Excursion 2

Neogene of the Styrian Basin

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Introduction

The Styrian Basin, as a subbasin of the Pannonian Basin System, established during the Neogene at the eastern margin of the Eastern Alps. It is about 100 km long, about 60 km wide and contains Neogene sediments of about 4 km thickness. The basin is divided into several small subbasins such as the Western Styrian Basin, the Mureck Basin, the Gnas Basin, and the Fürstenfeld Basin. It is separated from the Pannonian Basin by the South Burgenland Swell and is internally structured by the Middle Styrian Swell and the Auersbach Swell (Fig. 1). An overview of the tectonic evolution of the Styrian Basin is provided by Sachsenhofer (1996).

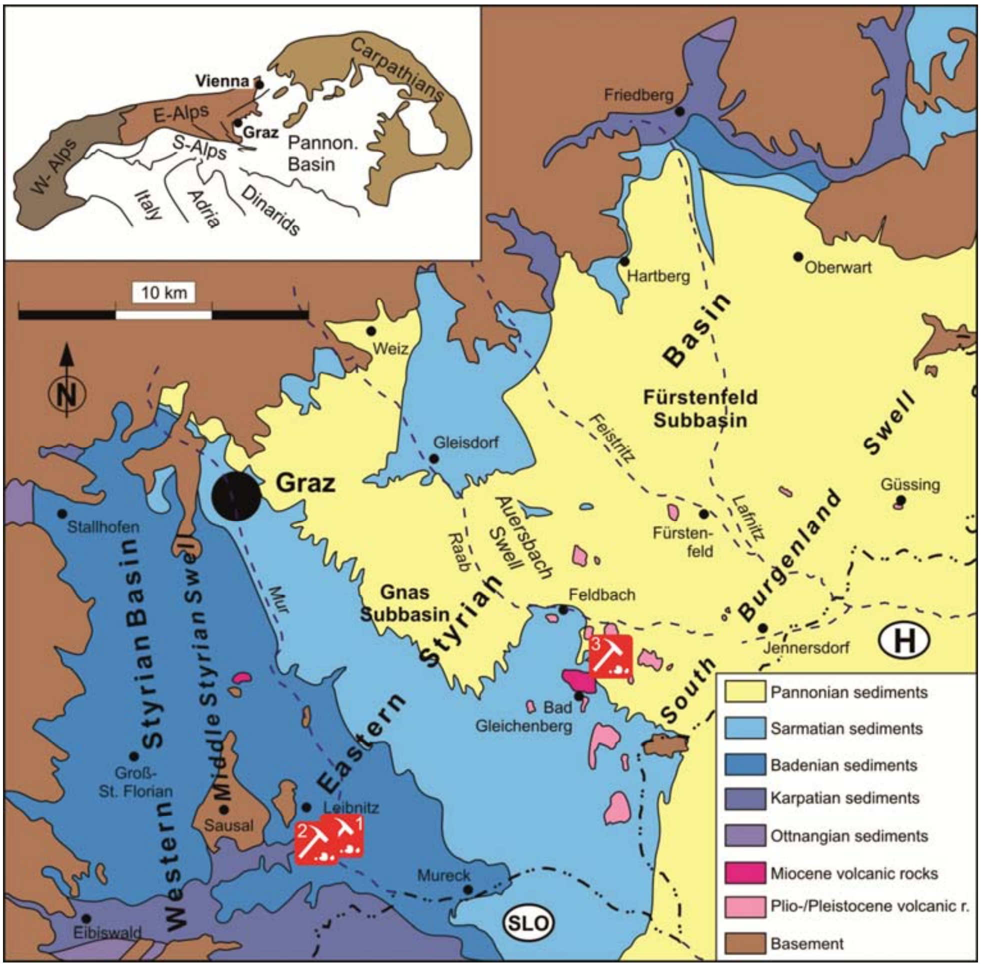

Figure 1: Geological map of the Styrian Basin (Gross et al., 2007) showing the position of the visited localities (hammer symbols: 1 = Brickyard Wagna, 2 = quarry Retznei, 3 = clay pit Mataschen.

