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Deltaic systems of the northern Vienna Basin: The lower-middle Miocene conglomerate bodies

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Abstract: The cross-border correlations of the Miocene lithostratigraphic units (Slovakia, Austria) remain poorly constrained, owing to the various sources of clastic material transported into the basin and the low stratigraphic resolution of conglomerates. Therefore, this study is focused on the Lower and Middle Miocene conglomerates in the deltaic systems of the northern Vienna Basin and assesses their implications for cross-border correlations, as well as for the evolution of the Eastern Alpine–Western Carpathians junction area. Revision of lithostratigraphy is based on analyses of the Zohor-1 and Lozorno-1 wells, including the reassessment of published data to account for any refined geochronology and paleogeography. The Lower Miocene Jablonica deltaic system is formed by deposits originating in the alluvial, deltaic, and littoral environments in the basin's northernmost tip. The conglomerates of exclusively Western Carpathian provenance pass into the deep-water marine mudstones of the Ottnangian Lužice and Karpatian Lakšáry fms. The Karpatian conglomerates at the base of the Zohor-1 well can be correlated both with the conglomerates of the Jablonica deltaic system in the north, as well as with the conglomerates of the Gänserndorf Mb. in the south. The overlying heterolithic flood plain of the deltaic sediments from the Závod Fm. most likely form the facies continuation of the alluvial to flood plain deposits of the Schönkirchen Mb. (upper part of the Aderklaa Fm.). The Middle Miocene deposition starts with terrestrial sediments along the slopes of the Malé Karpaty Mts. dated to 15.2 Ma. Later, along with the Devínska Nová Ves fan-delta, the presence of *Orbulina suturalis* developed in the marginal parts of the Vienna Basin during the Early Badenian. Despite having the same structural position, it is younger than the Rothneusiedl Formation in the southern part of the Vienna Basin. The provenance of clastics reflects the source in the Central Western Carpathian units and documents the Miocene uplift of the horst structure of the Malé Karpaty Mts.

Keywords: Miocene, Vienna Basin, sedimentology, biostratigraphy, provenance analyses, sedimentary petrography, cross-border correlation.

Introduction

The position of the Vienna Basin (VB) covering the territory of three states (Fig. 1) has led to multiple interpretations of the development of the entire system with an emphasis on each territory, in some cases with insufficiently substantiated spatio-temporal cross-border correlation of its formations and members (e.g., Buday 1955; Buday & Cicha 1956; Špička & Zapletalová 1964; Jiříček 1985; Wessely 1988, 1992; Mina-říková & Lobitzer 1990; Brix & Schultz 1993; Kováč 2000; Andrejeva-Grigorovič et al. 2001; Vass 2002; Kováč et al. 2004, 2007; Strauss et al. 2006; Kováčová & Hudáčková 2009; Harzhauser et al. 2011a, 2017, 2019, 2020; Beidinger & Decker 2014; Hohenegger et al. 2014; Lee & Wagreich 2017; Rupprecht et al. 2019; Siedl et al. 2020).

The aim of this study is to analyse the clast provenance in conglomerates of deltaic systems in the northern VB (Slovakia) in order to understand the paleogeographic changes between the Eastern Alps and Western Carpathians during the Karpatian–Badenian stage. The conglomerate bodies are informative about altitudinal changes of reliefs and can identify the periods of extensive erosion. Our research also concentrates on the relationship between the deltaic bodies and offshore marine sediments observed in the Slovak part of the VB and their analogues in the Austrian part of the basin. This evaluation is based on the sedimentological and biostratigraphical study of the Zohor-1 and Lozorno-1 well-core material, as well as the summarization and re-evaluation of previously published data, archival reports, and seismic lines. The relationship between the facies development, sedimentary architecture, and classification of sediments into groups, formations, or members was re-evaluated from previously-published data and based on data taken from the studied wells.

The refinement of the position and age of the sedimentary formations, as well as members defined using the micropaleontological biozonation in the past, needs to be approached carefully because the onset and duration of biozones can be diachronous or difficult to trace among distinct environments, especially when based on benthic taxa (e.g., Papp et al. 1978; Harzhauser et al. 2020; Siedl et al. 2020). Moreover, it can shift over time across the basin. Recent radioisotopic dates derived from tuffs (Sant et al. 2017, 2020; Rybár et al. 2019) require reassessment of the lithostratigraphy and unification of cross-border correlations between all parts of the basin. The local three-member division of the Badenian sediments into the Lagenidae, Spirorutilus and Bulimina–Bolivina zones used for the VB could not be extrapolated to other basins. In the Danube Basin, Badenian sediments can be divided only into the lower and upper parts (equivalents of the Langhian and early Serravallian) with a radiometrically $^{40}Ar/^{39}Ar$ dated boundary – 13.8 Ma (Šarinová et al. 2021).

It is possible to distinguish several depositional cycles in the VB sedimentary record conditioned by tectonics and sea-level fluctuations. Correlation of the sequences of boundaries documented on the seismic profiles and wells within the Slovak and Austrian territories resulted in a comparison of attributes of cycles that mutually occur in both states (Fig. 2; Kováč et al. 2004, 2018a; Strauss et al. 2006; Piller et al. 2007; Siedl et al. 2020; Kranner et al. 2021a, b). However, such correlations could be affected by regional differences in accommodation space (e.g., Popov et al. 2004; Piller et al. 2007; Kováč et al. 2017, 2018a; Sant et al. 2017; Hofmayer et al. 2019; Brlek et al. 2020; Ruman et al. 2021).

The Lower Miocene sediments of the VB include several 3rd order sequences, which are often incorrectly correlated with TB 2.1 and TB 2.2 sequences *sensu* Haq et al. (1988). In the VB, the global rise in sea level is reflected only in sediments above the Bur3 boundary in the Ottnangian and above the Bur4 boundary during the Karpatian (*sensu* Hardenbol et al. 1998). While signals of these two sea-level rises are significant and clearly recognisable (e.g., Harzhauser et al. 2019, 2020; Kranner et al. 2021a,b), the remaining fluctuations are distorted by local tectono-sedimentary changes. Such 4th order depositional cycles have been documented in the late Burdigalian – the lower Karpatian (K1) and the upper Karpatian–lowermost Badenian? (K2) cycle (Kováč et al. 2004) are discussed below.

The Middle Miocene sediments of the basin comprise depositional cycles that may be partly correlated with the transgressions above the Bur5/Lan1, Lan2/Ser1, Ser2, Ser3 boundaries (*sensu* Hardenbol et al. 1998) or TB 2.3., TB 2.4., TB 2.5., and TB 2.6. cycles (*sensu* Haq et al. 1988). However, the standard sequence stratigraphy cycles and observed local Middle Miocene depositional cycles are not fully synchronous. This discrepancy is caused by paleogeographic changes and multiple tectonically-controlled openings/closings of the marine connections between the Central Paratethys and the Mediterranean during the Badenian and Sarmatian (e.g., Kováč et al. 2004, 2007, 2008, 2018a; Piller et al. 2007; Hohenegger et al. 2014; Harzhauser et al. 2020; Kranner et al. 2021a,b). Based on the study of wells and seismic lines, three Badenian depositional cycles have been defined: B1, B2 and B3 (e.g., Strauss et al. 2006; Siedl et al. 2020); however, the correlation of these cycles with the traditionally used biostratigraphic zoning is uncertain (see above).

This study also points to the consequences of tectonic activity during the collapse of the wedge-top basin around the Lower/Middle Miocene boundary (~16.5–15.5 Ma), which led to paleogeographic changes that resulted in the immigration of plankton from both the Mediterranean and the Eastern Paratethys/Indopacific domain (Hofmayer et al. 2019; Hudáčková et al. 2020; Ruman et al. 2021).

Geological settings

The VB, which is situated between the Eastern Alps and Western Carpathians, is about 190 km long and 55 km wide, and the sedimentary fill reaches a thickness up to 5500 m (Killényi & Šefara 1989). The NNE–SSW striking basin was filled by deposits whose facies and quantity depended on active tectonics, as well as on the accommodation of space (Fig. 2). The basin development was closely related to the subduction of the Rhenodanubian and External Western Carpathian Flysch Belt basement below the font of the orogenic wedges of the Eastern Alps and Western Carpathians (Plašienka 2018). This process reached a continent–continent collision phase at the end of the Oligocene, resulting in the extrusion of ALCAPA (Eastern Alpine, Western Carpathian, and Transdanubian Range units) lithospheric fragment northeastward (Ratschbacher et al. 1991; Csontos et al. 1992; Kováč et al. 2016, 2017). Several distinct counterclockwise rotations of the movement trajectory of this crustal fragment during the Miocene are documented by structural measurements (e.g., Márton et al. 2016; Kováč et al. 2018b).

The Early Miocene formation of the wedge-top basin situated above the Flysch Belt nappe pile in the north-western part of VB (Fig. 2), which is coupled with the subsidence of its depocenters, led to synchronous evolution of the deep-water environment, as well as uplift and erosion of the orogen internal complexes (e.g., Plašienka & Soták 2015; Hók et al. 2016; Tari et al. 2021). The alluvial, deltaic, and shallow-marine coastal environments rimmed the southern margins of the subsiding basin. The basin development was controlled by constant transpression, and the source areas of clastics were in the Eastern Alpine and Western Carpathian units.

The rapid subsidence of the middle late Burdigalian wedge-top basin was replaced by inversion at the end of this period. The present structure of the Pieniny Klippen Belt was formed, the Flysch Belt nappe-pile started to be uplifted, and exhumation of complexes of the Eastern Alps and Central Western Carpathians accelerated (e.g., Danišík et al. 2004; Plašienka & Soták 2015; Hók et al. 2016; Králiková et al. 2016; Tari et al. 2021).

During the Middle Miocene, the initial rifting of the VB started with the graben-and-horst structure formation. The basin subsidence was controlled by a transtensional to an extensional tectonic regime (Royden 1993; Fodor 1995; Nemčok et al. 1998; Kováč et al. 2004; Strauss et al. 2006; Granado et al. 2016; Lee & Wagreich 2017). The oblique collision of the Western Carpathian orogenic segment with the European

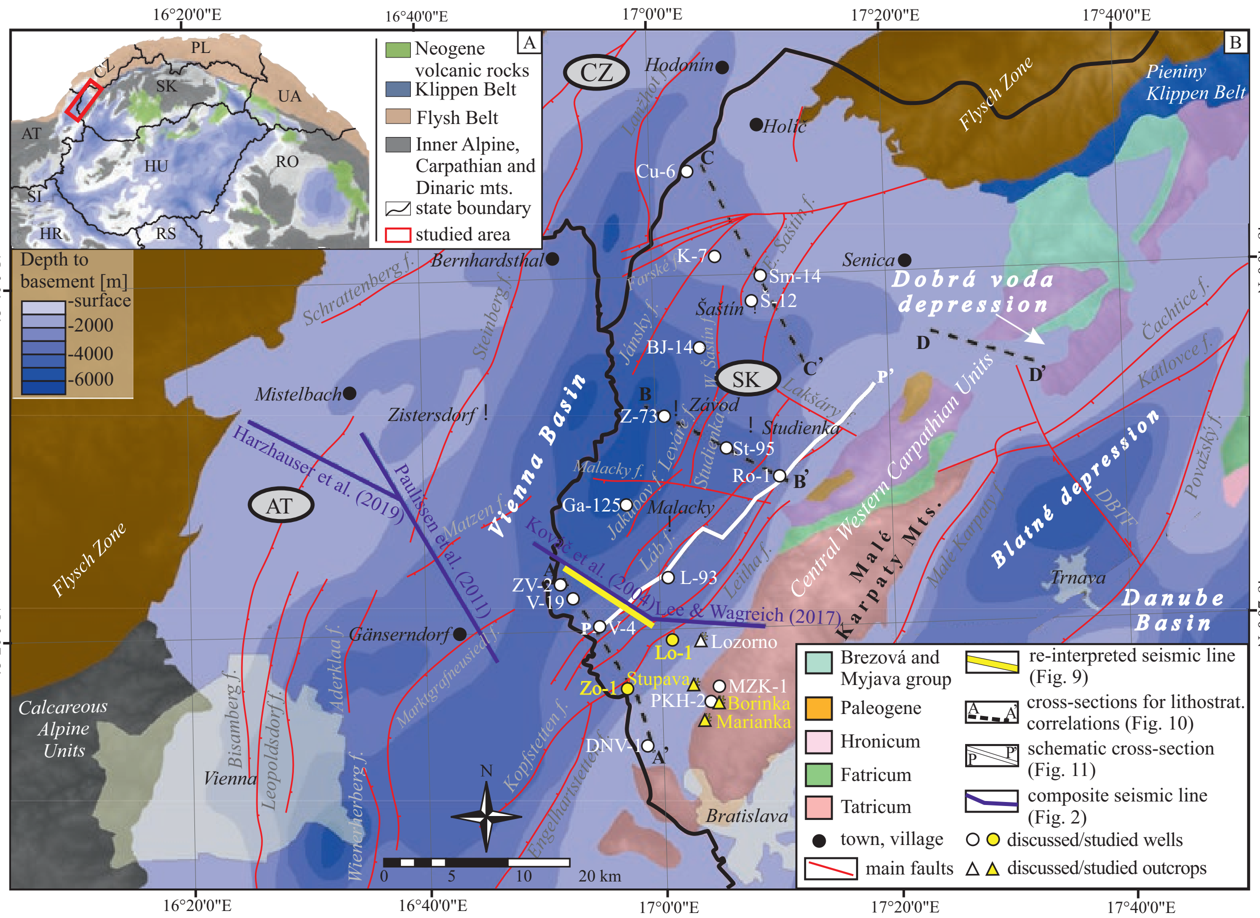


Fig. 1. A — Location of the studied area within the Pannonian Basin System; **B** — Location of the studied well, outcrops and seismic lines within the northern Vienna Basin (background map of pre-Cenozoic basement depth modified according to Fusán et al. 1987; Killényi & Šefara 1989; Horváth et al. 2015).

