Combined analysis of faults and deformation bands reveals the Cenozoic structural evolution of the southern Bükk foreland (Hungary)

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(structural geology, Cenozoic, Bükk Mts., stress field, deformation band, seismic profile, subsidence history)

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1. Introduction

The deformation mechanism of faulting in porous media is particularly interesting phenomena from tectonic point of view and therefore they have been applied efficiently in structural geological and petrological investigation for a couple of years. Firstly Aydin (1978) defined shear-related deformation structures without discrete fault surfaces in porous sandstone as deformation bands. By definition deformation bands are strain localization structures commonly developed in porous, granular rocks with mm to tens of centimetres offsets. Their length is usually less than 100 m and their thicknesses do not exceed few centimetres (Aydin and Johnson, 1978, 1983). Either they occur as single structures or as zones of several individual but connected features. The porosity and permeability along deformation bands are often reduced significantly forming seals or, along this barrier, extremely important pathway for fluid and hydrocarbon migration (Pittman, 1981; Antonellini and Aydin, 1994; Fossen and Bale, 2007). Deformation bands can be classified by their deformation mechanism as disaggregation bands, phyllosilicate bands, cataclastic bands, and dissolution bands (Fossen et al., 2007). Based on their kinematic attributes deformation bands were divided into simple shear bands and bands with dominant volumetric deformation: compaction, dilation bands and those with associated shear (Aydin et al., 2006).

Classification of deformation bands has revealed important conclusions for cementation and burial history of their host rock. Disaggregation bands form in unconsolidated rocks, where grain reorganization without fracturing the prevailing deformation mechanism. Although the cataclastic deformation bands also develop in unconsolidated rocks (Cashman and Cashman,

2000) but the more progressed cataclasis may refer to deeper burial state of the host rocks. In addition, more deeply buried rocks are more favourable for dissolution and cementation bands.

Due to previously described variable type of deformation bands, they are very important in reconstruction and timing of deformation events affected the studied rocks. However, there is no clear connection of the time interval separating the deposition and faulting; the important condition for deformation band development is the poor induration of sediment. This state could persist for long time if cementation retarded, because of hampered fluid flow and/or lack of overburden and deriving effect of compaction.

This is the reason that although observations on deformation bands can provide data on the relative age of bands and cementation events (indirectly on the thickness of cover sediments) but are, in most cases, insufficient alone to give relatively narrow time spans for the age of band formation, and their role in overall structural evolution. In our paper we intend to present the combination of various methods which together are capable of describing the evolution of fracture system in terms of orientation, stress, and timing. We compared information from deformation bands with field fault-slip data, paleostress analyses, seismic profile analysis and reconstruction of burial history. Fault-slip analysis were inevitable to understand the fault kinematics and paleostress evolution. The interpretation of seismic profiles and geological cross sections extended outcrop-scale fault-slip data into map-scale pattern. 1D subsidence modelling constrained the most important subsidence and erosion events and how much the deformation band-bearing Oligocene sequence might have been buried during fracturing.

2. Geological background

The study area is located in the eastern Hungary, south of the Bükk Mts (NHSSPB, Figs. 1, 2). The Paleogene history of the study area is strongly tied to the evolution of the NHSSPB which is interpreted as a flexural basin situtating in a retroarc position with respect to the Western Carpathian orogenic arc (Fig. 1) (Tari et al., 1993). The beginning of the basin formation is marked by an uppermost Eocene transgressive sequence including terrestrial to shallow-marine clastics and limestone (Báldi and Báldi-Beke, 1985; Nagymarosy, 1990). Due to tectonic subsidence they were covered by shallow bathyal marls, laminated claystone and thick siltstone of Early Oligocene age (Báldi, 1986; Less, 2005) (Fig. 3). The last phase in basin evolution is represented by Late Oligocene to earliest Miocene fine- to medium-grained clastics; however, the Miocene part seems to be missing in the study area. Sedimentation was influenced by eustatic sea level changes and in a minor way by local fault activity (Sztanó and Tari, 1993; Báldi and Sztanó, 2000). The largest such fault in the study area, the Darnó fault acted as a SE dipping reverse fault creating a deep flexural sub-basin (Figs. 1a, 4). The NW vergent reverse fault put the Mesozoic rocks onto the Paleogene sediments proven by borehole Bsz-51 (Figs. 2, 4). The evolution of the NHSSPB terminated by a regional unconformity and denudation in the middle part of the Early Miocene, ca. 23–19 Ma ago (Figs. 3, 4). During the Early Miocene the NHSSPB was dextrally displaced from its original continuation, from the North Slovenian Paleogene Basin by ENE–WSW trending Periadriatic Fault (PAF) and Mid-Hungarian Shear Zone (Fig. 1a) (Csontos et al., 1992, Fodor et al., 1999). Post-19 Ma evolution belongs to the formation, evolution of the extensional Pannonian Basin system (Royden and Horváth, 1988). The rock suites in the Pannonian Basin are divided into pre-rift, syn-rift and post-rift sediments. In the study area, the first rock suites of the basin fill are extended rhyolitic and dacitic pyroclastic rocks produced by many explosive eruptions of late Early to Mid Miocene times (Szakács et al., 1998; Lukács and Harangi, 2002; Lukács et al., 2002; Pentelényi, 2005). The age of lower, middle and upper volcanoclastic levels is 21-18.5 Ma, 17.5-16 Ma and 14.5.-13.5 Ma, respectively (Fig. 3) (Márton and Pécskay, 1998). The age of these volcaniclastic levels was used for timing of deformation phases. During the main rifting phase (17.5-14.5 Ma) several NW-SE trending troughs and few strike-slip related basins were developed in a NE-SW extensional (transtensional) stress field (Fig. 1b). Following the extensive volcanism, late Middle Miocene sedimentation was partly coeval with the third (so-called upper) volcanoclastic level. In late Mid-Miocene several roughly NE–SW trending depressions were formed such as the Vatta-Maklár Trough and the Felsőtárkány Basin south and southwest from the Bükk Mts., respectively (Fig. 2). The previous one was interpreted as a Mid-Miocene transtensional half–graben system bounded by an ENE–WSW trending strike-slip master fault in the south (Tari, 1988). Late Miocene sedimentation was long time considered as post-rift, but new structural data demonstrate ongoing faulting at least up to 8 Ma in the study area (Petrik and Fodor, 2013) and in the whole Pannonian basin (Fodor et al., 2013) (Figs. 1c, 4). The thick Late Miocene (Pannonian in local term) sedimentary unit consists of lacustrine marlstone, followed by basin floor sandy turbidites, clayey slope deposits, sandy deltaic and variegated fluvial suites (Magyar et al., 1999, Sztanó et al., 2013). In the study area these sediments were deposited in synsedimentary NE-SW trending Vatta-Maklár Trough due to NW-SE extension (Tari, 1988) (Figs. 1c, 4). The study area is dominated by N–S to ENE–WSW trending normal and/or strike slip faults, the latters are sub-parallel to the reactivated Mid-Hungarian

Shear Zone (Fig. 1). These faults are responsible for sudden thickness changes within the Miocene sequences (Tari, 1988, Pentelényi, 2005, Petrik, 2012).

The study area was affected by at least three main tilting events in the Cenozoic. The first one took place in the latest Oligocene-earliest Miocene and resulted in stratigraphic gap and erosional discordance between the Paleogene and Miocene sediments (Báldi and Báldi, 1985; Nagymarosy, 1990; Báldi and Sztanó, 2000). The second one was coeval with the deposition of the youngest 14.5-13.5 Ma volcanoclastics and caused wedge-shaped synsedimentary thickening in tilted blocks. The third one was in early Late Miocene time (Early Pannonian in local Parathetys term) when syn-tectonic wedges of early Late Miocene sediment units can be seen in troughs as opposed to small thickness on elevated highs (see details in the methods chapter on tilting) (Fig. 5A).

3. Methods

During the field work 75 outcrops were examined covering the whole study area. The summary of these observations were presented by Petrik and Fodor (2013) while a thorough analysis is under way. From these sites, 20 are presented here, because they are the most important in interpretation of deformation bands (Fig. 2). We focused on measuring brittle tectonic elements such as joints, faults with or without striae and deformation bands. We calculated stress tensors for striated surfaces using the software of Angelier (1984, 1990). When sufficient data was available we used automatic phase separation often combined with manual separation. During stress tensor calculation we took into account the average misfit angle (α) between measured striae and the calculated shear stress (τ) (ANG criterion) and the misfit between the direction and size of a maximum calculated shear vector and the unit vector along the striae (RUP criterion) (Angelier, 1984; Angelier, 1990). We accepted a fault being part of a stress field when these parameters were below 22.5° and 45%, respectively. Based on Anderson's assumption (Anderson, 1951) we estimated the main stress axes for faults without striae and joints.

Backtilting of measured structures was also applied to reveal whether the deformation might have taken place before, during or after a specific tilting event in order to determine the relative chronology among stress fields. We took into account the average dips of Eocene, Oligocene and Miocene volcanoclastic levels (Fig. 5a). We made two scenarios for tilting of beds because dip data show spatial variations even if measured from the same stratigraphic horizons; one of them supposes a more significant tilting (35° for Eocene, 25° for Oligocene sediments, 15° for lower volcanoclastic level and 7° for Late Miocene sediments). The other represents a more moderate scenario (altogether 25° for Eocene, Fig. 5a). The pre-19 Ma tilting of the Paleogene units resulted in a discordance and erosion of 23-19 Ma, although a gradual, syn-sedimentary Palaeogene tilting also occurred, which was not resolved in our tilt analysis. The syn-sedimentary tilting of volcanoclastic levels can be connected to thickening toward the main basin-margin faults (Fig. 5a). The early Late Miocene tilting also resulted in syn-tectonic thickening of the early Pannonian sediment pile, particularly the formations below the slope unit. The post-sedimentary Pliocene to Quaternary tilting was very modest, and can be estimated from the dip degree of the uppermost imaged Late Miocene horizons: they are fluvial beds supposedly horizontal at the time of deposition and now they dip less than 3°. The recent surface is gently lowering southward, and the dip degree is smaller than 1°. These last tilting events can be connected with the Pliocene-Quaternary uplift of the Bükk Mts. (Dunkl et al., 1994). In our analysis we coupled post-Miocene tilting to early Late Miocene tilting; 7° a complete backtilting led to horizontal bed position of base Pannonian level.

