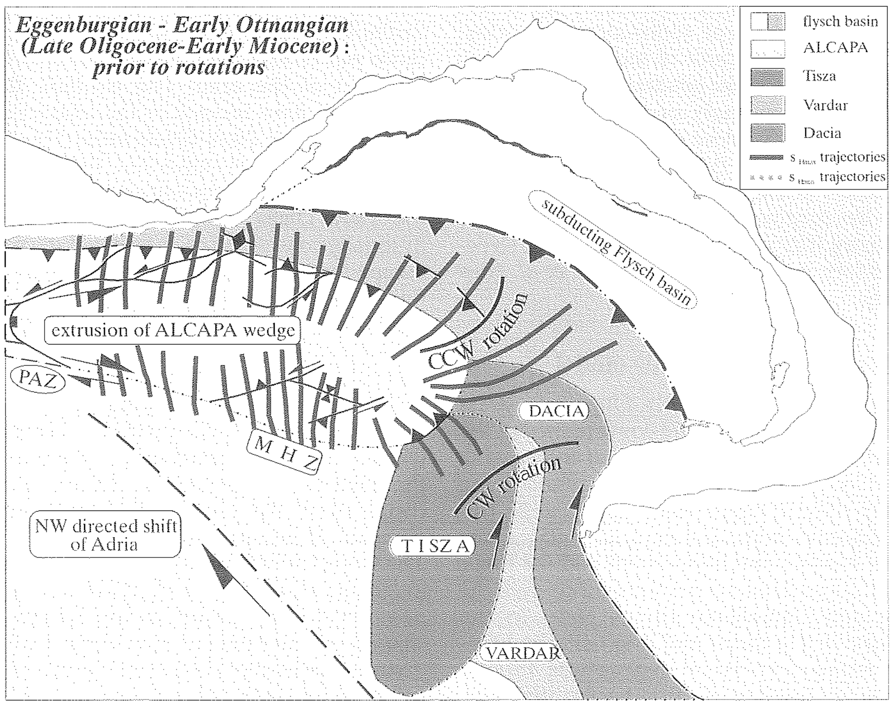
The Tisza–Dacia block had flysch troughs on its northern and northeastern external parts. The compressive stress field generated the Palaeogene flysch trough of the East Carpathians, which were overridden by the Dacides from the south. Palaeogene deposits between the Apuseni Mts and the East Carpathians can be interpreted as flexural basins on top of and in the hinterland of the prograding East Carpathian arc.

On major part of the Tisza–Dacia unit no Palaeogene deposits can be found. There are two exceptions. A small, but relatively deep continental basin near the Mecsek (Wéber 1985) and the more voluminous marine deposits of the Szolnok flysch trough. The former indicates that there was erosion and continental environment on most of the area. The Szolnok flysch is thought to have formed at the contact zone of the Alcapa and Tisza–Dacia units (Fig. 1.).

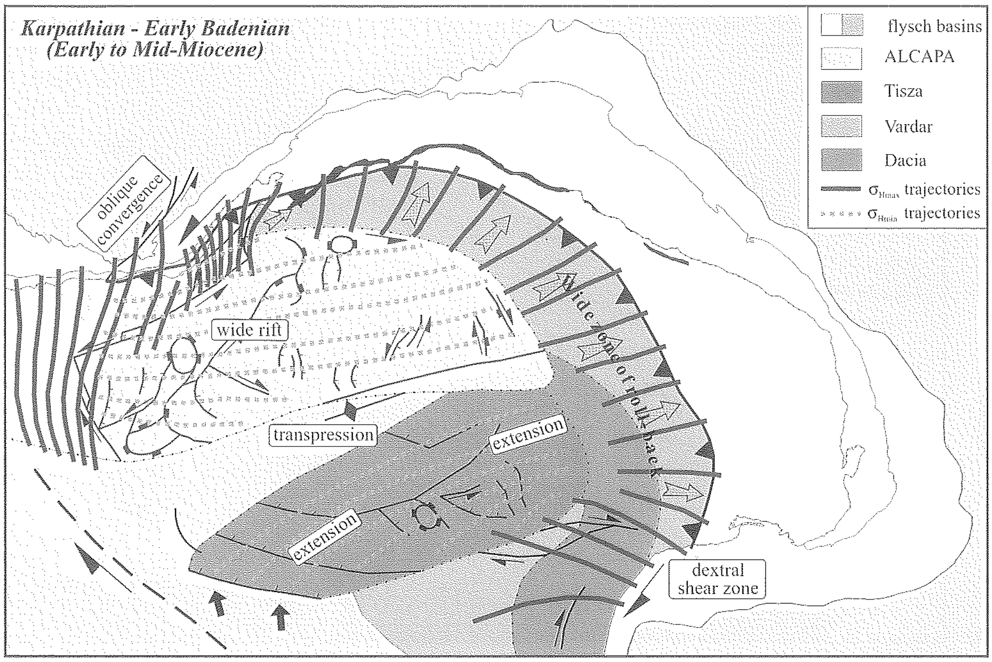
One of the most important Tertiary tectonic process was the juxtaposition of the Alcapa and the Tisza–Dacia terranes. This event was traditionally considered as an effect of the eastward escape (extrusion) of the wedge-shaped Alcapa unit (Fig. 6b), (Balla 1984; Kázmér & Kovács 1985; Csontos *et al.* 1992). Alcapa was bordered by a sinistral strike slip zone in the north, a dextral one in the south and a low-angle normal fault at the western side, along the eastern edge of the exhuming Tauern window (Neubauer & Genser 1990; Ratschbacher *et al.* 1991*a*). Due to the eastward extrusion, the Alcapa block was thrust upon the Tisza–Dacia unit, observed along the Botiza area (Györfi *et al.* this volume). The stress trajectories (corrected for the palaeomagnetic rotations) show relatively homogeneous N–S compression in the Alcapa. A fan shaped σ_1 trajectories may occur only at its eastern boundary along the contact with the southern unit (Fig. 6b).


Eocene - Early Oligocene
flysch basin
ALCAPA
Tisza
Vardar
Dacia
trajectories
trajectories
subducting Flysch basin
oblique thrusting
forearc flysch basin
ALCAPA
basins behind the orogenic arc
DACIA
TISZA

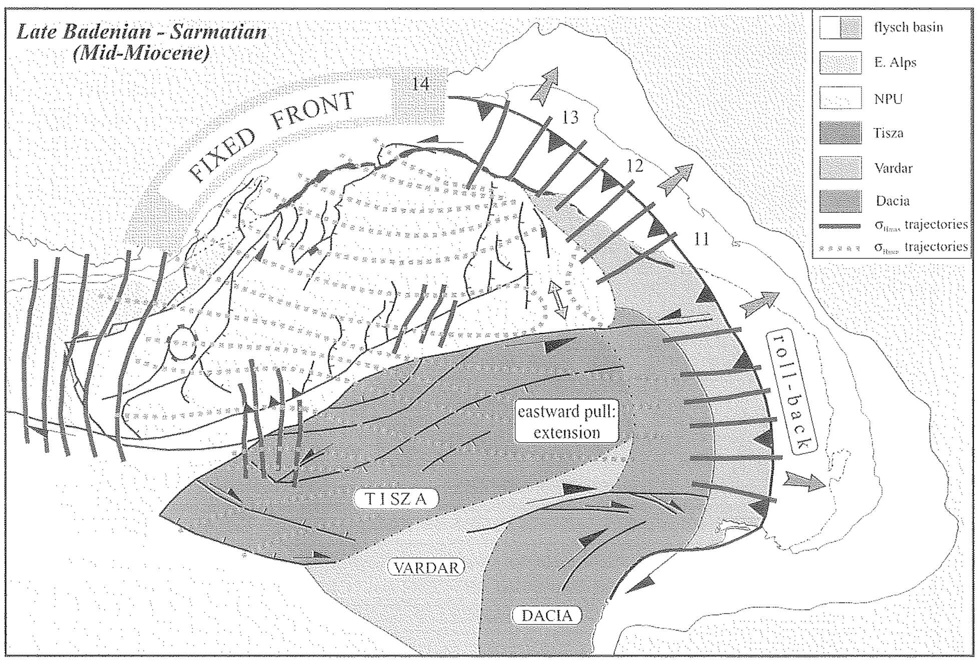
(a)


Eggenburgian - Early Ottnangian (Late Oligocene-Early Miocene): prior to rotations
flysch basin
ALCAPA
Tisza
Vardar
Dacia
trajectories
trajectories
subducting Flysch basin
extrusion of ALCAPA wedge
CCW rotation
PAZ
DACIA
M H Z
CW rotation
NW directed shift of Adria
TISZA
VARDAR