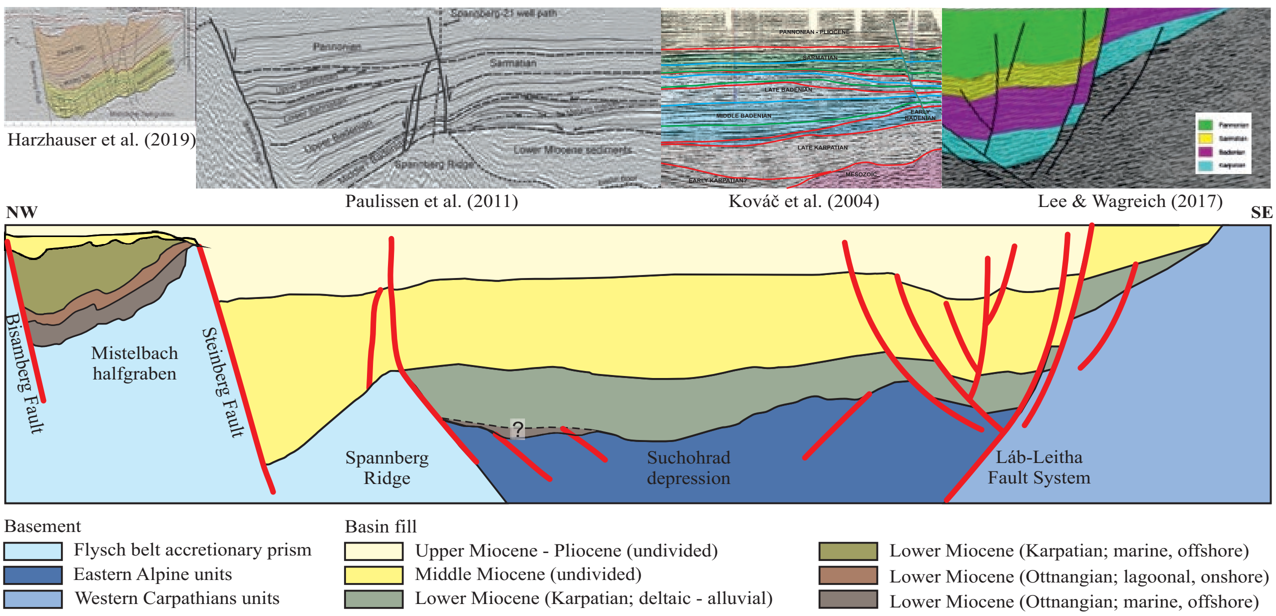


Fig. 2. Interpreted composite seismic line (cross-section) situated in the central part of the Vienna Basin depicting spatial distribution of strata (compiled from Kováč et al. 2004; Paulissen et al. 2011; Lee & Wagreich 2017; Harzhauser et al. 2018). For location, see Fig. 1.

Platform (Bohemian Massif) resulted in the openning of several pull-apart depocenters along the NE margin of the basin (Marko & Jureňa 1999). The following syn-rift stage promoted the subsidence of new depocenters concentrated in the basin's central and southern parts (Lankreijer et al. 1995; Kováč et al. 2004; Lee & Wagreich 2017; Harzhauser et al. 2020; Siedl et al. 2020).

Sedimentary record of the VB thus consists of remnants of the older (Lower Miocene) wedge-top basin, followed by the deposition of the Middle Miocene fill of the basin (Fig. 2), which is interpreted as a thin-skinned pull-apart to the extensional rift basin (e.g., Royden 1985, 1988; Kováč et al. 1993, 1996, 2004; Fodor 1995; Lankreijer et al. 1995; Decker & Peresson 1996; Kováč 2000; Strauss et al. 2006; Allen & Allen 2013; Beidinger & Decker 2014; Granado et al. 2016; Lee & Wagreich 2017). The oldest, latest Eggenburgian–Ottnangian shoreface sediments, which contain pebble and cobble material from the underlying rock complexes, were deposited in the NE part of the basin. Based on lithology, these conglomerates are classified to the Podbranč and Chropov conglomerate mbs. (e.g., Buday & Cicha 1956; Jiříček & Seifert 1990; Vass 2002). The offshore facies represent the marine deep neritic to shallow bathyal mudstones (*schlier*) of the Lužice Fm. (Jiříček & Seifert 1990; Wessely 1992; Vass 2002; Kováč et al. 2004, Harzhauser et al. 2020). In the southern part, the alluvial to lagoonal Bockfliess Fm. was defined (Wessely 1992; Weissenbäck 1995; Harzhauser et al. 2019 and references therein).

The Karpatian offshore facies are represented by the pelitic *schlier* deposits of the Lakšáry Fm. in the north (Špička & Zapletalová 1964; Jiříček & Seifert 1990; Wessely 1992; Vass 2002; Kováč et al. 2004), while in the southern part, braided to meandering river alluvial plain with lakes of the Aderklaa Fm. were deposited (e.g., Weissenbäck 1995; Harzhauser et al. 2020). Mudstones and sandstones of the alluvial to deltaic hyposaline Závod Fm. (with occasional marine ingressions), which proceeded from the SW, gradually pushed the marine Lakšáry Fm. towards the north and replaced it (Špička & Zapletalová 1964). A deltaic body passing basin-ward to the Lakšáry Fm. was described in the northern tip of the VB as the Jablonica conglomerate Mb. (Vass 2002; Kováč et al. 2004; Teťák 2017). In the area of the Dobrá Voda depression, it is parallelized with the upper part of the Ottnangian–lower Karpatian Planinka Fm. (Kováč et al. 1992). In the NW, Karpatian onshore shallow water sediments of the Lakšáry Fm. are represented by the Týnec and Šaštín sand mbs. (Špička & Zapletalová 1964; Vass 2002).

Tectonic re-arrangement of the basin took place around the Lower/Middle Miocene boundary (e.g., Kováč et al. 2004; Harzhauser et al. 2019, 2020). This process was associated with the incision of canyons into the older sedimentary fill (Dellmour & Harzhauser 2012; Siedl et al. 2020), followed by their repeated infilling.

The Middle Miocene marine transgressions permeated into the VB from the Pannonian back-arc basin system in several pulses (e.g., Sant et al. 2017; Ivančič et al. 2018; Brlek et al. 2020; Hudáčková et al. 2020; Ruman et al. 2021). In the south (Austria), beside/above the earliest Badenian (early Langhian) erosional boundary, the alluvial Rothneusiedl Fm. (previously referred to as the Aderklaa conglomerate Fm.) followed by the marine Mannsdorf Fm. began to deposit prior to ~15 Ma (Strauss et al. 2006; Harzhauser et al. 2020). In the NE (Slovakia), the Middle Miocene basin-fill starts with the terrestrial deposits containing the "Kuchyňa" rhyolite tuff dated by $^{40}Ar/^{39}Ar$ sanidine to 15.2 Ma (Rybár et al. 2019). In the NW, the Badenian Lanžhot Fm. basal conglomerates and pelites containing anhydrite (Kúty Mb.) are covered by marine mudstones (Špička 1966; Vass 2002). The basinal mudstones (*tegel*) of the NN5 Zone containing *Orbulina suturalis* are represented by the top of the Lanžhot and Jakubov fms. (Vass 2002; Fordinál et al. 2012; Siedl et al. 2020). Their biostratigraphic inference follows Grills' zonation where the mudstones of the Lanžhot Fm. correspond to the "Upper Lagenidae Zone", and the Jakubov Fm. represents "Spirorutilus carinatus Zone" (Grill 1943). The mudstones (*tegel*) of the Jakubov Fm. spread through the entire northern VB (Špička 1966; Vass 2002; Kováč et al. 2004). On the eastern margin of the VB, the lower/middle Badenian "Zohor" and DNV conglomerates were deposited (Vass et al. 1988; Vass 2002). Marginal Žižkov and Hrušky mbs. were deposited synchronously on the western margin (Vass 2002), while the Auersthal Fm. conglomerate and sand form its temporal equivalent in the south (Harzhauser et al. 2020). The Badenian sedimentary sequence is terminated by marginal clastics of the Sandberg Mb. which laterally pass through the marine mudstones of the Studienka Fm. (Špička 1966; Baráth et al. 1994; Vass 2002).

Methods and materials

Drilling of the Lozorno-1 (48.334649°N, 17.013539°E) well was performed in the years 1956–57 and the Zohor-1 well (48.281975°N, 16.957611°E) in 1995–96 for hydrocarbon prospection purposes. Both of the studied wells were selected based on their penetration into the subjected conglomerate bodies. Core samples taken discontinuously from the depth intervals depicted in Fig. 3 were stored in the repository of Nafta a.s. (Gbely, Slovakia). Stratigraphic terminology, if possible, inclines towards the use of regional stages instead of international ones.

For the purposes of provenance and determination of the transport mechanism, the cores were cut in half and scanned. Sedimentary structures in the individual well-cores were evaluated *sensu* Talling et al. (2012) and Rossi et al. (2017). The well log data (spontaneous potential – SP; resistivity – RT and gamma ray – GR) were evaluated based on Emery & Myers (1996) and Rider & Kennedy (2011). The degree of sorting was made using visual scale *sensu* Folk (1968) and Jerram (2001), and roundness description based on Pettijohn (1975). The lithology of clasts in conglomerate bodies was confirmed by thin sections under a polarizing microscope. A 2D seismic section was taken from a 3D reflection seismic

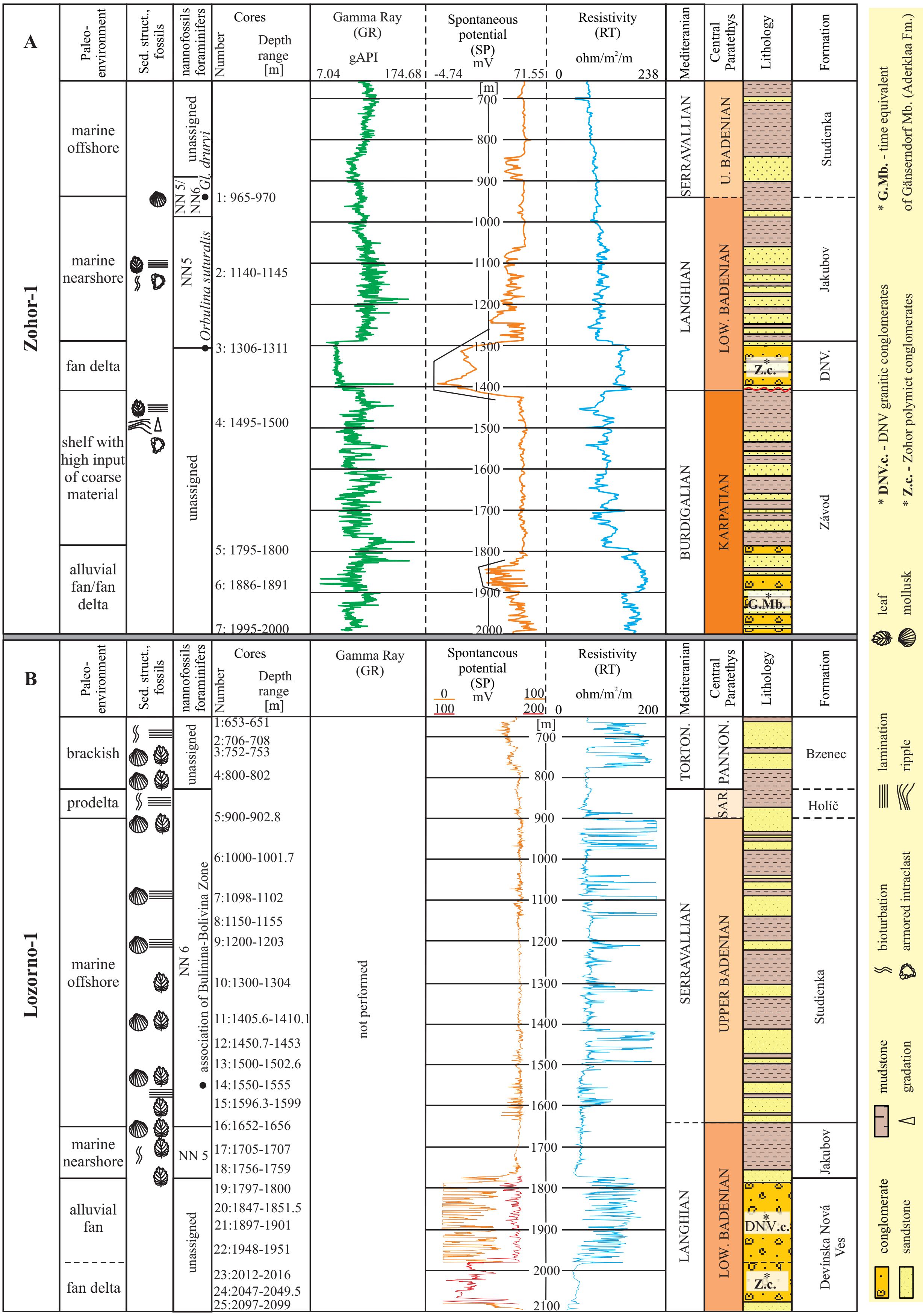


Fig. 3. Lithostratigraphy and well-logs of the Zohor-1 well (**A**) and Lozorno-1 well (**B**) with interpreted paleoenvironment.

survey obtained by the company, Nafta a.s. For the purpose of stratigraphic interpretation, the surrounding wells were correlated to the seismic survey by velocity information from checkshot surveys and velocity functions (the depth of seismic section is in $m.s^{-1}$ of TWT (Two Way Time)). The seismic data were re-interpreted based on Mitchum et al. (1977), Sangree & Widmier (1979) and Vail (1987).

Three outcrops situated on the SW slopes of the Malé Karpaty Mts. (Borinka – 48.2666°N, 17.0853°E; Marianka – 48.2491°N, 17.0650°E; Stupava – 48.2815°N, 17.0486°E; Fig. 1) that supplemented this analysis were cleaned and photo-documented. The main sedimentary structures and textures were described in the same manner as the well samples.

