DISSERTATION

Titel der Dissertation

„Quaternary landscape evolution of the Little Hungarian Plain"

Verfasser

Mag. András Zámolyi

angestrebter akademischer Grad

Doktor der Naturwissenschaften (Dr. rer. nat.)

Wien, 2015

Studienkennzahl lt. Studienblatt:	A 091 431
Dissertationsgebiet lt. Studienblatt:	Dr.-Studium der Naturwissenschaften, Studienzweig Geologie
Betreuerin / Betreuer:	Univ.-Prof. Dr. Bernhard Grasemann
	PhD. Székely Balázs

1. Introduction

The Little Hungarian Plain is a sub basin of the Pannonian basin and is located at the transition zone between the Eastern Alps and the Western Carpathians. It is an extensional basin that evolved along low angle normal faults due to the lateral extrusion of the Eastern Alps and subduction roll-back beneath the Carpathians (Horváth, 1993). While the Pannonian basin area experienced a Late Miocene compressional phase on a large scale (Tari, 1994; Horváth, 1995), the Little Hungarian Plain is characterized by on-going subsidence (Joó, 1992) since the Early Miocene (Royden and Dövényi, 1988). Upper Miocene strata in the central part of the Little Hungarian Plain reach a thickness of up to 2400 m as deducted from well data (Körössy, 1987; e.g. well Mosonszentjános 1). At the western margins, the thickness of Upper Miocene sediments is around 1500 m (Körössy, 1987; well Mosonszolnok 2) and 1250 m (Körössy, 1987; well Pinnye 2).

Due to ongoing subsidence the basin morphology of the central part of the Little Hungarian Plain is characterized by a very low relief. Only at the margins can we observe pronounced hills or remnants of distinct river terraces. If we compare the geomorphologic appearance of the Little Hungarian Plain to the adjacent Vienna Basin (Fig. 1), we see that in the Vienna Basin several terrace remnants can morphologically be observed in the central parts of the basin (Decker et al., 2005). This fact indicates Quaternary faulting which can be further supported by sedimentologic evidence (i.e. stratigraphic correlations of the downthrown parts, dating of strata by fossil finds, etc…). It is important to note that the fault planes of the Quaternary faults bounding the terraces can be observed close to the surface (Beidinger and Decker, 2011). In contrast, the morphologic expression of terraces in the central part of the Danube Basin is very minimal due to (i) the on-going subsidence and the capabilities of the rivers to transport enough sediment into the basin and because (ii) subsidence is accommodated along reactivated basement faults with no fault planes close to the surface (Fig. 1 and Appendix B).

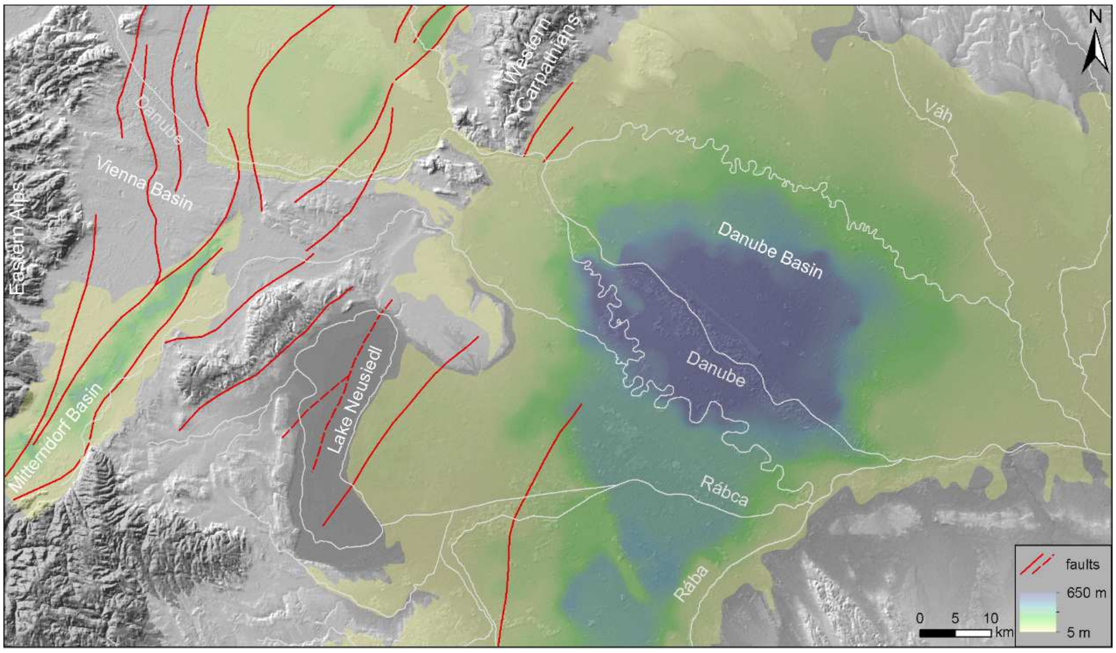

Fig. 1.: Quaternary thickness map of the Vienna Basin and the Danube Basin modified after Decker et al. (2005) and Scharek et al. (2000) on an SRTM hillshade. Faults after Schnabel (2002). Note the differences in relief and thickness of Quaternary sediments between the Vienna Basin and the adjacent Danube Basin.

The morphology of the basin margins is more pronounced. The main boundaries are formed by the Rust Hills, Leitha Mountains and the Male Karpaty Mountains in the West and by the Hungarian Mountain Range (including the Transdanubian Range and Vértes Hills) in the East.

At the end of the Middle Miocene the Hungarian Mountain Range formed a peninsula with roughly the same shape as present (Jámbor, 1980). This palaeogeographic setting changed in the early Late Miocene accompanied by the transition from shallow marine environment to the brackish Lake Pannon which covered the entire Hungarian Mountain range (Jámbor, 1980; Magyar et al., 2013). The uplift of the region can be divided into several phases and began at the end of the Middle Miocene. The present day uplift rate is +1 to 1.5 mm/year (Joó, 1992; Ruszkiczay-Rüdiger et al., 2005). As a result of uplift the pre-existing fault pattern within the crystalline rock is revealed by exogenous processes (see appendix D). Thus, the older fault pattern influences the orientation of modern valleys. As a further effect of uplift fluvial deposits can be observed in elevated positions along the eastern margin of the Little Hungarian Plain (Fig. 2): In the Vértes Hills 79 ka old fluvial sediments are located at 237 m a.s.l. (Thamó-Bozsó et al., 2010) in a quarry close to Süttõ 75.3 +/-4.7 ka old fluvial loess deposits lie at 240.5 m a.s.l. (Novothny et al., 2011). At Salföld the surfaces lying at presently 140 m a.s.l. were exposed at 287 ka (Ruszkiczay-Rüdiger et al., 2011).

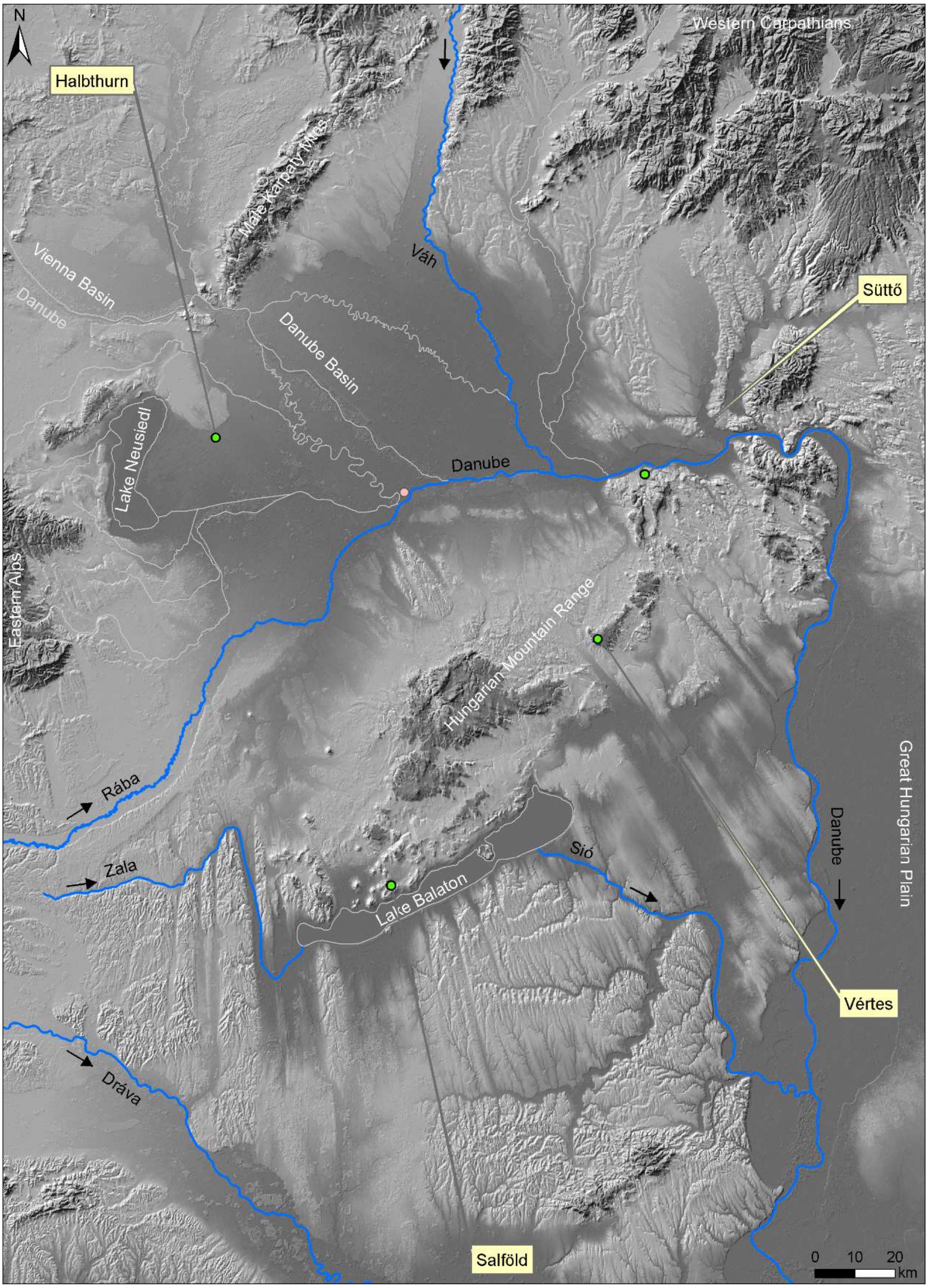

Fig. 2: Shaded SRTM DTM of the western part of the Pannonian Basin with certain sections of the present river courses highlighted. Note that the Danube, Rába and Zala Rivers flow around the rapidly uplifting Hungarian Mountain Range. The river courses of the Váh and Rába Rivers erode the eastern margins of their valleys indicating on-going subsidence in the depo-center of the Danube Basin (red dot). The callouts mark the locations at which following geochronologic data were derived: (i) **Halbthurn**: 75 ka fluvial sediments at 124 m a.s.l. (Appendix C), (ii) **Süttõ**: 75.3 ka fluvial loess at 240.5 m a.s.l. (Novothny et al., 2011), (iii) **Vértes**: 79 ka fluvial sediments at 237 m a.s.l. (Thamó-Bozsó et al., 2010), (iv) **Salföld**: surface presently at 140 m a.s.l. was exposed at 287 ka (Ruszkiczay-Rüdiger et al., 2011).

In the West, the Leitha Mountains emerged as a low-topography island or platform of metamorphic rocks during the formation of the Vienna Basin in the Karpatian (~16 Ma). Evidence for this is derived from the predominantly biologically dominated sedimentation of Middle Badenian limestone in the southwestern part of the Leitha Mountains (Wiedl et al., 2014). After the deposition of Upper Badenian limestones a major subsidence in the NE-part of the Leitha Mountains occurred (Wiedl et al., 2014). However, the highest Quaternary fluvial deposits (>300 ka) in the area are located on the Parndorf Plateau at an elevation of ~160 m a.s.l., and 75 ka old fluvial sediments lie at 124 m a.s.l. (Appendix C). Interestingly, lower Middle Miocene fluvial gravel deposits can be found in the Rust Hills at an elevation of ~206 m a.s.l. (Schönlaub, 2000).

Previous literature states that the Quaternary gravels on top of the Parndorf Plateau are genetically equivalent to the gravels in the Seewinkel Plain (e.g. Schnabel, 2002; Tauber, 1959). Current thesis suggests that there is a significant age difference in deposition between the gravels on the Parndorf Plateau and the Seewinkel Plain (Appendix C). The sediments located at higher elevations (~160 m a.s.l.) on the Parndorf Plateau are older (>300 ka) than the sediments at lower elevation of 124 m a.s.l. in the Seewinkel Plain (75 ka) (Appendix C).

2. Aim

The aim of the study is to provide possible answers to key questions in the landscape evolution of the western margin of the Little Hungarian Plain. For this, digital geomorphology was used as the basic approach, complemented by the integrated analysis of geomorphologic observations with seismic and well data.

The key questions were:

- Can we spot the surface expression of tectonic features in such a low relief area?
- Can we observe on-going subsidence and tectonic activity in recent river dynamics?
- How did such a conspicuous landscape feature as the Parndorf Plateau form? Is it tectonically influenced?
- What possibly caused the formation of the Neusiedler See?
- How are the gravel deposits in the Seewinkel Plain related to the gravels on the Parndorf Plateau?
- What is the provenance of the gravels?

The methods applied to develop hypotheses to solve these questions included the integration of already available data from literature with freshly derived results from new measurements. The location of the study area at the border of three different countries and a plethora of available data from international co-operations (i.e. 2000 DANREG project) indicated the crucial role of gathering of data in an adequate spatial database and integrated analysis. Based on the results new multidisciplinary methods were devised to acquire new data for the identified focus points.

The low relief of the study area requires adequate digital data. The technique of Airborne Laser Scanning (ALS) provides digital terrain models (DTM) with the necessary high resolution and precision. Available ALS data from 1999 and 2004 were merged and processed to retrieve a DTM for geomorphologic and structural geologic analysis (see Appendix A).

The focus of digital terrain analysis shifted towards the fluvial dynamics and river sinuosity. In this case, accurately georeferenced historic maps allowed for the analysis of river courses not influenced by anthropogenic activity (see chapter 6). The river sinuosity of the streams of the Little Hungarian Plain was not assessed and interpreted previously.

Both above mentioned digital geomorphologic studies revealed that the landscape is highly influenced by reactivation of older, subsurface structural features. Geophysical methods, especially shallow lake seismic measurements were used along with lake drilling to reveal the relationship between Early to Late Miocene geodynamics and modern landscape. Existing industrial seismic data and information from deep, as well as shallow exploration wells were utilized to complete the new measurements. Luminescence dating and provenance analysis were included and as consequence a model of the deposition sequence of Lower Miocene sediments and Quaternary fluvial deposits in the study area was developed (Appendix C).

3. Geomorphologic analysis

For the geomorphologic analysis of the low relief surface of the Little Hungarian Plain an Airborne Laser Scanning derived digital terrain model (ALS DTM) was used (Appendix A). Regarding geo- and environmental sciences, the ALS technique was first widely used in forest management for determining forest canopy heights or the estimation of biomass (e.g. Nelson, 1984). Flood management and topographic applications followed later (e.g. Hill et al., 2000) and by 2003 geomorphologic studies were conducted based on ALS data (e.g. Hooper et al., 2003). The ALS survey of the Seewinkel Plain was carried out in 2004 with first processing results published in 2006 (Attwenger and Chlaupek, 2006), followed by hydrological analysis (Bitenc, 2007) and geomorphological/structural geologic interpretation in 2008/2009 (Appendix A).

From the integrated analysis of the ALS DTM indications were derived for a neotectonic origin of the mapped linear elevated ridges (Appendix A). Taking this into consideration, a possible deep-seated influence of these linear features could not be excluded in such a basin environment and thus this topic was chosen as next investigation step. The specific geodynamic setting and depositional environment leading to this way of thinking is elaborated in the next chapters 4 and 5.

4. Basin evolution

The Danube Basin is regarded as a back-arc basin that formed during the extensional collapse of the overthickened Alpine nappe stack and coeval subduction roll-back beneath the Carpathians (Horváth et al., 2006). The initiation of basin formation can be defined by the first Ottnangian rhyolite tuff horizon that can be observed basin-wide (Horváth, 1995). The Danube Basin displays a well-defined succession of pre-rift basement, syn-rift (18-13 Ma), and post-rift (13-6 Ma) sequences (Tari, 1994; Horváth and Cloetingh, 1996; Fodor, 2005).
The pre-rift basement consists of Mesozoic or Palaeozoic rocks that were deformed during the Eoalpine orogenic phase in the Cretaceous (Tari, 1994; Horváth 1995). The architecture of the pre-rift substrata is of special importance since it includes a set of several major faults (Szafián et al., 1999 and citations therein) bounding the basement highs. Reactivation of major faults of Miocene age highly influences modern river dynamics (see also Appendix B).

A characteristic of not only the Danube Basin but the entire Pannonian basin is the fact that the syn-rift sedimentary sequence is relatively thin compared to the thick post-rift succession (Tari, 1994). In the following, the post-rift geodynamic evolution is of further interest: After a brief compressional phase at the end of the Middle Miocene (12-9 Ma) filling up of the basin continued until the latest, neotectonic compressional event (5.3 Ma to present) (Horváth, 1995; Horváth and Cloetingh, 1996). Due to the on-going subsidence in the Little Hungarian Plain (Joó, 1992) Pannonian (10.5 to 7.8 Ma) sediments show a continuous thickening towards the center of the basin (Tari, 1994; Appendix C).
As a consequence of large-scale deformation of the hot and rheologically weak lithosphere during the neotectonic compressional phase (Horváth and Cloetingh, 1996), the center of the Little Hungarian Plain (located around the city of Győr) is still subsiding while at the same time the flanks of the basin are subject to uplift (Joó, 1992; Tari, 1994; Appendix C).

5. River dynamics

At the end of the Middle Miocene the Vienna Basin and the adjacent Pannonian Basin both belonged to the Central Paratethys and were isolated from the Eastern Paratethys at around 11.6 Ma (ter Borgh et al., 2013) which coincides with the Sarmatian/Pannonian boundary. From approximately 10 Ma on sediment supply rates gradually exceeded the subsidence rates and the paleo-Danube shelf progressed continuously into the Lake Pannon, the brackish successor of the Central Paratethys (Magyar et al., 2013). Due to this gradual regression of the Lake Pannon, fluvial processes became more and more dominant in the Little Hungarian Plain. The terrestrial/fluvial environment remained dominant in the area until present (e.g. Harzhauser et al., 2007). In this row of events, regional uplift plays a major role: Ter Borgh et al. (2013) suggest regional uplift of the Carpathians and adjacent areas to cause the isolation of the Central Paratethys. Another uplift event affected the region at ~4 Ma (Wagner et al., 2010). The causes triggering this event are not entirely clear.

In the Pleistocene the area was not directly affected by glaciation (e.g. van Husen, 2011; van Husen and Reitner, 2011) and was part of the European periglacial zone during the Middle and Late Pleistocene (van Husen and Reitner, 2011; Homolová, 2012; Vanderberghe et al., 2014). The indications for periglacial conditions are preserved in cryoturbated sediments (Fig. 3), as well as ventifacts (Sebe et al., in press). In this epoch cool treeless-steppe periods alternated with a forest-steppe environment (Katona et al., 2012 and citations therein). Although aeolian processes are reported to play a major role in landscape formation during this time (Ruszkiczay-Rüdiger et al., 2011; Sebe et al., in press) in the eastern part of the Pannonian basin the existence of a long-lasting, large meandering river could be verified during 47 to 24 ka (Cserkész-Nagy et al., 2012). Taking into consideration the work of Nádor and Sztanó (2010) we can conclude that this is not an isolated observation and that rivers in the Pannonian basin were probably significantly larger during the Pleistocene than at present. From observations on fluvial deposits in the Netherlands, Germany and Poland Mol et al. (2000) deduce main phases of development of meandering rivers in the Early Glacial, Late Glacial and Holocene. Strong variations in discharges and thus incision or rapid aggradation are linked to strong permafrost, absence of vegetation (Mol et al., 2000) as well as to climatically unstable periods between glacials and interglacials (Ruszkiczay-Rüdiger et al., 2005). Aggradation and meandering occur in stable climate periods during both glacials and interglacials (Gábris, 1997; Mol et al., 2000; Vanderberghe, 2003). The major fluvial deposits on the Seewinkel Plain that could be dated to ~ 75 ka by Infrared Luminescence (Appendix C) indicate a stage of major sediment supply at the end of MIS 5 corresponding to the transition from the last interglacial to the last glacial (Helmens, 2014).


Fig. 3: Cryoturbated quaternary sediments at Friedrichshof in the central part of the Parndorf Plateau. The view is towards ESE, the folding rule of 1 m length serves as scale. View towards E.

Taking the above considerations into account, we can conclude that after ~75 ka the drainage pattern in the Little Hungarian Plain was in a stable enough state to be able to record tectonic influence. The assessment of river sinuosity was chosen as the appropriate method to investigate this possible influence (Appendix B).

6. Historic maps and river sinuosity

Application of historic maps in geomorphology and environmental sciences is crucial to gain the invaluable insight in spatiotemporal variations of the given study areas (Raper, 2000). The feasibility of using historic maps depends on various factors: (i) level of detail of the maps, (ii) precise drawing and interpretability of map elements, and (iii) possibility of accurate georeferencing. All these factors have to satisfyingly full-filled to be able to integrate historic maps. Due to these circumstances this approach is still on the rise (e.g. Timár and Rácz, 2002; Timár and Molnár, 2003; Tschegg, 2012; James et al., 2012; Petrovszki et al., 2012). For the Little Hungarian Plain this approach was first utilized during this thesis (Appendix B).

To be able to derive river sinuosity without the effects of the most recent anthropogenic impact, exactly geo-referenced historic maps were used to reconstruct the original channel geometries that in essence presumably prevailed since ~75 ka (Appendix B). The Little Hungarian Plain was covered by two comprehensive topographic mapping campaigns in historic times: The first military survey of the Habsburg Empire (1782-1785) and the second military survey (1819-1869) (Kretschmer et al., 2004). For digitization of the historic river courses and river sinuosity calculations the second military survey was chosen considering the level of detail of mapping and the accuracy of georeferencing (Appendix B).

7. Geophysics

A key part of the western margin of the Little Hungarian Plain is covered by the Lake Neusiedl which renders geologic investigations and the digital geomorphologic approach difficult. This area is important due to its transitional position between the Leitha Mountains and the Little Hungarian Plain. Thus, shallow lake seismic measurements were conducted to gain an insight into the subsurface geometry of the sediments beneath the lake (Appendix C).

Lake seismic measurements were applied to study the sedimentation and faulting processes by the 1980ies in the United States (e.g. Johnson, 1980). Van Rensbergen et al. (1998) used high resolution seismic measurements to investigate the Quaternary stratigraphy of Lake Annecy (France).

The Lake Balaton is the closest lake to the study area that was investigated between 1993 and 2007 using high resolution shallow lake seismic (Tóth et al., 2010). Different seismic facies, unconformities, tilted and deformed layers were clearly distinguishable in the lake seismic sections and the data quality and quantity allowed for the generation of comprehensive neotectonic maps (e.g. Bada et al., 2010). For most of the campaign the IKB-Seistec™ device was used with a signal generation frequency range of 1-10 kHz, a penetration of 20-40m and a resolution of 10-20cm (Tóth et al., 2010). Based on the good experiences at Lake Balaton and the vicinity the same equipment was used to study the sediment geometry beneath Lake Neusiedl (Appendix C).

8. Luminescence dating

Sediments samples were taken for Luminescence dating to clarify the question on the connection between the fluvial gravel deposits on top of the Parndorf Plateau and the gravels in the Seewinkel Plain. Also, scientific literature on this region starting from the first half of the 20th century (e.g. Szontagh, 1904; Szádeczky-Kardoss, 1938) up to more recent (e.g. Tauber, 1959) is not clear on the topic of tectonic versus erosive formation of the relatively steep dipping flanks of the Parndorf Plateau.

The samples derive from three locations: Nickelsdorf (Parndorf Plateau), Frauenkirchen and Halbthurn (both from the Seewinkel Plain). They were taken from sand lenses within fluvial gravels (Appendix C). The coarse grain feldspar fraction of the samples was dated with the infrared stimulated luminescence (IRSL) method (e.g. Preusser, 2003).

IRSL dating provides us with the date of last exposure to sunlight. If the signal is not re-set we can exactly determine the date of last deposition and burial of the sediment (e.g. Rittenour, 2008). Fluvial sediment can contain remnant signals of the previous deposition cycle due to partial bleaching of the mineral grains and thus the age of deposition can be over-estimated (e.g. Prescott and Robertson, 1997; Cunningham et al., 2015). With the application of the single-aliquot regenerative-dose (SAR) protocol (e.g. Wallinga et al., 2000; Wintle, 2008) sensitivity changes are monitored and corrected. Thus, the possible influence of partial bleaching is reduced.

9. Sedimentology

To complete the previously mentioned geophysical and geochronological methods, sedimentological analysis was conducted on lake drilling cores and in outcrops at the eastern shore of the Lake Neusiedl (Appendix C).

The lake drilling cores penetrated up to 3 m beneath the surface of the lake bottom. The description of the lake drilling cores followed standard procedures (Appendix C).

Heavy mineral analysis was carried out on samples of the lake drilling cores and of the IRSL samples (Appendix C).

In order to be able to support heavy mineral analysis key outcrops were sampled for component analysis (Fig. 4). Material between 11 and 14 kg was gathered and the fraction coarser than 5 mm was sorted according to lithology. Results are shown in table 1 and figure 5.

Rock type	Outcrops [g]			Outcrops [weight %]		
	Halbthurn (KIR1)	Bezenye	Nickelsdorf	Halbthurn	Bezenye	Nickelsdorf
Quartz	6480	4820	6030	45	41	51
polycrystalline Quartz	0	540	20	0	5	0
Quartzites	502	1010	960	4	9	8
Granite	40	320	140	0	3	1
Sandstone	1345	970	700	9	8	6
Gneiss	710	60	150	5	1	1
limestone, dolomite	330	260	0	2	2	0
Breccia	0	140	50	0	1	0
Basalt/Andesite	107	30	20	1	0	0
fraction <5mm	4740	3480	3800	33	30	32
	14254	11630	11870	100	100	100

Tab. 1.: Results of the component analysis of samples from key outcrops in dry weight [g] and %.

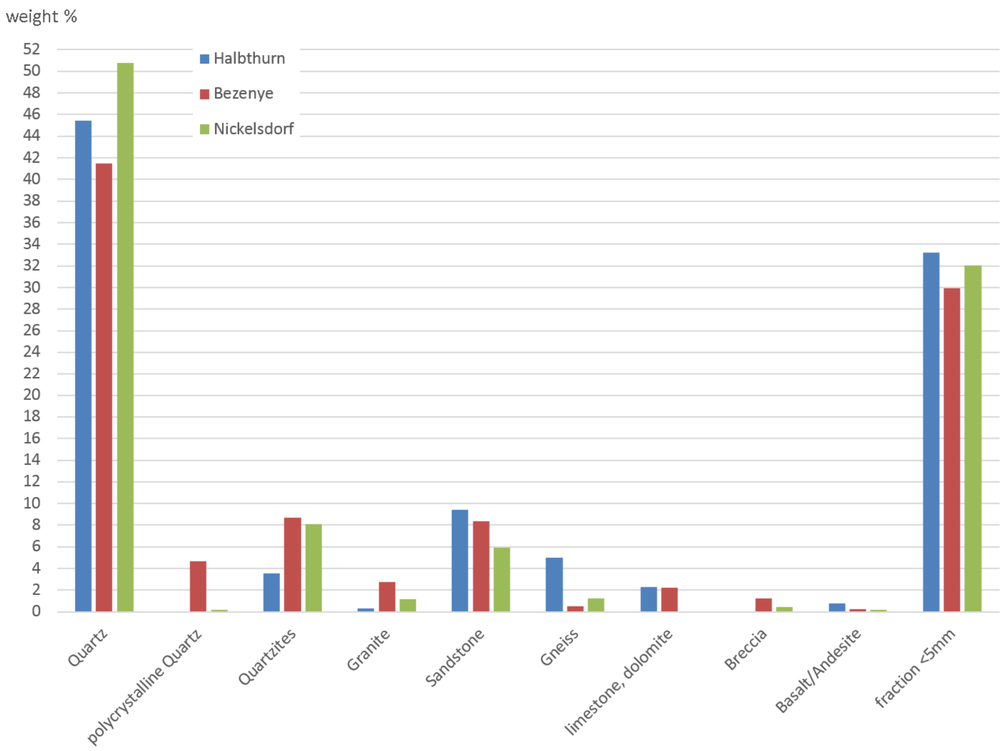

Fig. 4.: Results of the component analysis of samples from key outcrops in dry weight %.

A significant difference in the distribution of the components can be observed between the samples from Nickelsdorf on the Parndorf Plateau and the samples from the Seewinkel Plain (Halbthurn, Bezenye). The sample from Nickesldorf contained no carbonatic components. This lack of limestones and/or dolomites is either due to weathering (deposition of gravels at Nickelsdorf prior to 300 ka according to Appendix C) or because the gravels on the Parndorf Plateau derive from a different source area. Figure 5 and 6 show selected examples of the components found in the material from Nickelsdorf (Fig. 5) and Bezenye (Fig. 6).

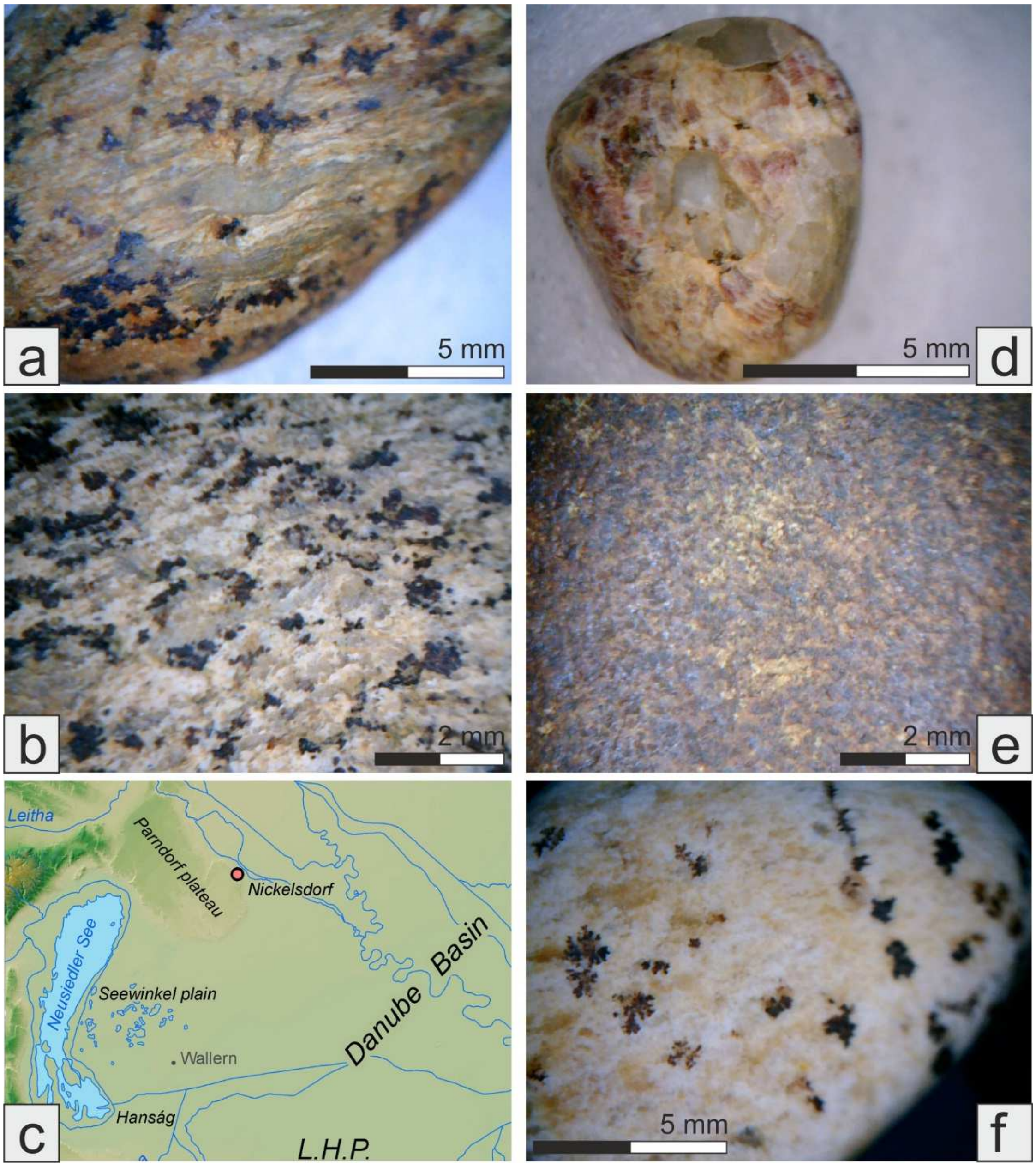

Fig. 5.: Some selected components from the Parndorf Plateau (outcrop at Nickelsdorf). 5a.:Mylonitic quartzite with well developed layering. 5b.: Granite. 5c.:Location map. 5d.: Breccia clast, possibly a remnant of a pegmatite. 5e.: Sandstone. 5f.: Quartz pebble.

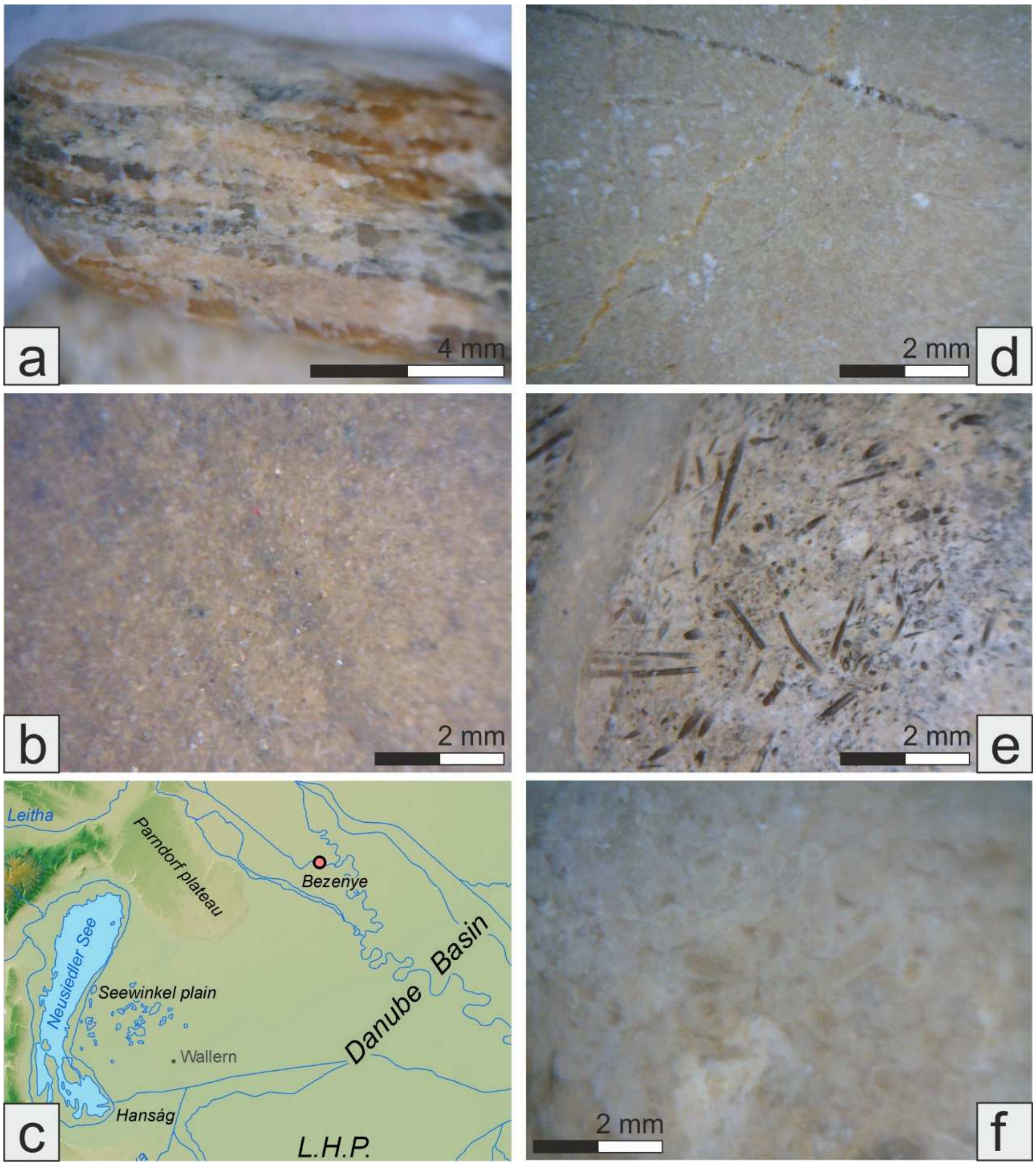

Fig. 6.: Some selected components from the plain below the Parndorf Plateau (outcrop at Bezenye). 6a.:Quartzite with recognizable layering. 6b.: Sandstone. 6c.:Location map. 6d.: Micritic limestone (Folk, 1962). 6e.: Biomicritic limestone (Folk, 1962). 5f.: Oomicritic limestone (Folk, 1962).

10. Summary

The western margin of the Little Hungarian Plain has been investigated by multidisciplinary methods using the digital geomorphologic approach as a starting point. The majority of the methods were not applied in the study area up to now and helped to gain new insights into the Late Miocene to Quaternary evolution of the area.

The key questions raised at the beginning of the study (chapter 2) were addressed by digital terrain analysis, river dynamic investigation, integration of historic maps, geophysical measurements and sedimentologic analysis. The detailed description of the methods and the results can be found in 3 published papers (Appendix A, B and D) and one submitted manuscript (Appendix C).

In the paper Székely et al. (Appendix A) the Seewinkel Plain, a key lowland area of the western margin of the Little Hungarian Plain, is investigated with the Airborne Laser Scanning (ALS) technique. This method is an emerging technique for geomorphologic and structural geologic analysis (see chapter 3) and has been applied for the first time in the Little Hungarian Plain. The method provides the necessary resolution and accuracy to distinguish features with a relief of up to only 2 m. Linear ridges arranged in an orthorhombic pattern were detected. These ridges fit into the local fault pattern which – along with other phenomena – indicates a probable neotectonic origin. The main axis of the orthorhombic pattern runs NE-SW and thus sub-parallel with the so called Lake Neusiedl Fault, a fault that was inferred by previous literature but is still debated. With the ALS method the micro-topographic expression of strike-slip and normal faulting was assessed.

The neotectonic activity of the faults in the Little Hungarian Plain has been further studied in the publication of Zámolyi et al. (Appendix B) by means of river dynamics. River sinuosity is a highly indicative method when investigating the tectonic influence on changes in channel geometries of alluvial rivers (e.g. Ouchi, 1985). The river sinuosity index was calculated for five smaller rivers of the western margin of the Little Hungarian Plain using different window sizes. This approach adheres to the scale-sensitivity of the method. Historic maps (chapter 6) were utilized to reconstruct the river courses least influenced by anthropogenic activity. Calculations with all window sizes yielded the essentially same results showing the influence of faults with a major normal movement component on the river courses. On-going channel pattern adjustment in the Little Hungarian Plain was demonstrated. It is most probably linked to the reactivation of major Miocene tectonic elements.

To place above mentioned findings into a basin-wide context, structural geologic analysis of the eastern margin of the Little Hungarian Plain was carried out (Appendix D). The study Zámolyi et al. focused on the orientation of geomorphologically significant faults versus faults without geomorphologic expression. The results show a slight but significant difference of 10-15° in fault pattern orientation between the uplifting Hungarian Mountain Range (Bakony Mountains in particular) and the surrounding areas. This can be interpreted as selective denudation of inherited fault patterns. When comparing the recent elevation of fluvial sediments on both margins of the Little Hungarian Plain we can observe that the rate of uplift at the eastern margin was higher than at the western margin (Chapter 1, Appendix C).

In the submitted manuscript Zámolyi et al. (Appendix C) key phases in the formation of the landscape of the western margin of the Little Hungarian Plain are assessed. Luminescence dating helps to pinpoint the time of deposition of the Quaternary gravels on the Parndorf Plateau and the lowlands of the Seewinkel Plain. High-resolution lake seismic measurements combined with lake drilling provide insight into the important area beneath the Lake Neusiedl. No continuous layer of Quaternary gravels could be detected in the lake. Thus, the interpretation of the data reveals the geometry of the underlying Upper Miocene (Upper Pannonian) strata: Constant thickening of the Upper Pannonian sediments starts eastward of the central axis of the lake. No thickening of sediments could be detected towards the NE, towards the Parndorf Plateau indicating either no major tectonic movement or post-sedimentary faulting. Accordingly and in line with results from provenance analysis, two geologic models are presented. The first model favors the deposition of the Quaternary gravels from local tributaries from the South and a Late Miocene subsidence of the Seewinkel Plain. The second model includes the deposition of gravels by the Danube River from the North and a Quaternary subsidence of the areas surrounding the Parndorf Plateau. In both cases earlier basin evolution history and reactivation of older tectonic features plays an important role in shaping the modern landscape of the western margin of the Little Hungarian Plain.

11. Zusammenfassung

Der westliche Rand der kleinen ungarischen Tiefebene wurde mit multidisziplinären Methoden untersucht. Als Ausgangsbasis diente die Sichtweise der digitalen Geomorphologie. Der Großteil der Methoden wurde bis jetzt in dem Untersuchungsgebiet noch nicht angewandt und lieferte neue Erkenntnisse auf dem Gebiet der spätmiozänen bis quartären Beckenentwicklung der kleinen ungarischen Tiefebene.

Die zentralen Fragen die zu Beginn der Untersuchungen aufgeworfen wurden (siehe Kapitel 2), wurden mit Hilfe der Analyse von digitalen Geländemodellen, der Untersuchung der Dynamik von Flussläufen, Integration historischer Karten, geophysikalischer Messungen und sedimentologischen Analysen behandelt. Eine detaillierte Beschreibung der Methoden und einzelnen Ergebnisse sind in drei publizierten Artikeln (Appendix A, B und D), sowie in einem eingereichten Manuskript (Appenidx C) dokumentiert.

Im Artikel Székely et al. (Appendix A) wurde der Seewinkel, eine wichtiges Flachlandgebiet am westlichen Rande der kleinen ungarischen Tiefebene mit Airborne Laser Scanning (ALS) untersucht. Diese Methode findet immer mehr Verbreitung in den Bereichen der Geomorphologie und Strukturgeologie (siehe Kapitel 3) und wurde in der kleinen ungarischen Tiefebene zum ersten Mal angewendet. Die Methode verfügt über die nötige Auflösung und Genauigkeit, um geomorphologische Landschaftsformen mit einem Relief von 2 m darzustellen. Es konnten gerade, in einem rhombischen Muster angeordnete Rücken identifiziert werden. Diese Rücken fügen sich in das lokale Störungsmuster ein, was – neben anderen Phänomenen – auf einen neotektonischen Ursprung hinweist. Die Hauptachse des rhombischen Musters verläuft NE-SW und subparallel zur sogenannten Neusiedlersee-Störung, einer Störung die in der früheren Literatur angenommen wird, dessen Existenz jedoch noch nicht eindeutig geklärt wurde. Mit der ALS-Methode konnte die mikro-topographische Spur von Seitenverschiebungen und Abschiebungen analysiert werden.

Die neotektonische Aktivität der Störungen in der kleinen ungarischen Tiefebene wurde weiterführend in der Publikation Zámolyi et al. (Appendix B) durch die Dynamik des Gewässernetzes untersucht. Die Sinuosität alluvialer Flussläufe spiegelt sehr genau ihre Beeinflussung durch Tektonik wider (z.B.: Ouchi, 1985). Der Sinuositätsindex von fünf kleineren Flüssen am westlichen Rand der kleinen ungarischen Tiefebenewurde mit verschiedenen Basislängen berechnet. Diese

Herangehensweise trug der Maßstabsabhängigkeit der Methode Rechnung. Historische Karten (siehe Kapitel 6) wurden für die Rekonstruktion von anthropogen möglichst unbeeinflussten Flussläufen herangezogen. Berechnungen mit allen Basislängen lieferten die im Wesentlichen gleichen Ergebnisse und zeigten den Einfluss von Störungen mit hauptsächlich abschiebender Komponente auf die Flussläufe und somit deren aktuell vor sich gehende Anpassung. Diese Anpassung ist wahrscheinlich an die Reaktivierung miozäner tektonischer Elemente gebunden.

Um die bereits erwähnten Resultate in einen beckenweiten Zusammenhang bringen zu können, wurde auch der östliche Rand der kleinen ungarischen Tiefebene strukturgeologisch untersucht (Appendix D). Das Kernthema der Studie Zámolyi et al. galt der gegenüberstellung der Orientierung von geomorphologisch bedeutenden Störungen versus Störungen ohne zuordenbare Oberflächenformen. Die Ergebnisse zeigten einen kleinen jedoch bedeutenden Unterschied von 10-15° zwischen dem von Uplift betroffenen ungarischen Mittelgebirge (insbesondere dem Bakonygebirge) und den umgebenden Bereichen. Dies kann als selektive Denudation von früheren, vererbten Störungsmustern interpretiert werden. Wenn man die heutige Höhe von fluviatilen Sedimenten auf beiden Seiten der kleinen ungarischen Tiefebene vergleicht, so kann man feststellen, dass die Uplift-Rate auf der östlichen Seite höher war, als auf der westlichen (Kapitel 1, Appendix C).