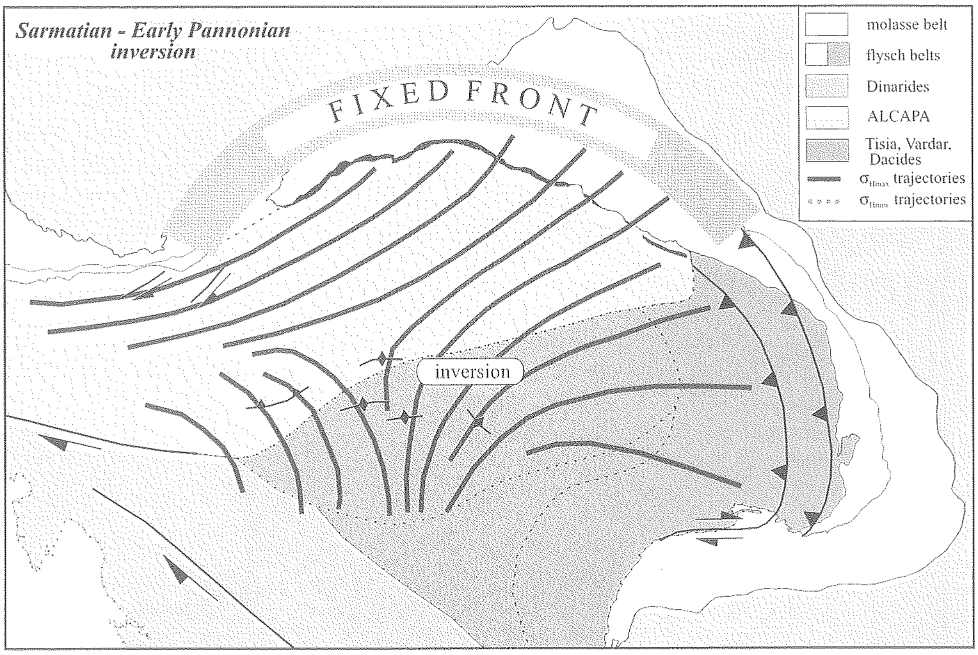
(b)


Karpathian - Early Badenian
(Early to Mid-Miocene)
flysch basins
ALCAPA
Tisza
Vardar
Dacia
σHmax trajectories
σHmin trajectories
oblique convergence
wide rift
transpression
extension
extension
wide zone of roll-back
dextral shear zone

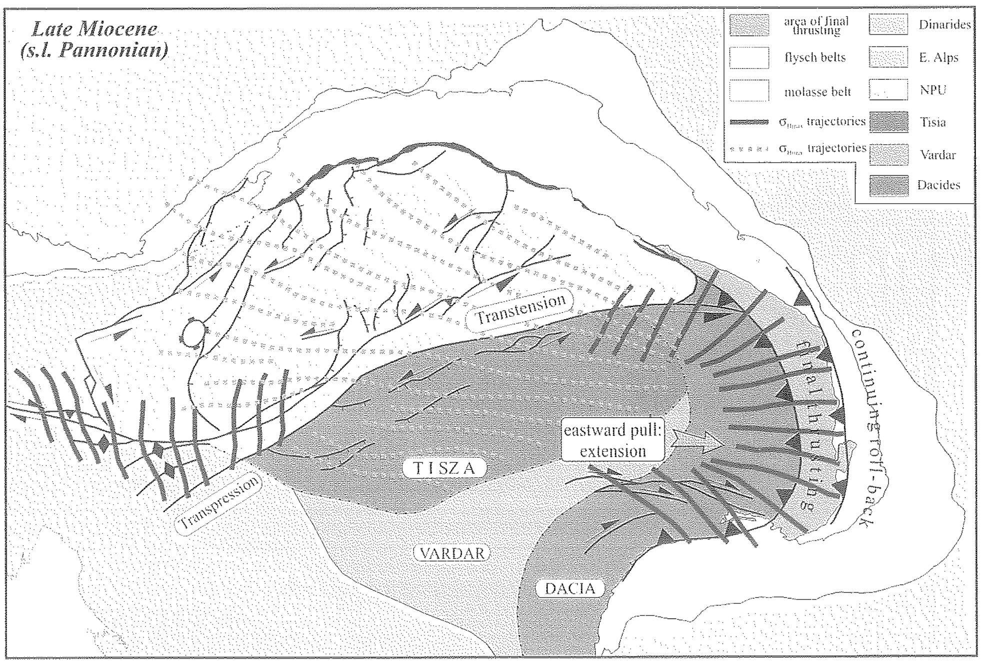
(c)


Late Badenian - Sarmatian
(Mid-Miocene)
flysch basin
E. Alps
NPU
Tisza
Vardar
Dacia
σHmax trajectories
σHmin trajectories
FIXED FRONT
14
13
12
11
roll-back
eastward pull:
extension
TISZA
VARDAR
DACIA

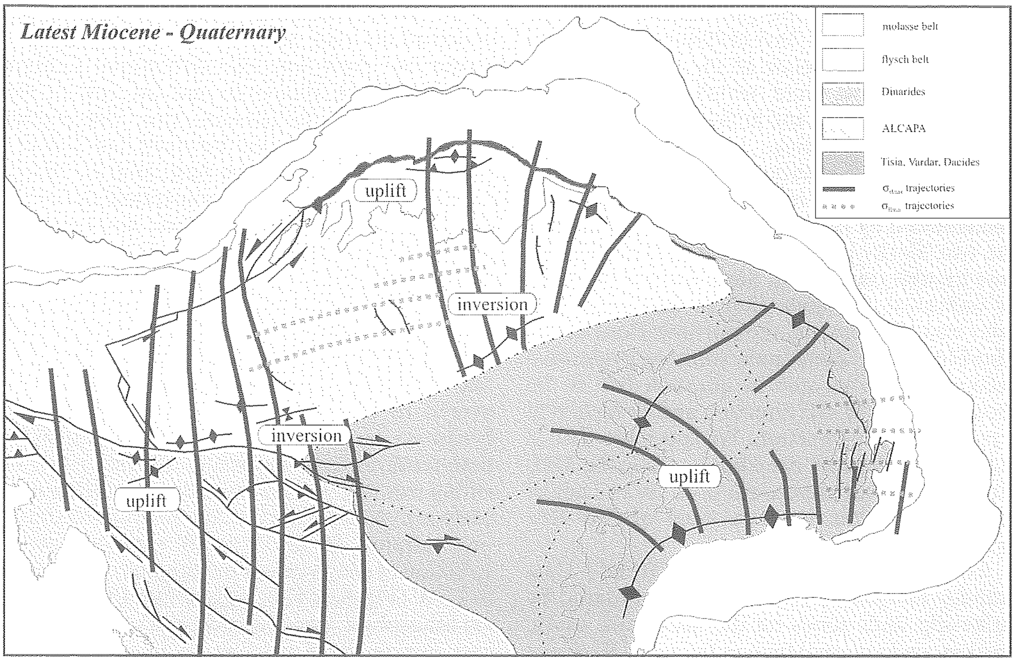
(d)


Sarmatian - Early Pannonian inversion
FIXED FRONT
inversion
molasse belt
flysch belts
Dinarides
ALCAPA
Tisia, Vardar, Dacides
σHmax trajectories
σHmin trajectories

(e)


Late Miocene (s.l. Pannonian)
area of final thrusting
flysch belts
molasse belt
σHmax trajectories
σHmin trajectories
Dinarides
E. Alps
NPU
Tisia
Vardar
Dacides
Transtension
eastward pull: extension
final thrusting
continuing roll-back
Transpression
TISZA
VARDAR
DACIA

(f)



(g)

Fig. 6. Evolution of major blocks of the Pannonian area: tentative model for the connection of stress field and deformation. Compare with Figs 4 and 5. Kinematic reconstruction of major blocks is partly after Balla (1984); Csontos (1995); Kováč *et al.* (1994), Ratschbacher *et al.* (1991*a*). Numbers along the thrust front indicate the age of the last thrust in Ma, after Jiříček (1979).

The final exhumation of the Tauern window can be connected with this process (Ratschbacher *et al.* 1989; Neubauer & Genser 1990), which could have started in the latest Oligocene (late Egerian). However, the bulk of displacement occurred only at the end of Eggenburgian or in the early Ottnangian, indicated by the disruption of the formerly unique Hungarian and Slovenian Palaeogene Basins (Csontos *et al.* 1992; Tari *et al.* 1995).

Sinistral strike-slip displacement along the northern boundary zone of the Eastern Alps is less than 100 km which is in striking contrast with the more important apparent dextral separation along the Periadriatic line, between the Southern Alps and the North Pannonian area. This is estimated between 300 to 400 km by Tari (1994). The solution is probably given by Tari *et al.* (1995) who emphasized that the dextral shift is the sum of the northwestward motion of the Adriatic promontory, the eastward extrusion of the Alcapa and the extension of the Pannonian area.

During and after the escape/extrusion event (around 18–17 Ma) the Alcapa and Tisza–Dacia units experienced important rotations, counterclockwise and clockwise sense, respectively (Márton & Mauritsch 1990; Márton & Márton 1996). Following Balla (1984) and Csontos (1995) our tectonic reconstruction incorporate the assumption of rigid-body rotations despite that small-scale domino-type block rotations are not excluded at some regions (Marko *et al.* 1991). Due to these rotations, the formerly linear PAL–MHZ was bent and the subsequent Mid-Miocene dextral shift along the southern boundary of the extruding Eastern Alps was transferred from the Periadriatic zone to other NW–SE oriented shear zones within the Dinarides.