Calcareous nannofossil samples were prepared by the common method of decantation using smear slides. Smear slides were prepared following the time-efficient method of sample preparation for quantitative analysis (Bown 1998). A small portion of sediment was mixed with distilled water on a glass slide to create a thick suspension. The suspension was smeared thinly across the coverslip with a toothpick and dried on a hotplate. Later, the coverslip was affixed using Norland 61 optical adhesive and glued under a UV lamp. Samples were observed using an Olympus BX50 microscope at 100× magnification, while oil immersion was used for microscopic evaluation. An Olympus Infinity 2 camera with QuickPHOTO CAMERA 2.3 software was used for the photographic record. Systematic identification of calcareous nannofossils was based on the taxonomy of Young (1998). Standard NN zonations (*sensu* Martini 1971) and regional zones (*sensu* Andrejeva-Grigorovič et al. 2001) were used for age determination. Calcareous nannofossils were counted at 300 fields of view. Foraminifera was obtained from 100g of sediment diluted by hydrogen peroxide and wet sieved (0.071 mm and 1 mm). A binocular stereoscopic microscope (Olympus SZ75) and a biological polarizing microscope were used for determination. The scanning electron microscope QUANTA FEG 250 was used for their imaging (Institute of Electrical Engineering, SAS). Residua were split into approximately 300 specimens (if possible). Determination of foraminifers was based on Loeblich & Tappan (1992), Cicha et al. (1998), Łuczkowska (1974) and Holbourn et al. (2013). Paleoecological parameters based on the presence and dominance of taxa exhibiting special environmental significance were evaluated for samples containing benthic foraminifers. For better interpretations of distributional patterns, species with similar environmental significance were grouped. Taphonomic analyses of associations were performed according to the methods described by Holcová (1997, 1999). Biostratigraphy was mainly inferred from Cicha et al. (1998). *Orbulina suturalis* appears in the World Ocean at 15.1 Ma and in the Mediterranean at 14.6 Ma (Abdul Aziz et al. 2008; Iaccarino et al. 2011; Wade et al. 2011). For the purpose of this study, it was set to ~14.9 Ma with respect to the regional works in the Central Paratethys, because here, the onset of *O. suturalis* coincides with the NN5 Zone starting at 14.9 Ma (e.g., Trakovice-1 well in Rybár et al. 2016; Rögl et al. 2008; Hohenegger et al., 2014; Bukowski et al. 2018; Kováč et al. 2018a; Ilieș et al. 2020).

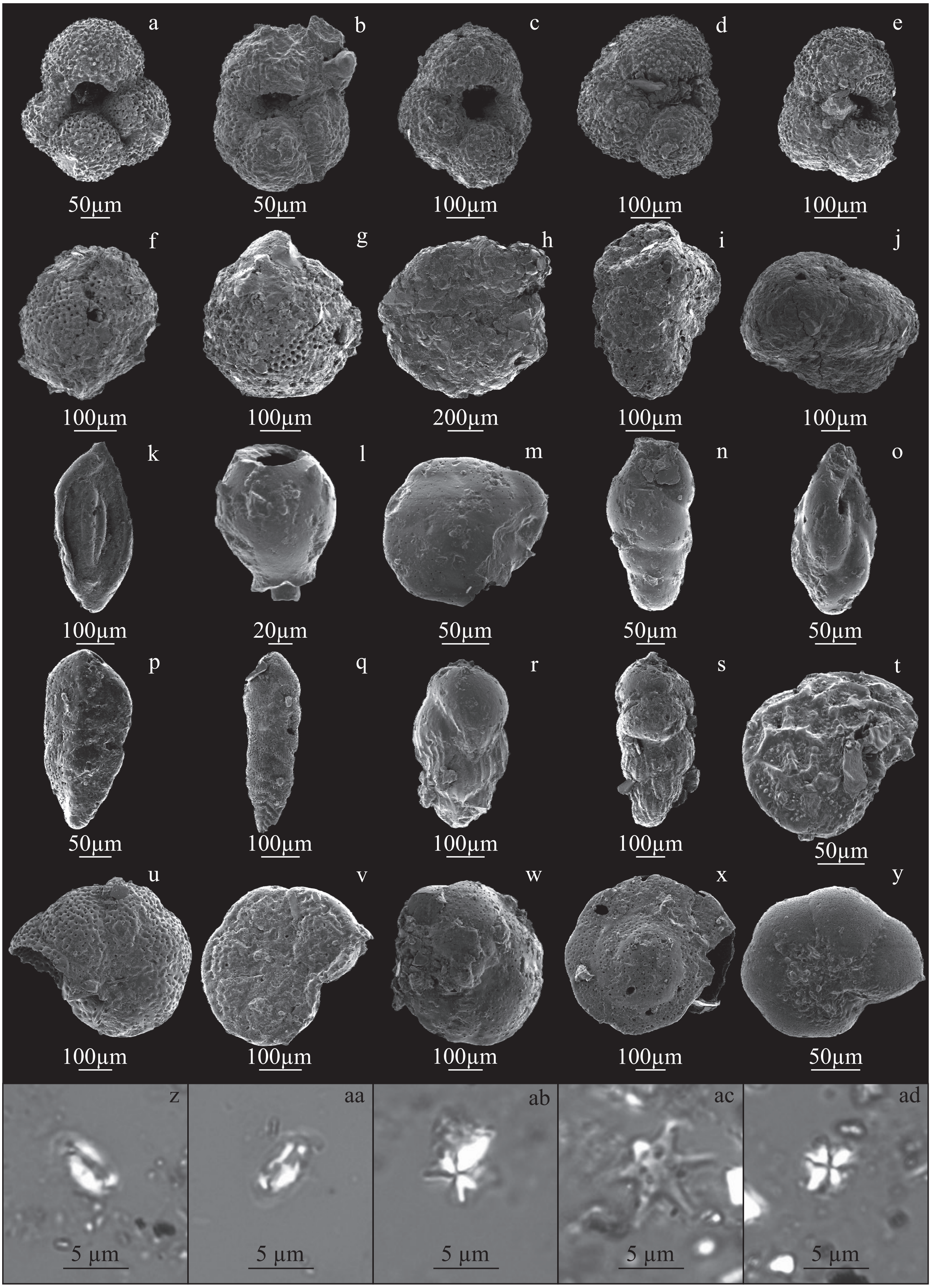
Results

Biostratigraphy

Zohor-1 well

Samples from the lower intervals (depth range 2000–1495 m; Fig. 3) are poor in recovery and lack index microfossils. These samples contain nannofossils reworked from the Cretaceous, Paleogene, and Lower Miocene deposits. Small, fragmented, planktic foraminifera together with *Plectofrondicularia* sp. are also observed at 1495–1500 m. The damaged index foraminifera *Orbulina suturalis* together with *Preorbulina glomerosa* appears at 1311–1306 m. The lower part of the interval at 1140–1145 m is poorly preserved and without index species. The upper part of this core contains the index nannoplankton species *Sphenolithus heteromorphus*, *Helicosphaera walbersdorfensis*, *H. waltrans*, and *Reticulofenestra pseudoumbilicus* (Fig. 4). Other common species are *Calcidiscus tropicus*, *Coronocyclus nitescens*, *Umbilicosphaera rotula*, *U. jafari*, and *Helicosphaera scissura* (Supplementary Table S1).

Fig. 4. Foraminifers and calcareous nannofossils of the Zohor-1 well: **a** – *Globoturborotalita druryi* (Akers, 1955), core 1, box 2 (966–967m); **b** – *Globoturborotalita druryi* (Akers, 1955), core 1, box 2 (966–967m); **c** – *Globoturborotalita druryi* (Akers, 1955), core 1, box 2 (966–967m); **d** – *Trilobatus* cf. *quadrilobatus* (d'Orbigny, 1846), core 1, box 3 (967–968m); **e** – *Trilobatus trilobus* (Reuss, 1850), core 1, box 3 (967–968m); **f** – *Praeorbulina* cf. *circularis* (Blow, 1956), core 2, box 3 (1142–1143m); **g** – *Orbulina suturalis* Brönnimann, 1951, core 2, box 3 (1142–1143m); **h** – ?*Haplophragmoides fragilis* Hoeglund, 1946, core 1, box 3 (967–968m); **i** – *Semivulvulina* sp., core 1, box 2 (966–967m); **j** – *Textularia gramen* d'Orbigny, 1846, core 1, box 3 (967–968m); **k** – *Spirosigmoilina tenuis* (Cžjžek, 1848), core 1, box 1 (965–966m); **l** – *Nodosaria* sp., core 1, box 1 (965–966m); **m** – *Cassidulina laevigata* d'Orbigny, 1826, core 1, box 1 (965–966m); **n** – *Bulimina elongata* d'Orbigny, 1846, core 1, box 1 (965–966m); **o** – *Globobulimina pyrula* (d'Orbigny, 1846), core 1, box 1 (965–966m); **p** – *Bolivina maxima* Cicha & Zapletalová, 1963, core 1, box 1 (965–966m); **q** – *Bolivina pokornyi* Cicha & Zapletalová, 1963, core 1, box 1 (965–966m); **r** – *Uvigerina semiornata* d'Orbigny 1846, core 1, box 1 (965–966m); **s** – *Uvigerina semiornata* d'Orbigny, 1846, core 1, box 1 (965–966m); **t** – *Elphidium* sp., core 1, box 1 (965–966m); **u** – *Lobatula lobatula* (Walker & Jacob, 1798), core 1, box 1 (965–966m); **v** – *Lobatula lobatula* (Walker & Jacob, 1798), core 1, box 1 (965–966m); **w** – *Heterolepa dutemplei* (d'Orbigny, 1846), core 1, box 1 (965–966m); **x** – *Heterolepa dutemplei* (d'Orbigny, 1846), core 1, box 1 (965–966m); **y** – *Porosononion granosum* (d'Orbigny, 1846), core 1, box 1 (965–966m); **z** – *Helicosphaera walbersdorfensis* Müller, 1974, core 1, box 2 (966–967m); **aa** – *Helicosphaera waltrans* Theodoris 1984, core 1, box 3 (967–968m); **ab** – *Sphenolithus heteromorphus* Deflandre 1953, core 1, box 2 (966–967m); **ac** – *Discoaster exilis* Martini and Bramlette, 1963, core 1, box 1 (965–966m); **ad** – *Sphenolithus abies* Deflandre in Deflandre and Fert, 1954, core 1, box 2 (966–967m).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5 µm

In foraminiferal associations, *Orbulina suturalis* co-occurs with *Dentoglobigerina altispira, Lenticulina* sp., *Valvulineria complanata,* and globigerinoideses.

The nannoplankton association from the uppermost cored interval (965–970 m) include *Discoaster exilis*, *Coronocyclus nitescens* (elliptical), *Sphenolithus abies*, *Helicosphaera carteri*, *H. walbersdorfensis*, *R. pseudoumbilicus*, *U. rotula*, *U. jafari*, *Discoaster* spp., *Calcidiscus premacintyrei*, and *Orthorhabdus serratus*. Foraminiferal association consists of *Globigerina bulloides*, *G. regularis*, *Globorotalia (Turborotalia) transsylvanica*, *Globoturborotalita druryi*, *Trilobatus quadrilobatus, Uvigerina* sp*., Cassidulina* sp., *Bulimina* spp., *Valvulineria* sp*., Textularia gramen*, *Semivulvulina pectinata*, ?*Haplophragmoides fragilis*, and *Bathysiphon* sp. (Fig. 4).

Lozorno-1 well

The cored intervals at 2099–1797 m were poor in recovery and lacked marker species (Supplementary Table S2). Calcareous nannofossil association from the interval 1759–1756 m includes, in addition to the marker species *S. heteromorphus* (scarce) and *H. waltrans*, *C. premacintyrei*, *Coccolithus pelagicus*, small reticulofenestrids (*R. minuta*, *R. haqii*), abundant helicospherids (*H. waltrans*, *H. carteri*, *H. mediterranea and H. scissura*), *R. pseudoumbilicus*, and reworked Cretaceous and Eocene nannofossils. Foraminiferal associations are dominated by agglutinated species, especially *Haplophragmoides vasiceki.* Planktic and calcareous benthic species were also present, e.g., *Trilobatus trilobus*, *Cassidulina carinata* and *Pullenia quinqueloba*. The nannofossil association from 1656–1550 m contains ACME *C. premacintyrei, C. leptoporus*, *C. tropicus, C. pelagicus*, *C. nitescens* (+elliptical form), *Discoaster deflandrei*, *H. carteri*, *H. mediterranea, H. walbersdorfensis*, *H. wallichii*, *Rhabdoshaera pannonica*, *R. haqii*, *R. minuta*, *R. pseudoumbilicus*, *S. abies,* S. *moriformis* and *Thoracosphaera* spp. Foraminiferal association of the Bulimina–Bolivina Zone (e.g., *Bolivina viennensis, Cassidulina laevigata*) is observed from the depth of 1552 m.

At 1453–900 m, marker species of the NN6 Zone, such as *H. wallichii* (1098.4 m), *R. pseudoumbilicus* >7 mμ (1098.4 m; 1000–1001.7 m), *Orthorhabdus rugosus* (1200–1203 m and 901.5 m), and *Orthorhabdus rioi* (901.5 m) are present. The uppermost core from this interval (902.8–900 m) contains nannofossil association with *C. leptoporus*, *C. tropicus, C. pelagicus*, *C. nitescens*, *Discoaster variabilis, H. carteri, H. walbersdorfensis*, *Orthorhabdus rioi*, *R. haqii, R. minuta*, *R. pseudoumbilicus*, *Rh. pannonica, S. abies, U. rotula, Orthorhabdus rugosus*, and Paleogene and Cretaceous reworking. In this interval (1453–900 m), foraminiferal associations, in addition to others, contain *Ammonia* ex gr. *beccarii*, *Elphidium* sp., and *Globigerinella obesa* (1303 m); *Elphidium* sp. and *Triloculina* sp. (1153 m); *Bolivina dilatata, Bolivina* sp. and *Ammonia* cf. *tepida* (1102–1098 m); *Porosononion granosum* and *Elphidium* cf. *microelegans* (1001 m); *Ammonia* ex gr. *beccarii*, *Porosononion granosum* and *Elphidium macellum* (900–902.5 m). Fragments of the ostracod *Hemicytheria omphalodes* are also present. The microfossil and nannofossil associations from the cored interval 802–650 m are poor in specimens and lack index fossils.

