1. Introduction

Lakes are ephemeral phenomena from a geological point of view because most of them exist for some (ten) thousands of years only (e.g., Martens et al., 1984). Nonetheless, they particularly mirror climatic and hydrologic changes, tectonics, pollution, etc. very well. These characters are detectable in the nearby past as well as in deep time (e.g., De Batist and Chapron, 2008). Lake deposits are, therefore, excellent high-resolution time archives of earth and life history (e.g., Martens et al., 1994). Nonetheless, they particularly mirror climatic and hydrologic changes, tectonics, pollution, etc. very well.

A multidisciplinary study was performed on a c. 30 m thick, limnic–deltaic sequence in the Styrian Basin (Austria). Geophysical (gamma ray activity, rock magnetism), geochemical (organic carbon, sulphur) and sedimentological observations were combined with palaeontological information (mainly ostracods). On this base, several ecological factors were deduced (terrigenous influx, salinity, water depth and oxygenation).

Based on integrated stratigraphy as well as on palaeomagnetic results the whole section is set to Chron C5r.2r–1n (11.308–11.263 Ma) and covers a period of less than 45 kyr. In addition to the long-term record, we analysed short-term changes by high-resolution sampling (5 mm sample interval; ostracods, magnetic susceptibility). This ~2.3 m-thick interval spans in total ~3500 yr with an individual sample resolution of a few years only. The combination of these data permit the description of the palaeoenvironmental evolution of the section in detail: at the base of the section, the development of a conifer swamp on the lakeside of Lake Pannon is documented. This almost freshwater swamp existed for some centuries before it became extinct within a few decades due to a rise of the water-table. The drowning is related to a transgression of Lake Pannon, which triggered the establishment of a brackish-water fauna as well as greigite formation in the lake sediments. In general, the ongoing transgression favoured benthic life due to increased salinity (up to mesohaline conditions). The high-resolution ostracod and magnetic susceptibility record reflect short-term fluctuations in bottom-water ventilation. These oscillations probably range in the order of centuries and decades and are possibly related to climatic shifts. Later, the successive deepening of the lake resulted in a significant faunal turnover. A meromictic system with a well-established, oxygen-depleted hypolimnion developed. Finally, the limnic phase was replaced by a prograding deltaic system, where the amplified input of coarse-grained material and freshwater also affected benthic life. Periodic changes in clay (illite) content were detected by the gamma ray-log throughout the entire section. The observed cycles (~5–12 kyr) of the gamma ray-log may reflect a super-ordinate modulation of the sedimentary record by climatically forced changes in precipitation or run-off.

© 2011 Elsevier B.V. All rights reserved.

One prominent example for a fossil ancient lake is Lake Pannon, which evolved in Central Europe as a remnant of the Paratethys Sea around the Middle/Late Miocene boundary (Kázmér, 1990; Magyar et al., 1999; Rögl, 1999; Harzhauser and Piller, 2007; Fig. 1a). Due to this marine heritage, in the early phase the lake water was brackish, slightly alkaline and slowly freshening (Harzhauser et al., 2007). At its maximum extent (~10–9 Ma) Lake Pannon covered an area of c. 290,000 km² with a water depth of up to 800 m (Magyar et al., 1999; Harzhauser and Mandic, 2008). Its palaeogeography and hydrology were largely shaped by tectonic activities in and around the Pannonian basin system (e.g., Kováč et al., 2000; Cloetingh et al., 2005; Csato et al., 2007). It remains a matter of debate if a glacio-
eustatic sea-level lowering or tectonic uplift of the Carpathians caused the isolation of Lake Pannon from the sea (Lirer et al., 2009; Vasiliev et al., 2010). However, climatic changes, possibly astronomically forced, may have modulated its development as well (Juhász et al., 1997; Sacchi and Müller, 2004; Jiménez-Moreno et al., 2005; Harzhauser et al., 2007, 2008; Lirer et al., 2009). Whereas Lake Pannon was initially influenced by the latest Middle Miocene dry climate and represented a meromictic system, it switched into a monomictic one due to increased humidity during the Middle Pannonian (Harzhauser et al., 2007; Böhme et al., 2008, 2010). Lake Pannon is well known for impressive radiation events among aquatic organisms (i.e., molluscs, ostracods; Papp et al., 1985; Geary et al., 2002). This stimulates the introduction of a regional stratigraphic stage “Pannonian” (~Tortonian) and provides the base for a high-resolution biostratigraphy (Papp et al., 1985; Magyar et al., 1999).

The present paper investigates the complex evolution of an ~30 m thick, limnic–deltic sequence, which was deposited at the western margin of early Lake Pannon about 11.3 Ma (Mataschen clay pit; SE-Austria; 15°57'16"E/46°54'15"N; Fig. 1b). We describe a complete transgression–regression cycle based on sedimentological, geophysical, geochemical and palaeontological data. Our study evaluates the long-term development of that outcrop (tens of thousands years) as well as short-time changes (decades–centuries) by high-resolution sampling of the most crucial part of the profile.

2. Section description

The section of the Mataschen clay pit starts with a >1.5 m thick succession of laminated sandy pelites and partly ripple-bedded fine-medium sands (Gross, 2004a; Fig. 2). The top of these strata is bioturbated by roots and represents the floor of the pit (= 0.0 m of the section). Here, 3–4 m high, autochthonous Glyptostrobus-tree trunks are regularly found at distances of ~10–15 m throughout the outcrop (~700×200 m).

From 0.0 to 0.3 m of the section densely packed, coalified plant fragments with clayey interlayers occur. They contain a low diverse, azonal plant assemblage and sporadically vertebrate remains (Daxner-Höck, 2004; Gross, 2004b; Kovar-Eder, 2004; Meller and Hofmann, 2004). Upsection (0.3–0.8 m) follows a layer of laminated clay in which the plant content decreases upwards. Infrequently remains of unionid bivalves, insects, cyprinid fishes as well as amphibian and bird fossils were found (Schultz, 2004; Tempfer, 2004; Engel and Gross, 2008). Towards the upper part of that layer (~0.4–0.8 m) an almost monospecific coquina of the dreissenid mussel Mytilopsis neumayri is observed (Harzhauser, 2004).