On Fig. 5b we give three different examples for tilt tests. The first tilt test considers small angle (7°) reflecting the earliest Pannonian (and later) tilting while the second one was greater (15-35°) and reconstructed all tilting events. The sign of tilting might be if striae and stress axes are oblique but they are close to the intersection of two sets of conjugate fault planes or of fault and bedding planes. If the intersection line of conjugate Mohr fractures is located on the tilted bedding plane this will also indicate the tilting. The framed stereograms indicate the positive result of tilt test where the symmetry planes of conjugate faults or joints are located in the theoretically vertical position, the intersection line is horizontal. These stereograms were accepted as final result of the stress tensor analysis.

4 sampling points were selected to observe and collect the best developed deformation bands The investigated rocks are porous and partly diagenised late Early Oligocene sandstone and conglomerate, Late Oligocene sandstone and Late Miocene sandstone. Thin sections were made from deformation bands parallel to dip direction, but some sections are subhorizontal and showing features along strike. We applied cathodoluminescence analysis in cases, where relict of multiphase carbonate cement was supposed to be present.

Faults and striated fault planes were compared to deformation bands in order to determine their correlation and to infer kinematics of the bands. The age of deformation was constrained based on the position of estimated stress axes, relative chronology between kinematic indicators and the relationship with tilting events.

Structural analysis was complemented by generalized subsidence model based on 3 boreholes (Eg-1, Dsz-2, Ns-2) located near to the sampling points (Fig. 2). We used a few calibration data (well temperature data, Rock-Eval data, effective porosity and vitrinite reflectance) in order to verify our subsidence models and to shed light on how much the investigated area might have been buried when deformation bands occurred and evolved. The subsidence history models were implemented by Petromod v11 1D software of Schlumberger-IES Company. When making subsidence history we used several parameters elaborated for the Pannonian Basin. Paleo-water depth estimations for Paleogene period were adopted from Báldi and Sztanó (2000) while for the Miocene we took into account the generalized paleo-water depth estimations of the Pannonian Basin (Dövényi and Horváth, 1988), and subaerial conditions of the Early-Mid Miocene volcanoclastic deposits (Lukács and Harangi, 2002; Lukács et al., 2002). Paleo-heat flow data calculated for the central part of the Pannonian Basin were taken from Horváth et al. (2004).

We also performed NW–SE geological cross sections through wells and surface outcrops to depict the main NE to ENE trending structures of the study area (Figs. 7a,b,c). Using average dip values the formation boundaries were projected above the surfacial sampling points to determine how much the burial depth might have been (Figs. 7a,b,c). These data supplement burial modelling, giving a direct geometrical constraint on overburden thickness.

Field structural analysis was supplemented by few 2D seismic reflection profiles together with geological cross sections. The structural interpretation of seismic profiles was compared to stress field evolution and fault pattern known from field work and surface map of Less et al., (2005). Major interpreted faults are displayed on the geological map (Fig. 1).

4. Results

4.1 Subsidence history

The age of fracturing and the style of deformation bands may have connections to burial history. In order to reveal the maximum burial of deformation bands 1D subsidence models were run by integrating 3 borehole data (Eg-1, Dsz-2, Ns-2 on Fig. 1) and a number of calibration data.

Three significant subsidence events can be seen on the generalized subsidence diagram (Fig.6). The first one correlates with the main and fastest subsidence of the NHSSPB (Tari et al. 1993) when the deposition of late Early Oligocene Kiscell Clay took place (Fig. 6). This subsidence continued during the deposition of Eger Formation although with lower rate (~150-200m, 27.5–23 Ma). Corresponding to the syn-rift phase of the Pannonian Basin the second main subsidence is marked by the deposition of Miocene volcanoclastic levels in late Early to Mid-Miocene (~250-300 m thick, 19-13 Ma) and the third one is the deposition of Pannonian sediments during the post-rift phase (~200-250 m thick) in Late Miocene (Fig. 6).

Three main erosion events were considered for the study area based on paleogeographical, sedimentological, geochronological (fission track) data and tectonic evidences (Báldi and Báldi, 1985; Nagymarosy, 1990; Dunkl et al., 1994; Báldi and Sztanó, 2000; Pelikán, 2005).

The first main erosion occurred in latest Oligocene-earliest Miocene, the second one was at the transition of Middle and Late Miocene and the last one could occur from latest Miocene to recent times (Fig. 6). This latter was a regional event and was due to the uplift of the Bükk Mts. to the north (Dunkl et al., 1994).

The NW–SE geological cross sections (Fig. 7a) depict the folded Mesozoic rocks covered discordantly by Paleogene clastics and limestone. The synsediment thickening of Eocene sediments NW from the site 8 can be coupled with north-westward dipping normal/transtensional faults (Fig. 7a). The thickening and northward pinching out of Oligocene siliciclastic rocks can be associated with a Late Oligocene folding and tilting induced by NW–SE compression. This gentle folding could be followed by earliest Miocene denudation of late Oligocene rocks in the southern part of the section. During the deposition of the oldest volcanoclastic level the Kőkötő Fault (KF) and Bogács-Szomolya Fault might have been active, because formation thickness is larger on the northern hanging wall block. During the Middle and Late Miocene these faults were rejuvenated as normal faults or sinistral transtensional faults resulting in syn-sediment thickening and southward tilting of Mid-Late Miocene volcanoclastic rocks and Pannonian sediments, respectively.

The reconstructed possible top of Mid Miocene on Fig.7b indicates that site 5 (Fig. 13) was not buried by more than 200m volcanoclastic rocks before the Late Miocene sedimentation. Site 8 was buried the most among our sampling points indicating 300m volcanoclastics cover in late Mid-Miocene. Cover of Pannonian sediments is possible although it is not easy to control by projected formation base.

The possible thickness of Pannonian sediments during Late Miocene (Fig. 7c) might have been 200-300m at Novaj and 200m at site 5 sampling points. According to the subsidence model and the reconstructed top of Mid and Late Miocene sediments (Figs. 6, 7b, c), the host rock of deformation bands might not have been buried more than 600–800 m and this was reached in the Late Miocene.

4.2 Deformation bands

4.2.1 Szőlőske

This gravel pit is built up of coarse sand, pebbly granule sand, matrix-, and clast-supported conglomerates, which are interpreted as submarine gravity flows – mainly high-density gravely turbidity currents and debris flows – operated during the last part of the early Oligocene (late Kiscellian) (Sztanó and Tari, 1991, Báldi and Sztanó, 2000). Beds dip to S-SE with 20-25°.

Band sets with different orientations could belong to separate phases; the sets are referred to by sample sign S1-S4. All of them are indurated by red ferrigenous cement which made the bands more resistant against weathering (Figs. 8). Two main types of band-like structures were found; one group is gently dipping and wavy (Fig. 8f); the other is sub-vertical or steeply dipping and roughly planar (Figs. 8a,b,f). While the origin of the first set is uncertain, the latter ones are considered as deformation bands. Pebbles dip more steeply than within the sedimentary beds, and varies depending on orientation of deformation band. These deformation bands contain intact and fractured pebbles, as well. Fractures are oriented and confined to rigid pebbles and they cannot be traced in the matrix indicating a relatively early syn-diagenetic deformation (Fig. 8c).

In thin section, deformation bands comprise of oriented elongated intact pebbles, smaller pieces of fractured pebbles, and fine grained quartz matrix cemented by reddish brown iron-oxide material. Although part of the fine grains could derive from fracturing of pebbles, the presence of fine-grained quartz as matrix is sedimentary. Few fractures could partly be inherited from syn-diagenetic pebble fracturing, because host rock also contains damaged pebbles outside the deformation bands. . Damaged pebbles mainly related to deformation band because they are more abundant towards the core zone of bands in conjunction with size reduction (Figs. 8d,e). The fractures of pebbles mostly show wedge shape which is typical for local extension. In many cases these wedge-shaped fractures contain fine grained sediment matrix.

The common point between bands is they are cemented by red ferruginous cement, which was probably a result of mixture of oxidizing meteoric water and ascending reduced acidic Fe-bearing groundwater, as shown in other examples by Chan et al. (2000). Along mixture zones iron oxide-hydroxide were precipitated, when deformation bands were the conduits for fluid flow. All of these indicate that pebble fractures were formed in situ in the sediment (and not carried fractures from pre-Oligocene phases) and were formed prior to iron-oxide cementation, respectively.

Some differences are also present in deformation bands, which can be correlated with their orientation. The main difference is in the ratio of disaggregation and cataclastic zones. In one type of deformation band (sample S4, Fig. 8f) the reorientation (rotation) of pebbles is the dominant mechanism (Fig. 12a), therefore these bands are considered as disaggregation type by the classification of Fossen et al. (2007). In other samples (sample S1, Figs. 8b,c,d, S3, Fig. 8a) zones of cataclasis alsodeveloped. Subtle but noticeable differenceis present within variably cataclastic types. In S1 sample the disaggregation zone is about 3-4 cm, and the cataclastic zone is only few mm (Fig. 8d). From reoriented pebbles toward cataclastic zone, pebble fractures are associated with sediment-filled wedge-shaped fractures. While in the core zone of cataclasis, the pebbles almost entirely crushed into cataclasite (Fig. 8d). To sum up our observations, although narrow cataclasis zone is present (Fig. 8d), the reorientation of pebbles is the prevailing deformation mechanism in the bulk of the deformation band. These characters refer to a transitional type between disaggregation and cataclastic band in the classification of Fossen et al. (2007). The narrow cataclasis may indicate later (mature) progressive step of deformation in which slip surface started forming. The cataclasis is surely prior to cementation, because there is no sign of damaged cement material. Kinematic indicators, like rotation of pebbles and wedge.shaped fracture indicate normal shear sense (Fig. 8d,e).