Im eingereichten Manuskript Zámolyi at al. (Appendix C) wurden die wichtigsten Phasen der Landschaftsentwicklung des westlichen Randes der kleinen ungarischen Tiefebene untersucht. Datierung mit Hilfe der Lumineszenzmethode konnte das Ablagerungsalter der quartären Schotter auf der Parndorfer Platte und dem Flachland des Seewinkels feststellen. Hochfrequente Seeseismik in Verbindung mit Seebohrungen lieferten neue Einblicke in das wichtige Gebiet unter dem Neusiedler See. Es konnte keine durchgehende Schicht von quartären Schottern im See entdeckt werden. So gibt die Interpretation der Daten Auskunft über die Geometrie der obermiozänen (oberpannonen) Schichten: Die stetige Mächtigkeitszunahme der oberpannonen Sedimente beginnt östlich der Hauptachse des Sees. Es konnte kein Mächtigerwerden der Schichten Richtung Nordosten, Richtung der Parndorfer Platte festgestellt werden. Dies weist entweder auf keine größere tektonische Aktivität oder eine postsedimentäre Bewegung entlang einer Störung hin. Dementsprechend und auch den Ergebnissen der Provenanzanalyse gemäß wurden zwei geologische Modelle präsentiert. Das erste Modell folgt der Annahme, dass die quartären Schotter durch locale Gerinne von Süden aus abgelagert wurden und dass sich der Seewinkel im späten Miozän abgesenkt hat. Im zweiten Modell werden die Schotter aus dem Norden von der Donau aus geschüttet und das Gebiet rund um die PArndorfer Platte snkt sich erst im Quartär ab. In beiden Fällen spielt die frühere Entwicklungsgeschichte des Beckens und die Reaktivierung älterer tektonischer Strukturen eine wichtige Rolle in der heutigen Landschaftsentwicklung des westlichen Randes der kleinen ungarischen Tiefebene.

12. Acknowledgements

I would like to thank my supervisors, Dr. Székely Balázs and Univ. Prof. Mag. Dr. Bernhard Grasemann for the extensive discussions and all their support during my work. They shared many ideas and projects with me and helped me to develop my scientific skills. A big thank you to the university staff of Eötvös Loránd University, Budapest, the University of Vienna and the University of Leeds: Dr. Timár Gábor, Dr. Molnár Gábor, Prof. Horváth Ferenc, Dr. Surányi Gergely, PD Mag. Dr. Erich Draganits, Dr. Ulrike Exner, Dr. Darko Spahic, Prof. Greg Housemann and a lot more with whom I could work together and who supported me. I want to mention one key moment as an example: The geophysical-geological field practice in the Lake Neusiedl region in 2008 with the cooperation of the three universities. To my knowledge an undertaking of this scale has been very rare in this area between Hungary and Austria and I feel very lucky to have been able to participate in the organization and the field work itself. I also want to thank the students and graduates of these three universities for all the fruitful field work, lectures and scientific (and less scientific) discussions and who I really enjoyed working with!

A special thank you I would like to say to Dr. Kurt Decker who provided me with a stable project background throughout my PhD studies at the University of Vienna. He and the Karpatian Tectonics team, Monika Hölzel, Andreas Beidinger, Eun Young Lee, Maria Hoprich – they were always helping me with ideas and discussions in the field and back at the university.

I am very thankful for the work and together with Bernhard Salcher, Michael Wagreich, Stefi Neuhuber and Markus Schwab. I am much obliged to my colleagues at work at OMV who were aware of my aspirations: They were all encouraging me to keep going with the thesis in my free time next to the demanding tasks and challenges facing me at work.

Last but not least I would like to thank my parents, my brother and sister who kept encouraging me and provided me a lot of help to work towards finishing my thesis.

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14. Appendix A

Székely, B., Zámolyi, A., Draganits, E., Briese, C. 2009. Geomorphic expression of neotectonic activity in a low relief area in an Airborne Laser Scanning DTM: A case study of the Little Hungarian Plain (Pannonian Basin). Tectonophysics, 474, 353-366.

Tectonophysics 474 (2009) 353–366

Contents lists available at ScienceDirect

Tectonophysics

journal homepage: www.elsevier.com/locate/tecto

Geomorphic expression of neotectonic activity in a low relief area in an Airborne Laser Scanning DTM: A case study of the Little Hungarian Plain (Pannonian Basin)

Balázs Székely [a,b,*], András Zámolyi [b,c], Erich Draganits [d], Christian Briese [a]

[a] Vienna University of Technology, Christian Doppler Laboratory for Spatial Data from Laser Scanning and Remote Sensing, Institute of Photogrammetry and Remote Sensing, Gusshausstr. 27-29, A-1040 Vienna, Austria

[b] Space Research Group, Department of Geophysics and Space Science, Institute of Geography and Earth Sciences, Pázmány Péter sétány 1/A, H-1117 Budapest, Hungary

[c] University of Vienna, Department of Geodynamics and Sedimentology, Althanstrasse 14, A-1090 Vienna, Austria

[d] Vienna University of Technology, Institute for Engineering Geology, Karlsplatz 13/203, A-1040 Vienna, Austria

ARTICLE INFO

Article history:
Received 28 February 2008
Received in revised form 26 September 2008
Accepted 21 November 2008
Available online 6 December 2008

Keywords:
Mur–Mürz–Leitha–Žilina fault system
Neusiedlersee Fault
Airborne Laser Scanning
Neotectonic activity
DTM
Pannonian Basin
Tectonic geomorphology
Deformation bands

ABSTRACT

The NW corner of the Little Hungarian Plain, which lies at the junction of the Eastern Alps, the Pannonian Basin and the Western Carpathians, is a neotectonically active region linking the extrusional tectonics of the Eastern Alps with the partly subsiding Little Hungarian Plain. The on-going deformation is verified by the earthquake activity in the region. An extremely flat part of the area, east of Neusiedlersee, the so-called Seewinkel, has been investigated with Airborne Laser Scanning (ALS, also known as airborne LiDAR) techniques, resulting in a digital terrain model (DTM) with a 1 m grid resolution and vertical precision of better than 10 cm. The DTM has been compared with known and inferred neotectonic features.
Potential neotectonic structures of the DTM have been evaluated, together with geological maps, regional tectono-geomorphic studies, geophysical data, earthquake foci, as well as geomorphological features and the Quaternary sediment thickness values of the Seewinkel and the adjacent Parndorfer plateau. A combined evaluation of these data allows several tectonic features with a relief of <2 m to be recognized in the DTM. The length of these linear geomorphological structures ranges from several hundred meters up to several kilometers. The most prominent feature forms a 15 km long, linear, 2 m high NE–SW trending ridge with gravel occurrences having an average grain size of ca. 5 cm on its top. We conclude this feature to represent the surface expression of the previously recognized Mönchhof Fault. In general, this multi-disciplinary case study shows that ALS DTMs are extremely important for tectono-geomorphic investigations, as they can detect and accurately locate neotectonic structures, especially in low-relief areas.

1. Introduction

Neotectonic evaluation of low relief areas, such as alluvial plains, are typically based on seismic profiles, because of the lack of outcrops. The active status of detected faults characterized by low, but present seismicity is an important issue for the safety of vulnerable industrial facilities, including hydroelectric and nuclear power plants.

The determination of the fault pattern helps to recognize principally uplifting and subsiding structural units, which is extremely important for long-term prognosis of sites of denudation in low-relief areas. The activity of the majority of the faults in alluvial regions cannot be proven by seismic studies, because the topmost part of seismic sections is typically muted. However, it is known that almost all active faults have a topographic expression (e.g., Persaud and Pfiffner, 2004), although normally it is difficult to observe this in the field, especially in densely populated alluvial plains with strong anthropogenic surface modifications. On topographic maps, these features are sometimes present, but the interval between the contour lines is typically spaced too far, making it hard to identify them. Digital terrain models (DTMs) may help, but their resolution is often not enough to allow satisfactory results.

Airborne Laser Scanning (ALS, also known as airborne LiDAR) is a relatively new and promising technology that is capable of providing very high quality DTMs with a vertical precision of better than 10 cm, at which scale active faulting can be detected, even in low relief areas (Cunningham et al., 2006). A further advantage of ALS technology is that DTMs can also be determined in vegetated areas if the leaves do not cover the ground fully, allowing some laser beams to reach the ground surface. Consequently, tectonic faults can be detected in these regions as well (Prentice et al., 2003).

To locate possible faults in the Seewinkel, a meadow and vineyard area east of Neusiedlersee (Lake Neusiedl or, in Hungarian, Fertő-tó), we analyzed an ALS DTM (Attwenger and Chlaupek, 2006; Attwenger et al., 2006) with a grid resolution of 1 m and vertical precision of

* Corresponding author. Vienna University of Technology, Christian Doppler Laboratory for Spatial Data from Laser Scanning and Remote Sensing, Institute of Photogrammetry and Remote Sensing, Gusshausstr. 27-29, A-1040 Vienna, Austria. Fax: +43 1 58801 12299.
E-mail address: balazs.szekely@ipf.tuwien.ac.at (B. Székely).

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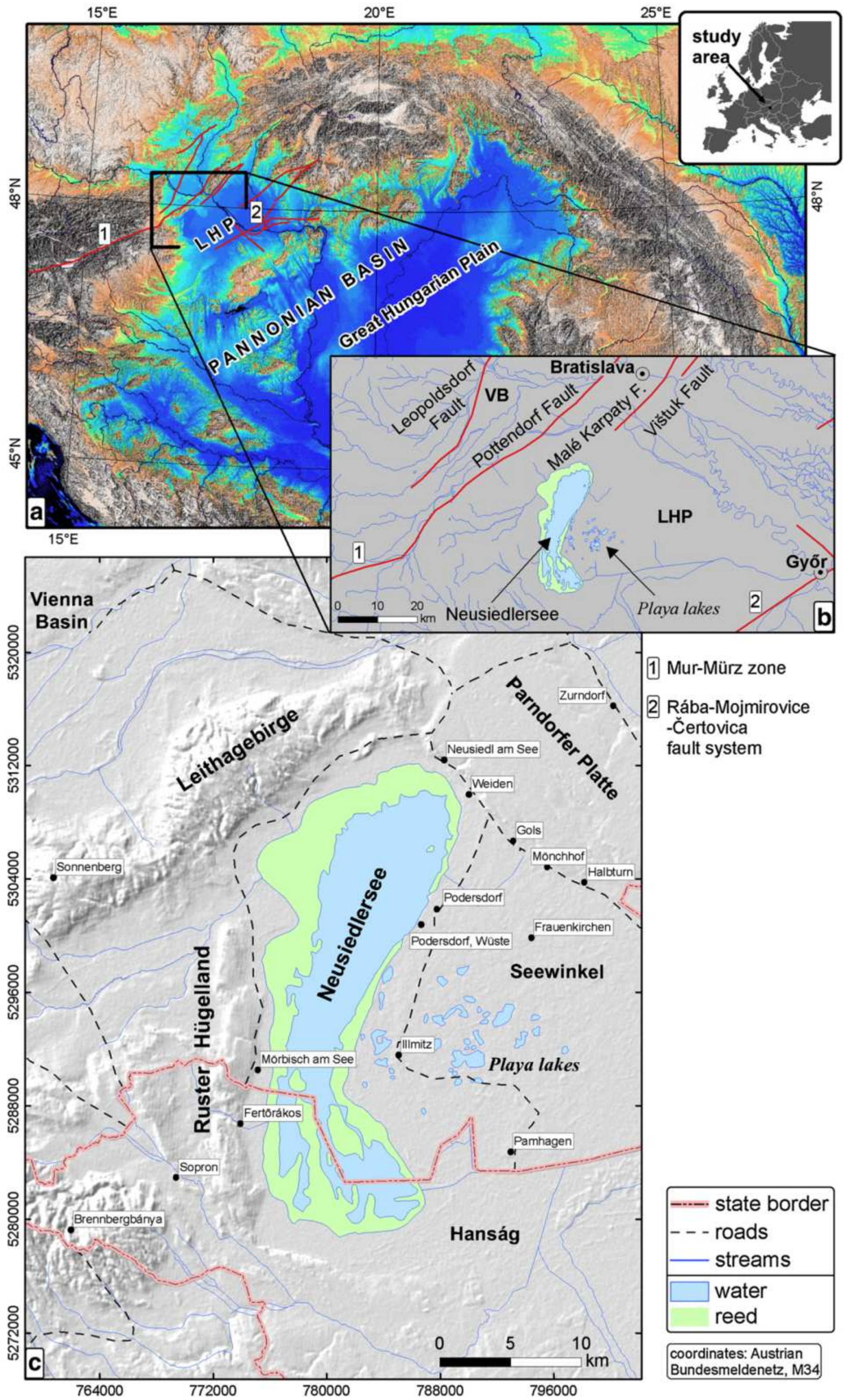
doi:10.1016/j.tecto.2008.11.024



15°E
20°E
25°E
study area
48°N
45°N
1
2
LHP
PANNONIAN BASIN
Great Hungarian Plain
a
Bratislava
Leopoldsdorf Fault
VB
Pottendorf Fault
Malé Karpaty F.
Vištuk Fault
LHP
Győr
Neusiedlersee
Playa lakes
0 10 20 km
b
Vienna Basin
Leithagebirge
Parndorfer Platte
Zurndorf
Neusiedl am See
Weiden
Gols
Mönchhof
Halbturn
Sonnenberg
Podersdorf
Podersdorf, Wüste
Frauenkirchen
Seewinkel
Neusiedlersee
Ruster Hügelland
Illmitz
Playa lakes
Mörbisch am See
Fertőrákos
Pamhagen
Sopron
Brennbergbánya
Hanság
5320000
5312000
5304000
5296000
5288000
5280000
5272000
764000
772000
780000
788000
796000
0 5 10 km
c
1 Mur-Mürz zone
2 Rába-Mojmirovice -Čertovica fault system
state border
roads
streams
water
reed
coordinates: Austrian Bundesmeldenetz, M34

better than 10 cm. The data were integrated with existing earthquake foci, geological maps, geomorphological investigations and geophysical data.

2. Geodynamic and geomorphic setting

2.1. Geodynamic environment

Due to its special geodynamic situation, the Pannonian Basin has been studied intensively in recent years (Gerner et al., 1999; Szafián et al., 1999; Kováč et al., 2002; Sperner et al., 2002; Fodor et al., 2005; Fodor, 2006; Bada et al., 2007; Lenhardt et al., 2007). Most studies mainly focussed on two aspects: large-scale consequences of the Alpine–Carpathian orogeny on the basin evolution and, at smaller scales, the effects of post-Miocene basin inversion. The post-Miocene and neotectonic deformation pattern was very inhomogeneous in both space and time, with some areas subsiding while others, despite their relatively low relief, seem to have been uplifted at unexpectedly high rates (Bada et al., 2007).

These very localized differential vertical crustal movements and the complex tectonic relationship between two extremely low relief areas, the Great Hungarian Plain (GHP) and the Little Hungarian Plain (LHP), separated by the rapidly uplifting (ca. 1 mm/year; Joó, 1992) hilly Transdanubian Range (TR), recommend the area for the application of ALS data for recognising tectonic structures and their interpretation (Fig. 1). Although the geomorphologies of the GHP and LHP seem to be similar, their post-Miocene evolution clearly indicate that the surface processes forming their landscapes were different; the tectonic regimes and the sedimentary environments of the two areas differed considerably during most of the last few million years.

The study area lies in the Neusiedlersee region, situated exactly at the boundary of three important European geodynamic domains; the Eastern Alps, the Western Carpathians and the Pannonian Basin (Fig. 1). This location, lying in the center of the ALCAPA region (Neubauer et al., 1997; Fodor, 2006), makes it an important site for investigating the interaction of Alpine (Tollmann, 1976; Peresson and Decker, 1997; Linzer et al., 2002) and Carpathian (Csontos et al., 1992; Sperner et al., 2002; Plašienka, 2003; Fodor et al., 2005) collisional tectonics, lateral extrusion (Ratschbacher et al., 1991), which essentially ended in the late Pannonian, the related pull-apart formation of the Vienna Basin (Decker, 1996; Decker et al., 2005) as well as the evolution of the Pannonian Basin (Royden, 1988; Fodor et al., 2005; Horváth et al., 2006), including the subsidence of the LHP (Höggerl, 1980; Joó, 1992).

The LHP, a sub-basin of the Pannonian Basin (Horváth, 1995; Lankreijer et al., 1995) has had a complex geological history since the early Paleogene. After the onset of the second major continental collision phase of the Eastern Alps, during Eocene–Oligocene times, the area east of the convergence zone was dominated by extension. Roll-back of the subducted oceanic slab beneath the European foreland provided a weak boundary, allowing the eastward lateral extrusion of the Eastern Alps in the Early Miocene (Ratschbacher et al., 1991; Linzer et al., 1997), enhanced by the gravitational collapse of the overthickened orogenic crust (Horváth, 1995). The evolution of the extensional sub-basins (e.g. LHP) started in the early Miocene along fault systems with a mainly normal component (Szafián et al., 1999). The northeastward movement of lithosphere in the Pannonian Basin and the linked Carpathian mountain belt stopped at the Tornquist–Teisseyre zone, triggering a new, compressive deformation phase which is also referred to as the inversion of the Pannonian Basin (Fodor, 2006). In contrast to adjacent areas, the LHP preserved its basin characteristics after this convergent deformation phase (Horváth and Cloetingh, 1996; Fodor et al., 2005). However, the major fault systems partly underwent a left-lateral strike-slip dominated reactivation. Recent measurements through the Central European GPS Geodynamic Reference Network indicate a movement of the Alpine and North Pannonian units towards the east (Grenerczy et al., 2005).

The study area lies between two major structural features that have been traced in seismic sections and earthquake activity: the Rába–Mojmirovice–Čertovica fault system and the Vienna Basin transform faults (Lenhardt et al., 2007; Figs. 1 and 2). The latter includes the Mur–Mürz–Leitha–Žilina fault system (MMLZ, also known as Mur–Mürz–Žilina, Mur–Mürz–Leitha [–Little Carpathian or –Malé Karpaty] fault; Reinecker and Lenhardt, 1999; Székely et al., 2002; Bada et al., 2007). At the westernmost margin of the LHP, predominantly left-lateral deformation has occurred from post-Miocene times up to the recent (e.g., Reinecker and Lenhardt, 1999; Bada et al., 2007). Whereas the MMLZ in the Alpine region was confined to a rather narrow zone, displacement in the foreland of the Leithagebirge has been accommodated by faults distributed in a 15–20 km wide zone. However, the exact position of these faults is poorly known because of the low relief (typically below 20 m over some 10 km radius). Fault activity since at least Miocene times is clearly indicated by structural geological field observations and earthquake focal mechanisms (Decker, 1996; Reinecker and Lenhardt, 1999, Tóth et al., 2002, Lenhardt et al., 2007). The Eastern Alps (an area of strike-slip displacement) are separated by MMLZ from the LHP; the latter has a tendency for reverse faulting since the Pliocene (Bada et al., 2007). However, the detailed neotectonic behavior and location of the faults within the LHP is difficult to decipher. Previous workers have faced considerable problems because of the scarcity and poor quality of the data and the lack of outcrop (e.g. Strausz, 1942). As a result, the evolving concepts of tectonic geomorphology and the development of high-resolution DTMs derived from ALS represent an important advance in the study of neotectonic structures in low relief areas.

Due to the complex fault pattern and their poor exposure, the nomenclature is often inconsistent and sometimes confusing. This is also partly due to Neusiedlersee, which has made seismic surveying difficult until recently (Hodits, 2006). Furthermore, the cold war made comprehensive, across-border surveys impossible. In this paper, where appropriate, we have followed the established nomenclature, but we have added clarifying names where necessary, to avoid confusion (Fig. 2).

Although some small faults, such as the Fertő, Ikva and Répce faults have been identified from subsurface data (Szafián et al., 1999), there are hardly any comprehensive tectonic studies from the Neusiedlersee area. This is in marked contrast to the much better investigated Vienna Basin to the west (Decker et al., 2005 and references cited therein) and the Pannonian Basin to the east (Horváth et al., 2006).

2.2. A review of the previous tectonic studies in the Neusiedlersee area

Early structural observations in the wider Neusiedlersee area originated from the strongly faulted Brennberg coal mines, in the western part of the Sopron Hills. There, NW–SE, as well as broadly E–W striking normal faults were documented (Wolf, 1870). Vendl (1933) applied tectonic observations from the Brennberg area to the Ruster Hügelland and described N–S trending faults. Szádeczky-Kardoss (1938) investigated the tectonics of the Neusiedlersee area and recognized their importance in lake formation. He described a fault along the southwestern boundary of the Parndorfer Platte, a NE–SW striking fault east of the Hackelsberg and a N–S trending fault at Rust. Küpper (1955) mapped NE–SW striking normal faults at the eastern boundaries of the Leithagebirge and Hackelsberg, and later those of the Ruster Hügelland (Küpper, 1957). Tollmann (1955), working in the southeastern part of the Leithagebirge, defined NE–SW trending faults and smaller WNW–ESE trending faults. Tauber (1959c) presented

Fig. 1. Location and overview map (a) of the study area with localities mentioned in the text (c). Red lines (a and b) indicate selected faults of the area according to Lenhardt et al. (2007). LHP: Little Hungarian Plain; VB: Vienna Basin. Background in (a): SRTM DTM (Farr et al., 2007). Coordinates: Austrian Bundesmeldenetz, M34.

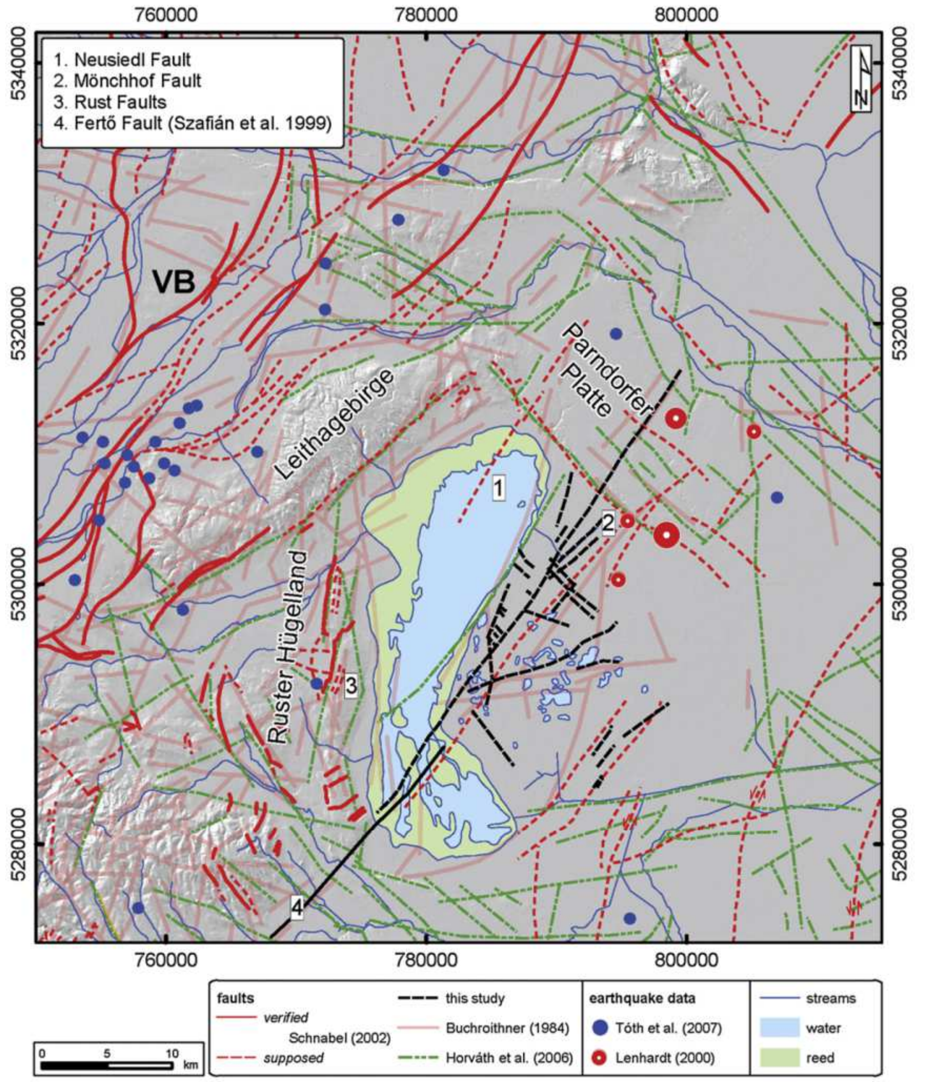


Fig. 2. A summary of lineaments and faults described by previous authors (Buchroitner 1984; Schnabel, 2002; Horváth et al., 2006) together with the earthquake pattern (Lenhardt, 2000; Tóth et al., 2007) and the linear features evaluated in this study. VB: Vienna Basin.

seven sections of the Seewinkel that were interpreted from well data and proposed offsets of 200 m and 100 m for the Neusiedl and Mönchhof faults, respectively. The active phase of the faults was presumed to have been between middle Pannonian and late- to post-glacial times.

The first tectonic overview of the central part of the Alpine–Carpathian–Pannonian region was presented by Küpper (1960). Schmid (1968) found several NE–SW trending faults at the eastern side of the Leithagebirge. Apart from NE–SW trending faults, Herrmann (1973) also proposed NW–SE trending faults. Decker and Peresson (1996) describe a roll-over structure that was interpreted as evidence of Sarmatian/Pannonian synsedimentary extension from a sand pit west of the Ruster Hügelland, although this has recently been questioned by Spahic et al. (2008).

The regional tectonic context and timing of Miocene to present-day deformation phases was summarized by Decker (1996). The main subsidence of the Vienna Basin was related to movements along sinistral strike-slip faults between the Karpatian (17 Ma) and late Pannonian (8 Ma). The main phase of the lateral extrusion ended in the late Pannonian (Decker, 1996).

Tari (1996) used seismic and well data to investigate the tectonic evolution of the NW part of the Pannonian Basin and observed a pattern of broadly NE–SW as well as NW–SE trending faults. In a NW–SE trending section, south of the Neusiedlersee, Tari (1996) and Szafián et al. (1999) both reported three normal faults, the Fertő, Répce and Ikva faults. Due to this normal faulting, the pre-Cenozoic sediments lie more than 3600 m below the surface in the SE part of the investigated area (Kröll et al., 1993). These 2D seismic studies, oriented at high angles to the strike of the faults cannot detect any strike-slip faulting component.

Although the main movement along faults is thought to have occurred during the late Miocene (Tari, 1996), the surface expression in late Pleistocene sediments (e.g. Fodor et al., 2005) as well as seismic activity (Tóth et al., 2002; Lenhardt et al., 2007) implies a late Quaternary reactivation. Similar observations have been described from the Vienna Basin, where moderate seismic activity, subcrop data and geomorphological features indicate that virtually all active structures in the Vienna Basin are reactivated Miocene structures (Decker et al., 2005). GPS data indicate slip rates around 1 mm/year along the strike-slip faults (Grenerczy et al., 2005; Caporali et al., 2008). The dominance of strike-slip movements is supported by the very small thickness (4–7 m) of Quaternary sediments in the Neusiedlersee area (Szontagh, 1904; Tauber, 1959c). In contrast, eastwards from the Seewinkel, in the Győr area, more than 250 m of Quaternary sediments have accumulated (Joó, 1992; Timár and Rácz, 2002).

2.3. Overview of the geomorphological units of the Neusiedlersee area

The present regional topography in the wider Neusiedlersee area is the result of five major processes, partly discussed above: (i) crustal thickening during the Alpine orogeny (Tollmann, 1976; Peresson and Decker, 1997; Linzer et al., 2002), (ii) lateral extrusion of crustal fragments towards the east (Ratschbacher et al., 1991), (iii) formation of the Vienna pull-apart basin by left-lateral strike-slip faults related to lateral extrusion (Decker, 1996; Decker et al., 2005), (iv) normal faulting connected to the formation of the Pannonian Basin (Horváth et al., 2006) and (v) terraces and other relict landforms formed by erosion and sedimentation of the Danube and its tributaries (Szádeczky-Kardoss, 1938; Küpper, 1955; Grill et al., 1968; Häusler et al., 2007). A combination of these geological processes has formed several distinct geomorphological units with characteristic elevation ranges (with reference to level zero at Trieste, Italy), representing important geomorphological features in the low-relief Neusiedlersee area.

a) Leithagebirge (Leitha Mountains) The Leithagebirge is a NE–SW trending, 33 km long, <10 km wide, hilly landscape, separating the Vienna Basin from the Neusiedlersee area. Its elevation ranges between 118 m and 484 m a.s.l. (Sonnenberg), rising abruptly from the surrounding, extremely low-relief areas. The Leithagebirge, which is a NE–SW striking horst (Schnabel, 2002), comprises Lower Austroalpine schists, gneisses and amphibolites, covered by early Mesozoic quartzites and dolomites. The metamorphic rocks are covered at the rim by Langhian to Tortonian clastic sediments and limestones (Pistotnik et al., 1993). The Leithagebirge shows some relict planation surfaces (e.g. Schmid 1968) which are dissected by intermittent and perennial rivulets at high angles to its longitudinal orientation.
b) Ruster Hügelland (Rust Hills) The Ruster Hügelland represents a N–S trending, narrow (22 km long, <3.5 km wide) hilly area (118–283 m a.s.l.) at the western boundary of the Neusiedlersee. It comprises schists, gneisses and amphibolites of the Lower Austroalpine tectonic unit that are almost completely covered by Burdigalian to Tortonian clastic sediments and limestones (Fuchs, 1965; Pistotnik et al., 1993). Analogous to the Leithagebirge, the Ruster Hügelland is also a tectonic horst; probable normal faults at the eastern side, towards Neusiedlersee, have been covered by young sediments (Fuchs, 1965; Küpper, 1957).
c) Neusiedlersee/Fertő-tó (Lake Neusiedl) Neusiedlersee, which lies across the Austro-Hungarian border, is the largest lake in Austria and the second largest steppe lake in Central Europe (285 km^2, including the adjacent reed areas), although the maximum depth is only about 1.8 m (Bácsatyai et al., 1997). Continuous measurements of the lake level have been made since 1932, during which the minimum lake level was 114.50 m a.s.l. (July 1949) and the maximum value reached 116.08 m (May 1941) with a mean elevation of 115.30 m (BYC, 2007). Due to its shallowness and the flat surrounding area, even small variations of the lake level have an impact on large areas. Artificial regulation of the lake level started in the 16th century and has continued with an ever rising intensity into recent time. Water-level variations before the 20th century range between 113.6 m (dried out between 1865 and 1870) to a maximum above ca. 117.7 m, indicated by lake sediments and geomorphologic features (Draganits et al., 2007). The lake basin comprises tectonically slightly tilted and faulted Pannonian clastic sediments (Fuchs and Schreiber, 1985; Hodits, 2006) that are only locally covered by lake sediments (Szontagh, 1904; Tauber, 1959a; Bácsatyai et al., 1997). It is important to note that the lake area lacks the thin cover of Quaternary fluvial gravel above the Pannonian sediments that is characteristic of the Parndorfer Platte, Seewinkel and Hanság (Tauber, 1959a).
d) Parndorfer Platte (Parndorf Plateau) The Parndorfer Platte (130–184 m a.s.l.) represents a fairly even relict surface that rises relatively steeply above the surrounding landscape by about 10–45 m, with the highest relief values between Neusiedl am See and Weiden (Fig. 1). The surface dips gently towards the SE (from 184 m in the NW to 144 m in the SE) corroborating sedimentological observations of imbricated gravel deposits (Szádeczky-Kardoss, 1938). In particular, the SW margin (towards the lake), and also the SE and NE boundaries, are steep and relatively straight, suggesting a tectonic origin (Schnabel, 2002). Noticeable features are systems of strikingly straight, NW–SE trending dry valleys in the southeastern part of the paleosurface (Fig. 3). Generally, the plateau consists of Pannonian freshwater sands and minor gravels, overlain by a thin layer of younger Danube gravel that only locally may reach up to 20 m thickness (Tauber, 1959c; Lipiarski et al., 2001). Due to the high infiltration capacity of the gravel cover, the plateau hardly shows any surface runoff. The geological map associates these gravels with the Günz, Mindel and Riss glaciations (Fuchs et al., 1985); however, this correlation lacks unequivocal geochronological data and should be treated cautiously.
e) Seewinkel The Seewinkel is bounded by Neusiedlersee to the west, by the Parndorfer Platte to the northeast and by the Hanság to the south. The Seewinkel is a very flat area, with elevations ranging between 118 and 130 m a.s.l. Similar to the Parndorfer Platte, the Seewinkel consists of slightly tilted and faulted relatively fine-grained, clastic, Pannonian sediments (Tauber, 1959c; Fuchs and Schreiber, 1985), covered by a thin layer of Quaternary fluvial gravel from the Danube (Szádeczky-Kardoss, 1938). The gravels thin out towards the northwest, almost completely disappearing below Neusiedlersee, but increase towards the southeast, reaching a thickness of about 20 m at the Austro-Hungarian border (Tauber, 1959a). Additionally to the fluvial gravel and sand, this area comprises loess, as well as mud from episodic shallow lakes (Tauber, 1959b; Husz, 1965). Relatively low precipitation (<600 mm/a) in combination with high evaporation rates during hot, continental summers result in locally high concentrations of salt in the episodic shallow lakes and surroundings.
f) Hanság/Waasen The Hanság is a low-lying, extremely flat area bounded by the Seewinkel to the north, by Neusiedlersee to the west and by the Ikva and Répce rivers to the south. In the investigated area, the altitude lies below 118 m a.s.l. Before it was drained by channels and dams, it was part of Neusiedlersee, forming an extensive swamp (Draganits et al., 2007). Today, the area is mainly dry. The peat on top of fluvial gravels reaches thicknesses of up to 2 m.

3. Data and methods

3.1. ALS: data acquisition and processing

The increasing availability of Airborne Laser Scanning (ALS), an active remote sensing technique, has revolutionized the field of topographic surveying because it allows a very dense and accurate sampling of the landscape (Kraus, 2007). ALS utilizes a narrow laser beam for a high frequency range determination to illuminate the surface of the objects, and the back-scattered signal is measured in various ways (cf. Wehr and Lohr, 1999; Wagner et al., 2004, 2006; Doneus and Briese, 2006; Pfeifer and Briese, 2007). In order to allow an extensive acquisition of the topography, the laser beam is continuously deflected across the flight path such that the landscape is acquired in strips. ALS strips are typically flown with an overlap, to guarantee complete coverage of the area.

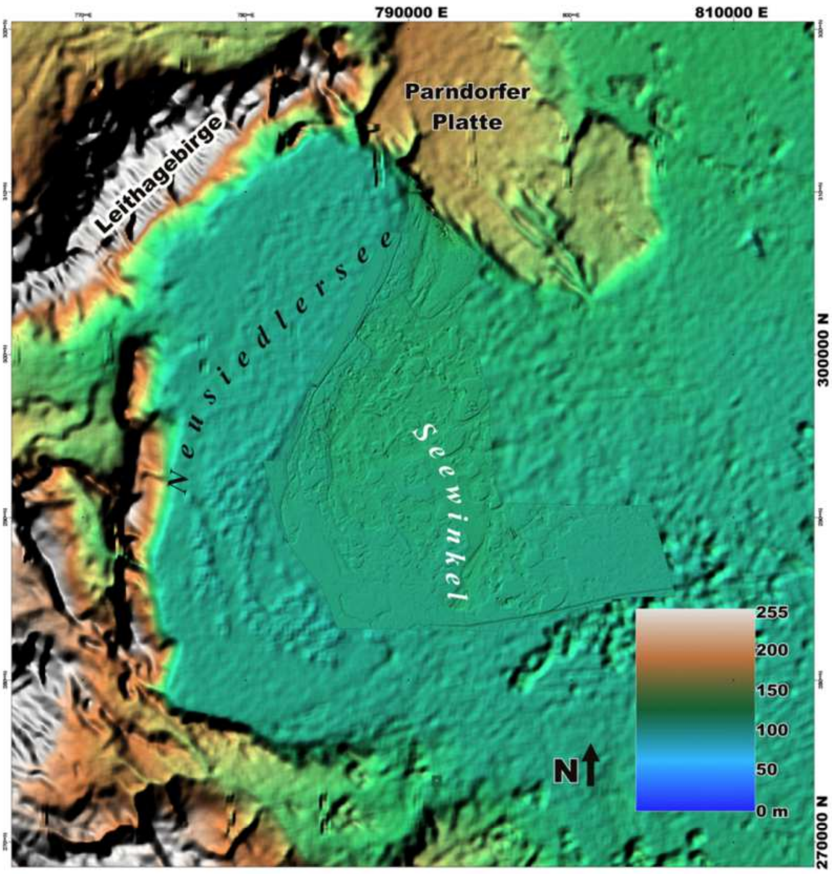


Fig. 3. A combination of SRTM DTM (Farr et al., 2007) and ALS DTM (Attwenger and Chlaupek 2006; Attwenger et al., 2006) colour coded between 0 and 255 m a.s.l. including shading. Note the parallel dry valleys in the Parndorfer Platte and the elongated features in Seewinkel. See further discussion in the text.

Based on the range measurement, the angle of deflection and the observation of the position and orientation of the aircraft, ALS allows a direct 3D determination of points on the object surface within one coordinate frame. The position and orientation of the aircraft is measured by a Position and Orientation System (POS) that allows the direct georeferencing of the dynamically moving sensor platform. The POS typically consists of a Global Navigation Satellite System (GNSS) receiver and an Inertial Measurement Unit (IMU). An improvement of the georeferencing of the data can be made using strip adjustment techniques (cf. Burman, 2000; Filin, 2002; Kager, 2004). These use the redundant observations present within the overlapping regions of the ALS strips and external pass information in order to improve the georeferencing (relative and absolute) of the ALS data.

The typical point density of current ALS systems operating at a flying height of 500 m to 1500 m is 1 point/m^2 or denser. In our survey, carried out in November 2004, the average point density lies well above 1 point/m^2, although in a few places the actual density fell to 0.9 point/m^2 (Attwenger and Chlaupek, 2006; Attwenger et al., 2006). The applied ALS sensor utilized a pulsed laser for the range determination and the first and last echo per emitted laser pulse has been determined.

3.2. Data accuracy

During ALS data acquisition, neither interpretation nor classification of the recorded echoes is performed. Consequently, the resulting point cloud represents reflections from different object surfaces, such as the terrain, building roofs and the vegetation. Based on the first reflections per emitted laser beam (first echo points) a digital surface model (DSM) can be determined representing the top level surface of all the objects present during data acquisition. However, for a detailed analysis of the terrain surface a DTM, describing the bare earth surface, is required. Even though ALS systems can distinguish between different echoes at different ranges, the resulting last echo point cloud still includes echoes from off-terrain objects. Thus the last echo points have to be classified into terrain and off-terrain points by "filtering" the data, for which several methods have been developed (cf. Sithole and Vosselman, 2004).

For high-accuracy DTM generation it is also essential that the georeferencing accuracy of the range measurements is maximised, prior to the determination of the surface models. Consequently, the direct georeferencing using the observations from the POS is followed by strip adjustment (cf. Burman, 2000; Filin, 2002; Kager, 2004). This process allows to check and improve the precision (relative accuracy) as well as the accuracy of the data. Subsequently, with the refined ALS data, the surface model generation can be performed. The typical resulting horizontal and vertical precision of the resulting surface models is 0.3 and 0.1 m, respectively.

In our study, these processing steps were all performed and a DSM was derived from the first echo dataset. For the generation of the DTM, the software package SCOP++ (SCOP++, 2007) was used to filter the data (i.e. the classification of the last echo data into terrain and off-terrain points) and to interpolate a DTM. Furthermore, an older ALS DTM, from 1999, was merged with the 2004 ALS DTM. The final DTM was derived with a horizontal resolution of 1 m. The absolute height accuracy is ca. 0.1 to 0.2 m, while the relative height accuracy (vertical positions relative to other points in the vicinity) is better than 0.1 m.

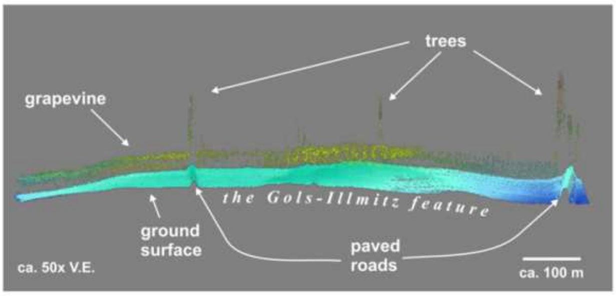


Fig. 4. 3D visualization of first echo point cloud of a small E–W strip north of Illmitz. The vertical exaggeration (V.E.) is ca. 50. The vegetation (here: grapevine) can be separated from the points of the ground. The positive topographic form is the Gols–Illmitz feature. Compare the relief of the ground to the height of the trees. The "narrow" positive ridges are paved roads. The profile is taken near to the standing point of Fig. 7a. The view direction is subparallel to the axis of the profile.

From a geomorphological viewpoint, the natural surface has been reconstructed and represented quite well by the DTM, although only data from a discrete echo ALS sensor system with limited first and last echo recording capability was available. As Fig. 4 demonstrates, reflections from the bare earth can be clearly distinguished from signals from the vegetation.

3.3. DTM data integration

To make the detected features comparable with other geomorphological characteristics of the area around Seewinkel, we integrated several other digital elevation data sets with our ALS-derived DTM. Three different data sets were used: (1) the 10 m resolution DTM of the Austrian National Mapping Authority (Bundesamt für Eich- und Vermessungswesen; BEV) created from stereo aerial photographs, (2) a small part of the DDM10, a 10 m resolution DTM of Honvéd Térképészeti KHT, Hungary, derived from contour lines of 1:25,000 topographic maps, and (3) SRTM (Farr et al., 2007) as an overview. These have been integrated into GIS for visualization and map generation purposes. Generation of the profiles used was based exclusively on the ALS DTM, because of its far higher accuracy.

4. Evaluation of ALS data and synthesis

4.1. Observed geomorphological features in the ALS DTM

The very low relief topography of the Seewinkel area is characterized by two types of features: (1) shallow depressions and (2) elongated ridges a few meters high (Fig. 5). Some of the depressions host very shallow, partly perennial and partly ephemeral playa lakes (Fig. 1). In the south, the depressions are considerably shallower than their north-eastern counterparts. The depressions, which vary from tens of meters to several hundred meters in length, typically have rounded or oval forms, although elongate depressions also occur. Some depressions are connected by reaches or artificial channels, although watercourses are very rare over the whole area.

The depressions are spatially rather scattered, without any obvious pattern. Riedl (1965) suggested that the shallow lakes formed as relict pingos, a common periglacial landform (French, 2007). A periglacial interpretation is supported by ice-wedge casts and other soft-sediment deformation structures in gravel pits in the Seewinkel (Riedl, 1965). Additionally, the artesian ground water conditions, their elevated salt contents and the existence of faults in the Seewinkel area (Tauber et al., 1958; Tauber, 1959c; Husz, 1965) are thought to support the formation of hydrologically open pingos (Weise, 1983). Alternatively, the depressions may be former thaw lakes that formed by thermokarst processes in a periglacial environment (French, 2007).

The other features of this area are elongate ridges a few meters high (Figs. 5 and 6) in two main orientations, hereafter descriptively referred to as 'long ridges' and 'short ridges' to avoid yet any implications on their formation or composition. Long ridges (from 1 to 15 km in length) are NE–SW or NNE–SSW oriented, whilst short ones are typically NW–SE directed. The latter, though often shorter than 1 km, sometimes seem to be spatially associated as if they are dissected parts of previously continuous, longer ridges. The long ridges are almost never cut by playa lakes or other depressions, while the shorter ones are broken up by closed, often dry depressions. In the area where the playa lakes are abundant, the ridges are rare; but if present, the short type occurs (Figs. 5 and 6).

One of the longest ridges is a Y-shaped, bifurcating positive form consisting of several branches of ridges splitting towards the north (Fig. 6). This feature extends from north of Illmitz to Gols and to Mönchhof, separating the area of playa lakes into eastern and western parts. This ridge is here called the Gols–Illmitz feature.

Both ridges types have widths of 300–500 m. Long ridges are relatively widely spaced, sometimes forming an undulating surface, while short ridges may be closely spaced and hence sometimes form small elevated areas (Figs. 5 and 6). The village of Podersdorf is situated on such an elevated small plateau.

In the field, it is difficult to document their actual extent and they are scarcely recognizable on topographic maps. However, long straight linear features, such as roads and vineyard rows enhances their optical appearance (Figs. 7 and 8). Their smooth surface is partly due to modern agriculture, although historical reports and maps, from when the area was used as pasture, indicate that the geomorphology was very similar then (see historical maps in Timár et al., 2006).