From 0.8 to 7.5 m massive to laminated (silty) clays with two sandy intercalations at 5.5 and 6.0 m follow. Partly associated fish

Fig. 1. Location of the Mataschen section at the western margin of Lake Pannon. (a) Extent of Lake Pannon during the earliest Late Miocene (after Magyar et al., 1999). (b) Facies map of the Styrian Basin (Lower Pannonian).
skeletons occur between ~1.5–3 m; articulated specimens of the large dreissenid bivalve Mytilopsis ornithopsis were found rarely between ~2 and 3 m. The sandy interlayers display turbiditic features (parallel lamination at the base followed by climbing ripples).

Between 7.5 and 27.0 m the sediments consist of alternations of clayey silts and fine sandy silts with sandy intercalations and display a general coarsening-upwards trend. Sandy beds are often rich in plant detritus and occasionally enclose diaspores (Meller and Hofmann, 2004). Close to the top (~26.5 m), fine sandy silt layers yielded a highly diverse macro-flora (Kovar-Eder and Hably, 2006).

The top of the mined pelite deposit is formed by a >2.5 m thick, large-scale cross-bedded medium to coarse sand, which is overlain by alternations of laminated fine sandy silt and ripple-bedded sand layers.

3. Materials and methods

In the field, natural gamma ray activity (Heger-probe) and magnetic susceptibility (Exploranium KT-9-probe) were measured at a sampling interval of ~50 mm. Twenty-eight bulk samples were
taken for micropalaeontological as well as geochemical analyses. Micropalaeontological results of 13 samples are discussed in Gross (2004a). Analyses of total carbon (TC) and total sulphur (TS) contents were performed on a LECO CS 300 elemental analyser (dried and homogenised samples, 0.10–0.15 g/sample). Prior to total organic carbon (TOC) analysis, samples were treated with 2-molar HCl at 20 °C. The concentrations obtained represent the arithmetic mean of three measurements and are expressed as percent of dry weight (wt.%).

Eighty-one palaeomagnetic standard cubes (20 × 20 × 20 mm) were subjected to stepwise demagnetisation (alternating field treatment from 2 to 140 mT). Palaeomagnetic data analyses included principal component analysis based on visual inspection of orthogonal projections. Stepwise saturation, measurements of the coercivity, demagnetisation of the saturation magnetisation and Curie-point determinations helped to identify the magnetic minerals. Measurements of the anisotropy of the magnetic susceptibility (AMS) were carried out to describe the sediment texture. Natural remanent magnetisation (NRM) was measured on a three-axis cryogenic dc-squid magnetometer with in-line degauser (2G Enterprises). AGICO MFK instruments were used for measuring low-field magnetic susceptibility and its anisotropy.

For high-resolution investigations the lower part of the section (profile-interval 1.04–3.31 m) was continuously cored with a percussion-drill (five sediment cores of ~0.5 m length each and 100 mm in diameter). Magnetic susceptibility of the cores was measured with a MS2E1 Bartington-probe (sample interval = 5 mm). One half of each core was sampled for micropalaeontological analysis (sample interval = 5 mm = 20 cm3 sediment). These samples were treated like the micropalaeontological bulk samples: dried (~40 °C, 24 h) and washed using diluted hydrogen superoxide for disintegration through standard sieves (H2O2:H2O = 1:5; 63/125/250/500 μm). Residuals ≥250 μm were picked out completely (cores) or were split in a some cases (bulk samples; ESM 1).

Grain size and mineralogical data from an earlier study by Rantitsch et al. (2004) are included here. Data points represent one section-metre of homogenised sediment and are adapted to our profile.

For statistical analyses, the software package PAST 1.92 was used (Hammer and Harper, 2008).

4. Results

4.1. Geophysical results

4.1.1. Natural gamma ray activity (γ)
The recorded γ-values decrease from the base to the top of the section. Pelitic proportions show elevated gamma ray activity in contrast to sandy/silty layers (Fig. 2).

4.1.2. Magnetic susceptibility (κ)
Measurements of κ directly the field recorded enhanced signals in lower part (up to ~7.5 m) with a prominent peak at around 2.2 m (Fig. 2). One additional, significant κ-exursion at 0.8 m is documented by palaeomagnetic laboratory analyses only due to inaccessibility of that profile part during outcrop-κ-logging (Fig. 3). Higher up in the section (>7.5 m) the magnetic signal is weak. In particular, the sandy layers are marked by very low κ.

The κ-log of the high-resolution sampling interval (Fig. 4) documents low values between from 1.04 to 1.76 m (= interval A), which progressively rise (1.76–2.19 m; interval B1) and reach a maximum at 2.19–2.31 m (= interval B2). Above, κ-values decrease markedly (2.31–2.61 m; interval B3), exhibit accentuated peaks (2.61–2.94 m; interval B4) and distinctly drop down at 2.94–3.31 m (= interval C).

4.1.3. Palaeomagnetic analyses
Anisotropy of low-field magnetic susceptibility proves an undisturbed primary sedimentary fabric: magnetic lineation (L=k1/k2) is typically below 1.01 and the average magnetic foliation (F=k2/k3) is 1.044 (ESM 2).

During demagnetisation, well-defined demagnetisation paths with one or two components of NRM and a good separation of the unblocking coercivity spectra could be obtained. Typically, the NRM decayed slowly at low demagnetising fields and the samples were fully demagnetised at 100 mT alternating-field strengths (Fig. 3a). The characteristic remanence components yield mostly normal polarity except in the lowermost part of the section, where indications for a polarity change are observable (Fig. 3b).

A significant deflection of the demagnetisation vector above 50 mT was observed in many samples, which can be attributed to the presence of the Fe-sulphide greigite (Fig. 3a). Accordingly, the isothermal remanence magnetisation (IRM) acquisition curves and backfield experiments indicated the dominance of low-coercivity magnetic carrier minerals with remanence coercivities below 50 mT. This is typical for magnetite and/or greigite.

4.2. Geochemical results

Total organic carbon (TOC) and total sulphur (TS) contents are low and almost co-vary through the section (Fig. 2, ESM 1; TOC: mean/max/min = 0.78/2.67/0.19 wt.%, TS: mean/max/min = 0.16/0.53/0.02 wt.%). All major positive excursions of TOC are clearly related to layers, which are rich in leaf-remains (samples at 0.30, 0.75 and 26.40 m) or coaly plant detritus (19.80 and 22.40 m). TS-data display a similar pattern with amplified values at the aforementioned layers. Additionally, elevated TS-content is measured between 2.60 and 5.50 m.

