Interpretation of subhorizontal crustal reflections by metamorphic and rheologic effects in the eastern part of the Pannonian Basin

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SUMMARY

The geologic origin of subhorizontal reflections, often observed in crustal seismic sections, was investigated by establishing metamorphic facies and strength of rocks in depth, and correlating these properties to seismic reflection sections from eastern Hungary. Estimation of the depths of metamorphic mineral stability zones utilized the principles developed by Fyfe *et al.* and known geothermal data of the area. The strength versus depth profile was derived by relating local seismic *P*-wave interval velocities to Meissner *et al.*'s activation energy. The results show that the series of subhorizontal reflections, observed in the Pannonian Basin, are a consequence of combined metamorphic and rheologic changes in depths. The synthesis of the integrated data set suggests that the retrograde alteration of the pre-Tertiary basement above the percolation threshold was made possible by the softening effect of shear zones and their water-conducting capacity. The subhorizontal reflections of highest energy, of the consolidated crust below the percolation threshold, originate in the depths of greenschist, amphibolite and granulite metamorphic mineral facies, which were formed in geothermal and pressure conditions similar to those existing today. These results imply the overprint of earlier (Variscan) metamorphic sequences of the crust by more recent retrograde metamorphic processes.

Key words: crustal structure, metamorphism, rheology, seismic structure.

1 INTRODUCTION

The current tasks of geophysical investigations have surpassed the original primary mandate of creating images of subsurface structures. The synthesis of geophysical data now incorporates conclusions considering quality of rocks, their fluid content, and specific physical properties. This investigation—while searching for the conditions, which can produce reflections in the basement and deeper in the crust—also attempts to find the factors influencing the state of rocks.

In the last decades a large number of papers have been published concerning the fundamental nature of the lithosphere and the associated seismic signatures:

Dohr & Meissner (1975) interpreted the lamellae of the lower crust as the result of either intrusions or crystallization seams or peeling of mantle material.

Klemperer (1987) concluded that in the consolidated crust just below the sedimentary cover, generally very few reflections can be found. Below this transparent zone, reflections appear where the temperature is reaching higher than 300°–400°C. As the upper boundary of the zone of rock ductility coincides with that of increasing reflectivity, he suggested that ductility may be the factor in inducing reflectivity.

Christensen (1989) described the origin of subhorizontal reflections, in Inner Piedmont (Southern Appalachian Mts., South Carolina), by investigation of rock samples from deep boreholes, which penetrated rocks of middle and lower crustal origin (upper amphibolite metamorphic facies). The synthetic seismograms, computed from density and velocity properties of drill cores, led to the conclusion that reflectivities are generated by 0.3–13.7 m variations in thickness of silicic and mafic layers, originating in the lower crust by metamorphism, most probably by ductile flow.

In the vertical seismic profiling (VSP) study of James & Silver (1988) in the corridor stack (Leary *et al.* 1988) of the VSP data, subhorizontal reflectors of spatially variable strengths appear to be related to zones of fracturing and hydrothermal alteration. In the Cajon Pass, California deep drill hole and in surface outcrops intense zeolitic alteration of hydrothermal origin is visible. The laumontite mineralization is pervasive within 1 km of the San Andreas fault, decreases in intensity away from the fault until none is present in surface exposures more than 4 km from the fault (Vincent & Ehlig 1988). It appears that the zeolites are zoned near the rehydrated big fractures and this retrograde zoning can be recognized by the subhorizontal reflections.

Lower crustal lamination has variously been ascribed to mafic sills, ductile stretching fabric, cumulate layering, trapped fluids, or some combination of all of these residing within the lower crust (Nelson 1991).

According to Pavlenkova (1996) the low-velocity and low-resistivity zones of high reflectivity, described by Western European authors within the lower crust, in areas of thick continental crust, can also be found in the middle crust. These middle crustal anomalous zones can be recognized by the replacement of the steeply dipping reflections of the upper crust with subhorizontal events. Deeper than 10 km the fluid-saturated fine fissuring causes the low velocity and high conductivity as well as ductility and contributes to intensive metamorphism.

The subhorizontal reflections of the crust are often interpreted as consequences of igneous intrusions, doleritic sills (Holbrook *et al.* 1992; Mooney & Meissner 1992; Juhlin 1990; Ross & Eaton 1997; Mandler & Clowes 1997, 1998), which are tabular intrusions breaking through the older rock complexes (Nelson 1991).

Borradaile *et al.* (1999) studying the Kapuskasing Structural Zone (KSZ) found that seismic reflections of consistent orientation did not conform to the gneissic layering of a 30 km thick Archean granoblastic gneiss bloc of granulite and upper amphibolite grade metamorphism, overthrust from the lower crust to the surface. The plane of the seismic reflectors could, however, be attributed to mineral orientation which manifested itself in magnetic lineation lines. The magnetic fabric is defined through anisotropy of low-field magnetic susceptibility (AMS), and by the anisotropy of anhysterestic remanence (AARM), which isolates the component of the magnetic fabric due to magnetite. Both AMS and AARM are subhorizontal and do not correspond with the visible lithologic layering. Thus, the magnetic fabric and the reflections must indicate the ancient extension direction at depth—originating before the emergence of the KSZ block.

Hurich *et al.* (2001) examining velocity and density data from the Grenville Province of eastern Quebec conclude, that in the upper crust composition dominates the velocity and density structure, while in the middle and lower crust, composition and metamorphism play equally important roles.

The fundamental concepts behind virtually all of the possible geologic settings, which may be responsible for the observed crustal reflectivity zones, are well documented by Mooney & Meissner (1992). Contrary to these theories, the clarification of the source of deep crustal seismic reflections is still a major challenge for seismology, due to lack of direct calibration of observations to known geology. Meanwhile, large-scale crustal seismic-profiling such as COCORP (USA), LITHOPROBE (Canada), BIRPS (United Kingdom), DECORP (Germany) ECORS (France), illustrated that the crust is reflective globally. However, recent developments in mining (hardrock) explorations (Drummond *et al.* 2000; Eaton *et al.* 2003), provide novel perspectives to several kilometres depth about petrophysical properties of crystalline rocks and their associated reflections. These studies, through correlation to boreholes at several locations, demonstrate that special mineral assemblages, lithotologic changes, alterations either of consequence of fluids flows or metamorphism, do generate seismically recognizable geologic environments. A comprehensive study of acoustic properties of minerals and host rocks (Salisbury *et al.* 2003) demonstrates that the acoustic impedance is a consequence of a special, favourable assemblage of minerals or rocks, which were created by a certain process(es).

In Hungary, series of subhorizontal reflections appear in shallow basement depth, and in variable settings. Many of these events were recognized and interpreted with industrial perspectives such as the structural study of the Mecsek mountains (Ráner *et al.* 1979) or the high amplitude subhorizontal reflections of the Transdanubian Central Range, with its associated magnetotelluric conductivity anomaly at a depth of 4–5 km (Pápa *et al.* 1990; Redlerné-Tátrai & Varga 2001). Subhorizontal reflectivity patterns are, however, also evident in other regions of the Pannonian basin (Posgay *et al.* 1995, 1996).

Some of the atypical characteristics of these seismically anomalous zones do not directly correspond to the above mentioned interpretive concepts, thus providing the impetus for the investigations presented here. These new results suggest that the reflectivity patterns are the consequences of metamorphic processes which are controlled by specific pressure and geothermal regimes of the different crustal depths. These recent deductions advance some novel views of the crust and outline a fresh interpretation philosophy for subbasement reflectors within the Pannonian Basin.

2 SUBHORIZONTAL REFLECTIONS OF THE PRE-TERTIARY BASEMENT

2.1 Geologic setting of the study area

The pre-Tertiary basement of the Pannonian Basin is consisted of tectonic ‘terranes’ (Fig. 1) formed in various parts of the Tethys sea. The region of this investigation falls on the Tisza terrane originating from the European (northern) side of the Tethys. Kovács *et al.* (2000) estimate that the pre-Mesozoic rocks of the Tisza composite terrane was subjected to the following three stages of metamorphism:

- between 440–400 Ma: P = 0.95–1.2 GPa and T = 600°–560°C
- between 350–330 Ma: P = 0.4–0.65 GPa and T = 640°–650°C
- between 330–270 Ma: P = 0.2–0.3 GPa and T = 680°–685°C.

The pre-Tertiary basement of the Pannonian Basin consists essentially of Variscan metamorphic rocks (different gneiss, mica schist and amphibolite varieties) and the overlying Mezosoic sedimentary cover formations. As shown by Pogácsás *et al.* (1989), Posgay & Szentgyörgyi (1991), Albu & Pápa (1992), Tari *et al.* (1992, 1999), Lörincz & Szabó (1993), Kovács *et al.* (2000) among others, the structural evolution of the metamorphic basement around the study area was rather complex since the Variscan orogeny especially due to the multiphase subsidence of the Pannonian Basin. Although some of these tectonic events involved only the cover formations, others also affected the basement rocks and contributed to the development of the juxtaposed blocks of diverse metamorphic evolution. At the present, as a result of the above processes, the basement is a mosaic of blocks of very much differing metamorphic evolutions and ages. Following the poly-metamorphic evolution the basement was exhumed and consequently retrogressed.

Secondary mineralization due to retrograde metamorphism and other alteration processes (e.g. propylization, K-metasomatism, weathering) is widespread in the basement rocks. Amphibolite usually contains secondary actinolite rim around hornblende, plagioclase is replaced by albite/oligocklase and ilmenite is rimmed by sphene. Garnet and biotite in different gneiss and mica schist types alter to chlorite, and sericitization of K-feldspar is a common process as well. In addition to the above minerals, micro fissures also contain epidote, clinozoisite, prehnite, pumpellyite, zeolite and carbonate minerals (Juhász *et al.* 2002; M. Tóth 1994). All these phenomena suggest that, although the rocks of Variscan basement are

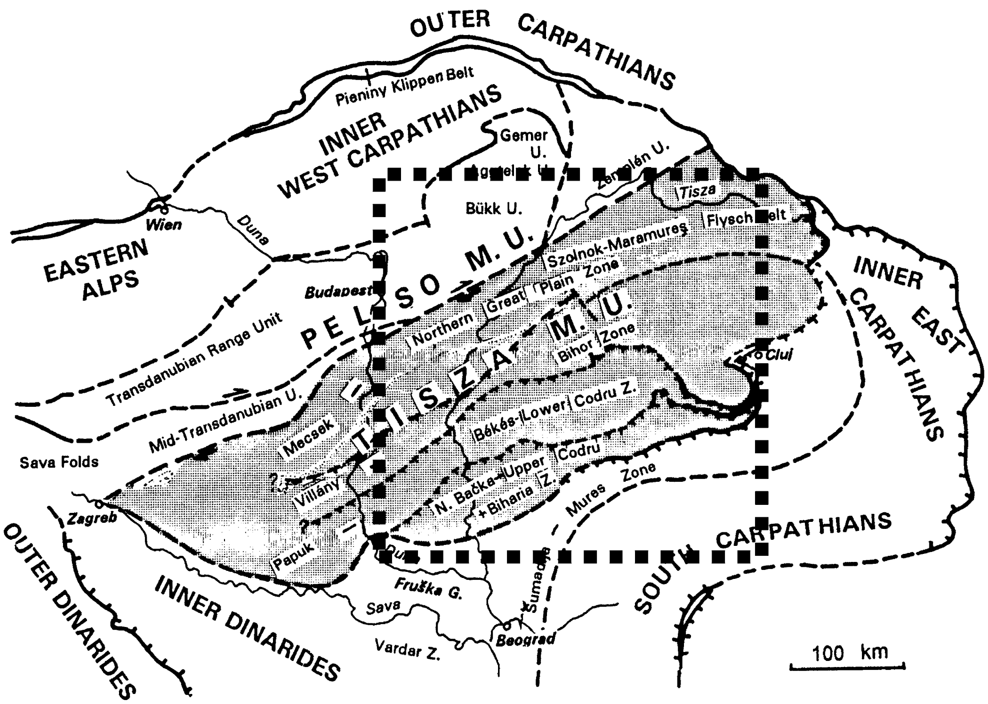

Figure 1. Location map and tectonic sketch of the study area: the Tisza Mega-unit (modified after Haas & Péró 2004).

relatively fresh, they all underwent significant retrograde alteration during the exhumation from the middle crust. Through increasing Fe^{3+} rimwards, chemical zoning of Ca–Al phases (epidote, pumpellyite, prehnite) suggest growth during decreasing temperature (M. Tóth 1994, and unpublished results). Al-in-chlorite thermometry also shows identical tendency. In addition to above, fluid inclusions confirm comparable development. Subsequent fracture filling minerals (calcite and quartz) contain fluid inclusions of decreasing homogenization temperature from >300°C down to 50°C (Juhász *et al.* 2002; Schubert & M. Tóth 2003).