On the other hand, the time of the rotation of the Alcapa unit is very close to the exhumation of the Rechnitz window (17 ± 1.9 Ma; Dunkl & Demény 1997). The differential rotation between the Alps and Pannonian domain resulted in the disruption of the formerly unique Alcapa block and the separation of the East Alpine and Pannonian–Carpathian domains.

The boundary can be put to the Rechnitz detachment fault of Tari (1994) at mid-crustal level. One consequence of the rotation is the decoupled stress field which was markedly different in the Alps and in the Pannonian Basin from the onset of rifting.

The opposite rotation of the Alcapa and Tisza–Dacia rises a serious problem, also encountered by Balla (1984). The area between the two blocks has to be reduced considerably (see triangle on Fig. 6b). Since there is no proof or indication of subduction in this sector, we infer that the area, i.e. the portions close to the Mid-Hungarian Zone underwent very intensive deformation. The area in question had to be shortened in NW–SE direction and elongated in a perpendicular direction. We speculate that at least part of the very deep basins along the Mid-Hungarian shear zone were born as a result of major elongation, with the help of NW–SE-oriented normal faults.

After the rotation, the unified Carpathian–Pannonian block was still bordered by strike-slip fault zones on the north and south, and low angle normal fault at the west (Rechnitz window), so one can speculate on continuing extrusion tectonics (e.g., Decker & Peresson 1996).

The rifting of the Pannonian Basin can be explained by the interplay of several driving mechanisms. One component is the convergence of the Adriatic microplate and Europe. The effect of this is clearly seen in the relatively stable stress field orientation of the Eastern and Southern Alps and along the Periadriatic fault system. The resulting tectonic process is the continuation of the uplift of the Eastern Alps and the formation of conjugate sets of strike-slip faults which led to the model of Miocene extrusion of the Eastern Alps (Ratschbacher *et al.* 1989; Neubauer & Genser 1990). Conjugate strike-slip faults permitted only minor N–S contraction while more important shortening was maintained along the transpressional Periadriatic zone and in southvergent thrusting of the Southern Alps (Boccaletti *et al.* 1990).

The other, probably dominant driving mechanism of rifting is the subduction roll-back along the outer Carpathians (Royden *et al.* 1982; Csontos *et al.* 1992; Royden 1993). The onset of calc-alkaline volcanism around 19 Ma was associated with the subduction. The migration of volcanic centres toward the thrust front was long-time connected to the retreat of the subducting plate toward the foreland (Balla 1981, 1984; Szabó *et al.* 1992; Lexa *et al.* 1995).

The third factor which might have been contributed to rifting is given by the CCW rotation of the North Pannonian Unit. It is because, this occurred not only relative to a fix geomagnetic frame, but there was also differential rotation between North Pannonia and the rest of the Alcapa unit, i.e. the Eastern Alps. This implies that the transition zone between the Eastern Alps and North Pannonia bent and the 'hinge zone' coincide with the demonstrated metamorphic core complexes. Therefore, we speculate that they are genetically related (Csontos 1995; Fodor *et al.* 1998).

Two main directions of tension can be differentiated within the majority of the Pannonian Basin system, a NE–SW and an E–W to SE–NW one. The main tectonic subsidence can be connected to only one tensional phase in some basins while in others the combined effects of both tensions were important. The first phase of rifting was in general oriented east-northeast or eastward. At least part of the determined direction suffered 25° to 35° CCW rotation, reconstruction of which brings the original orientation to E–W (Fig. 6c). This suggest that the Pannonian lithosphere was extended east or east-northeast during the Karpatian to Early Badenian first rifting phase. This orientation points in general towards the subduction front, which, at this time, incorporated the whole Carpathian front (Fig. 6c).

The thinning of the lithosphere occurred by low angle normal and accompanying high angle faults (Tari *et al.* 1992). High-angle normal faults are widespread far from the metamorphic core complexes. They probably developed in 'wide rift mode' (Tari 1994). The oppositely tilted half-grabens, the sudden change of extended and non-deformed blocks need the existence of transfer faults. The largest one could be the Mid-Hungarian shear zone but several smaller ones are also documented (Tari *et al.* 1992; Csontos 1995). While they are subparallel to the tensional direction, their kinematics is not always evident. Smaller sinistral and dextral separations are also documented along minor N–S to NNE–SSW and E–W faults, respectively, but strike-slip faulting seems to be minor with respect to normal faulting (Figs 4c, 6c).

At both the northwestern and southeastern ends of the front, at the junction with the Alpine and Balkan thrust-and-fold belts, important strike-slip zones developed to accommodate the steps of the Carpathian thrust front (Fig. 6c). At the southeastern corner earlier, sometimes curved dextral faults were reactivated (Balla 1984) and Oligo-Miocene Basins were inverted (Ratschbacher *et al.* 1993*b*). On the northwestern corner the sinistral opening of the Vienna Basin accommodated the northeastward step of the Carpathians with respect to the Alpine thrust belt. Strain partitioning occurred along this oblique front: at the Vienna Basin margin sinistral displacement occurred while N from the sinistral shear zone, in the external Flysch and Molasse zones thrusting was almost perpendicular to the arcuate thrust front (Fodor *et al.* 1995). At both corners the tensional stress field is replaced by strike-slip type stress, having fan shaped σ_1 trajectories (Fig. 6c).

After the phase of the first rifting, 20–30° CCW rotation occurred, affecting part of, or the whole North Pannonian–West Carpathian unit. The age of rotation is well-defined as earliest Mid-Miocene (early to middle Badenian) in central North Hungary (Márton & Fodor 1995). However, NE–SW tension was observed in rocks younger than the age of rotation, mainly in Badenian andesitic rocks of central North Hungary where rotation is proved to be absent or minor (Márton & Márton 1996). This suggests that the duration of the first rifting event overlaps the time span of the rotation, at least in the central and northeastern Pannonian Basin. The suggested scenario can be the following; initial E–W tension, rotated to NE–SW during the Early Badenian, renewed NE–SW tension (already non-rotated) during the Mid-Badenian.

After this rotation the stress field changed in the western Pannonian Basins (Vienna and Danube Basins) and the minimal stress axis (σ_3) rotated to ESE–WNW. The tension was still oriented toward E or NE in the central and eastern Pannonian Basins (Fig. 6d). This change in tensional direction was gradually younging eastward; Late Badenian in the central northern Pannonian Basin and late Sarmatian in the Tokaj–Slanec volcanic chain and in the East Slovakian Basin (compare Fig 6c, d & f).

To explain this inhomogeneous tensional direction, we suggest that the northeastward suction effect of the subducted slab beneath the northeastern and eastern Carpathians was modified by the lateral boundary conditions of the overriding slab. Spakman (1990) suggested a laterally propagating slab detachment beneath the Carpathians. The detached portion of the slab could exert no more drag effect on the overriding plates. In consequence, the drag was exerted on shorter and shorter sections and the orientation has also changed. The northeastward drag ceased after the Mid-Miocene and only the drag effect toward the Eastern Carpathians remained.

The other consequence of the gradual cessation of thrusting is that the overriding Western Carpathian–Northern Pannonian unit could not slip more (north)eastward and its northwestern boundary became gradually fixed. In consequence, the drag effect appeared along a curved path between the retreating (moving) and the fixed front of the Western Carpathians and northern Pannonian Basins (Fig. 6c, d, f). The centre of this curved path changed, explaining the rapidly changing tensional directions. The directional variation is more expressed in the central Western Carpathian depressions because they were situated closer to the freezed lateral boundary: here three episodes of tension could be differentiated (Nemčok *et al.* 1993; Hók *et al.* 1995). The E–W tension occurred during the latest Badenian and Sarmatian, when thrusting (and probably slab retreat) was active east of this region. From the latest Mid-Miocene the tension was oriented SE probably corresponding to the drag effect from the East Carpathians (Fig. 6f). In the northern central Pannonian Basins only two tensional direction were clearly separated. During the Late Miocene the tension was similar to that of the Sarmatian, because the ESE σ_3 direction approximates the probable direction of the drag (Fig. 6f).

Back-arc extension is well-expressed in the fast subsidence of the East Slovakian Basin during the Late Badenian and Sarmatian (Figs 4d & Fig. 6d). Because the thrust front was fixed just north of this basin, the extension was associated with moderate counterclockwise rotation of rocks and the thrust front itself. This rotation was observed in the Tokaj hills and in the flysch belt (Bazhenov & Burtman 1980; Balla 1984; Márton & Pécskay 1995; Orlicky 1996). The rotation was fast, occurring within the late Sarmatian (12–11 Ma). Because the area west of the Bükk hills did not rotated (Fig. 1), the area affected by extension and rotation had triangle shape, narrowing northward (Balla 1984). The rotation affected the earliest faults of the Tokaj hills. The original ENE–WSW tension (and perpendicular compression) rotated to NNE–SSW orientation. A younger ENE–WSW tension is already non-rotated and expresses directly the drag of the retreating subduction zone.