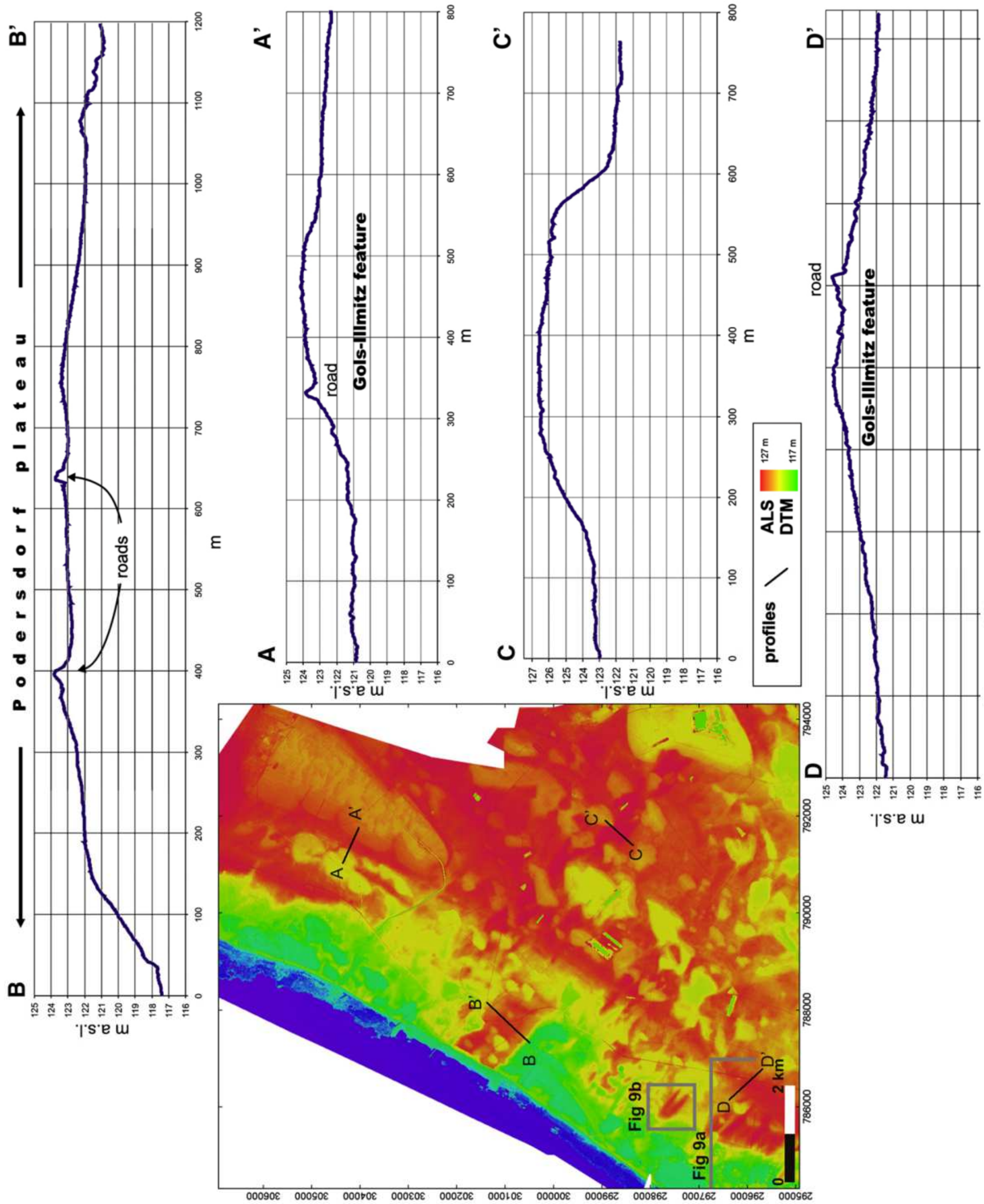
4.2. Field observations

Field work was carried out to characterize the ridge structure and composition. Attention was paid to the geomorphological appearance of the lineaments mapped in the ALS DTM, as well as noting the distribution of surface clastic material. As the area is very poorly exposed, house construction pits and animal holes provided the best opportunities to study the ridge material. No real outcrops, from which structural observations could be made, occur near the ridges. However, ploughing to depths of up to 50 cm allowed the observation of changes in the grain size of the topmost layer. At places, abundant gravelly material covers the vineyards while fine-grained material was excavated from animal holes a few hundred meters away.

Macroscopic observations indicate that gravels are abundant along the ridge crests and occur less frequently in lower level areas (Fig. 9). Generally, the gravels are polymict, with both siliciclastic and carbonate clasts.

4.3. Auxiliary data

Although extensive geophysical measurements have been made in the area during the last decade, most of the data are not publicly available. The gravity field shows a distinct NE–SW trending positive Bouguer anomaly of 4–6 mgal at the eastern margin of the Neusiedlersee, NE of Podersdorf (Zych et al., 1993). Figdor and Roch (1984) measured a 1 km long, NW–SE trending gravimetric profile SW of Frauenkirchen (approximately 16°52′09.49″, 47°49′36.75″, 118 m). Kohlbeck et al. (1993, 1994, 2000) and Szarka et al. (2003) provided maps and profiles of several geoelectric and VLF measurement campaigns. More recently, a high-frequency lake seismic survey, conducted in cooperation between the University of Vienna (H. Häusler) and the Eötvös University, Budapest (T. Tóth), provided important new insights on the continuation of the Neusiedlersee Fault within the lake (Hodits, 2006). The thickness of Quaternary sediments on the Parndorfer Platte was derived from aerial geophysical measurements (Lipiarski et al., 2001).


B
B’
P o d e r s d o r f   p l a t e a u
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A
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DTM
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117 m
Fig 9a
Fig 9b
2 km

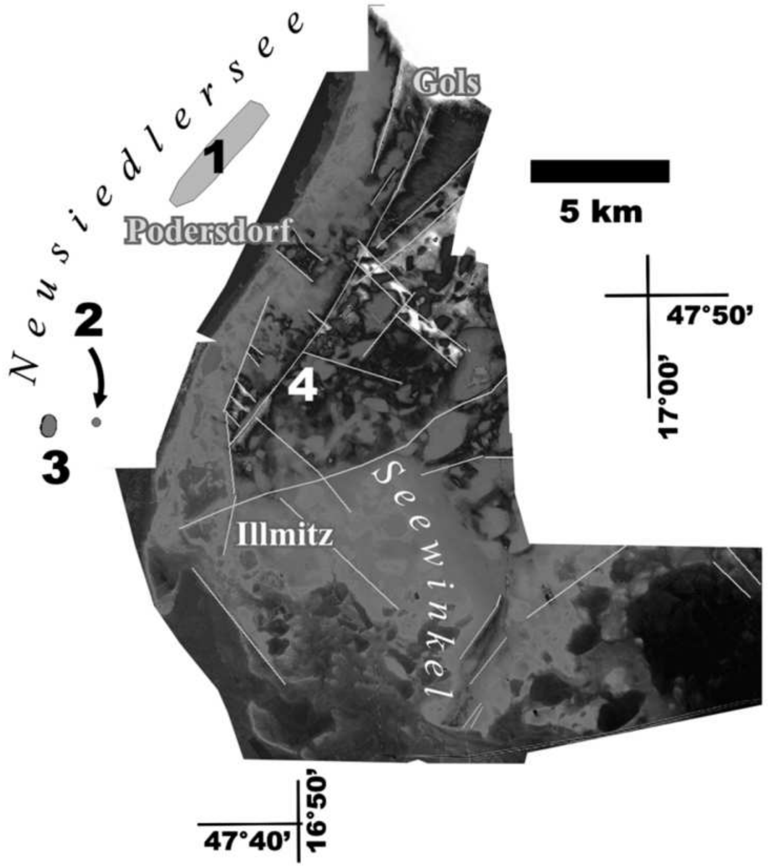

Fig. 6. Digitally enhanced image of the ALS DTM of Seewinkel with the identified linear features. Numbered features: 1: Geoelectric anomaly found by Szarka et al. (2003) in the lake; 2: Schotterinsel; 3: Fünfschoppen; 4: Gols–Illmitz feature. See discussion in the text (localities Gols, Illmitz and Podersdorf are also indicated for better orientation).

Bácsatyai et al. (1997) conducted a detailed geodetic survey and constructed a digital bathymetric surface of the lake bottom. Reinecker & Lenhardt (1999) and Lenhardt et al. (2007) provided a catalog of earthquake data in the Eastern Alps–Pannonian–Carpathian region from 1267 to 2004; Lenhardt (2000) gave a description of the relevant earthquakes in the Seewinkel region.

5. Discussion

Although an anthropogenic origin could be considered for the elongated topographic features visible in the ALS DTM, it is clear that they are far too large and linear (up to 15 km long, 500 m wide and 2 m high) to be man-made. Volumetrically it would be equivalent to the full load of 500,000 large trucks that makes this interpretation unlikely.

In winter, the predominant NW-oriented wind can push ice, causing it to pile-up sediments along the eastern shore-line, forming an ice-pushed ridge (*Seedamm*; Bernhauser, 1962), usually comprising sand of variable grain sizes, although sandy gravels are also common (Bernhauser, 1962; Husz, 1965). An analogous origin for some of the shorter NE–SW trending features has been proposed, assuming higher lake levels in Holocene times (Bernhauser, 1962). However, the overall shape and extremely linear character of the ridges compared to very recent ice-push ridges east of Neusiedlersee (Bernhauser, 1962) makes this unlikely.

Selective erosion by water or wind could contribute to ridge-formation. However, fluvial erosion can be excluded, because of the exceptionally low density of rivers and the abundant playa lakes. Eolian selective denudation by deflation of the fine-grain fraction of the sediment, leaving mainly gravel behind, may have occurred. The course of the longest and most prominent ridge is oriented almost normal and almost parallel to the two prevailing wind directions (NW and NE; Dobesch and Neuwirth, 1979). Despite this, the linear features lie in an angle of at least 100–110° to the typical orientation of topographic features in the Transdanubian area (e.g. Gerner et al., 1999) thought to have been formed by the prevailing Pleistocene katabatic wind direction funneling in the nearby water gap of the river Danube between the Little Carpathians and the Leitha Mts (see Gerner et al., 1999 for references). As the ridges in the Seewinkel have high gravel contents, they cannot be sand dunes.

The spatial coincidence of different phenomena, including sedimentological, seismological, geophysical and geomorphic observations argue for a neotectonic explanation.

5.1. Orientation of faults

Classic lineament mapping, based on satellite remote sensing data (Buchroithner 1984; Fig. 2), reveals NNE–SSW and NW–SE orientations in the area. Recently, Horváth et al. (2006) compiled a series of

Fig. 5. Selected profile locations in the colour-coded ALS DTM. The roads are clearly distinguishable from the ridges (here the Gols–Illmitz feature); the Podersdorf plateau shows different characteristics compared to the profile of the Gols–Illmitz feature. The profile shapes of the latter object changes along the lineament as well. The positions of the two panels of Fig. 9 are also indicated.

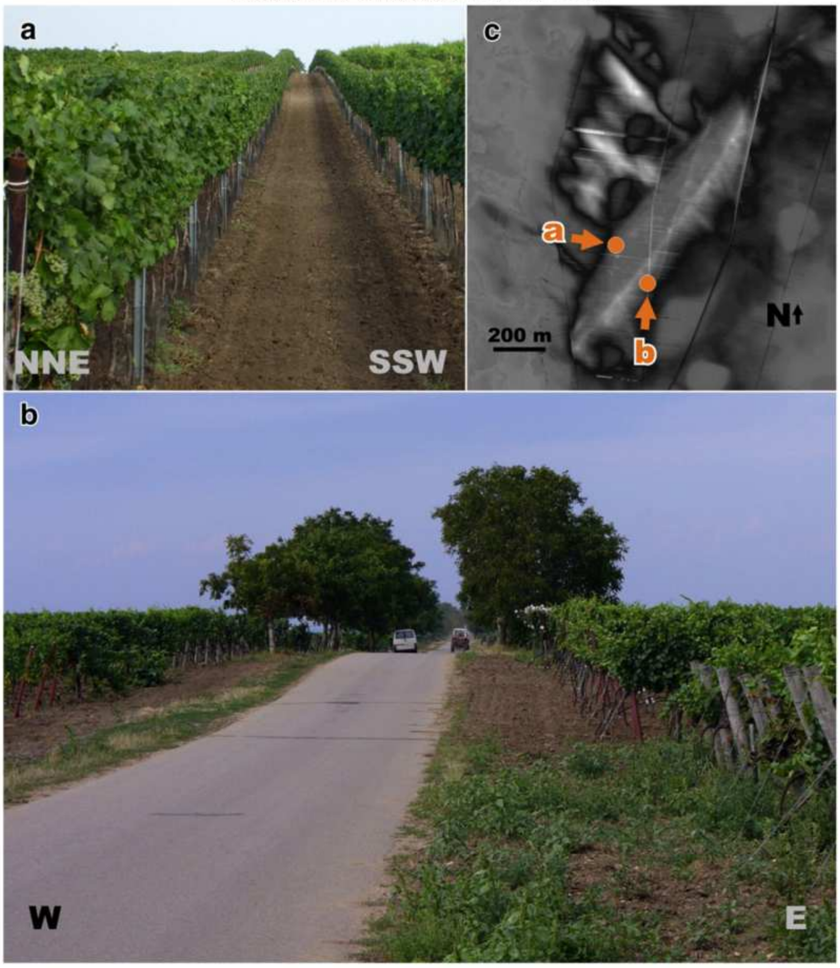

Fig. 7. Field observations of the Gols–Illmitz feature. (a) The vineyard rows enhance the positive topographic form. (b) An additional relief element is enhanced by the paved road (vehicles provide scale). (c) Sophisticated processing of the ALS DTM to enhance the section of the Gols–Illmitz feature: whitish bright shades indicate the positive form. Orange dots indicate the standing points and the view directions (orange arrows) of panels (a) and (b). Note the roads (thin linear features) and the ladder-shaped ridges to the W.

geodynamic maps of the Pannonian Basin and surrounding areas, including a map of morphostructural elements derived from global SRTM data (Farr et al., 2007). Despite the low relief of the area, the same lineament orientations were seen in the two interpretations. By analogy with structural observation in the former Brennberg lignite mines, SW of the Neusiedlersee, conjugate extensional faults, giving horst and graben structures (Vendl, 1933), as well as book-shelf-type extensional faults were geophysically recorded in Neusiedlersee (Hodits, 2006). Strike-slip faults, which are common in the western part of the LHP (Lenhardt et al., 2007), may also be expected in the area.

The lineament pattern shows several tectonic, and some non-tectonic directions, characterized by various ages. The majority of the mapped lineaments in the ALS DTM are NE–SW oriented, parallel to the MMLZ and are probably related to on-going Alpine lateral escape (Ratschbacher et al., 1991). This model is supported by earthquake foci data (Lenhardt, 2000; Tóth et al., 2002; Lenhardt et al., 2007).

Directions present within the feature-group mapped in the ALS DTM coincide with directions of tectonic features that have been observed in the Pannonian Basin in similar low relief areas, such as blind reverse faults, transpressional strike-slip faults and normal faults (Fodor et al., 2005).

In his aforementioned study, based on well data, Tauber (1959c) recognized two NE–SW trending faults in the Seewinkel area. According to his sections, the Neusiedlersee Fault dips at only 20–25° towards the southeast and the Mönchhof Fault, with a dip of 20–40°, is only slightly steeper. The significant difference in depth of equivalent, essentially horizontal Pannonian strata in the Podersdorf 1 and Podersdorf 2 boreholes (Fuchs and Schreiber, 1985), lying to the west and east of the Mönchhof Fault, may also indicate a normal faulting component in a generally left-lateral strike-slip regime.

A resistivity map by Fritsch and Tauber (1963) shows a NE–SW trending anomaly, parallel to the Mönchhof Fault, but slightly towards the NW, in a similar position as the lineaments (especially parts of the Gols–Illmitz feature) indicated by the ALS data. The increased resistivity was attributed to the flow of ground water with low salt contents along faults. Fodor et al. (2005) also recognize a left-lateral strike-slip fault in the area of the Mönchhof Fault. However, due to the lack of outcrop and the low-relief geomorphology of this structure, its exact location varies considerably according to different authors (Fig. 2).

Tollmann (1985) also inferred the Neusiedlersee Fault as a major structural feature, running parallel to the NW-rim of the Parndorfer Platte, from Neusiedl am See to Mörbisch am See. The exact location of the fault was poorly constrained before Hodits (2006) used lake seismic data to confine its position, verifying the direction proposed by Tollmann (1985) and Schnabel (2002). Faulting occurs in the uppermost strata, indicating that the fault is still active.

Fig. 8. One of the many abundant gravel occurrences on top of the Gols–Illmitz feature. Near to the sharp separation line of the high gravel abundance there is a small topographic escarpment. The inset shows the size distribution of the gravels.

Geoelectric mapping (Kohlbeck et al., 1993, 1994, 2000) and Szarka et al. (2003) in the lake, somewhat more to the east, off-shore of Podersdorf, also supports this structural direction. The resistivity pattern shows a NE–SW trending, elongated resistivity contrast (Fig. 6). Resistivity contrasts and varying resistivity patterns were also recorded in the southern part of the lake, south of the Illmitz–Fertőrákos line aligned with the aforementioned direction.

5.2. Location of lineaments

The linear elevated areas mapped in the ALS DTM occur in Pleistocene fluvial sediments. Consequently, if they are of tectonic origin, they must be younger than late Miocene, which was a structurally very active period in this region (Decker, 1996). According to the earthquake data (Lenhardt et al., 2007), higher seismic activity occurred at the eastern boundary of the Little Carpathians, which is the along-strike continuation of the investigated area, than at the western margin, along the Pottendorf Fault (Fig. 1), representing the southeastern boundary of the Vienna Basin. Recent seismic activity in the Seewinkel is further indicated by several historical, as well as instrumentally recorded earthquakes (Lenhardt, 2000). Of special importance was the Halbturn earthquake of 11th February 1989 (magnitude 4.1), which formed at ca. 10 km depth. Moment tensor solutions indicate the generation of a steeply dipping, NW–SE trending normal fault that forms the SW boundary of the Parndorfer Platte (Lenhardt, 2000).

Note that at its present-day water level, Neusiedlersee has a concave eastern part that coincides with the strike of the mapped lineaments. Although the water level has changed continuously with time, the position of this part of the shoreline has been relatively constant, because of the high relief along this side in comparison with the surrounding area. This supports the notion of a fault in this area. The small escarpment at the SW margin of the lake is also completely aligned with the features in question, though here the geological setting is very different to that of the Seewinkel (Fig. 10).

The spatial distribution of the lake bathymetry (Bácsatyai et al., 1997) also shows bimodality: in the southern part the water is shallower than farther the north, and the change is aligned with ALS DTM elongated features. A small gravelly sandbank (the *Schotterinsel*) lying in the middle of the lake formed an island showing abundant gravel on its top in 1964, when the lake level was extremely low. Further banks (the *Fünfschoppen*) lie farther to the west; these features, together with the

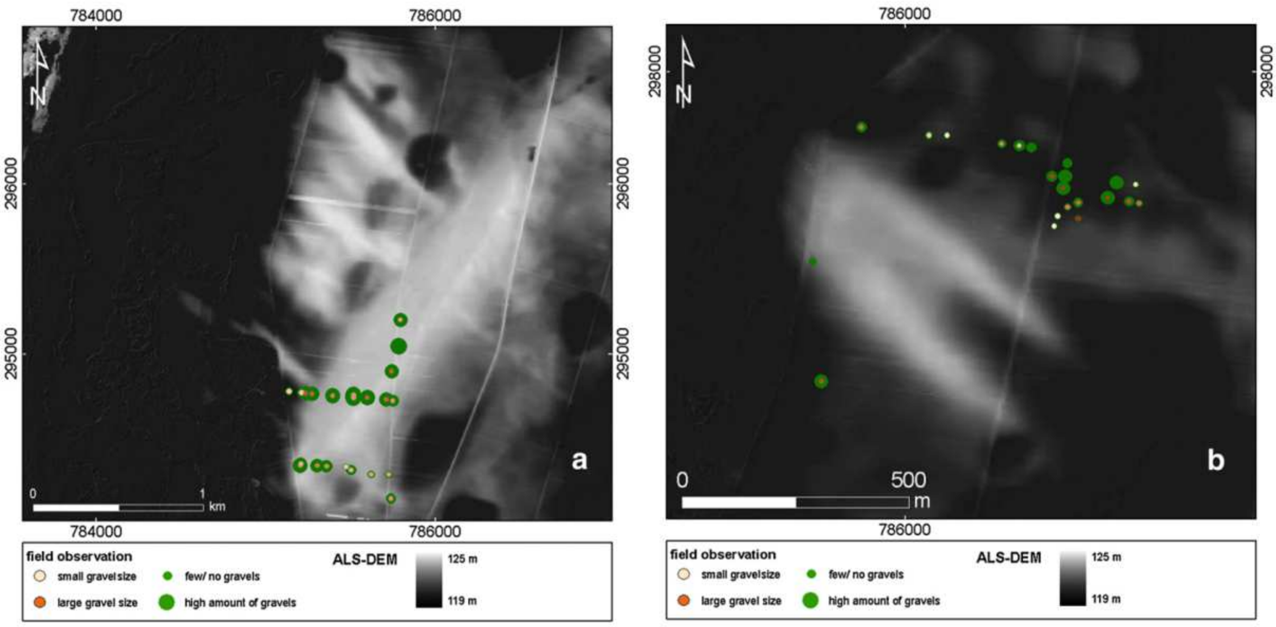

Fig. 9. The spatial distribution of gravel occurrences along a few sections. Green circles represent the abundance of the gravel on a relative (and somewhat subjective) scale. Pale to orange scale of the small inner dots indicate the maximum size of gravels observed. Note that the high gravel abundance appears also along topographically deeper areas. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

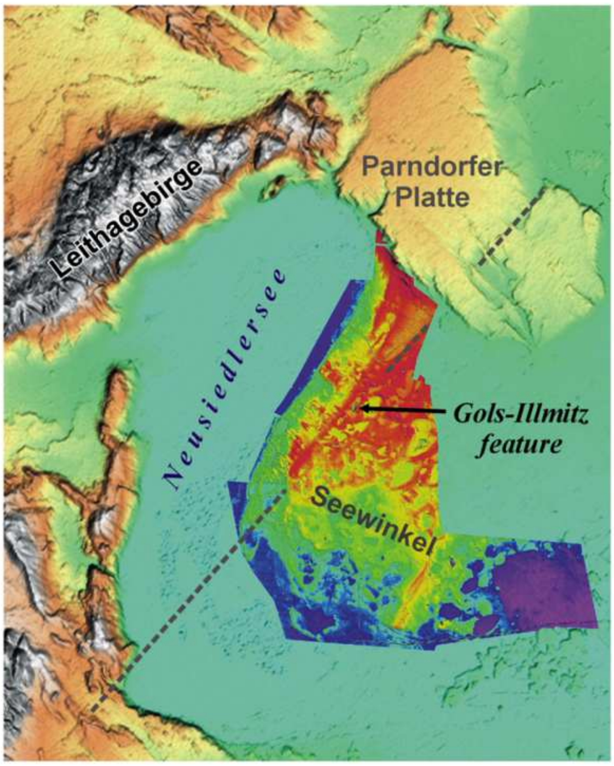

Fig. 10. An interpretative sketch of the correlation of the Gols–Illmitz feature and other tectono-geomorphic elements of the surroundings. The dashed straight line (at places masked out intentionally for clarity) connects the sudden geomorphic, and also sedimentological, changes of the Parndorfer Platte with topographic escarpment SW of the Neusiedlersee. The main part and a NE branch of the Gols–Illmitz feature (that can be identified at places as the Mönchhof fault) is almost perfectly aligned with the dashed line. The rainbow-coloured part is a processed version of the ALS DTM, the surrounding is the SRTM DTM. See further discussion in the text.

Schotterinsel, conform to the direction defined by the high resistivity anomaly mentioned above. An eolian formation from the gravelly spots in the area is disproved by the gravel abundance at the lake bottom.

Häusler et al. (2007) discuss the positional issue of the Mönchhof Fault. They cite Figdor and Roch (1984), who carried out gravity measurements along a 1 km long profile, found an NW–SE trending anomaly SW of Frauenkirchen that the latter authors interpreted as the possible position of the fault. This coincides with one of the features in our ALS DTM evaluation.

Although drainage networks are usually good indicators for active tectonics, this is scarcely possible in the present instance. The study area can be divided into two distinct parts from this viewpoint; a southern, extremely low relief area, practically without a drainage network, reflected in the existence of playa lakes and a northern area, in which the position of the rare perennial water courses has presumably been influenced by human activity since Roman times.

In contrast, the adjacent Parndorfer Platte has an interesting dry valley pattern, comprising elongated features trending NW–SE, showing an abrupt orientation change at a NE–SW trending line, hereafter referred to as Zurndorf–Gols zone. This is in line with the Gols–Illmitz feature to the south. Furthermore, the variation in thickness of Quaternary sediments on the Parndorfer Platte (Lipiarski et al., 2001) also changes approximately at the Zurndorf–Gols zone. Two instrumentally registered earthquakes during the 1980 occurred in this zone (Lenhardt, 2000). According to the online maps of the DANREG 2000 project (Scharek et al., 2000), the topmost layers in the northwest and southwest part of the Parndorfer Platte are covered by two slightly different lithological units. The boundary, again, coincides with the Zurndorf–Gols zone.

In the south easternmost part of the study area, a 5 km long and 300 m wide ridge has similar characteristics to the extensively analyzed features near the lake (Fig. 6). Although there is no comparable geophysical information, we suggest that neotectonic processes may have formed this ridge.

A number of elevated, 150 m wide, 600–800 m long ridges aligned in NW–SE direction have also been observed west of the Gols–Illmitz feature, forming a fishbone-like structure. Farther north, a 900 m wide elevated strip with the same strike as at Podersdorf (hereafter referred to as Podersdorf High) begins on the western lake shore and extends almost to the Gols–Illmitz feature, then, after a northward offset of 1.1 km, seems to continue further to the SE, with a characteristic width of 450 m. This latter feature is partly disrupted by depressions in a seemingly regular pattern.

The aforementioned pattern of ladder-shaped ridges (Fig. 6) observed in the ALS DTM belongs to one deformation system. One of the possible interpretations includes sets of conjugate faults. The small relief may also result from the slightly increased cementation of the faults zones due to fluid flow, increasing erosion resistance. A further possibility is the assumption that these faults within porous clastic sediments must represent deformation bands that — per definition — form zones of reduced porosity which characteristically show low erodibility (Aydin, 1978). This interpretation would endorse the selective erosional explanation of the Gols–Illmitz feature discussed above and would imply a precedence of faulting to the erosion.

6. Conclusions

A neotectonic interpretation of ALS DTM data has been undertaken in an extremely low relief area in the NW part of the Little Hungarian Plain. Linear elevated ridges, with a relief of only some 2 m, are the most prominent features in the area. Two meters microtopography seems insignificant, but it has a pronounced impact on microclimate and freezing/thawing in the intermediate season, factors extremely important for the local wine production.

In spite of intense agricultural re-working of the ground, the linear geometry of these elongated ridge-like features is clearly visible in the ALS data. Their length ranges from hundreds of meters up to 15 km. A comparison of these surface structures with field data, as well as geological maps, previous tectono-geomorphic evaluations, geophysical data, earthquake foci integrated in a GIS, supports their neotectonic origin.

Most of the structures, especially their spatial pattern, are only revealed by detailed analysis of the high-resolution ALS DTM. Generally, the longer features trend in NE–SW direction, with shorter ones oriented NW–SE. The longest NE–SW trending ridge, which extends from Gols to Illmitz, has been analysed in detail, showing that it coincides exactly with the Mönchhof Fault defined by previous borehole and seismic data. Recent earthquake foci suggest that the fault is tectonically active. So on-going surface deformation, or active deformation in the near geological past is possible. The determination of the exact surface location of this fault and the geomorphologic expression of its neotectonic activity is greatly enhanced by ALS data.

In summary, the Zurndorf–Gols zone and the Gols–Illmitz feature have a common origin which is actually the Mönchhof Fault postulated by Tauber (1959c). This fault is seismically active and geophysically proven at places; from the directional point of view parallel to the Neusiedlersee Fault and with the major strike-slips of the MMLZ.

The study also shows that this fault extends at least from Fertőrákos through Neusiedlersee up to Zurndorf at the northeastern edge of the Parndorfer Platte (Fig. 10). Interpreting the results of Tari (1996), who investigated the tectonic evolution of the NW part of the Pannonian Basin, based on seismic and well data, we conclude that his pattern of broadly NE–SW as well as NW–SE trending faults are related to the geomorphic features considered in this study. From the aforementioned Fertő, Répce and Ikva faults (Tari, 1996; Szafián et al., 1999) we interpret the Fertő Fault as SW continuation of the Mönchhof Fault (Figs. 2, 3 and 10). The Vištuk Fault (Lenhardt et al., 2007) and the Malé Karpaty Fault (Kováč et al., 2002) are possible candidates for the northeastward continuations of the Mönchhof Fault and Neusiedlersee Fault, respectively.

Acknowledgements

This work has been carried out in the framework of Christian Doppler Laboratory for Spatial Data from Laser Scanning and Remote Sensing (Vienna University of Technology, Austria) and projects OTKA T47104 and NK60445 of the Hungarian National Scientific Fund, as well as project P536001 Karpatian Tectonics (University of Vienna). The ALS data were acquired in the framework of *Spatial Information Systems for Transnational Environmental Management of Protected Areas and Regions in CADSES* (SISTEMaPARC) Community Initiative Programme INTERREG IIIB project, partly financed by the European Union. This paper is a contribution to the *TOPO-EUROPE — 4D Topography Evolution in Europe: Uplift, Subsidence and Sea Level Rise* project supported by the ESF. A. Zámolyi contributed as PhD student, received funding from the Stiftung Aktion Österreich-Ungarn; his contribution will be a part of his PhD thesis. L. Kárpáti, G. Takács (Fertő-Hanság National Park, Hungary), A. Herzig (Nationalpark Neusiedlersee-Seewinkel, Austria), F. Kohlbeck (TU Vienna) and L. Szarka (GGKI, Sopron) provided valuable help. A. Beidinger, G. Draganits, B. Salcher, and H. Thyringer are thanked for their support and discussions during fieldwork. H. Rice (Vienna), M. Vrabec (Ljubljana) and an anonymous reviewer are thanked for their constructive remarks and linguistic improvements.

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15. Appendix B

Zámolyi, A., Székely, B., Draganits, E., Timár, G. 2010. Neotectonic control on river sinuosity at the western margin of the Little Hungarian Plain. Geomorphology, 122, 231-243.

Geomorphology 122 (2010) 231–243

Contents lists available at ScienceDirect

Geomorphology

journal homepage: www.elsevier.com/locate/geomorph

Neotectonic control on river sinuosity at the western margin of the Little Hungarian Plain

A. Zámolyi [a,b,*], B. Székely [a,c], E. Draganits [d], G. Timár [a]

[a] Space Research Group, Department of Geophysics and Space Science, Institute of Geography and Earth Sciences, Eötvös University, Pázmány Péter sétány 1/A, H-1117 Budapest, Hungary
[b] Department of Geodynamics and Sedimentology, University of Vienna, Althanstrasse 14, A-1090 Vienna, Austria
[c] Institute of Photogrammetry and Remote Sensing, Vienna University of Technology, Gusshausstr. 27-29, A-1040 Vienna, Austria
[d] Institute for Engineering Geology, Vienna University of Technology, Karlsplatz 13/203, A-1040 Vienna, Austria

ARTICLE INFO

Article history:
Received 2 April 2008
Received in revised form 15 February 2009
Accepted 16 June 2009
Available online 17 July 2009

Keywords:
Neotectonics
River sinuosity
Little Hungarian Plain
Vienna Basin
Second Military Survey
Alpine–Carpathian–Pannonian Junction

ABSTRACT

This study investigates the influence of neotectonic activity on river channel patterns in low-relief areas. Our study area, the westernmost part of the Little Hungarian Plain, belongs to the Danube catchment in the transition zone between the Eastern Alps and Western Carpathians. This area evolved within the Pannonian back-arc basin during the Neogene and was also affected by the major lateral tectonic extrusion of the Eastern Alps.

Water course analysis has been carried out on the Leitha, Répce, Rábca, Ikva and Wulka rivers, to detect a possible relationship between their river courses and any on-going tectonic activity that is otherwise difficult to detect in this poorly exposed low-relief area. In order to derive channel geometries hardly modified by human activity (i.e. prior to the major river control works of the last 150 years), calculations of river channel properties were based on georeferenced historical map sheets of the Second Military Survey of the Habsburg Empire. These recorded the channel patterns and geomorphologic situation around 1840.

Classic sinuosity values from the reconstructed river courses have been derived using several window sizes. The calculated values show surprisingly strong local variations, considering the low-relief and lithological homogeneity of the area. The spatial distribution of the pronounced sinuosity variations coincides with the location of Late Miocene faults well-known from seismic data. On-going active tectonic activity along these faults is further indicated by the local earthquake record and geomorphic parameters derived from high-resolution digital elevation models. In conclusion, river sinuosity calculations represent a sensitive tool for recognizing neotectonic activity in low-relief areas.

1. Introduction

The assessment of possible tectonic activity in low-relief, densely populated areas is of special importance when estimating infrastructure vulnerability. However, detection in such areas is very difficult because very few applicable methods exist, especially if the area is characterized by low seismicity. A possible method is provided by geomorphologic evaluation: analysis of alluvial river dynamics has been found to be an appropriate tool (Ouchi, 1985). Even the smallest changes in the topography affect the sinuosity of low gradient rivers (Holbrook and Schumm, 1999), providing hints on on-going micro-topographic changes.

The study area, the Little Hungarian Plain (LHP, Kisalföld in Hungarian), situated at the westernmost margin of the Pannonian Basin (PB), belongs to the Danube River catchment area (Fig. 1a). It is characterized by a very low-relief energy, of less than 10 m/km² (Góczán, 2002) and a relatively homogenous cover of clastic sediments (Scharek et al., 2000a). Due to the location, at the boundary of the Eastern Alps, Western Carpathians and Pannonian Basin, on-going tectonic activity is expected, but difficult to locate.

Under certain conditions, alluvial rivers tend to evolve as single meandering channels. This behaviour is influenced by tectonic movements, reflected in river channel parameters. Within a given range of channel gradients, the meander pattern changes as vertical tectonic movements influence the valley slope (Fig. 3). This process is largely independent of river size, once the fluvial system enters the meandering stage. In this way, not only large rivers are suitable for analysis, but – as shown in this study –smaller creeks and reaches can also be evaluated, so long as they are essentially free of human influence.

In this paper, we analyze the smaller water courses of the LHP to find indications of on-going vertical movements. As the topography of the study area has been subject to anthropogenic influence since the Neolithic, including flood-control measures since the beginning of the 19th century, the river channels seen today are far from being in their

* Corresponding author. Space Research Group, Department of Geophysics and Space Science, Institute of Geography and Earth Sciences, Eötvös University, Pázmány Péter sétány 1/A, H-1117 Budapest, Hungary.
E-mail address: andras@zamolyi.info (A. Zámolyi).

0169-555X/$ – see front matter © 2009 Published by Elsevier B.V.
doi:10.1016/j.geomorph.2009.06.028

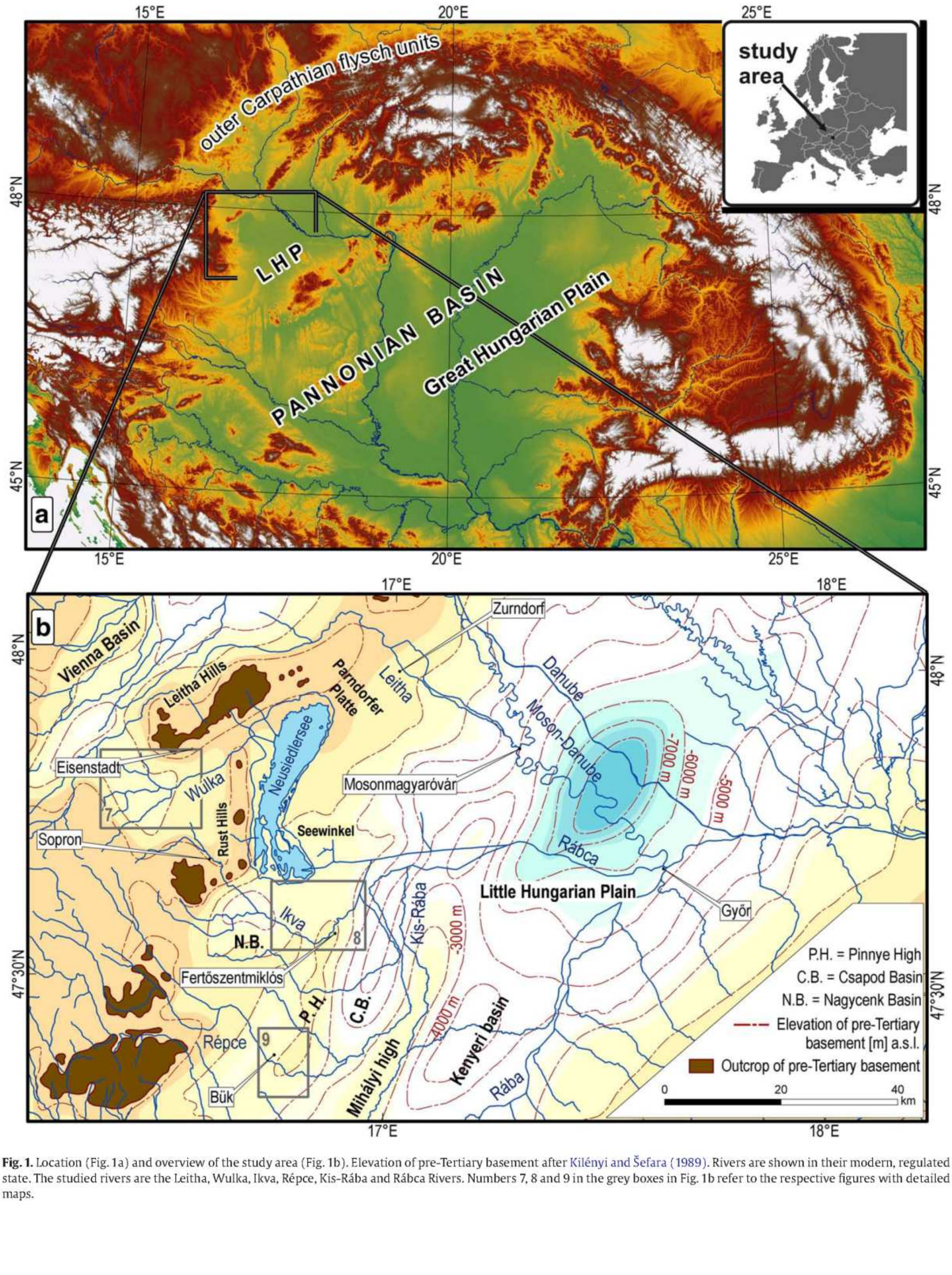

Fig. 1. Location (Fig. 1a) and overview of the study area (Fig. 1b). Elevation of pre-Tertiary basement after Kilényi and Šefara (1989). Rivers are shown in their modern, regulated state. The studied rivers are the Leitha, Wulka, Ikva, Répce, Kis-Rába and Rábca Rivers. Numbers 7, 8 and 9 in the grey boxes in Fig. 1b refer to the respective figures with detailed maps.

natural condition. To reconstruct an almost uninfluenced state, historical maps made during the Austro–Hungarian Empire were used to delineate river courses.

Classic sinuosity values were calculated for six streams in the LHP and compared with the results from industrial seismic sections (from ELTE database, originated from MOL national archive; see also Horváth, 1995; Tari, 1994; Rumpler and Horváth, 1988). In this novel approach, data from historical maps and geological–geophysical observations were combined to reveal the interaction of surface processes with basin evolution.

2. Tectonic and geomorphic setting

The WNW-part of the Pannonian Basin is bordered by the sinistral Vienna Basin Transfer Fault zone; this is geomorphologically accompanied by hilly and tectonically elevated horst structures (Leitha and Rust Hills; Schnabel, 2002; Fig. 2). A crystalline complex, the easternmost outcrop of the Eastern Alps, the Sopron Hills, forms the western margin of the LHP. The study area within the LHP is bounded in the east by the confluence of the Danube and the Rába rivers at Győr (Fig. 1b). To the south, it stretches as far as the southernmost tip of the Répce River. The Leitha River, within the Vienna Basin, marks the boundary to the west. Since the majority of the studied rivers originate in Austria, but reach their recipient in Hungary, there are both German and Hungarian names for the rivers; Table 1 provides a summary of river names in both languages and their hydrological data. Of the five rivers studied (Leitha, Répce, Rábca, Ikva and Wulka), the Wulka, Ikva and the Répce originate in the foothills of the Eastern Alps (Figs. 1b and 2). The Wulka River crosses the Eisenstadt basin and forms an estuary to Neusiedlersee. The Rábca, representing the continuation of the Répce, flows into the southern branch of the Danube at Győr (Fig. 1b). The Leitha River is more complex, since it is formed by the confluence of two rivers of almost equal size, one coming from the Northern Calcareous Alps (Schwarza River), the other from Lower Alpine tectonic units (Pitten River). The alluvial part of the river course is divided by a water gap at the northern tip of the Leitha Hills, forming an antecedent valley. Upstream from this point, the Leitha drains a narrow band of the SE Vienna Basin; in the downstream direction, it crosses the LHP and also reaches the southern branch of the Danube (Fig. 1b).

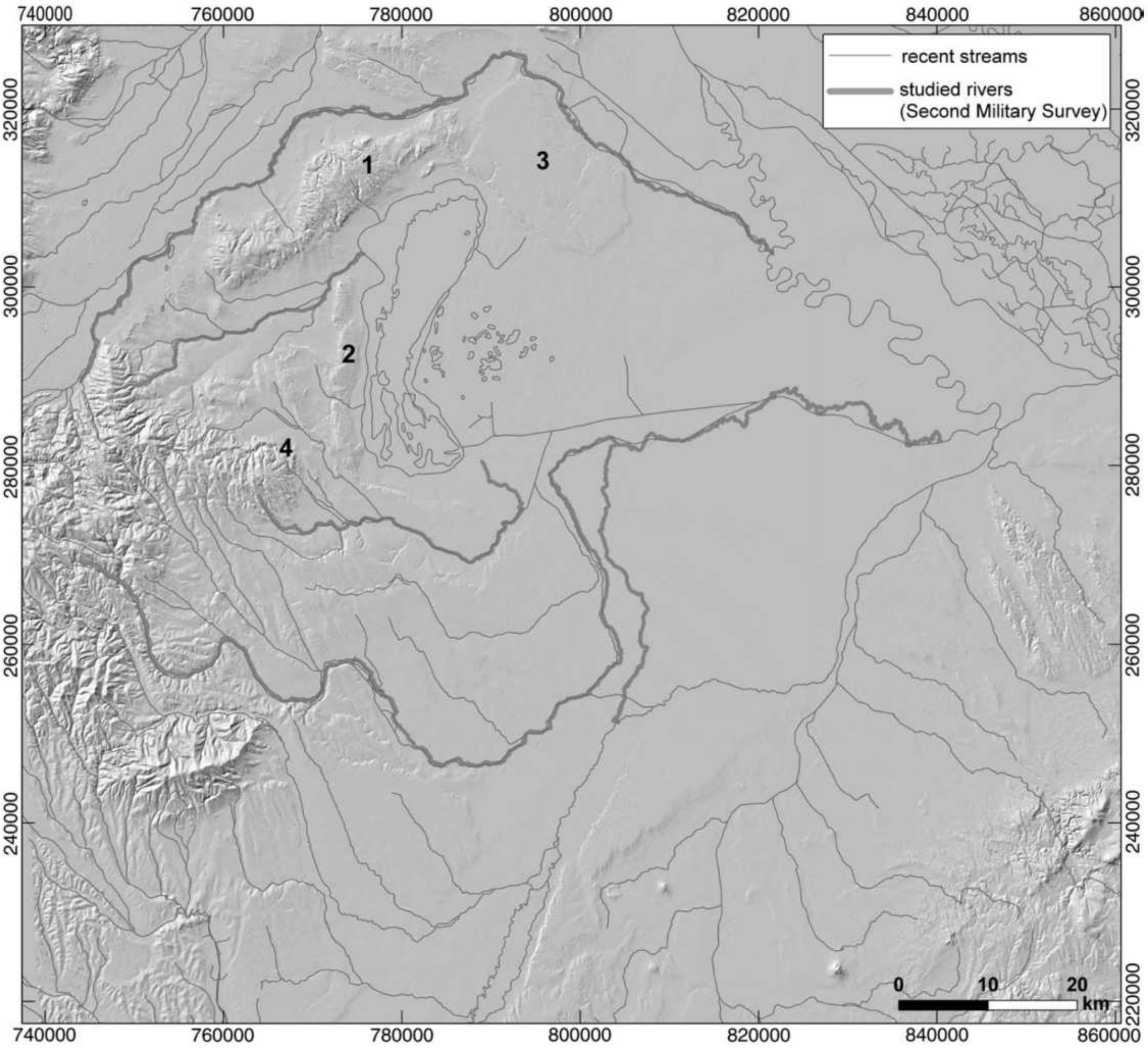

Fig. 2. Shaded DEM of the study area without interpretation for geomorphological overview. Recent streams are shown in thin grey, thick grey lines represent the river courses digitized from historical maps. 1: Leitha Hills, 2: Rust Hills, 3: Parndorf Plateau, 4: Sopron Hills (coordinates: Austrian BMN-grid).

Table 1
Summary of river basin characteristics of the rivers in the study area. Hungarian data: ÉDUKÖVIZIG (2008), Austrian data: BMLF, (2003).

River names		Total length [km]	Total catchment area [km^2]	Mean annual discharge [m^3/s]	Gauge station	Elevation [m] a.s.l.	
Hungary	Austria					Origin	River mouth
Lajta	Leitha	186	2379	10.59	Mosonmagyaróvár	906	118
Wulka	Wulka	39	400	1.16	Donnerskirchen/ Neusiedlersee	482	115
Ikva	Spittelbach[a]	64	1188	1.03	Pamhagen	467	115
Répce Rábca	Rabnitz[b]	199	4816	15	Győr	700	110
Kis-Rába	Kleine Raab	41	–	–	–	140	114

[a] Magda (1832).
[b] Downstream of the confluence of the Répce River with the Kis-Rába tributary, the Répce is called Rábca. The naming convention in Austria does not follow this rule.