At present, in the uppermost part of the basement penetrated by wells, a whole sequence of retrograde parageneses can be examined. They clearly formed at subsequent depth intervals when uplifting from the middle crust and now overlap in the most exhumed, and most intensely altered rocks. Petrological study of rock samples from the pre-Tertiary basement infers retrogression in a considerable segment of the crust of the basin.

Based on the stratigraphic and facies character of those strata, five structural units can be distinguished. They are, from the bottom upwards: Mecsek–northern part of the Great Hungarian Plain zone, Villány–Bihar zone, Papuk–Békés-Lower Codru zone, Northern Bácska–Upper Codru zone, and Nagybihar zone (Fig. 1). Those structural units contain both Mesozoic and metamorphic Palaeozoic and older series, except the Nagybihar zone, where no Mesozoic formations are known. The upthrusting of those units occurred in the upper Cretaceous (Middle–Upper Turonian).

Tectonic processes leading to the formation of the Pannonian Basin (system) started in the Late Oligocene and/or in the Early Miocene and culminated during the Middle Miocene. The early and major synrift phase can be timed at 24–14 Ma, a consequent thermal subsidence phase is defined at 14–4 Ma (Hámor *et al.* 2001). The crust, which was significantly thicker prior to those developments, extended 1.5 to 2 times of its original extension (Horváth *et al.* 1988; Posgay *et al.* 1996) with considerable up swell of the mantle lithosphere at certain locations (Posgay *et al.* 1995).

Low-viscosity basalt volcanism occurred along faults (Hámor *et al.* 2001). The activity started during the Pannonian times (11.5–3.5 Ma) and ended in the Quartenary. Its total volume is small. Lower Pannonian basalt was encountered in several wells on the Tisza terrane and in seismic sections as well (Posgay *et al.* 1996).

2.2 Seismic images of the Pre-Tertiary basement

Figs 2 and 3 present seismic images of segments of this complex upper crust region. The data originated from oil exploration surveys in eastern Hungary. Their geographic location is marked A and B in Fig. 4. (Data acquisition parameters: vibratory source, 240 fold coverage, 30 m seismometer array spacing, unattenuated natural frequency of seismometers 10 Hz, 12 s recording time, 4 ms sampling rate. Data processing: DMO, wave form and phase corrections, 2-D wave equation migration, deconvolution and depth transformation.)

In the central part of profile A of Fig. 2a dipping shear zone can be observed (*a–a*), which is disturbed in the depth range of 4.5–5.5 km by a series of subhorizontal reflections (*b–b*). Comparable events (*c–c*) at similar depth are evident in profile B of Fig. 3. The origin of those subhorizontal reflections, which appear to be younger than the dipping events, were investigated considering specific metamorphic and rheologic conditions in the above-mentioned upper crustal depths.

Fig. 5 shows the approximate spatial distribution of mineral facies as derived from experimentally determined mineral stabilities—(modified after Fyfe *et al.* 1978). The metamorphic facies are normally named after one of the characteristic assemblages found in metabasalt. In the same figure marked also is the average temperature versus depth curve of the region from the vicinity of the seismic profiles (Dövényi *et al.* 1983; Cermák & Bodriné-Cvetkova 1987;

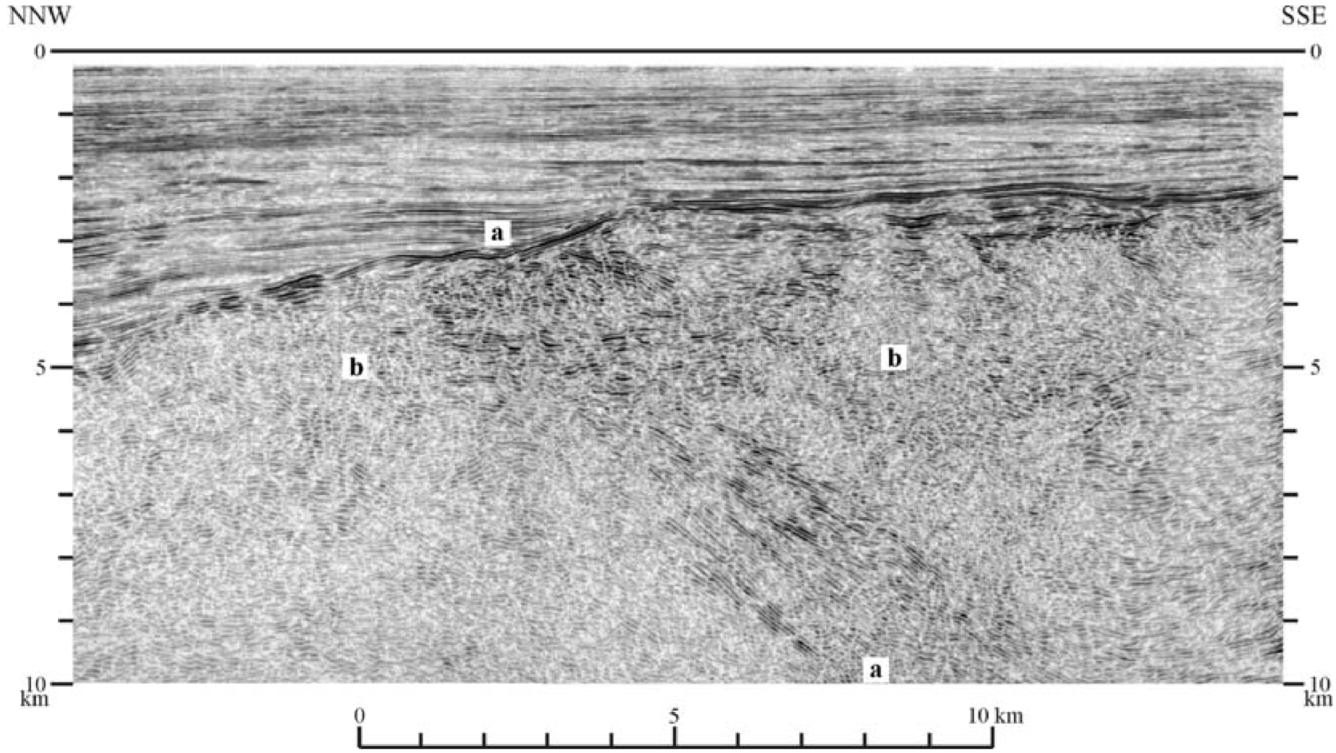

Figure 2. A part of an oil-exploration seismic depth section from eastern Hungary. Its location is marked A in Fig. 4. In the depth range of 4–5 km, subhorizontal reflections *b–b* overprint the shear zone *a–a* within the basement. We suppose that above the percolation threshold the water content of the shear zone made the retrograde alteration of the metamorphic complex possible.

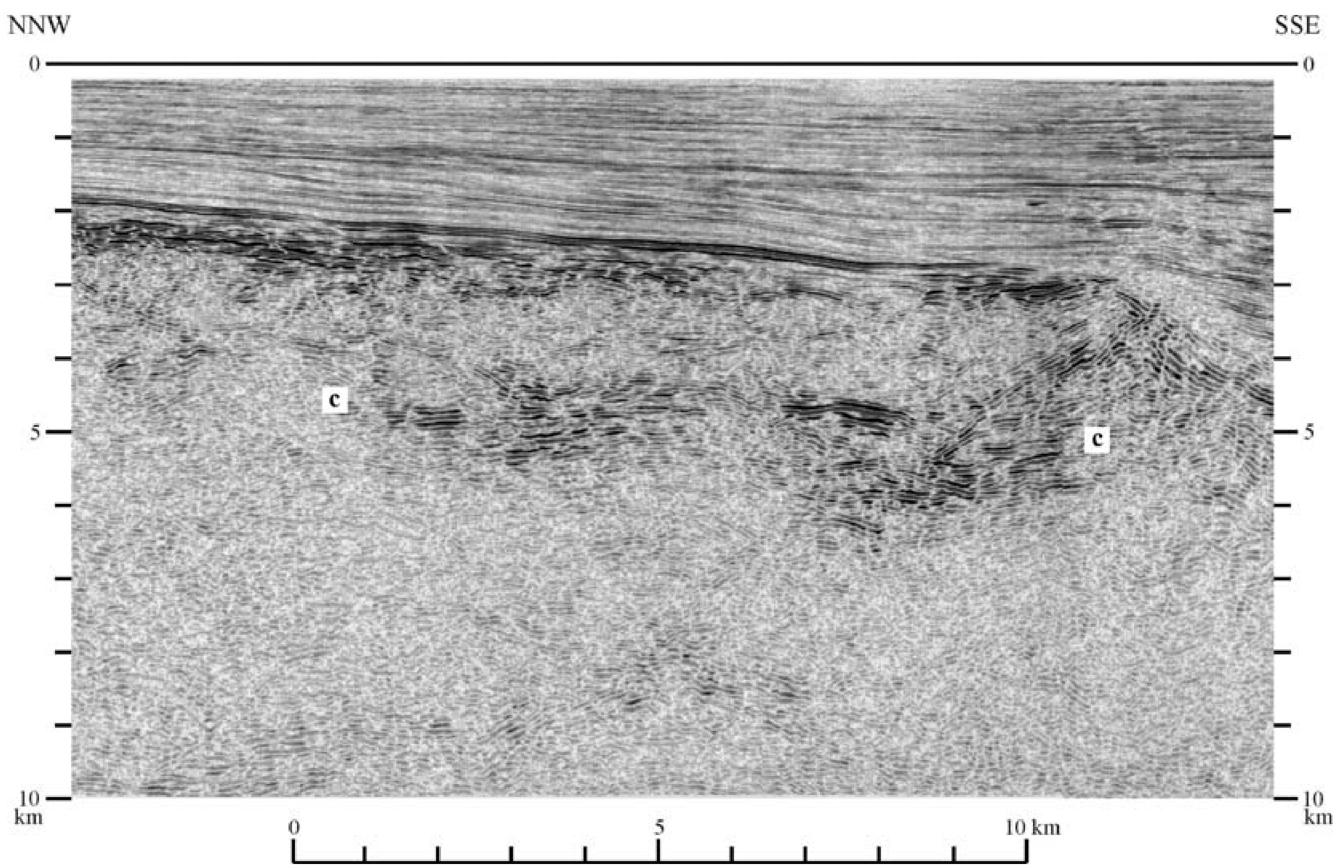

Figure 3. A part of an oil-exploration seismic depth section, marked B in Fig. 4. Subhorizontal reflections *c–c* can be observed in the same depth range as *b–b* in Fig. 2.