Paleogeographically the Styrian Basin was part of the Central Paratethys. Reaching from Bavaria in the West to the Carparthian Mountains in the East, this shallow epicontinental sea originated during the latest Eocene and Early Oligocene due to the rising Alpine island changes which acted as geographic barriers (Rögl, 1998). Geodynamic changes related to the convergence of the Afro-Arabian and Eurasian plates superimposed by sea level fluctuations initiated a complex pattern of changing seaways and landbridges between the Central, Western and Eastern Paratethys, the Mediterranean Sea and the Indian Ocean. This caused the biogeographic separation of the Central Paratethys and required the establishment of regional chronostratigraphic stages (Fig. 2). Times of open connections of the Paratethys with adjacent oceans (e.g., middle Miocene Badenian regional stage) are reflected by a very low rate of endemism (Harzhauser and Piller, 2007). During these phases, the exchange of plankton allows a biostratigraphic correlation with coeval Mediterranean areas. In contrast, phases of total or partial isolation coincide with considerable endemisms and usually also with a near-complete breakdown of all biostratigraphically relevant planktonic groups. In the Central Paratethys, the Sarmatian and Pannonian regional stages represent phases of apparently complete isolation; their correlation to Mediterranean records has been controversial since decades (see Papp et al., 1974, 1985; Stevanović et al.,1990; Lirer et al., 2009 for discussions). A summary of the lithostratigraphic units in the Styrian Basin and their chronostratigraphic correlation is given in Fig. 2.

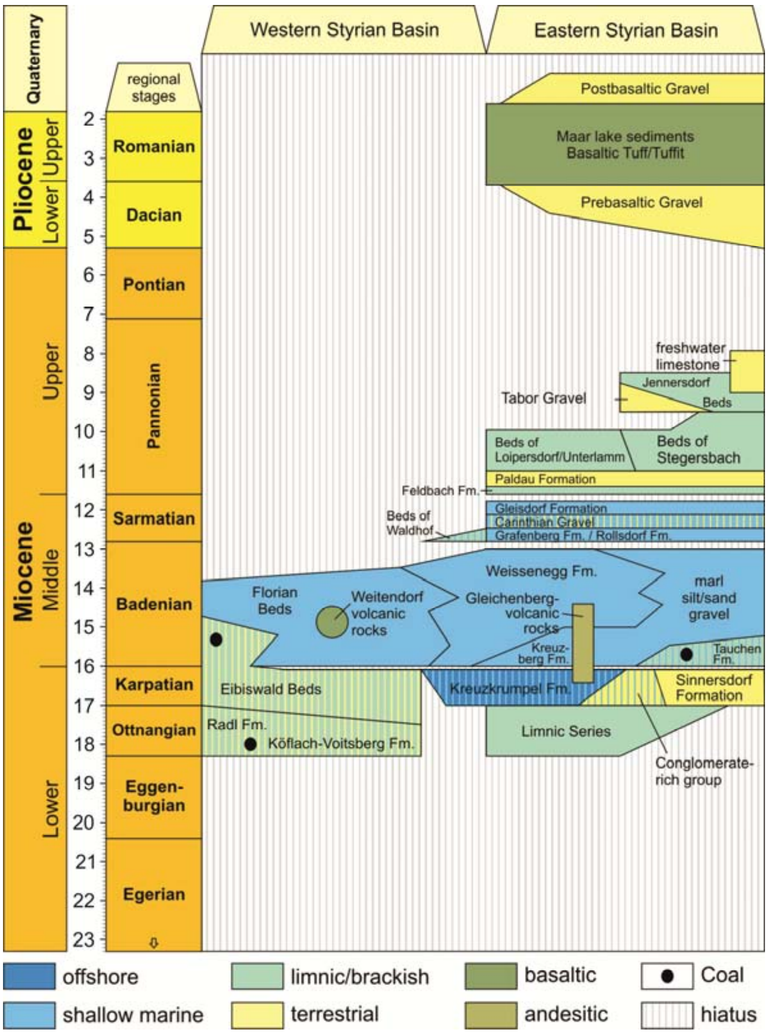

Figure 2: Stratigraphic chart of the Styrian Basin (Gross et al., 2007).