4.3. Ostracods

The ostracod fauna (28 species in total) is dominated by the genera Cyprideis, Loxoconcha, Hemicytheria and by candonids (mainly Caspiolla, Fabaeformiscandona, Lineocypris and Typhlocypris; Fig. 2). The population structure (adult and numerous juvenile specimens) as well as the preservation of ostracods (translucent, no signs of corrosion or decalcification) point to fast burial and to almost autochthonous assemblages (Whatley, 1988).

In the lowermost part of the outcrop (up to 0.75 m), poor ostracod faunas are reported, which comprise Darwinulina, candonids and ilyocyprinids (samples at ~0.95 and 0.30 m) or are barren of ostracods (samples at ~0.55, 0.10 m). The bulk samples at 1.25 and, especially at 2.60 m yielded rich and diverse ostracod faunas that largely consist of Cyprideis, Loxoconcha, candonids and Hemicytheria; darwinulids and ilyocyprinids are absent (Fig. 2).

Results of high-resolution ostracod sampling are similar to the general trend of the high-resolution κ-log (Fig. 4). Ostracods are rare in intervals A and C (mean = 20 valves/sample) and comparably rich in interval B (mean = 106 valves/sample). Considering the four major ostracod groups, which comprise 98.1% of the total fauna, all display similar, small-scale fluctuations in abundance throughout this record (δHonset = 27,728 valves; Cyprideis = 55.8%, Loxoconcha = 27.8%, candonids = 9.9%, Hemicytheria = 4.6%). However, some significant taxonomic shifts are evident: within interval A, co-varying Cyprideis and Loxoconcha contents progressively increase, whereas candonids and Hemicytheria remain at a low level. Cyprideis and Loxoconcha increase up to interval B3. Both genera make up ~90% of the total fauna at intervals B1–B3 (δHonset = 16,361; Cyprideis 59.4%, Loxoconcha 29.9%; candonids 4.6%, Hemicytheria 4.7%). Above that peak (interval B4), Hemicytheria almost vanishes and candonids (Lineocypris, Typhlocypris) become a more important component of the fauna (δHonset = 6928 valves; Cyprideis 60.1%, Loxoconcha 23.8%, candonids 14.9%, Hemicytheria 0.6%).
In interval C, all groups noticeably decline. Both, *Hemicytheria* and *Loxoconcha* are virtually absent. Numbers of *Cyprideis* distinctly decrease and candonids are an essential faunal element (total = 1546 valves; *Cyprideis* 50.3%, candonids 45.7%, *Loxoconcha* 3.2%, *Hemicytheria* 0.7%).

Bulk samples above the high-resolution sampling interval demonstrate a gradual decline of the ostracod fauna (Fig. 2). While *Cyprideis* and candonids prevail with distinctly reduced numbers within the record, the amount of loxoconchids and especially of hemicytherids is considerably limited. Samples between 9.20 and 16.30 m lack ostracods or contain only scattered candonid valves. At 13.65 m, a limited fauna is documented. All samples above are barren of ostracods.

5. Discussion

5.1. Gamma-log

The γ-log mirrors the general coarsening-upward trend of the section (Fig. 2). The increase in grain size depends in the lower part (up to ~7.5 m) mainly on rising amounts of silt in respect to clay. Upsection (>7.5 m) increased sand content is responsible for the coarsening of the sediments (Rantitsch et al., 2004).

The gamma ray activity clearly corresponds to variations in clay contents and is linked to the quantity of illite, which is the principal clay mineral at Mataschen. Other clay minerals (chlorite, smectite, mixed-layer minerals) and feldspars (anorthite, albite) are present in low concentrations only (Rantitsch et al., 2004). We assume that largely K⁴0 of the illite accounts to our γ-signal (Cripps and McCann, 2000; Potter et al., 2005). Thus, at Mataschen, the γ-log is a valuable estimator of the illite/clay content and a robust recorder of changes in sedimentation.

5.2. Palaeomagnetics

Field measurement of magnetic susceptibility roughly reflect the lithological subdivision of the outcrop into a lower clay-rich (up to 7.5 m) and an upper, more silty–sandy part (>7.5 m; Figs. 2 and 3).
A comparison of the susceptibility obtained by laboratory measurements with the intensity of saturation remanence yields a discrimination of two sample groups: greigite (Fe₃S₄) is the dominant magnetic mineral in the lower portion of the section, while the upper and very bottom part are characterised by a magnetic phase with lower saturation remanence values (SIRM), indicative of relatively coarse grained, detrital magnetite. The same trend can be observed in the SIRM/κ-ratio (Fig. 3b; ESM 3). A SIRM/κ-ratio in the range of 70,000 A/m is regarded as typical for greigite, whereas magnetite exhibit ratios between zero and 20,000 A/m (Maher and Thompson, 1999).

Greigite can be formed in situ under oxygen-depleted conditions during diagenesis (Berner, 1984; Roberts and Weaver, 2005; Rowan et al., 2009). Especially in settings with low dissolved sulphate concentrations, limited organic carbon supply and readily available reactive iron, greigite develops as a precursor to pyrite (Kao et al., 2004; Rowan et al., 2009). However, greigite is also produced by anaerobic magnetotactic bacteria around the oxic/anoxic transition zone (Mann et al., 1990; Pan et al., 2004; Kopp and Kirschvink, 2008).

In both cases high sedimentation rates and limited post-depositional sulphur flux in the pore waters favour its preservation due to restricted post-burial oxidation as well as pyritisation (Kao et al., 2004; Rowan and Roberts, 2006; Rowan et al., 2009).

Several studies have demonstrated that greigite is the dominant magnetic mineral in sediments of Lake Pannon and the Late Miocene–Pliocene Eastern Paratethys (Babinszki et al., 2007; Vasiliev et al., 2008, 2010). Its occurrence is shown to be mainly related to early diagenesis and/or to the preservation of magnetosomes. In the first case, greigite forms authigenetically deeper within the sediment (up to several metres) with an uncertain lag in time (up to kyr). In the second case, greigite is created within the sulphidic zone already during sedimentation (Roberts and Weaver, 2005; Rowan and Roberts, 2006; Babinszki et al., 2007; Vasiliev et al., 2007, 2008; Hüsing et al., 2009; Rowan et al., 2009).