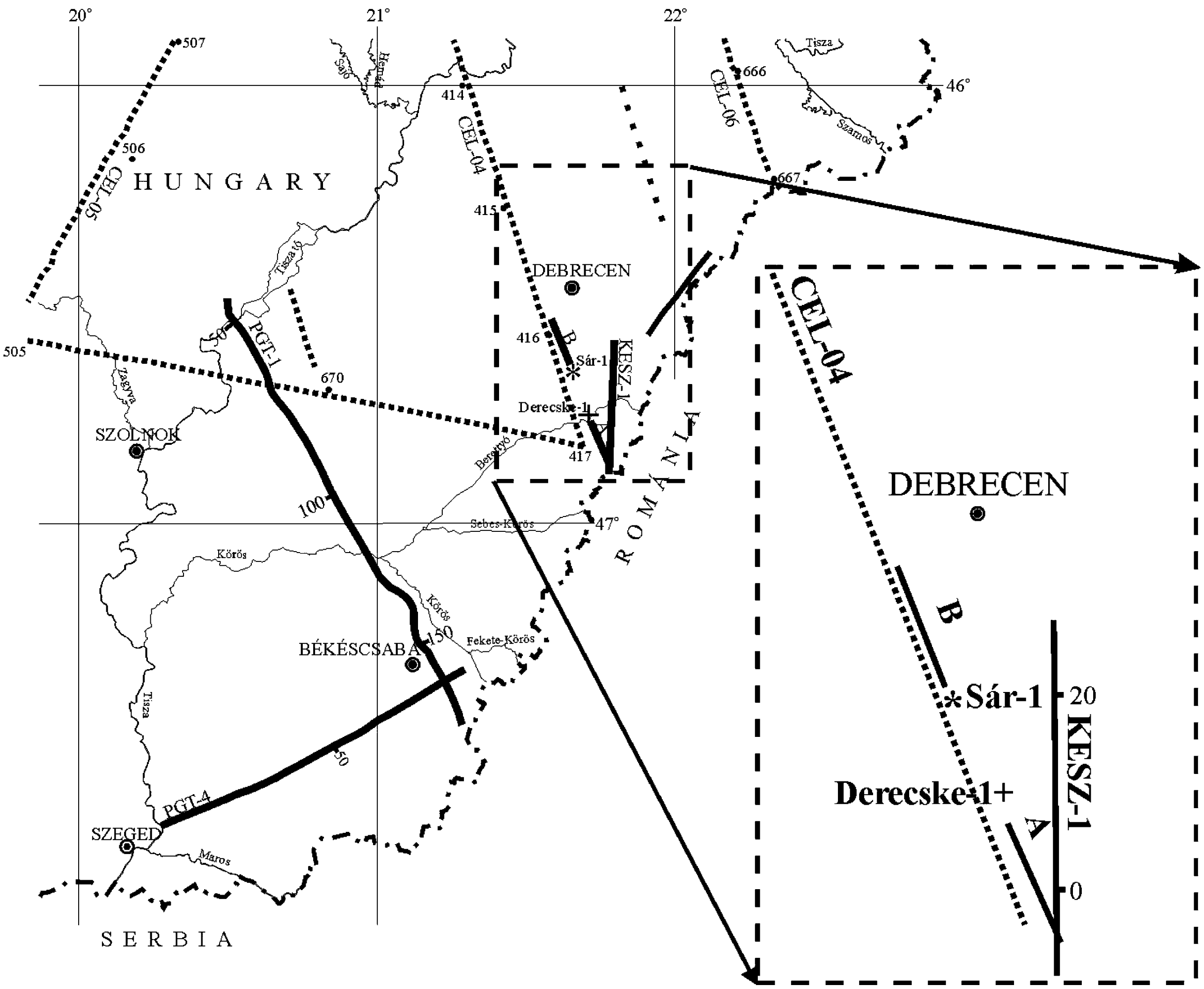

Figure 4. Location map of deep seismic profiles (solid lines) and deep boreholes in Eastern Hungary. The dotted lines of the CELEBRATION experiment (Guterch *et al.* 2000) are used to determine the velocity and ray path density section of profile CEL 04 by 3-D tomography (Fig. 6).

Lenkey 1999) and the interval velocity-depth values determined from KESZ-1 (Fig. 4) deep seismic survey (Posgay *et al.* 1981, 1986).

According to Fyfe *et al.* (1978) prograde metamorphism is accompanied by loss of water as well as by compaction. If the rock is subjected to low *P–T* conditions after suffering high-grade metamorphism, it can again absorb a small amount of water but retains its original structural fabric. Fyfe *et al.* conclude that 'if dry high-grade rocks are pervasively rehydrated, overthrusting may be indicated'. Consequently, in the current case, the zone of reflectivity in the central part of Fig. 2, *a–a*, suggests overthrusting, and the bundle of high-energy subhorizontal reflections *b–b* is connected with the rehydration zone.

Borehole temperatures, in the vicinity of seismic lines A and B are also available to considerable depth. At the southern end of seismic line B, in borehole Sár-1 (marked with * in Figs 4 and 5) at a depth of 4800 m the temperature is 290°C (Árkai *et al.* 1998), meanwhile in the borehole Derecske-1, close to line A, (marked with + in Figs 4 and 5), at a depth of 5200 m the bottom temperature is 250°C (Dövényi & Horváth 1988). These temperature observations are higher than the average temperature values given by the temperature-depth curve of Fig. 5. This significant anomaly is most probably caused by the overlying more than 6500 m thick young sediments of the Derecske trough (Kilényi *et al.* 1991), since the sedimentary fills thermal conductivity is much lower than that of the basement rocks; consequently the temperature near the bottom of the trough is significantly higher than in similar depths of areas where the sedimentary cover is much thinner. Borehole Derecske-1 penetrated 5000 m Neogene sediments while Sár-1 was located on the edge of the Derecske trough, but evidently it is still within the same temperature regime.

In view of the observed data and the presently probable stability zones of mineral facies given by Fig. 5, the subhorizontal reflections of Figs 2 and 3 in the depth range of 4.5–5.5 km are associated with metamorphic phases of the prehnite–pumpellyite zone. It seems, however, expedient to study also these results from the point of view of rheology, as well to interpret the data by considering both approaches.

3 RHEOLOGY STUDIES

Stress regimes originating within the Earth may continue from a few seconds to several hundred million years. Rheologic response of rocks to these stresses of varying magnitude and duration may extend from elastic to ductile reactions (Ranalli 1995). Until the oscillations of atoms do not surpass the potential threshold of the cohesive forces, the response of the rock will be elastic (generally in the case of the effect being shorter than 10 million years). Within this elastic regime the deformation is proportional to stress. Tectonic

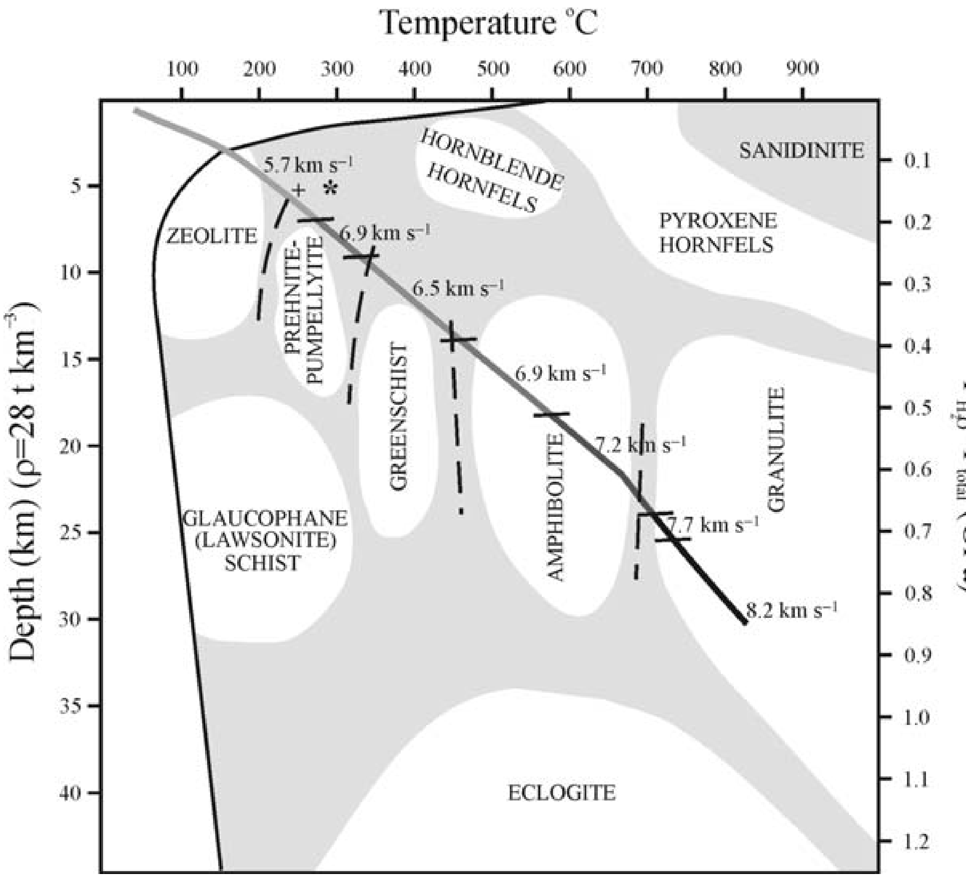

Figure 5. Approximate locations of mineral facies in terms of experimentally determined mineral stabilities (modified after Fyfe *et al.* 1978). Facies are normally named after one of the characteristic assemblages found in metabasalt. The interval velocities (Posgay *et al.* 1981) and the average temperature versus depth diagram characteristic of the area (Dövényi *et al.* 1983; Cermák & Bodriné-Cvetkova 1987; Lenkey 1999) are also plotted. Bottom-hole temperatures of oil-exploration boreholes Sár-1 (Árkai *et al.* 1998) and Derecske-1 (Horváth *et al.* 1988) are marked with * and +, respectively.

force depends on direction, that is, involvement of maximum (σ_1) and minimum principle stress (σ_3) components. If the differential stress ($\sigma_1 - \sigma_3$) surpasses the yield stress point of the rock, it suffers permanent failure. Beyond this stress range of elastic behaviour the material either fractures or undergoes plastic flow.

Rheology of the lithosphere can be characterized by a strength versus depth diagram. Strength, in this diagram, is described by the differential stress [or by its logarithm: log ($\sigma_1 - \sigma_3$)] at which the cohesion of the rock ceases to exist. Constructing this diagram, for several orders of magnitude laboratory steady-state data (Chen & Molnár 1983; Strehlau & Meissner 1987) were extrapolated.