1.1. Early Miocene

1.1.1. Ottnangian

Basin fill started in the Early Miocene with limnic-fluvial sediments (red soils, breccias, marls with coal seams and conglomerates) of the "Limnic Series" and alluvial fan and delta sediments (e.g., Radl Formation, „Lower Eibiswald Beds", „Beds of Naas", „Breccia of Zöbern") deposited in proximal settings (e.g., Bay of Eibiswald, Weiz, Friedberg-Pinkafeld; Kollmann, 1965, Stingl, 1994; Fig. 3A). These continental deposits are poorly dated except of coal-bearing, limnic-fluvial sediments in the Bay of Stallhofen (Köflach-Voitsberg Formation), which can be assigned to the Ottnangian regional stage by integrated bio- and magnetostratigraphy (Haas et al., 1998; Steininger et al., 1998). The Köflach-Voitsberg Formation contains also the oldest tuffs of the Styrian Basin (Ebner et al., 2000).

1.1.2. Karpatian

The Karpatian was a time of increased tectonic activity, which caused the differentiation in an eastern and a western subbasin by the uplift of the middle Styrian and Leibnitz swells. Strong subsidence led to rapid drowning of the Eastern Styrian Basin resulting in several hundred meters thick marine mud- and siltstones with sandy, turbiditic intercalations ("Styrian Schlier" or Kreuzkrumpel Formation; Friebe, 1990; Schell, 1994; Rögl et al., 2002; Fig. 3B). The Trans-Tethyan Trench Corridor provided a marine connection with the Western Tethys/Proto-Mediterranean Sea via Slovenia at this time (Rögl, 1998). Eruptive volcanism also occurred in the Eastern Styrian Basin in response to extensional tectonics (Balogh et al. 1994; Slapansky et al. 1999), producing a volcanic island complex. Its major eruptive center was a 500 km^2 by 1,000-m-thick shield volcano in the area of Bad Gleichenberg (Fig. 3B).
In contrast, limnic-fluvial sedimentation continued in the Western Styrian Basin (Fig. 3B). Fluvial fan sediments (Sinnersdorf Formation; Nebert, 1985) dominate the Bay of Friedberg - Pinkafeld and are supposed to extend into the Fürstenfeld Subbasin (Goldbrunner, 1988). Limnic-deltaic sediments north of the Bay of Stallhofen („Conglomerate of Stiwoll"; Flügel, 1975) and fine-clastics with bentonites in the Bay of St. Florian are also questionably assigned to the Karpatian (Kollmann, 1965; Ebner and Sachsenhofer, 1991). Distal delta slope environments at the transition of Western and Eastern Styrian Basin are characterised by subaquatic mass flows („Arnfels Conglomerates", „Leutschach Sands"; Winkler, 1927a).
The tectonic activity increased at the end of the Karpatian and caused block tilting and unconformities in shallow water areas (Wagna, Retznei, Katzengraben/Spielfeld) across the Early/Middle Miocene boundary (Styrian Tectonic Phase; Stille, 1924).

1.2. Middle Miocene

1.2.1. Badenian

During the Badenian a stable seaway (Trans-Tethyan Trench Corridor) via Slovenia as well as intermittent seaways into eastern directions connected the Pannonian Basin System with the Mediterranean Sea (Piller et al., 2007). These connections enabled three marine transgressions into the Central Paratethys that correlate with the TB 2.3–2.5 sea level cycles of Haq et al. (1988) (Strauss et al., 2006). Facilitated by the open seaways and warm climate of the Middle Miocene Climate Optimum (ca. 17–15 Ma), tropical coral reef ecosystems