Petrography and sedimentology

Zohor-1 well

The depth interval between 2000–1795 m consists of conglomerates to sandstones (Fig. 5). The coarse-grained character is also confirmed by negative excursions on the SP log and the high resistivity recorded by the RT log (Fig. 3A). This interval consists of three cores. The lowermost core (2000–1995 m) predominantly contains poorly to moderately sorted, clast supported conglomerates (orthoconglomerates *sensu* Pettijohn 1975), with clasts over 10 cm which repeatedly pass to the matrix supported (paraconglomerates *sensu* Pettijohn 1975), sandy conglomerates or conglomeratic sandstones (gradation intervals). The groundmass is formed by a sandy matrix and carbonate cement. Chalcedone cement is observed in thin sections in the conglomerate at 1999–1998 m (Fig. 6B). Pebbles and cobbles are sub-rounded to rounded, imbrications of clasts are observed locally (Fig. 5). These transitions, from poorly to moderately sorted orthoconglomerates with a sandy matrix and carbonate cement to conglomeratic sandstones, are also observed in the overlying core (1891–1886 m). Carbonate and metamorphic (schist) granules and pebbles are well-rounded, as opposed to sub-angular to sub-rounded granitoid pebbles to cobbles (Fig. 5). In the subsequent core (1800–1795 m), multiple fining-upward layers are observed. Besides the poorly to moderately sorted conglomerates with sub-angular to sub-rounded pebbles and carbonate cement, the core contains several intervals grading from fine-grained conglomerates to sandstones and siltstones, often with an erosional base. In the siltstones, lamination is also observed. The composition of all conglomerate clasts from this interval (2000–1795 m) consists of the same lithologic types, while the differences are in their ratio (Table 1). The clasts are mainly of carbonate rocks (micrite, sparite, recrystallized limestone, biomicrite, biosparite, peloosparite and dolomites), granitoid rocks, biotite paragneisses with garnet and staurolite, chlorite schists, quartz arenites, shales, metasandstones (quartzites), while clasts of basalts and cherts are rare (Fig. 6). The mineral composition of granitoides involves quartz (Qz), plagioclase (Pl, sericitized), potassium-feldspar (Kfs), often chloritized biotite (Bt), and rare muscovite (Ms; Fig. 6). Basalts with intersertal structure consists of plagioclase crystals and chloritized groundmass. In general, the carbonate clasts dominate in the basal part of the interval (2000–1795 m), while the granitoid clasts become dominant in its upper part (Table 1). The sandy matrix in conglomerates is composed of monocrystalline quartz, feldspar, Ms, Bt, clasts of carbonates and metamorphic rocks (medium sand fraction). Upwards, the carbonate admixture content within the matrix increases. Locally, crystals of carbonate cement are present.

Fig. 5. Longitudinal cross-cuttings of the studied cores from the Zohor-1 well and Lozorno-1 well.

The transition to finer-grained sediments is observed between 1795–1410 m as documented by the serrated trend (shale baseline) on the SP log, while the GR and RT logs display high amplitude excursions (Fig. 3A). The core from this interval (1495–1500 m; Fig. 5) is formed by alternation of medium- to coarse-grained sandstones and heterolithic sediments composed of mudstone and fine-grained sandstone with lenticular bedding (asymmetric ripples), and contain carbonized plant detritus as well. The normal gradation and soft-sediment deformations are also observed (convolute bedding, water escape, flame structures). Sandstones contain quartz (mono- and polycrystalline), K-feldspar, plagioclase, muscovite,

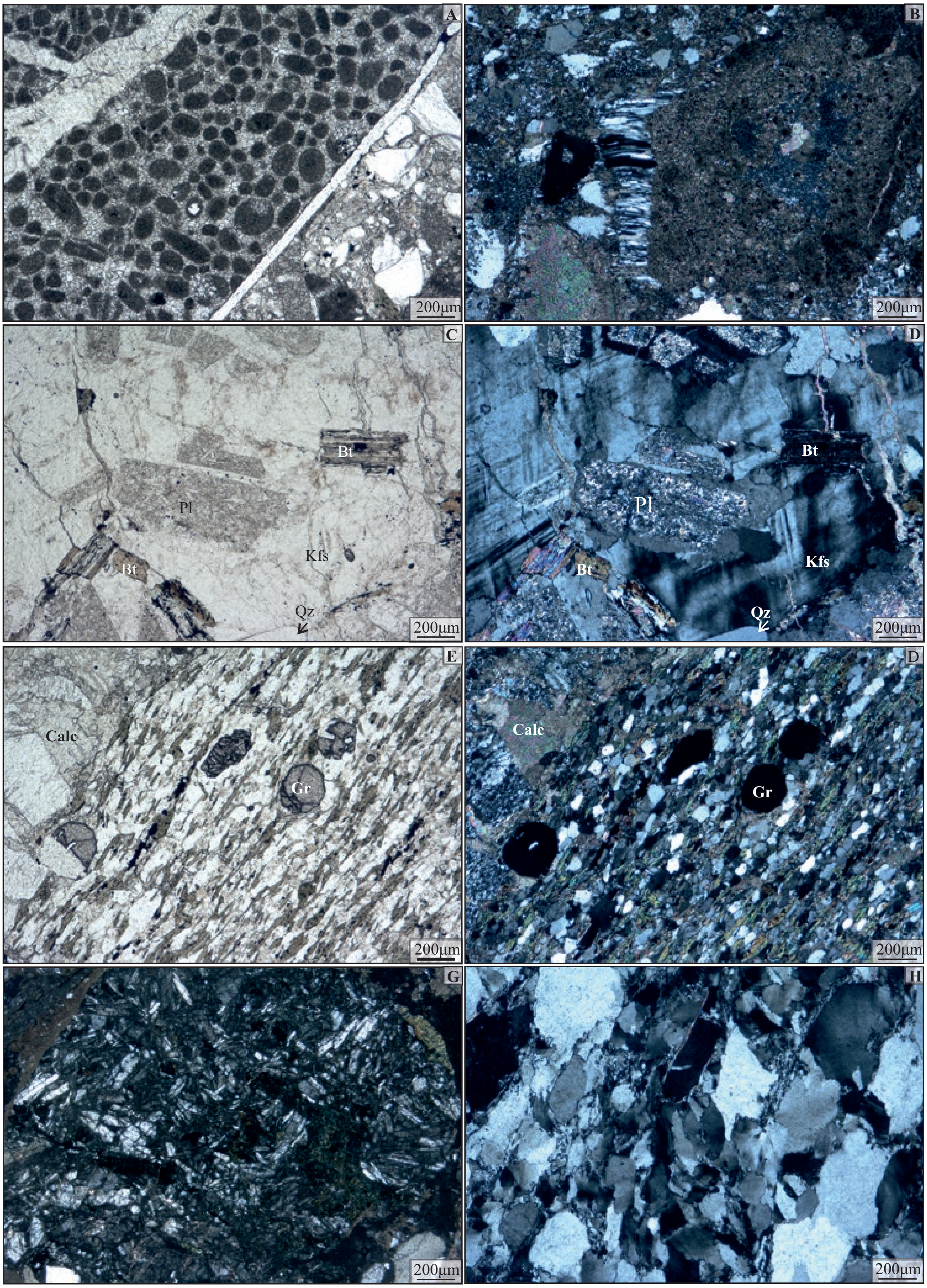


Fig. 6. Thin sections of conglomerates from the Zohor-1 well (PPL=plane polarised light, XPL=crossed polars). **A** — Oosparite clast (core 7, box 4, 1306–1311m, PPL); **B** — Clast of micrite with chalcedone cement rim (core 7, box 4, 1306–1311m, XPL); **C, D** — Clast of granitoide rock (core 7, 1995–2000m, C – PPL, D – XPL); **E, F** — Clast of biotite-paragneiss with garnet (core 7, 1995–2000m, E – PPL, F – XPL); **G** — Paleobasalt clast (core 6, 1886–1891m, XPL), **H** — Quartz arenite clast (core 6, 1886–1891m, XPL).

Table 1: Clast composition of studied conglomerates from the Zohor – Plavecký Mikuláš graben (number of calculated clasts for each sample was limited by core samples dimensions.) Legend: ***Z.c.** – Zohor polymict conglomerates of Devínska Nová Ves Fm. (DNV Fm.); ***DNV.c.** – DNV granitic conglomerates of Devínska Nová Ves Fm. (DNV Fm.); ***G.Mb.** – time equivallent of Gänserndorf Mb. (Aderklaa Fm.)

Core (depth)	Carbonate	Granitoide	Metamorph.	Qz-arenites	Paleobasalts	Number of clasts	Formation
Zohor-1							
7 (1995–2000m)	75 %	9 %	14 %	2 %		127	*G.Mb.
6 (1886–1891m)	65 %	16 %	9 %	7 %	3 %	300	*G.Mb.
5 (1795–1800m)	33 %	43 %	13 %	11 %		341	*G.Mb.
3 (1306–1311m)	80 %	6 %	9 %	5 %		152	*Z.c.
Lozorno-1							
23 (2016–2012m)	46 %	24 %	30 %	+		201	*Z.c.
22(1951–1948 m)		83 %	17 %			55	*DNV.c.
Borinka		84 %	3 %	13 %		100	*DNV.c.
Marianka	2 %	87 %	5 %	6 %		100	*DNV.c.

biotite, fragments of paragneiss, phyllite, granitoid rocks and authigenic glauconite.

The interval 1410–1295 m is formed by coarse-grained sediments, which is confirmed by high amplitude negative excursions on the SP log, low values on the GR log, and high resistivity on the RT log (Fig. 3A). Core from this interval (1311–1306 m) consists of coarse- to medium-grained polymict orthoconglomerates with indistinct normal gradation. Clasts are sub-angular to sub-rounded and rarely well-rounded (clasts up to 5 cm). Compared to the lower one (from the depth of 2000–1795 m), this conglomerate is generally better sorted (Fig. 5); even though it consists of the same lithotype clasts, it has a higher content of carbonates (Table 1). A diagenetic lithostatic compression made the matrix hard to determine. Clasts of siliciclastic rocks are carbonatized.

The depth interval 1295–1060 m is characterized by several bell, funnel, and symmetrical trends with highly negative SP excursion. The GR and RT logs predominantly display pronounced positive excursions (Fig. 3A). Several repetitions of the graded layer, with the fine-grained conglomerate to coarse-grained sandstone at its base, to laminated fine-grained sandstone at its top occur between 1145–1140 m (Fig. 5). Moderately-sorted arkosic sandstones contain carbonized plant detritus and a large amount of mica. The SP and RT logs in the interval 1060–900 m display a low-amplitude serrate trend (shale baseline; Fig. 3A). The GR is characterized by medium amplitude positive excursions. The core at 970–965 m is represented by massive siltstones and mudstones, which also contain carbonized plant detritus. Siltstones to mudstones from the last two cores are bioturbated.

Lozorno-1 well

The basal cored interval (2099–2097 m) is formed by gray, calcareous, bioturbated, and laminated sandstones to siltstones. They pass into the conglomeratic interval (2080–1980 m) determined by negative SP excursions and high resistivity values on the RT log (Fig. 3B). The core at 2049.5–2047 m consists of an alternation of orthoconglomerates and sandstones with a fining upward trend. Poorly to moderately sorted conglomerates with a sandy matrix and carbonate cement without observable gradation or imbrication is observed at 2012–2016 m (Fig. 5). Clasts are sub-rounded to rounded and dominantly consist of a wide range of carbonates (e.g., Calpionella limestone with echinoderm fragments and calcite veins, micrite, pelitic limestone with ooids, and milliolid foraminifers), granitoids (composed of quartz, muscovite, and plagioclase), Bt-paragneisses with garnet, fylites, metasandstones, a small number of Qz-arenites, and cherts (Table 1; Fig. 7). The interval 1980–1770 m is defined by a significant shift to a higher excursion in SP and RT (Fig. 3B). The core at 1948–1951 m from this interval consists of thin sandstone layers (with faint lamination) and of poorly-sorted paraconglomerate layers with sub-rounded clasts of granitoids, including less frequent metamorphic rocks (Table 1). In the conglomerate from the core at 1947–1851.5 m, granitoids dominate. The Bt-paragneisses consisting of sericitized plagioclase and strongly chloritized biotite, as well as a small amount of quartz arenite clasts are also present. The sandy matrix is also chloritized. The depth interval between 1770–1610 m shows a serrated funnel shape on the well logs (coarsening upward trend). Cores (c. 18–15) from this interval mainly consist of claystones and sub-horizontally laminated, calcareous siltstones with intercalations of fine-grained sandstones. The interval 1610–900 m is characterized by several intercalations of siltstones to fine-grained calcareous sandstones containing abundant mollusks with faintly laminated mudstones. Carbonized plants and fossil fragments are also present. Upwards (900–650 m), indistinctly laminated mudstones with carbonized plant fragments and bioturbations are observed. Fossil association is dominated by ostracods and fish fragments.

Outcrop data

All described sites supplement analysis of the Devínska Nová Ves granitic conglomerates. In general, these granitic conglomerates crop out in the western slopes of the Malé Karpaty Mts. between Bratislava and the village of Kuchyňa. Basal boundaries of conglomerates are not exposed, and tops of the sections were formed by recent soil.