In the elastic part of the lithosphere, rheology is described by the value of frictional failure (Sibson 1974; Byerlee 1978), taking into account the pressure of the pore fluid as well (Ranalli & Murphy 1987). For the ductile segment of the lithosphere, where deformation takes place in the form of steady-state creep, the stability computed by a power-law empirical function (Kirby 1985) is lower than the one computed by principles based on friction, therefore, the power-law function is implemented.

The models of the crust published in the literature consists only one or two layers (Chen & Molnár 1983; Strehlau & Meissner 1987; Kusznir & Park 1987; Bodri 1995). Fundamental limitations of these models are that they divide the crust into too few isotropic layers and the characteristic physical parameters of these layers are derived from velocities obtained by laboratory measurements of samples under high pressure and temperature (e.g. Christensen 1979). Consequently the models could not account for intralayer parameter variations.

In an attempt to minimize the errors arising from neglecting those internal layer properties, the empirical relation between the longitudinal interval velocities and the activation energy of rocks,—$V_p - E$ diagrams—(Meissner 1989; Meissner *et al.* 1991) is considered in parameter computations. The lithosphere, in the area of interest, is divided into a number of physically distinct intervals. The division is based on the interval velocities, derived from local seismic data (Posgay *et al.* 2001). The empirical relations of Meissner *et al.* (1991) made it then possible to estimate the rock parameters utilizing *in situ* velocity–depth curves instead of generalized laboratory data. The physical basis of Meissner *et al.*'s diagram is the presumption that increasing packing of minerals increases both activation energy and acoustic wave velocity. Up to now we have found no reference in the literature that considered these principles to generate a rheologic model of the lithosphere.

For the upper crust, we have computed the frictional stress by Sibson's (1974) relation:

$$\sigma_1 - \sigma_3 = \beta\rho g z(1 - \lambda), \tag{1}$$

where

σ_1 is the largest, and

σ_3 the smallest principal stress, Pa, in the extension phase of the Pannonian Basin (Huismans *et al.* 2001) when subhorizontal

Table 1. The strength envelope curve at point X of seismic line A (Fig. 7) was computed by utilizing the parameters of certain rows at the belonging depth shown in the first column.

H km	V_{int} km s^{-1}	E KJ mol^{-1}	A MPa^{-n} s^{-1}	n	T K$^{\circ}$	Reference	Material
3.5	5.2	125	2×10^{-4}	1.9	475	Ranalli & Murphy (1987)	wet granite
4	5.7	137	2×10^{-4}	1.9	500	Ranalli & Murphy (1987)	wet granite
6	5.5	132	2×10^{-4}	1.9	570	Ranalli & Murphy (1987)	wet granite
7	5.7	137	2×10^{-4}	1.9	600	Ranalli & Murphy (1987)	wet granite

reflectors were formed, according to the classical theory of Anderson (1951) σ_1 was vertical (see point 6.3),

β is a numerical parameter depending on the type of faulting (our procedure was carried out using $\beta = 0.75$, which means normal faults),

ρ is the average density, kg m^{-3},

g is the gravity acceleration, m s^{-2},

z is the depth, m,

λ is the pore fluid factor (in the upper crust, supposing the pore pressure to be hydrostatic, $\lambda = 0.36$).

In the ductile zone we used the so-called Dorn relation (Ranalli 1995) with the necessary modification:

$$\sigma_1 - \sigma_3 = \left(\frac{\varepsilon'}{A_D}\right)^{1/n} \exp\left(\frac{E}{nRT}\right), \tag{2}$$

where

ε' is the creep rate, s^{-1}, (our procedure was carried out using $\varepsilon' = 10^{-15}$ s^{-1}, Cloetingh *et al.* 1995; Posgay *et al.* 1996),

A_D is a material constant (Dorn parameter), Pa^{-n} s^{-1},

E is the creep activation energy, J mol^{-1},

n is the stress exponent,

R is the gas constant,

T is the temperature, $^{\circ}$K.

To get an indicative model the activation energy E was estimated for the consecutive depth intervals from seismic interval velocities by the help of Meissner *et al.*'s (1991) $V_p - E$ diagram.

From published laboratory data of Hansen (1982), Hansen & Carter (1982), Ranalli & Murphy (1987) and Ranalli (1995) such A_D and n data were selected for what approximately the same E values were determined. A further criteria in selecting a parameter from the wide range of laboratory data was to find those rocks which provided similar velocity values under high pressure and temperature in laboratory tests as those determined in profile CEL 04 and KESZ-1, and whose occurrence in the given interval is probable (e.g. Christensen & Mooney 1995). At boreholes that reached the basement, the associated rocks were also taken into consideration. The creep parameter values of crustal materials (Tables 1–3) are only indicative, and quantitatively assessed error limits can be large (Ranalli 1995). In the literature dealing with rheological calculations numerous papers describing the possible errors of those calculations can be found (e.g. Chen & Molnár 1983; Kusznir & Park 1987; Strehlau & Meissner 1987; Scholz 1988; Ranalli 1995; Fernández & Ranalli 1997). The absolute values on the stress scale are not better than indicative, but the relative values of the strength envelope curve (Figs 7, 8 and 10) show systematic vertical variations comparable to the approximate locations of mineral facies in terms of experimentally determined mineral stabilities (Fig. 5) and also to the seismic sections showing likely the compositionally layered crust (Figs 7–10).

4 INTERPRETATION OF SUBHORIZONTAL REFLECTIONS IN THE PRE-TERTIARY BASEMENT

Implementing the new background perspectives for the interpretation of the high-amplitude subhorizontal reflections of Figs 2 and 3, required the following steps:

(1) Determination of the necessary lateral and depth distribution of velocity and ray path density along the relevant portion of CEL 04 seismic profile (Fig. 6). The best fit values are derived through 3-D tomographic inversion of the first-arrival traveltimes (Zelt & Barton 1998) of northeastern Hungary. The location of this survey line is nearly coincidental with lines A and B of this investigation (Fig. 4). The CEL 04 line (Fig. 4) is a part of the Central European Lithospheric Experiment Based on Refraction, 2000 (CELEBRATION 2000) project (Guterch *et al.* 2000; CEL Org. Com. and Exp. Team 2001). This major multinational survey was targeted to investigate the structure and evolution of the complex collage of dominant tectonic features in the region of the Trans-European suture zone (TESZ), the southwestern portion of the East European craton, the Carpathian Mountains, the Pannonian Basin, and the Bohemian massif. It was also complimented by specially designed experiments (e.g. Hajnal *et al.* 2004).

The determination of the spatial velocity field of northeastern Hungary was started with larger cell dimensions ($10 \times 10 \times 5$ km). To increase resolution of the tomographic reconstruction the inversion continued with gradually reduced cell sizes ($5 \times 5 \times 2$ km, $2 \times 2 \times 1$ km). The 3-D tomography inversion permitted the display of a velocity profile (upper part of Fig. 6) and the plotting

Table 2. The strength envelope curve at point Y of seismic line B (Fig. 8) was computed by utilizing the parameters of certain rows at the belonging depth shown in the first column.

H km	V_{int} km s^{-1}	E KJ mol^{-1}	A MPa^{-n} s^{-1}	n	T K$^{\circ}$	Reference	Material
2.5	5.7	137	2×10^{-4}	1.9	424	Ranalli & Murphy (1987)	wet granite
3.75	6.9	235	$3, 3 \times 10^{-4}$	3.2	500	Ranalli & Murphy (1987)	anortozite
5	6.1	173	$3, 2 \times 10^{-2}$	1.9	540	Hansen (1982)	wet quartzite
7	5.9	167	$3, 2 \times 10^{-2}$	1.9	640	Hansen (1982)	wet quartzite
8.5	6.1	173	$3, 2 \times 10^{-2}$	1.9	643	Hansen (1982)	wet quartzite

Table 3. The strength envelope curve of seismic section presented by Fig. 10 was computed by utilizing the parameters of certain rows at the belonging depth shown in the first column.

H km	V_{int} km s^{-1}	E KJ mol^{-1}	A MPa^{-n} s^{-1}	n	T K °	Reference	Material
3.6	5.7	137	2×10^{-4}	1.9	420	Ranalli & Murphy (1987)	Wet granite
6.9	6.9	235	$3, 3 \times 10^{-4}$	3.2	523	Ranalli & Murphy (1987)	Anortozite
9.3	6.5	220	$1, 3 \times 10^{-3}$	2.4	623	Hansen (1982)	quarz diorite
14	6.9	293	$3, 2 \times 10^{-1}$	2.4	773	Ranalli (1995)	ortopyroxene
18.2	7.2	335	15.7	2.6	843	Ranalli 1995	clinoporoxene
24.1	7.7	445	$1, 4 \times 10^{-4}$	4.2	1010	Ranalli 1995	mafic granulite
25.5	8.2	540	4×10^{6}	3	1030	Ranalli & Murphy (1987)	olivine

of a ray path density profile (lower part of Fig. 6) along the CEL 04. The ray path density shown belongs to a cell size of 5 × 5 × 2 km.

(2) The projection of the velocity isolines of the appropriate segment of the line CEL 04 onto the seismic profiles A and B (Figs 7 and 9). The comparatively small (5600–5700 m s^{-1}) values of the velocity field within the pre-Tertiary basement (Fig. 7) suggest the rocks having been exposed to strong stress and strain effects.

(3) The strength envelopes curve at point X of the seismic profile A (Fig. 7) and at point Y of profile B (Fig. 8) were determined utilizing the velocity distribution and the parameters of Table 1 for A and those of Table 2 for B. For both profiles to consider the smoothing effect of the tomography and to simplify the computations the velocity curves were approximated by stepped velocity models. Their breakpoints are shown by Tables 1 and 2. Considering the temperature curves measured in the boreholes Sár-1 and Derecske-1 the presently probable stability zones of mineral facies are presented in accordance with the diagram of Fig. 5.

The structural fabric of the seismic section suggests that after development of the shear zone *a–a* the basement rocks have suffered an additional retrograde alteration phase by receiving water from the intersecting shear zone. The existing near-vertical principal maximum stress and geothermal gradient is leading to subhorizontal foliation *b–b* and overprinting of the shear zone. This interpretation is in harmony with the zonation of zeolite mineralogy as described by James & Silver (1988) and findings of Kozlovsky (1987) in the ultra-deep borehole of the Kola peninsula where retrograde metamorphism was found in fault zones combined with disintegrated amphibolite and granitic gneiss zones.

Below the subhorizontal reflections a low-velocity domain (<5600 m s^{-1}) $d - d$ is found. The origin of it is partly connected to temperature effects (Holbrook *et al.* 1992) and partly to the overpressure of the pore fluids below the zone of recrystallization (Hyndman 1988). The high-velocity domain (>5800 m s^{-1}) $e - e$, near to the basement-sediment contact, may represent the most dens segment of the rocks (Meissner 1986a, p. 34) subjected to retrograde hydrothermal alteration.