In the clay-rich portions at Mataschen, where greigite is recorded, rapid burial and restricted post-sedimentary pore water flux can be assumed based on the preservation of ostracods and the autochthonous
tree trunks (Sections 4.3 and 6.2). Consequently, greigite is probably of nearly syndepositional origin and the preserved NRM is acquired during deposition of those sediments (Vasiliev et al., 2010).

5.3. Organic carbon and sulphur contents

Elevated TOC-values are clearly related to plant-rich layers and hence reflect the amount of land-derived material and terrigenous input respectively (Fig. 2; Meyers and Ishiwatari, 1993; Cohen, 2003; Meyers, 2003; Potter et al., 2005). This is in accordance with low hydrogen- and high oxygen-indices throughout the section, which point to detrital vascular-plant matter as the major contributor to TOC (Rantitsch et al., 2004; ESM 1).

TS-values are low and co-vari with TOC-contents. This suggests that much sulphur is related to organic compounds and the TS-record roughly mirrors changing organic matter input (Holmer and Storkholm, 2001). Only four samples in the lower part of the section display low TOC/TS-ratios (samples at 2.60, 4.05, 4.45 and 5.50 m).

In lakes, dissolved sulphate is typically short in supply, as microbial sulphide production is limited and results in comparably low sulphur concentrations in limnic sediments. Organic matter contents, however, remain relatively high (Berner, 1984; Holmer and Storkholm, 2001). Therefore, TOC/TS-ratios are supposed to be a rough estimator of palaeosalinity (Cohen, 2003; Potter et al., 2005; Russell and Werne, 2009). By accepting an insignifican loss of carbon during diagenesis, low TOC/TS-ratios in Matachener hint to elevated salinity between 2.60 and 5.50 m (Draxler et al., 1994; Rantitsch et al., 2004). There, waters that are more saline raised the TS-contents in contrast to under- and overlying strata, where detrital organic matter input controls TS.

5.4. The long-term ostracod record

The rare ostracod faunas obtained from bulk samples of the lowermost layers (up to 0.75 m; Fig. 2) with commonly freshwater dwelling darwinuilids, candonids and illyocypridids indicate a fluvial to marginal lacustrine environment (Gross, 2004a).

Rich and diverse ostracod associations at 1.25 and 2.60 m (mainly *Cyprideis, Loxoconcha*, canondids and *Hemicytheria*) refer to a rise in salinity up to mesohaline waters (Gross, 2004a; Rundić, 2006; compare lowering of TOC/TS-ratios between 2.60 and 5.50 m discussed in Section 5.3). Extant *Cyprideis* is able to cope with fluctuating salinities, however, it mainly occurs in mesohaline waters (see compendium in Meisch, 2000). Endemic Panonian species of *Hemicytheria* and *Loxoconcha* as well as of *Cyprideis* are supposed to be a heritage of the marine Paratethys Sea and are probably most sensitive to a lowering of salinity (e.g., Jiříček, 1985; Gross, 2004a; Rundić, 2006; Gross et al., 2008; Starek et al., 2010). In addition, some candonid taxa (*Fabaformiscandona*, *Lineocypris*, *Thyphlocypris*) are thought to tolerate brackish waters (Gross, 2004a; Rundić, 2006; Cziczer et al., 2008). In these layers, a “typical” Lake Pannon ostracod community (e.g., Jiříček, 1985) replaced the freshwater ostracod fauna from the preceding samples (below 0.75 m).

Upsection, abundance and diversity of ostracods notably decrease (bulk samples above 4.05 m). The number of loxococonids and, in particular, of hemicytherids drops significantly, which argues for a gradual lowering of salinity (Gross, 2004a; Rundić, 2006). However, the lack of noded *Cyprideis*-values implies that salinity never went down below oligohaline conditions (Keyser, 2005; Gross et al., 2008). The recorded loxococonids are at least partly phytoplilous and *Cyprideis* mainly occurs in shallow waters (Meisch, 2000; Harzhauser et al., 2008). In combination with the relative increase of sublittoral candonids (*Lineocypris, Thyphlocypris; Cziczer et al., 2008*), a deepening of the environment is suggested.

Above 7.5 m, the input of coarser grained sediments might have distinctly affected ostracod populations in a negative way (“siltation”; Cohen, 2003; Cohen et al., 2005). Upon 16 m, the almost complete lack of calcite (Rantitsch et al., 2004) matches with the disappearance of ostracod valves. Diagenetic dissolution most likely modulated the fossil record there.

5.5. The high-resolution sampling interval

Based on the high-resolution χ-log, three intervals are distinguished in the core samples. Interval B can be further subdivided (Fig. 4). The general trend of the χ-log and the abundances of ostracods are similar. Both show low values at intervals A and C and elevated ones at interval B.

The ostracod record of intervals A–B3 refers to the gradual development of a shallow, brackish environment, which is related to a transgression (compare Section 5.4). Most probably, the successively increased influx of saline waters delivered more dissolved sulphate and stimulated high-susceptibility greigite formation, which is expressed by elevated χ-values. Thus, the principal environmental parameter for both, the χ- and the ostracod-log, is a transgression of Lake Pannon.

At interval B a remarkable faunal turnover is observed (decrease of *Cyprideis* and *Loxoconcha*, almost disappearance of *Hemicytheria*, increase of candonids; compare Section 5.4), which is probably linked to slightly decreasing salinity and a rising lake-level. At interval C, the marked decrease in *Loxoconcha*, the sharp decrease of *Cyprideis* and the relative increase of sublittoral candonids reflect a continuation of that deepening trend upwards. Due to increased water depth, bottom-waters possibly became more oxygen-depleted and thus, unfavourable for benthic life. However, interval C still reflects a brackish water environment (*Cyprideis* without nodes; sublittoral candonids). Relatively high sulphate supply due to the ongoing transgression and more extensive dysoxic/anoxic conditions probably favoured the formation of low susceptible pyrite instead of greigite (Berner, 1984; Rowan et al., 2000). This might cause the breakdown of the χ-log at interval C synchronously to the ostracod record.