The rivers in the study area have mean annual discharge values in the range of 2 to 10 m³/s and a total relief of 300–700 m, of which only 30–50 m are located in the LHP sections (Table 1). They flow through different lithologies, including crystalline and metamorphic rocks, carbonates and Neogene sediments. After entering the Neogene Vienna Basin and the Little Hungarian Plain, they evolve into alluvial rivers flowing on gravel deposits of mainly Würmian age. The Wulka and the Ikva have their main catchment areas within Pannonian sediments. Tributaries with a mean discharge of the same order of magnitude as the studied rivers are atypical in the study area, because the former area of the Neusiedlersee (Draganits et al., 2006, 2008) and the neighbouring Seewinkel are practically free of any water courses. Furthermore, the rivers crossing the surrounding area (Leitha, Répce, Rábca, Ikva and Wulka) have no tributaries SE of the axis of the Gols–Illmitz topographic feature (Fig. 5; Székely et al., 2009). North of the Seewinkel, an elevated area, the Parndorf Plateau (Parndorfer Platte), stretches in a NW–SE direction, dissected by SE-oriented dry valleys. The plateau rises to approximately 10–45 m above the surrounding area and forms a relic fluvial surface gently tilted towards the SE. The main body of this plateau consists of Pannonian freshwater sand deposits with a thin cover of Quaternary gravels of the Danube.

The current sedimentological environment developed from a previous palaeogeographic setting (see Magyar et al., 1999 and Kováč et al., 2006 for overviews). The evolution of the LHP started in the Lower Miocene with the formation of pull-apart basins in the northern part (Rumpler and Horváth, 1988; Vass et al., 1990). The basin was covered by Lake Pannon from approximately 10.8 to 9.5 Ma. The lake then retreated towards the SE, due to large delta-systems filling-up the lake with clastic sediments (Vakarcs et al., 1994). At around 9 Ma, the study area was dominated by alluvial plains of the early Danube and its tributaries. At 5 Ma, the lake shoreline lay at the southern rim of the Great Hungarian Plain (Magyar et al., 1999).

Due to the pull-apart geometry of the LHP, on-going subsidence was accommodated along transfer faults and major high- and low-angle normal faults. Those faults that lie within the deeper parts of the LHP have been clearly revealed by industrial seismic surveys (Tari, 1994; Horváth and Cloetingh, 1996; Szafián et al., 1999). Simultaneously with the deltaic filling-up of the basin, differential uplift characterized the evolution of the region. The basement topography (Fig. 1b) reflects earlier vertical displacements marked by horst and graben structures. This differential uplift continues now: vertical crustal movements with values of up to −2.2 mm/a have been recorded in the northern part of the Little Hungarian Plain near Győr (Joó, 1992; Joó et al., 2006). Towards the east, terraces of the Danube identified with geomorphological and geochronological methods indicate rapid incision by the Danube (Ruszkiczay-Rüdiger et al., 2005) into the uplifting northeastern part of the Transdanubian Range (reaching ca. +1.0 mm/a; Joó, 1992) .

3. Data and methods

The mosaic of uplifting and subsiding zones has considerably affected the river courses in the region: they react as sensitive sensors, modifying their courses according to the meso- and micro-topographic changes due to the differential uplift/subsidence patterns.

The detection of their geomorphic expression in low-relief areas can be improved by interdisciplinary methods. Analysis of high-resolution LiDAR DTM data shows a pattern of long NE–SW trending ridges cross-linked by shorter NW–SE trending ridges in the western part of our study area. Both of these features are of tectonic origin (Székely et al., 2009). The tectonic activity of at least some parts of these faults has been revealed by structural field data; this highlighted the influence of strike-slip faulting on the recent stress field (Csontos et al., 1991).

3.1. Fault data

The following fault and lineament datasets have been used: the 1:200,000 scale neotectonic map of the DANREG 2000 project (Scharek et al., 2000b), the 1:500,000 scale map of morphostructural elements of the Pannonian Basin (Horváth et al., 2006), the faults mapped on the 1:200,000 scale maps of the Austrian geologic survey at the western margin of the study area (Schnabel, 2002) and the 1:500,000 scale map of Landsat lineaments (Buchroithner, 1984; Figs. 5, 7, 8, and 9a). In the southern Vienna Basin, data from Hinsch et al. (2005a) and Salcher et al. (2008a,b) were integrated. Additionally, industrial seismic sections from earlier hydrocarbon explorations have been used extensively (Figs. 8 and 10). Deeper parts of the sections indicate fault activity during the Miocene, but the uppermost 300 m of the seismic sections used for hydrocarbon exploration are usually muted and thus scarcely usable to document neotectonic or active tectonic movements on faults. Fault indications in the remaining part of the sections were used in the evaluation.

3.2. Topographic data

In this study, river sinuosity has been used for the detection of vertical movements possibly induced by neotectonic faulting. In areas with a long tradition of water regulation, river courses that have not been influenced by anthropogenic activity must be used in the calculations. Georeferenced historical maps are one source of such data. The map sheets of the Second Military Survey of the Habsburg Empire (Jankó et al., 2005) represent the first comprehensive cartographic work conducted on a predefined geodetic basis and projection system (Timár and Molnár, 2003). The map sheets were continuously produced for the territory of the entire Empire within a time frame of fifty years (1806 to 1869) at a scale of 1:28,800

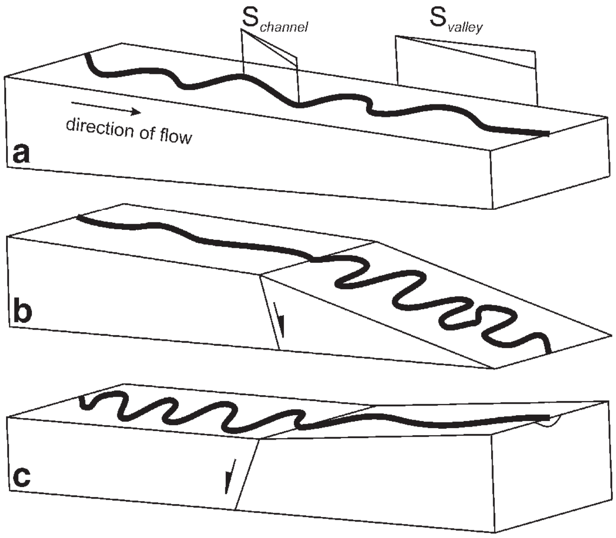

Fig. 3. Channel pattern adjustment of a meandering river affected by vertical tectonic movements. The self-organizing behaviour of the river tends to keep the channel slope constant (modified after Ouchi, 1985; Timár, 2003) resulting in increased sinuosity when the valley slope increases in down-stream direction (b) and leading to channel straightening and incision when the valley slope increases in the opposite direction (c).

(Kretschmer et al., 2004). The georeferenced versions of the maps (Timár et al., 2006) were used for analysis. Errors have arisen for several reasons: (1) in the 19th century, no geodetic adjustments of the high accuracy triangulation measurements were made; (2) the projection used by the engineers during surveying was not exactly the Cassini projection, although this formed the basis of rectification for integration into a geographic information system; (3) several minor error sources lie in the folding and deformation of the paper sheets during storage. The general error has been set at ca. 50 m in the study area (Timár et al., 2006). The topography on the map sheets was displayed using Lehmann-hachures. Steep slope angles are shown as dark, broad stripes, whereas gentle slopes appear as light, thin strokes; slope angles were measured with a goniometer (Jankó et al., 2005). This method provides a plastic impression of the landscape but lacks exact elevation data. However, river courses, along with their anthropogenic influences, were accurately mapped and the influence of possible errors from the projection on channel morphology is considered to be very low. Thus the 150 year old georeferenced river courses can be reconstructed with reasonable accuracy (see Fig. 9b).

3.3. The applied method of sinuosity analysis

Generally, a fluvial system reacts to tectonic influences by changing its longitudinal and cross-sectional profiles, channel pattern and/or sediment discharge. Several studies have documented the influence of vertical crustal movements on the channel pattern (e.g., Ouchi, 1985; Jorgensen, 1990; Holbrook and Schumm, 1999). Changes of the valley slope above a certain threshold may cause the channel pattern to shift between a braided, meandering, anastamosing or straight channel pattern. If conditions allow a meandering state in the river, minor intra-pattern variations occur, increasing the sinuosity of the channel (Friend and Sinha, 1993). Faults influencing the valley slope will affect the sinuosity of the channel, since the river tries to keep the channel slope constant in a self-organized manner (Fig. 3). A normal fault downthrowing in the upstream direction causes a straightening of the channel whilst downstream downthrowing faults result in increased meandering (Ouchi, 1985; Keller and Pinter, 1996; Holbrook and Schumm, 1999; Bridge, 2005).

A quantitative measure of the variation of the meandering pattern is the classic sinuosity index (SI; Leopold et al., 1964), given by the ratio of along-channel distance to the shortest path-length. However, tributaries, increasing the discharge and sediment load, may influence the channel pattern.

The SI values have been calculated here using the digitized planforms of water courses using the ESRI ArcGIS package. An appropriately small distance, d (here 50 m) has been chosen to define the overall resolution of the study. We inferred that with shorter distances than d no change occurs, or is detectable. Thus the water courses have been equidistantly resampled along their tracks, with the new vertices spaced at d distance (Fig. 4).

Since the sinuosity analysis is scale-sensitive, i.e. the size of the reference shortest path-length, termed as the window size, does matter (e.g. Lancaster and Bras, 2002), several window sizes were used. However, in our study, instead of keeping the reference straight line constant, the along-track distance was used as the window size. The window sizes (s_k) were multiples of distance d, with k set at values of 10, 20, 30, 50, 70 and 100; thus if $k=20$, then $s_{20}=1000$ m). The distance l_i between the associated vertices of the ith segment was determined, and the sinuosity indices $\mathrm{SI}_k(i)=s_k/l_i$ were calculated. The $\mathrm{SI}_k(i)$ values were assigned to the midpoint of the along-track distance (situated at $(k/2+1)$th vertex); in other words, the reference point is always a point (and also a resampled vertex) of the original channel.

4. Results of the sinuosity analysis

In this study, we focus on the changes in sinuosity in (i) the vicinity of faults and/or morphostructural elements, (ii) areas with high sediment accumulation rates and (iii) areas with a thin or no Quaternary cover. In the following, the sinuosity pattern of the analysed rivers is evaluated in the light of known tectonic features. At the end, possible tectonic influence on the river planforms is assessed.

4.1. The Leitha River

The Leitha River, which runs across the southwestern part of the Vienna Basin, is formed south of Lanzenkirchen at the confluence of the Schwarza, originating in the Northern Calcareous Alps, and the Pitten, with its source in the crystalline area of the Lower Austroalpine units (BMLF, 2003). The studied section of the river starts at the

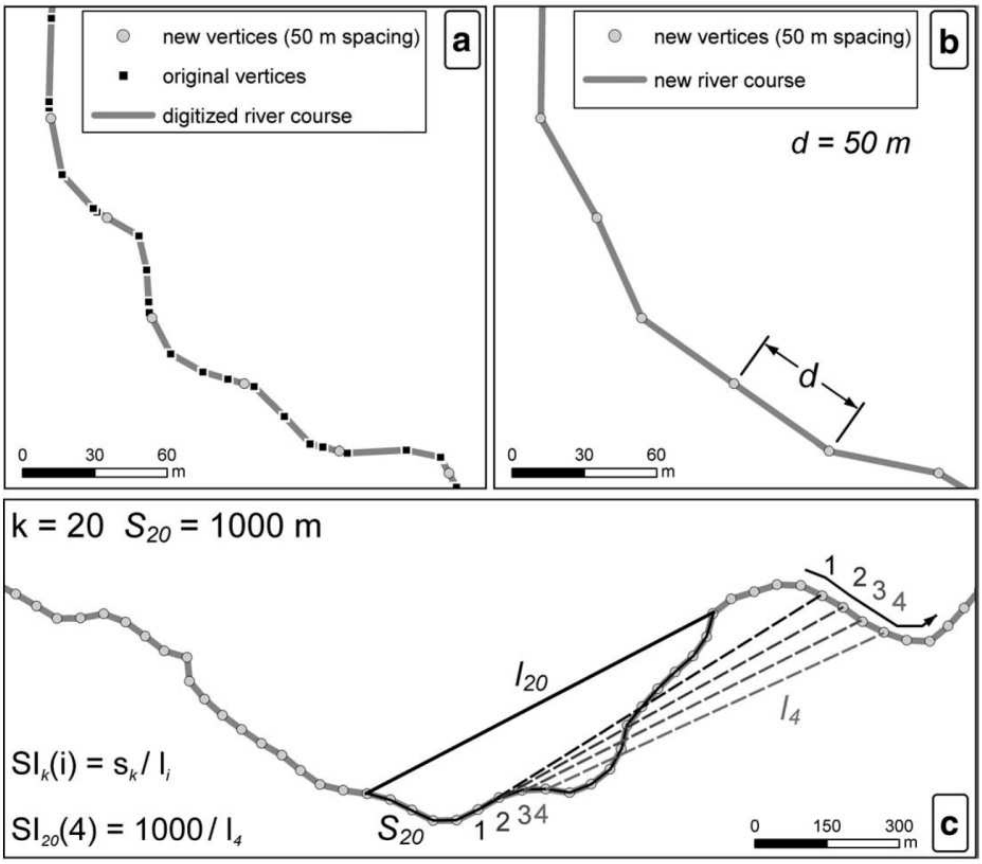

Fig. 4. Applied method of sinuosity analysis. Fig. 4a shows the original river course digitized from historical maps. In Fig. 4b the resampled rivercourse is displayed resulting from the new vertices placed with d spacing along the rivercourse. The shown section is the same as in Fig. 4a. Fig. 4c summarizes the calculation of the SI-values for $k = 20$. Using larger window sizes results in an increasing smoothing effect; sections with high sinuosity will be difficult to detect.

confluence and ends where the river joins the Moson–Danube. In the upper part of the river, several intersections with mapped normal faults occur (Schnabel, 2002) at two of which an influence on the river sinuosity is clearly visible (Fig. 5a). At location 1, the normal fault dips upstream and causes a decrease in SI values in a section with otherwise relatively high sinuosity (Fig. 5a). At location 2, the increase of sinuosity following sections with rather low SI values is probably the effect of a supposed normal fault (Fig. 6). The earthquake epicenters from the years 1995 to 2007 (Tóth et al., 2007) document on-going faulting at these locations.

Further downstream, where the Leitha River crosses the surface trace of the Vienna Basin Transfer Fault (Decker et al., 2005; Hinsch et al., 2005a,b; Lenhardt et al., 2007), a very sinuous section occurs, characterized by SI_{30}-values between 1.2 and 3.5, with most values between 1.6 and 1.8 (Fig. 5b). After turning 90° to the SE, which represents the typical direction of streams draining the LHP and which is also parallel to the NE-rim of the Parndorf Plateau, the sinuosity falls to below 1.2, sometimes approaching unity. The only exception is a 4.5 km long section near Zurndorf, where it exceeds 1.6, locally reaching 4.4. This section coincides with morphostructural features recognized by previous authors; this is further analysed in the discussion.

4.2. The Wulka River

The Wulka River, which is the smallest of the investigated rivers, originates in low grade metamorphic schists and crosses the Eisenstadt Basin, comprising almost 1500 m of Tertiary and Quaternary deposits (Kilényi and Šefara, 1989). On entering the basin, the sinuosity values along the river show three distinct regions with high values (Fig. 7). Two of these are located at the basin margin (Fig. 7: Locations 1 and 3) and coincide with mapped morphostructural elements (Horváth et al., 2006). Furthermore, location 1 coincides with mapped Landsat lineaments (Buchroithner, 1984). The increase in sinuosity in the center of the basin lies in the vicinity of the mapped Landsat lineaments. However, verification was not possible at this location (Fig. 7: location 2).

4.3. The Ikva River

The fault data of Schnabel (2002) and Scharek et al. (2000b) show a supposed normal fault W of Fertőszentmiklós that dips upstream. Upstream, the SI values form a peak, apparently related to an inferred fault (Scharek et al., 2000b; Schnabel, 2002; Fig. 8: location 1). Further downstream, at location 2 (33 km), SI values correspond to an intersection with a morphostructural element (Horváth et al., 2006). The decrease in sinuosity here suggests a normal fault dipping upstream (Fig. 8: location 2). In seismic sections, normal faulting of pre-Pannonian strata can be observed. The continuation of this normal fault zone into Pannonian and younger deposits is not clearly visible, due to the poor resolution of the seismic survey, but river dynamic data point towards possible fault activity (Fig. 8). In the surrounding region downstream, the SI values calculated with different window sizes show somewhat contrasting behaviour: overall, the values of the SI_{30} curve (corresponding to a 1500 m window) increase almost continuously to 38.5 km, while SI_{10} (500 m) exhibits an undulating

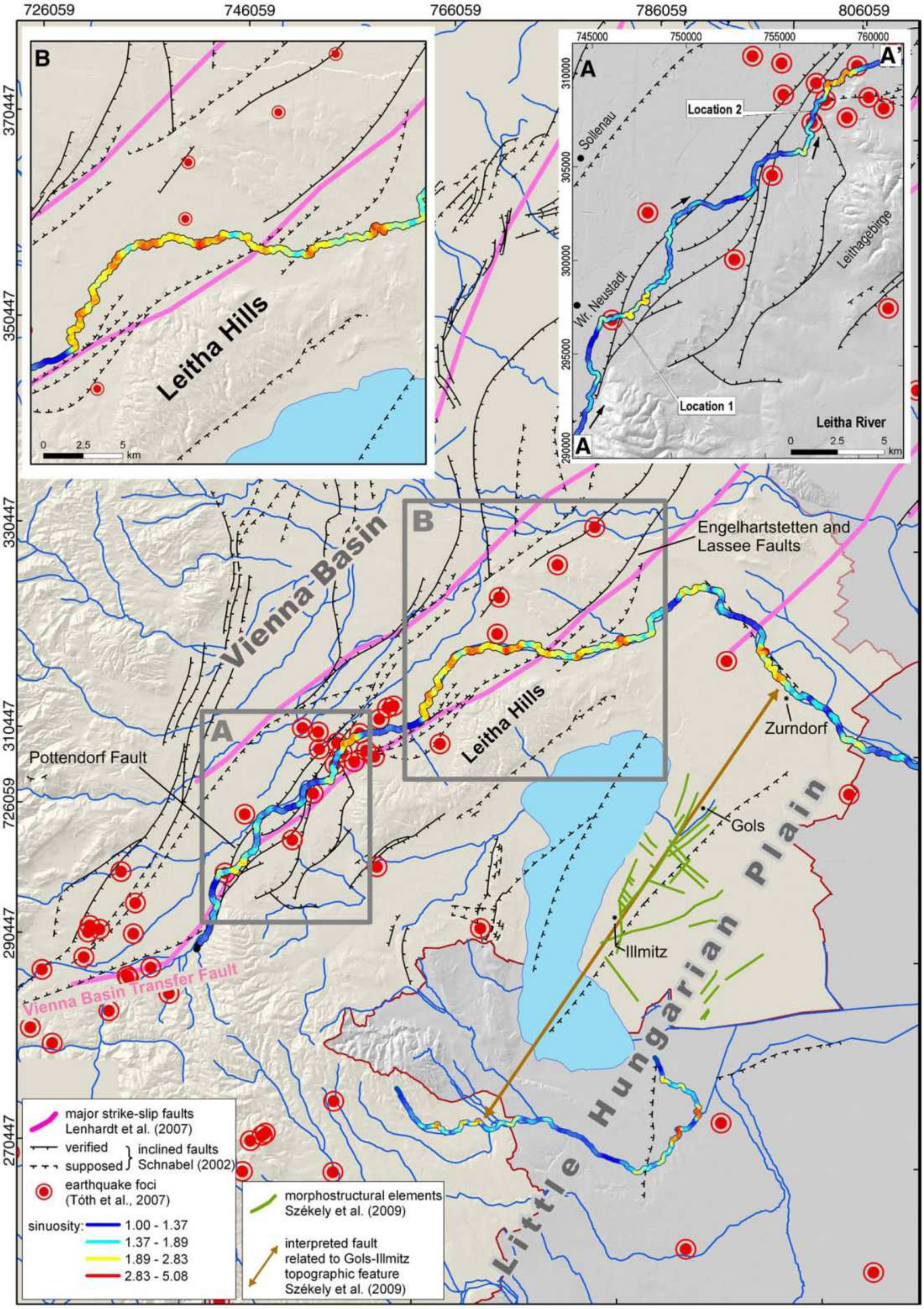

Fig. 5. Index map of river sinuosity values along a studied section of the Leitha River. Note the changes of river sinuosity at the indicated locations. Earthquake epicenters after Tóth et al. (2007). A shaded DEM serves as background (coordinates: Hungarian EOV-grid). For further explanations see main text.

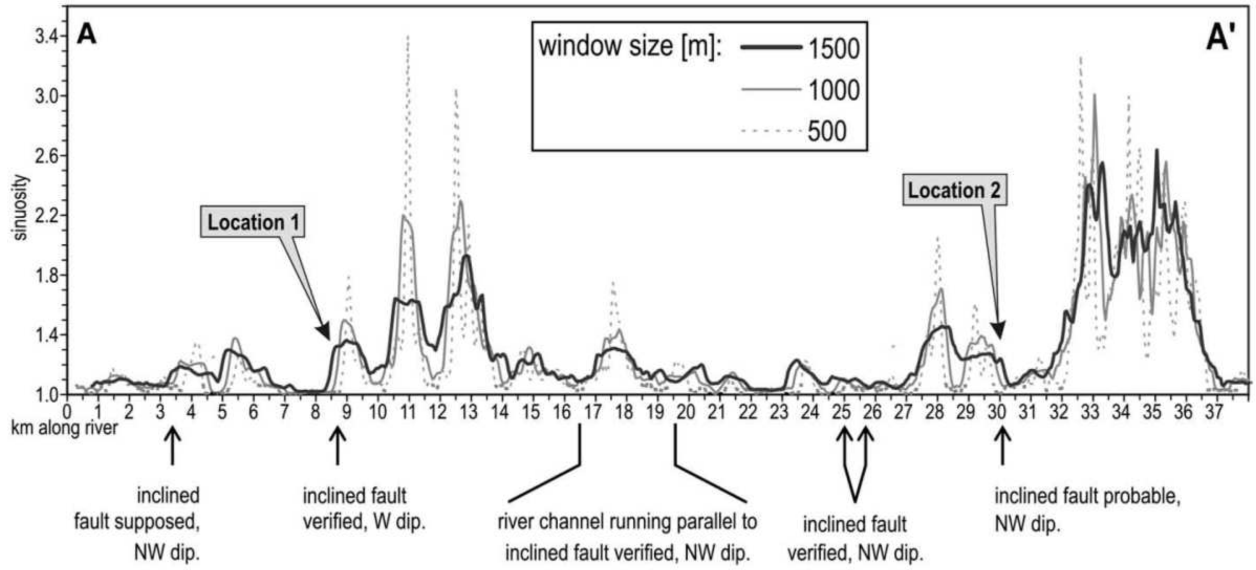

Fig. 6. Diagram of river sinuosity values from the studied section along the Leitha River shown in the index map (Fig. 5a). All fault data indicated are from Schnabel (2002).

pattern, with four marked domains of higher sinuosity (1.2 to 2.0) separated by ca. 1 km long sections with lower sinuosity values (ca. 1.1). This difference also underlines the importance of the window sizes: while the long-wavelength component clearly indicates the generally more sinuous zone, the individual features are outlined by the signal of SI_{10}. At location 3, where the behaviour of the curves of the SI values are the most contrasting, another morphostructural element underlies the Ikva River at a section with relatively high SI_{30} values of ca. 2.0 (Fig. 8).

4.4. The Répce River

The Répce River, which forms at the confluence of the Spratzbach and Thalbach creeks in Austria, flows through gneissic units of the foothills of the Eastern Alps and crosses the Répce Fault Zone, a major sinistral strike-slip structure (Fülöp, 1990; Tari, 1996; Szafián et al., 1999) forming two prominent wide bends. In Fig. 10, a marked correlation of sinuosity values can be observed above a distinctive horst, the Pinnye high, exceeding 500 m relief in the basement. A zone

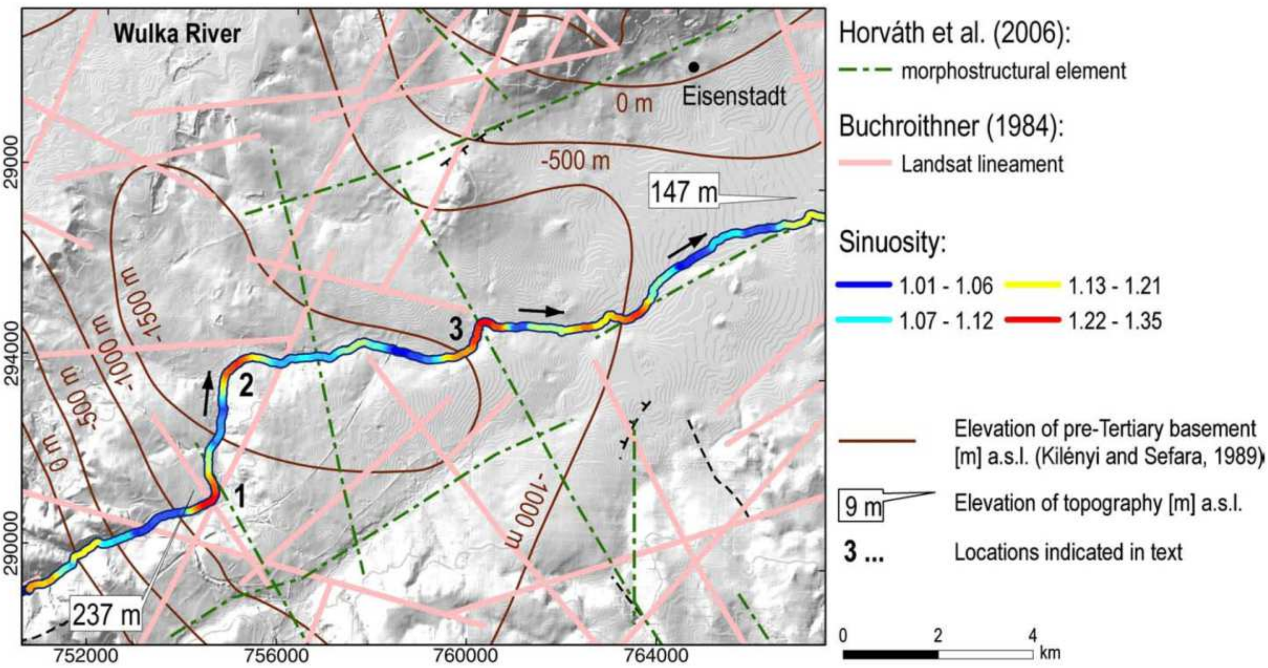

Fig. 7. Shaded DEM overlain with calculated sinuosity values of the Wulka River. Three locations with abrupt changes in sinuosity can be separated as the river crosses the Eisenstadt basin (coordinates: Hungarian EOV-grid).

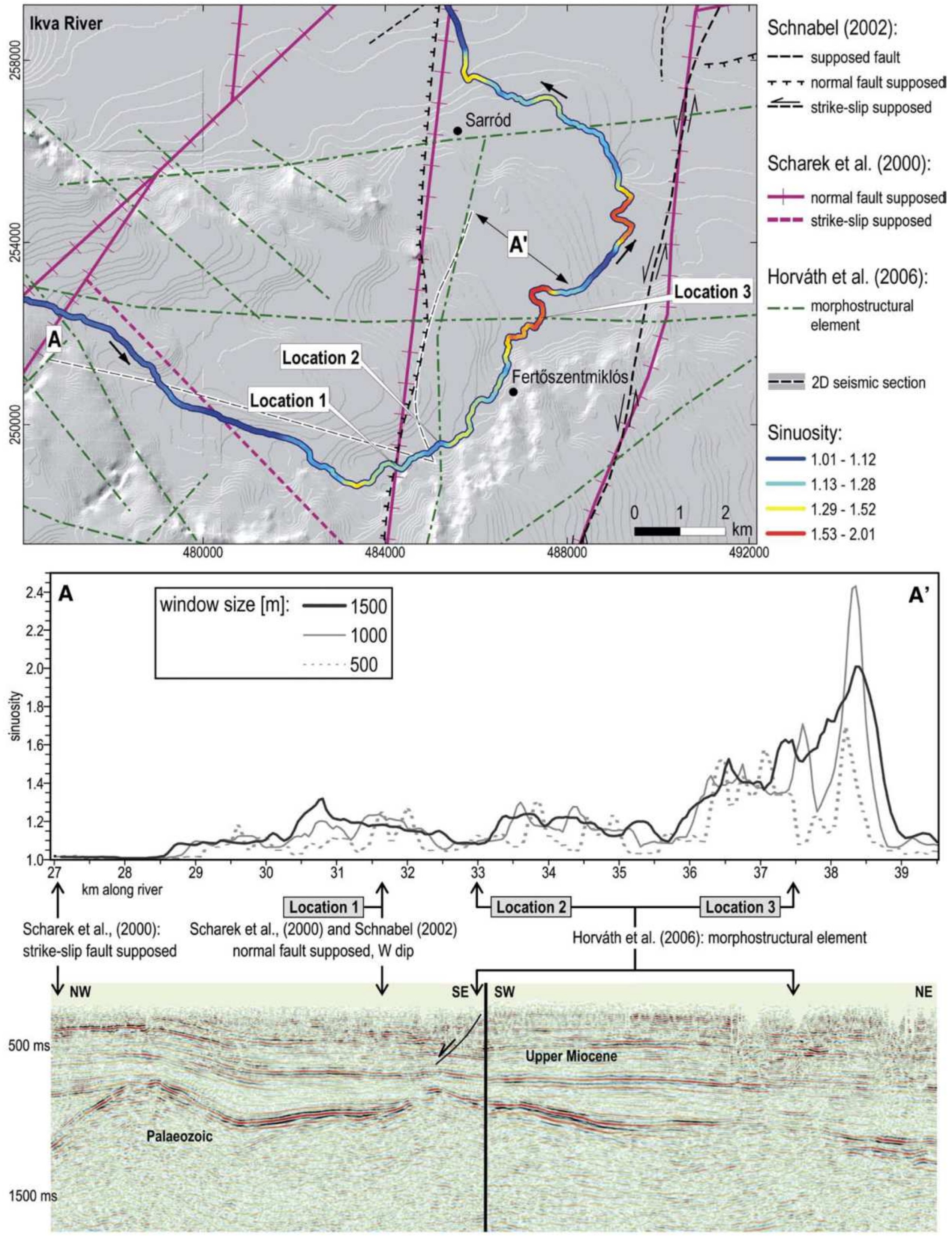

Fig. 8. Index map of a selected section of the Ikva River on a shaded DEM. Below the respective sinuosity profile correlated with an industrial seismic section (coordinates: Hungarian EOV-grid).

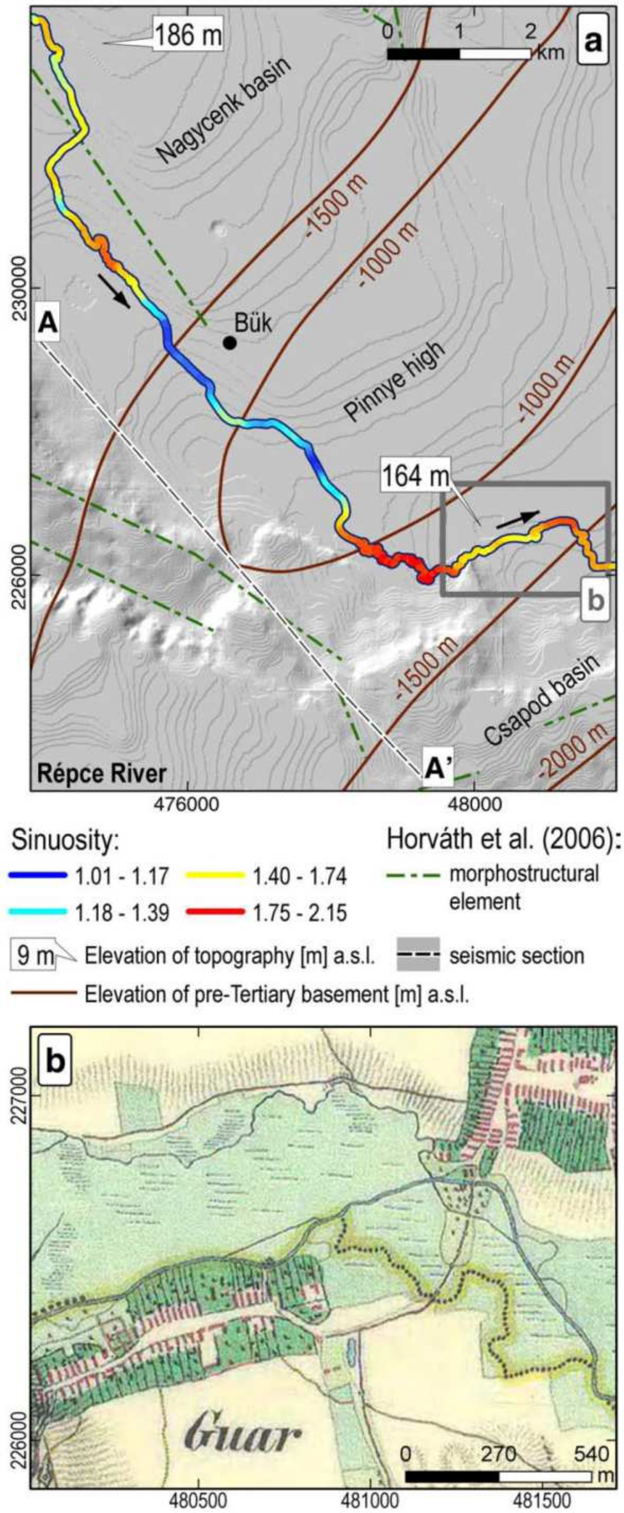

Fig. 9. Calculated SI-values of a section of the Répce River on a shaded DEM. Elevation of pre-Tertiary basement after Kilényi and Šefara (1989). Grey box indicates location of Fig. 9b. An example of a part of a map sheet of the Second Military Survey of the Habsburg Empire is displayed in Fig. 9b. The map sheets were constructed in the original scale of 1: 28800. Note the extensive wetlands and the detailed mapping of water elements, streets and settlements. The Lehman-stripes render a plastic impression of the geomorphology of the area (coordinates: Hungarian EOV-grid).

of high sinuosity on both edges of this pre-Neogene high (SI_{30} = 1.6 to the NW and SI_{30} = 2.0 to the SE) can be observed (Fig. 9).

4.5. The Rábca River

Downstream of the confluence of man-made channels and the Kis-Rába River, the Répce River is called the Rábca River (Table 1). This part of the river is strongly influenced by anthropogenic activity in the immediate surroundings of the confluence. The river course derived from the topographic map sheets of the Second Military Survey shows this influence clearly and thus it is not suitable for quantitative analysis. However, further downstream, the river system tends to regain its natural state and changes in sinuosity values, possibly correlatable with the mapped morphostructural elements identified by Horváth et al. (2006).

Summarizing the variations in sinuosity values, note that the majority of sinuous sections, except for the upstream sections of the Répce and Leitha rivers, are (S)SW–(N)NE oriented. This may imply a general, possibly tectonic trend. Since tectonic movements cannot be separated clearly into purely vertical and horizontal components, the planform pattern of the studied rivers is not only influenced by differential uplift, but also by strike-slip movements. It should be emphasized that in some cases angles of intersection of the fault and the river may also be subject to tectonic influence (e.g. river channel deflections along transform fault zones). In the study area, this occurs at several places. For instance, the course of Ikva River changes through almost 90° at two locations that closely coincide with tectonic faults (Fig. 8). Certain rivers run parallel to supposed faults and morphostructural elements: upon crossing the Pinnye high the Répce River changes its course, which can be attributed to an increase in valley slope due to on-going subsidence in the juxtaposed Csapod (SE) and Nagycenk basins (NW; Fig. 9, points marked by A and A′).

5. Discussion

Results of extensive studies carried out in recent decades show that the planform of meandering alluvial rivers is controlled by several factors. Beside the aforementioned tectonic effects and discharge, the sediment load plays an essential role (Schumm and Khan, 1972). Although grain size coarsening of the transported sediment is generally interpreted as the effect of tectonic activity along the river course, an approach based merely on this observation can be erroneous. In certain cases, the grain size is an indicator for sediment supply as the river crosses a lithological boundary or an indicator for depositional areas where the river leaves the coarse-grained sediment and only the fine-grained load is transported farther (Holbrook and Schumm, 1999).

Since relevant tributaries in most of the area of the LHP are lacking, sediment load can change only gradually: abrupt increases along the channel do not occur and hence variations in the SI values cannot be related to changes in discharge or sediment load. Therefore, in the study area it is reasonable to assume that sinuosity variations, especially if they appear along confined sections, are related to, or strongly coupled with differential uplift/subsidence pattern.

A good example for obvious tectonic control on the river pattern is the Répce near Bük (Fig. 9). The overall relief of the area is low, 22 m (186 m to 164 m) within 8 km, while, as we stated before, the Pinnye High rises more than 500 m in 1 km in the NW and drops even more in the SE. Rumpler and Horváth (1988) and Tari (1994) interpreted these basement faults as Miocene structures. The variations in SI values and the opposing deflections strongly correlate with the buried antiform. Furthermore, a few earthquake foci aligned with the flanks of the Pinnye high (Tóth et al., 2007) indicate tectonic activity. Consequently, we conclude that, at this point, the river channel responds to on-going faulting that re-activates Miocene fault structures by bending and changing its sinuosity.

Another, more striking, although more complex example is the SW–NE directed upper reach of the Leitha River. The river flows within the Vienna Basin Transfer Fault zone, a major strike-slip dominated fault zone (Decker, 1996; Decker et al., 2005; Hinsch et al., 2005a; Lenhardt et al., 2007) that forms the SE-margin of the Vienna Basin pull-apart. This master fault, which is offset by left-lateral step-over structures, continues up to the outer Carpathian flysch units and has proven neotectonic activity (Hinsch et al., 2005a; Lenhardt et al., 2007). Thus, within the southern Vienna Basin, geomorphology is influenced by on-going deformation along the master fault and its associated fault splays (Hinsch et al., 2005a,b; Zámolyi et al., 2006, Hölzel et al., 2008; Salcher et al., 2008a,b) resulting in slightly tilted regions (Zámolyi et al, 2006), fault scarps and hanging valleys (Hinsch et al., 2005b).

The spatial pattern and the high number of earthquake foci support the interpretation of increased sinuosity at location 2 (Fig. 5): here, the Leitha River crosses an actively deforming zone linked to the Pottendorf Fault.

The continuation of the Pottendorf Fault, also known as the Engelhartstetten and Lassee Faults (Decker et al., 2006; Beidinger et al., 2007), coincides with the generally highly sinuous section of the Leitha that runs down to the turning point of the river, where it leaves the Vienna Basin Transfer Fault zone.

The only short sinuous section downstream from this point is near Zurndorf, which is, in our interpretation, due to the river crossing the continuation of the Gols–Illmitz morphostructural feature (Székely et al., 2009; Fig. 5).

Although the Wulka River runs sub-parallel to the upstream reach of the Leitha River, the tectonic setting here, on the opposite side of the Leitha Hills, is very different (Fig. 7). High SI values at locations 1, 2 and 3 may indicate fault activity. However, verification with additional data is not possible in a satisfactory manner.

Further to the SE, within the LHP, the almost perpendicular deflections of the Ikva River fit well into a NW–SE and SW–NE oriented fault pattern. This agrees with the overall large-scale evidence of on-going faulting and the relative movements of the differentially uplifting areas (e.g. Fodor et al., 2005; Horváth et al., 2006; Joó et al., 2006). The similarly low SI values ($SI_{10} = 1.1$) of the lower reach of the Leitha River and of the entire alluvial section of the Ikva and Répce rivers, despite their differing discharge values, also indicate a congruent hydrological regime. In most cases, short sections with increased sinuosity can be reliably linked to either faults detected in seismic sections or mapped morphostructural elements. Furthermore, the spatial pattern of these sinuous sections shows a coherent picture, suggesting that they formed by on-going deformation.

In the Pannonian Basin, rivers with such small discharge values have not yet been studied with the methods documented here. However, the results are very similar to observations in the

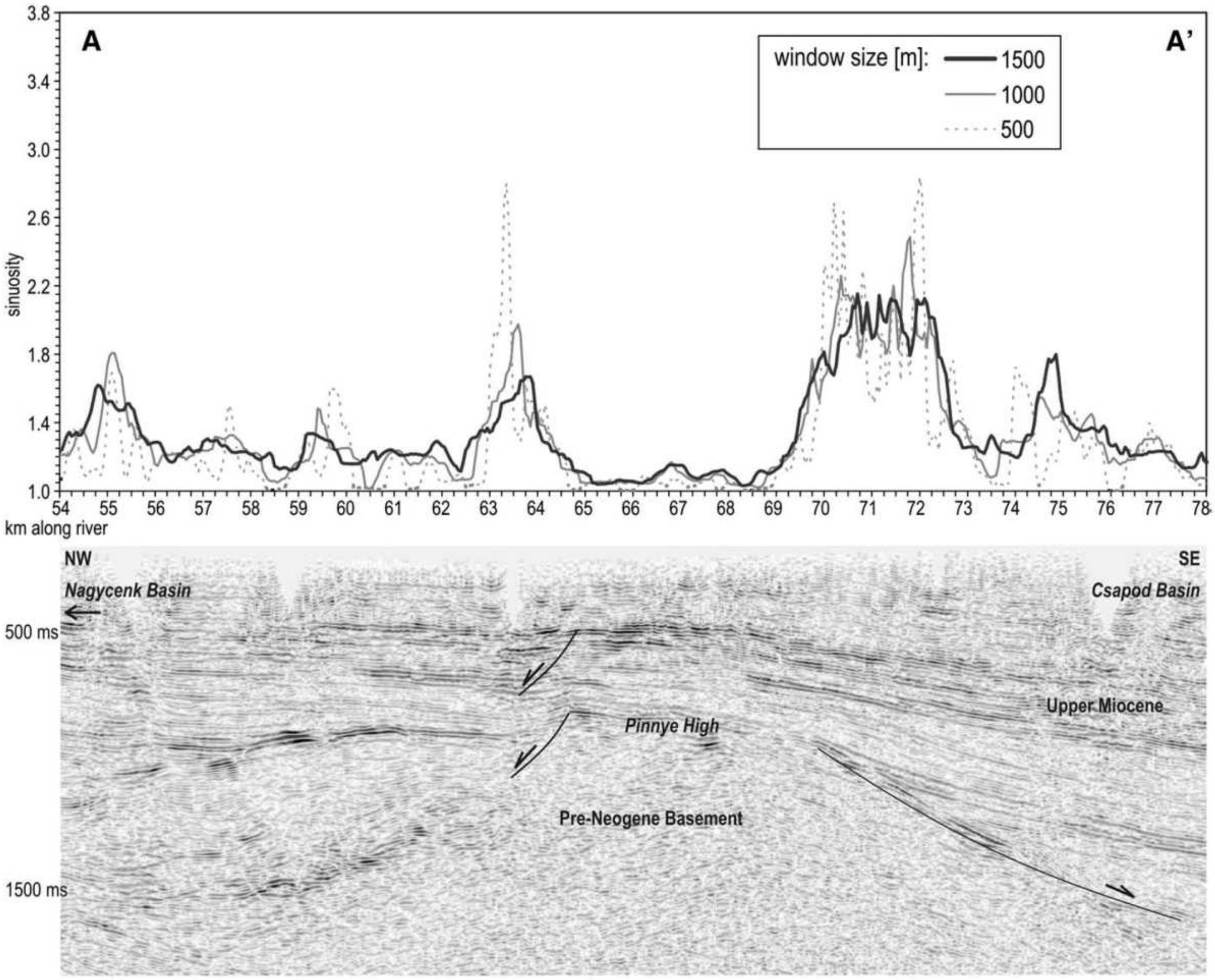

Fig. 10. Sinuosity diagram of the Répce River superposed on an industrial seismic section showing marked variations at the edges of the pre-Neogene Pinnye high.

geomorphologically comparable Great Hungarian Plain, where the Tisza River drains a large part of the eastern Pannonian Basin (Mike, 1975; Timár, 2003; Timár et al., 2005). Timár (2003) demonstrated a correlation between sinuosity changes in the entire alluvial part of the river with vertical crustal movements. The results gave insights into the geomorphical responses to tectonic forcing along the channel of a major river in one-dimension, paving the way for river sinuosity analysis in the Pannonian Basin.

6. Conclusions

Correlation of river sinuosity variations using verified fault data from geological and geophysical mapping demonstrated channel pattern adjustment in the LHP.

The calculated SI values, depending on the window size, are found to be representative for the investigated rivers in the study area. Comparison of the sinuosity distribution along the river courses with geological and geomorphological features reveals that changes in sinuosity values reflect neotectonic features. Therefore, the neotectonic origin of some morphostructural elements mapped earlier is supported by the sinuosity analysis. The change of river sinuosity at the boundaries of major tectonic elements of Miocene age indicates and supports the idea of reactivation of these structures, as previously suggested by seismic data interpretation (Szafián et al., 1999; Tari, 1994; Rumpler and Horváth, 1988).

This technique only reveals fault segments with vertical movement. However, large-scale river channel deflections highlight strike-slip dominated faults. Their existence is not surprising, in this tectonically complex boundary area between the Eastern Alps, Pannonian Basin and Western Carpathians.

This investigation considerably improved our tectonic-geomorphological understanding of the W Pannonian Basin and the location of active faults, providing a link between the small-scale hydrological processes and the tectonic regime at the sub-basin scale. The importance of our contribution is the recognition of a pattern of active tectonic faults in a badly exposed, low-level area by the integration of calculated SI values of several separate rivers.

Acknowledgements

This work was part of projects OTKA T47104 and NK60445, supported by the Hungarian National Scientific Fund. Interpretation of seismic data was conducted in the framework of the Landmark-Geographix University Grant Program. This paper is also a contribution to the *TOPO-EUROPE – 4D Topography Evolution in Europe: Uplift, Subsidence and Sea Level Rise* project, supported by the European Science Foundation. AZ received financial support from the Stiftung Aktion Österreich-Ungarn (Osztrák-Magyar Akcióalapítvány) and the ÖAD (Österreichischer Akademischer Austauschdienst) as part of his doctoral work. Support from Project 536001 Karpatian Tectonics, funded by the Österreichische Mineralölveraltungsgesellschaft (OMV) is greatly acknowledged. ED has been partly supported by project H990-2004 of the Hochschuljubiläumsstiftung der Stadt Wien. We thank the two anonymous reviewers for their helpful and accurate work.