(4) The subhorizontal reflections *c–c* of profile B are again characterized by higher-velocity zones (Fig. 8). Associated with the zone of subhorizontal reflections *c–c*, also here similarly to profil A (Fig. 7), there are anomalously high-velocity segments (>6500 m s^{-1}) marked *f* and *g* (Fig. 8). Those are again regarded (Meissner 1986a, p. 34) as the most progressive densification effects of the retrograde alteration. On the velocity distribution based strength envelope profile the depth range of the subhorizontal reflections zone *h* is identified as the brittle–plastic transition zone. Kohlstedt *et al.* (1995) consider it as a semi-brittle region.

(5) Considering the velocity values within the *b–b* and the *c–c* zones, the temperature values of boreholes Derecske-1 (Dövényi & Horváth 1988) and Sár-1 (Árkai *et al.* 1998), and the appropriate mineral stability zones (Fig. 5), the most-likely mineral stability zone of the anomalous *b–b* and *c–c* segments were indicated on Figs 7 and 8. It can be seen that the subhorizontal reflections fall in the presently existing prehnite–pumpellyite metamorphic facies zone. On the right-hand side of the Fig. 8 are also presented the major geologic stratigraphic units of the borehole Sár-1. From the borehole samples in the depth interval of 2925–3846 m Árkai *et al.* (1998) established that the Variscan or pre-Variscan (300–400 Ma) prograde metamorphism altered the polymetamorphic overthrust block to the amphibolite facies. The retrograde metamorphism of prehnite–pumpellyite and subordinately greenschist facies of this block is terminated with mylonitization and formation of cataclasites. (On the basis of crystallinity measurements the approximated temperature was of *ca.* 300°C, a thermal effect between anchizone and epizone).

The underlying Mesozoic—probably Triassic—rocks (recognized between 3846 and 4800 m) are para-autochtonous (Árkai *et al.* 2000). Their upper portion suffered a low-temperature anchizone prograde metamorphism, while the lower segment belongs to the epizone (greenschist facies) and the transition zone of anchi metamorphism and epizone metamorphism. The K–Ar isotope geochronological results suggest, that this event and the retrograde metamorphism of the overthrust block occurred during or slightly preceding the Austrian and/or Subhercynian compressional events (80–105 Ma) which are widespread in the region.

Subhorizontal reflections (*c–c*) appear slightly to the NNW of the borehole Sár-1 and deeper than the bottom of the borehole. However, the structural dips of the core samples are comparable to events *i–i* when projected into the section of Fig. 8.

The two SSE dipping shear zones, $j - j$ and *k–k* (Fig. 8), bordering the high-velocity blocks *g* and *f* are recognized regional tectonic trends. Kovács *et al.* (2000) describe the trend as an Austrian and/or sub-Hercynian compressional event. These two zones are overprinted by the younger retrograde alteration generating subhorizontal reflections. The strong reflectivity contrasts are indications, that the retrograde recrystallization has changed the *P*-wave velocity and stability of the rocks. This analysis is supported by the detailed, velocity distribution determined by tomography and, furthermore by the relative maximum value of the strength profile (elastic–brittle to ductile transition zone) presumed in this depth.

5 INTERPRETATION OF SUBHORIZONTAL REFLECTIONS IN THE MIDDLE AND LOWER CRUST

Subhorizontal, high-amplitude reflections ($l - l$, *m–m*, *n–n*, *o–o*, *p–p*), below the percolation threshold, can also be recognized on deep sounding seismic profiles with focus to investigate the lithosphere and asthenosphere. (Acquisition and processing data are given in

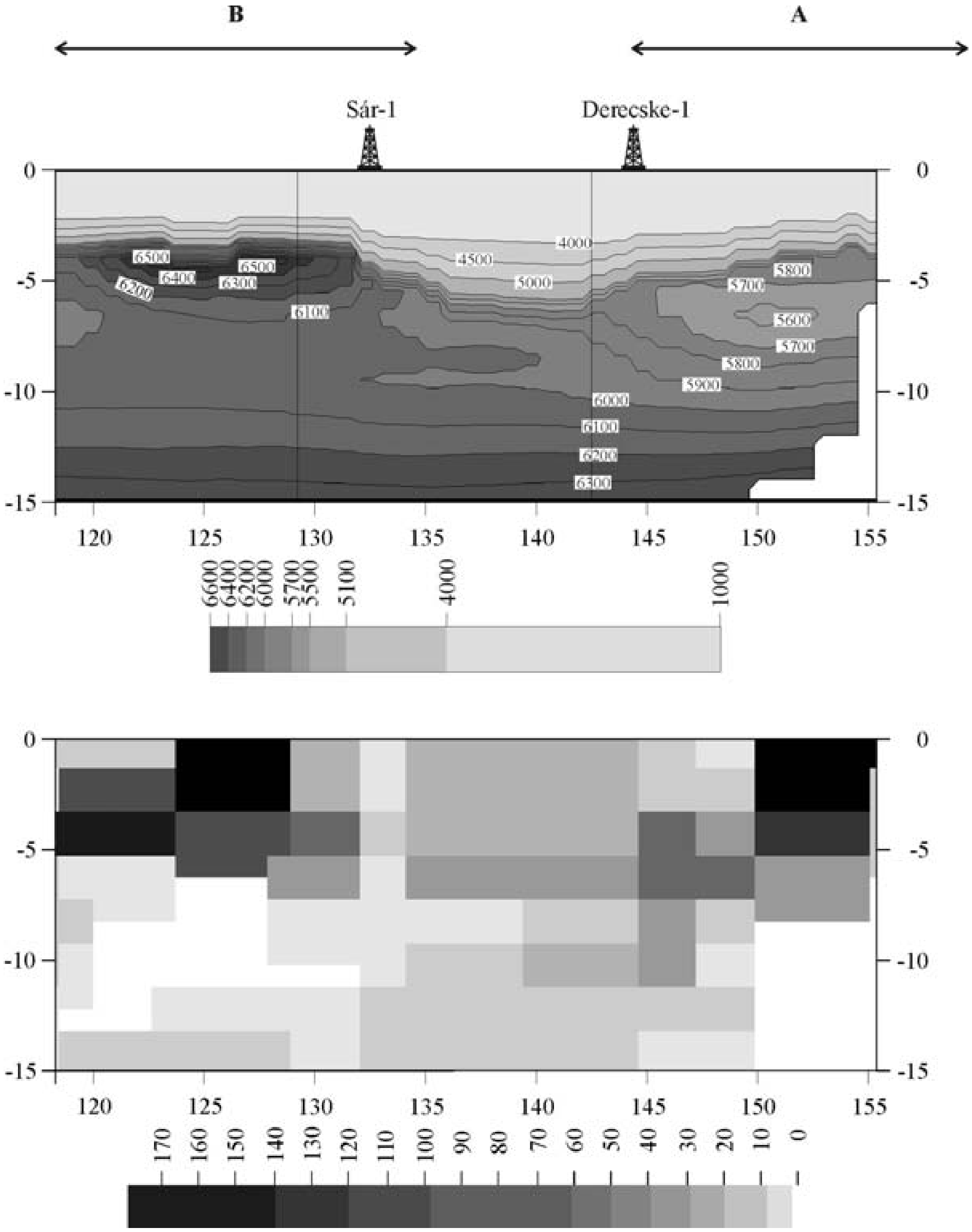

Figure 6. Tomographic velocity and ray path density section of profile CEL 04 derived through 3-D tomographic inversion for the first-arrival traveltimes of northeastern Hungary. Lines A and B mark the projected positions of oil-exploration seismic profiles A and B. The profiles of CELEBRATION experiment indicated by dotted lines in Fig. 4 are also used by 3-D tomography inversion. The scale of the upper profile represents velocity in m s^{-1}. The lower scale is ray-density per unit area. The diagram indicates, that within the region of interest along profiles A and B the ray density is high.

Posgay *et al.* 1995, 1996). Fig. 9 presents a part of the reflection depth section of profile PGT-1 (modified after Posgay *et al.* 1995). Its location is shown in Fig. 4. The pre-Neogene basement is marked by *B–B* and the crust-mantle boundary by *M–M*. In the consolidated crust several steeply NNW dipping zones (*q–q*, *r–r*, *s–s*) are evident through their low amplitude signal characteristics. The low signal levels of these zones are attributed to local heterogeneous velocity intervals within complex displacement zones.

Fig. 10 displays a part of the migrated depth section of the same profile (Posgay *et al.* 1995). The specific interval velocities of the crust (left side of the diagram) were obtained from the data of KESZ-1 deep seismic profile (Posgay *et al.* 1981, 1986). The strength envelope-depth profile computation utilized those velocities and the parameters are listed in Table 3. The diagram of Fig. 10 marks also the zone of 300°–400°C temperature interval, which—according to Klemperer (1987)—highlights the beginning of the reflecting lower crust, and a regionally determined layer of high conductivity (Ádám 1987). The depth of the latter was calculated by Ádám's empirical relation (1983):

$$H = 1718.7q^{-1.09}, \qquad (3)$$

where

H is the depth, km,
q is the heat flow, mW m^{-2}.

A medium strength reflectivity event ($l - l$ in Fig. 10), in an approximately 7 km depth, between horizontal distance markings 83 and 90 indicates the beginning of the prehnite–pumpellyite grade

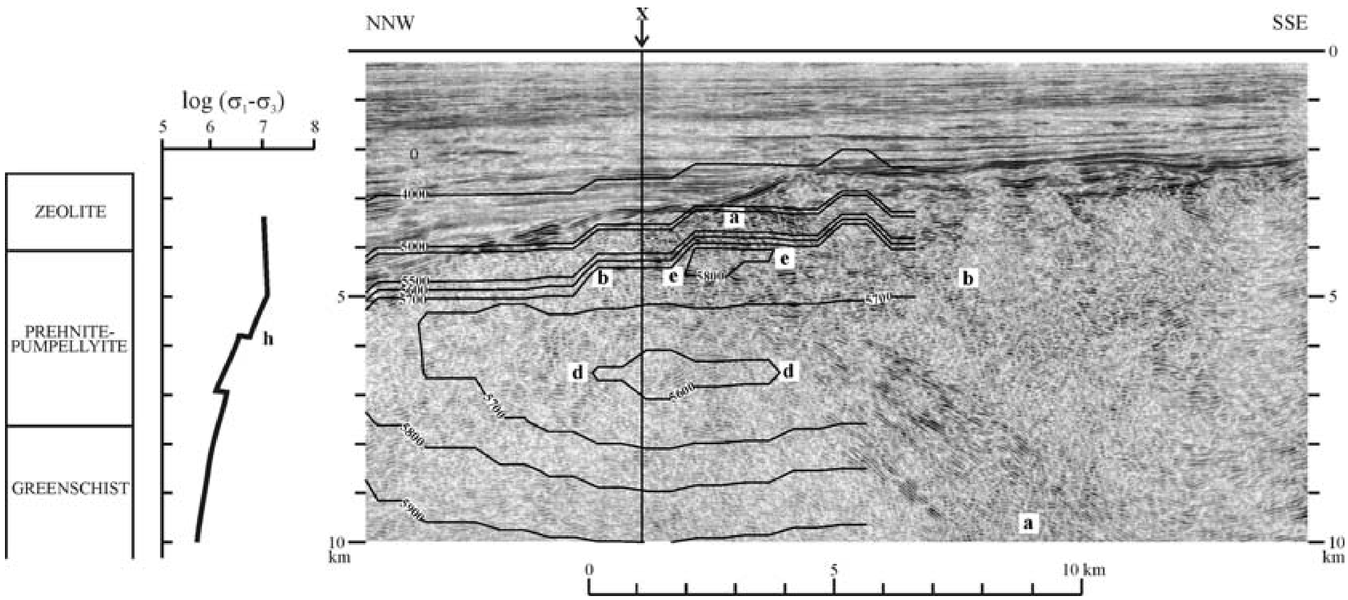

Figure 7. Velocity distribution of profile CEL 04 projected into the depth section of profile A (Fig. 2). Approximate locations of mineral facies in terms of experimentally determined mineral stabilities are presented on the left side (to determine it Fyfe *et al.*'s (1978) results shown by Fig. 5 and the temperature measurements carried out in the Derecske-1 deep borehole (Horváth *et al.* 1988) were used). The indicative strength envelope curve was determined utilizing the velocity distribution at point X and the parameters of the Table 1. The subhorizontal reflections *b–b* appear to be related to hydrothermal alteration connected to the fracture zone *a–a* partly overprinted by the later retrograde events.