The other explanation for the inhomogeneous stress field may be the collision of the overriding plates with buoyant, non-subductable material of the European margin. This caused halting of the thrusts and forelandward propagation of the orogene, consequently the hampering of back-arc extension. The westernmost Carpathians were blocked first in the Badenian, therefore the back-arc region could extend only in an E–W to SE–NW direction.

In the eastern Pannonian Basin no major difference occurs between the first and second phase of rifting (Fig. 6d, f). The southern boundary zone of the Tisza–Dacia unit continued the dextral slip during the Mid–Late Miocene. The retreating subducted slab beneath the Eastern Carpathians was far enough to create local differences, arcuate strain trajectories developed only close to thrust fronts and the lateral boundaries of the eastward moving unit (Linzer 1996).

The tensional deformation was coeval with, or interrupted by, strike-slip or true compressional deformation occurring mainly within the external thrust belt. Such a deformation is natural in fold and thrust belts, but well-dated stress data are scarce (Figs 5c, d, e and 6c, d, e). One example can be the stress field along the Pieniny Klippen belt (Ratschbacher *et al.* 1993*a*), where the timing is poor. In the Pannonian region the NE–SW to E–W compressional stress data can also be regarded as a far field effect of compression in the northeastern Carpathians. Although generally not well-dated, we tentatively connect this event with the termination of subduction along this front (Fodor *et al.* 1990; Peresson & Decker 1997). Since this termination of thrusting was time-progressive, the same might be the case for the strike-slip or compressional event. However, the resolution of timing is not good enough to support this idea. The best documented inversion structures are located in the central-southern Pannonian Basin and the strike of the structures would indicate N–S compression. In that way the σ_1 trajectories would show an arcuate pattern, starting from NE or ENE direction in the external thrust belt and rotate gradually to N–S approaching the Dinarides (Fig. 6e). The age of this inversion is latest Sarmatian or early Pannonian (11–8? Ma).

From the latest Miocene the continuing northwestward convergence of Europe and Africa induced N–S to NNW–SSE compression in the main part of the Pannonian Basin (Fig. 6g). This effect was amplified during the latest Miocene–earliest Pliocene and continued up to recent times (Gerner *et al.* this volume). Important shortening occurred in an E–W trending zone, from the Venetian Alps to the Southern Carpathians. This shortening was propagating in time through the Miocene, from the tip of the Adriatic promontory, from the Lombardian Alps (Massari 1990). This suggests that this compressive belt was gradually propagating eastward. Dextral shift of Adria and its deformed foreland with respect to the Pannonian area was maintained by dextral transpression along the Periadriatic zone, Sava folds and other NW-striking dextral faults within the Dinarides (Fig. 6g). Large-scale thrusting seems to lack in the Pannonian Basin, indicating eastward decrease in the magnitude of inversion. Folding and thrusting are progressively replaced by sinistral and dextral transpressional reactivation of a number of earlier faults. The subduction finished all along the Eastern Carpathians and the Pannonian Basin has undergone inversion and tectonic reactivation since the Pliocene (Horváth 1995; Gerner *et al.* this volume).

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Appendix: references for stress and structural data, including maps

Eastern Alps (Austria)

Structural data – whole: Fuchs (1984); Neubauer & Genser (1990); Ratschbacher *et al.* (1989, 1991*a*, *b*).

Structural data – regional: Decker *et al.* (1993, 1994); Fodor, (1995); Flügel & Neubauer (1984) (1:200 000); Flügel *et al.* (1988) (1:200 000); Fuchs & Grill (1984) (1:200 000); Hamilton *et al.* (1990); Linzer *et al.* (1995); Nemes *et al.* (1995*b*); Polinski & Eisbacher (1992); Ratschbacher *et al.* (1990); Schmid *et al.* (1989).

Palaeostress – calculations: Decker *et al.* (1993, 1994); Fodor (1995); Fodor *et al.* (1990); Linzer *et al.* (1995); Nemes *et al.* (1995*a*, *b*).

Palaeostress – estimations, microtectonic data: Polinski & Eisbacher (1992); Sauer *et al.* (1992).

Western Carpathians (Czech Republic, Slovak Republic, Poland)

Structural data – whole: Fusán *et al.* (1967; 1:500 000); (1987; 1:500 000); Jiříček (1979); Mahel *et al.* (1984) (1:500 000); geol. maps of 1:200 000.

Structural data – regional: Birkenmajer (1985); Konečny & Lexa (1984) (1:50 000); Kováč *et al.* (1989, 1990, 1994, 1995); Lankreijer *et al.* (1995); Plašienka, (1991); Ratschbacher *et al.* (1993*a*); Salaj (1995).

Palaeostress – calculations: Fodor (1995); Fodor *et al.* (1990, 1995); Marko *et al.* (1991); Nemčok (1993); Nemčok *et al.* (1993); Ratschbacher *et al.* (1993*a*).

Palaeostress – estimations, microtectonic data: Hók *et al.* (1995); Kováč *et al.* (1989, 1990, 1995); Kováč & Hók (1993); Nemčok *et al.* (1989); Nemčok & Lexa (1990).

Eastern Carpathians (Romania)

Structural data – whole: Săndulescu *et al.* 1981

Structural data – regional: Berza *et al.* (1988*a*, *b*).

Palaeostress – calculations: Ratschbacher *et al.* (1993*b*); Linzer 1996; Györfi *et al.* this volume.

Southern Alps–Dinarids (Slovenia–Italy)

Structural data – whole: 1:500 000 geological map of Yugoslavia.

Structural data – regional: Boccaletti *et al.* (1990); Massari (1990); 1:200 000 sheets of Yugoslavia; Premru (1976).

Structural data – local: Brezigar *et al.* (1987); Vrabec (1994).

Palaeostress – calculations: Fodor *et al.* (1998); Nemes *et al.* (1995*a*).

Palaeostress – estimations, microtectonic data: Poljak (1984).

Southern Pannonian basin (Croatia)

Structural data – whole: Dragičević *et al.* (1983) Prelogović *et al.* (1995); Tari & Pamić (1998).

Structural data – regional: Jamičić (1983, 1995); Pamić (1995).

Pannonian basin (Hungary)

Structural data – whole: Balla (1984, 1988); Brezsnyánszky & Haas (1990) (1:500 000); Csontos *et al.* (1992); Csontos (1995); Kilényi *et al.* (1991); Szafián (unpublished); Tari (1994); Tari *et al.* (1992); Maps of 1:200 000.

Structural data – regional: Bada *et al.* (1996); Balla (1989); Balla & Dudko (1989); Balla *et al.* (1987); Balogh (1964) (1:100 000); Benkovics (1991); Csontos (1988); Dudko *et al.* (1989, 1992); Fodor *et al.* (1992, 1994); Grill *et al.* (1984); Gyarmati *et al.* (1976) (1:50 000); Györfi (1992, 1993); Györfi & Csontos (1994); Hámor (1985); Horváth *et al.* (1998); Kőrössy (1989, 1990); Lőrintz & Szabó (1993); Mészáros (1982); Mészáros & Tóth (1981); Radócz (1966) (1:50 000); Rumpler & Horváth (1988); Tari *et al.* (1993); Tari (1988, 1992); Telegdi-Roth (1951); Vass *et al.* (1979, 1993); Vadász (1935); Wein (1967).

Structural data – local: Bence *et al.* (1991); Bergerat *et al.* (1983); Csontos & Nagymorosy (1998); Kókay (1976, 1989, 1990); Maros (1988).

Palaeostress – calculations: Bada *et al.* (1993); Benkovics (1991; 1997); Bergerat (1989); Bergerat *et al.* (1984); Bergerat & Csontos (1987, 1988); Csontos (1988); Csontos & Bergerat (1992); Csontos *et al.* (1991); Fodor *et al.* (1992, 1994); Györfi (1993); Györfi & Csontos (1994); Maros (1988); Márton & Fodor (1995); Sztanó & Fodor (1997).