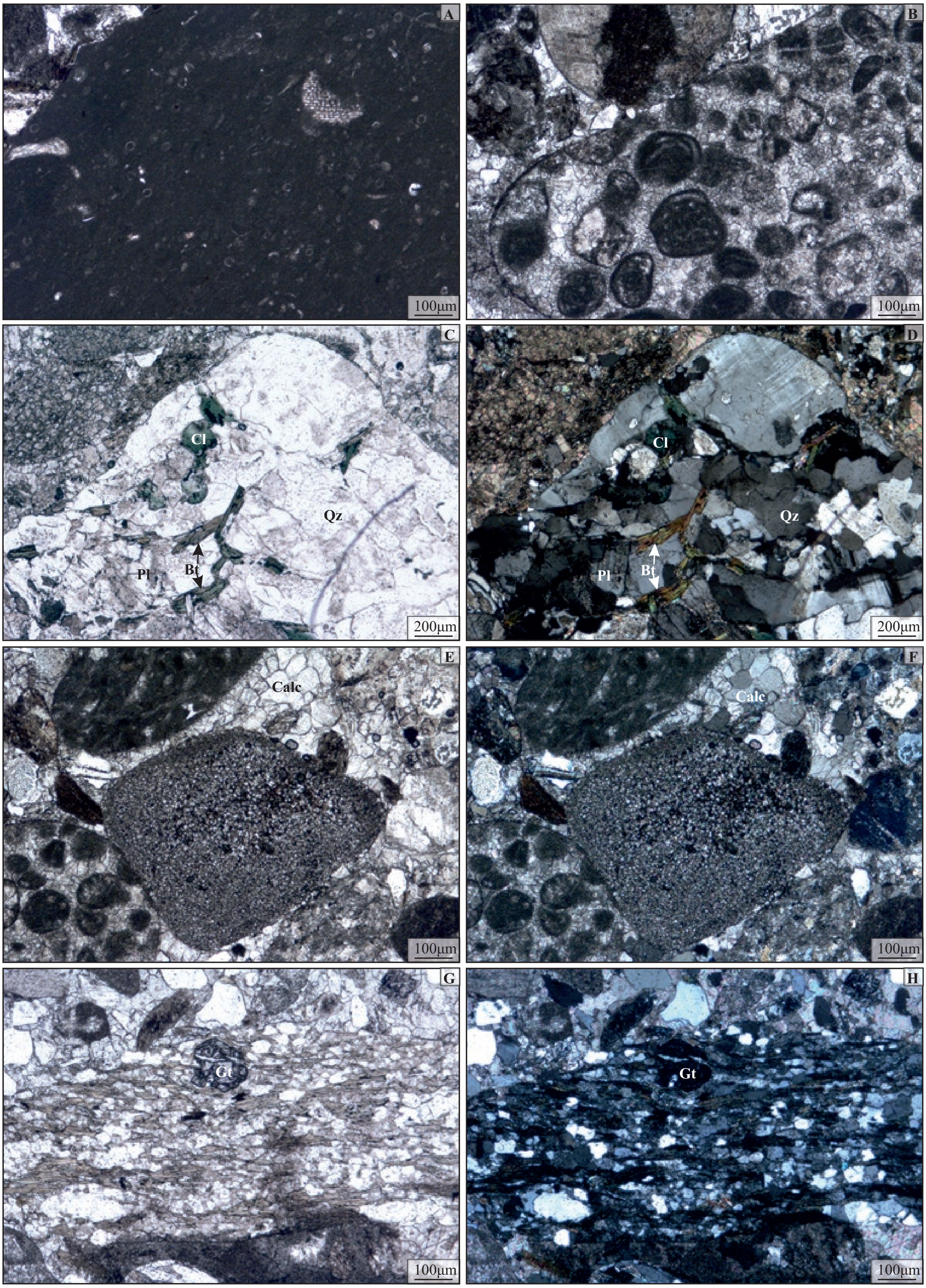


Fig. 7. Thin sections of conglomerate from the Lozorno-1 well (core 23; 2012–2016m; PPL=plane polarised light, XPL=crossed polars). **A** — Clast of the Calpionella biomicrite (PPL); **B** — Pelbiosparite clast (PPL); **C, D** — **C**last of biotite paragneiss with garnet (C – PPL, D – XPL); **E, F** – Detail of carbonate cement between various carbonate clasts (E – PPL, F – XPL).

The Borinka outcrop originally described by Vass et al. (1988; Fig. 8A–D) is an approximately 70 m long section containing several conglomerate layers with a sandy matrix. The exposed part of the section starts with 1 m-thick, clast-supported conglomerates with indistinct gradation (Gg). Clasts are represented mostly by moderately-sorted granitoid cobbles and boulders (the biggest clast is 60 cm in diameter), accompanied by well-rounded, strongly weathered metamorphic rocks and Qz-arenites (Table 1). The profile continues with 5 m-thick massive, monomict granitoid conglomerates (Gm; Fig. 8B). Predominantly angular-shaped to sub-rounded clasts are poorly sorted. The overlying layer is formed by a 30–50 cm-thick layer of heterolithic sediment with rhythmic planar lamination of yellowish fine-grained sands and clays (Sh; Fig. 8C). The section terminates with matrix-supported conglomerates with faint gradation (Gmg, Fig. 8D). This layer is located just below the recent soil; the deluvial origin of this layer could not be excluded.

The Marianka outcrop originally described by Baráth et al. (2015; Fig. 8E, F) is situated in the central part of the village. The exposed part of the section starts with a 3 m-thick layer of indistinctly graded, poorly-sorted, polymict conglomerate with a sandy matrix (Gmg) and rounded boulders to cobbles. Planar bedded gravels (thickness of layer varies between 0.3–1 m) with pebble-sized sub-angular clasts (Gh) follow. The overlying, approximately 1 m-thick gravel layer with planar cross-stratification contains sub-angular shape granules and pebble-sized clasts (Gp). Granitoid clasts dominate (Table 1), although pebble to cobble-sized clasts of sandy limestones are also present. The granule-sized fraction is supplemented by metamorphic rocks and Qz-arenites. The top of the outcrop terminates with 0.5–1 m mudstone and recent soil.

The Stupava outcrop (Fig. 8G) is located on a private property in the center of the town of Stupava. The section is 1.5 m-thick and is formed by a matrix supported by massive, monomict conglomerates that are composed of well-rounded granular to boulder-size granitoids and a sandy matrix (Gmm).

Seismic line

The observed change in seismic facies on the bottom of the seismic line (Fig. 9) indicates a transition between the pre-Neogene basement (chaotic, discontinuous, low-amplitude reflexes) and the Neogene basin fill (semi-parallel, semi-continuous, high-amplitude, and a high to low frequency of reflexes). This shift demonstrates that the conglomerates located at the base of the Zohor-1 well (2000–1795 m) overlie the pre-Neogene basement. Rare imbrications, transitions between grain-supported and matrix-supported conglomerates, as well as poor to medium sorting indicate transport by gravity flows (Nemec & Steel 1984; Nichols 2009). Microfossil associations that mainly contain redeposited specimens are present in the upper part of this interval. Well log data (SP, RT, GR; Fig. 3), in combination with the reflexes on the seismic line (Fig. 9), document the transition to the overlying strata without any major unconformity. Due to the sorting and roundness of clasts, these conglomerates can be interpreted as an alluvial fan body passing to an aquatic environment (Fig. 3) represented by a 400 m-thick interval formed by the alternation of sandstones and heteroliths with lenticular bedding (lithology extrapolated based on the SP and RT log in the entire interval 1795–1410 m; Figs. 3A, 5). On the seismic section, this interval is represented by oblique, semicontinuous, low amplitude, and low to medium frequency seismic reflexes with truncation at the top (Fig. 9). Graded sandstone intercalations may indicate the presence of the gravity flow mechanism (turbidity current, grain flow); however, they can be most likely interpreted as event beds deposited by fluvial traction transport as well, which commonly occur in fluvial environments (Rossi et al. 2017).

The second conglomeratic body in the Zohor-1 well (1400–1300 m) shows the high amplitude reflexes arranged in a prograding clinoform downlapping onto the underlying interval, indicating that the conglomerates were deposited in a deltaic environment as seen by the gravelly topset to sloping foreset beds.

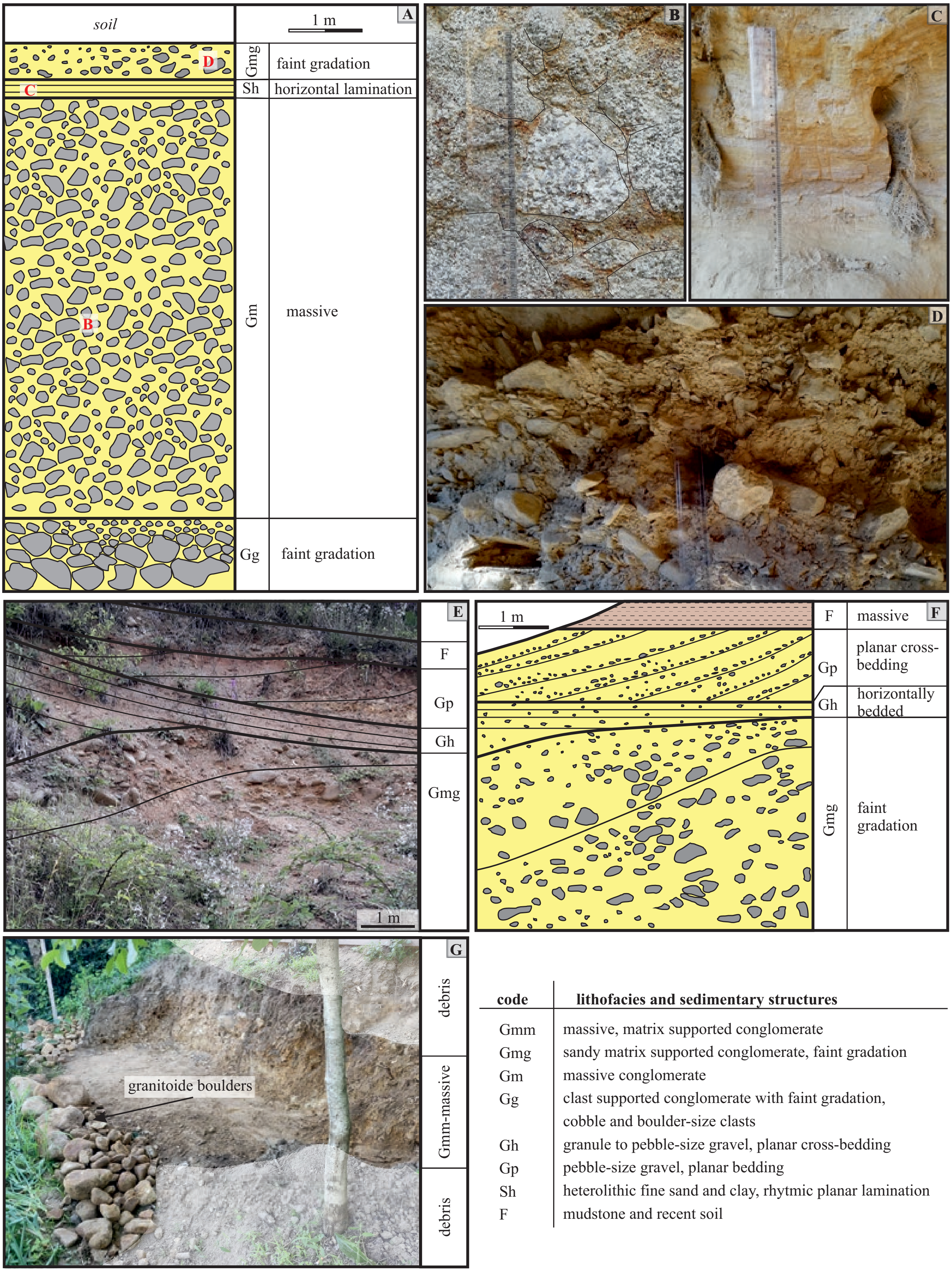
This interval is correlated with conglomerates in the Lozorno-1 well (2100–1790 m; Fig. 5; Table 1), where the transition from polymict conglomerates to granitic conglomerates is also observed. While the granitic conglomerates are cropped out at the surface in the marginal part of the basin (Marianka and Borinka sections), they are overlain by marine sediments of the Jakubov Fm. in the deeper part of the basin (Zohor-1 and Lozorno-1 wells; Fig. 3).

Discussion

Biostratigraphy

In the Zohor-1 well (Fig. 10), conglomerates and heterolithic sediments from the base up to 1311 m do not contain any index microfossils. *Orbulina suturalis* at 1311–1306 m indicates an age younger than 14.9 Ma (see Methods), i.e., the Early Badenian. Association poor in microfossils (containing sponge spicules and fish scales) and lithology of the interval between 2000–1410 m resembles the appearance of the Karpatian Závod Fm. in the close vicinity of this well (Špička & Zapletalová 1964).

The association of the NN5 Zone observed at 1145–1140 m is documented by *S. heteromorphus*, *H. walbersdorfensis*, *H. waltrans* and *R. pseudoumbilicus*. The common occurrence of *H. walbersdorfensis* (18) according to Fornaciari et al. (1996), Di Stefano et al. (2008), and Iaccarino et al. (2011) points to the upper part of the NN5 Zone (MNN5b and MNN5c). The presence and/or high abundance of *S. heteromorphus* together with other supporting taxa (*C. nitescens* + (elliptical), *H. scissura*, *S. heteromorphus*, *U. rotula*, and *U. jafari)* in all samples from the depth of 965–970 m also incline to the NN5 Zone *sensu* Andrejeva-Grigorovič et al. (2001). On the other hand, the common presence of *S. abies* accompanied by *D. exilis* and *C. nitescens* (elliptical) can



code	lithofacies and sedimentary structures
Gmm	massive, matrix supported conglomerate
Gmg	sandy matrix supported conglomerate, faint gradation
Gm	massive conglomerate
Gg	clast supported conglomerate with faint gradation, cobble and boulder-size clasts
Gh	granule to pebble-size gravel, planar cross-bedding
Gp	pebble-size gravel, planar bedding
Sh	heterolithic fine sand and clay, rhytmic planar lamination
F	mudstone and recent soil

Fig. 8. Studied outcrops with marked facial codes. **A** — Schematic sketch of the Borinka section; **B, D** — Details of the section (for localization see 8A); **E** — Marianka outcrop; **F** — Schematic sketch of the Marianka outcrop; **G** — Massive, matrix supported monomict (granitic) conglomerates of the Stupava outcrop.