In S3 sample (Fig. 8a) the cataclasis was more progressed. It is hard to recognise reorientation of pebbles without damage, and grain size reduction is more intense. The zone of cataclasis is cm in scale, thicker than in the S1 type, where it is only mm. This type is a classical cataclastic deformation band (Fossen et al., 2007) and indicates more indurated sediments during their formation.

In chronologic order the S4 disaggregation deformation band (Fig. 12a), which generally dips southeastward represents the earliest phase taken place during early diagenesis of conglomerates. The transitional type (S1) disaggregation deformation band with weak cataclasis represented by NE-SW and NW-SE trending bands (Figs. 8b,d) was the second, which precedes the generation of more intense ENE-WSW trending cataclastic deformation bands S3 (Fig. 8a). This conclusion based on the more effective cataclasis indicates more indurated host rock during burial diagenesis and resulted in greater displacement.

It is important to note that both deformation bands indicate burial depths upon their formation which cannot exceed 1km. In case of disaggregation band, the type of deformation mechanism suggests <1km covering sediments (Fossen et al., 2007), in case of younger cataclastic bands fault related iron oxide-hydroxide precipitation shows oxidizing, shallow depth of burial upon deformation.

4.2.2 Andornaktálya

This sand pit is located between Eger and Andornaktálya and exposes poorly cemented sandstone and intercalated clays (Báldi, 1986; Sztanó and Tari, 1991; Sztanó et al., 1991). The age of the section is Late Oligocene (Egerian) (Báldi, 1986; Hably, 1993). Between the dominantly cross-bedded sandy units clays are interlayered representing barrier and inlet deposits in the upper part of the Eger Formation (Sztanó et al., 1991). Beds dip to S, SE with 20°.

The poorly indurated host sandstone consists of mainly quartz, glauconite and subordinately elongated biotite fragments with very poor limonitic cementation. Measured data and macroscopic observations on this sandstone suggest at least two generations of deformation bands (Fig. 9). Sample A1 was taken from a NW-SW trending single deformation band (Fig. 9d), which occurs as a positive relief in poorly indurated sandstone with cm scale normal displacement. In thin section, the macroscopically single deformation band consists of mm-scale dark coloured bands, in which the grains form aggregates (Fig. 12c). In these bands the grain density is significantly greater, than in the host rock and is associated with slight grain size reduction by friction at contact points of grains. Another indicator of deformation is the mica flakes. They are usually straight and originally deposited parallel to bedding, but their orientation parallel to deformation bands indicate their rotation during deformation. The curving of flakes also indicates deformation that appeared in squeezed position between grains (Fig. 12d). All these features prove that the deformation band was formed by compactional shear mechanism (Aydin et al., 2006), where the grains are packed forming aggregates with weak grain size reduction at contact points. Within the band the cohesive fine-grained material derives from incipient cataclasis taking place at grain contacts. Formation of bands is resulted in porosity reduction. Small amount of limonite precipitation along, but out of band boundaries can relate to fluid flow migration along, but out of reduced porosity zone of the band.

A2 sample was taken from NW-SW trending anastomosing deformation bands which form a 10-15 cm thick deformation zone (Figs. 9a,b),occured in conjugate orientation with respect to A1, but in slightly finer grained sandstone. Along this band the maximum displacement is 15 cm (Fig. 9a). In thin section this is a typical cataclastic band, where fine grained material is the result of cataclasis; but main grains (with size of host rock grains) are also preserved intact within the band (Fig. 12e). The main grains showing crude orientation parallel to band boundary refers to shear. This band is considered as cataclastic shear band and accommodated normal displacement.

The third sample (A4), made from WNW-ESE trending deformation bands (Fig. 9c) which have different orientation relative to A1 and A2 samples. A4 bands form a 10-15 cm thick clustered anastomosing zone, but here a discrete slip surface with two generation of striae also developed. In thin section A4 sample show the most intense cataclasis (Fig. 12f) without preserved main grains. Consequently, the grain size reduction is more significant than in the others. These anastomosing bands of A4 sample are considered as cataclastic shear bands. Along but out of deformation band brownish material, presumably iron-oxide filled the porous space, indicating fluid flow along but outside the band, where band behaved as barrier for lateral fluid flow and gave pathway for along dip fluid migration.

To sum up of our observations, we found three slightly different deformation bands, in which the main difference is the intensity of cataclasis. A1 deformation band represented by aggregate-like structures shows the weakest cataclasis. A2 displays thicker and more intense cataclasis, but here main grains can be present floating among smashed grains. The A4 samples show the most concentrated and intense cataclasis, where main grains are entirely crashed into cataclasite. Although there are some differences between A1 and A2 samples, they are considered to be engaged because they show conjugate geometry and NW-SE strike. A1 sample formed in the same, but coarser grained sandstone as A2. Smaller grain size resulted in more developed induration of host-rock, which favour greater amount of total displacement for A2. In case of A4 sample the cataclasis is the most progressed and the orientation of deformation bands also deviate from A1 and A2 (Fig. 9d stereogram).

4.2.3 Eger, Wind brickyard

This clay pit is located in the eastern part of Eger town exposing the holostratotype of the Oligocene Egerian Stage and the Eger Formation (Báldi, 1973). The exposed sandstone is similar as we described at previous outcrop. Beds dip to S or SE (25°). The whole section is overlain unconformably by Ottnangian (ca. 19.0–17.3 Ma) massive rhyolite tuff which belongs to the lower volcanoclastic level.

At least two types of deformation bands occur in this brickyard, both of them appear in the sandy part of sequence. At first glance the main difference between these two types is in the aspect of cementation and their thicknesses. The poorly indurated and thinner bands are clearly related to NW-SE trending normal faults and were not studied in detail. The other typeis well cemented by calcite (Fig. 10a). These latter well indurated bands compose an anastomosing zone of 5cm maximum width (Fig. 10a) with 15 cm of vertical displacement. Band direction isNNW-SSE. In thin section, we can see that the deformation zone and host rock mainly differ in their appearance (Fig. 12g). Host rock is a layered sandy material with minor calcite matrix, while the deformation zone contains host sandstone grains, which are floating in fine grained calcite material (Fig. 12h). The boundary between band and host rock is not sharp, and partly marked by brownish organic matter present in pore spaces of the bounding host rock along, but out of deformation zone. We interpret this deformation zone as a rare, dilation type deformation band (Du Bernard et al., 2002). Within the deformation band the material and main grain size are the same as that in host sandstone and grains are not fractured. The difference is in the smaller abundance of main grains in contrast of host layers (Fig. 12g). It seems that the larger pore spaces have grown from initial pores by grain displacement, and then these increased pores were almost entirely filled with fine calcite matrix (micrite) (Fig. 12h). Although this band may seem to be a syn-sedimentary dyke (infilled tension crack), but we have 4 reasons why we regarded this structure as rarely described syn-sedimentary dilation band. Firstly (1), in field this band has no aperture, it is a few cm thick calcareous zone, which both up and downwardly become more and more narrow and finally disappears. Secondly (2) there is no sharp boundary between the band and the host rock, which refer to dilation band of Du Bernard et al. (2002). (3) We assumed that the fine grained calcite material filled continuously the increasing void during dilation because the band contains significant amount of host rock grains in equal distribution floating in the matrix. The syn-sedimentary calcareous infiltration into the subvertical band helped preserve this structure. Last but not least (4) there is evidence for ductile deformation of the carbonate matrix. After infiltration of fine matrix into small pores, calcite cement might have started to precipitate. Dilation continued, because fibrous calcite has grown on grain walls in the direction of incremental extension (Fig. 12h). These extensional features associated with no fracturing in matrix indicate that matrix still behaved plastically during deformation; but not enough plastically to fill the dilational voids. Therefore the deformation of matrix occurred prior to lithification of the band.

Our interpretation suggests that this deformation zone can be considered as dilation deformation band evolved as dilational fault segment (Ferrill and Morris, 2003) in very poorly consolidated sand. The dilational fault segment occurs, when failure cut mechanically different layers. In this case the failure angle in the mechanically stronger layer (sand) is smaller than weaker layer (clay). During faulting steep dilation segments have been produced in sandy part, which were filled by sediments. The infilling fine grained calcite matrix is present within the Egerian host sandstone. Moreover this sedimentary unit covered by thick terrestrial volcanoclastic levels therefore band infillings can only derive from the same host sedimentary unit.. Consequently this type of deformation band formed before the end of deposition of host sedimentary unit, which is a very important result for an absolute deformation chronology.

4.2.4 Novaj, sand pit

This sandpit is made up by poorly cemented Lower Pannonian sandstone alternating with siltstone deposited between 11.6-9 Ma. The layers dip to the south with 10-15°. Numerous deformation bands with cm scale offsets form single or locally anastomosing features at this location. Occasionally, in the vicinity of deformation bands water-escape (pipe and bowl) structures can be seen, their development can be related to temporal and local overpressure condition in soft sediment (Fig. 11).

In thin section (Fig. 12b) host sandstone contains mainly sub-angular quartz, feldspar grains and rhyolitic volcanic clasts. The deformation band (N1) can be divided into a main finer grained and a bordering ocher-coloured cement-filled zone (Fig. 12b). Within the bands, grain size reduction is present due to cataclasis resulting in porosity reduction. The few elongated grains show orientation parallel with band boundaries and deviate from sedimentary beddings; this geometry indicates shearing. These southward dipping deformation bands are cataclastic bands, which are often developed into minor faults. The presence of limonite indicates that deformation band, as reduced grain sized zone with reduced porosity, behaves like a barrier on fluid flowsalong which Fe content of fluids were precipitated as Fe-hydroxide cement in pores just adjacent to bands. Presence of limonite in cement also refers to near-surface initiation of fault.

4.3. Stress field evolution

8 stress fields were determined in the studied area from Early Paleogene to Late Miocene by means of field structural observations, deformation band analysis and interpreted seismic profiles (Fig. 13). **The first stress field (D1)** is predominantly NE–SW compression but less typically NW-SE extension also occurred (Fig. 13). This phase is mainly characterized by NW–SE trending conjugate reverse faults in association with folds. However, the synsediment thickening of Late Eocene clastic rocks and limestone along NE–SW trending normal/transtensional faults can be tied to this D1 phase (Fig. 7a). Joints and normal faults in sites 8 and 10 were also associated with this deformation phase (Fig. 13). NE–SW trending disaggregation bands in site 8 (sample S4) presumably also belong to this phase (Fig. 6).