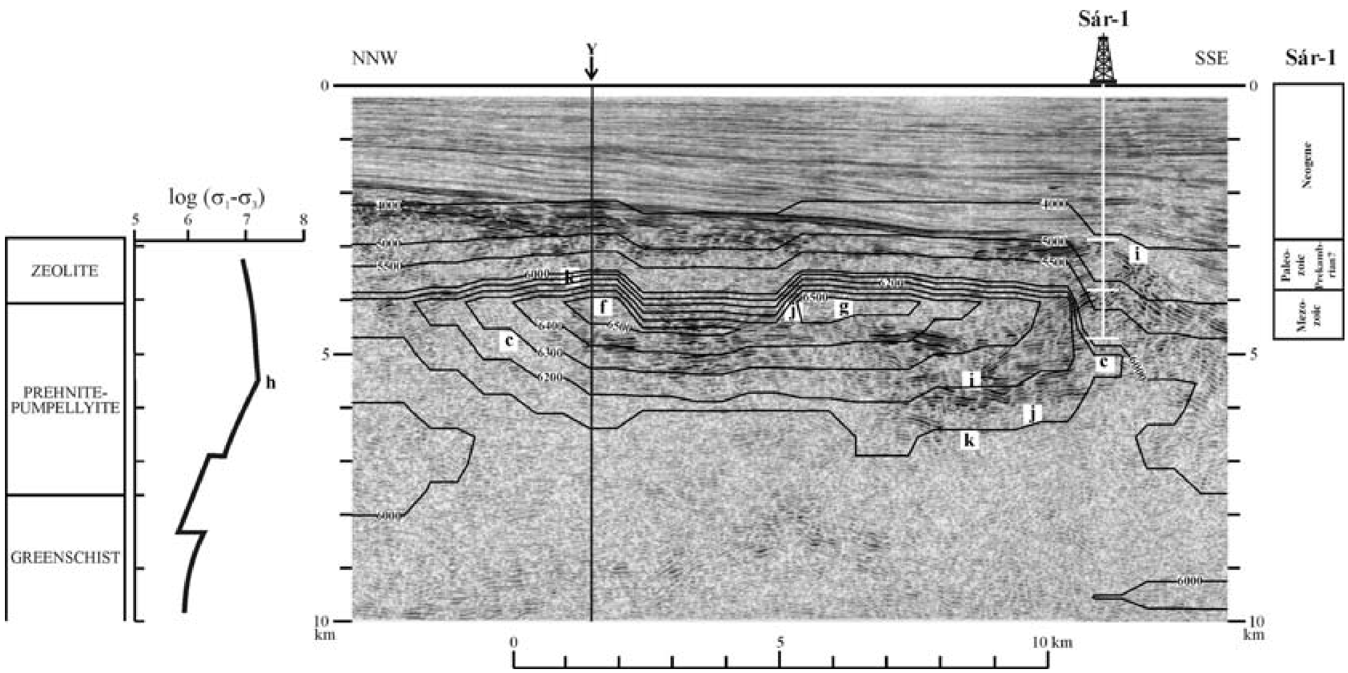

Figure 8. Velocity distribution of profile CEL 04 projected into the depth section of profile B (Fig. 3). Approximate locations of mineral facies in terms of experimentally determined mineral stabilities are presented on the left side (to determine it Fyfe *et al.*'s (1978) results shown by Fig. 5 and the temperature measurements carried out in the Sár-1 deep borehole were used). The indicative strength envelope curve was determined utilizing the velocity distribution at point Y and the parameters of the Table 2. Above subhorizontal reflections *c–c*, the $V > 6500$ m s^{-1} (*f* and *g*) velocity domain may mark the most solid rocks originating during retrograde alteration. Recognized shear zones marked with *i–i*, $j - j$ and *k–k*, are partly overprinted by later retrograde events.

metamorphic band, coincidental with the upper boundary of a zone of relatively high stability (and high velocity).

The dominating subhorizontal reflections (from *m–m* to $m_1 - m_1$ in Fig. 10) in a depth range of 8.5–11 km outline the Klemperer's zone, and the upper part of the greenschist metamorphic facies section. This depth range is also referred to by Kohlstedt *et al.* (1995) as brittle–ductile transition or semi-brittle zone. Associated with subhorizontal reflections $m_1 - m_1$ are some gently dipping events ($\sim 17°$, $m_2 - m_2$). The amplitudes of these events are similar to the subhorizontal ones. We suggest that tectonic influences—occurred

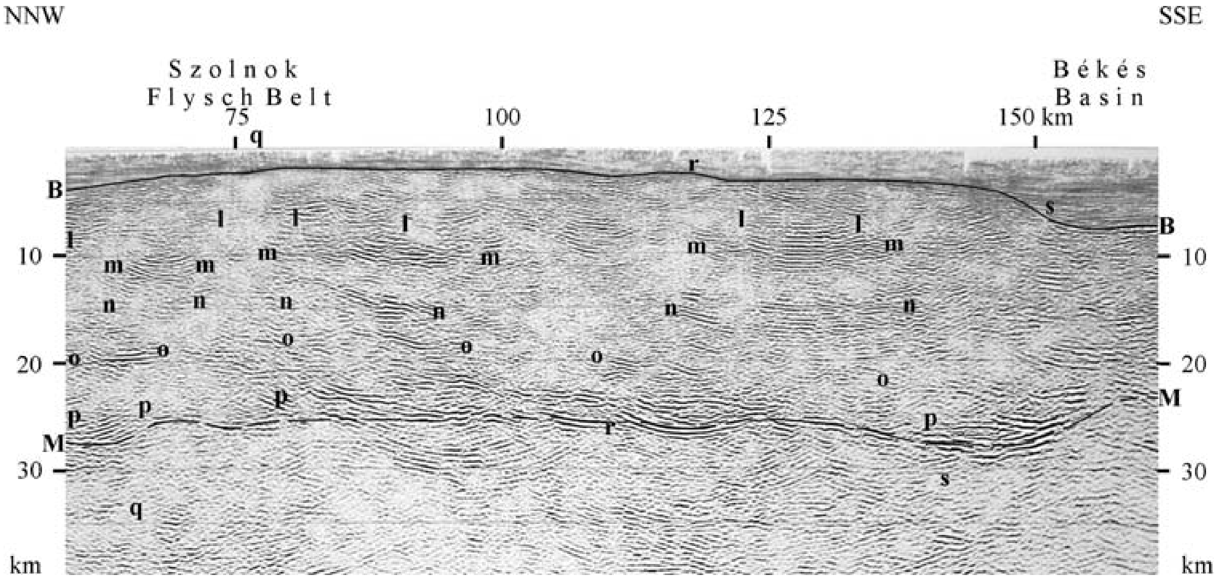

Figure 9. A part of the seismic depth section of PGT-1 (modified after Posgay *et al.* 1995). The deep displacement zones, *q–q*, *r–r*, *s–s* disturb the continuance of subhorizontal reflections. The pre-Neogene basement *B–B*, the crust-mantle boundary *M–M* are also indicated.

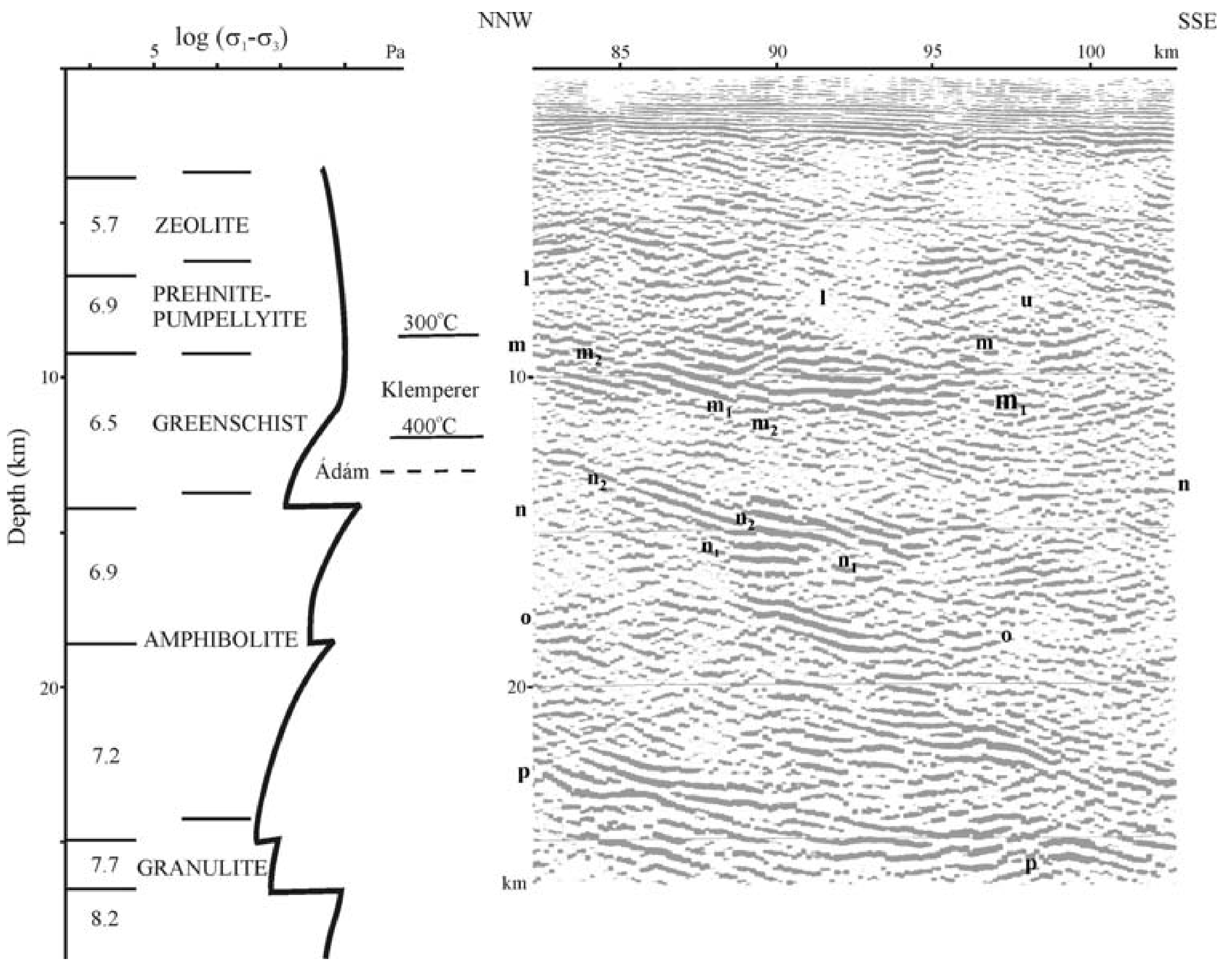

Figure 10. The upper part of a portion of migrated depth section of PGT-1 (modified after Posgay *et al.* 1995). Furthest left there are the interval velocities (Posgay *et al.* 1981, 1986) followed by the estimated metamorphic facies and the strength envelope profile. Also marked are the 300°–400°C temperature range which is according to Klemperer the beginning of reflecting lower crust, and the depth of regional conductive layer of Ádám (1987). Subhorizontal reflections *l – l*, *m–m*, *n–n*, *o–o* and *p–p* mark zones of mineral stability, and coincide with the depth of changes of the strength envelope profile, respectively.

nearly in the time of the retrograde metamorphism- are responsible for these dipping structures.