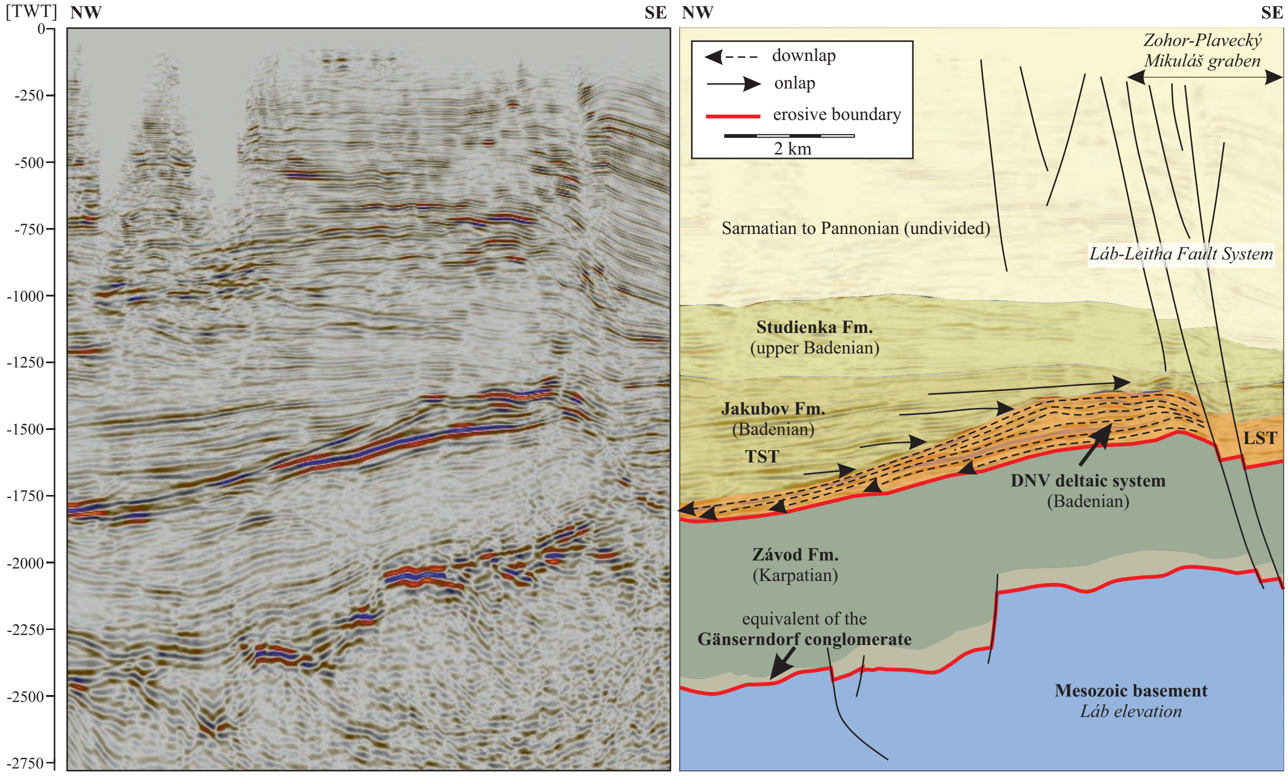


Fig. 9. Re-interpreted seismic line (cross-section) near the Lozorno-1 well. For location, see Fig. 1.

indicate the NN6 Zone, which resemble the nannofossil associations of the Studienka Fm. of the VB (Andrejeva-Grigorovič et al. 2001). The FO of *S. abies* occurs near the NN5 top (Bartol 2010; Bergen et al. 2017) and slightly preceeds the onset of the NN6 Zone. The foraminifers of the Bulimina–Bolivina Zone also suggest that these samples belong to the NN6 Zone. Therefore, the association at 965–970 m corresponds to the uppermost part of the NN5 and to the NN6 Zone respectively (Fornaciari et al. 1996; Andrejeva-Grigorovič et al. 2001; Ćorić & Hohenegger 2008; Di Stefano et al. 2008; Bartol 2010; Iaccarino et al. 2011; Bergen et al. 2017; Vlček et al. 2020). In conclusion, conglomerates from 1410 to ~1300 m belong to the Devínska Nová Ves Formation. Subsequently, Jakubov Fm. sediments extend up to a depth of 966 m. Sediments occurring higher up were deposited around and above the NN5/NN6 transition (Studienka Fm.; Fig. 10).

In the Lozorno-1 well (Fig. 10), co-occurrence of *S. heteromorphus* and *H. waltrans* in mudstones at 1759 m implies the NN5 Zone of the Badenian Jakubov Fm. Nannofossil associations without *S. heteromorphus* in all samples at and above 1656 m (c. 16) allow for the assigning of this interval to the NN6 Zone. The event with a high abundance of *C. premacintyrei* at 1656–1652 m (c. 16) correlates well with the MNN6 Zone interval, which is characterized by common *C. premacintyrei* above the LO of *S. heteromorphus sensu* Fornaciari et al. (1996). Occurrences of the marker species *H. wallichii* (c. 16, 1656–1652 m; c. 14, 1551.5–1550 m and c.7, 1098.4 m), *Orthorhabdus rioi* (c. 5, 901.5 m), *R. pseudoumbilicus* >7 µm (c. 7, 1098.4 m; c. 6, 1001.7–1000 m) and *Orthorhabdus rugosus* (c. 9, 1203–1200 m and c. 5, 901.5 m) also point to the NN6 Zone at 1656–901.5 m (Perch-Nielsen 1985; Young 1998; Marunteanu 1999; Chira & Vulc 2003; Bartol 2009; Jamrich & Halásová 2010). This inference is supported by an association typical for the upper Badenian Bulimina–Bolivina Zone at 1553 m. The fragments of the ostracod *Hemicytheria omphalodes* at 900.2 m indicate the Sarmatian age. Therefore, conglomerates and mudstones at 2100 to ~1660 m belong to the Badenian, and sandstones and mudstones between 1660–901.5 m to the upper Badenian Studienka Fm. The overlying sandstones and mudstones of the Sarmatian Holíč Fm. pass at 830 m to the Pannonian strata (Fig. 10).

Provenance and depositional settings of conglomerate bodies

All described clasts obtained from the two wells and surface outcrops can be derived from rock complexes exposed nowadays on the western slopes of the Malé Karpaty Mts. (Polák et al. 2012). The source of granitoids can be associated with granites and granodiorites exposed in the Bratislava Massif (Bratislava nappe). Chlorite–sericite schists, phyllites, and Bt-paragneisses are presented in the crystalline complex of the Pernek and Pezinok groups. Based on the facies similarity, the Mesozoic clasts are expected to be mainly sourced by sedimentary cover of the Tatric Unit, although it is possible that they had also been sourced by the Fatric or Hronic units. Specifically, the quartz arenites may have been derived from the Lúžna Fm. (Lower Triassic); dark carbonates with calcite

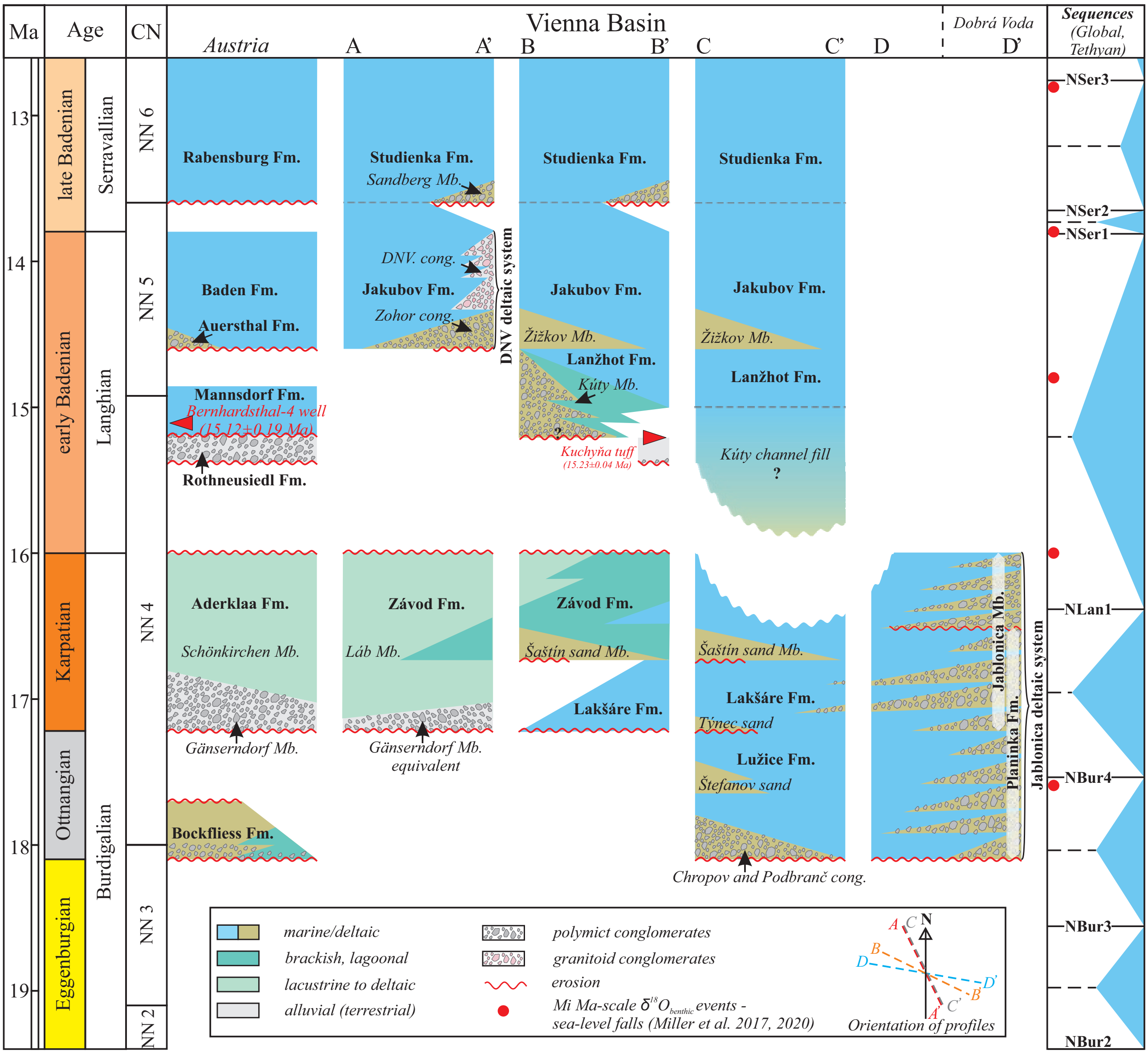


Fig. 10. Cross-border correlation of the Vienna Basin's lithostratigraphic units. Sequences are based on the Time Scale Creator 8). For location of sections, see Fig. 1 (dashed lines). The Austrian part of lithostratigraphy is based on Harzhauser et al. (2020); The Blatné depression of the Danube Basin has been modified according to Rybár et al. (2016), Csibri et al. (2018) and Šujan et al. (2021).

veins without fossils and dolomites from the Gutenstein Fm. (Middle Triassic); sandy limestones from the Prepadlé Fm. (Lower Jurassic); shales from the Jurassic Marianka/Slepý fms. From the Hronic Unit, the pelitic and pelitic-oolitic limestones of the Upper Triassic Norovice Fm. and the basalts of the upper Paleozoic Ipoltica Group (Malužiná Fm.) can be derived. As a consequence, the Central Western Carpathian source for all processed conglomerates can be assumed. The same sources of basal conglomerates from the Zohor-1 well were investigated by Mišík & Reháková (2004). The shift in composition, from polymict conglomerates dominated by carbonate clasts to almost exclusively granitic conglomerates upwards, reflects the tectonic activity in the vicinity. Increasing clast size (boulder to cobble) in granitic conglomerates also reflects the ongoing formation of the Middle Miocene graben and horst structure resulting in the steep morphology of the Malé Karpaty Mts. (e.g., Vass et al. 1988; Fordinál et al. 2010; this study).

The Karpatian conglomerates of the gravity flow origin observed at the base of the Zohor-1 well (2000–1795 m) represent an alluvial fan to fan delta body. They share local provenance of clasts with the Jablonica conglomerates of the northern part of the VB and the Dobrá Voda Depression (e.g., Mišík 1986). These conglomerates can represent a lateral equivalent of the Gänserndorf Mb. of the Aderklaa Fm. in the Austrian part of the basin (Kreutzer 1992; Weissenbäck 1996; Harzhauser et al. 2020). The Gänserndorf Mb. was formed by conglomerates and sandstones to silty clay sediments, which had been deposited in an alluvial fan to the braided river system and in wetland ponds. The Gänserndorf Mb. passes without unconformity into the meandering river system of the Schönkirchen Mb. (Aderklaa Fm.; Harzhauser et al. 2020). Similarly, conglomerates at the base of the Zohor-1 well gradually pass into a 400 m-thick heterolithic interval characterized by intercalations of sandy and muddy layers/laminas. On the basis of a poor microfossil association, it is assigned to the Závod Fm., deposited mainly in deltaic-lacustrine environments with rare marine ingressions (Špička & Zapletalová 1964; 1795–1410 m; Fig. 3). Euryhaline conditions are implied by the presence of *Ammonia* ex. gr. *beccarii*).

The second conglomerate body in the Zohor-1 well (1400–1300 m) arranged in a prograding clinoform downlapping onto the underlying Závod Fm. point to a fan delta environment. The presence of *Orbulina suturalis* confirms the Badenian age and also documents that the conglomerates originated in the marine environment. The presence of the co-occurrence of *O. suturalis* and NN5 Zone nannofossils is also described from pelitic intercalations within the polymict conglomerates in the Devínska Nová Ves-1 well (DNV-1) where they are covered by granitic conglomerates of the DNV Fm. (Zlinská 1987; Vass et al. 1988). It should be noted that originally, the Badenian polymict conglomerate was informally assigned by Vass et al. (1988) as the Zohor conglomerate, while Mišík & Reháková (2004) considered the Zohor conglomerate to be a different body occurring in the depth interval of 1850–2000 m of the Zohor-1 well. Therefore, to maintain the age assignment (Badenian), the deeper occurring conglomerate body of the Karpatian age should be considered to be the temporal equivalent of the Gänserndorf conglomerate.

The transition from polymict conglomerates to granitic DNV conglomerates described by Vass et al. (1988) from DNV-1 is also observed in the studied Lozorno-1 well (2100–1790 m; Fig. 5; Table 1). Such granitic conglomerates outcrop in the Lozorno, Marianka, and Borinka sections (Fig. 8; Vass et al. 1988; this study). On the basis of the DNV-1, MZK-1 and PKH-2 wells situated nearby the studied outcrops (Fig. 1), the conglomerates overlie the Pre-Neogene basement. The sorting and grain-size, together with the absence of fossils, allow us to interpret the granitic conglomerates origin in mass flow of an alluvial to colluvial fan, in accordance with Vass et al. (1988) and Fordinál et al. (2010).

The overlying granitic conglomerates were originally assigned to the Devínska Nová Ves Mb. of the Jakubov Fm. (Vass et al. 1988; Vass 2002). Later, they were grouped with other terrestrial deposits, including Kuchyňa tuff, to the Devínska Nová Ves Fm. (DNV Fm.; Fordinál et al. 2010). The recent study of Kuchyňa tuff (Rybár et al. 2019), which covered the terrestrial mudstones, shows a $^{40}Ar/^{39}Ar$ sanidine age of 15.2 Ma. Therefore, either: (*i*) the polymict Zohor conglomerates and the granititic Devínska Nová Ves conglomerates should be both incorporated into the Devínska Nová Ves Fm., whose deposition started at ~ 15.2 Ma; or (*ii*) the Kuchyňa tuff with the underlying continental mudstones should be excluded from the DNV Fm. We incline to treat the mentioned Badenian sedimentary bodies as individual members (see Fig. 10).