Late Eocene limestone is the oldest rock in which conjugate reverse faults of D1 phase were identified. The youngest rock affected by this compression was the lower part of Eger Formation (early Late Oligocene, site 6 on Fig. 13). We have a clear upper time limit for the deformation: tilt test proves that all structures formed at nearly sub-horizontal layer position, before the latest Oligocene–earliest Miocene post-sedimentary tilt event. The onset of the phase is uncertain but faults in Late Eocene may suggest that this deformation occurred already before Oligocene syn-sedimentary tilting event.

The second stress field (D2) was divided in two sub-phases or events; they are similar in the direction of the main stress axes but different in style. The **D2a** is characterized by NE–SW extension (in sites 5, 6, 8, 9 Fig. 13). The extensional deformation is expressed as normal faults predating the first major tilt event (Fig. 13). The A1 aggregates-like band and A2 cataclastic band (in site 5) belong to this phase. The W5-2 dilational deformation band (site 6) indicates early practically synsedimentary deformation prior to the end of deposition of the late Oligocene sandstone. For event 2a, fractured pebbles at site 8 prove early syn-diagenetic deformation because fractures within rigid pebbles (site 8) cannot be traced in the matrix and predate any tilting event.

The **D2b** is basically a NW–SE compression with NE–SW trending conjugate reverse faults and oblique dextral strike slip faults (in sites 9, 10, 11, Fig. 13) which accommodated small scale offsets. This phase might be associated with latest Oligocene-earliest Miocene tilting because site 9 indicates pre-tilt deformation as opposed to sites 10, 11 which clearly show post-tilt deformation.

In the northwestern part of the seismic profile (Figs. 14a,b) Paleogene sediments are folded and dissected by Mid to Late Miocene normal faults. The toplaps of Late Oligocene sediments indicate an important discordance between these sediments and Early Miocene volcanites (Figs. 14a,b); this erosion phase can be connected with latest Oligocene-earliest Miocene tilting event (Fig. 5a). North-westward the Paleogene sediments are thinning and pinching out on the flank of an elevated high which indicates syn-sedimentary tilting to the SE during the Oligocene (Figs. 7a, 14a,b). This paleohigh can be correlated with Kis-Eged Hill along the supposed strike to NE. (Figs. 2, 7a, 14ab). This thickening can be interpreted as part of a fold. We postulate that folds are connected to reverse faults which remained blind below the Paleogene sediments; they are schematically figured on the seismic line (Figs. 14a,b). Less typically SE vergent reverse faults can also be suspected on seismic profiles in the vicinity of Verps-3 borehole causing small offset in Paleogene sediments (Figs. 14a,b).

The syn-depositional thickening of Kiscell and Eger Formations within folds and in front of propagating blind reverse faults also suggests Oligocene time for this deformation phase (Figs. 14a,b). The reduced or completely missing Oligocene sediments on elevated highs (Fig. 7a), toplaps in Eger Formation may suggest that the D2 phase was still active after Paleogene sedimentation, during the earliest Miocene erosion. Because all sites of 2a event pre-date any tilting, it may predate, at last partly the event 2b, which continued after the tilt of Oligocene units. However, a precise timing of events could not be determined.

The **D3 stress field** was developed after latest Oligocene-earliest Miocene tilting because it was observed on already tilted Paleogene layers (Fig. 13, sites 5, 7, 10, 11, 18, 20) and also on Early Miocene volcanites (site 4). The D3 is characterized by E–W compression and perpendicular extension marked by E–W trending normal faults, NE-SW and NW–SE trending dextral and sinistral strike-slips, respectively (Fig. 13). We observed conjugate fractured pebbles in late early Oligocene (late Kiscellian) conglomerate (site 7 in Fig. 13).

Deformation band with intense cataclasis (A4) evolved during this phase (site 5). In addition to deformation bands, fault slip surface coated with 1cm clay smear formed which bears several sets of striae: the dip slip striae were formed in D3 phase, but the normal fault surface was reactivated in two other stress fields; their younger relative ages are suggested by tilt test. On geological map a larger ENE–WSW trending normal fault (Kőkötő Fault, KF in Fig. 2) can be seen close to site 5. The thickness of these lower volcanoclastic level increases to the north, toward the hanging wall - when crossing the Kőkötő Fault – from a few meters to several hundred meters. The southward syn-sediment thickening of this unit can also be observed along NE–SW oriented normal faults on Fig. 14a,b. A stress field similar to D3 was recorded in Early Miocene sediments west from our research area, along the Darnó Zone (Fodor et al., 2005b).

On the basis of regional observations this stress field is present in the lower rhyolitic volcanoclast level (site 4) but it cannot be observed in the middle dacitic volcanoclast level indicating, together with the tilt test, a latest Early Miocene age for the phase (early Ottnangian, ca. 19-18 Ma). It is also corroborated by the presence of fractured pebbles, which indicate an “early” phase of deformation, prior to maximal burial reached in the late Miocene. This stress field is similar to the earliest syn-rift extensional event of the Pannonian Basin (Fodor et al., 1999).

According to paleomagnetic measurements on volcanoclastic rocks, there were two counterclockwise rotations during volcanism (Márton, 1990; Márton and Márton, 1996). The first took place between ca. 18.5-17.5 Ma (after the deposition of the lower volcanoclastic level) and caused 40-50^o CCW rotation; this deformation occurred after the D3 phase. The stress field evolution can be correlated with paleomagnetic data (Márton and Fodor, 1995). The change in the principal stress axes between the D3 and D4 phases is similar in magnitude than the vertical axis rotations, and opposite in sense. Thus, the change is stress field is apparent and connected to rotations.

The **D4 stress field** shows NE–SW extension with NW–SE trending conjugate faults and joints (Fig. 13). This stress field was observable in sites 4, 5, 15-18, 20 (Fig. 13). In site 5 NE–SW extension (D4 phase) is indicated by conjugate and only slightly tilted NW–SE trending joints and deformation bands (A4). In the same site some formerly created faults and deformation bands (formed in D3 phase) were reactivated by NE-SW extension of the D4 phase. Relative age of this deformation event is proved by tilt test which shows symmetrical fractures in present-day position. In site 4 the dominant structures were NW–SE trending conjugate normal faults. Tilt tests proved that part of the normal faults were developed in sub-horizontal position because faults were more symmetrical when restoring the total dip angle (15°). The other fault set shows NE–SW extension with conjugate normal faults which rejuvenated and newly striated in post-tilt (present day) position (see phase 8 in Fig. 13).

On NE-SW trending seismic profiles some syn-sedimentary normal faults can be seen suggesting NE–SW extension which resulted in the thickening of the volcanic units. The age of this stress field is Karpatian–Early Badenian (17.5–14.5 Ma). This phase represents one of the most typical extensional deformations, the major syn-rift phase of the Pannonian Basin (Fodor et al., 1999).

The second blockrotation was 25-30° in CCW direction and took place between D4 and D5 deformation phases in the time span of 16-14.5Ma (Márton, 1990; Márton and Márton, 1996). The change in the principal stress axes between D4 and D5 is the same but in opposite sense as indicated by paleomagnetic measurements. Therefore the change in stress field is apparent. The **D5 stress field** of WNW–ESE extension induced the formation of prominent, map-scale faults with ENE–WSW strike (Fig. 13). On seismic profiles and geological cross sections (Figs. 7, 14a,b) most of the NNE–SSW trending normal faults dip to the WNW were responsible for southward thickening of Mid Miocene volcanoclastic levels. The development of these transtensional faults was partly coeval with the second major synsedimentary tilting in Mid-Miocene.These normal faults were later rejuvenated because they even offset the Late Miocene sediments (Figs. 14a,b). On the other hand, the ENE-WSW trending faults have sinistral strike slip kinematics and associated with outcrop-scale NNW–SSE trending dextral faults (Fig. 13 sites 12-15, 17).

Transitional type conjugate deformation bands (S1 type samples) in Szőlőske (site 8) indicates N-S compression and perpendicular extension and we attributed these structures to the D5 phase. Shear could be responsible for the reorientation and weak fracture of pebbles along these deformation bands.

In Late Triassic limestone (site 11) this stress field created NNE–SSE trending joints and fault zones which promoted flow of meteoric water and formation of karstic holes during a much younger (Quaternary?) dissolution event.

A major NNE–SSW trending normal fault bounds the Felsőtárkány Basin from the east (Fig. 2) which resulted in the downfaulting of all Mid Miocene volcanic and clastic formations.

The normal and left later strike-slip faults of this phase were responsible for the initial opening of the Vatta-Maklár Trough. Increasing thickness of the upper volcanoclastic level (14.5-13.5 Ma) and intercalated sediments toward basin-margin faults give a time constraint on this deformational phase. The age of this deformation phase is late Badenian–Sarmatian (Petrik and Fodor, 2013) which corresponds to the second rifting phase of the Pannonian Basin (Fodor et al., 1999).

The **D6 stress field** indicates NE–SW to ENE–WSW compression (Fig. 13). N-S trending dextral and E-W trending sinistral strike-slip zones characterize this stress field together with NW–SE trending conjugate reverse faults (Fig. 13 sites 2, 6, 12-14, 19). The N-S trending right lateral strike slip zone in site 2 (upper volcanoclastic level) is 0.5-1m wide and exhibits numerous Riedel shears of NNE–SSW directions with dextral striae (Fig. 13, site 2). The secondary faults probably rotated within the shear zone; this is the reason why the two events can be separated by computer stress calculations (Fig. 13 site 2), although this separation could be artificial and reflect only progressive rotations of small faults.