Above the percolation threshold, *a–a* (Fig. 2) and *i–i*, $j - j$ as well as *k–k* (Fig. 3) are considered as tectonic interfaces. In many places, these were overprinted by presumably much younger subhorizontal reflections. Below the percolation threshold, the horizontal and slightly dipping events *m–m*, *n–n*, *o–o* and *p–p* (Fig. 10) appear essentially by the same amplitudes. It is highly probable, that above the percolation threshold the duration of retrograde metamorphism was shorter than that below the percolation threshold because of the more significant fracturing of rocks and the quantity of volatiles present. The reason that below the percolation threshold the traces of tectonic events appear in the same zone as the horizontal reflections is an outcome of the presumably longer lasting retrograde processes. The dipping interfaces are inferred to be:

(1) either horizons which were formed in an early phase of the retrograde process where their northern parts were subsequently uplifted, while their southern parts remained in original position were blending with subhorizontal horizons created by later stages of the same process,

(2) or horizons, which were formed by earlier movements creating a fragmented segment permitting the flow of volatiles possible.

The approximate depth of Adam's (1987) conducting layer coincides with the ductile zone and with Klemperer's metamorphic process generated free-fluids zone.

Starting at about 14 km depth, bundles of subhorizontal and gently dipping reflections *n–n* with prominent energy levels are visible in Figs 9 and 10. They extend laterally to considerable distances. These reflections mark the upper limit of amphibolite grade metamorphism. Similarly to Mueller *et al.* (1987) and Mueller (1991) we suggest, that reflections in this portion of the crust are an effect of positive change in the acoustic impedance (sudden increase in interval velocities). This bundle of reflections can also be divided into subhorizontal ($n_1 - n_1$) and dipping bands ($n_2 - n_2$).

Hurich *et al.* (2001) by determining velocities and densities of gabbro and metagabbro samples (from greenschist to eclogite phase metamorphism) conclude that reflectivity significantly increases at the transition of the lower to upper amphibolite metamorphic facies. Similarly the reflectivity at around 18 km depth (*o–o*) can also be attributed to sudden velocity increase as a result of the above proposed metamorphic facies.

Earlier investigations in this region (Posgay 1993) credited the reflectivity around 24 km depth (*p–p*) to the remnants of a former crust/mantle boundary and interpreted here as a progress of metamorphic phase to granulite grade. AVO inversion in the vicinity of the Mohorovičić discontinuity (Takács & Hajnal 2000) along PGT-4 deep seismic profile (Fig. 4) recognized a trend comparable to the strength envelope profile of Fig. 10 in the 21–26-km-depth range.

6 DISCUSSION

6.1 Development model of subhorizontal reflections

The depth range of 5–10 km (as confining pressure increases to about 200 Mpa) can be regarded as the uppermost regional limit where reflectivity generated by retrograde metamorphism originates as an influence of residual water. In this depth range, not considering anomalous tectonic disturbances, the original lithostatic pressure (Holbrook *et al.* 1992) closed the primary microcrack systems. Above this percolation threshold (Kornprobst 2002) surface waters circulate in the tectonic zones, where a relatively low extent of fissuring increases the possibility of retrograde alteration by providing the necessary water and increasing permeability. The overthrusts of the discussed area are regarded as the mechanism responsible for fluid migration and initiation of retrograde alteration (Figs 2, 3, 7 and 8). Reflectivity discontinues and the alteration process is ceased where permeability of rocks approximates the state of a former prograde metamorphism.

To understand whether retrograde reactions in the continental crust may or may not be responsible for formation of the subhorizontal reflection packages, intensity of these metamorphic processes should be modelled. To do so, first spatial distribution of the important variables, like temperature, reactive minerals, as well as amount of water available for reactions are to be studied. The metamorphic basement fundamentally consists of diverse gneiss and amphibolite types of medium to high-grade metamorphic rocks. Due essentially to the fast uplift, most mineral phases of these rock types, at the depth in question, become unstable under lower temperature conditions and are not in equilibrium with water. Therefore, they usually tend to be transformed to more stable, hydrous phases.

The upper continental crust contains about 40 per cent plagioclase, which is also the main constituent of the known lithologies of the study area. Investigating H_2O-fixing retrograde reactions of feldspar, deductions can be made about the progress of the evolution of an uplifting continental crust. The most significant mineral reactions of this type in the corresponding Ca–Al–Si–O–H chemical system are the following:

$$4\ \mathrm{An} + \mathrm{H_2O} = \mathrm{Epi}\,(+\mathrm{SiO_2}), \tag{4}$$

$$\mathrm{An} + \mathrm{H_2O} = \mathrm{Prh}\,(+\mathrm{Al_2O3} + \mathrm{SiO_2}), \tag{5}$$

$$8\ \mathrm{An} + 7\ \mathrm{H_2O}\,(+\mathrm{Mg2+}) = \mathrm{Pump}\,(+\mathrm{Al_2O3} + \mathrm{SiO_2}), \tag{6}$$

$$\mathrm{An} + 4\ \mathrm{H_2O}\,(+\mathrm{SiO_2}) = \mathrm{Lmt}, \tag{7}$$

$$\mathrm{An} + 7\ \mathrm{H_2O}\,(+\mathrm{SiO_2}) = \mathrm{Stb}, \tag{8}$$

(An—anorthite; Epi—epidote; Prh—prehnite; Pump—pumpellyite; Lmt—laumontite and Stb—stilbite). In addition to the plagioclase alteration reactions, Ca-Fe-Mg phases tend to react forming chlorite and amphibole, (tremolite) K-feldspar alters to sericite among many others. Besides hydration, also appearance of secondary carbonate phases is common.

The fracture network of the brittle upper crust can be considered water saturated. On the other hand, however, both pore space and permeability (Stober & Bucher 2004) supported by open, water-filled fissures decreases progressively with depth. Long-term hydraulic tests also confirm that hydraulic potential in crystalline basement rocks is continuously decreasing with depth (Stober & Bucher 2004) and does not reach steady state (Schulze *et al.* 2000). From such behaviour, Stober & Bucher (2004) conclude, 'water tends to migrate from shallow levels into deeper parts of the crystalline basement'. They also propose that migration of water to depth and the maintenance of a corresponding hydraulic gradient could result from consumption of H_2O by hydration reactions at depth. Because of the huge amount of reactive plagioclase as well as decreasing fractured porosity, the limiting factor of the mineral reactions is the amount of available H_2O, while the amount of plagioclase is unlimited.

In a numerical model of plagioclase alteration along the upper crust (M. Tóth & Schubert 2005), first the boundary conditions should be given. All mineral reactions (eqs 4–8) are discontinuous

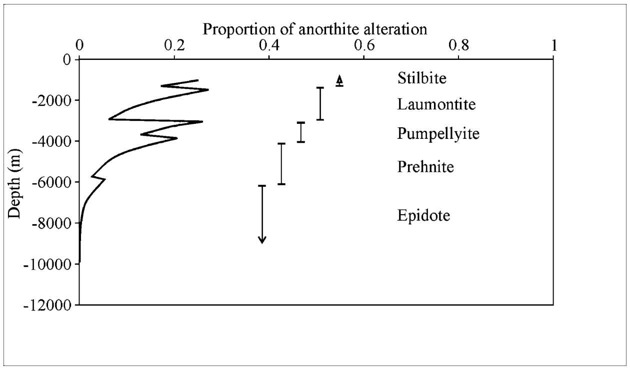

Figure 11. A numerical model of plagioclase alteration of the uppermost part of the crust.

with the following approximate (depending on the pressure) stability fields: stilbite: <80°C; laumontite: 80°–160°C; pumpellyite: 160°–200°C; prehnite: 200°–300°C; epidote: 300°C. Geothermal gradient of the upper part of the crust in the Pannonian Basin varies around 50°C km^{-1}. In our calculations fracture porosity decreases exponentially with depth down to 10 km in the range of 5–0 per cent. From the above simple statements one can conclude that within the stability field of any newly formed retrograde phase, the amount of the altered feldspar decreases with depth. If the amount of consumed plagioclase is plotted with depth, a monotonous curve must be derived, as the amount of water decreases exponentially. At the border of two stability fields, however, the amount of altered plagioclase increases abruptly, because subsequent phases need immediately less water for each mole of plagioclase. Proportions of fixed water in the range of the new phases are the following: Epi 1:4 < Prh 1:2 < Pmp 7:8 < Lmt 4:1 < Stb 7:1. The final model curve for plagioclase consumption rate with depth comes from the product of the exponential porosity curve and the above constants representing each stability field (Fig. 11).

In a real chemical system all mineral phases involved in the above reactions appear as solid solutions. The reactions are therefore continuous in the sense of occurring in a certain temperature (depth) range instead of a sharp T (depth) datum and so clearly result in a bundles of reflections.

Retrograde alteration is the most effective at the borders of mineral stability fields what appear as subhorizontal isotherms in the crust and coincide well with the strongest horizontal reflection packages shown on Fig. 10. Taking also into account that reaction progress increases with temperature, minerals show increasing ability to use gradually decreasing amount of water for retrograde reactions. Similar process may lead to more characteristic retrograde horizons, and consequently reflection packages below greenschist facies conditions.

It is recognized that the crust below the percolation threshold zone (5–10 km) can store residual water of a former prograde metamorphism (Fyfe *et al.* 1978; Kozlovsky 1987). Below this depth range water necessary for retrograde metamorphism is coming from the residual water.

Below depths of 7–10 km, the basement of the region is constituted mainly of pre-Mesozoic crystalline rocks (Kovács *et al.* 2000). This deduction implies that the seismic signatures of the subsurface, from this subsurface interval, are associated with different stages of metamorphic alterations. Comparing the metamorphic grades with those of the regional velocity-related metamorphic facies of Fig. 5, there is good indication that at later stages this segment was also subjected to retrograde metamorphic processes.

Several evidences can be considered to estimate the porosity and fluid content of these rocks:

(i) The high velocities found in the depth range of 5–10 km (Fig. 10) suggest that in this depth the microfissures of the rocks are closed (Holbrook *et al.* 1992).

(ii) Kozlovsky (1987) and his team stated that below the depth of closing microfissures (4.5 km) the water content (and its pore pressure) is generated through prograde metamorphism of the rocks. They presumed that the rock preserved this water for more than 1 Ga. They found foliation and zonality, the latter being of complex hydrophysical character, depending on the rock type and the discharged fluids.