The Vienna Basin deltaic systems: correlations, paleogeography and geodynamics

The Lower Miocene wedge-top "Vienna" basin

The preserved sedimentary record of the older, wedge-top basin, which lasted from the latest Eggenburgian/Ottnangian to the Karpatian, contains several conglomerate bodies. In addition to the basal conglomerates of the Lužice Fm., marginal conglomerate bodies were developed in the NE. Owing to the lack of geochronological and biostratigraphic data, these marginal Planinka Fm. and Jablonica Mb. of the Lakšáry Fm. were poorly constrained in terms of the lithostratigraphic classification. The Planinka Fm. (Kováč et al. 1992) documented in the Dobrá Voda depression was originally proposed to include the Ottnangian to lower Karpatian mudstones, siltstones to conglomerates that overlie either the Eggenburgian sediments, or the pre-Neogene basement. The interruption of sedimentation around the Ottnangian/Karpatian boundary was not detected here (Fig. 10). The Jablonica Mb. of Lakšáry Fm. was assigned to the lower Karpatian (Buday 1955; Buday & Cicha 1956; Jiříček 1988; Vass 2002) and was equalized with the lower Karpatian part of the Planinka Fm. (Vass 2002). On the other hand, the Planinka Fm. in the Dobrá Voda depression is covered by a younger conglomerate body, which is considered the Jablonica Mb. by Kováč et al. (1992) and therefore attributed to the late Karpatian (Kováč et al. 1991, 1992). Identical conditions of the origin and the lack of fossils inhibit accurate stratigraphic assignment of the Planinka and Jablonica conglomerates as well as their connection to the basinal Lužice and Lakšáry fms. (e.g., Buday 1955; Buday & Cicha 1956; Kováč 1985, 1986; Vass 2002; Kováč et al. 2004). While the lower conglomerates (Planinka) consist mainly of clasts from local Mesozoic complexes (Kováč et al. 1991, 1992), the upper conglomerates (Jablonica), in addition to the Mesozoic clasts, also contain Paleozoic rocks (Buday & Cicha 1956; Kováč 1985; Mišík 1986; Fordinál et al. 2012). Thus, we hypothesize that all the Otnangian-Karpatian conglomerates from the NE margin of the VB belong to a single depositional system grouped to the Jablonica deltaic system (Fig. 10). Marginal alluvial to deltaic Jablonica Mb. vertically and laterally passes to offshore turbidites of the Prietž Mb. of the Lakšáry Fm. (Vass 2002; Teťák 2017) that were deposited up to the end of the Karpatian. The Karpatian part of the Jablonica Mb. passes into the deep-water schlier deposits of the Lakšáry Fm. in the north. In the west, coastal sands of the Týnec Mb. are its lateral equivalent (Špička & Zapletalová 1964).

A braided river, as well as alluvial to delta plain sediments of the Gänserndorf Mb. of the Aderklaa Fm. were deposited in the southern part (Figs. 9, 10, 11). These coarse-grained fluvial sediments prograde from the SW towards the NE (e.g., Weissenbäck 1995; Harzhauser et al. 2020). In the Slovak part of the VB, the alluvial to fluvial conglomerates proceed along

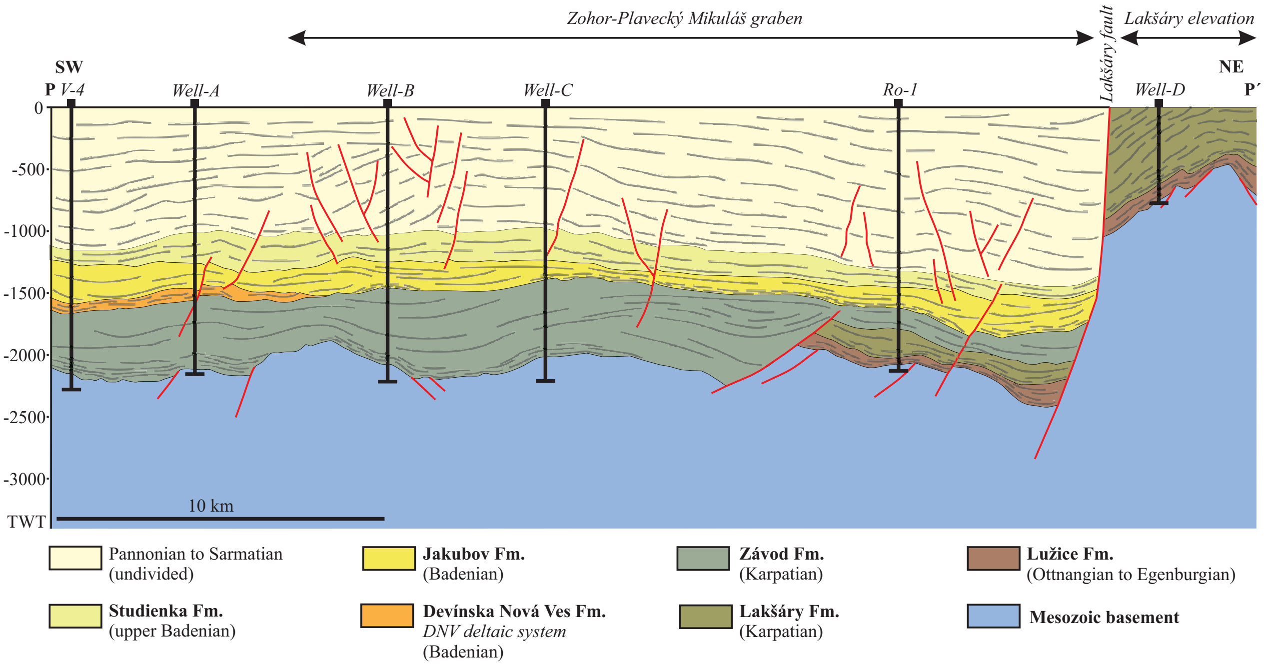


Fig. 11. Composite 2D seismic section along the eastern margin of the Vienna Basin. For location, see Fig. 1.

its eastern margin. In contrast to the clasts of the Gänserndorf Mb. sourced in the Eastern Alps, conglomerates at the base of the Zohor-1 well were derived from the Western Carpathian complexes. They represent the temporal equivalent of both the Jablonica Mb. and the Gänserndorf Mb. conglomerates. In the study area, these conglomerates are covered by a heterolithic flood plain to deltaic sediments of the Závod Fm. (Figs. 3, 5).

Another common misconception and inconsistency in cross-border correlation is rooted in the spatial and temporal assignment of the Laa, Lakšáry, and Závod fms. In Slovakia, the Karpatian strata are divided into the K1 and K2 cycle (Kováč et al. 2004) by local angular discordances that are visible on the seismic section above which the Šaštín sand Mb. is deposited. The tectonic background of this boundary is supported by the extent of the deltaic Závod and marine Lakšáry fms. (Figs. 1, 10, 11). From the south towards the Malacky fault, the alluvial to hyposaline deltaic environment of the Závod Fm. is set on the pre-Karpatian strata, while in the space between the Malacky and Lakšáry faults, the Závod Fm. with marine intercalation (K2) overlies the marine (K1) Lakšáry Fm. (Hlavatý 1996). Unlike in the south, the marine sedimentation continued above the Šaštín sand Mb. in the north (Fig. 10; Špička & Zapletalová 1964). Harzhauser et al. (2019) redefined the Lakšáry and Závod fms. to become members of the marine Laa Fm. without considering the original Závod Fm. description as mostly hyposaline to freshwater deltaic deposits (Špička & Zapletalová 1964). Additionally, the facies development in the south reflects the change from the braided river system of the Gänserndorf Mb. to the meandering river flood plain with channels and lakes of the Schönkirchen Mb. (Fig. 10) of the Aderklaa Fm. (Weissenbäck 1995; Harzhauser et al. 2020). In the Slovak territory, the lacustrine Láb Mb. represents the lateral continuation of the Schönkirchen Mb. (e.g., Vass 2002; Harzhauser et al. 2020).

The spatial distribution of the Lakšáry and Závod fms. hints to paleogeographical changes around the boundary of the “lower” and “upper” Karpatian (Kováč et al. 2004). The full marine environment of the Lakšáry Fm. (Figs. 10, 11), as well as marine ingressions into the Závod Fm. were influenced by active tectonics in the area of the Alpine–Carpathian–Dinaridic orogenic chains affecting the closing and re-opening of marine connections (gateways) towards the Mediterranean or the Eastern Paratethys (Kováč et al. 2018a). This hypothesis is supported by (*i*) the closure of a marine connection between the Central Paratethys with the Mediterranean in front of the Alps at around ~16.5 Ma (Hohenegger et al. 2014; Garefalakis & Schlunegger 2019; Hofmayer et al. 2019), and by (*ii*) subsequent re-openings of the “Dinaridic” or “Carpatian” gateway around the latest Burdigalian–earliest Langhian (*sensu* e.g., Bistričić & Jenko 1985; Rögl 1988; Sant et al. 2017; Ivančič et al. 2018; Kováč et al. 2018a; Hudáčková et al. 2020; Ruman et al. 2021). The presence of a marine connection at ~16.5–16 Ma is supported by the occurrence of deep-water marine sediments in the Gbely-100 well dated by foraminiferal $^{87}Sr/^{86}Sr$ ratio to 16.3–15.9 Ma (Hudáčková et al. 2003). A similar Sr-based age of 16.1–15.3 Ma was determined from the Cerová-Lieskové site (Less et al. 2015), a locality characterized by the presence of stenohaline fauna and flora, the NN4 Zone, and the foraminifera *Uvigerina graciliformis* and *Trilobatus bisphericus* (Harzhauser et al. 2011b; Hyžný & Schlögl 2011; Hudáčková et al. 2020). This site resembles the Wagna site where the sediments containing *Trilobatus bisphericus* are correlated with C5Cn.3n-C5Cn.2r chron (Hohenegger et al. 2014). The presence of *T. bisphericus* and the absence of praeorbulinas were also utilized for the determination of the late Karpatian age of the Laa Fm. This formation is overlain discordantly by the Ginzersdorf Fm. channel fill (~16.3–16 Ma; Ćorić et al. 2004; Harzhauser et al. 2019). Sediments of the Ginzersdorf Fm. show lithological similarities with the Cerová–Lieskové site and also contain deep shelf (even upper bathyal) mollusk associations (Harzhauser et al. 2011b, 2017, 2019). Canyons filled by the Ginzersdorf Fm. sediments served as a precursor for the earliest Badenian drainage system from the VB into the Alpine–Carpathian Foredeep Basin (Harzhauser et al. 2019). Although such canyons cutting into the Lower Miocene strata documented in the VB and surroundings (e.g., Adámek 2003; Dellmour & Harzhauser 2012; Harzhauser et al. 2019) are generally linked with a fall in sea level, the incision can be triggered possibly by the tectonic inversion of the wedge-top (piggyback) basin induced by the collision of the orogenic system with the platform. Due to large-scale erosion, a considerable thickness of the upper Karpatian–lowermost Badenian sediments is likely absent (e.g., Jiříček 1979, 1985; Jiříček & Seifert 1990; Harzhauser et al. 2019, 2020). This tectonic event led to the origin of a distinct unconformity separating the Lower Miocene from the Middle Miocene sequences (Ćorić & Rögl 2004; Kováč et al. 2004; Harzhauser et al. 2020). In addition to the VB, a significant erosional boundary below the extensive Badenian (mid-Langhian) marine flooding (~15 Ma) has been documented in the Danube and Novohrad-Nógrád basins as well (e.g., Kováč 2000; Kováč et al. 2004, 2005, 2007; Harzhauser et al. 2020; Hudáčková et al. 2020; Šujan et al. 2021).

The Middle Miocene extensional Vienna Basin

The earliest Badenian (B1) depositional cycle (*sensu* Strauss et al. 2006) can be partly correlated with the short-lasting marine flooding of the VB and Carpathian Foredeep above the Bur5/Lan1 boundary (e.g., Harzhauser et al. 2020). The cycle starts during a period of erosion enhanced by tectonics and followed by the filling-up of these canyons incised in the older deposits, mainly of Karpatian age (Siedl et al. 2020). In the Austrian part and the adjacent area of the Alpine–Carpathian Foredeep, these sediments form the Iváň and Grund fms. (Harzhauser et al. 2019, 2020). The Iváň Fm., which was originally described as the Iváň Mb. with assemblages belonging exclusively to the NN4 Zone and without orbulinas, is bracketed by angular unconformities and overlain by the sediments contained in the NN5 Zone (Adámek 2003; Adámek et al. 2003; Tomanová Petrová & Švábenická 2007; Brzobohatý & Stráník 2011). Similarly, the Grund Fm.contains assemblages typical of the "Lower Lagenidae Zone", and the NN5 Zone (Spezzaferri 2004; Ćorić & Švábenická 2004) overlies the basal clastic with the NN4 Zone (Ćorić & Rögl 2004; Ćorić et al. 2004). Analogous incised canyons are visible in seismic sections from the Kúty depression in the northern (Slovakia) part of the VB (Fig. 10), with overall fill-thickness reaching up to 400 m (3D seismic block performed by Nafta a.s.).

In general, the Badenian (B1) sedimentary record of the VB starts with the alluvial Rothneusiedl Fm. (previously labelled the Aderklaa conglomerate Fm.; e.g., Strauss et al. 2006; Siedl et al. 2020; Harzhauser et al. 2020) in the central and southern parts of the VB (Fig. 10). The formation is overlain by marine sediments of the Mannsdorf Fm. (Harzhauser et al. 2020) and dated by tuff from the Bernhardsthal-4 well to 15.12±0.19 Ma (Harzhauser et al. 2020; Sant et al. 2020). A similar age of 15.2 Ma was obtained from the terrestrial tuff of the Kuchyňa site (Rybár et al. 2019) located at the eastern margin of the VB. The Kúty Mb. formed by the littoral basal clastics of the Lanžhot Fm. (Vass 2002) can be a lateral equivalent of the early Badenian Mannsdorf Fm. These conglomerates, sandstones, and motley pelites with anhydrite reaching a thickness of 500 m pass into the marine clays and silts of the Lanžhot Fm. (Špička 1966; Vass 2002). The presence of *Orbulina suturalis* and the NN5 Zone in the upper part of the Lanžhot Fm. indicates that this formation is younger than the Mannsdorf Fm. (Harzhauser et al. 2020); therefore, it likely belongs to the next Badenian sequence (*sensu* Harzhauser et al. 2020; Siedl et al. 2020).