In site 6, this phase incorporates reactivated older fracture planes which determine relative chronology between phases. D6 reverse slip offsets the former normal faults of phase D2a (Fig. 13 site 6, Fig. 10b). The reverse displacement was accommodated by folding in the less competent clay. The fold axes dip to SSE perpendicular to the main compressional axis. This compressional phase might be associated with late Sarmatian to earliest Pannonian inversion of the Pannonian Basin creating small scale folds and erosion of Mid Miocene sediments (Horváth, 1995, Fodor et al., 1999, Csontos et al., 2002).

The **D7 stress field** is a NNW–SSE extension in which the most significant structures are ENE–WSW trending normal faults (Fig. 13). This stress field characterises the Early Pannonian tilting to the south-southeast. This phase can be observed in numerous outcrops of southern Bükk foreland (Fig. 13 sites 1, 3, 5, 8-14, 16, 19).

More progressed cataclastic deformation bands in site 8 (S3 sample) with ENE–WSW trend indicates NNW–SSE extension which resulted in several meters of normal offset. These deformation bands can be tied to D7 deformation phase (Fig. 13). The general ferrigenuous cementation of all types of deformation bands in site 8 can also be related to fault activity of D7 phase. Cataclastic bands (N1) in site 1 also indicates N-S extension (D7 phase) together with other normal faults with E-W trend (Fig. 13).

D7 phase gives prominent structures resulting in the duplication in map view of Oligocene to Miocene sequences along NNW dipping normal faults (Less et al. 2005) namely the Szomolya-Bogács and the Kőkötő faults (Figs. 2,7). The development of these normal faults was associated with the third major tilting event resulting in the synsediment thickening of Late Miocene sediments (Figs. 1, 14a, b. c, 14a,b). Important subsidence occurred during this deformation phase in the main part of the Vatta-Maklár Trough where more than 700 m thick Lower Pannonian sediments were deposited before the arrival of the shelf slope prograding from the north. Using magnetostratigraphical age calibration of the southerly located Tiszapalkonya-1 well (Elston et al., 1984) the deformed, pre-slope lacustrine sediment package can be bracketed between 11.6-9.2 Ma (Petrik and Fodor, 2013). This D7 phase can be tied to the post-rift deformation of the Pannonian Basin and resulted in NE–SW trending half-grabens filled with Late Miocene sediments of the Lake Pannonian. This subsidence is related to the thermal contraction of the mantle lithosphere (Royden and Horváth, 1988).

The **D8 stress field** is characterized by NW–SE compression and perpendicular extension with WNW-ESE trending dextral and NE–SW trending reverse faults (Fig. 13 sites 1-3, 12-13, 20). The NW-SE compression affected the upper volcanoclastic level where conjugate ENE-WSW trending reverse faults produced 10-15 cm offsets (Fig. 6 site 3). This stress field could be responsible for regional tilting of the Late Pannonian sediment in the whole Vatta-Maklár Trough. In our interpretation this stress field must be older than the recent stress field which is characterized by NE–SW compression (Bada et al., 2007).

5. Discussion

5.1 Main deformation phases and their interpretation

8 stress fields were separated by using fault slip and deformation band analysis (Fig. 13). Deformation bands contributed to specify more precisely the time of deformation and fault evolution. Most of the stress fields can be fitted in the Cenozoic deformational phases of the Pannonian Basin (Csontos, 1995, Fodor et al., 1999). However, stress field **D1** of NE–SW compression and perpendicular extension is a new phase although its interpretation is problematic (Fig. 13). The age of this stress field is rather controversial: it seems to be pre-tilt in different formations, up to the early Late Oligocene, which would mean an age of late Eocene to early Late Oligocene. On the other hand, the phase D2 may have similar syn-sedimentary character during the Paleogene and earliest Miocene but the compression would be perpendicular to D1 compression. The upper age of D1 is also obscure but the lower part of the Egerian rocks was affected by this stress field. Csontos (2000) also identified this NE–SW compression based on microfolds in the late Early Oligocene Tard Clay Fm. (probably the same as our site 9). He presumed Eocene-Oligocene age of this stress field (Csontos, 2000). The integration of this deformation phase into the Paleogene history is controversial because in other parts of the North Hungarian Paleogene Basin this stress field is missing and the deformation is identic to our D2 phase.

The other stress state which can have a regional relevance is the NE–SW extension (**D2a**) before latest Oligocene-earliest Miocene tilting (Fig. 13). The age of this deformational phase is well-constrained by dilational bands which indicate syn-sedimentary deformation of Egerian (late Oligocene) age. Many normal faults and conjugate deformation bands show the NE–SW extension before the latest Oligocene-earliest Miocene tilting. However, we assume that the NW–SE compression (**D2b**) was also present at the same time. Our explanation for the coexistence of both stress states is that the extensional structures accommodated a noticeably elongation along the axis of the compressional folds. The other alternative explanation is that the stress field was actually transpressional one with E–W trending dextral faults and NE–SW trending reverse faults. The reverse fault-bounded sub-basins were connected by dextral tear faults. However, in the extensional bending of strike-slip faults NE–SW extension might have been the dominant deformation. Such scenario is known 200 km west from the study area (Fodor et al., 1992a, 1999).

The D2b NW–SE compression is the same stress field identified in other parts of North Hungary for Paleogene. The dominant structures are NE–SW trending folds, reverse faults and E–W trending dextral faults (Fodor et al., 1992; Fodor et al., 1999), which were connected to transpressional asymmetric basins. Tari et al. (1993) presumed that the Paleogene basin was developed as a retroarc flexural basin by southward propagating backthurst along the northern margin of the NHSSPB. However, Palotai and Csontos (2010) identified NW vergent Late Oligocene reverse faults ca.75km to the SW, along strike of the basin. Their interpretations are in good agreement to ours; because we also identified similar Oligocene reverse faults in the northern proximity of the Mid-Hungarian Shear Zone (MHSZ) (Figs. 1, 14a,b). These observations revealed that the Paleogene Basin was bordered on the south by northwest-vergent reverse faults during the Late Oligocene.

The D3 phase is dominated by N–S extension. This phase precedes the first CCW blockrotation in northern Hungary and already affects the lower volcaniclastic level. D3 probably marks the earliest syn-rift deformation of the Pannonian Basin (Fodor 2010). The synsediment thickening of lower volcaniclastic level toward major normal faults indicates syn-volcanic late Eggenburgian-Ottnangian deformation (21-18.5 Ma). This early extension was just recently proposed (Fodor, 2010) and our study provide the first map-scale structures which belongs to this early extensional deformation.

The D4 is dominated by NE–SW extension. In the study area the Darnó Fault acted as a left lateral strike-slip during this time span (Fodor et al., 2005b). NW–SE trending half grabens and pull-apart basins are widespread in the whole Pannonian Basin (Fig. 1) (Royden and Horváth, 1988; Csontos 1995; Fodor et al., 1999). This phase belongs to the “classical” syn-rift deformation (Royden and Horváth, 1988). The extension is triggered by the roll-back mechanism along outher Carpathian subduction zone (Royden and Horváth, 1988). D4 already affects the middle volcanoclastic level follows the first CCW rotation (Márton and Fodor, 1995) indicating Karpatian-early Badenian deformation (17.5-14.5 Ma) (Fig. 13).

The D5 is a transtensional phase and is dominated by WNW–ESE extension and perpendicular compression (Fig. 13). In the study area this phase caused the development of Felsőtárkány subbasin and the Vatta-Maklár Trough (Figs. 2, 4, 7). The ENE–WSW trending faults were acted as transtensional left lateral strike slip zones resulting in the synsediment deposition of upper volcaniclastic levels (Fig. 7). According to Tari (1988) the Vatta-Maklár Trough evolved as a transtensional half-garben during Mid-Miocene. This deformation was responsible for synsediment deposition of late Badenian and Sarmatian suites in many subbasins in the Pannonian Basin (Fodor et al., 1999); the basin system of the study area also represents this prominent deformation. This phase is regarded as the second major syn-rift deformation after the Mid-Miocene blockrotation (Fodor 2010). The change in extensional direction corresponds to the southeastward propagation of the subduction front (Fodor et al. 1999) and related roll-back mechanism along Eastern Carpathian (Royden 1993).

The D6 NW–SE extension and perpendicular compression is often mentioned as late Sarmatian inversion (Horváth 1995). This phase induced inversion of earlier basins and might have caused the erosion of Sarmatian sediments (Horváth 1995). In the study area this phase seems to be divided into a rather transpressional and a clearly extensional one (Fig. 13). In the latter case the NNE–SSW trending oblique normal faults reactivated the earlier normal faults of D5 phase (Fig. 13). The change in extension is not significant but the oblique reactivation clearly shows a younger deformation (Fig. 13). The D6 phase also resulted in some small scale folds and reverse faults (site 6 in Fig. 13) which might correspond to the postulated inversion. In our interpretation, the D6 is a strike-slip deformation wich might be responsible for varying deformational patterns.The **D7** phase is also a very significant and new observation in southern Bükk foreland which is characterized by NW–SE extension (Fig. 13). This stress field was coeval with the earliest Late Miocene tilting and caused syn-sediment thickening of the early part of the Late Miocene sediments along the Vatta-Maklár Trough (Petrik, 2012). In other parts of the Carpathian Basin this stress field is also present and creates many transtensional sub-basins and fault scarps (Fodor et al., 1999; Fodor et al., 2005a; Palotai and Csontos, 2010).

The D8 nearly NE–SW extension and perpendicular compression is post-tilt deformation. It appears also in Late Miocene sediments. This phase has not been revealed in the Pannonian Basin thus regional map scale structures can not be attributed to it. The contemporaneous deformation of the Pannonian Basin is NE–SW compression proven by stress field data from boreholes (Bada et al. 2007). Thus D8 should be older in age and can be acted during latest Miocene (Fig. 13).

5.2 Connection of deformation bands, burial history and displacement

In this chapter we discuss the possible connection between the type of deformation bands, their deformation mechanism, and the presumed burial depth of the sample at the time of deformation.