In detail, taxonomical shifts (see Section 4.3) obstruct a plain statistical correlation of ostracod groups. Nevertheless, highly co-varying abundances of the main ostracod taxa are evident (e.g., *Cyprideis* and *Loxoconcha* at interval B; correlation of log-transformed data: \( r = 0.82, p < 0.001 \)). Similar high-frequency oscillations are visible in the χ-log.

The ostracod taxa discussed here are epibenthic, only penetrating the sediments some millimetres dependent on the position of the redox-front (Corbari et al., 2004; Frenzel and Boomer, 2005). Although some of them (i.e., *Cyprideis* or some candonids) might be able to survive hypoxia for some time, they certainly prefer oxic settings (Gamenick et al., 1996; Jahn et al., 1996; Modig and Ölabfsson, 1998; Corbari et al., 2004; Starek et al., 2010). Apparently, the growth of the hingement and of water currents may affect preservation of articulated valves (Whatley, 1988). However, the lack of closed ostracod carapaces at Mataschen suggests that they largely lived upon the better oxygenated sediment surface. The “butterfly” position of lymnocaridid bivalves points to the same direction. Thus, the observed small-scale changes of ostracod abundances, regardless of their taxonomic affiliation, indicate short-term alternations of the oxygen status of bottom-waters (Boomer et al., 2005).

The results of palaeomagnetic investigations (Section 5.2) imply that the χ-record largely depends on the presence of early diagenetic and/or bacterial greigite. While authigenesis may occur with some delay, bacterial greigite can be created within the sulphidic zone only a few millimetres below the flourishing, epibenthic ostracod populations (Cziczer et al., 2008; Vasiliev et al., 2008). Most likely, both greigite producing mechanisms contribute to our record. These shape the χ-signal at varying rates and hamper further comparisons with the ostracod-log (e.g., correlation of χ and *Cyprideis* at interval B: \( r = -0.22, p = 0.001 \)). However, if we consider the dense sample intervals of 5 mm and assume limited post-depositional pore water
flux in these pelitic sediments, we suggest that the $\kappa$-log mainly reflects high-frequency changes of the redox-front and of bottom-water ventilation respectively (Larrasoňa et al., 2003).

Based on the palynological analyses of interval B conspicuous short-term shifts in mean annual precipitation are implied (Kern et al., submitted). Hence, climatic oscillations (wet and dry periods, which cause variations in run-off) could be the trigger of the observed changes in oxygenation, e.g., due to variations in nutrient input or episodic mixing of the water column.

6. Age model, sedimentation rate and possible cyclicities

6.1. Dating of the Mataschen section

The pelitic interval of the Mataschen section belongs to the Lower Pannonian Feldbach Formation (Myilopsis ornithopsis zone; Gross, 2003, 2004a; Harzhauser, 2004; Fig. 5). Seismic data gives evidence for a remarkable gap to underlying Middle Miocene deposits, which is widely used for intrabasinal correlation within the Pannonian Basin (e.g., Magyar et al., 1999; Sacchi and Horvath, 2002; Harzhauser et al., 2004; Kováč et al., 2004, 2008; Schreilechner and Sachsenhofer, 2007). The Feldbach Fm. is unconformably overlain by fluviolacustrine sediments of the Paldau Formation (Myilopsis hoernesi zone). There, the three-toed equid Hippotherium – a cornerstone of continental biostratigraphy – is recorded for the first time in the Styrian Basin (Mottl, 1970; Bernor et al., 1988; Mein, 1999; Aguilà et al., 2004).

Recently, Lirer et al. (2009) proposed dating the Sarmatian/Pannonian-boundary in the Vienna Basin to 11.42 Ma and the first occurrence of Hippotherium to 11.26 Ma. Based on striking similarities (lithology, biotic contents) between the Styrian and the Vienna Basin (Papp, 1951; Kolmann, 1965; Harzhauser et al., 2004), these dates bracket the age of the Mataschen section to about 11.42–11.26 Ma.

Palaeomagnetic data document normal polarity throughout the Mataschen section. Indications for reversed polarity are only observed at its very base (Section 4.1). For Myilopsis–bearing outcrops close-by, reverse and normal polarity is recorded (personal observation, R.S.). This suggests an onset of the lowermost Pannonian sedimentation in the Styrian Basin close to a magnetic reversal. If we consider the aforementioned time window, a correlation to the short Chron C5r.2r–1n (11.308–11.263 Ma) is strongly implied (Hüsing et al., 2007; Vasiliev et al., 2010). This reduces the maximum time range of our section to about 45 kyr. Deposition at Mataschen probably started close to the offset of that Chron but comprises most likely only the lower part of it because normal polarity is still observed in the lowest part of the overlying Paldau Fm. (Fig. 5). Thus, it is possible that Mataschen comprises only about half of the duration (–22 kyr) of Chron C5r.2r–1n.

However, if we assume a maximum time range of <45 kyr and claim a relatively constant sedimentation (which is admittedly not the case in such a marginal area), the average sedimentation rate is >0.7 mm/yr (maximal ~1.4 mm/yr). This is in coincidence to estimations for Lake Pannon (Lirer et al., 2009: 0.65 mm/yr) and lacustrine–deltaic sequences in the close-by Dacian Basin (Vasiliev et al., 2004: 0.8–1.4 mm/yr).

6.2. Evidence of chronology derived from fossil tree trunks

The most impressive feature of Mataschen are in situ preserved Glyptostrobus–tree trunks, which represent one of the oldest and largest non-petritified fossil forests (Kázmér, 2008; Erdei et al., 2009).

Counting of tree rings indicates an age of several centuries for stumps of ~1.5 m in diameter (measured 1 m above their base). This renders the existence of the swamp forest to some 100s of years. Conspicuously, growth rings become very dense in the outermost ~10–15 mm, which represent approximately ~10–30 yr and hint at suppressed growth at the end of their life (Hunter et al., 2006).

Recent Glyptostrobus–swamps commonly grow in shallow waters (~0.5–1 m depth) but an excessive rise of the water level heavily reduces their vitality (Avaryanov et al., 2009). At Mataschen, such a swamp seems to have withstood a rising water table for a few decades (water depths <1 m) before it died (water depths >1 m) and became completely drowned. Because almost all excavated stumps are of similar height (~3–4 m), the water level must have risen quickly up to this level at least in order to minimise wood degradation. Avaryanov et al. (2009) showed an almost complete decomposition of subaerially exposed parts of dead Glyptostrobus–stumps just within 25 years.