The second, Badenian (B2) pronounced depositional cycle spread throughout the entire VB. This cycle can be approximately correlated with a transgression above the Lan2/Ser1 boundary (*sensu* Hardenbol et al. 1998). The base of this cycle is represented by the polymict conglomerates preserved in the Zohor-1 (1306–1311 m) and Lozorno-1 (2012–2016 m) wells (Fig. 3) and also detected in deep wells along the basin's eastern margin (e.g., Záhorská Ves-2, Vysoká-4, Vysoká-19, Láb-93; Biela 1978; Fig. 1). Seismic data show clinoforms of the conglomerate fan lobes prograding from the SE to the NW (Fig. 9). The Badenian age of this alluvial to the fan-delta Devínska Nová Ves (DNV) system (Fig. 10) is supported by nannofossils of the NN5 Zone in the DNV-1 well, and *Orbulina suturalis* in the DNV-1 and Zohor-1 wells (Zlinská 1987; Vass et al. 1988; this study). These polymict conglomerates pass upwards to conglomerates dominated by granitoid clasts (DNV conglomerates) derived from the uplifting Malé Karpaty Mts. horst. However, the stratigraphic position of polymict conglomerates is similar to the position of the older Rothneusiedl Fm. lying above an erosional contact with the Karpatian strata (Figs. 9, 11). On the other hand, *O. suturalis* shows that the polymict conglomerates are rather coeval with conglomerates and sands of the coastal marine delta of the Auersthal Fm. (Harzhauser et al. 2020; Fig. 10). On the basis of the presence of *O. suturalis* and the NN5 Zone, Harzhauser et al. (2020) hypothesized that the Auersthal Fm. is temporally equivalent to the Kúty Mb. and lagoonal Žižkov Mb. of the Jakubov Fm. in the west (Fig. 10). The Žižkov Mb. located mainly on the basin's western margin separates the basinal Lanžhot and Jakubov fms. (Buday 1955; Špička 1966; Vass 2002; Kováč et al. 2004; Fig. 10). The Lanžhot Fm. (without its Kúty Mb.) to the "Upper Lagenidae Zone" with *O. suturalis*, as well as the Jakubov Fm. assigned to the "Spirorutilus carinatus Zone" are equivalents of the marine Baden Fm. defined in the Austrian part, as well as the deltaic Hrušky Fm.

defined in the Czech Republic (Harzhauser et al. 2020). Other Badenian VB marginal facies are bound to structural elevations along with the Leitha-Láb fault system. In Austria, these shallow-water limestones are referred to as the Leitha Fm., which is correlated with the carbonates of the Stupava Mb. of the Jakubov Fm. in Slovakia (Kováč et al. 2004; Harzhauser et al. 2020).

However, the original "middle Badenian" Spirorutilus carinatus Zone is currently assigned to the late Badenian (e.g., Hohenegger et al. 2014), and so the Jakubov Fm. together with the basal Žižkov Mb. may represent deposition during the fall in sea level around 13.8 Ma. In some parts of the Central Paratethys, the sea-level fall at the Langhian/Serravallian boundary (early-late Badenian) is coupled with a salinity crisis (~13.8–13.6 Ma; de Leeuw et al. 2010). In the western part of the VB, this event is correlated with a major relative sea-level drop by about 200 m, resulting in a rapid shift from marine depositional environments to coastal and freshwater swamps (Harzhauser et al. 2018). On the eastern margin of the VB, the Badenian sediments with the NN5 Zone are conformably overlain by deposits of the "upper" Badenian NN6 Zone (Vass 2002).

The upper Badenian (B3) depositional cycle (13.8–12.6 Ma) can be correlated with a transgression above the Ser2 and represents the last marine flooding of the eastern margin of the VB. This event can be correlated both with the sea-level rise in the Mediterranean (Kováč et al. 2004, 2007, 2008), as well as the inflow from the Eastern Paratethys (Palcu et al. 2015; Kováč et al. 2017, 2018a). The cycle is represented by the marine Studienka Formation in the NE part of the VB (Vass 2002; Kováč et al. 2004; Fig. 10).

Conclusions

The reinterpretation of older data and new observations have improved the cross-border correlation and specify the influence of local tectonics and sea-level changes on the sedimentary record in the area between the Eastern Alps and Western Carpathians during the Karpatian–Badenian stages.

The basal conglomerate in the Zohor-1 well forms an alluvial to delta fan body, covered by the Karpatian hyposaline deltaic Závod Fm. Thus, it represents the temporal equivalent of the Gänserndorf Mb. (Aderklaa Fm.). These conglomerates are composed of clasts derived from the Central Western Carpathian units (Tatric super-unit – crystalline basement plus sedimentary cover and Hronic nappe Unit). These studied conglomerates possess similar provenance as the co-eval Jablonica deltaic system, which passes in the NE to the marine sediments of the Lakšáry Fm. The relationship between the Závod and Lakšáry fms. is influenced by the tectonic activity, topography, and bathymetry of the Lower Miocene wedge-top basin, as well as by changes in CP marine connections. The alluvial plain sediments of the Aderklaa and Závod fms. are preserved in the south, while marine sediments of the Lakšáry Fm. are preserved in the north. Above the intra-Karpatian boundary, these hyposaline and marine sediments alternate in the central part of the VB.

The second studied conglomerates belong to the DNV deltaic system, which is typical by its change from the polymict Zohor conglomerates to the granitic Devínska Nová Ves conglomerates. Both types can be sourced from the rocks of the Malé Karpaty Mts. Change in the lithology and the presence of boulder-sized clasts in the upper DNV conglomerate reflect the formation of graben and horst structures during the evolution of the Middle Miocene VB depocenters, as well as point to a steep morphology of slopes in the study area. The presence of *Orbulina suturalis* and nannofossils of the NN5 Zone in the polymict conglomerates, including radioisotopic data from the Kuchyňa tuff, document that the DNV deltaic system began deposition after 14.9 Ma. Thus, although having a similar structural position to the Rothneusiedl Fm. (Aderklaa conglomerate Fm.), they still represent an equivalent to the Austrian Auersthal Fm. Marine sediments of the Lanžhot and Jakubov fms. (jointly equivalent to the Austrian Baden Fm.), which form the basinal equivalent of the DNV deltaic system, contain nannofossils of the NN5 Zone and *O. suturalis* as well.

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Supplement

Table S1: Nannoplankton assemblages of the Zohor-1 well. (Abbreviation: PRES – presence, ABN – abundant, SABN – super abundant, (rew.) – reworked)

Depth (core)	box	Zone/ Subzone	Event
965–970m (core 1)	1	NN6/?NN5	TOP PRES *Helicosphaera walbersdorfensis*, *Discoaster exilis*, *Reticulofenestra pseudoumbilicus*, *Sphenolithus abies*, *Coronocyclus nitescens* (elliptical), *Umbilicosphaera rotula*, *U. jafari*, SABN *Coccolithus pelagicus*, *Reticulofenestra minuta*, *R. haqii*, ABN *Helicosphaera carteri*, *H. scissura* (?rew.), *Sphenolithus heteromorphus* (?rew.), early Miocene, Paleogene and Cretaceous reworking;
965–970m (core 1)	2	NN6/?NN5	PRES *S. abies*, *H. walbersdorfensis, D. exilis, Triquetrorhabdulus* spp., ABN *Reticulofenestra minuta*, *R. haqii*, *S. heteromorphus* (?rew.), early Miocene, Paleogene and Cretaceous reworking;
965–970m (core 1)	3	NN6/?NN5	PRES *S. abies*, *H. walbersdorfensis*, *H. carteri*, *C. nitescens* (elliptical), *Discoaster musicus, Triquetrorhabdulus* spp., *T. milowii*, *U. rotula*, *U. jafari*, SABN *R. minuta*, ABN *R. haqii*, *C. pelagicus*, *S. heteromorphus* (?rew), Early Miocene, Paleogene and Cretaceous reworking;
1140–1145m (core 2)	1	Unassigned	Barren, pyrite;
1140–1145m (core 2)	2	Unassigned	PRES *C. pelagicus*, pyrite; poor sample;
1140–1145m (core 2)	3	NN5 NN5b/c	PRES *H. walbersdorfensis, H. waltrans*, *S. heteromorphus*, *Calcidiscus tropicus*, *C. nitescens*, *R. pseudoumbilicus*, *U. rotula*, *U. jafari*, SABN *R. minuta*, ABN *R. haqii*, *C. pelagicus*, Paleogene and Cretaceous reworking;
1140–1145m (core 2)	4	NN5 NN5b/c	PRES *H. walbersdorfensis, H. scissura*,*S. heteromorphus*, ACME *C. tropicus*, *C. nitescens* (elliptical), *U. rotula*, *U. jafari*, *R. pseudoumbilicus*, SABN *R. minuta*, *R. haqii*, *C. pelagicus*, Paleogene and Cretaceous reworking;
1140–1145m (core 2)	5	?NN5	PRES *S. heteromorphus*, *C. pelagicus*, *H. carteri*, poor sample;
1495–1500m (core 4)	1	Unassigned	PRES *C. pelagicus*, *Cyclicargolithus floridanus*, *R. haqii*, Diatomaceae, Early Miocene, Paleogene and Cretaceous reworking; poor sample;
1495–1500m (core 4)	2	Unassigned	PRES *C. pelagicus*, *C. floridanus*, *S. abies*, *S. moriformis*, *R. haqii*, Paleogene and Cretaceous reworking; poor sample;
1495–1500m (core 4)	3	Unassigned	PRES *C. pelagicus*, *C. floridanus*, *H. carteri*, Early Miocene, Paleogene and Cretaceous reworking; poor sample;
1495–1500m (core 4)	4	Unassigned	PRES *C. pelagicus*, *C. floridanus*, *R. haqii*, Paleogene and Cretaceous reworking; poor sample;
1495–1500m (core 4)	5	Unassigned	PRES *C. pelagicus*, *C. floridanus*, *H. carteri*, *Holodiscolithus macroporus*, early Miocene, Paleogene and Cretaceous reworking; poor sample;
1886–1891m (core 6)	1	Unassigned	PRES *Discoaster* spp., *H. macroporus*, *R. minuta*, *Triquetrorhabdulus* spp.; poor sample;

Table S2: Nannoplankton assemblages of the Lozorno-1 well.

Depth (core)	box	Zone/ Subzone	Event
650–655m (core 1)	2	Unassigned	Small reticulofenestrids, Paleogene and Cretaceous reworking;
~	1	Unassigned	Small reticulofenestrids, Paleogene and Cretaceous reworking;
901.5m (core 5)	2	NN6	*Calcidiscus tropicus, C. leptoporus*, *Coccolithus miopelagicus*, *C. pelagicus*, *Coronocyclus nitescens*, *Discoaster variabilis, Helicosphaera carteri, H. walbersdorfensis*, *Orthorhabdus rioi*, *Reticulofenestra haqii, R. minuta*, *R. pseudoumbilicus*, *Rhabdosphaera pannonica, Sphenolitus abies, S. moriformis, Syracosphaera pulchra, Umbilicosphaera rotula, Triquetrorhabdulus rugosus*; Strong Paleogene and Cretaceous reworking;
1000–1001.7m (core 6)	1	NN6	*C. pelagicus, C. nitescens*, *R. pseudoumbilicus* (> 7mµ), *R. haqii*, *R. minuta*, *U. rotula*; poor sample;
1098.4m (core 7)	1	NN6	*C. pelagicus, C. nitescens*, *Cyclicargolithus floridanus, H.carteri, R. pseudoumbilicus* (>7mµ), *Pontosphaera multipora*, *R. haqii*, *R. minuta*, *U. rotula*; poor sample;
1101.8m (core 7)	4	Unassigned	*C. pelagicus*, *C. floridanus, C. nitescens*, *H. carteri, H. wallichii*, *R. pseudoumbilicus*, *R.haqii, R. minuta*, *U. rotula*, Paleogene reworking;
1200–1203m (core 9)	1	Unassigned	*C. pelagicus*, *C. nitescens*, *H. carteri*, *R. pseudoumbilicus*, *R. minuta*, *R. haqii*, *T. rugosus* ; poor sample;
1405.6–1410.1m (core 11)	1	Unassigned	*C. pelagicus*, *C. floridanus*, *D. variabilis*, *R. minuta, R.haqii*, *R. pseudoumbilicus*, Paleogene and Cretaceous reworking; poor sample;
1450.7–1453m (core 12)	1	NN6	*Braarudosphaera bigelowii*, *C. pelagicus*, *C. floridanus, H. carteri*, *R. pseudoumbilicus*, *Thoracosphaera* spp., *U. rotula*, Paleogene and Cretaceous reworking;
1550–1555m (core 14)	1	NN6	*C. pelagicus*, *C. floridanus, H. macroporus*, *H. carteri, H. mediterranea, H. wallichii, H. walbersdorfensis, Micrantholithus* spp., ABN *Rh. pannonica, R. haqii and R. pseudoumbilicus*, *P. multipora*, *Micrantholithus* spp., *Thoracosphaera* spp., Paleogene and Cretaceous reworking; ABN pyrite;
1596.3–1599m (core 15)	1	NN6	*Calcidiscus premacintyrei, C. leptoporus, C. pelagicus*, *C. floridanus, C. nitescens*, *Discoaster deflandrei*, *H. macroporus*, *H. carteri*, *H. wallichii, Rh. pannonica*, *S. abies,* S. *moriformis*, *Thoracosphaera* spp., *U. rotula*, ABN small reticulosfenerids (*R. haqii*, *R. minuta*), *R. pseudoumbilicus*, Paleogene and Cretaceous reworking; pyrite, carbonized plant fragments
1652–1656m (core 16)	1	NN6 MNN6	*Braarudosphaera bigelowii parvula*, ACME *C. premacintyrei, C. tropicus, C. pelagicus*, *C. nitescens* (elliptical form), *C. floridanus* , *H. carteri*, *H. walbersdorfensis*, *H. wallichii*, *P. multipora*, *Rh. pannonica*, ABN small reticulosfenerids (*R.haqii*, *R. minuta*), *R. pseudoumbilicus*, Paleogene and Cretaceous reworking;
1756–1759m (core 18a)	1	NN5, NN5a	*C. premacintyrei*, *C. pelagicus*, *C. floridanus*, small reticulofenestrids (*R. haqii, R. minuta*), ABN helicospherids (*H. carteri*, *H. mediterranea, H. scissura, H. walbersdorfensis*, *H. waltrans*), *P. multipora*, *R. pseudoumbilicus*, *Rhabdosphaera* and scarce *S. heteromorphus* together with Cretaceous and Eocene reworkings;
2097-2099m (core 25)	1	Unassigned	*C. pelagicus*, *H. carteri*, *R. haqii;* Silicoflagellate fragment;