extended northwards into the Central Paratethys Sea during the Badenian for the only time in the Neogene (Esteban, 1996; Perrin and Bosellini, 2012). These marginal reef coral communities are generally low diverse (usually less than 5 genera at the same site) and characterized by a low framework-building capacity. Non-framework forming coral communities and coral carpets dominated while higher diverse (up to 10 coral genera at the same site) coral patch reefs formed just briefly during the climax of the Middle Miocene Climate Optimum along the western coast (Styrian, Slovenian, Vienna basins) and spatially restricted to areas sheltered from siliciclastic input like the middle Styrian Swell or the isolated Leitha Mountains carbonate platform in the Vienna Basin (Riegl and Piller, 2000a, b; Perrin and Bosellini, 2012; Wiedl et al., 2013).
During the Badenian marine sediments reached its largest extent in the Styrian Basin despite a reduced subsidence (Friebe, 1990; Rögl, 1998; Kováč et al., 2004; Fig. 3C). In the Western Styrian Basin limnic-fluvial ("Eibiswald Beds", "Beds of Rein", Stallhofen Formation) and lagoonal ("Beds of St. Florian") sedimentation prevailed at this time (Rolle, 1855; Hilber, 1878; Ebner and Gräf, 1979; Ebner and Stingl, 1998; Hiden and Stingel, 1998; Ebner et al., 2000; Gruber et al., 2003; Fig. 3C). Coarse-clastics ("Schwanberg Beds") at the western margin of the basin point to the uplift of the basement (Nebert, 1989). At the northeastern margin of the basin (Bay of Friedberg-Pinkafeld, Fürstenfeld Subbasin) conglomerates, corallinacean limestones and paralic coals (Tauchen Formation) formed in shallow marine-deltaic environments (Nebert, 1985; Fig. 3C).
In areas of low terrigenous sedimentation, such as the Middle Styrian Swell, South Burgenland Swell and around shield volcanoes, the early Badenian transgression promoted the wide-spread development of coralline algal limestones and coral patch reefs (Fig. 3C). These carbonates and associated shallow marine siliciclastics are integrated in the Weissenegg Formation (Friebe, 1990) and interfinger with coarse-siliciclastic, deltaic deposits of the Kreuzberg Formation. Deeper water sedimentation of marine muds and turbidites characterizes central parts of the basin. The Gleichenberg volcano remained active in the Early Badenian but the eruptive center shifted to the north (Ilz-Walkersdorf). Another important shield volcano, extending ca. 125 km^2 and 200–300 m thick, formed on the Middle Styrian Swell at this time (Weitendorf volcanics; Slapansky et al., 1999; Fig. 3C).
A regression at the Badenian/Sarmatian, which corresponds with a global sea-level fall (Harzhauser and Piller, 2004a, b), caused the erosion and the progradation of fluvial ("Eckwirt Gravels") and deltaic sediments (Dillach Member of the Weissenegg Formation; Friebe, 1990).

1.2.2. Sarmatian

During the upper Middle Miocene Sarmatian Stage the Paratethys Sea formed a huge inland sea which was nearly completely disconnected from the Mediterranean Sea. This strong isolation caused serious environmental changes, which were critically evaluated in the last years (Latal et al., 2004; Harzhauser and Piller, 2004a; Piller and Harzhauser, 2005). Traditionally, the Sarmatian was interpreted as transitional from the marine Badenian Sea towards the temperate-freshwater environments of Lake Pannon (Papp, 1954, 1956). This interpretation was mainly based on the absence of stenohaline biota such as radiolaria, planktic foraminifera, corals and echinoderms (Steininger and Wessely, 2000), which disappeared at the Badenian/Sarmatian boundary. New microfacial, palaeontological and

geochemical data, however, clearly point to marine waters for the entire Sarmatian along the western margin of the Pannonian Basin System (Piller and Harzhauser, 2005).
During the Early Sarmatian siliciclastic sedimentation prevailed in the Styrian Basin and the basin margin was affacted by the drainage systems from the Alps. The gastropod *Mohrensternia* and the bivalve *Crassostrea* flourished in resultant hyposaline coastal environments (Rollsdorf Formation, Harzhauser and Piller, 2004b). For the central part of the basin small bryozoan-serpulid buildups that developed in the Bay of Friedberg-Pinkafeld (Grafenberg Formation) or close to the South Burgenland Swell (Klapping/St. Anna) indicate, however, normal marine conditions (Harzhauser and Piller, 2004a; Fig. 3D). The absence of many stenohaline marine biota (e.g., corals and echinoderms) at this time may be related to the sea level drop at the Badenian/Sarmatian boundary, which together with tectonic activities probably interrupted seaways into the Mediterranean/Indo-Pacific (Rögl, 1996) and prohibited re-immigration of these biota (Piller and Harzhauser, 2005). The "Carinthian Gravel" at the top of the Grafenberg Formation indicates a regressive phase following the formation of the carbonates (Winkler, 1927b; Skala, 1967).
During the Late Sarmatian a highly productive carbonate factory of oolite shoals, mass-occurrences of thick-shelled molluscs and larger foraminifera, as well as marine cements clearly point to shallow, normal marine to hypersaline, carbonate supersaturated conditions (Piller and Harzhauser, 2005). The fully marine to hypersaline conditions in the Late Sarmatian seem to be related with an opening of a seaway into the Mediterranean Sea as indicated by the sudden appeareance of certain molluscs (*Gibbula buchi*, *Jujubinus turricula*, *Mitrella agenta*) in the Central Paratethys Sea (Piller and Harzhauser, 2005). In the Styrian Basin this episode is represented by the Gleisdorf Formation (Friebe, 1994), which comprises cyclic successions of silts, sands and oolites (Waltra Member), and marly limestones (Löffelbach Member). Alluvial fan sediments (basal parts of the "Puch Gravels") and limnic-fluvial, partly coal-bearing deposits in the Bay of Weiz ("Lower coal-bearing Beds of Weiz") and north of Graz are doubtfully assigned to the upper Sarmatian (Flügel, 1975; Moser, 1986; Krainer, 1987a).