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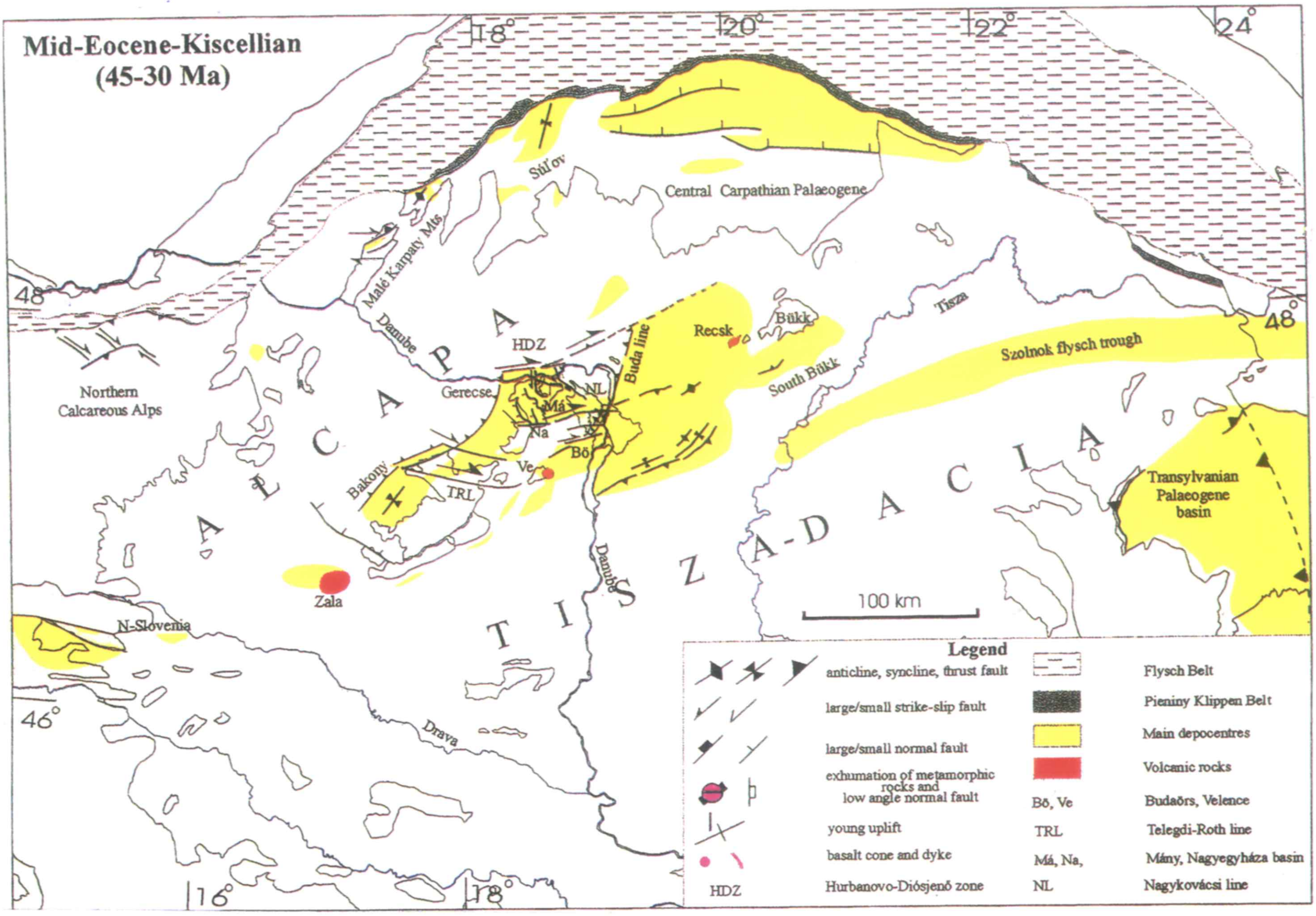
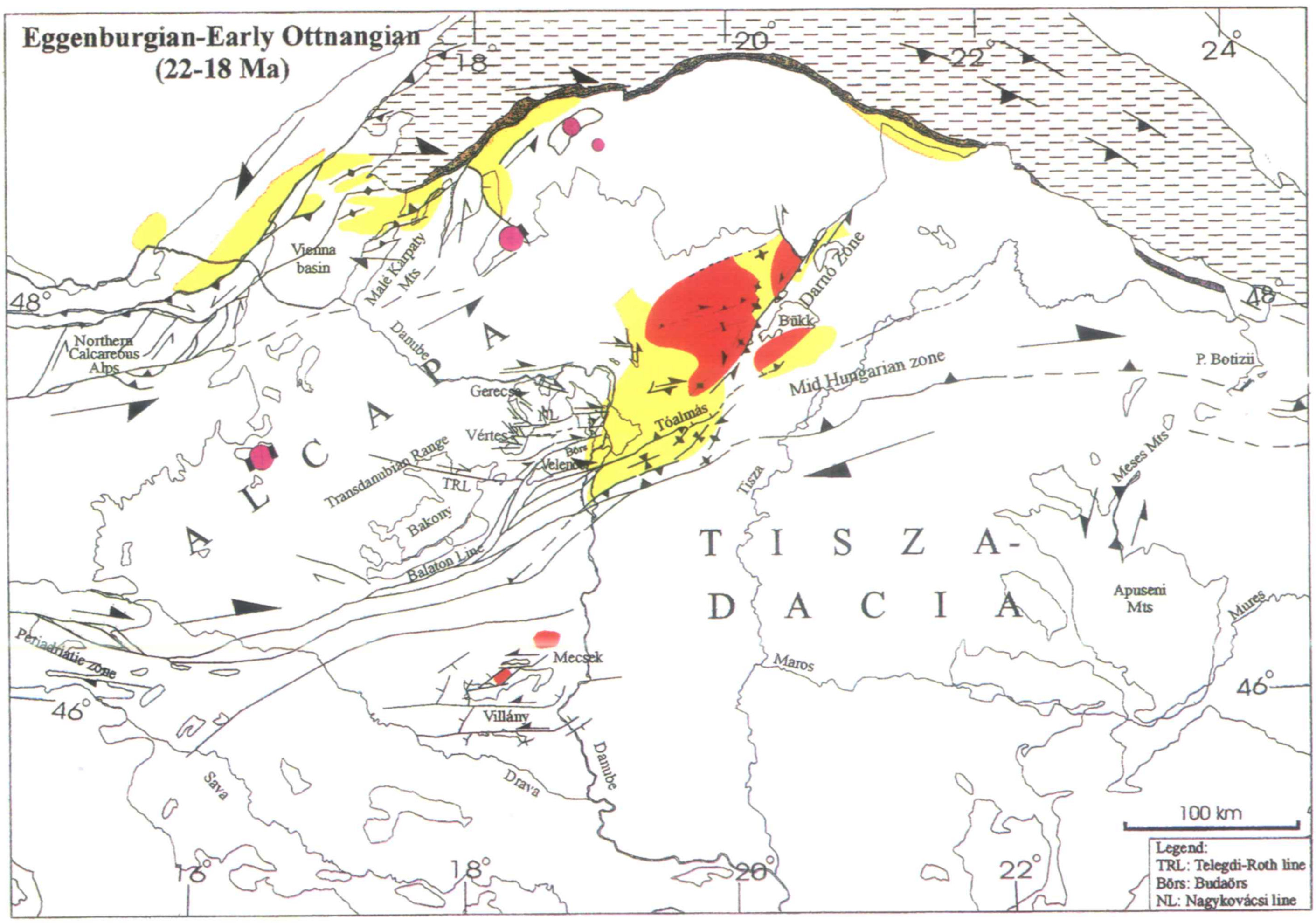