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16. Appendix C

Zámolyi, A., Salcher, B., Draganits, E., Exner, U., Wagreich, M., Gier, S., Fiebig, M., Lomax, J., Surányi, G., Diel, M., Zámolyi, F. (submitted to “International Journal of Earth Sciences”). Latest Pannonian and Quaternary evolution at the boundary between Eastern Alps and Pannonian Basin: New insights from shallow lake seismic and well data.

(submission date: 11th of April 2015)

Latest Pannonian and Quaternary evolution at the boundary between Eastern Alps and Pannonian Basin: New insights from shallow lake seismic and well data

Zámolyi, A.[1)], Salcher, B.[2)], Draganits, E.[3),4)], Exner, U.[5)], Wagreich, M.[3)], Gier, S.[3)], Fiebig, M.[6)], Lomax, J.[7)], Surányi, G.[8)], Diel, M.[9)], Zámolyi, F.[10)]

[1)] OMV Austria Exploration & Production GmbH, Protteser Strasse 40, 2230 Gänserndorf, Austria

[2)] Department of Geography and Geology, University of Salzburg, Hellbrunnerstrasse 34, 5020 Salzburg, Austria

[3)] Department of Geodynamics and Sedimentology, University of Vienna, UZA II, Althanstrasse 14, 1090 Vienna, Austria

[4)] Department of Prehistoric and Historical Archaeology, University of Vienna, Franz-Klein-Gasse 1, 1190 Vienna, Austria

[5)] Geological-Paleontological Department, Natural History Museum Vienna, Burgring 7, 1010 Vienna, Austria

[6)] Institute of Applied Geology, University of Natural Resources and Life Sciences Vienna, Peter-Jordan-Straße 70 1190 Vienna, Austria

[7)] Department of Geography, Justus-Liebig-Universität Gießen, Senckenbergstraße 1, 35390 Giessen, Germany

[8)] MTA-ELTE Geological Geophysical and Space Science Research Group 1117 Budapest, Pázmány P. s. 1/c, Hungary

[9)] 2120 Dunakeszi, Madách Imre u. 2/E, Hungary

[10)] Bautechnisches Institut, Karl Leitl Strasse 2, 4048 Puchenau bei Linz, Austria

Keywords: Upper Pannonian sediments, Quaternary geodynamic evolution, Pannonian Basin, Neusiedler See, Shallow lake seismics, Lake drilling, IRSL dating, Provenance

Abstract

The western margin of the Little Hungarian Plain is a key boundary and interference zone between the Eastern Alps, Pannonian Basin and Western Carpathians. Geological investigation of this low-relief area is hampered by the scarcity of outcrops and thus direct observation of sedimentological or tectonic features. This region is characterized by gentle hills, plateaus and depressions, of which several are covered by lakes – including one of Austria’s largest lakes, Lake Neusiedl. In this study we combine local data from shallow-lake drilling and seismic investigation with regional data from industrial seismics and core data to gain new insights into the latest Pannonian (Late Miocene) and Quaternary evolution of this area.

Shallow lake seismic data show the erosionally truncated Pannonian sediments dipping and thickening towards southeast, towards the modern depocenter of the Little Hungarian Plain near Győr. Overlying Quaternary fluvial sediments show a very similar thickening trend, thus indicating continuous subsidence. Strike-slip and extensional faults are common in the Pannonian deposits, but hardly visible in Quaternary sediments. Drill cores from locations along the lake seismic lines were analyzed concerning their lithostratigraphy, grain-size, mineralogy and heavy minerals and compared with outcrop samples from the surrounding plains and the nearby plateau, to decipher the complex tectonic and fluvial depositional history of the region.

Significant differences in the elevation of the top of Pannonian sediments between the surrounding plains and the plateau indicate post-Pannonian vertical tectonic movements creating the present morphology of the region. Luminiscence ages of samples from the Quaternary fluvial gravels on top of the Pannonian sediments indicate a significantly higher age compared to the gravels in the plain, suggesting ongoing tectonic subsidence and associated effects on river dynamics.

Introduction

The transition between the easternmost Alps and the Western Carpathians (Fig. 1) forms a complex deformation zone which has been active since the Miocene (Decker and Peresson, 1996; Szafián et al., 1999; Sperner et al., 2002; Fodor et al., 2005; Fodor, 2006; Kováč et al., 2006; Bada et al., 2007; Lenhardt et al., 2007, Székely et al., 2009, Brückl, 2011). In this period the tectonic processes in this area, resulting from the continental collision of the Eurasian and African plates (McClusky et al., 2003), are dominated by the lateral escape of crustal blocks (Ratschbacher et al., 1989) and extensional collapse (Sperner et al., 2002). Subduction retreat along the Carpathian mountain range, acting as a weak lateral boundary, provided the setting for the formation of two sedimentary basins in this area: the Vienna Basin and the Danube Basin (Fig. 1). A marked difference between these basins is their diverse character of basin formation. While the Vienna Basin is interpreted as a thin-skinned pull-apart basin (Royden, 1988), the Danube Basin is regarded as a back-arc basin that formed along low-angle normal faults in the Early Miocene (Tari, 1994, 1996; Kováč et al., 2011).

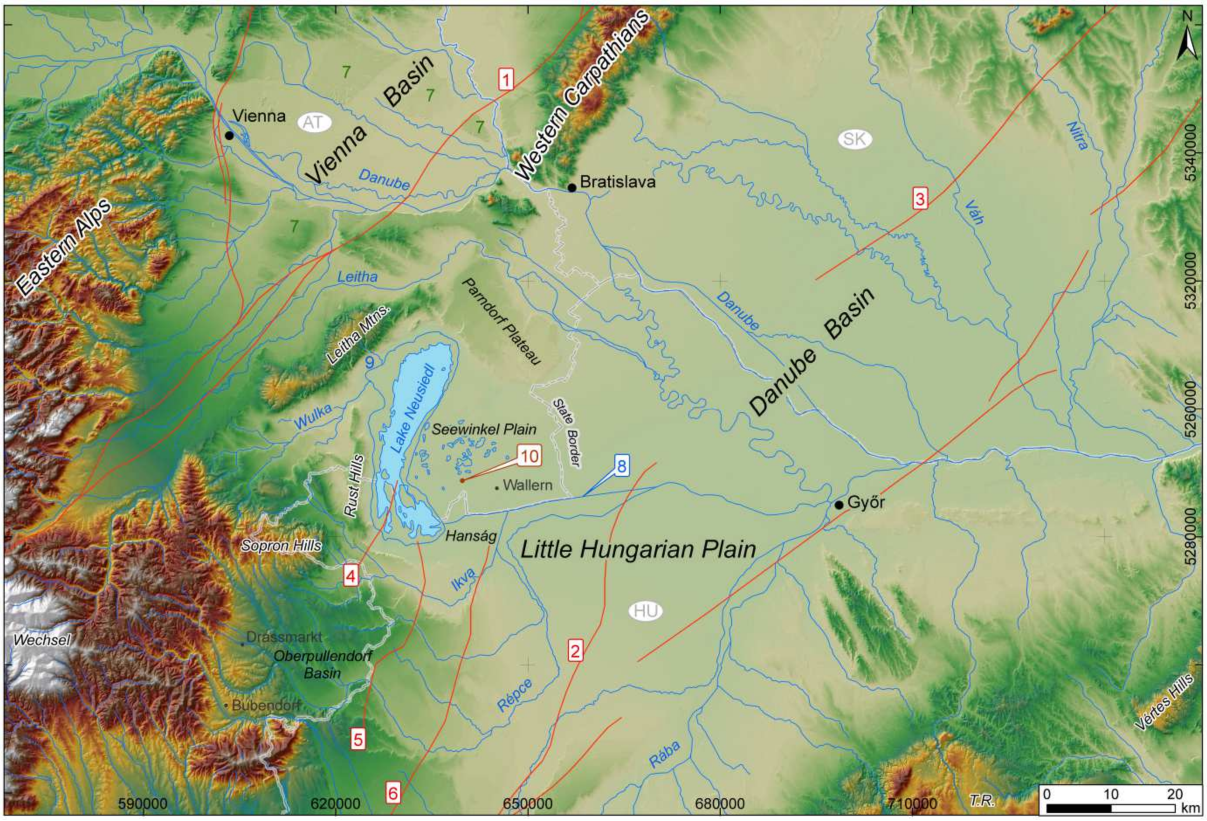

Fig. 1: Location of the study area in the transition zone between Eastern Alps and Western Carpathians. Note the distinct shape of the Parndorf Plateau and its location relative to the Lake Neusiedl and the Seewinkel Plain (background: SRTM digital elevation model; coordinate grid: UTM 33N). T.R.: Transdanubian Range, 1: Vienna Basin Transfer Fault, 2: Rába Fault, 3: Mojmirovice-Čertovica fault system, 4: Fertő Fault, 5: Ikva Fault, 6: Répce Fault, 7: River terraces with elevated margins, 8: Artificial drainage channel (Einser-Kanal or Hansági főcsatorna), 9: Angerbach rivulet, 10: Deepest topographic point of Austria (Huber-Bachmann et al., 2012).

Here, we focus on the impact of the younger, mainly Late Miocene to Quaternary evolution of the Danube basin which belongs to the Pannonian Basin system (Royden, 1988; Horváth, 1993). Our study area is located in the southern Danube Basin, an area referred to as the Little Hungarian Plain. It comprises the following regions (Fig. 1): (i) the Lake Neusiedl, one of Austria’s largest lakes; (ii) the Hanság area, a historical continuation of the Lake Neusiedl to the southeast; (iii) the Seewinkel Plain located to the east of the lake; (iv) the Parndorf

Plateau, an elevated area to the northeast of the lake; and (v) the Leitha Mountains and Rust Hills to the northwest and west of the lake.

Up to approximately 10 Ma, the study area was covered by the Lake Pannon, a brackish remnant of the Middle Miocene Paratethys Sea (Magyar et al., 1999). The depositional processes of the Lake Pannon and the Danube and its tributaries, considerable tectonic activity as well as erosive features show complex interactions preserved in the sedimentary record and landscape. So far, no consistent landscape evolution model of the western margin of the Little Hungarian Plain has been worked out yet – probably because of the complex tectonic and sedimentological history, the lack of well constrained age data as well as poor outcrop situation. An additional challenge is the difficult data access and evaluation due to its location in the border area of three different countries (Fig. 1) resulting in different languages, terminologies and relative age interpretations.

The aim of this study is the investigation and interpretation of the latest Pannonian (Late Miocene) and Quaternary evolution of the study area based on the integration of geomorphologic, geophysical, tectonic and sedimentological data. We address the following key points: (i) origin and age of the sediments of specific geomorphological units (e.g. Parndorf Plateau, Seewinkel Plain), (ii) geological processes controlling the modern sediment pattern and the formation of the Lake Neusiedl, (iii) timing of deformation and deposition of the main sedimentological units in the study area.

Geodynamic setting

The westernmost margin of the Danube Basin (Fig. 1) is probably still affected by the last major phase in the evolution of the Pannonian Basin system, an ongoing NE-SW compression (Horváth, 1995). This deformation results in subsidence in the central part of the basin and uplift at its flanks dividing the basin into elevated areas and depressions (Rumpler and Horváth, 1988; Horváth, 1995; Gábris and Nádor, 2007; Kováč et al., 2011). The Little Hungarian Plain was filled up with Pannonian (Upper Miocene) sediments during the post-rift subsidence phase of the Pannonian Basin (Magyar et al., 1999 and Magyar et al., 2013). Due to this on-going subsidence the thickness of the Pannonian and Quaternary sediments (Fig. 2) increases towards the East in the study area (Tauber, 1959a; Joó, 1992; Timár and Rácz, 2002). This Quaternary local subsidence occurs along partly reactivated strike-slip fault systems, such as the Rába-Mojmirovice-Čertovica fault system in the Danube Basin and the Vienna Basin transfer fault system in the Vienna Basin (Fig. 1) (Lenhardt et al., 2007; Székely et al., 2009; Zámolyi et al., 2010; Salcher et al, 2012). Additionally to local subsidence Pliocene to Quaternary long wavelength vertical movements (Decker and Peresson, 1996; Horváth and Cloetingh, 1996; Wagner et al., 2010) affect also the Danube Basin (Ruszkiczay-Rüdiger, 2007). As a consequence uplifted areas with incised river valleys can be found next to sediment-filled lowlands at the eastern margin of the Danube Basin (Ruszkiczay-Rüdiger et al., 2005). Here, the incision of the Danube into the Transdanubian Range and Vértes Hills (see Fig. 1 for location) provides a concise time marker for the onset of the Pliocene to Quaternary drainage pattern and landscape evolution of the Danube Basin after the regression of the Lake Pannon. Based on compiled geochronologic data (Ruszkiczay-Rüdiger et al., 2005) the onset of incision is estimated to 1.8 Ma before present. From this time onwards fluvial deposition and erosion of the Danube and its tributaries (Szádeczky-Kardoss, 1938; Küpper, 1955; Grill et al., 1968) contributed considerably to the landscape evolution.

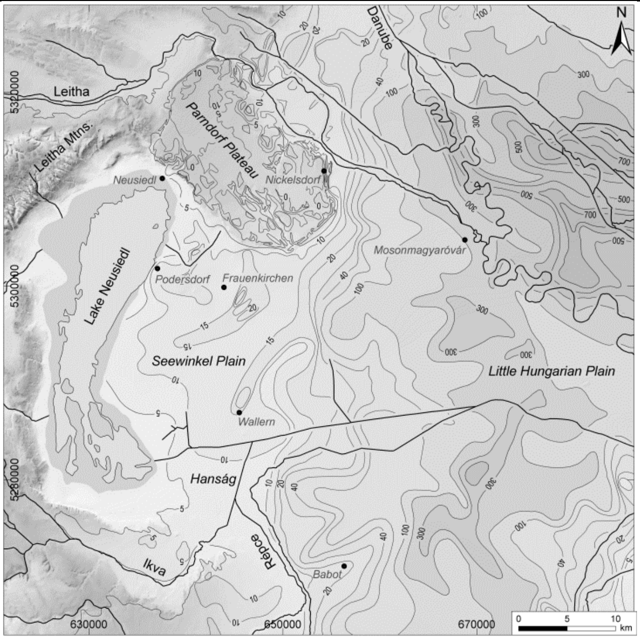

Fig. 2: Thickness of Quaternary sediments in the study area, which comprise mostly fluvial gravels, occasionally overlain by loess deposits of various thickness (e.g at Mönchhof). While the gravel deposits continuously increase in thickness towards the East in the Seewinkel Plain and adjacent Little Hungarian Plain, their thickness on the Parndorf Plateau shows no trend of thickening in any direction. This sediment geometry supports the view that the plateau was an elevated area before significant subsidence in the Little Hungarian Plain started. Note that significant thickening can only be observed East of the line Nickelsdorf – Wallern. Data have been compiled from Tauber (1959a), counterflush well data (Bohrarchiv, Austrian Geologic Survey), Fahrion (1944), DANREG 2000 project (Scharek et al., 2000b), Don (1991, 1993).

River terraces within the Vienna Basin are highly influenced by neotectonic faulting (Decker et al., 2005; Hinsch et al., 2005; Beidinger and Decker, 2011; Beidinger et al., 2011) forming tilted and vertically displaced terraces with distinct morphological scarps (Fig. 1). Active tectonics in the Little Hungarian Plain is also suggested from vertically separated gravel deposits of similar relative age (Schnabel, 2002) that cover a high-relief topography of Pannonian sediments. Ongoing tectonic activity in this area influences the drainage pattern and channel morphology of local tributaries (Zámolyi et al., 2010), as well as micro topographic features (Székely et al. 2009).

The most prominent deformation structures of the Little Hungarian Plain consist of NE-SW trending normal faults that initiated during the Karpatian (Early Miocene) by the reactivation of Cretaceous thrust faults (Tari, 1996). Several of these faults (e.g. Fertő-, Rába-, Ikva- and Répce Faults, Fig. 1) were traced on seismic sections in the area south of the Lake Neusiedl region (Tari, 1994; Tari, 1996; Szafián et al., 1999). The direction of extension in the Karpatian was ESE-WNW and rotated towards NNE-SSW in the Badenian (Middle Miocene; Fodor, 1996). Present extension in the area is oriented NNW-SSE and is a consequence of the inversion phase of the Pannonian Basin (Horváth, 1995). This phase started in post Sarmatian times (Late Miocene, i.e. younger than 11.63 Ma) and is still on-going with small variations in intensity and extension direction (Csontos et al., 1992; Bada et al., 2007). Geological maps (Scharek et al., 2000a; Schnabel, 2002) display a NE-SW- and NW-SE-oriented fault pattern in the Lake Neusiedl region (Fig. 3). Some of these faults can be linked to seismic activity (Lenhardt et al., 2007; Tóth et al., 2007) and some linear geomorphologic features probably are the expression of neotectonic processes (Székely et al., 2009). Normal faults with E-W and NW-SE extension have been observed in Badenian sediments and basement rocks of the Rust Hills (Spahic et al., 2011; Rath et al., 2011; Häusler et al., 2014) and at the SW margin of the Leitha Mountains (Fodor, 1991).

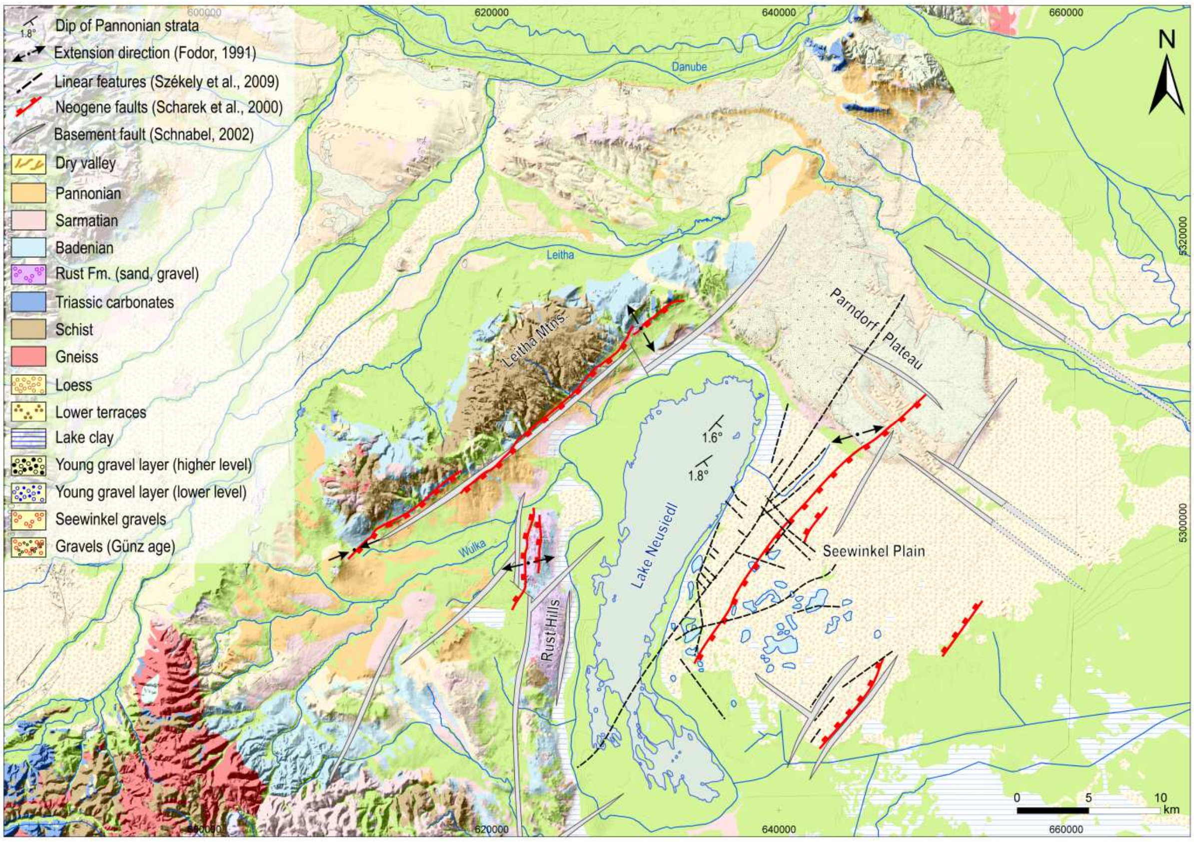

Fig. 3: Geological overview map of the study area after Schnabel (2002). According to this map the Parndorf Plateau is covered by three different gravel deposits: Günz age gravels and young gravels of the higher and the lower level. In contrast, the gravel layer deposited on the Seewinkel Plain is shown as a homogeneous unit. Basement and Neogene faults are arranged in a NE-SW- and NW-SE-oriented pattern. Black arrows indicate Late Miocene to Pliocene extension directions (Fodor, 1991). The black dashed lines represent tectono-geomorphic linear features derived from the analysis of an airborne laser scanned digital terrain model (Székely et al., 2009).

Geomorphologic overview

During the Pleistocene, the study area was not directly exposed to glaciation, which did cover major parts of the European Alps (e.g. van Husen, 1987; van Husen, 2011). However, fluvial, coarse grained deposits and periglacial features do suggest strong climatic impact on sedimentation and geomorphology during distinct cold periods (e.g. Heinrich et al., 2000; Salcher and Wagreich, 2010; Fábián et. al., 2014). Anthropogenic impact primarily involved the drainage of wetlands (Draganits et al., 2007).

In general the landscape of the study area (Fig. 1) can be divided into three main morphological categories: (i) elevated regions with a hilly relief comprising crystalline basement rocks and their Mesozoic cover, (ii) fluvial terraces (plateaus), and (iii) depressions (partly filled with lakes).

Elevated regions

The first category includes the Leitha Mountains (Leithagebirge), the Rust Hills (Ruster Hügelland), and further to the south, but still of some importance for this study, the Sopron Hills (Ödenburger Gebirge). The Leitha Mountains are a NE-SW trending, 33 km long and approximately 10 km wide, hilly landscape, separating the Vienna Basin from the Lake Neusiedl area (Fig. 1). Elevations range between 118 m to 484 m above sea level (a.s.l.), rising abruptly from the surrounding low-relief areas. Geologically, the Leitha Mountains represent a NE-SW striking horst comprising Lower Austroalpine schists, gneisses and amphibolites overlain by Triassic quartzites and dolomites (Schnabel, 2002). At the rim of the Leitha Mountains these metamorphic rocks are covered by Badenian to Sarmatian clastic sediments and limestones (Pistotnik et al. 1993; Wiedl et al., 2014).

The Rust Hills form a narrow, N-S trending, 22 km long, <3.5 km wide ridge (118-283 m a.s.l.) at the western boundary of the Lake Neusiedl (Fig. 1). They comprise schists, gneisses and amphibolites of the Lower Austroalpine tectonic unit that are almost completely covered by Karpatian to Sarmatian clastic sediments and limestones (Fuchs, 1965; Pistotnik et al., 1993). Analogous to the Leitha Mountains, the Rust Hills are also regarded as a tectonic horst, bordered by N-S trending normal faults which affect both the Neogene cover and the basement rocks (Fuchs, 1965; Scheibz, 2010; Spahic et al., 2011; Häusler et al., 2014).

The Sopron Hills (Fig. 1) represent the easternmost outcrop of Eastern Alpine rocks mainly bearing Lower Austroalpine gneisses and mica schists, covered by Tertiary sediments at the rim (Küpper, 1957). Topographic elevations range between 250 m and 520 m a.s.l. The metamorphic rocks as well as the Neogene cover rocks in the Sopron Hills are intensely deformed by brittle faults. Most dominant are NW-SE striking high angle faults, which probably formed as dextral strike slip faults, later reactivated as normal faults; another important fault set are E-W striking high-angle faults (Draganits, 1996). Both fault orientations are still reflected in the drainage system of the Oberpullendorf Basin. These brittle faults have been documented best in the former coal mines in the western part of the Sopron Hills at Ritzing and Brennbergbánya. Mining documentation shows the abundance of normal faults in Upper Miocene sediments in this area, with displacements of up to ca. 80 m (Kisházi and Ivancsics, 1977).

Fluvial terraces and plateaus

The Parndorf Plateau (Parndorfer Platte, 130-184 m a.s.l.) represents a fairly even surface comprising Pannonian fine clastic sediments with a thin cover of fluvial deposits. It is elevated about 25-45 m above the surrounding lowland (Fig. 1) and gently dips towards the SE (from c. 184 m in the NW, to 144 m a.s.l. in the SE) following the overall trend of the Pannonian sediments in the Little Hungarian Plain (Lipiarski et al., 2001). Towards the NW, the Parndorf Plateau has a narrow connection to terraces of similar altitude in the Vienna Basin, while in all other directions it drops along steep slope breaks (commonly >11°) towards the surrounding Upper Pleistocene to Holocene surfaces (Fig. 3). Based on the elevation of fluvial sediments, Szádeczky-Kardoss (1938) considered the Parndorf Plateau as the direct continuation of the River Danube´s highest (and thus oldest) terrace level in the Vienna Basin, the Laaerberg Terrace. Tauber (1959b) corroborated Szádeczky-Kardoss’s (1938) considerations concerning the provenance and Early Pleistocene age of the gravel layers of the Parndorf Plateau, based on the relatively high content of garnet in the heavy mineral fraction compared to lower terrace levels. Noticeable geomorphological features are straight, NW-SE trending dry valleys dissecting the southeastern part of the plateau (Fig. 4).

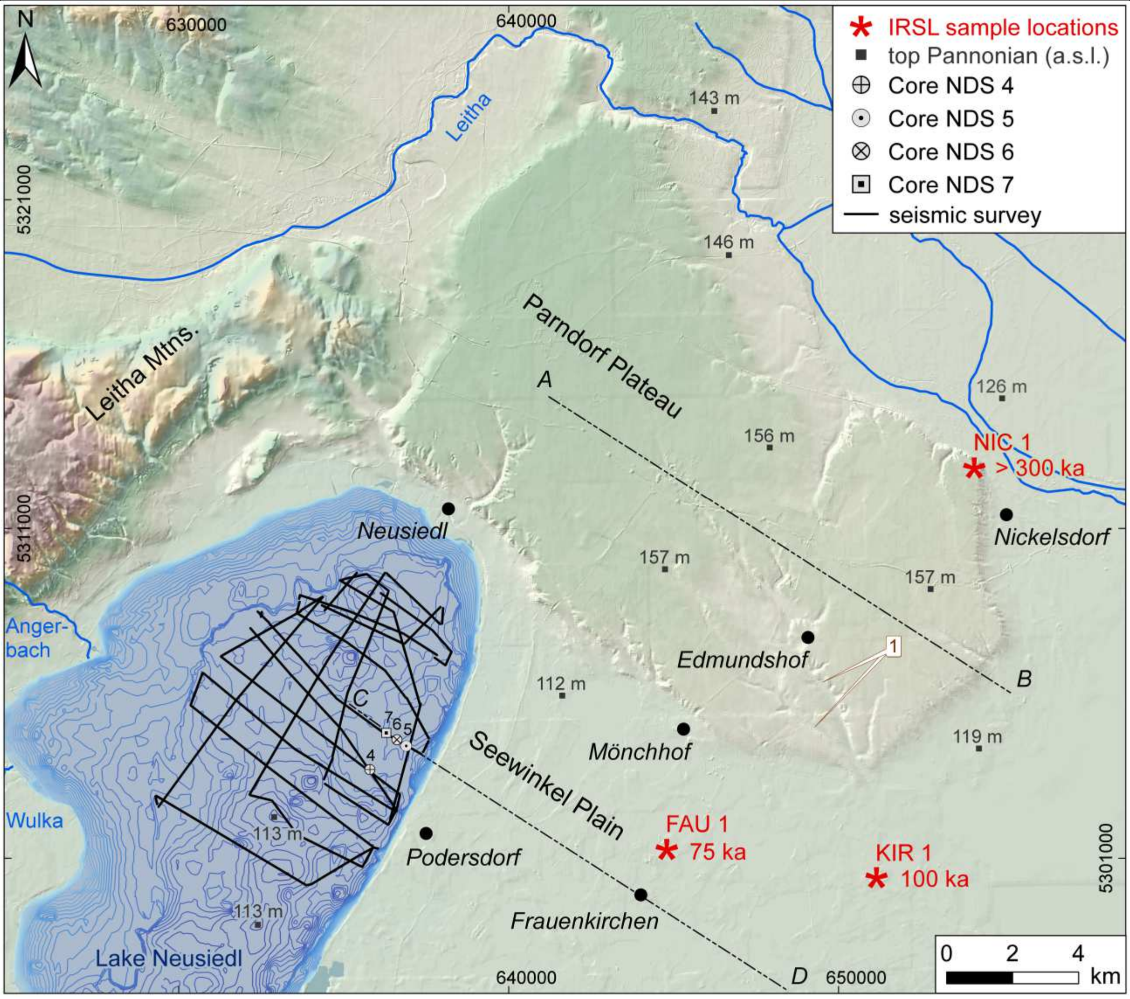

Fig. 4: Layout of lake seismic measurements and locations of IRSL samples. The profile lines A-B and C-D refer to profiles in Fig. 11. The contour lines of the lake bottom (after Bácsatyai et al., 1997) are shown in blue and have an interval of 0.1 m. The deepest point of the lake bottom lies at 113 m a.s.l. Background: SRTM digital elevation model; coordinate grid: UTM 33N. 1: Dry valleys in the southeastern part of the Parndorf Plateau.

The Seewinkel Plain (114-130 m a.s.l.) is a very low relief area, bounded by the Lake Neusiedl to the west, by the Parndorf Plateau to the northeast and by the Hanság depression to the south (Fig. 1). Similar to the Parndorf Plateau, the area is characterized by fluvial deposits on top of Pannonian sediments (Fahrion, 1944; Tauber, 1959a; Tauber, 1959c; Häusler, 2007). Gravels pinch out towards the west and diminish around the eastern shore of Lake Neusiedl (Fig. 2 and 3). The thickness of the Quaternary deposits increases towards the east to up to 25 m at the boundary to Hungary (Fig. 2; Tauber, 1959a), following the same thickening trend as the underlying Pannonian sediments (Fuchs and Schreiber, 1985).

Depressions

The Lake Neusiedl (Neusiedler See or Fertő-tó, Fig. 1), situated at the Austrian-Hungarian border, currently has an area of approximately 285 km^2 (Schmidt and Csaplovics, 2011; Király and Márkus, 2011) and a lake level of 115.5 m a.s.l. (Bácsatyai et al., 1997). The water level of the lake varied considerably in the past with a historical maximum around 117.7 m a.s.l. (Draganits et al., 2007). The entire lake is very shallow (1-1.7 m maximum, Fig. 4), with the deepest part of the lake bottom at around 113 m a.s.l. (Bácsatyai et al., 1997). The lake has no natural outlet, but for about 100 years the water level has been regulated by an artificial drainage channel (Fig. 1). At present the main tributaries are the Wulka River and to a very limited extent the Angerbach rivulet (Fig. 1). The lake is surrounded by an extensive reed belt except for the eastern shore. According to Tauber and Wieden (1959) hardly any recent sedimentation takes place in the open water area of the lake. Sediments are rather transported from the open water area to the reed areas to build up freshwater sapropel of up to several decimeters in thickness. Fine-grained sediments of Pannonian age underlie the freshwater sapropels (Tauber, 1959a and this study). The deepest part of the lake forms a roughly NNE-SSW depression ending close to the southernmost tip of the Rust Hills (Bácsatyai et al., 1997).

The Hanság depression is a low-lying, extremely flat area bounded by the Seewinkel Plain to the north and by the Lake Neusiedl to the west (Fig. 1). This area was part of the Lake before anthropogenic drainage and formed a continuous, L-shaped water body with the modern Lake Neusiedl (Draganits et al., 2007).

Methods

In order to investigate the geology and depositional history of the study area, we applied a suite of geophysical and geological methods: (i) shallow lake seismic measurements and interpretation, (ii) lake drilling and sedimentological characterization of the drill cores, (iii) acoustic velocity measurements of the sediments from the drill cores, (iv) heavy mineral investigations and (v) infrared stimulated luminescence (IRSL) dating. The data were consecutively analyzed together with outcrop descriptions and existing information compiled from the literature and scientific reports (including provenance studies, bore logs and industrial seismics).

Shallow lake seismic data

We use shallow lake seismic data to study the stratigraphy and deformation structures in a study area with extremely low relief, poor outcrop situation and a large lake. Lake seismic data were acquired in September 2008 during an international geophysical field course (Timár et al., 2009) using an IKB-Seistec™ single-channel boomer source (1 to 10 kHz) equipped with eight hydrophones arranged vertically in a cone receiver having a fixed offset. The grid of lake seismic measurements covers the NE part of Lake Neusiedl with a spacing of approximately 1 km (Fig. 4). The orientation of the grid has been planned orthogonal to a suspected NE-SW trending fault system within the lake (Tauber, 1959d; Tollmann, 1985; Schönlaub, 2000; Häusler, 2007) as well as to faults of similar orientation localized in tectonic-geomorphological studies (Székely et al., 2009). The seismic navigation file was created by a GPS tracking device measuring every four seconds. Processing of the seismic data in the time domain included the removal of the low-frequency trend, correction of minor static shifts due to the motion of the boat and equipment by waves, and the removal of artifacts due to the impedance contrast between water and the lake bottom. Changes in the speed of the boat towing the source/receiver arrangement can cause significant changes in the geometry of the reflectors (e.g. flattening due to decreasing speed, Tóth, 2003). However, the studied sections are not affected by such velocity effects, because the speed of the boat in these sections was very constant. Therefore, corrections in this regard were not considered to be essential and have not been carried out. Lake seismic data were interpreted in the time domain using the Petrel standard seismic interpretation software (www.slb.com). Characteristic marker horizons were mapped in time domain.

In order to be able to correlate the sedimentary bedding interpreted on the lake seismic sections with the stratigraphy at the Seewinkel Plain, industrial seismic sections (OMV, 1970) were combined with counterflush (CF) well log data (Fahrion, 1944; provided by the drill core archive, Austrian Geological Survey). The CF wells were placed systematically across the entire Seewinkel Plain during May 1943 and April 1944. The results of these wells that were part of a general prospection in the area were documented in a report by Fahrion (1944). In the present paper these wells were used for the evaluation of the Quaternary thickness map of the region (Fig. 2).

Lake drilling and drill core sedimentology

Interpretation of lake seismic data helped to locate key sites for a lake drilling campaign, which was conducted in 2009 in cooperation with the company UWITEC (www.uwitec.at). A floating aluminum sampling platform was used with a base plate placed on the bottom of the lake to ensure stable positioning and drilling. Sediment cores were retrieved by a “Niederreiter” piston corer applying rising and dropping hammer action. The bolt tip allowed for continuous sampling and minimized the effect of crumbling sediment. The cores were taken in 1 m long parts with core diameters of 9 cm or 6 cm. After acquisition, the cores were stored within metal cylinders and later extruded by a hydraulic pump into Plexiglas pipes. Drill cores were sampled in the eastern part of the lake near the village of Podersdorf, reaching a maximum depth of 3 m (Fig. 4).

Documentation of the drill cores was based on standard sedimentologic methods including lithofacies characterization, analysis of fossil content, color description by Munsell color charts, grain size measurements using a Sedigraph ET5100 and heavy mineral analysis of samples from given locations in the core. Table 1 summarizes the sampling locations.

Bulk and clay mineralogical compositions were established by x-ray diffraction using a PanalyticalX’Pert PRO diffractometer (CuKα radiation, 40 kV, 40 mA, step size 0.0167, 5 s per step). The bulk samples were analyzed as oriented powders. The clay fraction (< 2 µm fraction) was separated by sedimentation and analyzed as oriented clay films on glass slides.

core no.	sample no.	from cm	to cm	water loss [%]	grain size [%] sand	silt	clay	TOC [%]
NDS-4-1	NDS-4-1-1	4.0	6.0	33	13.7	34.7	51.6	1.38
	NDS-4-1-2	9.0	12.0	28	48.1	18.9	33.0	1.19
	NDS-4-1-3	41.5	45.0	20	74.5	16.2	9.3	0.47
	NDS-4-1-4	78.0	81.0	18	96.9	2.5	0.6	0.22
NDS-4-2	NDS-4-2-4	8.0	14.0	17	92.3	6.1	1.6	0.14
	NDS-4-2-3	18.0	22.0	20	90.8	7.6	1.5	0.26
	NDS-4-2-2	42.0	45.0	20	93.9	5.3	0.8	0.30
	NDS-4-2-1	93.0	97.0	17	94.6	4.7	0.7	0.26
NDS-4-3	NDS-4-3-1	14.0	17.0	17	93.4	5.7	0.9	0.21
	NDS-4-3-2	63.0	66.0	24	32.1	60.8	7.1	0.41
	NDS-4-3-3	82.0	84.5	26	3.7	76.2	20.1	0.76
NDS-5-1	NDS-5-1-1	17.1	20.0	20	0.7	62.5	36.9	0.34
	NDS-5-1-2	51.5	54.0	24	3.0	52.0	45.0	0.29
	NDS-5-1-3	77.5	80.7	19	29.1	49.1	21.8	0.16
	NDS-5-1-4	88.3	91.3	19	5.2	72.6	22.3	0.26
NDS-5-2	NDS-5-2-1	2.0	7.5	19	46.2	25.2	8.4	0.29
	NDS-5-2-2	24.7	28.0	20	21.7	67.6	10.7	0.18
	NDS-5-2-3	41.8	44.6	22	3.3	70.7	26.0	0.29
	NDS-5-2-4	78.5	81.0	21	0.1	76.6	23.4	0.38
NDS-5-3	NDS-5-3-1	12.5	16.0	22	0.0	63.7	36.3	0.34
	NDS-5-3-2	29.3	33.0	20	0.5	57.9	41.6	0.30
	NDS-5-3-3	40.0	43.3	16	11.6	50.8	37.6	0.18
	NDS-5-3-4	55.0	59.5	17	1.7	68.1	30.2	0.23
NDS-6-1	NDS-6-1-1	0.4	3.5	17	66.8	4.5	3.6	0.97
	NDS-6-1-2	19.5	21.7	33	6.8	32.5	60.7	2.17
	NDS-6-1-3	54.5	57.2	20	1.0	67.1	31.9	0.35
	NDS-6-1-4	82.0	84.5	22	14.3	32.5	53.2	0.15
NDS-6-2	NDS-6-2-1	11.2	15.5	22	17.3	34.5	48.2	0.13
	NDS-6-2-2	46.0	50.2	17	22.2	47.5	30.4	0.36
	NDS-6-2-3	86.2	90.0	17	65.0	26.7	8.4	0.14
NDS-6-3	NDS-6-3-1	12.0	16.5	18	56.4	35.0	8.6	0.23
	NDS-6-3-2	33.0	37.3	15	58.3	29.0	12.7	0.47
	NDS-6-3-3	44.5	49.0	18	17.4	66.5	16.1	0.30
	NDS-6-3-4	57.0	60.7	21	7.0	70.0	23.0	0.28
	NDS-6-3-5	71.0	74.8	19	11.8	61.0	27.2	0.32
	NDS-6-3-6	79.5	83.0	18	29.0	48.1	22.9	0.12
	NDS-6-3-7	86.5	90.0	16	35.1	53.7	11.2	0.14
NDS-7-1	NDS-7-1-1	11.0	14.0	20	1.8	36.5	61.7	n/a
	NDS-7-1-2	16.0	20.0	19	25.8	39.9	34.3	n/a
	NDS-7-1-3	41.0	44.0	19	43.2	40.0	16.8	n/a
	NDS-7-1-4	55.5	59.0	21	12.5	63.7	23.8	n/a
	NDS-7-1-5	86.0	92.0	19	50.6	44.7	4.6	n/a
NDS-7-2	NDS-7-2-1	3.5	6.2	24	9.4	73.5	17.1	0.22
	NDS-7-2-2	9.5	12.5	24	47.1	44.4	8.5	0.19
	NDS-7-2-3	30.3	33.0	27	1.4	74.0	24.5	0.30
	NDS-7-2-4	47.0	50.0	26	0.4	79.4	20.2	0.29
	NDS-7-2-5	70.5	73.0	24	4.3	80.5	15.2	0.19
	NDS-7-2-6	82.5	86.0	21	45.5	51.4	3.1	0.13
NDS-7-3	NDS-7-3-1	1.5	4.5	23	16.2	51.8	32.0	n/a
	NDS-7-3-2	9.0	11.5	23	7.9	84.4	7.6	n/a
	NDS-7-3-3	21.0	24.0	17	31.4	65.3	3.3	n/a
	NDS-7-3-4	28.5	32.0	21	25.3	71.1	3.6	n/a
	NDS-7-3-5	42.5	46.5	9	70.7	27.5	1.8	n/a
	NDS-7-3-6	50.5	52.5	21	8.4	83.4	8.2	n/a
	NDS-7-3-7	58.0	61.0	18	31.5	63.0	5.5	n/a
	NDS-7-3-8	67.0	69.5	21	7.4	77.7	14.9	n/a
	NDS-7-3-9	77.0	80.0	21	2.7	84.5	12.7	n/a
	NDS-7-3-10	90.0	93.0	18	39.0	57.7	3.2	n/a

Table 1: Overview of sample locations and results of grain size measurements from the lake drilling cores.

Acoustic velocity measurements

In order to obtain a detailed velocity profile to link seismic to well data, ultrasonic measurements were carried out on the core samples using a Proceq TICO ultrasonic instrument (measurement layout shown in Fig. 5). This device has two metal cylinders with approximately 6 cm diameter. One cylinder acts as an emitter of ultrasonic waves at 50 kHz, and the other records the travel time of the emitted waves. Proceq TICO uses the pulse velocity method to provide information on the uniformity of the investigated material. The measurements were carried out in a tandem arrangement (Gräfe, 2009) with a coupling paste. A standard material (concrete block) and the air temperature were measured at the beginning of each major measuring phase. Several measurement runs were conducted for the same section of core and the most robust velocity measurement was used for the depth to time conversion.

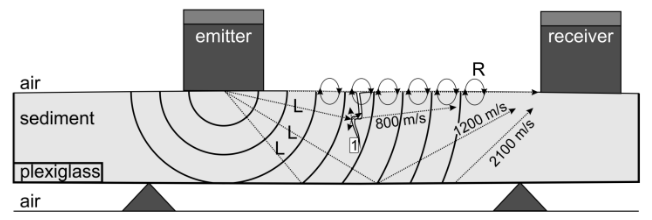

Fig. 5: Arrangement of acoustic velocity measurements showing the different set of acoustic velocities derived and their possible causes. Velocities measured depend on the type of wave involved. For simplicity, two types of waves are depicted: longitudinal wave (L) and Rayleigh wave (R). The acoustic axis of each cone was below the 1st critical angle, thus the longitudinal wave was recorded. Additional factors that lead to the recording of different velocities are scattering and absorption as well as noise due to the grain size of the sediment and air within the pore space (Gräfe, 2009). For the depth to time conversion the most robust velocity value derived from several measurement runs was used.

Heavy mineral investigation

The samples for heavy mineral analysis were derived from three different regions: (i) samples from the Upper Pannonian strata of the lake drilling cores near Podersdorf (Fig. 4), (ii) one sample from a sand body within the gravels of the Parndorf Plateau (NIC 1), and (iii) two samples from few dm thick sand lenses within the gravels of the Seewinkel Plain (FAU 1, KIR 1) (Fig. 4). The sand bodies within the sediments on the Parndorf Plateau and in the Seewinkel Plain are both embedded within the uppermost gravel layers. The sample NIC 1 was taken at 147 m a.s.l, 8 m below surface, FAU 1 was taken at 123 m a.s.l, 1.5 m below surface, and KIR 1 (Fig. 6c) was taken at 125 m a.s.l, 2 m below the surface. The fraction between 63 and 400 µm was used to investigate the heavy mineral assemblage of the samples. Heavy mineral separation and counting followed standard procedures (Mange and Maurer, 1992; Wypyrscyk et al., 1992).

For provenance analysis, our heavy mineral data were compared with those from the Seewinkel Plain (Husz, 1965), from Danube terraces (Csapó, 1998; Frasl, 1955), from the catchment of the Répce, also called Rabnitz (Nebert et al., 1980; Schocklitsch, 1962), and from Danube tributaries in the Vienna Basin (Szabó, 1961).

Infrared stimulated luminescence dating (IRSL)

Samples NIC 1, FAU 1 and KIR 1 (Fig. 4) were used for IRSL dating of sand bodies within the gravel layers on the Parndorf Plateau and in the Seewinkel Plain. NIC 1 was taken 10 m below the topographic surface of the Parndorf Plateau, FAU 1 was located 1.5 m and KIR 1 2 m below the modern topographic surface of the Seewinkel Plain. The samples were dated in the luminescence laboratory in Vienna using a single-aliquot regenerative-dose infrared stimulated luminescence protocol (Wallinga et al., 2000; Blair et al., 2005) of the potassium-rich, coarse grain feldspar fraction. Dose rates were determined by laboratory gamma spectrometry. The quartz grains were not suitable for OSL measurements due to very low strength of the luminescence signals.