A number of possible factors influence which type of deformation bands form. The most important factors are the physical properties of rock bodies (mineralogy, texture, grain size, roundness etc.) which more or less constant for a given sedimentary rock layer (Fossen, 2010). However, burial depth, confining pressure, tectonic environment can vary from time to time resulting in the formation of different type of deformation bands. This must be a reason why we cannot find certain type of deformation bands at a given burial depth. In case of disaggregation type of bands, we suggest that they can be tied to early deformation taken place prior to or during early diagenesis. In our case, the diagenesis of host rock strongly correlates with burial depth. When the burial depth and differential stress had been greater, cataclastic type of deformation bands have been formed.To demonstrate the evolution of bands we have chosen specific sampling points, where either more generation of bands or specific bands are present from chronological point of view. In another line, we consider also the eventual connection between deformation band type and final cumulative displacement. All these aspects are shown on figure 15.

In the Wind Brickyard the W5-2 syn-sedimentary dilation band indicates the bands type formed under very low lithostatic pressure. In this case, time interval between deposition of sediment and band formation (D2 phase) is the narrowest. In S4 disaggregation band without grain destructive deformation mechanism could already indicate shallow burial (max. ca. 100-200m from subsidence analysis, Figs. 6, 15b), but even narrow time span is between age of deposition (Ma) and deformation (D1 phase). The weakest grain destructive bands are from A1 sample, which is a single band. These compactional shear bands without significant grain size reduction indicate that the depth of the sediment did not exceed 200m following the subsidence analysis (Figs. 6, 15b) during their formation. The conjugate A2 sample indicates more compacted sediments either due to their finer grained size and subsequent lower porosity or a bit later formation of these bands upon burial. A4 bands show more progressed cataclasis than previous ones manifested by higher degree of grain–size reduction. These anastomosing bands are reactivated during the younger D4 phase indicated by striae on the surface of deformation bands. This trend in progression of cataclasis correspond to wider time span between the age of host rock and of the deformation, which means more indurated rock due to progression of burial diagenesis (depths exceed 200-250m).

In case of Szőlőske S1 sample similar evolution trend can be observable, but in conglomerates. S1 type band is transitional between diasaggregation and a cataclastic band. The development of a narrow cataclastic zone indicates more indurated host rock than S4 diasaggregation type, which is in connection with larger, cc. 300-400m burial depths at the time of its formation during the Mid-Miocene. S3 band shows even more advanced cataclasis taking place in thicker zone. This most destructive band suggests the largest burial depth; relating to D7 phase, it could be 600-800 m. Here band reactivation by faulting might have played a similar role as in case of A4 band.

These trends are also supported by modified Cam cap model for the formation of deformation bands (Schultz and Siddharthan, 2005). The q-p diagram (Fig. 15c) shows that dilation band (1) and dilation with shear band (2) can be formed at smaller values of confining pressure (p), where volume increase dominates the deformation. Shear band (3), which is corresponding with disaggregation bands by mechanism of deformation (Fossen, 2007) means transition stage with no volume change toward compactional types (4-5), where volume decrease is associated with deformation. 15d figure shows possible loading pathways to demonstrate which type of deformation band is favoured at variable q-p conditions. At lower confining pressure, shear bands with dilatancy are formed (Fig. 15d); in our case this correspond to W5-2 band. Disaggregation band S4, which corresponds with shear band with no volume change can be formed at relatively medium values of confining pressure. At higher lithostatic pressure compactional shear (volume decrease) is the prevailing – in our cases A1, N1, A2, S1, S3, A4 belong here in order to increasing cataclasis. Pathways for reactivated (faulted) bands could hardly be reconstructed but are schematically approached by B-C lines on the q-p diagram (Fig. 15d).

Based on these trends, the following conclusions can be drawn. The earlier the deformation band in the deformation history, the less destructive band type evolved (Fig. 15). The more indurated rocks show more cataclastic deformation. The more cataclastic rocks are more capable to evolve into a discrete slip surface. In close association with the A4 cataclastic deformation band even a clay smear and multiple slip evolved along the discrete fault surface. The most evolved cataclastic deformation bands (S3) is the youngest and probably indicates the most significant burial depth. The strongly cataclastic deformation mechanism itself could suggest 1-3 km overburden, although less intense cataclastic deformation can occur in shallower burial depth (less than 1 km) where poorly consolidated, well-rounded and good sorted grains are observable (Cashman and Cashman, 2000; Fossen et al., 2007). The subsidence modelling predicts 600-800m of cover at the presumed time of deformation. In summary, going from D1 to D7 deformation phase the burial depth is increasing which resulted in hardening of sediments in line with gradually changes in deformation mechanism of bands toward more intense cataclasis (Fig. 15a).

Concerning the displacement along the bands, we summarise our observation as follows. The less cataclastic type of deformation bands (S4, W2, A1, A2) are associated with less displacement (Fig. 15). Within the cataclastic type bands, the more developed the cataclasis, the greater total displacement can be observed, presumably due to the fact that bands evolved in a fault (Fig. 15). Concerning site 8 the youngest type of cataclastic deformation bands (S3) fits to this trend and indicates the greatest cumulative displacement.. However, in case of the S4 and S1 transitional type of deformation bands, the displacement is not increasing in time which can be explained by different deformation mechanism (Fossen 2007) rather than the result of increasing cumulative displacement. In conclusion, although the amount of displacement could influenced by the deformation mechanism, namely greater displacement can accumulate along disaggregation band than single cataclastic band (Fossen 2007), we suggest that within cataclastic type of bands the burial depth (and associated degree of diagenesis) was the decisive factor. This is in agreeement with our subsidence modelling and independent classification of bands into deformation phases and also can be reconciled with physical models for band formation.

6. Conclusion

Combination of fault slip data and deformation band analysis permit to separate 8 deformation phases (D1-D8), the most of which fit into Cenozoic evolution of the Pannonian Basin. The combined method is more powerful than any separated analysis, while they mutually cross-check the time of formation, kinematics of certain brittle elements.

2 new stress fields were observed (D1, D2a) for the Paleogene time but their integration into the structural evolution of the Pannonian Basin is not always unequivocal. The D1 is characterized by NE–SW compression and probably related to an early Paleogene deformation. The D2a indicates NE–SW syn-sedimentary extension of Late Oligocene age by dilational bands. The third new stress field (D7) of late Miocene age is NW–SE extension and

caused syn-sediment fault activity and the second main tilting of the study area in earliest Late Miocene. This D7 phase resulted in the formation of 700m thick Pannonian wedge in the NE–SW trending half grabens of the Vatta-Maklár Trough during 11-9 Ma just in the southern vicinity of the study area.

In addition to understand the local structural evolution, the deformation band and fault slip analysis combined with subsidence history allowed us to draw more general conclusion concerning the evolution of brittle deformation in poorly consolidated sediments. The less destructive type of deformation bands are the oldest ones, were formed at shallowest burial. The more intense cataclasis correlates with the deeper burial depth. With progressive burial, deformation bands evolved into discrete fault slip surface. This suggested trend can be crosschecked in other areas of brittle deformation of porous sediments.

7. Acknowledgements

Our investigation was supported by the Hungarian National Research Found (OTKA No. 83150). We are grateful to Hungarian Horizon Ltd. and MOL Ltd. for providing us with seismic profiles. Digital geological map of the wider study area was kindly provided for PhD studies of A. Petrik by the Geological and Geophysical Insitute of Hungary. We thank the support of and discussion with Pál Pelikán, András Nagymarosy, Orsolya Sztanó, and Kinga Hips. Comments of two anonymous reviewers greatly improved the earlier version of this paper.

References

Anderson, E.M., 1951. The dynamics of faulting and dyke formation with application to Britain. Oliver & Boyd, Edinborough, 2nd edition.

Angelier, J., 1984. Tectonic analysis of fault slip data sets. J. Geophysical Research 89, B7, 5835–5848.

Angelier, J., 1990. Inversion of field data in fault tectonics to obtain the regional stress. III. A new rapid direct inversion method by analytical means. Geophys. J. International, 103, 363–373.

Antonellini, M., Aydin, A., 1994. Effect of faulting on fluid flow in porous sandstones: petrophysical properties. AAPG Bulletin, 78, 355–377.

Aydin, A., Johnson, A.M., 1978. Development of faults as zones of deformation bands and as slip surfaces in sandstones. Pure and Applied Geophysics, 116, 931–942.

Aydin, A., Johnson, A.M., 1983. Analysis of faulting in porous sandstones. Journal of Structural Geology, 5, 19–31.

Aydin, A., Borja, R.I., Eichhubl, P., 2006. Geological and mathematical framework for failure modes in granular rock. Journal of Structural Geology, 28, 83–98.

Bada, G., Grenerczy, Gy., Tóth, L., Horváth, F., Stein, S., Cloething, S., Windhoffer, G., Fodor, L., Pinter, N., Fejes, I., 2007. Motion of Adria and ongoing inversion of the Pannonian Basin: Seismicity, GPS velocities and stress transfer. Geol. Society of America Spec. Paper, 425, doi: 10. 1130/2007.2423(16).

Báldi, T., 1973. Mollusc fauna of the Hungarian Upper Oligocene (Egerian). Studies in stratigraphy, palaeoecology, palaeogeography and systematics. Akadémia Press, Budapest.

Báldi, T., 1986. Mid-Tertiary stratigraphy and paleogeographic evolution of Hungary. Akadémia Press, Budapest, 201.

Báldi, T., Báldi-Beke, M., 1985. The evolution of the Hungarian Paleogene Basins. Acta Geol. Hungarica 28, 1–2, 5–28.

Báldi, T., Sztanó, O., 2000. Gravity mass movements and paleobathymetric changes in the marine Oligocene deposits of the Bükk Mts. Bull. of the Hung. Geol. Society, 130, 3, 451–496. (in Hungarian).

Cashman, S., Cashman, K., 2000. Cataclasis and deformation-band formation in unconsolidated marine terrace sand, Humboldt County, California. Geology, 28, 111-114.

Chan, M.A., Parry, W.T., Bowman, J.R., 2000. Diagenetic hematite and manganese oxides and fault-related fluid flow in Jurassic sandstones, southeastern Utah. AAPGBulletin, 84, 1281–1310.

Csontos, L., 1995. Tertiary tectonic evolution of the Intra- Carpathian area: a review. Acta Vulcanologica, 7, 1–13.