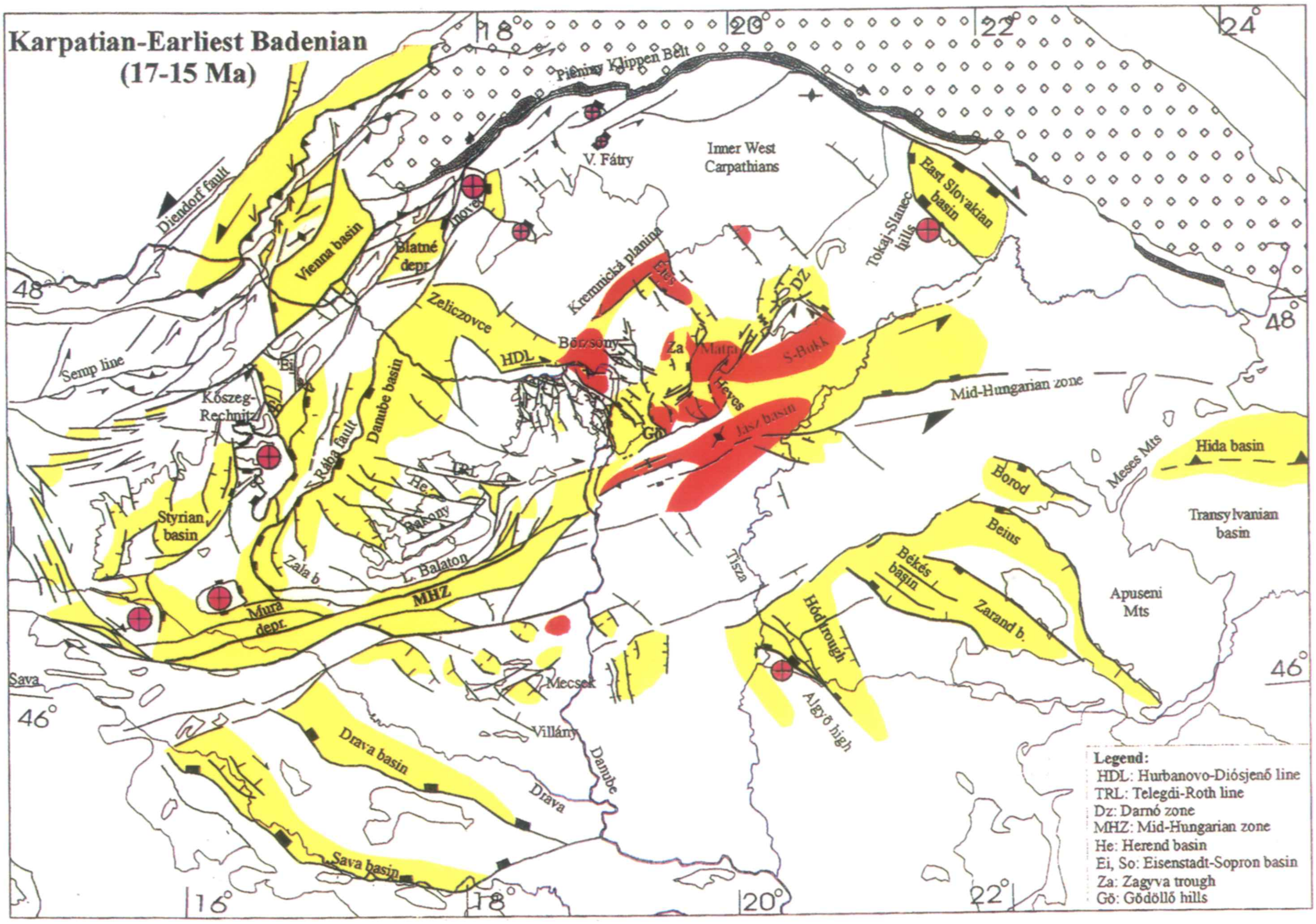
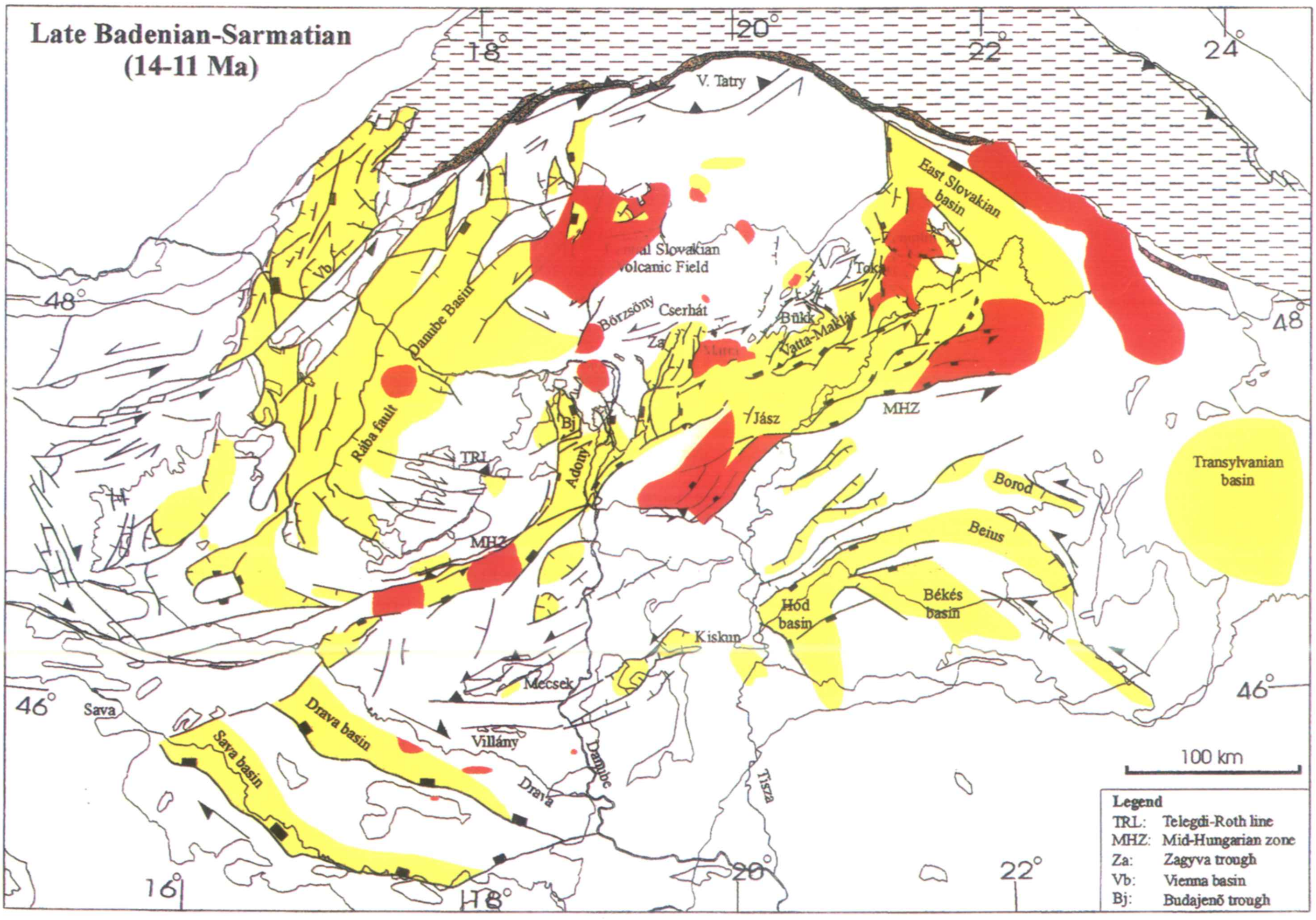
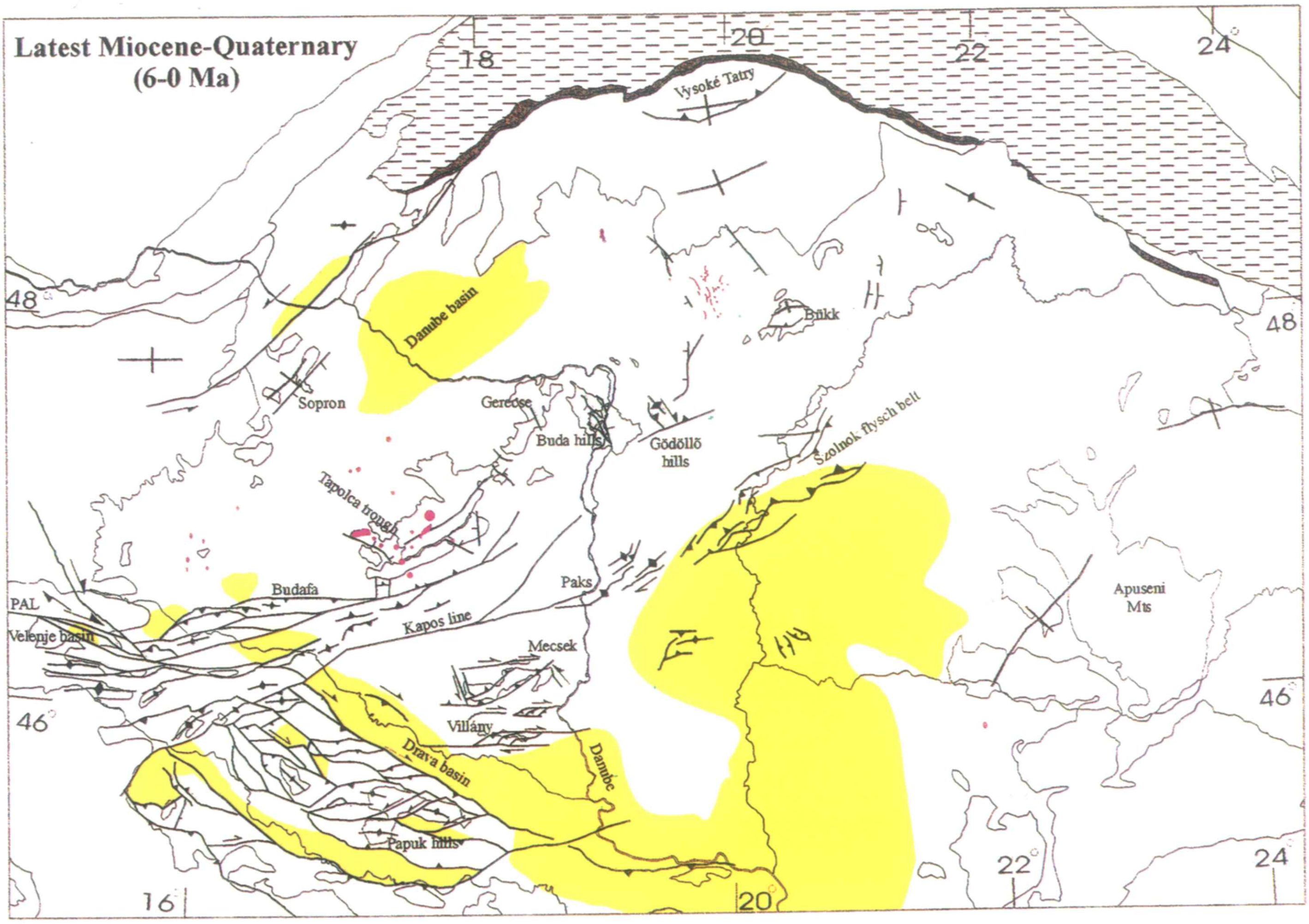