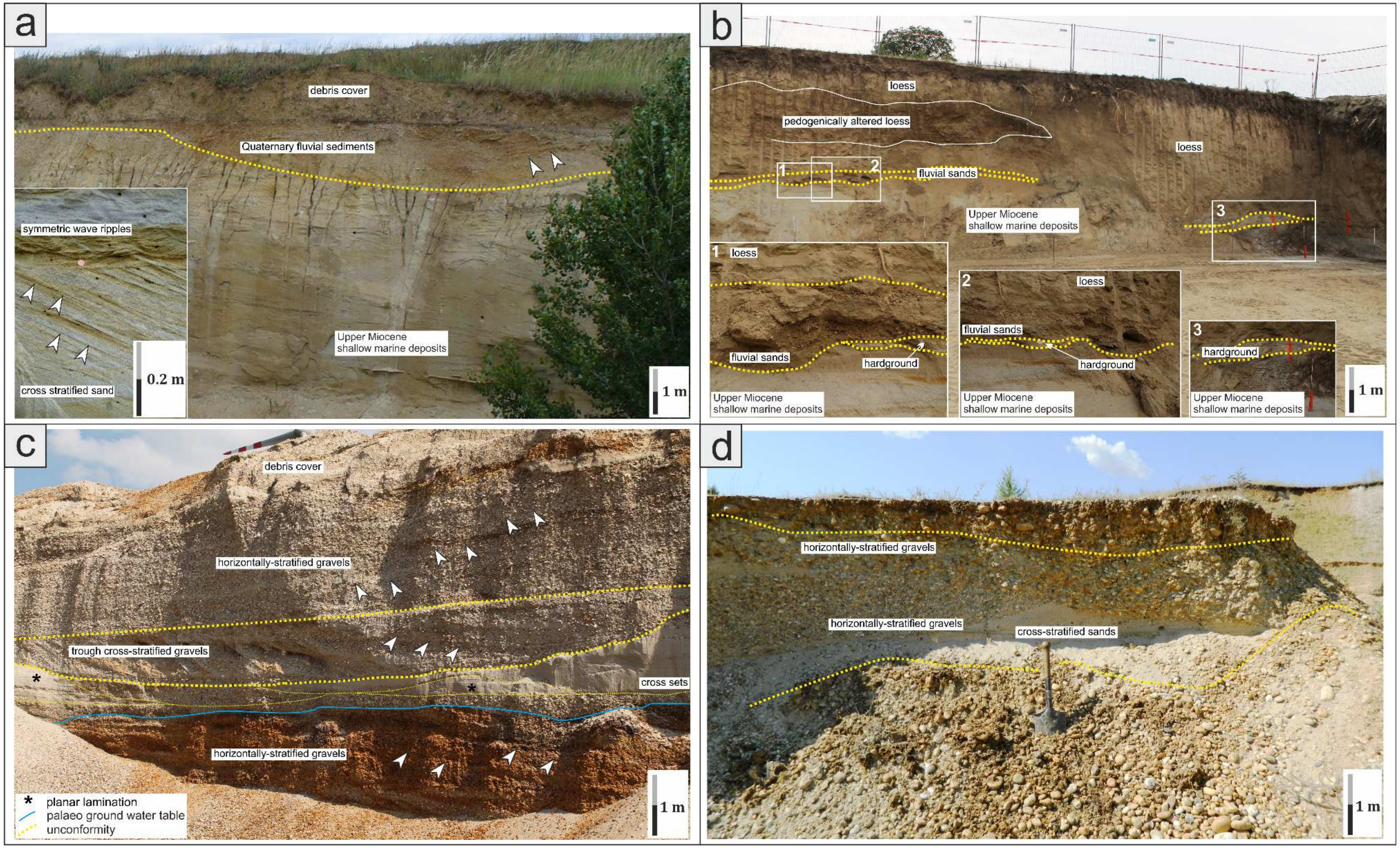

Fig. 6: Overview of outcrops in the study area. a) Quaternary channel incising in Upper Miocene strata at Edmundshof (Parndorf Plateau). Note the otherwise very thin Quaternary cover of the Upper Miocene shallow marine deposits. b) At Mönchhof, loess deposits are juxtaposed to the Upper Miocene shallow marine deposits. c) The outcrop at Halbturn (Seewinkel Plain) shows sand cross sets from which the luminescence sample KIR 1 was taken. d) Sample location for IRSL dating and heavy mineral analysis at the outcrop of Nickelsdorf. The sample NIC 1 was taken from the cross-stratified sands at the spade (10 m below the topographic surface

Infrared stimulated luminescence dating (IRSL)

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Results

Seismic interpretation

In total more than 110 km of seismic lines of the sediments beneath the Lake Neusiedl have been measured (Fig. 4). Using the acoustic velocity measurements, it was possible to convert drilling core depth in seismic time and thus correlate distinct sedimentary successions of the lake drilling cores with characteristic reflectors of the lake seismic sections (Fig. 7). The conversion of the thickness of the sedimentary layers from meters to seconds was done by stretching or squeezing the thickness value with the respective measured acoustic velocity. For layers that yielded no conclusive velocity measurement results (i.e. due to open cracks within the drill core) velocities from analogous layers were used.

A multitude of sedimentological and tectonic features could be documented in the lake seismic data, including (i) presence/absence of gravels, (ii) thickness variations, (iii) erosional truncation of beds, (iv) channel structures, (v) tilting of sedimentary bedding and (vi) normal faults (Figs. 7 and 8). However, the major fault postulated by Tauber (1959d) in the northern part of the lake could not be detected.

The dip of the Pannonian strata is generally low. No consistent tilting of the Pannonian beds could be observed in SW-NE orientation (Fig. 8a and 8c). In one section, oriented along the longitudinal axis of Lake Neusiedl, strata dip towards SW, away from the Parndorf Plateau (Fig. 8a). In a parallel section, 2 km to the SE, no significant tilting can be observed. In a perpendicular NW-SE section the sedimentary bedding dips towards the SE and terminates with a marked angular unconformity towards the bottom of the lake (Fig. 7b to c). The dip of reflectors in the close surroundings of the lake drilling locations was converted to true dip based on the results of the acoustic velocity measurements (see section 4.3). Accordingly, the dip data for the bedding in that area are around 2° towards SE (Fig. 3). Further to the west, the tilted bedding changes to a sub-horizontal undulating surface.
Across the entire area covered with seismic data, no indication of significant thickening of sediments was detected (Figs. 7 and 8). The top of the entire stack of Pannonian sediments displays erosional truncation and toplaps (Figs. 7 and 8). In some parts, especially close to the western shore, the sedimentary bedding is cut by channels of various sizes (Fig. 7c and 7d). Figure 7d shows indications of a point bar system. The seismic data together with the lake drilling samples, and Quaternary thickness maps from literature (Figs. 7 and 8, Fig. 2) support the observation that no significant Quaternary gravel layer exists in the northern part of Lake Neusiedl.

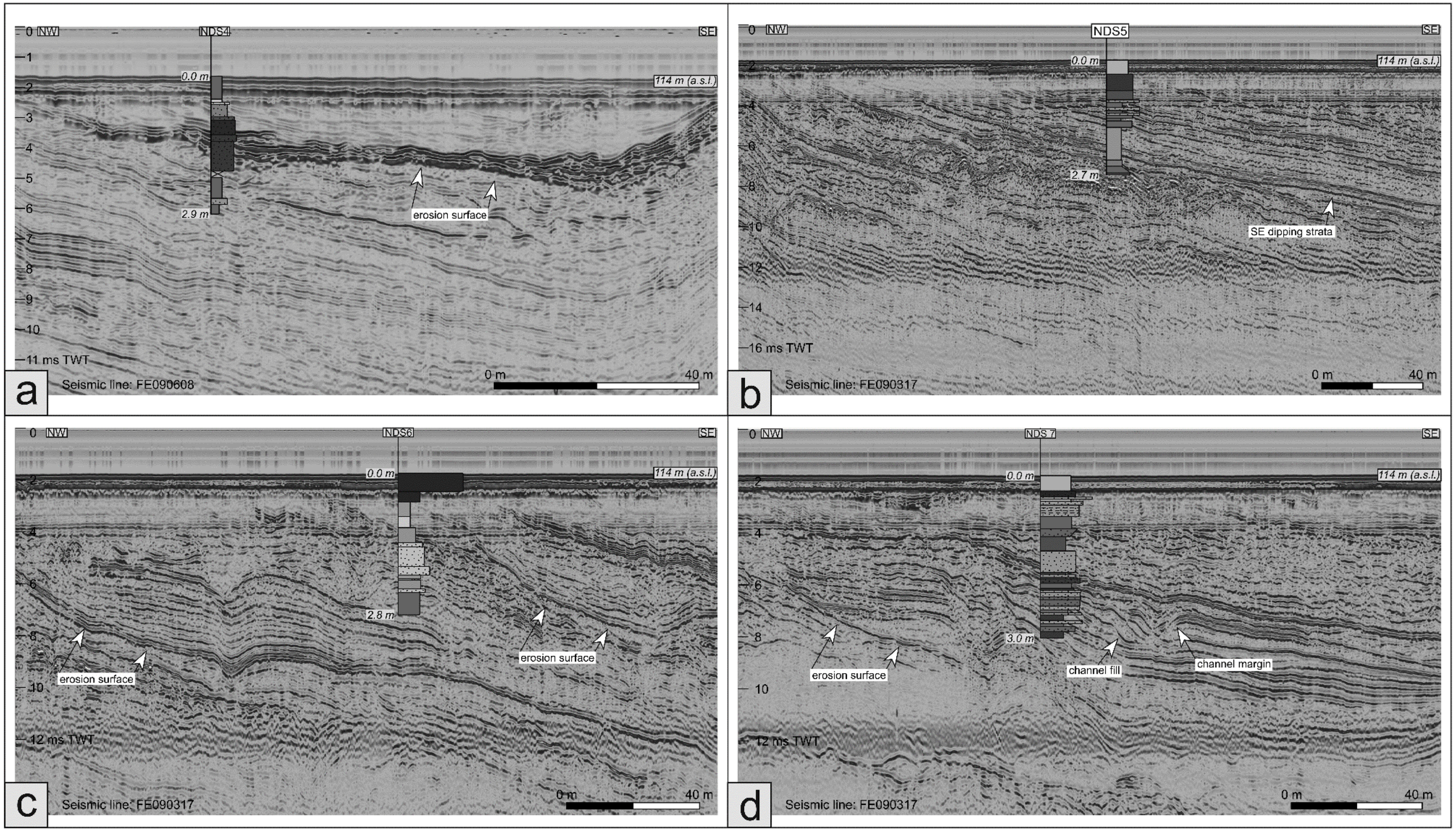

Fig. 7: Lake seismic sections in time domain shown with the correlated lake drilling sections. The correlation was done by depth to time conversion based on the results of the acoustic velocity measurements. Individual sections with uniform acoustic velocities were stretched or squeezed *en bloc*. The lithologic description of the lake drilling cores is shown in Fig. 9.

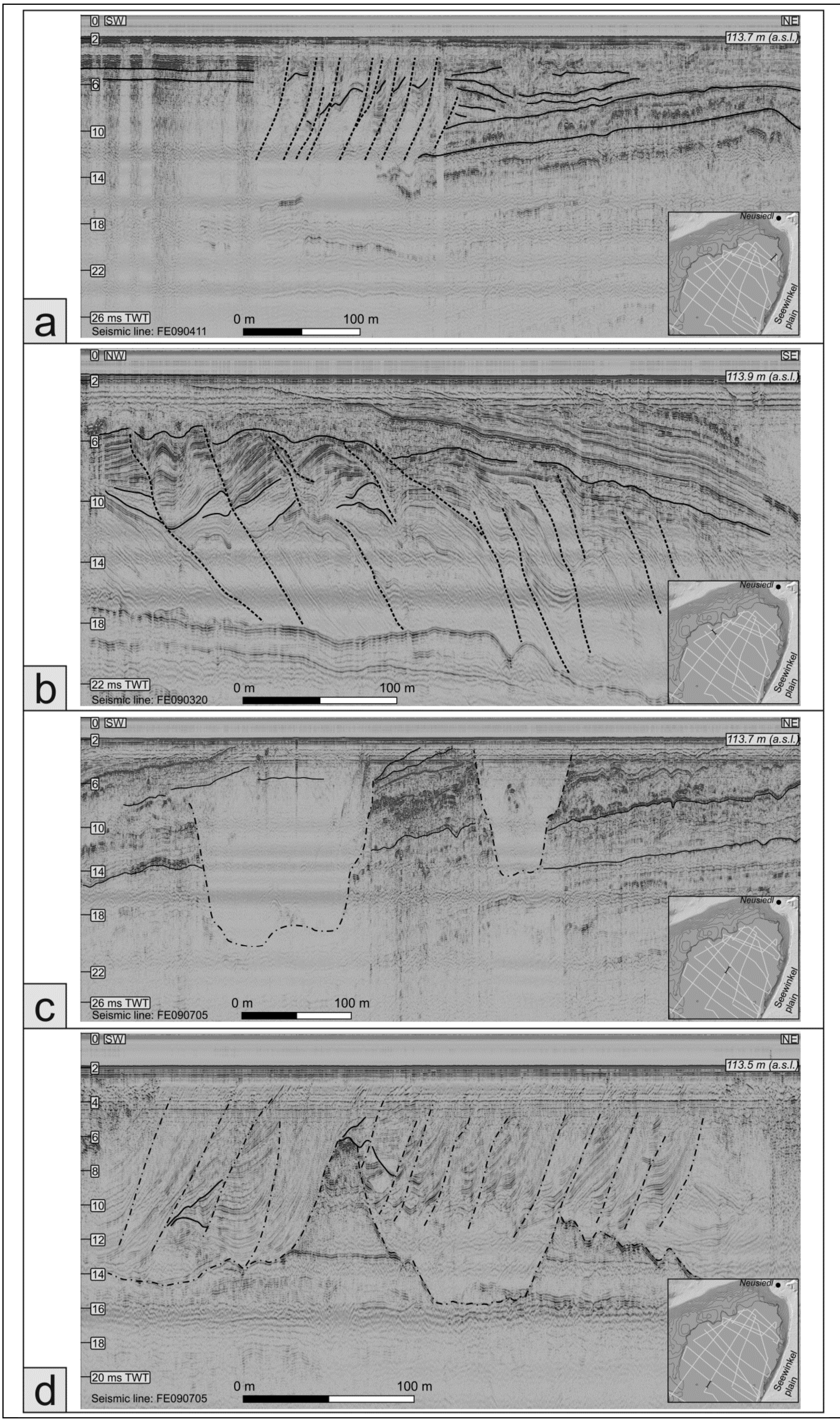

Fig. 8: Selected lake seismic sections showing key sedimentologic and tectonic features beneath the Lake Neusiedl. (a) No consistent tilting could be observed towards the southern margin of the Parndorf Plateau, only minor normal faults. (b) Similar minor faults were identified elsewhere in the lake. (c) Symmetric channels and (d) a point bar system indicate syn- to post-Pannonian fluvial erosion. Continuous lines indicate sedimentary bedding; dashed lines are interpreted faults and

Description of lake drilling cores

The four lake drillings at the lake bottom of Lake Neusiedl (NDS 4 – NDS 7) reached depths between 2.6 and maximum 3 m due to high compaction of the sediments. Table 1 provides an overview of the analyzed samples and summarizes the results of the grain size measurements. No significant amount of gravels could be observed in any of the drill cores, supporting the lack of significant gravel deposits within the lake as also observed in the lake seismic data. Paleontological analysis yielded no age indicative fossils in the sediments of the lake drilling cores (pers. comm. M. Harzhauser, 2011). However, the combination of seismic sections with well data and the calculation of true dip values as described in chapter 5.1 allowed the correlation of the reflectors onto adjacent onshore industrial seismic sections (OMV, 1970) as well as nearby CF well data in the Seewinkel Plain (Fig. 11). This correlation indicates a probable Late Pannonian age of the sediments beneath the lake bottom, which is also supported by our mineralogical findings and data by Szontagh (1904), Blohm (1974) and Fuchs and Schreiber (1985).

The mineralogy of drilling NDS 7-1 was investigated in detail. Eight samples were taken at 0-4 cm, 11-14 cm, 16-20 cm, 28-31 cm, 41-44 cm, 55.5-59 cm, 68-71.5 cm and 86-92 cm depth. The uppermost 20 cm of the drilling is dark gray, silty clay, followed by light gray to greenish clayey silt interbedded with layers of fine to medium sand. The bulk samples consist of muscovite, chlorite, quartz, feldspar (albite and microcline), calcite and dolomite throughout the profile. However, the uppermost two samples are slightly different, showing very broad calcite and dolomite peaks with a pronounced magnesium-calcite peak in between. Calcite and dolomite show better crystallization downhole, and the magnesium-calcite peak disappears at a depth of 60 cm. Furthermore, the upper two samples contain less feldspar than the rest. The clay fraction (< 2 µm fraction) gives even more evidence that the uppermost samples are different. Illite and chlorite are the dominant clay minerals until a depth of 20 cm. The deeper samples show, additional to illite and chlorite, pronounced peaks of smectite and some minor kaolinite.

The Pannonian deposits below the uppermost 20 cm display cm to maximum 1.3 m thick layers of clay to coarse sand, including isolated cm-sized gravel clasts (see Fig. 9 for detailed sedimentological logs). The colors range from grey to greenish grey; irregular orange oxidation spots are common. Sub-horizontal lamination is the only visible sedimentary structure. Relatively large detrital white mica flakes and low calcite contents are characteristic for the Pannonian sediments.

Heavy mineral data

The heavy mineral data represent assemblages from lacustrine Upper Pannonian strata (lake drilling cores) as well as from Quaternary fluvial sediments (Parndorf Plateau and Seewinkel Plain; Table 2).

	Zircon	Tur.	Rutile	Apatite	Garnet	Chloritoid	St.	Epidote	Chromium Spinell	Hbl. (grey)	Amp.	Sil.	Zoisite	Kyanite	Titanite	Hbl. (brown)	Biotite	
NDS 4-2-2	5	4	1	13	130	2	12	33	0	49	5	10	5	4	1	1	1	pcs.
	2	1	0	5	47	1	4	12	0	18	2	4	2	1	0	0	0	%
NDS 5-2-2	3	12	0	70	76	2	46	25	1	1	4	0	19	4	0	0	0	pcs.
	1	5	0	27	29	1	17	10	0	0	2	0	7	2	0	0	0	%
NDS 6-3-3	4	14	2	37	130	2	27	35	0	2	1	0	11	11	1	0	3	pcs.
	1	5	1	13	46	1	10	13	0	1	0	0	4	4	0	0	1	%
NDS 7-3-7	3	8	1	36	110	2	50	43	0	4	0	1	8	5	0	0	1	pcs.
	1	3	0	13	40	1	18	16	0	1	0	0	3	2	0	0	0	%
KIR 1	8	7	3	0	75	0	21	34	0	133	2	6	7	9	2	0	0	pcs.
	3	2	1	0	24	0	7	11	0	43	1	2	2	3	1	0	0	%
FAU 1	1	1	2	4	90	0	17	15	0	80	0	3	12	2	1	0	0	pcs.
	0	0	1	2	39	0	7	7	0	35	0	1	5	1	0	0	0	%
NIC 1	5	2	1	1	50	1	10	13	0	12	0	8	2	3	3	0	0	pcs.
	5	2	1	1	45	1	9	12	0	11	0	7	2	3	3	0	0	%

Table 2: Results of the heavy mineral analysis of the samples from the Late Miocene lake drilling cores (NDS 4 to 7) and from the Quaternary outcrop samples in the Seewinkel Plain (KIR 1, FAU 1) and Parndorf Plateau (NIC 1). Tur. = Tourmaline, St. = Staurolite, Hbl. = Hornblende, Amp. = Amphibole, Sil. = Sillimanite, pcs. = pieces.

Heavy mineral assemblages from the lake drilling cores (Table 2) show high abundances of garnet, apatite, staurolite and epidote. Some common heavy minerals include zircon, tourmaline, zoisite/clinozoisite, kyanite and transparent amphibole. This means that the heavy mineral spectrum is dominated by metamorphic minerals. Minerals of the ultrastable ZTR (zircon, tourmaline, rutile) group in regard to transport and weathering (Hubert, 1962) range up to a few percent in maximum (Fig. 10a). Chloritoide, chromium spinel, titanite, brown amphiboles are also present but in very low quantities. It is important to note that staurolite dominates over chloritoid and that sillimanite and kyanite are quite abundant. All in all, heavy minerals from a low- to higher grade metamorphic source area dominate the assemblages.

In the ternary plot (Fig. 10a) the 3 samples from the Quaternary gravels plot very near the Upper Pannonian samples, indicating either reworking or the same, mainly metamorphic hinterland.

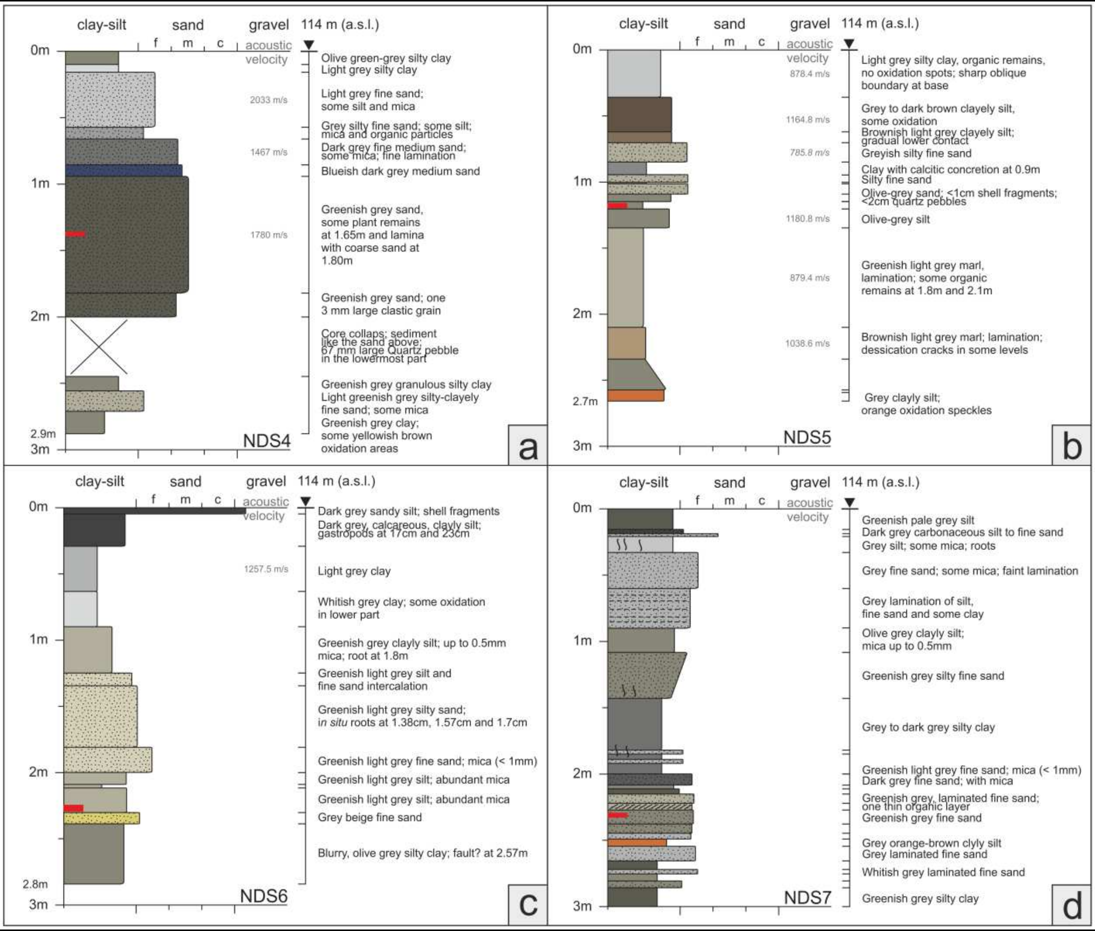

Fig. 9: Sedimentological logs of the lake drilling cores. Note that no significant gravel layer could be observed in the lake drilling cores. The red rectangle shows the sample locations for heavy mineral analysis. Results of the acoustic velocity measurements are indicated in light grey for each section.

The sample from the Parndorf Plateau (NIC 1) shows high abundance of garnet, staurolite, epidote. Green hornblende is also common, but in significantly lower amounts compared to the samples from the Seewinkel Plain (KIR 1, FAU 1). Additionally, zircon, tourmaline, fibrolithic silimanite, zoisite/clinozoisite, kyanite are common. Rutile, transparent amphibole, titanite and apatite are present in very low amounts (Table 2). In comparison with the Parndorf Plateau, the heavy mineral assemblages from the Seewinkel Plain (KIR 1, FAU 1) are dominated by green hornblende and garnet, and contain higher amounts of fibrolithic sillimanite and titanite, but less amphiboles (Table 2).

Plotting logratio-diagrams of the main heavy minerals epidote group and chloritoide versus garnet, as well as green amphibole versus staurolite (Figs. 10b and 10c) results in a separation of Quaternary samples from Pannonian samples. The amount of green hornblende is higher or equal to the amount of staurolite in the Quaternary samples, whereas in the Pannonian samples the amount of green hornblende is less or equal to the amount of staurolite, except for one sample.

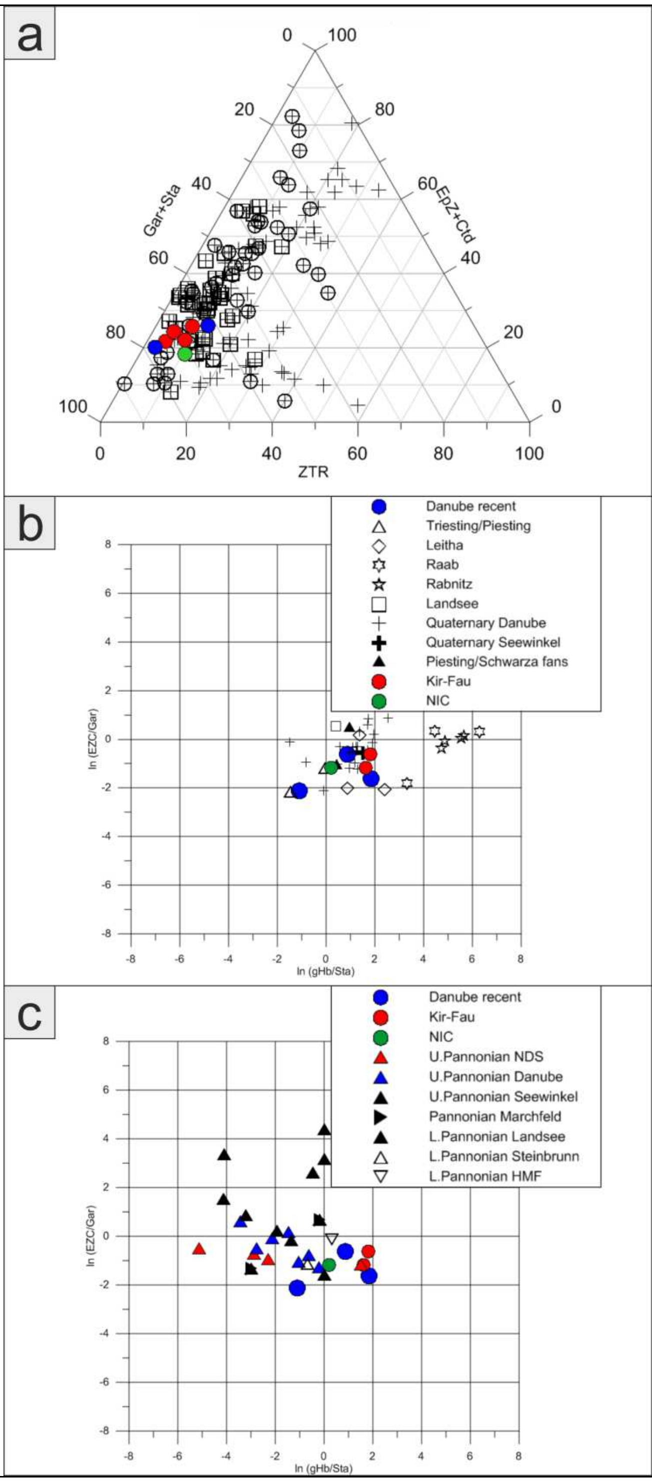

Fig. 10: Heavy mineral analysis diagrams showing a) A ternary plot ZTR (zircon, tourmaline, rutile) versus low-grade metamorphic (epidote group and chloritoide) versus higher-grade metamorphics (garnet and staurolite), b) logratio-diagram of the main heavy minerals epidote group+chloritoide versus garnet and green amphibole versus staurolite of recent samples and c) of Upper Miocene samples. Data of this paper (KIR 1, FAU 1, NIC 1 and NDS sampl es) are compared to data from literature (Frasl, 1955; Wieseneder and Maurer, 1958; Szabo, 1961; Schoklitsch, 1962; Husz, 1965; Fuchs, 1974).

Quaternary fluvial sediments of the Parndorf Plateau and the Seewinkel Plain

In the gravel layers of the Parndorf Plateau and the Seewinkel Plain three main lithofacies classes were recognized: (a) poorly sorted massive gravels, (b) laminated and cross bedded sands and (c) alternating open- and filled framework gravels.

The fluvial sediments of the Seewinkel Plain generally consist of loose, sandy, medium to coarse gravels often intercalated of cm to few dm thick and some couple of meters long bodies of medium to coarse sand. These sand bodies appear commonly laminated often showing thin stringers of pebbles. Gravel clasts appear generally rounded to well-rounded. Periglacial features within the fluvial sediments of the Seewinkel Plain are rare and represent involutions within the top decimeters. Loess deposits are virtually absent in the Seewinkel Plain. Paleo-groundwater levels were frequently observed in outcrops of the Seewinkel Plain at a few meters depth, represented by distinct, yellow to dark brown and blackish layers of grain-supported gravels (Fig. 6c).

Lithological logs from CF wells in the Seewinkel Plain highlight the clear lithological contrast between the fluvial Quaternary strata and the Upper Pannonian sediments (Fahrion, 1944). The Quaternary sediments comprise sand lenses and gravels with an average thickness of 10 m and with a thickening trend towards the East (Fig. 2). The CF wells encountered Upper Pannonian strata below the Quaternary but did not pass through the entire Pannonian strata. The Upper Pannonian is formed by a sequence of clays, clayey marls, and sand layers with varying thickness. Precise biostratigraphic dating of the above mentioned strata was not successful due to the lack of fossils (Fahrion, 1944). Two selected drillings near Neusiedl (CF N 6) and Frauenkirchen (CF FR 28) are shown in Figure 11 to delimit the top of Upper Pannonian sediments.

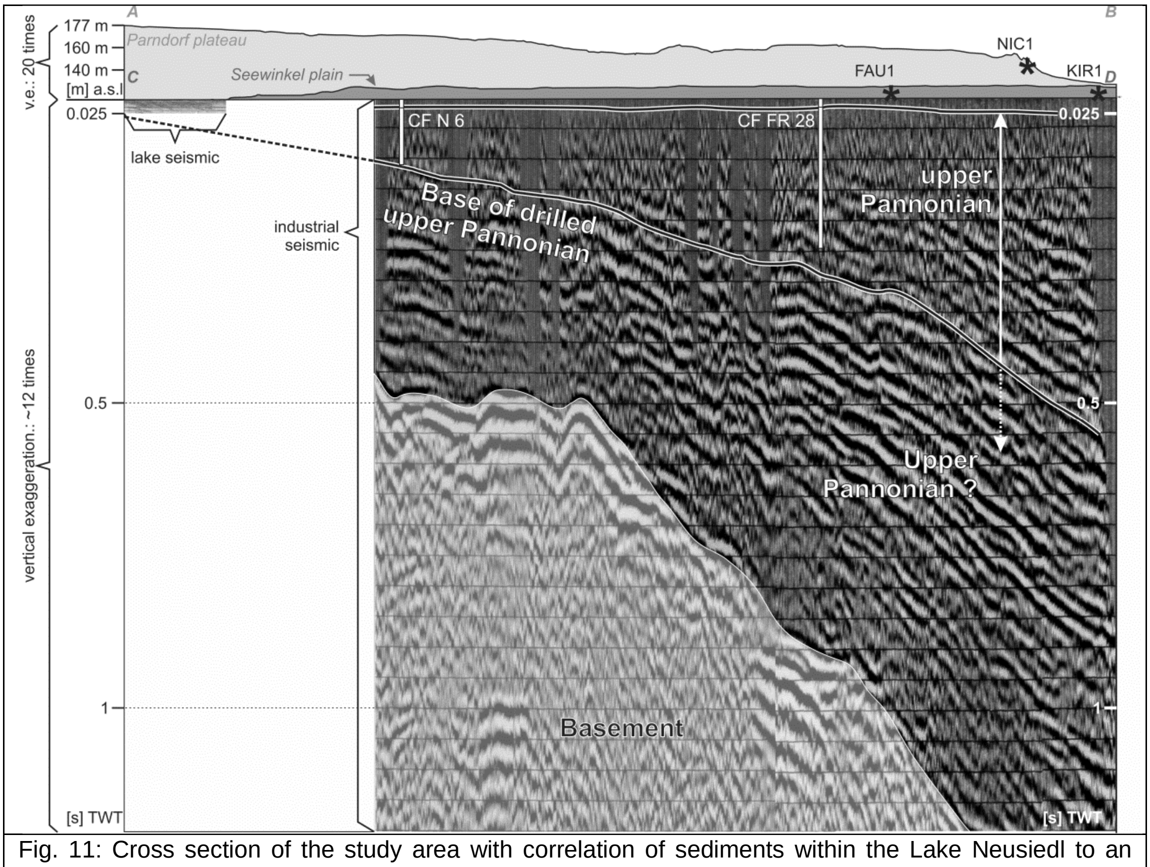

Fig. 11: Cross section of the study area with correlation of sediments within the Lake Neusiedl to an onshore industrial seismic section (OMV, 1970) rendered feasible by the acoustic velocity measurements and the link of seismic to well data. Sediments in the lake correspond to Upper Pannonian strata onshore. The deepest reflector that was still identified as upper Pannonian (Upper Miocene) with the help of the counterflush well CF N 6 is shown as “Base of the drilled upper Pannonian”. The counterflush well CF FR 28 also ended within upper Pannonain strata. Note that two

topographic sections (indicated in Figure 4) and the IRSL sample locations have been projected onto the cross section.

On the Parndorf Plateau, Quaternary sediments appear as a thin layer of only up to few meter thickness on top of Upper Miocene strata. Fluvial gravels are characterized by a similar fluvial depositional character as those observed in the Seewinkel Plain, but show a higher matrix content, a higher grade of chemical weathering and frequent, large-scale periglacial features (such as involutions or sand wedges). The higher grade of weathering is supported by analysis of the individual components which shows that limestone or dolomite clasts are almost completely absent. The crystalline components from the Parndorf Plateau include a granulite clast with garnets bearing a biotite rim and fine-grained granites (pers. comm. F. Neubauer, 2014). Similar to the sediments of the Seewinkel Plain, they appear generally loose and are not cemented. Meter-thick loess deposits were observed for example at the southern scarp of the plateau (Fig. 6b and 4). If occurring, they appear clearly thinner on top of the plateau.

IRSL ages of Quaternary fluvial sediments of the Parndorf Plateau and the Seewinkel Plain

IRSL-dating (Fig. 4, Table 3) of the sample NIC 1 from the Parndorf Plateau indicates an age in excess of the method (> 300 ka). The luminescence signal of this sample was in saturation and thus only yielded a minimum actual equivalent dose (De) and age. The samples from the Seewinkel Plain show potassium-feldspar IRSL ages of 102 ± 11 ka (FAU 1) and 76 ± 8 ka (KIR 1) (Table 3).

Sample ID		**NIC 1**	**KIR 1**	**FAU 1**
UTM Coordinates	x (m)	654116	651159	644800
	y (m)	5312846	5300405	5301258
De (Gy)		>600	228	221
uncertainty			10	11
Measured radionuclide concentrations				
% K		1.08	1.06	1.40
error (%K)		0.02	0.01	0.02
Th (ppm)		2.00	5.14	8.83
error (ppm)		0.07	0.15	0.24
U (ppm)		0.38	1.23	2.28
error (ppm)		0.01	0.03	0.04
Total dose rate, Gy/ka		**1.97**	**2.23**	**2.92**
error		**0.19**	**0.21**	**0.27**
AGE (ka)		**> 300**	**102**	**76**
error			**11**	**8**

Table 3: Overview of IRSL age measurement results.

Discussion and conclusions

Age and provenance of the sedimentary units

The IRSL and mineralogical data presented in this study, together with heavy mineral spectra and lithological information of certain pebbles from the gravel layers are used to identify the age and possible provenance areas of the investigated sedimentary units. Previously, age correlation was almost exclusively based on the altitude of terraces, diagenetical alteration of the sedimentsand on heavy mineral composition (e.g. Szádeczky-Kardoss, 1938; Tauber, 1959a; Fuchs et al., 1985; Wessely, 2006). This is, however, especially problematic in a tectonically active area such as the Little Hungarian Plain (Székely et al., 2009).

The distribution of sediments in the study area can be outlined as follows: Pannonian (Upper Miocene) strata occur throughout the study area, with a gradual thickening towards the depocenter of the Danube Basin further to the East (Mattick et al., 1996). Detailed mineralogical analysis of one drill core (NDS 7) from Lake Neusiedl indicates that only the uppermost few centimeters represent Holocene lake sediments. Higher smectite and feldspar contents, in addition to well crystallized carbonates suggest a significantly older, i.e. Pannonian age of the strata below. These findings together with the seismic data indicate the absence of continuous gravel deposits on top of the tilted sedimentary strata beneath the Lake Neusiedl.

In contrast, continuous Quaternary sediments, predominately sandy gravels, are found on both the Parndorf Plateau and the Seewinkel Plain. The measured IRSL ages from sand bodies within these gravels indicate Late Pleistocene deposition of the (upper) gravels in the Seewinkel Plain and a Middle Pleistocene or older deposition of the gravels on the Parndorf Plateau. Häusler (2010) provided an OSL age of 95 ± 10 ka from sediments in the Seewinkel Plain from the upper part of a gravel pit east of Wallern (Fig. 1), supporting our geochronological data.

Bernhauser (1962) distinguishes three main types of gravel deposits depending on elevation above sea level. The younger layers above 120 m a.s.l. comprise gravels with predominantly quartz components. The layers below 120 m a.s.l. also contain calcitic components. The third main type is observed only below 120 m a.s.l. and contains a variety of crystalline components most probably from the Leitha Mountains or the Rust Hills, but possibly also from the Sopron Hills (Bernhauser, 1962). A lithological inspection of individual components of the gravel horizons was performed in this study to retrieve indications on the provenance of the Quaternary units. Certain components from the KIR 1 sample were identified as gneisses and greenschist facies mylonites similar to Lower Austroalpine metamorphic rocks from the Wechsel unit (pers. comm. F. Neubauer, 2014) in the southwest of the study area (Fig. 1). However, similar greenschist facies mylonites are also found in the Mesozoic basement rocks in the Leitha Mountains (Erkmen, 2012). Thus, the inspection of gravel components from the chosen samples does not give clear indications on the exact source area.

Heavy mineral assemblages from (Upper) Pleistocene sediments indicate high amounts of garnets and staurolite, minor epidote and chloritoid, and varying, but significant amounts of green hornblende, attesting to a significant higher grade metamorphic source. The high abundance of garnet in the samples from the Seewinkel Plain (FAU 1, KIR 1) accompanied with the presence of fibrolithic silimanite and kyanite point to upper greenschist metamorphic to lower amphibolite facies rocks and minimizes the possibility of the Leitha Mountains as a source for the sediments of this area (Fig. 1). The Leitha Mountains and adjacent Rust Hills also contain mica schists, but almost no garnet (Tauber, 1959a). The presence of staurolite and kyanite in all samples principally could derive from the metamorphic rocks of the Sopron Hills, which contain staurolite, andalusite and kyanite schists in higher tectonic levels (Draganits, 1998), and include certain occurrences of fibrolithic silimanite within the Óbrennberg-Kaltes Bründl Series. A southern provenance would be further supported by the lack of chromium spinel that could originate from Gosau units in the Eastern Alps (Wagreich and Marschalko, 1995) and the minor amounts of epidote, which predominate the Upper Pleistocene fill of the Mitterndorf basin within the southern Vienna Basin (Salcher, 2008). However, the composition of these samples fall also into the variation of recent and Pleistocene sands of the Danube River (e.g. Frasl, 1955; Salcher, 2008).

The recent sands of the southerly derived Répce (Rabnitz) and Rába (Raab) Rivers show similar high amounts of garnet, but plot further to right in the logratio-diagrams due to higher amounts of green hornblende and/or lower amounts of staurolite. Therefore, no clear distinction between Danube derived and southerly-derived (Répce/Rába Rivers) deposits can be made using heavy minerals. Reworking from older, i.e. Pannonian sediments (see below) cannot be excluded.

Upper Pannonian sediments from the strata below the Lake Neusiedl show a rather similar heavy mineral assemblage (Fig. 10c), with high amounts of garnet and staurolite together with epidote and varying amounts of green hornblende, and significant traces of kyanite and silliminate. Green hornblende is significantly lower than staurolite, forcing almost all Pannonian samples to plot to the left of the zero-line. Therefore, a clear distinction can be made between Upper Pannonian and Quaternary strata, indicating changes in sediment transport and provenance during the last 5-6 Ma.

Lithologically, the central area of the Sopron Hills with exposures of Austroalpine basement units of higher metamorphic grade represents a source area compatible with this heavy mineral spectrum (Draganits, 1998). Investigations from the Oberpullendorf Basin (Fig. 1) provide largely similar heavy mineral spectra from these Neogene sedimentary units (Schoklitsch, 1962). Pannonian sediments from the catchment of the Répce at Bubendorf/Drassmarkt also show a rather similar heavy mineral assemblage strongly dominated by garnet, but higher epidote percentages (Nebert et al., 1980; Schoklitsch, 1962). In the eastern part of the Oberpullendorf basin, directly southwest of the Lake Neusiedl, Schoklitsch (1962) distinguishes two heavy mineral assemblages within the Pannonian sediments. A tourmaline-zircon dominated assemblage in younger Pannonian sediments, and a garnet and staurolite dominated assemblage in older Pannonian sediments, which correlates to the assemblages from the drilling cores from the Lake Neusiedl.

A major difference between the Upper Pannonian and the Quaternary samples in our study area is that the latter have more abundant hornblende and fibrolithic silimanite. This may indicate changing amounts of amphibolites within the source area, but may be also related to lower stability of hornblende during transport (Nebert, 1980; Mange and Maurer, 1992). It should also be noted that staurolite is present in all the sediments of the Danube-terraces further downstream in the Little Hungarian Plain (Csapó, 1998).

The role of (active) tectonics in shaping the geomorphology of the Eastern Little Hungarian Plain

Slow, regional vertical crustal movements are still affecting the western margin of the Little Hungarian Plain (Höggerl, 1989; Wolfartsberger, 2011). A regional uplift (i.e. 100 ka to 1 ma) is indicated by staircase terrace formation and stream incision. The uplift rate is suggested to be in the order of 100-200 m/ Myr (Salcher et al., 2012) which is very similar to adjacent regions (Wagner et al., 2010; Gusterhuber et al., 2012). On top of this uplift, local tectonics plays a major role in the study area. This is suggested e.g. by distinct, straight scarps and horst-and-graben like features represented by the Lake Neusiedl depression and the elevated highs limited by straight scarps surrounding the Lake Neusiedl (e.g. Székely et al. 2009; Rath et al., 2011).

The top of the Upper Miocene (Pannonian) strata in the lake area and the adjacent Seewinkel Plain is located at around 110 m a.s.l. The top Pannonian of the Parndorf Plateau is up to 50 m higher, clearly showing a dip towards SE. Székely et al. (2009) speculated about a neotectonic influence on linear features in the Little Hungarian Plain (see also Fig. 3). In fact, NNE-SSW trending grabens which were discovered in the Tertiary basement in a depth of more than 1500 m (Kilényi and Sefara, 1989; Tari, 1994) are parallel to the straight SE margin of the Parndorf Plateau. Further evidence for neotectonic activity was provided by Zámolyi et al. (2010) who showed that planform geometries of local Danube tributaries in the Little Hungarian Plain are affected by normal faulting. Young tectonic activity in that region is related to the onset of basin inversion phase at approximately 4.5 Ma years ago in the adjacent, more eastern parts of Pannonian Basin (Fodor et al., 2005 and citations therein). However, inversion did not affect the whole basin at the same time (Fodor et al., 2005 and citations therein). It is likely that the Parndorf Plateau and the Seewinkel Plain at the western margin of the Little Hungarian Plain are subject to a similar extensional deformation regime as the eastern margin. This is especially true when considering the ongoing subsidence of the Little Hungarian Plain (Joó, 1992; Timár and Rácz, 2002). Cloetingh et al. (2005) suggested that the regional-scale depressions and horst-like structures in the Pannonian Basin originate from stress induced deflection of the lithosphere. Analogous to this assumption, the elevated areas and depressions in the Little Hungarian Plain may be interpreted as horst and graben structures, which are the consequence of this extensional regime even though the resulted topographic wavelength is much shorter. However, it has to be noted that in contrast to other locations (e.g. the Lassee Fault in the Vienna Basin; Beidinger and Decker, 2011), outcrop evidence of normal faults displacing the Quaternary sediments at the margins of the Parndorf Plateau has not been found (yet).

The overall dip of Upper Pannonian strata around 2° towards the East (observed in the lake seismic and in industrial seismic sections; Fig. 3 and 11) are likely to be the consequence of the early/middle Miocene extensional to transtensional regime (Decker and Peresson, 1996). Fodor et al. (2005) report such a deformation style at the eastern margin of the Little Hungarian Plain (Vértes Hills, Transdanubian Range; Fig. 1) which initiated in the Late Miocene (Sarmatian, 11-13 Ma) and persisted up to Middle- to Late Pliocene. Thus, differences in elevations between the Parndorf Plateau and the Seewinkel Plain might be regarded as the result of local vertical crustal movements and post-Pannonian normal faulting.