In Mataschen, burial certainly started at the beginning of submergence when the trees were still alive. But, even if the ambient waters were generally not very well aerated, final burial must have happened within a few millennia at the most. Otherwise, the trunks would have been heavily affected, if not completely destroyed by microbial activity (e.g., Kim and Singh, 2000). Unfortunately, a dendrogeomorphologically based estimation of the sedimentation rate (e.g., Hupp and Morris, 1990) was not possible because compaction–induced slickensides along the trunk surfaces sheared off eventually existing adventitious roots.

Nevertheless, by considering a sedimentation rate of >0.7 mm/yr and the preserved height of the trunks (~3–4 m), complete burial must have been taken place within less than 5 kyr. This underlines the fast deposition at Mataschen.

6.3. Possible cyclicities

We are aware that many autogenetic (e.g., channel/delta avulsion, flood events) as well as allogetic processes (e.g., variations of basin subsidence or morphology of the catchment area) contribute to the sedimentary sequence in such a marginal area of Lake Pannon. Due to probable variations and the broad range of estimated sedimentation rates, a conversion of computed cycles into time remains uncertain. Here we only aim to present the observed periods and to outline their possible time scale.

Time series analyses performed on the $\gamma$-log reveal statistically significant periodicities at ~7.7 m, ~5.6 m and ~3.3 m (Fig. 6a). Based on the approximated sedimentation rate, these periods range...
in the order of 11.5 kyr, 8.4 kyr and 5 kyr. Previously, the interference of millennial-scale (“sub-Milankovitch”) climatic oscillations on Neogene lake sediments has been discussed, which could affect the palaeoenvironmental development of Mataschen as well (e.g., Willis et al., 1999; Cosentino et al., 2005; Harzhauser et al., 2008; Utescher et al., 2009). Power spectra of the high-resolution record of the two most frequent ostracod taxa at interval A—B.4 exhibit significant periods at ~0.48 m and 0.36 m for *Cyprididae* and two almost significant peaks at ~0.51 m and 0.19 m for *Loxoconcha* (Fig. 6b). Further peaks are present in both Lomb periodograms, but remain below the 95% confidence level (0.15, 0.13, 0.07, 0.06 m). For the high-resolution-ν-log only a period of 0.37 m is likely significant (Fig. 6b).

If the assumed sedimentation rate is correct, these periods range in the order of centuries (~0.50 m ≈ 750 yr; ~0.36 m ≈ 540 yr; ~0.19 m ≈ 285 yr), whereas weakly expressed cycles of higher order cover a decadal time scale (0.07 m ≈ 105 yr; 0.06 m ≈ 90 yr). Comparable (quasi-)periodic signals are recurrently related to solar activity fluctuations (Yu and Ito, 1999; Muñoz et al., 2002; Becker, 2005; Cosentino et al., 2005; Garcin et al., 2006; Harzhauser et al., 2008). Although it remains speculative, such cycles could be the reason for the observed short-term variations of bottom-water oxygenation.

7. Palaeoenvironmental evolution

7.1. Stage 1 — before the rise of Lake Pannon (~1.5–0.0 m; duration: unknown)

The basal, sandy–silty layers were deposited in a fluvo-lacustrine freshwater environment as indicated by freshwater ostracods and lithology (Fig. 7a). Reports of exploration wells document hardly traceable gravels and thin coal seams a few metres below, which hint at a highly variable wetland (Gross, 2004a).

7.2. Stage 2 — development of a taxodiacean-swamp (0.0–0.3 m; duration: hundreds of years)

A rising groundwater table, which anticipates a transgression of Lake Pannon, caused the establishment of a *Glyptostrobus*-swamp. This conifer-forest existed a few centuries. Beside water pines, several other trees (e.g., Juglans, Pinus, Carya, Quercus), shrubs (e.g., Myrica, Salix), various freshwater marsh taxa (e.g., Cyperaceae, Poaceae) and aquatic plants (e.g., Trapa, Potamogeton) document a vertically as well as laterally highly structured swamp, which was inhabited by semi-aquatic reptiles and mammals (e.g., Emydidae, Castoridae; Figs. 7a and 8). Based on the palaeoecological composition warm-temperate to almost subtropical conditions are suggested.

7.3. Stage 3 — drowning of the swamp (0.3–0.8 m; duration: hundreds of years)

The swamp forest became extinct within a few decades by a rising water table (>1 m water depth), with sediment changes from pelitic, coaly deposits (0.3–0.8 m) to laminated clays (0.3–0.8 m). Abundant plant remains (e.g., leaf litter, twigs), rare unionid clams, ostracods, insects, cyprinid fishes, amphibians and bird fossils indicate the close-by lakeside and almost freshwater conditions. Mass-occurrences of *Mytilopsis neumayri* occur between ~0.4 and 0.80 m. This dreissenid bivalve probably dwelled byssally attached to the submerged tree trunks and refers to salinities around 2.3–3.5 PSU (Harzhauser and Mandic, 2004). Between 0.3 and 0.75 m the abundance of land-derived plant material decreases (mirrored by declining TOC), while the content of derived plant material decreases (mirrored by declining TOC), while the content of *Mytilopsis neumayri* increases. Because this mussel is supposed to avoid oxygen-depleted environments, temporarily better oxygenation can be assumed (but not necessarily at the lake’s bottom; Figs. 7a and 8). However, around 0.80 m this bivalve abruptly disappears. This hints at a low aerated episode and/or an initial pulse of increased salinity in combination with an accelerated transgression. A first peak of the magnetic susceptibility at 0.80 m is related to greigite formation, which refers to reduced conditions and the influx of saline waters likewise.

7.4. Stage 4 — transgression of Lake Pannon (0.8–7.5 m; duration: thousands of years)

Upsection, the land-derived influx declines considerably (fewer plant fossils, decreasing TOC). The scarce mollusc fauna comprises mainly brackish water lymnocardiid bivalves. Opportunistic brackish water ostracods (e.g., *Cyprididae*, candonids) but also *Loxoconcha* and *Hemicytheria* start to shape the benthic microfaunas. Calcareous nannoplankton is recorded at 1.25 m for the first time (Ćorić and Gross, 2004). Dinoflagellates and brackish water fishes can also be found here (Meller and Hofmann, 2004; Schultz, 2004). These palaeoecological evidences as well as decreasing TOC/TS-ratios signify...
the influx of saline waters (~18 PSU), which is related to a transgression of Lake Pannon (Fig. 8).