Mid-Eocene-Kiscellian
(45-30 Ma)
18°
20°
22°
24°
48°
46°
16°
Súl'ov
Central Carpathian Palaeogene
Malé Karpaty Mts
Danube
HDZ
Buda line
Recsk
Bükk
Tisza
Szolnok flysch trough
Northern Calcareous Alps
Gerecse
NL
Má
Na
Bö
Ve
South Bükk
Bakony
TRL
Transylvanian Palaeogene basin
A L C A P A
T I S Z A - D A C I A
Zala
N-Slovenia
Danube
Drava
100 km
Legend
anticline, syncline, thrust fault
large/small strike-slip fault
large/small normal fault
exhumation of metamorphic rocks and low angle normal fault
young uplift
basalt cone and dyke
HDZ
Hurbanovo-Diósjenő zone
Flysch Belt
Pieniny Klippen Belt
Main depocentres
Volcanic rocks
Bő, Ve
Budaörs, Velence
TRL
Telegdi-Roth line
Má, Na,
Mány, Nagyegyháza basin
NL
Nagykovácsi line


Eggenburgian-Early Ottnangian
(22-18 Ma)
18°
20°
22°
24°
48°
46°
16°
Vienna basin
Malé Karpaty Mts
Northern Calcareous Alps
Danube
A L C A P A
Gerecse
NL
Vértes
Transdanubian Range
TRL
Bakony
Balaton Line
Velence
Börs
Tóalmás
Bükk
Darnó Zone
Mid Hungarian zone
P. Botizii
Tisza
Meses Mts
T I S Z A - D A C I A
Apuseni Mts
Mures
Maros
Periadriatic zone
Mecsek
Villány
Sava
Drava
Danube
100 km
Legend:
TRL: Telegdi-Roth line
Börs: Budaörs
NL: Nagykovácsi line