Quaternary landscape evolution models

The recent configuration of the Lake Neusiedl region is mainly the result of processes related to local uplift and subsidence and associated processes of erosion and accumulation. The role of climate for deposition is difficult to decipher. Ages from the exposed, upper levels of the Seewinkel Plain (e.g. FAU 1, KIR 1) exclusively fall into the early stages of the last glacial period. However, we cannot exclude the occurrence of older remnants. Ages and the stratigraphic context (i.e. lack of unconformities) suggest that the sedimentation of these upper levels of the Seewinkel Plain took rather place during a relatively short depositional phase. Such a distinctive and limited time period for deposition (i.e. few tens of thousands of years) is regionally uncommon for an active basin setting, were generally longer records of climatic and/or tectonically driven sequences are preserved (e.g. Decker et al., 2005; Salcher and Wagreich, 2010). Remarkably, the deposits do not seem to reach the area of the Lake Neusiedl hosting only much older, Upper Miocene sediments. The sediment distribution roughly coincides with the shore on the lake’s eastern side. In accordance, the regional distribution of gravel shows that the fluvial sediments where already deposited on the eastward tilted Pannonian sediments, progressively limiting accumulation space of any fluvial stream to the west. The lack of gravels beneath Lake Neusiedl cannot be explained by erosional processes. This view is also shared by Tauber (1959a).

The gravels on the Parndorf Plateau are significantly older than the upper levels of the gravel deposits in the Seewinkel Plain and they are located approximately 45 m higher (measured from Top Pannonian/Base Quaternary). However, Quaternary sediments are that old that no exact IRSL age could be derived from the plateau. An interpretation as Danube terrace can clearly be derived from sediment provenance and depositional geometry (Szádeczky-Kardoss, 1938; Tauber, 1959b) and is also suggested by the connection to terraces of similar altitude across the Leitha River. Elevation and tilt suggests that it might be the continuation of the highest and oldest terrace system preserved in the Vienna Basin more to the west (Szádeczky-Kardoss, 1938). We suggest that the steep, southeast margins of the Parndorf Plateau are scarps of reactivated basement faults; a scenario recently described from Danube terraces in the Vienna Basin (e.g. Decker et al., 2005; Beidinger et al., 2011).

Fluvial sediments on top of the Parndorf Plateau might not have reached the area of modern Lake Neusiedl. In fact, outcrops at or near the scarp’s slope (e.g. at Edmundshof and Mönchhof) suggest no or only very minor (few dm-thick) coverage of fluvial sediments (Fig. 6a and 6b) unconformably resting on top of Pannonian strata. The former floodplain of the Danube might therefore not have been extensive enough to cover the area of the modern lake. This view is supported by the lack of a continuous gravel layer within the lake. However, channel features are most likely present in the Upper Miocene strata within the lake as interpretation of the shallow lake seismic sections indicates. Similar features were also mentioned by Hodits (2006).

The age of scarp initiation and formation along the Parndorf Plateau remains even vaguer. A paleosol complex interbedded with thick loess deposits that outcrop along the scarp slope may suggests an age which clearly exceeds the Late Pleistocene (Fig. 6b).

While the provenance of the fluvial deposits on the Parndorf Plateau can clearly be attributed to the catchment of the Danube, provenance is less clear for the younger sediments in the Seewinkel Plain.

In conclusion two possible scenarios for the deposition of the Quaternary sediments in the Little Hungarian Plain are presented suggesting either (1) an origin from the South (e.g. Répce) or (2) from the North, the Danube.

Scenario (1): The Danube deposited its sediments on the area of the recent Parndorf Plateau and shifted towards the north (Wessely, 2006 and citations therein; Szádeczky-Kardoss, 1937) (Figs. 12a and b). The northward shift of the Danube into the gate of Devin (Fig. 12b) can be explained by a major incision phase probably linked to the uplift of the Western Carpathians and the Leitha Mountains (Wessely, 1961; Wessely, 2006). Caused by subsequent post-Pannonian normal faulting and subsidence the Parndorf Plateau formed an elevated region, probably at the time of the deposition of the gravels in the Seewinkel Plain (Fig. 12b). In this model it is assumed that the gravel layer of the Seewinkel Plain was deposited by local tributaries from the South similar to the modern Ikva or Répce Rivers (Figs. 12a to c). Fuchs (1974) states that the composition and depositional geometry of gravels within the Seewinkel Plain point towards an origin from the South, which would fit with the occurrence of staurolite (this study), gravel components from the

Wechsel unit (this study) and the fact that the quartz of the Seewinkel sediments is not suitable for OSL dating, which is in contrast to samples from the Danube River.

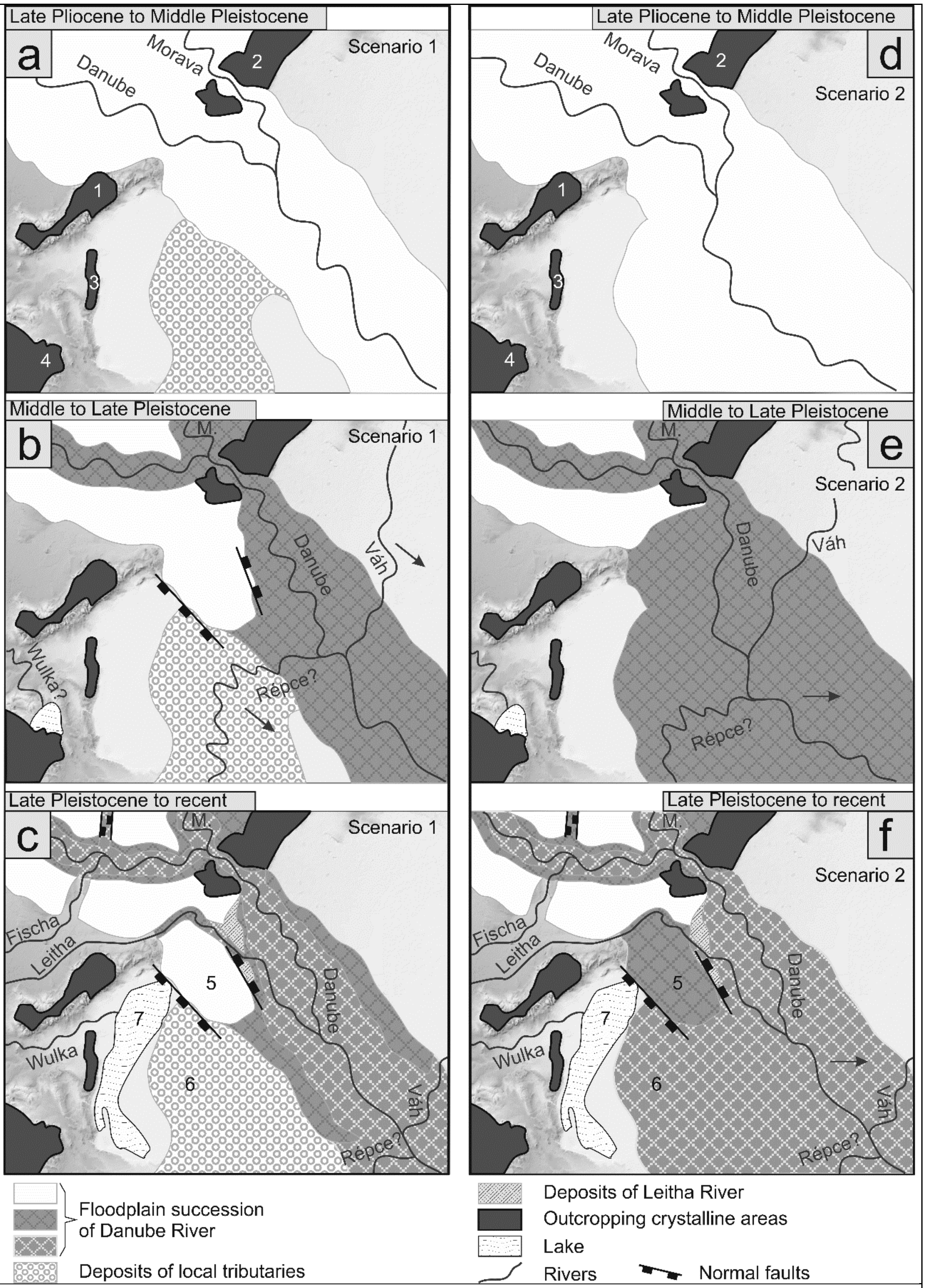

Fig. 12: Map view of the two possible evolution models of the study area. The age constraint for the beginning of the first phase is derived from data on the filling up of the Lake Pannon (Kováč et al., 2006; Magyar et al., 2013). The onset of the second phase (~640 Ka) is controlled by the northward migration of the Danube from the Gate of Bruck to the Gate of Devin (Wessely, 2006). The last phase is set to ~100 Ka due to the age of the gravel deposits in the Seewinkel Plain (this paper). The grey arrows indicate the migration of the river channels towards the East, towards the modern depocenter of the Little Hungarian Plain. 1: Leitha Mountains, 2: Western Carpathians, 3: Rust Hills, 4: Sopron Hills, 5: Parndorf Plateau, 6: Seewinkel Plain, 7: Lake Neusiedl.

Later on, upon deepening of the depocenter around Győr (Joó, 1992, Gábris and Nádor, 2007, Lovász, 2007) (Fig. 1) the main channels of the tributaries migrated towards the East to their current positions (Fig. 12c, see also Lovász, 2007) and the distinct orientation of the modern drainage pattern evolved.

Scenario (2): The Danube developed a wide floodplain spreading across the whole Seewinkel Plain and on the area of the not yet elevated Parndorf Plateau (Fig. 12d) up to the Late Pleistocene. Thus, in this scenario the gravel deposits on the Parndorf Plateau are younger than in scenario (1). This scenario is based based on foreset orientations and component analysis of Szádeczky-Kardoss (1938), Tauber (1959b) and Husz (1965). Szádeczky-Kardoss´s (1938) measurements indicate a WSW to SE directed palaeo-current direction. During the deposition of the gravels in the Seewinkel Plain in the Late Pleistocene (100 ka) local normal faulting created the present-day relief including the Parndorf Plateau (Fig. 12f).

Based on current data none of the two scenarios can be excluded. Further age dating of critical sediments and a detailed provenance analysis including pebble types and single grain mineral chemistry may solve this question.

Acknowledgements

This study was supported by the Austrian Science Fund (FWF projects P20092-N10 and V151-N22) and University project FA 536001 [Karpatian Tectonics II]. Field work of the lake drillings and lake seismics was kindly supported by the Biologische Station (A. Herzig), UWITEC, and students of the University of Vienna, Eötvös Loránd University, Budapest and the University of Leeds. Many thanks to Maria Heinrich for the access to the counterflush data in the drill core archive of the Austrian Geological Survey and to OMV Austria Exploration and Production GmbH for the permission to use seismic lines in the Seewinkel area. Magda Bottig’s accurate heavy mineral counting is highly acknowledged. The authors thank Franz Neubauer very much for the petrographic review of outcrop samples and Mathias Harzhauser for the paleontologic investigation of the lake drill cores.

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17. Appendix D

Zámolyi, A., Kovács, G., Székely, B., Timár, G. 2010. A morphometric analysis of the fault pattern of the Bakony Mountains: some tectonic geomorphological implications. Földtani Közlöny, 140(4), 439-453.

Magyarhoni Földtani Társulat
Földtani Közlöny
Hungarian Geological Society
140/4, 439–453., Budapest, 2010

A Bakony vetőmintázatának morfometriai vizsgálata és az ebből levonható néhány tektonikus geomorfológiai következtetés

ZÁMOLYI András[1,2], KOVÁCS Gábor[1,3,4], SZÉKELY Balázs[1,5], TIMÁR Gábor[1]

[1] ELTE Földrajz- és Földtudományi Intézet, Geofizikai és Űrtudományi Tanszék, H–1117 Budapest, Pázmány Péter sétány 1/c
[2] Department für Geodynamik und Sedimentologie, Universität Wien, Althanstrasse 14, A–1090 Wien
[3] ELTE Földrajz- és Földtudományi Intézet Természetföldrajzi Tanszék, H–1117 Budapest, Pázmány Péter sétány 1/c
[4] ELTE IK Térképtudományi és Geoinformatikai Tanszék, H–1117 Budapest, Pázmány Péter sétány 1/a
[5] Institut für Photogrammetrie und Fernerkundung, Technische Universität Wien, Gusshausstr. 27–29, A–1040 Wien

A morphometric analysis of the fault pattern of the Bakony Mountains: some tectonic geomorphological implications

Abstract

The unique tectonic setting of the Pannonian Basin was recognised as early as 1900. In particular, the evolution and geologic characteristics of the outcropping pre-Cenozoic mountain ranges have been subject to detailed studies and prolonged discussions. Among these efforts, the Transdanubian Range occupies an important position because its well-preserved outcrops, (which are rare in the Pannonian Basin) make it possible to collect direct measurements of the fault striations and fault plane orientations. The tectonic position of the Transdanubian Range provides the key to the link up of the Carpathian– Pannonian system to the Eastern Alps. It is interpreted as the uppermost thrust sheet of the Alpine nappe system that was emplaced in the Cretaceous.

A key advance was made by the pioneering approach of MÉSZÁROS, who compiled a 1:100,000-scale structural and economic-geologic map of large parts of the Transdanubian Range in 1982 (northern and southern Bakony and Balaton Highlands). This detailed work puts a major emphasis on the mapping of structural features, distinguishing their sense of movement and constraining the time of their activity. Unfortunately, the map itself has never been published. However, parts of the map which were important for mining activities taking place at that time were produced in a summary paper (MÉSZÁROS 1983). This study focuses on the correlation of the fault pattern with geomorphologic features. Integration of the original map of 1982 into a GIS environment and subsequent tectonic geomorphological analysis reveals the close relationship between tectonics and the landscape evolution of the Bakony Mountains (PÉCSI 1987). A selected set of study areas provides results for further characterisation of the structural elements.

A total of five hundred faults were subdivided and, segmented into two basic groups: one group consists of faults with spatially-related geomorphologic surface expression and the other group comprises tectonic elements not directly related to characteristic morphological features. These groups were further analysed on the basis of the properties assigned to them by MÉSZÁROS (direction of fault movement and age); the orientation patterns were compared to fault datasets from different authors for the whole of Transdanubia.

Orientation analysis of various classes of faults reveals a surprisingly uniform angular distribution: the overall orientation pattern in the Bakony shows a slight but significant difference with respect to the general orientation in Transdanubia. Azimuthal distribution of the faults mapped by MÉSZÁROS is bimodal and can be interpreted as orthorhombic sets of faults. Additionally, the strike of geomorphologically significant faults differs from the strike direction of geomorphologically non-visible faults by 15 to 20 degrees.

Keywords: Transdanubian Range, Bakony Mountains, structural geologic map, tectonic geomorphological analysis, azimuthal distribution

Összefoglalás

A Pannon-medence egyedi tektonikus környezetét már az 1900-as években felfedezték. Kiváltképp a kibukkanó pre-kainozoos középhegységek földtani fejlődése és felépítése részletes tanulmányokra és hosszú vitákra adott alkalmat. A középhegységek közül a Dunántúli-középhegységnek jut kiemelkedően fontos szerep, többek között, mert jól hozzáférhető feltárásai vetősíkok és vetőkarcok közvetlen mérését teszik lehetővé. A Dunántúli-középhegység szerkezetföldtani helyzete kulcsfontosságú információt szolgáltat a Kárpát–Pannon térség és a Keleti-Alpok közötti átmeneti zóna kutatásában. Ezt ma az alpi takarók legfelső, krétában áttolódott egységének tartják.

MÉSZÁROS 1982-es előremutató szerkezetföldtani összeállítása egy fontos alaptérképet szolgáltat kutatásaink számára. Az 1: 100 000-es léptékű szerkezet- és gazdaságföldtani térkép a Dunántúli-középhegység nagy részét lefedi, főleg az Északi- és a Déli-Bakonyra és a Balaton-felvidékre összpontosítva. Ez a részletes munka külön figyelmet szentel a szerkezeti elemek osztályozásának, nemcsak különbséget téve sokféle vetőtípus között, hanem azoknak korolását is dokumentálva. Magát a térképet sosem publikálták, viszont a folyamatban lévő bányászati tevékenység számára fontos részleteket és részlettérképeket egy összefoglaló cikkben tette közzé (MÉSZÁROS 1983). Jelen munkában a vetők és geomorfológiai elemek közötti kapcsolatra összpontosítottunk. Az 1982-ben készült eredeti térkép térinformatikai integrációja és tektonikus geomorfológiai elemzése a Bakonyvidék (PÉCSI 1987) szerkezetföldtani és felszínfejlődése közötti szoros összefüggésre enged következtetni. A kiválasztott elemzési területek a szerkezetföldtani elemek további jellemzését teszik lehetővé.

Összesen félezer vetőt osztályoztunk két alapvető szempont szerint, szükség esetén a vető vonalát felszabdalva: az első csoport olyan vetőket foglal magába, melyeknek geomorfológiai megnyilvánulásuk dokumentálható; a második csoport nem mutat közvetlen összefüggést jellegzetes geomorfológiai változásokkal. Ezeket a csoportokat tovább bontottuk a MÉSZÁROS által hozzárendelt tulajdonságok (vető mozgásának iránya és kora) alapján és csapásirányukat összehasonlítottuk más szerzők szerkezetföldtani adatbázisaival az egész Dunántúli-középhegység területére.

A különböző vetőosztályok csapásirány-elemzése meglepően egyveretű eloszlásokat mutat: az általunk elemzett vető-adatbázisokban a Bakonyban fekvő összes vető enyhe, de jól kimutatható iránybeli eltérést mutat az egész Dunántúlon megfigyelhető eloszláshoz képest. A MÉSZÁROS-féle térképen dokumentált vetők csapásirány-eloszlása bimodális és romboéderes térbeli elrendezést mutat. A geomorfológiailag megnyilvánuló MÉSZÁROS-féle vetők csapásiránya az egyéb vetők csapásirányától 15–20 fokot tér el.

Tárgyszavak: Dunántúli-középhegység, Bakony, szerkezetföldtani térkép, tektonikus geomorfológiai elemzés, csapásirány-eloszlás

Bevezetés

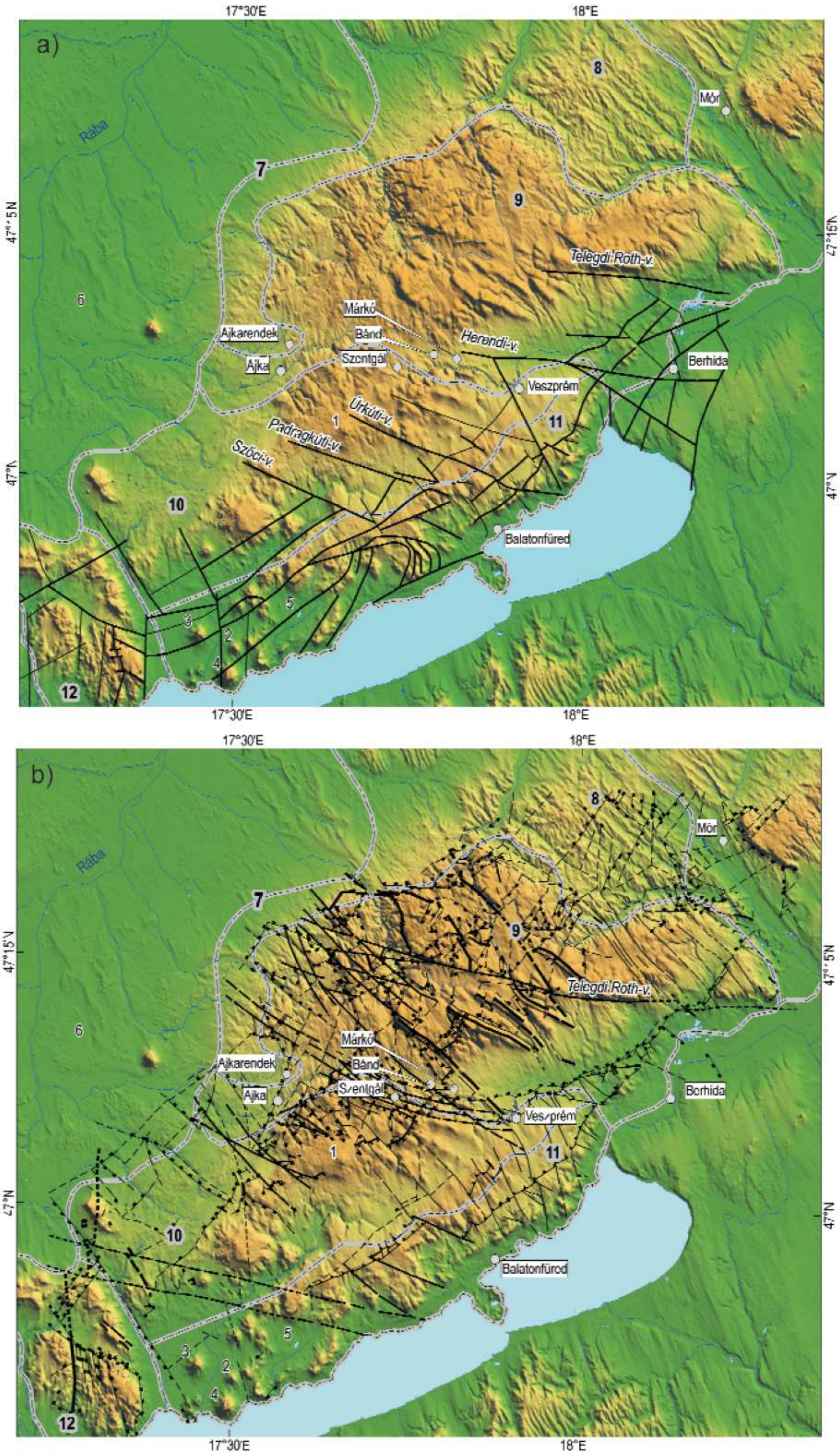
A Dunántúli-középhegység (*1. ábra*) és azon belül a Bakonyvidék (PÉCSI 1987) takarós felépítését már a múlt század elején felismerték (UHLIG 1907, STRAUSZ 1942). Később ezeket az ismereteket bővítették. Így például KÁZMÉR (1984) és KÁZMÉR & KOVÁCS (1985) a Dunántúli-középhegység kelet felé irányuló kiszökése mellett érveltek a Periadriai-vonal és a Defereggental–Anterselva–Valles-vonal mentén 400 km-re eredeti elhelyezkedésétől. Szeizmikus szelvények kiértékelése és felújult, miocénben keletkezett mélyszerkezeti elemek térképezése (pl. RUMPLER & HORVÁTH 1988, TARI 1996) a tektonikusan aktív Bakonyvidék képét támasztják alá. Recens vertikális kéregmozgások meghatározására irányuló geodéziai mérések eredményei a Bakonyvidék területére 0,3 és 0,5 mm/év kiemelkedést mutatnak (JOÓ 1992).

Ez a földtani környezet messzemenően alkalmas a szerkezetföldtani folyamatok geomorfológiai megnyilvánulásainak vizsgálatára. A MÉSZÁROS által összeállított szerkezetföldtani térkép (*2. ábra*) nemcsak a részletessége és területi kiterjedése miatt bizonyult alkalmas alaptérképnek, hanem a vetők osztályozása miatt alcsoportok képzésére is alkalmas. A vetők fekvése, futása és kinematikája a külszíni fejtésekben és bányavágatokban végzett megfigyelések alapján nagy térbeli pontossággal dokumentált (MÉSZÁROS 1983).

A tanulmányban nagy hangsúllyal szerepel az oldaleltolódások jellemzése és korolása (*1. táblázat*). Főleg a bauxit-előfordulások elhelyezkedésében látta az oldalelmozdulások meghatározó szerepét (MÉSZÁROS 1983). A NyÉNy–KDK-i csapású jobbos oldaleltolódások keletkezését TARI (1996) az albaira teszi. Felújulásukat több szerző a miocénbe helyezi (CSONTOS et al. 1992, FODOR et al. 2005, SASVÁRI et al. 2007). Az oldaleltolódásokat szakításos

1. táblázat. MÉSZÁROS által kor és jelleg szerint megkülönböztetett vetőtípusok.
Table 1. *Types of faults from the map of MÉSZÁROS differentiated by age and fault type*

Szerkezeti elem	Kor	Jelleg
Szinklinális		
Antiklinális		
Vető általában		megállapított/ feltételezett
Törésvonal	jura	megállapított/feltételezett; széles dörzsbreccsa-zónával rendelkezik
Oldaleltolódás		megállapított; nem egyértelmű irányú
	késő-neogén	megállapított
	késő-neogén	megállapított; jobbos
	késő-neogén	feltételezett, balos
	késő-neogén	feltételezett, jobbos
	késő-neogén	fedett balos
	késő-neogén	fedett jobbos
		megállapított/feltételezett; balos szerkezetet tagoló
		megállapított/feltételezett; jobbos szerkezetet tagoló
		fedett; jobbos/balos szerkezetet tagoló
		megállapított/feltételezett jobbos/balos szerkezetet megszabó
		fedett jobbos/balos szerkezetet megszabó
	szubhercini	megállapított/feltételezett; fiatalabb mozgások során vetővé felújult jobbos /balos
Feltolódás		megállapított/feltételezett
Áttolódás		



1. ábra. a) A kutatási terület áttekintő térképe (háttér: árnyékolt DDM10 domborzatmodell) a Bakony déli részének főbb posztpaleozoos szerkezeti elemeivel (DUDKO 1991) és a cikkben előforduló fontosabb helyrajzi nevekkel

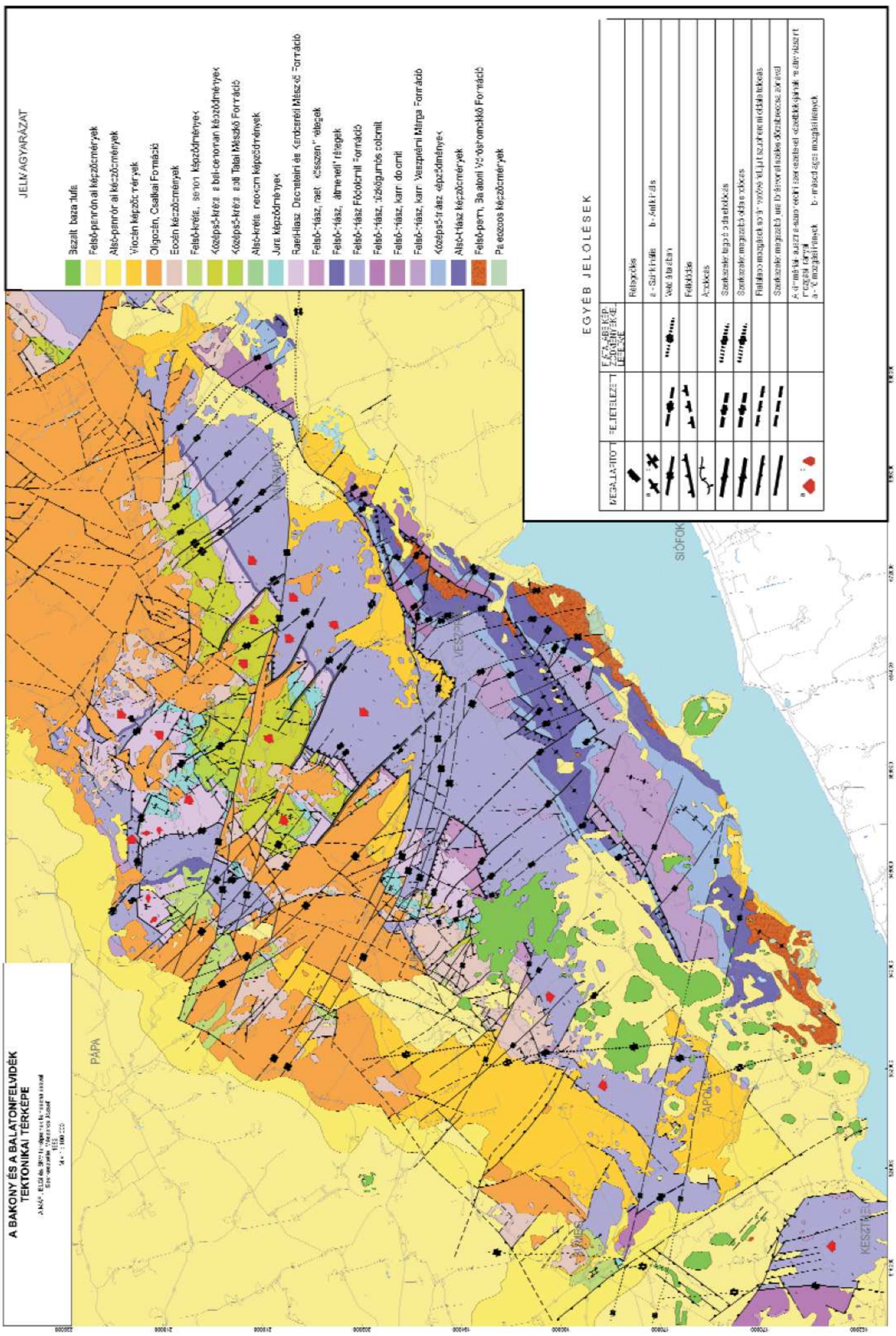
1 = Kab-hegy, 2 = Gulács, 3 = Szent György-hegy, 4 = Badacsony, 5 = Káli-medence, 6 = Marcal-medence. Szürke háttérrel rendelkező számok: 7 = Dunántúli-középhegység, 8–12 = Bakony-vidék, 8 = Bakonyalja, 9 = Északi-Bakony, 10 = Déli-Bakony, 11 = Balaton-felvidék, 12 = Keszthelyi-hegység.

b) MÉSZÁROS (1982) szerkezeti vonalai ugyanazon a háttéren összehasonlításképpen

***Figure 1.** a) Overview map of the study area (background: hillshaded DDM10 digital elevation model) with the main post-palaeozoic structures (DUDKO 1991) and important location names mentioned in the text*

1 = Kab Hill, 2 = Gulács, 3 = Szent György Hill, 4 = Badacsony, 5 = Káli Basin, 6 = Marcal Basin). Numbers with grey background: 7 = Transdanubian Range, 8–12 = Bakony Mountains, 8 = Bakony Promontory, 9 = Northern Bakony, 10 = Southern Bakony, 11 = Balaton Highlands, 12 = Keszthely Mountains

b) Structural features of MÉSZÁROS (1982) on the same background for comparison



2. ábra. MÉSZÁROS (1982) szerkezetföldtani térképének térinformatikai rendszerbe integrált vektoradatbázisa
Figure 2. *Vector database of the structural geologic map of MÉSZÁROS (1982)*

törésekként értelmezi, melyek az eoalpi felső-kelet-alpi takarórendszeren belül nyesődtek le (TARI 1996).

A Bakonyvidék részterületein már korábban végeztek részletes geomorfológiai elemzéseket. JORDÁN et al. (2003, 2005) szisztematikus digitális domborzati modell elemzést hajtottak végre a Káli-medence területén (*1. ábra*). Megállapították, hogy a Káli-medence alapja egységesen DNy-felé billen. Lejtőkitettség-elemzésük É–D és ÉNy–DK irányítottságú szerkezeti elemekre enged következtetni (JORDÁN et al. 2005).

Földtani háttér

A Bakonyvidéket számos szerző az észak-pannóniai terrénhez sorolja (BALLA 1984) és kipréselődését az alpi területről a kora-miocénben nagyméretű eltolódások (Rába-vonal, Balaton-vonal) mentén tekinti valószínűnek (KÁZMÉR & KOVÁCS 1985, TARI 1991, CSONTOS et al. 1992). Eredeti elhelyezkedése miatt a Bakony mind a Déli-Alpok, mind az Északi-Alpok szerkezeti jegyeivel rendelkezhet, vagyis pikkelyes és takarós szerkezeti felépítése is lehet (BUDAI et al. 1999; *1. ábra*).

A Bakonyvidék szerkezetileg (mostani elhelyezkedését alapul véve) az ÉK–DNy-i csapású, hosszanti alpi feltolódásokat elvető, K–Ny-i csapású oldaleltolódások által tagolt és miocén medencékkel szabdalt terület. Ezek a harántirányú oldalelmozdulások a Bakonyvidék markáns és fontos szerkezeti elemei. MÉSZÁROS (1983) térképén kiemelt szerepet juttat a harántirányú, túlnyomóan jobbos eltolódásoknak. Néhány ilyen eltolódásos zónát már korábban felfedeztek (pl. Litéri-vető, Telegdi Roth- és Herendi-vető, lásd *1. ábra*), fekvésük és elhelyezkedésük bizonyítottnak tekinthető (PÁVAI-VAJNA 1930, TELEGDI ROTH 1935, KÓKAY 1976). Néhány egyéb jobbos oldaleltolódás MÉSZÁROS térképén ábrázolt futását viszont nem sikerült későbbi szerzőknek megerősíteniük. Így például a Padragkúti-, illetve a Szőci-vonal BUDAI et al. (1999) térképén sokkal rövidebb, mint a MÉSZÁROS-féle térképen található hasonló futású és jellegű vetők. FODOR et al. (2005) vizsgálatai alapján a Padragkúti-vonal a litéri áttolódásos szerkezetben végződik el. Az Úrkúti-vonal DUDKO (1991) és BUDAI et al. (1999) szerint végigkövethető a Balaton partvonaláig. Az oldaleltolódások mellett a feltolódásos szerkezetek Balaton partvonalával szinte párhuzamosan futnak és egy viszonylag keskeny zónára korlátozódnak. Fiatalabb szerkezetföldtani fejlődését tekintve a pannóniai üledékeket is érintő deformáció egy ÉNy–DK-i extenziós feszültségteret mutat, amely normál vetők mentén alakult ki (BUDAI et al. 1999, KISS & FODOR 2007).

A felszínfejlődés fontosabb állomásai

A Dunántúli-középhegység domborzata összetett tektonikus fejlődése miatt erősen tagolt és több geomorfológiailag különböző egységre oszlik. Ezek között szerepelnek (i) maradványfelszínek, (ii) etchplainek (CSILLAG 2004), (iii) karsztfelszínek, (iv) sasbérc- és árok-szerkezetek, (v) késő-miocén bazaltvulkanizmushoz köthető felszíni formák (JUHÁSZ 2002).

Jelenleg a késő-permig lehet a felszínfejlődési folyamatokat visszavezetni (CSILLAG 2004). Geomorfológiailag jelentős felszínformát a kréta és a középső-eocén során kialakult etchplainek képeznek, melyek szubszekvens eróziós folyamatok során részben lepusztultak (CSILLAG 2004).

A kainozoos tektonikus mozgásokhoz köthető felszíni formák az eocén tengeröböl kialakulása során részben eltemetődtek. Az öblöt körülvevő szárazulatokon a lepusztulás különböző mértékben hatott. Maga a terület egésze vetőkkel tagolt sasbérc és árok jellegűnek képzelhető el (DUDICH & KOPEK 1980). Az oligocén pedimentációs és eróziós folyamatok során ezek a sasbércek átformálódtak (CSILLAG 2004). A miocénben a meglévő felszínformák eltolódások mentén (pl. Telegdi Roth-vető; SASVÁRI et al. 2007) elnyíródtak (KÓKAY 1996), és transzpressziós szerkezetek mentén jelentős kiemelkedés is végbement (CSILLAG 2004). A szarmata során a kiemelkedés folytatódott, ami erősen megnövekedett eróziós folyamatokhoz vezetett. A késő-pannóniaiban következő lepusztulási folyamatot (KÓKAY 1996, CSILLAG 2004) a pliocén vulkanizmus jelentősen befolyásolta. E vulkanizmus első fázisa során a lávafolyások már erodált felszíni formákat töltöttek ki és így részt vettek a felszín aktív formálásában. A második fázisban maguk a vulkáni felépítmények pusztultak le (CSILLAG 2004).

Fontos megjegyezni, hogy a kréta–középső-eocén etchplainek olyan mértékben átalakultak, hogy már nem mutatkoznak a mai geomorfológiában (CSILLAG 2004). A terület csaknem 1 mm/ év jelenkori emelkedését figyelembe véve (JOÓ 1992) az eróziós folyamatok feltehetően egy, a neoalpi szerkezeti mozgások által feldarabolt ősfelszínt takarnak ki a szarmata–pannóniai üledékfedő alól (JÁMBOR 1980, CSILLAG 2004). A terület relatív kiemelkedése (a Kisalföldhöz és a Balatontól délre eső területekhez képest, lásd JOÓ 1992) a völgyek fokozatos hátravágódását eredményezi, és hozzájárulhat egyes morfostrukturális elemek kihangsúlyozódásához.

Alapadatok

MÉSZÁROS 1982-ben szerkesztett térképének színes, eredeti, kéziratos példányát szkennelés után megfelelő számú illesztőponttal EOV koordinátarendszerbe illesztettük. Az ezen a raszteres alapon készült digitalizált vektoradatbázis magába foglalja a litológiai poligonokat, a szerkezeti elemek polivonalait és a rétegdőlések pontszerű adatait a hozzátartozó attribútumokkal együtt (*2. ábra*). MÉSZÁROS bakonyi szerkezetföldtani térképét TARI szeizmikus szelvényekből levezetett szerkezetföldtani térképével (TARI 1996), és a zalai terület presenon felszínével és szerkezeti vonalaival (JOCHA-EDELÉNYI 2005) vetettük össze. Az Északi- és Déli-Bakony és a Balaton-felvidék területén

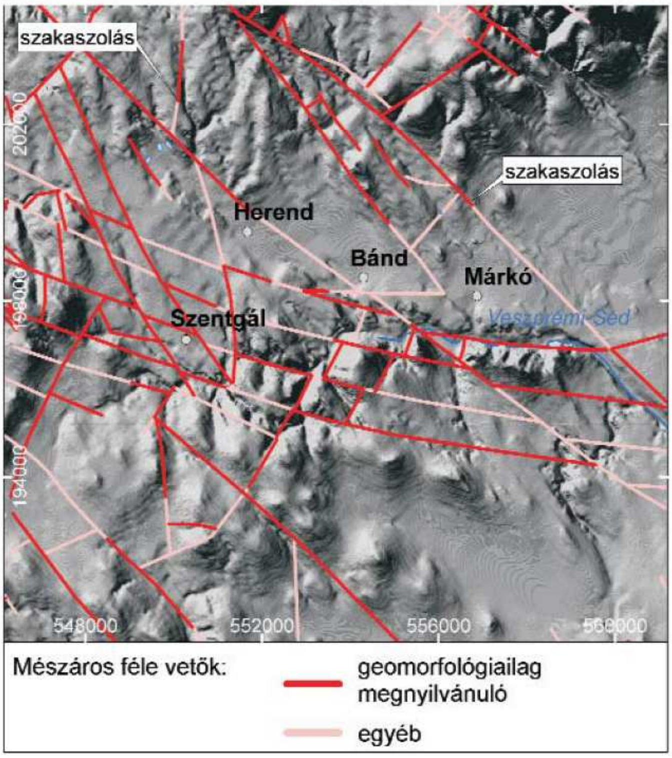
DUDKO (1991) által szerkesztett szerkezetföldtani térképpel egészítettük ki a csapásirány-elemzéseket (*1. ábra, a*).

Az árnyékolt globális SRTM domborzati modell alapján készült morfostrukturális elemek térképét (HORVÁTH et al. 2006a, b) összehasonlítási adatbázisként használtuk. Mivel az idézett két mű szempontunkból azonos adatbázist ír le (HORVÁTH et al. 2006a magát az adatbázist és az eredményeket, HORVÁTH et al. 2006b pedig a digitális képeket tartalmazza).

A geomorfológiai elemzés kiindulási adataként a terület 10 m felbontású digitális domborzati modellje (Honvéd Térképészeti KHT, Magyarország) szolgált, amely 1:50 000-es szintvonaltérképek adatainak interpolációjából készült.

Alkalmazott módszerek

A terület elemzése során több szempontot is figyelembe vettünk: (i) a terület földtani és geomorfológiai fejlődéstörténetét, (ii) geomorfológiai jellemzők halmozódásának vizsgálatát és (iii) tisztán eróziós és szerkezetföldtanilag meghatározott felszíni formák megkülönböztetését. Ezek közül a harmadik szempontot tekintettük a jelen kutatásban a legfontosabbnak. E megkülönböztetés során feltételeztük, hogy MÉSZÁROS szerkezetföldtani térképe terepi megfigyeléseken alapszik, és mint ilyet adattartalmát tényként fogadjuk el. Ebből kiindulva a domborzatmodell elemzése kifejezetten az ezen a térképen található vetők osztályozására és nem a vetők helyének vizsgálatára irányult (*1. ábra, b*). Továbbá e szerkezetföldtani térkép tartalmát használtuk vonatkoztatási alapként, amelyhez a többi szerző szerkezeti elemeit hasonlítottuk. Ezáltal többek között számszerűsített adatot tudunk szolgáltatni a térkép minőségének megítélésére. Az összes, MÉSZÁROS 1982-es szerkezetföldtani térképén található vetőt, tekintet nélkül a vélhetően terepen megfigyelt elmozdulási jellegükre, két csoportba osztottuk: (i) geomorfológiailag megnyilvánuló és (ii) egyéb. Szükség esetén a vetőt két vagy több szakaszra bontottuk (*3. ábra*). A MÉSZÁROS-féle vetők darabolási alapjaként egy kompozitképet használtunk, mely a terület lejtőszögtérképéből, az árnyékolt domborzatmodellből és a magassági szintvonalakból állt. A *3. ábrán* jól kivehető, hogy egy vető mentén a domborzat jelentősen változhat, a vető egy darabon völgyben, illetve annak közvetlen közelében fut, később



3. ábra. A MÉSZÁROS-féle vetők darabolásának bemutatása a kutatási terület egy példaként kiemelt részterületén

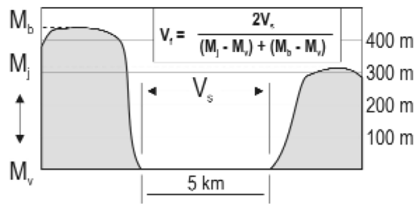
Ezt a geomorfológiailag megnyilvánuló és meg nem nyilvánuló szakaszokra történő darabolást az egész Bakony és Balaton-felvidék területére elvégeztük és az így keletkezett kétféle szakaszcsoportot a továbbiakban külön elemezzük. Háttér: árnyékolt DDM10 domborzatmodell és hozzátartozó szintvonalak 10 m szintközzel

***Figure 3.** Splitting scheme of the MÉSZÁROS-faults shown as an example for a part of the study area*

This split into segments with and without geomorphologic surface expression was done for the entire Bakony and Balaton Highland area. The two distinct groups were further analysed separately. Background: hillshaded DDM10 digital elevation model and related contour lines with 10 m vertical spacing viszont nem köthető futásához semmilyen jellegzetes geomorfológiai felszínforma. Mint később látni fogjuk, ez a megfigyelés jól mutatkozik a két kategóriához tartozó lejtőszögeloszláson is. A geomorfológiailag megnyilvánuló vetőszakaszok lejtőszögeloszlása bimodális trendet mutat, vagyis mind a völgyfenékhez köthető alacsony lejtőszögeket, mind a völgy oldalaihoz köthető meredek lejtőszögeket tartalmaz. A geomorfológialag meg nem nyilvánuló vetőszakaszok lejtőszögeloszlása sokkal egyveretűbb, túlnyomóan sík területre utaló alacsony lejtőszögértékeket tartalmaz. A kompozitképen ezáltal megbízhatóan ki lehetett választani a megfelelő kategóriába eső vetőszakaszokat. A döntéshozatal JORDÁN et al. (2003) eljárása alapján történt, ami feltételezi, hogy szerkezeti elemek, illetve szerkezetföldtanilag meghatározott felszínformák jellegzetesen lineáris elemekként mutatkoznak. Ilyen elemek lehetnek eltolódások esetében egyenes gerincvonalak, egyenes völgyek, egyenes, váltakozó meredekségű lejtők. Normál vetők esetében egyenes, meredek lejtők, vagy a lejtő meredekségében bekövetkező változás, amely szintén egy lineáris trendet követ (JORDÁN et al. 2003). A többi szerző vetőadatbázisát nem hasonlítottuk össze a kompozitképpel, mivel jelen dolgozatnak e vizsgálat nem képezi tárgyát.

Kiválasztott területeken, geomorfológiailag megnyilvánuló vetőszakaszok mentén a digitális domborzati modell alapján a völgyek keresztmetszetét vizsgáltuk a V_f–index segítségével (KELLER & PINTER 1996). A V_f– index a völgy keresztmetszeti alakját számszerűsíti (*4. ábra*). A völgykeresztmetszet alakjának egyik szélső esete a teknővölgy és KELLER & PINTER (1996) szerint a völgy fenekén kanyargó és oldalirányban erodáló folyóra utal. A másik szélső eset a szurdokvölgy, mely a völgyben lezajló erős bevágódásra és egyúttal a terület erős (relatív) emelkedésére utal.

A litológiai egységek geomorfológiai jellemzését az egész térképszelvény területére alkalmazott magasság-hisztogramokkal egészítettük ki, mivel ez a geomorfológiai elemzés egyik alapvető technikája. Ezt a vizsgálatot érdemes megtenni, mert egyes kőzettípusok (pl. karbonátos



4. ábra. A V_f-index számítása (KELLER & PINTER 1996 után)

Az ábra egy sematikus völgykeresztmetszetet mutat két különböző magasságú völgyoldallal. M_v: völgyfenék magassága, M_j: jobb völgyoldal magassága, M_b: bal völgyoldal magassága, V_s: völgyfenék szélessége

Figure 4. *Calculation of the V_f-index or valley floor width to height ratio (modified after KELLER & PINTER 1996)*

The figure shows a schematic cross section of a valley with valley flanks showing different elevations. M_v: elevation of valley floor, M_j: elevation of right valley flank, M_b: elevation of left valley flank, V_s: width of the valley floor

kőzetek, alacsony viszkozitású lávából keletkezett vulkanikus kőzetek) esetén fontos jelenségekre utalhatnak a magasság-hisztogramok tulajdonságai. Ez még akkor is igaz, ha — mint esetünkben — a területet jelentős részben érintette és érinti a bevezetésben már említett differenciális kiemelkedés. A magasság-hisztogramokat a digitális domborzatmodellből számoltuk MÉSZÁROS térképén ábrázolt egységes litológiájú területekre. A domborzatmodellből a kiválasztott litológiai egység területére eső részt kivágtuk és az azon a területen található magassági értékeket ábrázoltuk hisztogram formájában.