(iii) According to Fyfe *et al.* (1978), during prograde metamorphism the water content and pore pressure rise suddenly and later decrease gradually. This process may vary both in time and its course depending on the metamorphic facies. Both the water content and the permeability of rocks will decrease significantly within 10^8–10^9 yr.

The consequences of these additional specific circumstances, the development of the metamorphic regimes of the basement during the formation of the Pannonian Basin (Horváth & Royden 1981) can be reassessed as the following: the thinning of the lithosphere (and the increase of the heat flow) led to reduction of the depth ranges of the various mineral stability limits. The decreasing rock volume above these stability zones prompted the decrease of lithostatic pressure (and also the effective pressure), while the pore pressure relatively increased. Thus the pore volume and permeability increased, promoting the formation of a new metamorphic facies. Because of the vertical temperature gradient, the layers with macro-scale orientation formed horizontally. These macro-scale orientation then contribute to the formation of high amplitude reflections of the crustal seismic sections.

The retrograde metamorphism in the new mineral stability zones may have limited the upward migration of residual water thus initiating the formation of water-containing zones of relatively high pore pressure (Hyndman 1988). High pore pressure decreased both velocity and stability of rocks, in accordance with Christensen (1989) and Mueller (1991) who interpreted the low-velocity layers in the continental middle crust as the result of pore water—resulting from metamorphism—and relatively high pore pressure as compared with lithostatic pressure.

6.2 Discontinuity of subhorizontal reflections

A prevailing property of the high-amplitude subhorizontal reflections below the percolation threshold is that their lateral continuity does not exceed 10–20 km. This lack of regional persistence of these events can be attributed to tectonic interferences or local perturbations in the metamorphic processes (Posgay *et al.* 2000).

Specifically, along the seismic profile PGT-1 (Fig. 9)—which crosses the Szolnok Flysch Belt—a number of NNW-dipping displacement zones (*q–q*, *r–r*, *s–s*) disturb the continuance of subhorizontal reflections (Posgay & Szentgyörgyi 1991; Posgay *et al.* 1996, 1997). The development of these displacement zones created higher permeability and allowed migration of the accumulated waters of the prograde metamorphism processes (300–400 Ma) creating regions of dehydrated fissures (Fyfe *et al.* 1978). The subsequent reactivation of these strike-slip zones prevented the cracks and fissures from clogging. Subsequently the isopressure levels in the displacement zones deviated from that of the intact parts of the surrounding rocks, therefore the depth ranges of the metamorphic facies stability regions deviated in the two regions. It is also possible, that within the displacement zones, the frequent variability of the pressure and temperature did not allow sufficient time for development of a new stable metamorphic zone.

The disruptive reflection pattern across the subhorizontal reflections on the right-hand side of Fig. 10, between 93 and 103 km and at a depth of 7 km (*u*) is associated with a significant magnetic anomaly (Túrkeve anomaly: Posgay 1967) and is interpreted as the root of an intrusive body.

6.3 Estimation of the age of subhorizontal reflections

The shear zone *a–a* of Fig. 2 is interpreted, based on its direction and structural setting, as an Upper Cretaceous (80 Ma) overthrust, a characteristic of the well established Tisza tectonic unit in the Pannonian Basin (Kovács *et al.* 2000). The bundle of subhorizontal reflections *b–b* overprint zone *a–a*, therefore it must be younger than 80 Ma. Systematization of Alpine metamorphic events in space and time, in the Pannonian basin, is a significant result of the last decade (Árkai 2001). Very few investigations are available in the literature regarding younger alterations (e.g. Mezo–Alpine: 30–40 Ma: Balogh *et al.* 1990; Árkai *et al.* 2000; Árkai 2001; Balogh & Pécskay 2001) and the associated tectonothermal history. Age determination of alteration, on borehole samples, along seismic profiles where subhorizontal reflections were penetrated by drilling is still not available. Extrapolation of published data, however, suggests that the origin of the relatively shallow subhorizontal reflections may be placed in time prior to the formation of the Pannonian Basin.

Model studies of subsidence and thermal history (Dövényi 1994) suggest an initial fast phase followed by an era of slow thermal progression. The thermal isolines show greater perturbations in the initial stage, and consequent slight alterations of their depth position. In the last 2–10 Ma the rate of change of the geothermal gradient did not appear to be significant. In reality, the fast deposition of sediments of the deep basins has modified this model. Heating up of basin sediments was relatively late phase of the tectonic events, for two reasons:

(i) The ascending asthenosphere had to increase first the temperature of the lithosphere (Lachenbruch & Sass 1977),

(ii) Continuous deposition of cold sediments delayed the development of steady-state conditions (Stegena & Dövényi 1983). In certain deep basins thermal equilibrium has still not set in (Posgay *et al.* 1995; Lenkey 1999). Under this thermal subsidence conditions formation of some new stable mineral assemblages are favoured and these are the main origin of the observed subhorizontal reflections in the Pannonian Basin system and in other young extensional areas (Meissner 1986; Mooney & Meissner 1992).

6.4 Potential alternate provenance of the subhorizonal reflectivity

This is the first known investigation attempting to irradiate the origin of the horizontal reflections of the deep subsurface in the Pannonian Basin. A number of geologic models are already documented in the literature, with the same intentions to explicate the origin of seismic subhorizontal refection images from the crust. Brief assessment of some of these currently recognized hypotheses as potential geologic environments, which may be applicable in the Pannonian Basin, inspired the following conclusions:

A Intrusions and underplating by mafic diabase sills is one of the frequently considered geologic models (Dohr & Meissner 1975; Holbrook *et al.* 1992; Mooney & Meissner 1992; Juhlin 1990; Ross & Eaton 1997; Mandler & Clowes 1997, 1998). Within the Tisza unit of the basin, above the Curie temperature ($T_c = 600^\circ$ C), the lower crust could be modelled with significant susceptibility ($K = 1600 \times 10^{-6}$; Posgay *et al.* 1995). Notwithstanding the above, strong subhorizontal reflections were observed in the upper crust above the magnetically modelled zone. As well, from the period when temperature and pressure conditions were comparable to present setting, mafic intrusions are known, but only in small localities and marginal quantity (Hámor *et al.* 2001). Moreover, the reflective zones, based on the current analysis, are associated with amphibolites.

B The cumulate layering hypothesis (Christensen, 1989), built on laboratory observations, suggests that fine-scale layering of altered rocks can produce observable reflection amplitudes in the crust. Essentially this model is comparable to the current inference, by establishing acoustic impedances considering metamorphic processes. Based on its original perspectives, it is a conceivable hypothesis for upper crustal reflectivity. If, however, the original reflective horizons were formed in the lower crust and uplifted at a later stage, it is questionable, that their original horizontal positions could be retained. Since the presentation of the original Christensen model was based on investigation of rock samples, it did not take into consideration the importance of fluids although the importance of these factors was recognized by others (Fyfe *et al.* 1978; Jones 1992). Concurrently, Warner (1990) had difficulty to reconcile the importance of fluids; they required porosity at significant depths.

C The origins of many of the observed lower crustal reflectivity images are rationalized through the stretching fabric model (Phinney & Jurdy 1979; Matthews & Cheadle 1986; Smithson 1986; Allmendinger *et al.* 1987; McCarthy & Thompson 1988; Reston 1988). Although these models have some differences in detail, all consider ductile flow, through a stretching process where minerals are aligned to generate impedance layering. Albeit involvement of extensional forces was observed in the Pannonian Basin, the regional extensional strain suffered by the crust is estimated to be less than two (Horváth *et al.* 1988; Posgay *et al.* 1996). Transposed lithologic layering and micaceous foliation in rocks require orders of higher strain (Nelson 1991) than the documented levels of the basin.

Considering all the prior possibilities and the available geophysical and geological information, our most logical resolution, for the origin of the subhorizontal reflections in the Pannonian Basin, are the retrograde metamorphic processes. Though the question may arise: 'while it appears that the eastern Pannonian Basin shows a good correlation between crustal reflectivity and metamorphic changes with depth, why are there plenty of crustal-scale seismic reflection data that do not show such a correlation?'.

It is likely, that in the Pannonian Basin the series of interfaces evolved in the basement of the Neogene basin by retrograde metamorphism can be recognized and studied relatively easily because the series of mineral facies evolved by retrograde metamorphism took shape under comparatively quiet conditions roughly similar to the presently existing ones and its interfaces are even today nearly horizontal.

The dominant prograde metamorphism took place in the period of 440–270 Ma. Parts of the volatiles originating from these changes were preserved in the crust and provided the possibility for retrograde metamorphism. Geologic evidences indicate that the balance of mineral phases stability, which developed by prograde metamorphism prior to development of the basin, may have been altered during 24–4 Ma. At this time thick sedimentary layers were deposited in rapid succession in the eastern part of the basin. Therefore the evolved series of retrograde mineral facies zones in the basement were able to maintain their nearly horizontal positions.

It can be attributed to the difficulties of recognizing and studying them, that no attempt was made up to now to compare seismic deep reflection data against metamorphic lithology, rheology, geochemical, magnetotelluric and tectonic data within the same study. Though several paper can be found in the literature which demonstrate that experiences by mutual application of different areas have started. Beyond the ones cited in this paper we mention yet the paper of Burwash *et al.* (2000) giving an account of a transcrystallization process observed in a tectonic zone. We assume that similar studies will appear more and more frequently, because they provide not only scientific (e.g. palaeoseismic, palaeotermic, geochemical, metamorphic, etc.) but also industrial (e.g. prospecting of hydrocarbons or minerals, environmental protection, nuclear waste deposit site studies, etc.) results.

7 CONCLUSIONS

The integrated synthesis of the data sets reveals that the reflective horizons of the consolidated crust were mainly generated by temperature and pressure conditions comparable to the presently existing environment with some alterations during the formation of the Pannonian Basin. The depth coincidence of subhorizontal reflections with well defined stability zones of metamorphic facies, and with the relative maximum values of an indicative strength profile, suggest that these subhorizontal events were developed by overprinting after the formation of the Pannonian Basin.

In the crystalline basement, above the percolation threshold, in around 5 km depth, water necessary for the retrograde alteration to generate subhorizontal reflections was provided by intersecting dipping shear zones.

The investigated area lies within the Kunságia terrane of the Tisza structural unit. Deep borehole studies in this region suggest that the basement, below 7–10 km depths, is formed mainly by pre-Mesozoic crystalline rocks. These rocks were metamorphosed during the Variscan orogen at temperatures above 650°C. The presently observable temperature and pressure conditions indicate that these rocks of relatively low permeability below the percolation threshold were subjected to retrograde processes for which the necessary waters were provided by the remnant fluids of the Variscan metamorphism. The new layering with subhorizontal macro-scale orientation is the result of the vertical gravity and geothermal gradient.

Discontinuities of subhorizontal reflections (mainly below the percolation threshold) along steeply dipping displacement zones of the crust, may suggest the absence of retrograde transitions.

The novel ideas presented here may raise significant interest to investigate the subhorizontal reflecting interfaces in areas where the palaeotectonic and palaeothermic evolutions differ from those in the Pannonian Basin, or perhaps study subhorizontal reflections in depth of crust-mantle transition and mantle lithosphere together with xenolites of the same depth intervals.

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