Csontos, L., 2000. Structural outline of the Bükk Mts (N Hungary). Bull. of the Hung. Geol. Society, 129, 4, 611–651 (in Hungarian). Csontos, L., Nagymarosy A., 1998. The Mid-Hungarian line: a zone of repeated tectonic invesrions. Tectonophyiscs, 297, 51–71.

Csontos, L., Nagymarosy, A., Horváth, F., Kovác, M. 1992. Cenozoic evolution of the Intra-Carpathian area: a model. Tectonophysics 208, pp. 221-241.

Csontos, L., Márton, E., Wórum, G., Benkovics, L., 2002. Geodynamics of SW-Pannonian inselbergs (Mecsek and Villány Mts, SW Hungary): Inferences from complex structural analysis. In: Cloetingh, S. (Ed.), Stephan Mueller Special Publ. Series, 3, pp. 227-245.

Dövényi, P., Horváth, F., 1988. A review of temperature thermal conductivity and heat flow data from the Pannonian Basin. In: Royden, L., Horváth, F. (Eds.), The Pannonian Basin. Ass. Amer. Petr. Geol. Mem., 45, pp. 195-233.

Du Bernard, X., Eichhubl, P., Aydin, A., 2002. Dilation bands: A new form of localized failure in granular media. Geophysical Research Letters 29, 24, 2176.

Dunkl, I., Árkai, P., Balogh, K., Csontos, L., Nagy, G., 1994. Thermal modelling based on apatite fission track dating: the uplift history of the Bükk Mts. Bull. of the Hung. Geol. Society, 124, 1, 1–24. (in Hungarian).

Elston, D.P., Lantos, M., Hámor, T., 1984. High resolution polarity records and the stratigraphic and magnetostratigraphic correlation of Late Miocene and Pliocene (Pannonian, s.l.) deposits of Hungary. In: Teleki, P.G., Mattick, R.E., Kókai, J. (Eds.), A case study from the Békés basin, Hungary. Kluwer Academic Publishers, Basin Analysis in Petroleum Exploration, pp. 111–140.

Ferrill, A.D., Morris, A.P., 2003. Dilational normal faults. Journal of Structural Geology, 25, 183-196.

Fodor, L., Magyari, Á., Kázmér, M., Fogarasi, A. 1992. Gravity-flow dominated sedimentation on the Buda paleoslope (Hungary): Record of Late Eocene continental escape of the Bakony Unit. Geologische Rundschau 81, 3, 695–716.

Fodor, L., Csontos, L., Bada, G., Györfi, I., Benkovics, L., 1999. Tertiary tectonic evolution of the Pannonian Basin system and neighbouring orogens: a new synthesis of paleostress data. In: Durand, B., Jolivet, L., Horváth, F., Séranne, M. (Eds.), The Mediterranean Basins: Tertiary extension within Alpine Orogen. Geological Society, London, Special Publications 156, pp. 295–334.

Fodor, L., Bada, G., Csillag, G., Horváth, E., Ruszkiczay-Rüdiger, Zs., Palotás, K., Síkhegyi, F., Tímár, G., Cloetingh, S., Horváth, F., 2005a. An outline of neotectonics structures and morphotectonics of the western and central Pannonian basin. Tectonophysics, 410, 15–41.

Fodor, L., Radócz, Gy., Sztanó, O., Koroknai, B., Csontos, L., Harangi, Sz., 2005b. Post-conference excursion: tectonics, sedimentation and magmatism along the Darnó Zone. GeoLines 19, 142–162.

Fodor, L.I., Sztanó, O., Magyar, I., Törő, B., Uhrin, A., Várkonyi, A., Csillag, G., Kövér, Sz., Lantos, Z., Tőkés, L., 2013. Late Miocene depositional units and syn-sedimentary deformation in the western Pannonian basin, Hungary. In: Schuster R. (Eds.), 11th Workshop on Alpine Geological Studies & 7th European Symposium on Fossil Algae. Abstracts & Field Guides. Schladming, 99, pp. 33–34.

Fossen, H., 2010. Fracture and brittle deformation. In: Fossen H. (Eds.), Structural Geology, Cambridge Univ. Press, New York, pp. 119–148.

Fossen, H., Bale, A., 2007. Deformation bands and their influence on fluid flow. AAPG Bulletin, 91, 1685–1700.

Fossen, H., Schultz, R., Shipton, Z., Mair, K. 2007. Deformation bands in sandstone – a review. Journal of the Geological Society, London 164, 755–769.

Fusán, O., Plancár, J., Ibrmajer, J. 1987. Tectonic map of basement of Tertiary in Inner Western Carpathians. Dionyza Stúra, Geological Institute of Czechoslovakia.

Hably, L., 1993. Egerian macroflora from the Andornaktálya sandpit (Hungary). Studia Botanica Hungarica, 24, 5–23.

Hohenegger, J., Coric, S., Khatun, M., Pervesler, P., Rögl, F., Rupp, C., Selge, A., Uchmann, A., Wagreich, M., 2009. Cyclostratigraphic dating in the Lower Badenian (Middle Miocene) of the Vienna Basin (Austria): the Baden-Sooss core. Int. J. Earth. Sci., 98, 915–930.

Horváth, F., 1995. Phases of compression during the evolution of the Pannonian Basin and its bearing on hydrocarbon exploration. Mar. Petr. Geol., 12, 837-844.

Horváth, F., Bada, G., Windhoffer, G., Csontos, L., Dövényi, P., Fodor, L., Grenerczy, Gy., Síkhegyi, F., Szafián, P., Székely, B., Timár, G., Tóth, L., Tóth T., 2004. Atlas of the present-day geodynamics of the Pannonian Basin: Euroconform maps with explanatory text. Eötvös Loránd Univ., Dept. of Geophysics.

http://geophysics.elte.hu/atlas/geodin_atlas.htm

Less, Gy., 2005. Palaeogene. In: Pelikán, P. & Budai, T. (Eds.), Geology of the Bükk Mountains. Geol. Inst. of Hungary, pp. 204–211.

Less, Gy., Gulácsi, Z., Kovács, S., Pelikán, P., Pentelényi, L., Rezessy, A., Sásdi, L., 2005. Geological map of the Bükk Mountains 1:50 000. Geological Institute of Hungary.

Lourens, L., Hilgen, F., Shackleton, N.J., Laskar, J., Wilson, D., 2004. The Neogene period. In: Gradstein, F.M., Ogg, J.G., Smith, A.G (Eds.) A geologic time scale 2004. Cambridge Univ Press, Cambridge, 409–440.

Lukács, R., Harangi, S., 2002. Petrogenesis of the Miocene silicic magmas in the Pannonian Basin: A case study in the Eastern Bükkalja Volcanic Field, Northern Hungary. Geol. Carpathica, 53, 13–14.

Lukács, R., Czuppon, G., Harangi, S., Szabó, C., Ntaflos, T., Koller, F., 2002. Silicate melt inclusions in ignimbrites, Bükkalja Volcanic Field, Northern Hungary – texture and geochemistry. Acta Geol. Hungarica, 45, 319–336.

Márton, E., 1990. Paleomagnetic studies on the Miocene volcanic horizons at the southern margin of the Bükk Mts. Annual Report of the Eötvös Loránd Geophy. Inst. of Hungary for 1988-89, 211-217, 307–309.

Márton, E., Fodor, L., 1995. Combination of paleomagnetic and stress data – a case study from North Hungary. Tectonophysics, 242, 99–114.

Márton, E., Márton, P., 1996. Large scale rotations in North Hungary during the Neogene as indicated by paleomagnetic data. In: Morris A. & Tarling D. (Eds.), Paleomagnetism and tectonics of the Mediterranean Region. Geol. Society Spec. Publ., 105, 153–173.

Márton, E., Pécskay, Z.,1998. Complex evaluation of paleomagnetic and K/Ar isotope data of the Miocene ignimbritic volcanics in the Bükk Foreland, Hungary. Acta Geol. Hungarica, 41, 4, 467–476.

Nagymarosy, A., 1990. Paleogeographical and paleotectonical outlines of some Intracarpathian Paleogene Basins. Geol. Carpathica, 41, 3, 259–274.

Palotai, M., Csontos, L., 2010. Strike-slip reactivation of a Paleogene to Miocene fold and thrust belt along the central part of the Mid-Hungarian Shear Zone. Geol. Carpathica, 61, 6, 483–493.

Pelikán, P., 2005. Miocene Formations of the western and northern forelands. In: Pelikán, P., Budai, T. (Eds.), Geology of the Bükk Mountains. Geol. Inst. of Hungary, 215–230.

Pentelényi, L., 2005. Miocene pyroclastic beds in the Bükkalja. In: Pelikán, P., Budai, T. (Eds.), Geology of the Bükk Mountains. Geol. Inst. of Hungary, 212–215.

Petrik, A., 2012. The structural evolution of Bükkalja based on field works and seismic section analysis. The Meeting of Young Geoscientists, Geological Society of Hungary, Tatabánya, Abstract volume, 23–24.

Petrik, A., Fodor, L., 2013. The structural evolution of Bukkalja (NE Hungary) based on field works and seismic section analysis. Tectonic Studies Group Annual Meeting, Leeds, 63.

Pittman, E.D., 1981. Effect of fault-related granulation on porosity and permeability of quartz sandstones, Simpson Group (Ordovician) Oklahoma. AAPG Bulletin, 65, 2381–2387.

Popov, S.V., Akhmetev, M.A., Zaporozhets, N.I., Voronina, A.A., Stolyarov, A.S., 1993. Evolution of the Eastern Paratethys in the Late Eocene - Early Miocene. Strat. Geol. Corr., 1/6, 10–39.

Royden, L.H., 1993. Evolution of retreating subduction boundaries formed during continental collision. Tectonics, 12, 3, 629–638.

Royden, L.H., Horváth, F., 1988. The Pannonian basin - A study in basin evolution. Amer. Assoc. Petrol. Geol. Mem. 45, 394.