Diszkusszió

Magasság-hisztogramok

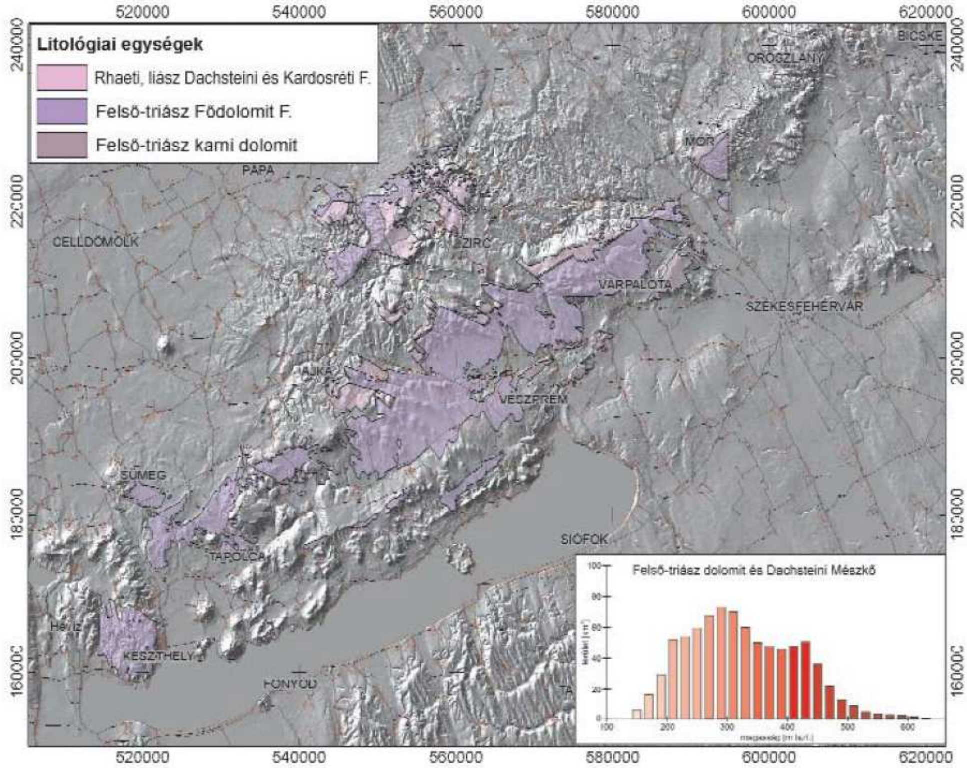
A magasságeloszlásokat kiválasztott, főbb litológiai egységekre készítettük el: a felső-triász (Fődolomit, Dachsteini Mészkő Formáció), alsó-jura (Kardosréti Mészkő) karbonátokra, felső-triász márgarétegekre (Veszprémi Formáció), oligocén képződményekre (Csatkai Formáció), miocén képződményekre és bazaltos egységekre.

A karbonátos litológiai egységek (*5. ábra*) esetében jellegzetes bimodális eloszlás mutatkozik: az eloszlás csúcsai 300 és 400 m tengerszint feletti magasságok körül fekszenek. Ez ugyan szerkezetföldtanilag befolyásolt elrendezés — ugyanis Balatonfüredtől nyugatra a Fődolomit egy ÉK–DNy tengelyű szinklinális magjában települ (BUDAI et al. 1999) — viszont nem köthető az általunk vizsgált vetőmintázathoz.

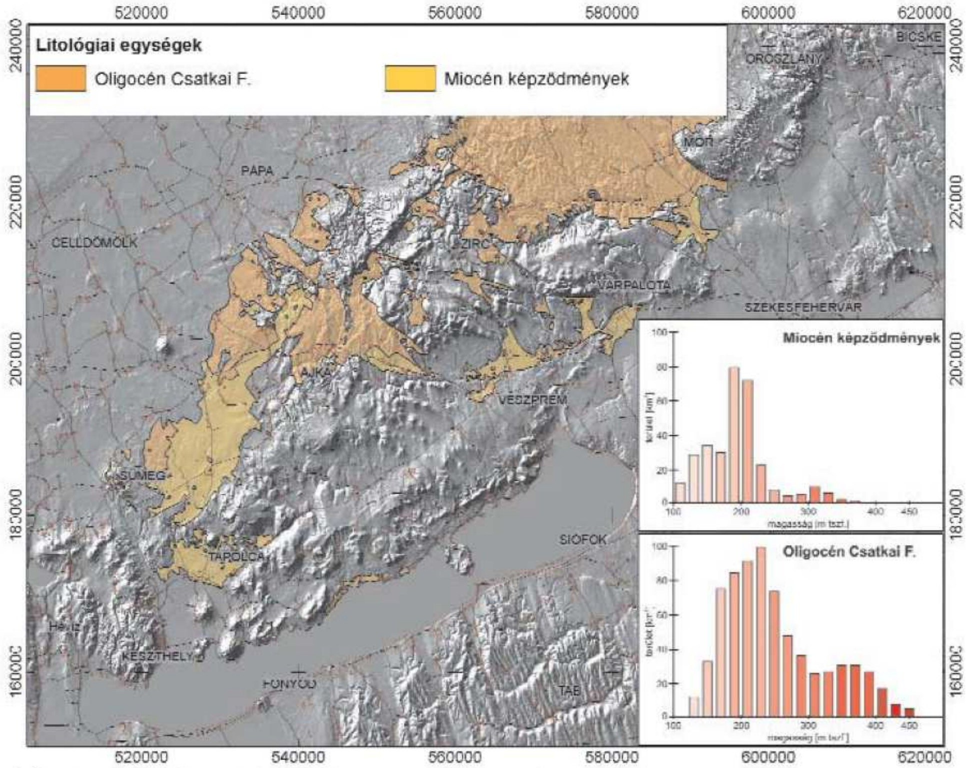
Az oligocén és miocén képződmények magasságeloszlása (*6. ábra*) szintén enyhén bimodális. Az egységek fő magassági tartománya 200 és 230 m között fekszik, de egy alárendelt csúcs 300 m, illetve 350 m tengerszint feletti magasságnál világosan elkülönül. Az oligocén képződmények MÉSZÁROS térképén megoszlanak a Bakony ÉNy-i szélén fekvő Marcal-medence alacsony térszínei (*1. ábra*) és a Bakonyvidék dombságának magasabb fekvésű területei között, amit a differenciális kiemelkedésnek tulajdonítunk. A miocén esetében megfigyelhető, hogy a Bakonyvidéken található félárkokban települ. Egy szép példa erre a Veszprém északi határában található miocén előfordulás.

A márgarétegek (*7. ábra*) 150 és 320 m tengerszint feletti magasság között egyenletesen oszlanak el. Elterjedésük a Balaton-felvidék középső részén jellegzetes; a délkeleties lejtés és a kőzettípus nagyobb eróziós érzékenysége határozza meg a hisztogram alakját.

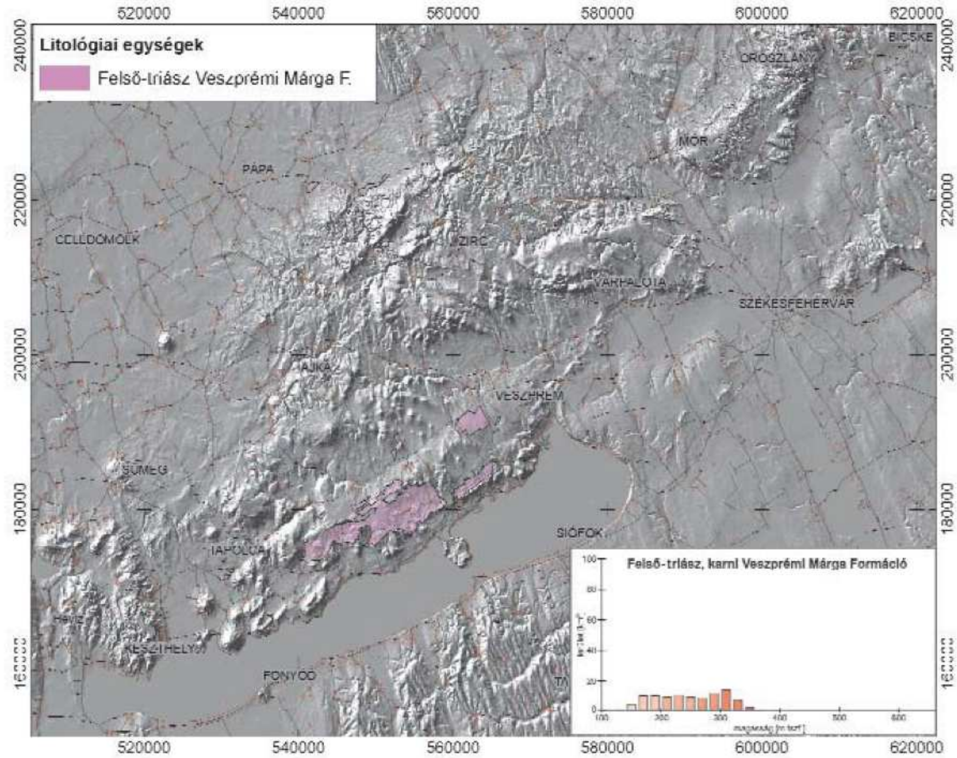
Noha a bazalt (*8. ábra*) az előbbiekben tárgyaltaknál jelentősen kisebb területet foglal el, hisztogramja széles, lényegében egycsúcsú, de enyhén mégis bimodális jelleget mutat. A csúcs a tanúhegy-jelleghez köthető, míg a széles eloszlás és a másodlagosan gyakori magasságok a vulkáni képződmények korbeli és genetikai sokféleségéhez kapcsolhatók. Így például a 100 és 200 m közé eső magassági tartományt a Tihanyi-félsziget, valamint a



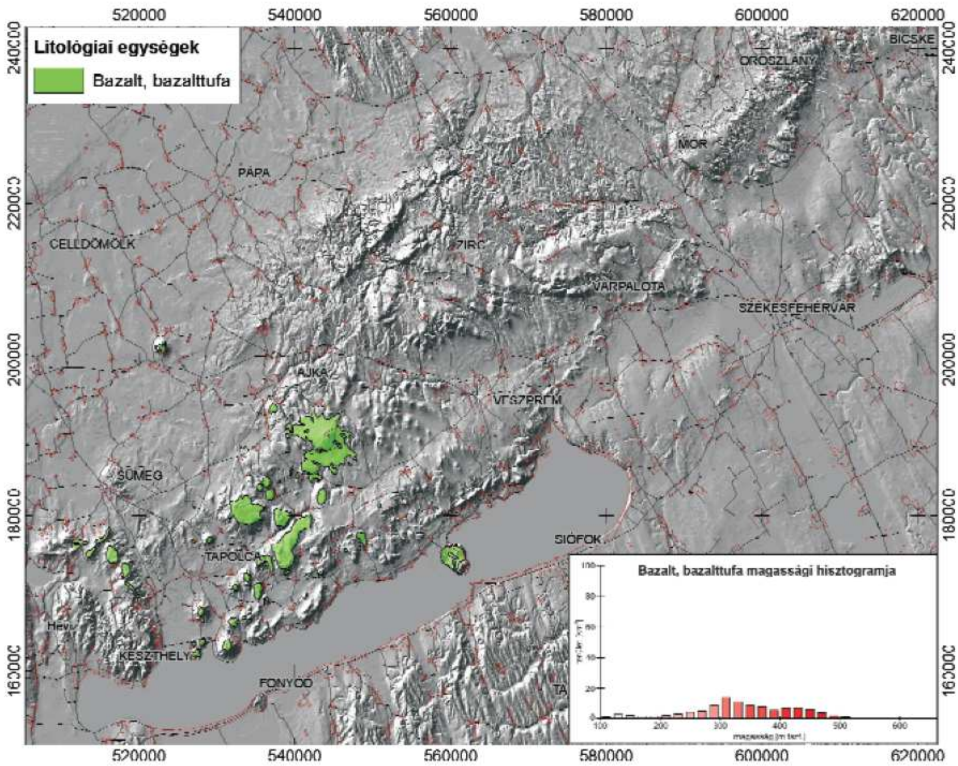
5. ábra. A karbonátos kőzetek elhelyezkedése és magassági hisztogramja
Figure 5. *Position and elevation histogram of carbonate lithology*



6. ábra. A márga jellegű kőzetek elhelyezkedése és magassági hisztogramja
Figure 6. *Position and elevation histogram of marly lithology*



7. ábra. Az oligocén és miocén képződmények elhelyezkedése és magassági hisztogramja
Figure 7. *Position and elevation histogram of Oligocene and Miocene formations*



8. ábra. A bazalt, bazalttufa elhelyezkedése és magassági hisztogramja
Figure 8. Position and elevation histogram of basaltic units

Szigliget alacsonyan fekvő bazaltos térszínei alkotják. 280 és 460 m között nagyjából egyenletes az eloszlás, ami a szabályos kúp alakú Kab-hegy (599 m; *1. ábra*) következménye.

Vetők csapásirány-elemzése

A következőkben a csapásirány-vizsgálat eredményeit elemezzük. A könnyebb áttekintés érdekében egy rövidített nevezéklistát vezettünk be, amelyet a *2. táblázat* összegez. A *3. táblázat* áttekintést nyújt az elemzésben szereplő MÉSZÁROS-féle vetők szakaszainak hosszáról és darabszámairól, mutatva a statisztikai elemzés robusztusságát. A feldolgozás és kiválasztás után a felszínformához köthető vetők csoportjába 695 szegmens tartozott 1420 km összhosszal. Az egyéb vetők kategóriájába 429 szegmens volt sorolható 867 km összhosszal.

H06B, H06T: egész Dunántúl és Bakonyvidék. Már a H06T és a H06B jelű csoportok összehasonlítása is tanulságos eredményt hoz. Mindkét csoport keletkezéstől független, viszont geomorfológiailag megnyilvánuló lineamensek, melyek egy adatbázisból származnak, csupán területileg osztottuk fel őket. Az Északi- és Déli-Bakonyra leszűkített lineamens-rendszerben az ÉÉNy–DDK kissé keletiesebb és egy elkülönülő ÉNy–DK-i komponens is megjelenik (*9. ábra, b*).

H06T, M83T: egész Dunántúl és MÉSZÁROS-féle vetők. Az M83T érdemi eltérést mutat a H06T-től: a H06T-ben kevésbé jellemzőek a konjugált irányok (*9. ábra, a* és *c*). A H06T fő ÉÉNy–DDK irányú lineamenseihez képest az M83T túlnyomórészt ÉNy–DK (alárendeltebben NyÉNy–KDK), konjugált iránynak pedig ÉÉK–DDNy csapású lineamenscsoport tekinthető. Ez utóbbi a H06T-ben rendkívül alárendelt. Az M83G irányeloszlása még sokkal inkább csúcsosabb, itt ÉNy–DK irány dominál és erős a konjugált iránynak interpretált ÉÉK–DDNy-i irány.

M83T, M83G, M83E: a MÉSZÁROS-féle vetők és a domborzat összefüggése. Az M83T és M83G lineamensek irányeloszlásában a feltolódások érdemben befolyásolják az irányeloszlás konjugált voltát (*9. ábra, d–f*). Az irodalom a geológiai szerkezetfejlődés során a bakonyi feltolódások aktív fázisát eoalpinak tekinti (TARI 1995), az oldaleltolódások elvetik a feltolódások szerkezeti vonalait (MÉSZÁROS 1983, DUDKO 1991). Ilyenformán a feltolódások képződése és fő aktív fázisa az oldalelmozdulásokhoz képest sokkal régebbi.

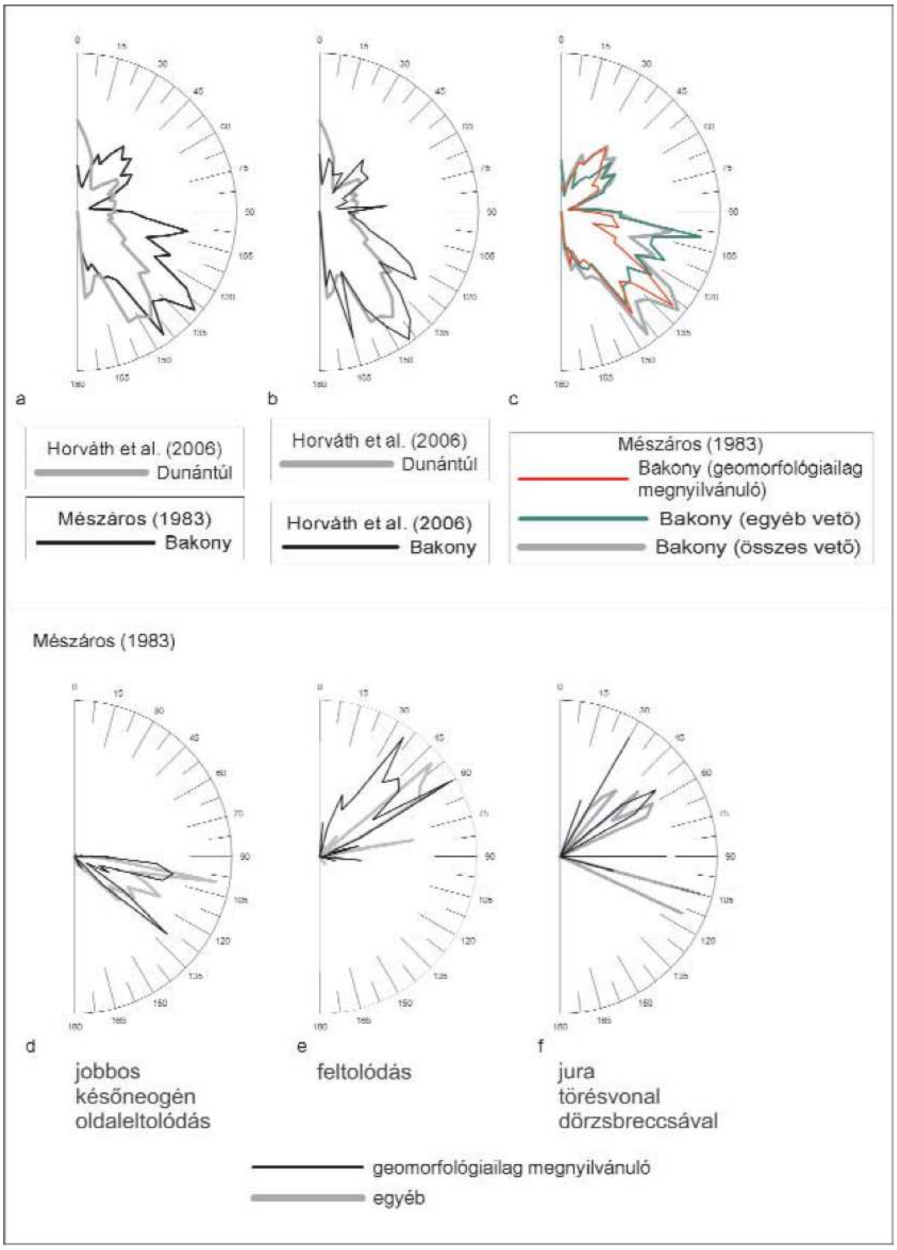
Feltételezzük, hogy a geomorfológiailag aktív vonalak vagy fiatalabbak, vagy valamilyen okból felújultak és kipreparálódtak. A kipreparálódást nem tartjuk véletlenszerűnek a vonalak erősen irányított irányeloszlása miatt. A Bakonyvidék kiemelkedése következtében az erózió által lepusztított felszínen kirajzolódnak az egykor fedett vetőrendszerek. A feltolódások geomorfológiai megnyilvánulását vagy a feltolódások felszíni folyamatok általi kipreparálódásának köszönhetjük, vagy a feltolódások mentén felújult tektonikus mozgásnak. A geomorfológiailag

2. táblázat. A felhasznált vetők tartalma és a szövegben használt rövidítése
Table 2. *Descriptions and used abbreviations of the investigated faults*

Tartalom	Eredeti adat területe, térképi méretaránya	Hivatkozás	Részcsoport	Rövidítés
Szerkezetföldtani térképezés	Bakony, 1: 100 000	MÉSZÁROS (1983)	összes vető geomorfológiailag megnyilvánuló egyéb	M83T M83G M83E
SRTM-DDM -en térképezett morfostrukturális elemek	Pannon-medence, 1: 100 000 – 1: 1 000 000	HORVÁTH et al. (2006)	az egész Dunántúl területére a Bakony területére	H06T H06B
Szeizmikus szelvények értelmezése alapján	Dunántúl, 1: 1 000 000	TARI (1996)	feltolódások, normál vetők, jobbos/balos eltolódások	T96
Hidrológiai térképezés	Zala, 1: 500 000	JOCHA-EDELÉNYI (2005)	vetők általában	J05T
Szerkezetföldtani térképezés	Balaton-felvidék, 1: 363 636	DUDKO (1991)	feltolódások, normál vetők, jobbos/balos eltolódások	D91

3. táblázat. A MÉSZÁROS-féle vetők statisztikailag jelentős osztályokra bontott darabszáma és az ezekbe az osztályokba tartozó vetők összhossza
Table 3. *Numbers and total length of segments, which fall within statistically meaningful classes of the MÉSZÁROS-faults*

Típus	M83T	M83G	M83E
Balos fiatalabb mozgások során vetővé felújult szubhercini oldaleltolódás	6 db 33,36 km	5 db 27,18 km	2 db 6,18 km
Feltolódás	47 db 75,31 km	33 db 55,45 km	17 db 20,3 km
Jobbos késő neogén oldaleltolódás	91 db 396,04 km	53 db 163,04 km	65 db 234,25 km
Vetőket megszabó jura törésvonal széles dörzsbreccsa zónával	15 db 26,61 km	8 db 10,25 km	9 db 16,35 km



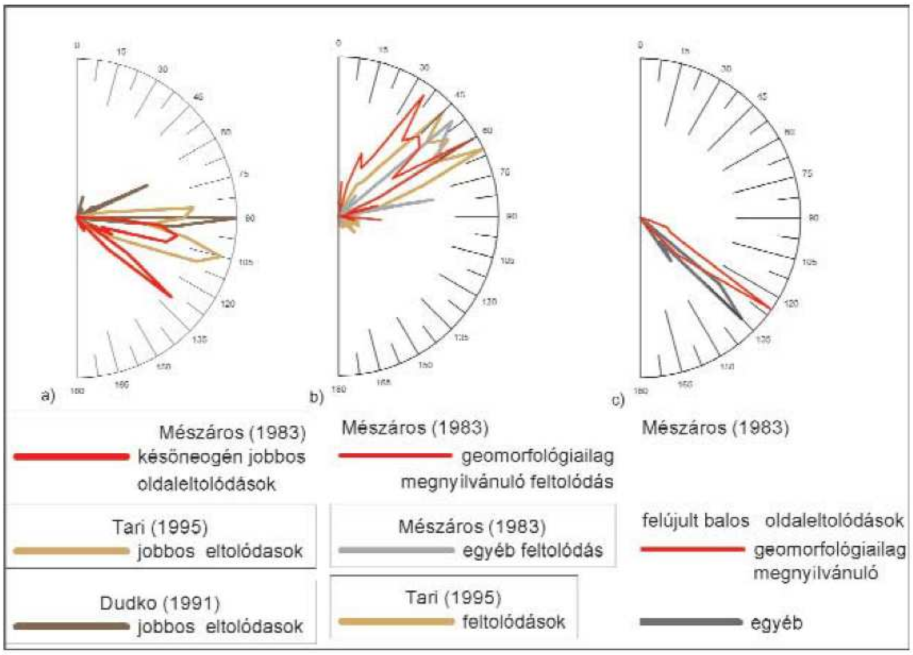
9. ábra. A feldolgozott MÉSZÁROS-féle vetők rózsadiagrammjai és az összehasonlításhoz felhasznált morfostrukturális elemek (HORVÁTH et al. 2006) elemek irányeloszlása 0–180° között, 5°-os felosztásban

***Figure 9.** Rose diagram of the analysed MÉSZÁROS-faults and the morphostructural elements (HORVÁTH et al. 2006) between 0 and 180°, bin size is 5°*

megnyilvánuló és a többi vető között jól kimutatható különbség van. A feltolódások esetében ez 20°-os, a jobbos késő-neogén vetők esetében 5–10°-os eltérés mutatkozik.

M83G, T96, D91: a MÉSZÁROS-féle vetők és más szerzők által térképezett vetők. Ebben az összeállításban további osztályokra bontottuk a MÉSZÁROS-féle vetőket és összehasonlítottuk őket más szerzők szerkezetföldtani adatbázisaival (*10. ábra*).

MÉSZÁROS geomorfológiailag megnyilvánuló késő-neogén jobbos oldaleltolódásainak csapásiránya jó összhangban van TARI (1996) által szeizmikus szelvényeken interpretált jobbos vetőivel (*10. ábra, a*). Fontos megfigyelni, hogy az M83G osztályon belül a jobbos vetők két csapásirányra oszlanak: egy KDK és egy DK irányú csoportra. Ugyanakkor a T96-os csapásirányaiban is megfigyelhető egy enyhén bimodális eloszlás (K és KDK). A D91-be tartozó vetők főleg a T96-os osztály keleti irányú vetőivel esnek egybe és az M83G KDK csapásirányú vetői



10. ábra. A MÉSZÁROS-féle vetők összehasonlítása más szerzők szerkezetföldtani elemeivel

***Figure 10.** Comparison of the MÉSZÁROS-faults with fault interpretations from other authors*

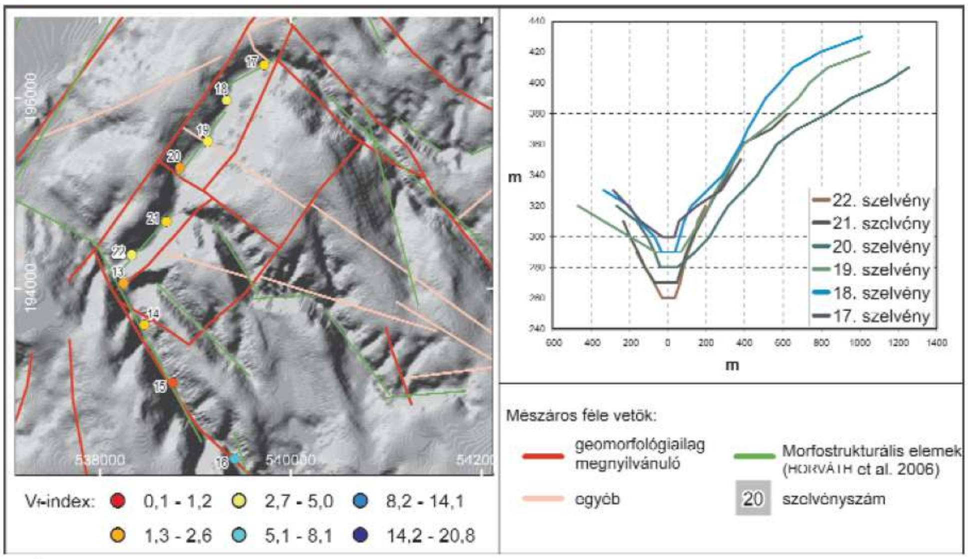
egy körülbelül 12°-os elfordulást mutatnak a T96 és D91 osztályokba tartozó vetőkhöz képest.

Az M83G, M83E és T96-os osztályokba tartozó feltolódások egyveretű eloszlást mutatnak (*10. ábra, b*), viszont itt is észrevehető egy 5–7°-os eltérés az M83G és a többi osztály között.

Végül az M83G és M83E felújult balos vetői között ez az 5–7°-os eltérés megint mutatkozik (*10. ábra, c*).

V_f-index

A völgykeresztmetszetre számolt V_f-index értékei 0,06 és 20,8 között váltakoznak. A kiválasztott területeken két fő trend figyelhető meg: egyrészt az alacsony, szurdokvölgyre tipikus V_f-értékek szimmetrikus völgykeresztmetszeteknél mutatkoznak (pl. Ajkától keletre, a Csinger-patak mentén; *11. ábra*). Másrészt a magas, teknővölgyekre jellegzetes



11. ábra. A Csinger-patak völgyének domborzati keresztszelvényei és azok fekvése az árnyékolt domborzatmodellen

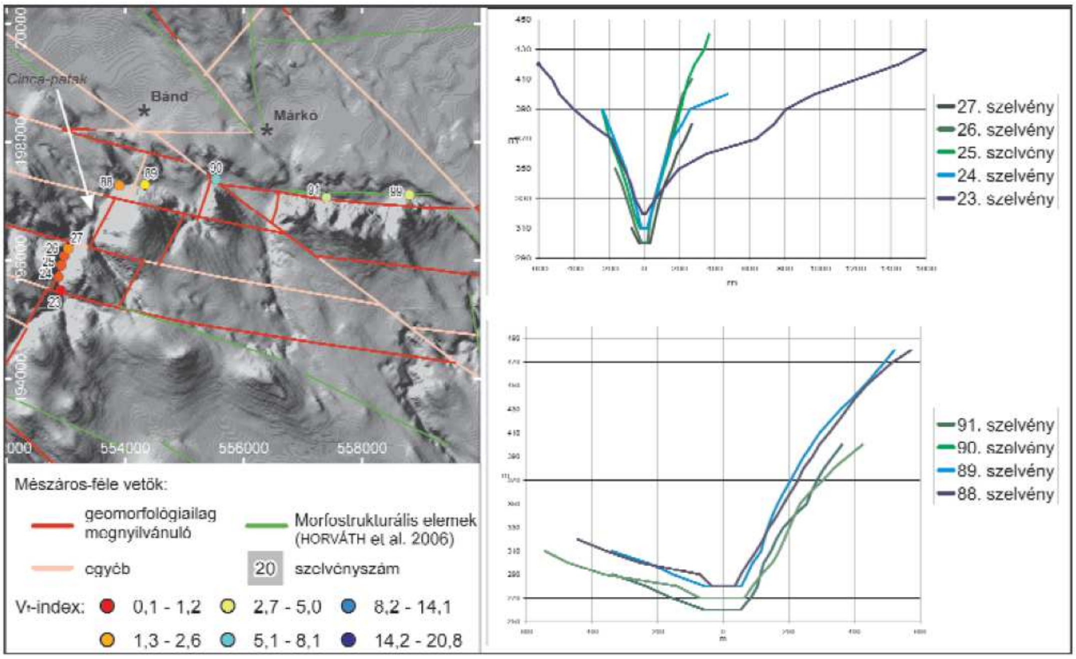
A keresztszelvények számai a térképi nézeten is megjelennek az arra a szelvényre számolt V_f-index színnel kódolt értékével együtt

***Figure 11.** Cross sections of the valley of the Csinger Stream. The location of the cross sections is indicated on the hillshaded digital elevation model*

The numbers of each cross section are also shown on the map view along with the color-coded V_f-index values értékek aszimmetrikus völgykeresztmetszetekkel párosulnak (pl. Szentgál közelében a Cinca völgyében és Ajkarendek közelében a Borsod-völgy mentén; *12.* és *13. ábra*). Bár a völgyaszimmetria egyes szerzők szerint utalhat deflációs eredetre is, jelen tanulmányban kifejezetten a MÉSZÁROS által észlelt, illetve interpretált vetőket és az azokhoz köthető formákat vizsgáljuk. Így definíció szerint a formákat tektonikus eredetűnek tekintjük, nem zárva ki azt a lehetőséget, hogy egyes esetekben a formák kialakulásához eolikus folyamatok is hozzájárulhattak.

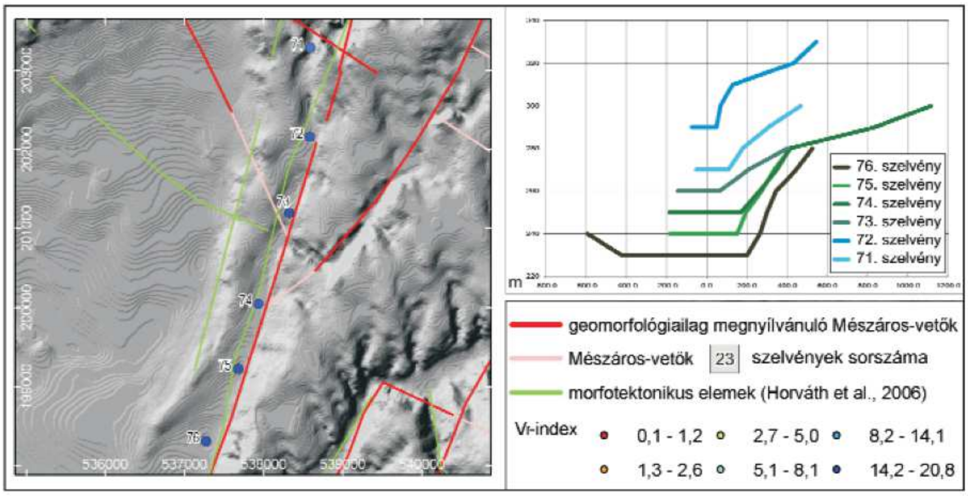
Következtetések

A MÉSZÁROS-féle szerkezetföldtani térképből előállított térinformatikai vektoradatbázis további szerkezetföldtani és



12. ábra. A Cinca völgyének domborzati keresztszelvényei és azok fekvése az árnyékolt domborzatmodellen
A keresztszelvények számai a térképi nézeten is megjelennek az arra a szelvényre számolt V_f-index színnel kódolt értékével együtt

Figure 12. *Cross sections of the valley of the Cinca Stream. The location of the cross sections is indicated on the hillshaded digital elevation model*
The numbers of each cross section are also shown on the map view along with the color-coded V_f-index values



13. ábra. A Széles-völgy domborzati keresztszelvényei és azok fekvése az árnyékolt domborzatmodellen
A keresztszelvények számai a térképi nézeten is megjelennek az arra a szelvényre számolt V_f-index színnel kódolt értékével együtt

Figure 13. *Cross sections of the Széles Valley. The location of the cross sections is indicated on the hillshaded digital elevation model*
The numbers of each cross section are also shown on the map view along with the color-coded V_f-index values

geomorfológiai vizsgálatok elvégzésére megfelelő pontosságú. A térkép a Bakonyvidék kutatásának egy fontos állomását képviseli. A MÉSZÁROS által alkalmazott vetőosztályozás a maga korában előremutató jellegű volt, noha a térkép földtani tartalma nem korunk modern felfogásait tükrözi.

A térinformatikai vektoradatbázis és más szerzők szerkezetföldtani, ill. morfostrukturális adatbázisainak összehasonlításából kimutatható, hogy az Északi- és Déli-Bakony szerkezeti elemeinek fő csapásiránya az egész Dunántúl morfostrukturális elemeinek csapásirányához képest 10–15°-kal elfordul (H06B és M83T jelű csoportok a *9. ábra a* részén). MÉSZÁROS geomorfológiailag megnyilvánuló jobbos eltolódásai is legalább egy ilyen nagyságrendbe eső mértékkel (15°) vannak elforgatva más szerzők a Bakonyvidék nyugati előterében fekvő jobbos vetőivel szemben (*10. ábra, a*). A morfostrukturális elemek irányeloszlását tekintve fontos szem előtt tartani, hogy ezen elemek keletkezése részleteiben nem tisztázott — lehet a tektonika, vagy más hatások eredménye. Jelen tanulmányban az irányok közötti eltérést indikáció szintűnek tekintjük, mélyebb okát további földtani, geomorfológiai, illetve morfometriai vizsgálatokkal kell kutatni.

A MÉSZÁROS-féle térképen található vetők és a domborzat között a Bakonyvidék területén statisztikailag kimutatható összefüggés létezik. Geomorfológiailag megnyilvánuló és nem megnyilvánuló szerkezeti elemek jól elkülöníthető csoportokat alkotnak (*9. ábra, c* és *10. ábra*). Ezért a szerkezeti elemek kipreparálódása sem tekinthető teljesen véletlenszerűnek.

A V_f-index esetében a teknővölgyekre jellemző értékek csupán akkor figyelhetők meg, ha a völgy egyik oldala jóval alacsonyabb a másiknál és így erősen aszimmetrikus völgykeresztmetszetek keletkeznek. Az összes többi vizsgált völgy szurdokvölgyekre jellemző értékeket mutat. Az értékek alapján a legjellegzetesebb szurdokvölgyek karbonátos kőzetekben és a MÉSZÁROS térképén ábrázolt vetők mentén fordulnak elő.

Köszönetnyilvánítás

A kutatást a T47104 számú OTKA-pályázat keretében végeztük. Köszönjük HORVÁTH Ferencnek a MÉSZÁROS-féle térkép eredeti példányának rendelkezésünkre bocsájtását. ZÁMOLYI Andrást a Bécsi Egyetem FA536001 *[Karpatian Tectonics]* projektje támogatta. SZÉKELY Balázs a kutatómunka egy részét Békésy György posztdoktori ösztöndíjasként végezte. Köszönjük KISS Adrienn és CSILLAG Gábor szakszerű és gondos bírálatát.

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Kézirat beérkezett: 2010. 03. 30.

18. Curriculum Vitae

Name: András Zámolyi

Academic Degree: Mag. rer. nat.

Date of Birth: 24.07.1980

Citizenship: Hungarian

Postal address: Auhofstrasse 231-237/ 4/ 7

1130 Wien

eMail address: andras@zamolyi.info

Mobile phone: +43 699 11009744

Education:

January 2013:	Initiation of cumulative PhD thesis.
8th of February, 2010:	Finalization of PhD lectures at Eötvös University, Budapest (Abszolutorium).
September, 2007:	Admission to PhD program of study at University of Vienna (joint degree contract with Eötvös University, Budapest)
September, 2006:	Admission to PhD program of study at Eötvös University, Budapest
	Proposed PhD thesis title: “Quaternary landscape evolution of the Little Hungarian Plain”
4th of July, 2006:	Graduation at University of Vienna, diploma thesis title:
	“Tectonic geomorphology of Serifos (Cyclades, Greece).
	Specialization field: structural geology, metamorphic petrology, GIS.
1st of October, 2003 – 1st of January, 2004:	ERASMUS – exchange semester at the Eötvös University, Budapest faculty of natural sciences, Department of General and Historical Geology.
	Scientific coordinator: Dr. Csontos László. Passing of five subjects related to the field of the diploma thesis (Structural Geology, Geomorphology, GIS, and Remote Sensing).
18th – 27th of July, 2003:	Participation on the 5th School of the European Mineralogical Union and ERASMUS Intensive Program on Ultra High Pressure Metamorphism, at Eötvös University, Budapest.
1st of December, 2000:	First partial exam on the subject of earth sciences at University of Vienna. Specialization: structural geology, metamorphic petrology, geomorphology.
1st of October, 1998:	Immatriculation at University of Vienna, faculty of natural sciences, earth sciences class.
16th of June, 1998:	A – level exam („Fadinger“ grammar school Linz, Austria).

Zámolyi András: “Quaternary landscape evolution of the Little Hungarian Plain”

Professional Experience (University-related):

Tutor at 1st FP7 EUFAR Training Course

ADvanced Digital Remote sensing in Ecology and earth Sciences Summer School, 2010 Tihany (Hungary) Tasks: defining geomorphologic/geologic targets, flight planning, organization of field data acquistion

Lecturing tutor at University Vienna

University lecture: “GIS and 3D modelling” (creation of geological maps/databases in ArcGIS)

Geological mapping practice: 2005 St. Michael, 2003 Alt Aussee (Austria), 2002 Sondershausen (Germany)

Lecturing tutor at Eötvös University, Budapest

University lecture:	“Digital elevation models (DEM) in geology” (creation and analysis of DEMs)
Geophysical field practice:	2006, 2007, 2008 Balaton Highlands (Central Hungary)
	2009 Zala Hills (SW-Hungary)

Professional Experience (General):

January 2012 to current: Junior Geoscientist at OMV Austria Exploration & Production, Production Geology Department, Gänserndorf.

March 2011 – January 2012: Employee of University of Vienna, scientific collaborator in the project Qartaba Anticline (Structural evolution of the Qartaba Anticline, Lebanon) funded by OMV (Geologic and geomorphologic analysis of study area in ArcGIS environment based on remote-sensing data, digital elevation models and field work).

March 2007 – March 2011: Employee of University of Vienna, scientific collaborator in the projects Karpatian Tectonics I and II (Miocene tectonic evolution of Vienna Basin and adjacent areas) funded by OMV (Creation of a tectonic model from seismic, borehole and outcrop data).

January 2006: Professional co-translator of the school-atlas “Geography from Space” (“Iskolai Űratlasz – Földrajz a világűrből”), published by the Hungarian Space Office (Magyar Űrkutatási Iroda), GeoData Services Ltd. and Geospace Austria.

January 2005

– April 2005: Scientific collaborator in the Austrian Research Fund (FWF) project “Celts in the Hinterland of Carnuntum” of the Department for Prehistory and Early History, University of Vienna (Geomorphological aerial photo interpretation and drainage pattern evolution analysis in the Vienna Basin using ArcView software).

March 2004

– March 2007: Employee of GeoData Services Ltd.; field of work:

- Participation in the project “Remote-Sensing Control of area-based subsidies” of the European Union for parts of the member state Germany. Satellite image orthorectification, spatial database creation and maintenance.

- Digitalization of municipial facilities maps (City of Mainz, Germany), resolving of inconsistencies in the dataset, digital map drawing, and translations.

Language Skills:

Hungarian (native language), German (native language), English (fluent writing and speaking)

Other Skills: Driving licence (category B), Software: ArcGis Desktop 9.1 to 10, ArcView 3, Landmark Geographix 2007.2, Erdas Imagine, ENVI, CorelDRAW & PHOTOPAINT, MS Office, Surfer, Bentley Microstation.

Project Reports:

Decker, K., Beidinger, A., Hoprich, M., Lee, E.Y., Zámolyi, A.(2011): “Tectonics of the Slovak part of the Vienna Basin and the adjacent Western Carpathians during the Lower Miocene.” OMV Final Report Karpatian Tectonics II.

Beidinger, A., Hölzel, M., Hoprich, M., Zámolyi, A., Decker, K. (2009): “Tectonic evolution of the Vienna Basin and the Waschberg Zone during the Early Miocene.” OMV Final Report Karpatian Tectonics I.

Selected Publications:

Zámolyi, A., Draganits, E., Doneus, M., Fera, M. (2012): Paläoflusslaufentwicklung der Leitha (Ostösterreich) – eine Luftbildperspektive. In: Doneus M. & Griebl M., Die Leitha – Facetten einer archäologischen Landschaft. Archäologie Österreichs Spezial, 3, 11-23.

Dorninger, P., Székely, B., Zámolyi, A., Roncat, A. (2011): Automated Detection and Interpretation of Geomorphic Features in LiDAR Point Clouds. Österreichische Zeitschrift für Vermessung und Geoinformation (VGI), begutachteter Beitrag, 99, 2; 60 - 69.

Zámolyi, A., Székely, B., Draganits, E., Timár, G. (2010): Neotectonic control on river sinuosity at the western margin of the Little Hungarian Plain, Geomorphology, 122, 231-243.

Hölzel, M., Decker, K., Zámolyi, A., Strauss P., Wagreich, M. (2009): Lower Miocene structural evolution of the central Vienna Basin (Austria), Marine and Petroleum Geology, 27, 666-681.

Székely, B., Zámolyi, A., Draganits, E., Briese, C. (2009): Geomorphic expression of neotectonic activity in a low relief area in an Airborne Laser Scanning DTM: A case study of the Little Hungarian Plain (Pannonian Basin), Tectonophysics, 474, 1-2, 353-366.

Selected Conference Abstracts:

Zámolyi, A., Lee, E.Y., Beidinger, A., Hoprich, M., Strauss, P., Decker, K. (2010): Miocene deformation of the central Vienna Basin (Austria-Slovakia). Geophysical Research Abstracts, 11, 10097, ISSN 1029-7006.

Beidinger, A., Decker, K., Zámolyi, A., Lee, E.Y., Hoprich, M., Strauss, P. (2010): Oligocene to Miocene kinematics of the Outer West Carpathians and the Vienna Basin area. Geophysical Research Abstracts, 11, 6804, ISSN 1029-7006.

Zámolyi, A., Székely, B., Biszak, S. (2010): Assessing the accuracy of the Second Military Survey for the Doren Landslide (Vorarlberg, Austria). Geophysical Research Abstracts, 12, 9974, ISSN 1029-7006.

Zámolyi, A., Székely, B. (2009): Structural geological environment of the Doren landslide (Vorarlberg, Austria) derived from LiDAR DTM analysis. Geophysical Research Abstracts, 11, 12903, ISSN 1029-7006.

Beidinger, A., Decker, K., Zámolyi, A., Hölzel, M., Hoprich, M., Strauss, P. (2009): Palinspastic reconstruction of the Alpine thrust belt at the Alpine-Carpathian transition - A geological Sudoku. Geophysical Research Abstracts, 11, 12015, ISSN 1029-7006.

Zámolyi, A., Decker, K., Hölzel, M., Strauss, P., Wagreich, M. (2008): Variations in deformation characteristics along the front of the Alpine-Carpathian wedge (Waschberg-Zdanice Unit, Austria-Czech Republic). 6th Meeting of the Central European Tectonic Studies Group (CETeG) – 13th Meeting of the Czech Tectonic Studies Group (ČTS), Proceedings and Excursion Guide, ISBN 978-80-89343-01-0.

Selected Presentations:

“GIS supported seismic mapping – Examples and experiences”. Invited oral presentation, Department of Geological Sciences, Masaryk University, Brno, 2010.

“Tectonic Evolution of the Vienna Basin and the Waschberg Unit during the Early Miocene”. Invited oral presentation, Austrian Geologic Society (ÖGG) lectures at University of Vienna, 2008.