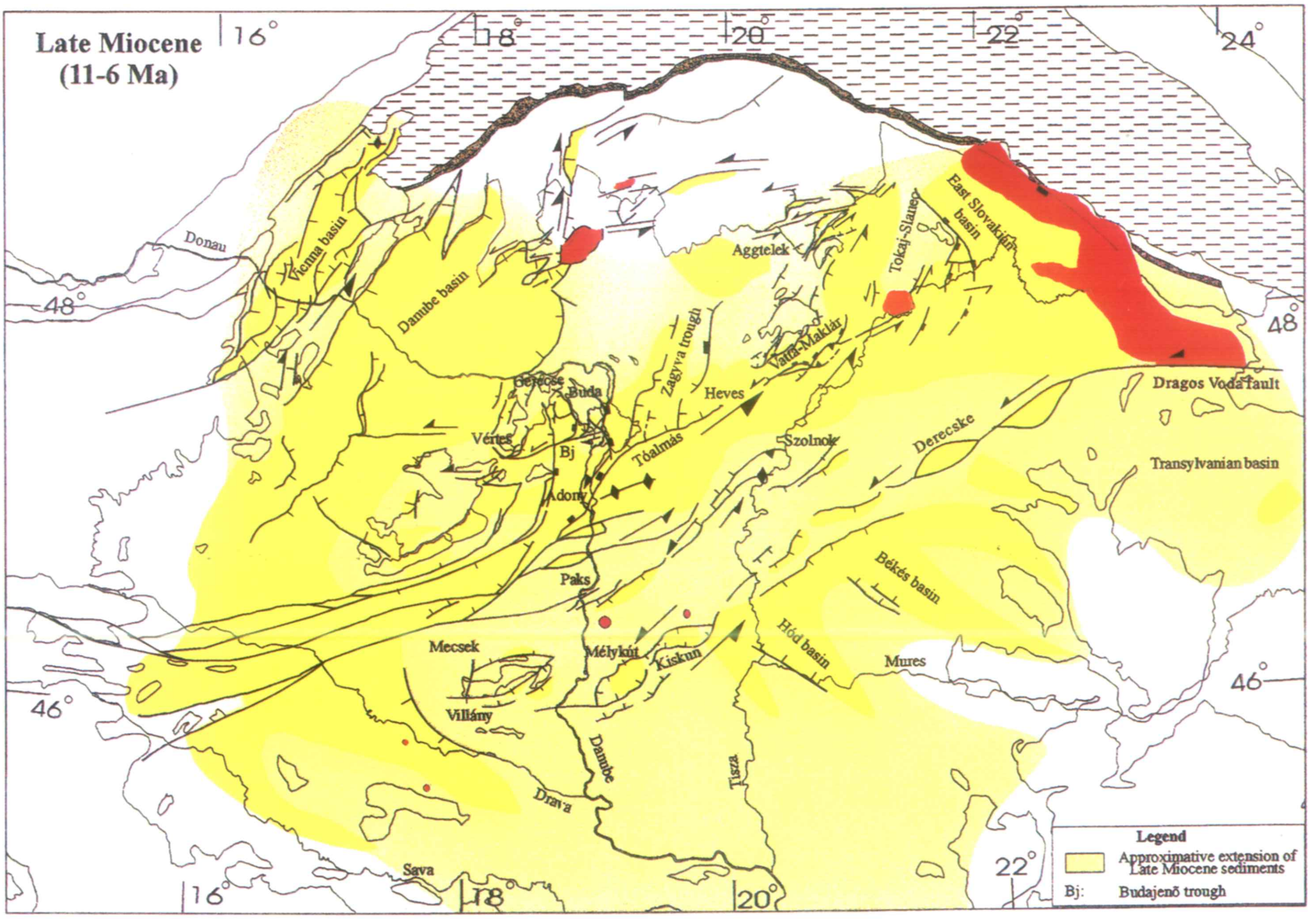




Fodor et al., Tertiary stress field of the Carpatho-Pannonian area

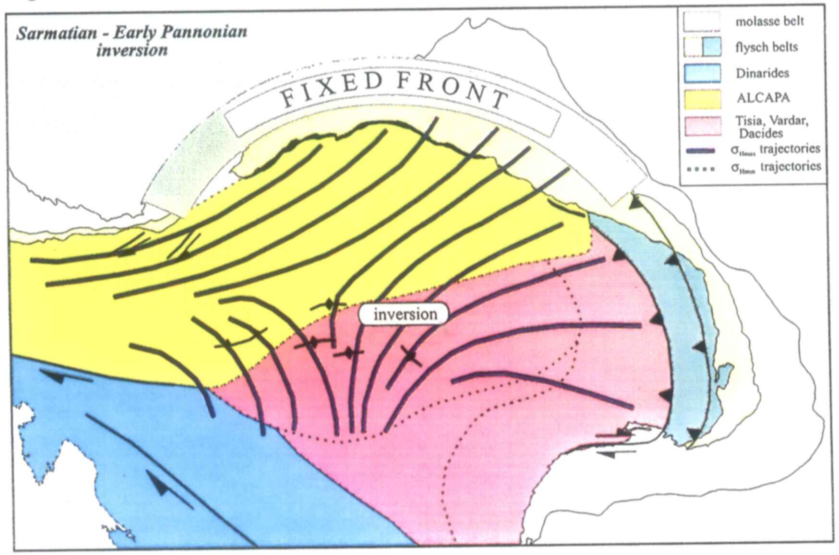


Fodor et al., Tertiary stress field of the Carpatho-Pannonian area



Fodor et al., Tertiary stress field of the Carpatho-Pannonian area

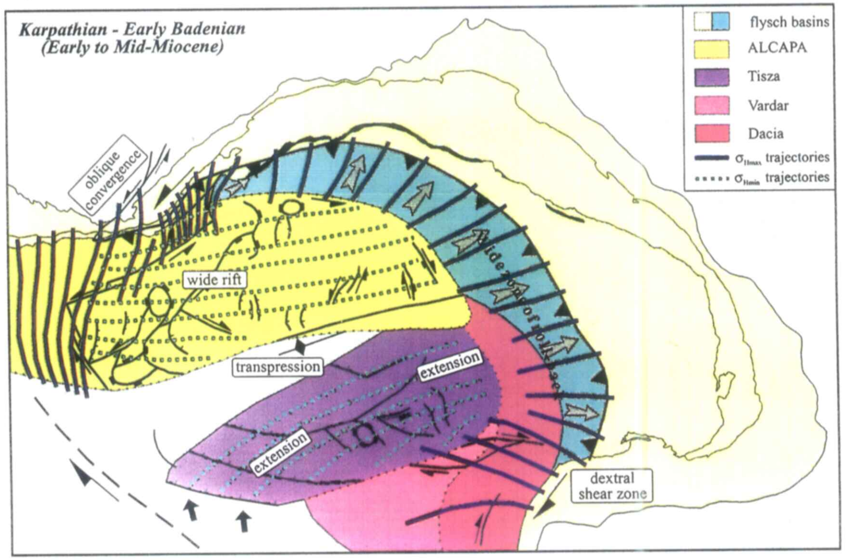
Figure 6e



Fodor et al., Tertiary stress field of the Carpatho-Pannonian area

AV 14_6E

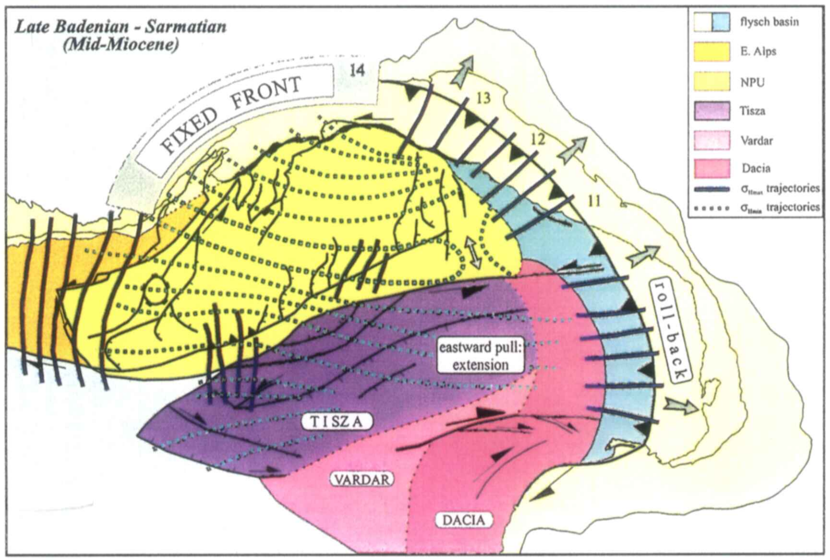
Figure 6c



Fodor et al., Tertiary stress field of the Carpatho-Pannonian area

AV 14_6C

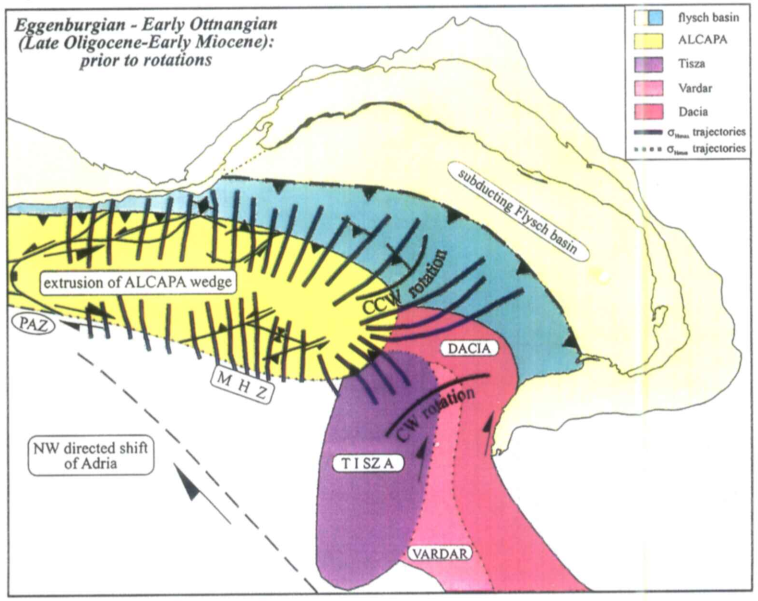
Figure 6d



Fodor et al., Tertiary stress field of the Carpatho-Pannonian area

AV 14_6D

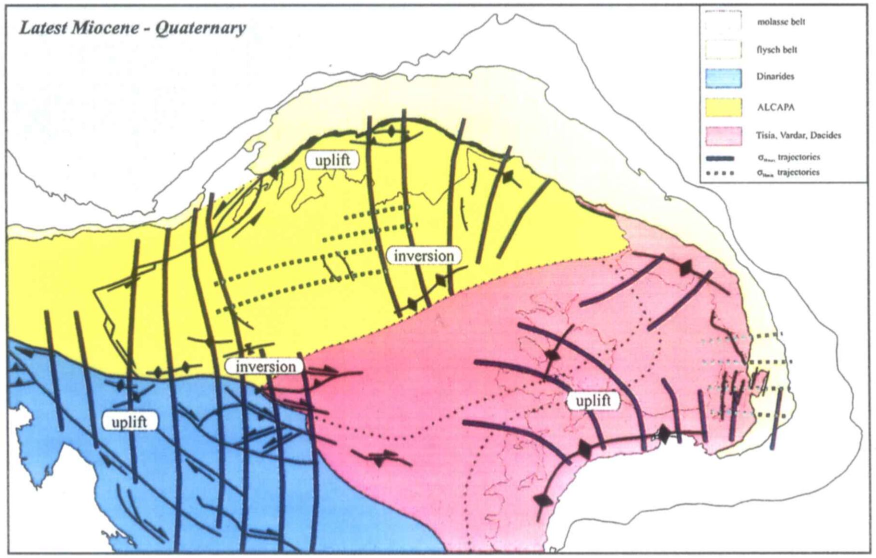
Figure 6b



Fodor et al., Tertiary stress field of the Carpatho-Pannonian area

AV 14_6B

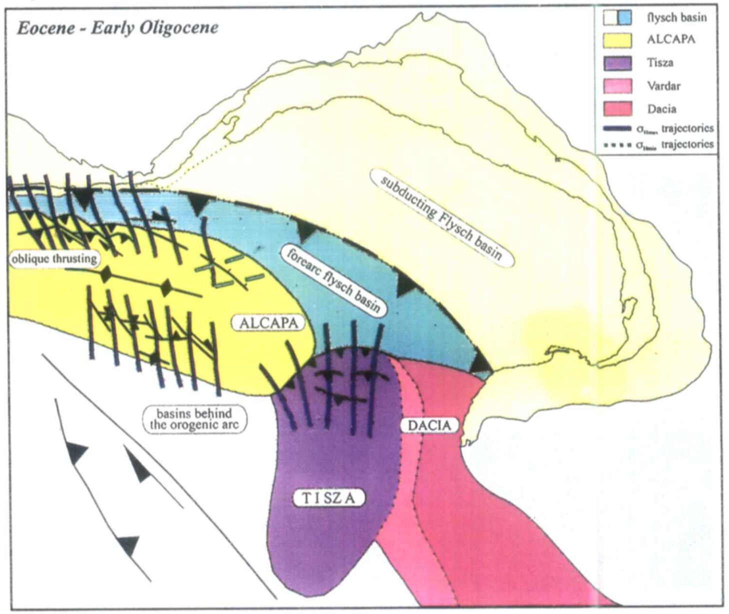
Figure 6g



Fodor et al., Tertiary stress field of the Carpatho-Pannonian area

AV 14_6G

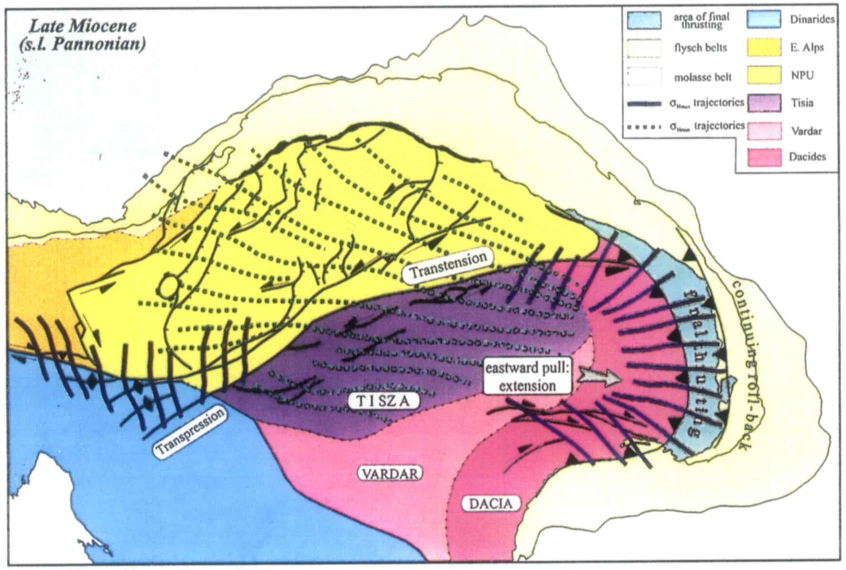
Figure 6a



Fodor et al., Tertiary stress field of the Carpatho-Pannonian area

AV 14_6A

Figure 6f



Fodor et al., Tertiary stress field of the Carpatho-Pannonian area

AV 14_6F