Schultz, R.A., Siddharthan, R., 2005. A general framework for the occurrence and faulting of deformation bands in porous granular rocks. Tectonophysics,411, 1, 1-18.Steininger, R.R., Berggren, W.A., Kent, D.V., Bernor, R.L., Sen, S., Agusti, J., 1996. Circum Mediterranean Neogene (Miocene and Pliocene) Marine-Continental Chronologic correlations of European Mammal Units and Zones. In: Bernor, R. L., Fahlbisch, V., Rietschel, S. (Eds.): Later Neogene European Biotic Evolution and Stratigraphic Correlation. Columbia University Press, New York, 7-46.

Sztanó, O., Tari, G., 1991. Parasequences in a Late Egerian sand-pit at Andornaktálya. International Litosphere Prog. Task Force, Field trip & Workshop, Mátraháza, 46–49.

Sztanó, O., Tari, G., 1993. Early Miocene basin evolution in Northern Hungary: Tectonics and Eustacy. Tectonophysics, 226, 1–4, 485–502.

Sztanó, O., Tari, G., Varga, P., 1991. High frequency sea level changes in a late Oligocene barrier island – lagoon complex, Eger, northern Hungary. First Int. Meeting of Young Geol., Abstract volume, 63–64.

Tari, G., 1988: Strike-slip origin of the Vatta-Maklár trough. Acta Geol. Hungarica, 31, 101–109.

Tari, G., Báldi, T., Báldi-Beke, M., 1993. Paleogene retroarc flexural basin beneath the Neogene Pannonian Basin: a geodynamic model. Tectonophysics, 226, 433–455.

Figure caption

Fig.1. Schematic structural evolution maps of the Carpathian Basin depicting the main deformational phases of Cenozoic era. The major depocenters for each time span are also indicated. Insets show the local structures of the study area at the time of D2 (A), D4 (B) and D7 (C) deformational phases.

Fig.2. Simplified geological map of the study area with the position of investigated sites, seismic profile and geological cross sections. Note the dominance of NW dipping normal faults which mainly acted during Mid Late Miocene.

Fig.3. Lithostatigraphic chart of the study area for the Cenozoic era. Standard stages and their absolute ages are based on Lourens et al. (2004). Central Paratethys stages and their absolute ages are based on Hohenegger et al., (2009a); Popov et al., (1993); Steininger et al., (1996) The K/Ar radiometric ages of volcaniclastic levels are taken from Márton and Pécskay (1998). Deformation band sampling points are also indicated along lithostatigraphic column (site 1,5,6,8 see in Fig. 2).

Fig.4. Schematic geological NNW-SSE oriented cross section indicating the different aged subbasins and their rock piles for the NHSSPB. Major fault systems are also displayed. Note the normal reactivation along Darnó Zone. Note the synsediment thickening of Paleogene rocks in front of propagating reverse faults and their pinching out to north-westward direction. Post-rift rock suites only appear in southeast in the Vatta-Maklár Trough.

Fig. 5. Schematic models for tilt tests (1) and their application on field measurements (2).

(A) Schematic models are depicting two scenarios for tilting of Cenozoic stratigraphic units. The tilting events in case of bigger dips of beds (1). The tilting events in case of more moderate dips of beds (2).
(B) Examples for tilt tests for sites 5, 6 (see Fig. 2 for location, Fig. 13 for complete final interpretation). Different structures are displayed on lower hemisphere Schmidt stereonet. Circles indicate the position of intersection line of conjugate fractures. Rectangles indicate the oblique position of striae. Dashed framed stereograms indicate the position when fractures were formed.

Fig. 6. The generalized subsidence history of the study area based on 3 boreholes and a few calibration data. Orange rectangles indicate the possible burial depth of deformation bands at the time of their formation. Black dashed lines show the end of main erosion periods. Black dotted lines indicate the subsidence periods. Short black line indicates the confidence interval

of the burial depth of deformation bands. A4 deformation bands are indicated twice because of their reactivation in later deformation phase (see details in chapter 4.2).

Fig. 7. NW-SE oriented geological cross sections with preserved and possible top and base of Mid Miocene (B) and Pannonian sediments (C). Site 1, 5, 8 are the sampling points of deformation bands (see in Figs. 2, 13). Blue dashed lines on B and C sections indicate the possible top of Mid Miocene and Pannonian, respectively. Black dashed lines on B and C sections indicate the preserved top of Mid Miocene and Pannonian, respectively. Dotted lines on B and C sections show the base of Mid Miocene and Pannonian, respectively.

Fig. 8. Deformation bands at site 8 (Szőlőske) with stereograms, field photos and thin section.

A) B) more indurated deformation bands standing out from less indurated conglomerates

C) Closer view of S1 deformation band (from location of photo B), in which oriented pebbles can be seen with some fractured pebbles (black lines).

D) Macroscopic and microscopic photo of thin section were made from S1 samples show the contact of sedimentary layering and deformation band in which narrow cataclasis occurs with reorientation of pebbles indicating shear.

E) Near horizontal wavy layers were dragged along S4 deformation band.

Fig.9. Deformation bands at site 5. (Andornaktálya) with stereogram and field photos.

A) and B) pictures show the same, clustered A2 deformation bands (continuous). In photo A, deformation bands are shown as single offset with 15 cm displacement in fine grained sandstone, while deformation bands in photo B, display thickened zone of anastomosing bands with max few cm offsets (see inset of a deformed trace fossil) in coarser grained sandstone.

C) A4 deformation bands and related fault with slip surface. Pink dashed line indicates a clayey fault surface which bears several striations. Inset shows a closer view of zone, where densely clustered deformation bands are associated with discrete fault. White colour of bands indicates posterior bleaching of zone.

D) A1 single deformation band with few cm offset marked by a faulted clay clast.

Fig. 10. Deformation bands at site 6. (Eger, Wind brickyard) with stereogram, photos and cross section.

A) More indurated deformation band standing out in less indurated sandstone with decimetre scale offset. This thickened deformation band is interpreted as a dilatation fault segment developed at the boundary of mechanically different host rock.
B) Small reverse faults (related to D6 phase) offset normal faults of D2a phase
C) Folds in clay accommodate reverse separation.
D) NE-SW oriented geological cross section of Wind Brickyard with main deformation structures including D2a normal faults and D6 reverse faults.

Fig. 11. Photo of deformation bands at site 1. (Novaj) with their stereogram. White dashed lines indicate deformation bands in sandstone have decimetre scale offset (blue dashed lines indicate a faulted a marker layer).

Fig. 12. Thin sections of deformation bands.

A) Reoriented pebbles with incipient pebble fracturing considered as disaggregation band in S4 samples from site 8. conglomerates (under plane polars).
B) Deformation bands from site 1. (under crossed and plane polars) indicated by well sorted finer grained zone by cataclasis in contrast of weakly sorted coarser grained host sandstone. Thin section was made parallel to beddings. Limonite precipitated out but along deformation bands.
C) Grain aggregates mark the deformation bands in sample A1 (from site 5.), arranged in parallel bands. Aggregates formed by shear related compaction, weak cataclasis are present at contacts of grains (white line shows the sedimentary bedding).
D) During normal shear, mica flake (under crossed polars) has curved in A1 deformation bands (site 5.) following the shapes of their neighbours (white line shows the sedimentary bedding).
E) Deformation band A2 (from site 5.), show more intense cataclasis than A1, but lot of main grains remained intact floating in cataclastic matrix. Crude orientation of main grains (intact grains) parallel to band indicates shear (shear orientation marked by black dashed line).
F) Anastomosing deformation band of A4 sample (from site 5.), show more even more progressed cataclasis than A2, with no intact main grains. Along but out of bands ferrigenuous cement precipitated.
G) Dilation (deformation) band (white arrows) from site 6. represented by zone of host rock grains floating in dark fine grained micrite matrix. Transition zone between host

rock and dilation band shows crude vertical orientation of main grains indicating shear.

H) Within deformation band (site 6.) needle cement on grains (under crossed polars) shows further dilation in fine grained carbonate matrix. Pores of micrite matrix started to fill with cement causing more induration of matrix. Therefore, matrix cannot able to fill the newly forming voids during further dilation, and precipitated needle cement. It was happened prior to micrite matrix lithification, because no traces of brittle fracture.

Fig. 13. The summary table of stress fields and main deformation phases. For legend see Figure 5. RUP (in %), ANG (in degree) criteria and phi value are indicated for stress tensor calculation. After the value of criteria, the number of unfitted data is also displayed.

Fig. 14. Uninterpreted (a) and interpreted seismic profile (b) of the study area. The location of the seismic profile see in Fig. 3. Note the blind reverse faults of D2b phase. Normal faults dipping NW belong to the D5, D7 phases (see in Fig. 13).

Fig. 15. The evolution of deformation bands with respect to displacement and burial depth. A) deformation mechanism (after Fossen et al. 2007) and deformation phases for the studied bands. B) Burial history and possible depth of deformation band formations (orange diamonds). Thick black line indicates the burial depth of the base Kiscell Fm. (Early Oligocene). Black dotted lines indicate the end of deformation phases. The blue dashed lines show the end of main subsidence periods while the red ones indicate the erosional periods. The increasing burial depth favours the development of cataclastic type deformation mechanism which is associated with greater displacement. C) shows q-p diagram applied for porous media by modified Cam cap model (Schultz and Siddharthan, 2005). $p = I_1$, $q = (I_1^2 - 3 I_2)^{1/2} = \sigma_1 - \sigma_3$) where $I_1 = \sigma_1 + \sigma_2 + \sigma_3$, $I_2 = \sigma_1\sigma_2 + \sigma_2\sigma_3 + \sigma_1\sigma_3$ and the curve represents the yielding surface. Normal vectors to yield surface indicate shear induced dilation at lower confining pressure (p), and compaction at higher confining pressure. Moving outward from yield surface, strain hardening associates with deformation band formation. d) Dashed lines indicates possible loading pathways. A shows initiation of deformation bands on yield surface, the location of A points to how the yield surface will move with increasing strain and which kind of band will form. B indicates maximum frictional strength of newly formed bands by strain hardening. Between B and C small discrete faults nucleate at bands. The damage zone is growing due to faulted bands until reach the point of through-going fault (C).