Overall, stage 4 is characterised by limited oxygenation of bottom waters. Rare benthic mollusc faunas, partly associated fish skeletons and limited bioturbation emphasise this assumption (Cziczer et al., 2008). Shallow burrowing, probably dysoxic environments tolerating lymnocardiids are dominant, accompanied by rare specimens of Mytilopsis ornithopsis. Commonly, lymnocardiids are found in "butterfly" preservation, which hints at their death at the sediment surface and a redox-front close to the water/sediment interface (Briggs and Crowther, 2001). The rare occurrence of intact ostracod carapaces suggests that these animals avoided the low aerated sediment as well.

The high-resolution interval (1.04–3.31 m) demonstrates that after a first phase of settlement (interval A), Cyprideis and Loxoconcha were the most successful taxa (Fig. 7b). Both experienced a bloom at interval B, where salinity is supposed to be highest. Up to 2.6 m the lake floor was still within the photic zone with only a few metres of water depth (~5–10 m). This can be deduced from shallow water ostracods (Cyprideis, Loxoconcha) as well as from the mollusc record (lymnocardiids, M. ornithopsis; Magyar, 1995; Meisch, 2000; Harzhauser and Mandic, 2004; Cziczer et al., 2008). During intervals B.4–C, salinity slightly dropped (loss of Hemicytheria) and water depth faintly increased (decline of Cyprideis and Loxoconcha, relative ascent of candonids). This deepening is also signalled in the dinoflagellate spectrum (Kern et al., submitted).

Short-term oscillations of bottom-water oxygenation were recorded by the as well as by the ostracod-record throughout the high-resolution interval (Fig. 4). The sparse mollusc fauna obviously lived close to its oxygen limit and was not able to recover as fast as the ostracods in times of better ventilation. Quasi-periodic changes in precipitation (Kern et al., submitted) possibly triggered bottom-
Meromictic system was established, which perturbed benthic life. Laminited; ostracod- as well as bivalve-contents significantly declined. Only occasionally sandy layers document turbiditic events with hypopycnal behaviour at 5.5 and 6 m.

7.5. Stage 5 — delta progradation (7.5–30 m; duration: a few ten thousands of years)

Above 7.5 m the sediments become successively coarser (Fig. 7a). Silt content as well as the frequency of turbiditic sandy intercalations increase. This is clearly mirrored by the χ-log. Elevated TOC-values at sandy layers document the enhanced input of terrestrial plant debris as well. Here, the progradation of a deltaic system terminated the limnic phase.

Molluscs are completely missing in stage 5 and the ostracod content is notably reduced. The large amount of sandy-silty sediment in combination with a decrease in salinity significantly hampered ostracods to colonise this environment (but not below 5 PSU up to 13.6 m = last occurrence of un-noded Cyprideis). Additionally, by 16 m, the lack of calcite (Rantitsch et al., 2004) can be linked to diageneric dissolution, which possibly affected the preservation of ostracod valves. Conditions for greigite formation had already ended by ~7.5 m and largely detrital magnetite accounts to the χ-record in this stage. Highly diverse, but fluvially transported leaf assemblages from layers close to the top (~26.5 m) document the existence of nearby evergreen broad-leaved to mixed mesophytic forests, still within a humid Cfa climate (MAT ~15–19 °C, MAP ~1280–1950 mm; Kovar-Eder and Hably, 2006).

On top of the pelitic sequence (~27–30 m), large scale cross-bedded and wave ripple-bedded, silty–sandy deposits indicate the change to a delta-front environment. Hence, the return to a terrestrial environment became complete and finished the sedimentary cycle.

8. Conclusions

We delineate the evolution of a transgressive–regressive sedimentary sequence, which developed at the margin of long-lived Lake Pannon within a period of less than 45 kyr. The sequence comprises: 1. A badly exposed fluvio-lacustrine phase; 2. The development of a swamp forest, which subsisted a some centuries; 3. The drowning of the swamp due to a transgression within a few decades; 4. The establishment of limnic, increasingly deeper and more saline conditions (up to 18 PSU and ?10–15 m water depth); this phase lasted a few millennia and is marked by short-term (centuries and decades) fluctuations of bottom-water oxygenation; and, 5. A gradual progradation of a deltaic system within a few ten thousands of years.

The bundle of applied methods enabled us to decipher the influence and time scale of different ecological factors:

- On outcrop-scale, the γ- and κ-log as well as the ostracod faunas are useful indicators of changing environments.
- γ-log: limnic, pelitic vs. fluvial/deltaic, psammitic sedimentation;
- κ-log: mainly fluvial/deltaic, detrital magnetite vs. largely limnic–brackish, syndepositional greigite;
- ostracods: freshwater vs. shallow–brackish and sublittoral–brackish.

Organic carbon and sulphur contents are primarily related to land-derived plant material. Only in the limnic phase, elevated sulphur-
values depend on the influx of saline waters due to a transgression of Lake Pannon.

- Cyclocities obtained by spectral analysis of the δ-18 log are possibly a recurrent theme in gastropod morphological evolution. Paleobiology 28, 208–221.

Acknowledgements

The Austrian Science Fund (FWF-project P21748) financed this work. We are grateful to Andrea Kern, Mathias Harzhauser (both Natural History Museum Vienna; FWF-project P21414), Ali Soliman, Dan Danielopol and Klaus Minati (all University of Graz; FWF-project P17738) for discussions and interchange of results. We like to thank two anonymous reviewers for their constructive comments and suggestions.

References

Aguilar, J.P., Berggren, W.A., Aubry, M.-P., Kent, D.V., Clauzon, G., Benammi, M., Mix, A., J. 2004. Mid-Neogene Mediterranean marine–continental correlations: an alternative interpretation. Palaeogeography, Palaeoclimatology, Palaeoecology 252, 626–636.

Albrecht, C., Witte, T. 2008. Ancient Lake Ohrid: biodiversity and evolution. Hydrobiologia 513, 105–140.

Averyanov, L.V., Phan, K.L., Nguyen, T.H., Nguyen, S.K., Nguyen, T.V., Pham, T.D. 2009. Preliminary observation of native Glyptostrobus penilis (Taxodiaceae) stands in Vietnam. Taiwania 54, 191–212.

Babinszki, E., Márton, E., Márton, P., Kiss, L.F. 2007. Widespread occurrence of greigite from the Clay Pit Mataschen (Pannonian, Styria). Joannea Geologie & Paläontologie 88, 161–185.

Berner, R.A., 1984. Sedimentary pyrite formation: an update. Geochimica et Cosmochimica Acta 48, 605–615.

Berner, R.L., Koyar-Eder, J., Lipscomb, D., Rögl, F., Sen, S., Töhen, H. 1988. Systematic, stratigraphical and palaeoenvironmental contexts of first-appearing Hippurion in the Vienna Basin. Austria. Journal of Vertebrale Paleontology 8, 427–452.

Böhme, M., Ig, A., Winklhofer, M. 2008. Late Miocene “washhouse” climate in Europe. Earth and Planetary Science Letters 275, 393–401.

Böhme, M., Winklhofer, M., Ig, A. 2010. Miocene precipitation in Europe: temporal trends and spatial gradients. Palaeogeography, Palaeoclimatology, Palaeoecology. doi:10.1016/j.palaeo.2010.09.028.

Boomer, I. 1998, Sedimentary parry formation: an update. Geochimica et Cosmochimica Acta, 48, 605–615.

Borer, N.R., Kavar-Eder, J., Lipscomb, D., Rögl, F., Sen, S., Töhen, H. 1988. Systematic, stratigraphical and palaeoenvironmental contexts of first-appearing Hippurion in the Vienna Basin. Austria. Journal of Vertebrale Paleontology 8, 427–452.

Böhme, M., Ig, A., Winklhofer, M. 2008. Late Miocene “washhouse” climate in Europe. Earth and Planetary Science Letters 275, 393–401.

Boomer, I. 2004. Late Miocene “washhouse” climate in Europe. Earth and Planetary Science Letters 275, 393–401.

Boomer, I. 2005. The use of ostracods from marginal marine, brackish waters as bioindicators of modern and Quaternary environmental change. Palaeogeography, Palaeoclimatology, Palaeoecology 225, 68–92.

Geary, D.J., Stanley, A.W., Müller, F., Magara, L. 2002. Iterative changes in Lake Pannon Melanosipus reflect a recurrent theme in gastropod morphological evolution. Palaeobiology 28, 208–221.

Gamerick, I., Jahn, A., Vopel, K., Gieße, O. 1996. Hypoxia and sulphide as structuring factors in a macrzoobenthic community on the Baltic Sea shore: colonisation studies and tolerance experiments. Marine Ecology Progress Series 144, 73–80.

Garcín, V., Williamson, D., Tait, M., Vincens, A., Mathé, P.-E., Majiak, A. 2006. Centennial to millennial changes in maar lake deposition during the last 45,000 years in tropical Southern Africa (Lake Malawi, Tanzania). Palaeogeography, Palaeoclimatology, Palaeoecology 230, 334–354.

Gross, M. 2003. Beitrag zur Lithostratigraphie des Oststeirischen Beckens (Neogen/ Pannonium; Österreich) und eines Zwerghamsters aus der Wissenschaten, Schriftenreihe der Erddwissenschaftlichen Kommissionen 16, 11–62.

Gross, M. 2004a. Contribution to the ostracode fauna (Crustacea), palaeocology and stratigraphy of the clay pit Mataschen (Lowert Pannonian, Styrian Basin, Austria). Joannea Geologie & Paläontologie 5, 121–147.

Gross, M. 2004b. Pond turtles (Emydymopus turnauensis (Meyr., 1847); Bataguridae) from the Clay Pit Mataschen (Pannonian, Styria). Joannea Geologie & Paläontologie 5, 149–161.

Harzhauser, M., Mandic, O., 2004. The muddy bottom of Lake Pannon—a challenge for dreissenid settlement (Late Miocene; Bivalvia). Palaeogeography, Palaeoclimatology, Palaeoecology 201, 111–120.

Harzhauser, M., Mandic, O., 2008. Neogene lake systems of Central and South Eastern Europe: faunal diversity, gradients and interrelations. Palaeogeography, Palaeoclimatology, Palaeoecology 260, 417–434.

Harzhauser, M., Piller, W.E., 2007. Benchmark data of a changing sea—palaeogeography, palaeobiogeography and events in the central Paratethys during the Miocene. Palaeogeography, Palaeoclimatology, Palaeoecology 253, 8–31.

Harzhauser, M., Baxxor-Hock, G., Piller, W.E. 2004. An integrated stratigraphy of the Pannonian (Late Miocene) in the Vienna Basin. Austrian Journal of Earth Sciences 95 (96), 6–19.

Harzhauser, M., Latal, C., Piller, W.E., 2007. The stable isotope archive of Lake Pannon as a mirror of Late Miocene climate change. Palaeogeography, Palaeoclimatology, Palaeoecology 249, 335–350.

Harzhauser, M., Kern, A., Soliman, A., Minati, K., Piller, W.E., Danielopol, D.L., Zuschin, M. 2008. Central European Cenozoic. Palaeogeography, Palaeoclimatology, Palaeoecology 270, 102–115.

Holmer, M., Storkholm, P. 2004. Soil reduction and sulphur cycling in lake sedi- ments: a review. Freshwater Biology 46, 431–451.

Hey, R.G., Morris, E.E., 1990. A dendroecological approach to measurement of sedimentation in a forested wetland, Black Swamp, Arkansas. Wetlands 10, 107–124.

Hütting, S.K., Hilgen, F.J., Abdul Aziz, H., Krijgsman, W., 2007. Completing the Neogene time scale between 14.5 and 12.5 Ma. Earth and Planetary Science Letters 253, 340–358.

Hütting, S.K., Dekkers, M.J., Franke, C., Krijgsman, W. 2009. The Tortonian reference section at Monte dei Corvi (Italy): evidence for early remanence acquisition in gyttja-bearing sediments. Geological Journal International 179, 125–143.

Hunter, R.D., Panyushkina, I.P., Leavitt, S.W., Wiedenhoeft, A.C., Zawiskie, J. 2006. A multiproxy environmental investigation of Holocene wood from a submerged cypress forest in Lake Huron, USA. Quaternary Research 66, 67–77.