1.3.1. Late Miocene

1.3.2. Pannonian

The complete restriction of the Central Paratethys around the Middle/Late Miocene boundary gave rise to the Lake Pannon, which covered an area of c. 290,000 km^2 by a maximum water depth of ca. 800 m at its maximum extent (ca.10–9 Ma; Kázmér, 1990; Magyar et al., 1999; Rögl, 1999; Harzhauser and Piller, 2007; Harzhauser and Mandic, 2008). However, it remains controversial if a glacio-eustatic sea level fall or tectonic uplift of the Carpathians caused the isolation of Lake Pannon (Lirer et al., 2009; Vasiliev et al., 2010). In the early phase (Fig. 3E) the lake water was brackish, slightly alkaline and slowly freshening due to its marine origin (Harzhauser et al., 2007). Influenced by the dry latest Middle Miocene climate the lake initially represented a meromictic system, but switched into a monomicic one during the middle Pannonian due to increasing precipitation (Harzhauser et al., 2007; Böhme et al., 2008, 2011). Astronomically forced climatic changes have modulated this development (Juhász et al., 1997; Sacchi and Müller, 2004; Jiménez-Moreno et al., 2005; Harzhauser et al., 2007, 2008; Lirer et al., 2009). A few highly euryhaline mollusc (Dreissenidae, Lymnocardiidae) and ostracod (Cytherideidae, Hemicytheridae) groups managed to survive the radical environmental change from Central Paratethys to Lake Pannon. These faunal relics and new freshwater immigrants gave rise to endemic lineages, which are used for regional biostratigraphic zonation (Papp, 1951; Kollmann, 1960; Daxner-Höck, 1996; Müller et al., 1999; Gross, 2000).

The Pannonian sedimentary succession starts with coarse-clastics with some coaly interbeds, ("Mühldorf Gravel", "Lignites of Feldbach", "Sandy bed with *Melanopsis impressa*"), which are discordantly overlying Sarmatian deposits (Stiny, 1918; Winkler, 1927c; Winkler-Hermaden and Rittler, 1949). Above follow limnic-brackish muds ("*Congeria* Marls", "Ostracod Marls"; Eisengraben Member) and limnic-deltaic, mud-sand-alternations with coal seams (Sieglegg Member) of the Feldbach Formation (Gross, 2000, 2003). Only at the northern basin margin alluvial ("Puch Gravel") and limnic-fluvial sedimentation continued ("Upper coal-bearing bed of Weiz"; Flügel, 1975; Krainer, 1987a; Fig. 3E).

A regression phase in the upper Lower Pannonian caused erosion and initiated a predominately fluvial sedimentation regime. Alluvial fans developed close to the northern basin margin ("Puch Gravel") and passed into braided and meandering rivers (Paldau Formation) ending in deltaic environments of the south-eastern Styrian Basin (Winkler, 1927c; Krainer, 1987a, b; Gross, 1998a). Ostracod and mollusc faunas in the lower Paldau Formation as well as a terrestrial flora, which differs from the azonal vegetation of meandering rivers, document a short-term ingression of the Pannonian Lake in the Styrian Basin (Kovar-Eder and Krainer, 1990, 1991; Gross, 1998b, 2000).

The middle Pannonian is represented by coal-bearing alternations of mud, sand and gravel ("Beds of Loipersdorf and Unterlamm", "Beds of Stegersbach" (Sauerzopf, 1952; Kollmann, 1965), while coarse-clastics ("Tabor Gravel", "Gravels of the Millstone Quarry") and associated fine-siliciclastics and sands ("Beds of Jennersdorf") are doubtfully assigned to the upper Pannonian (Winkler, 1927b; Kollmann, 1965). Late Pannonian fissure and cave fillings as well as gastropod-bearing freshwater opals are noticed from the South Burgenland Swell in the area of Eisenberg (Kümel, 1957; Bachmayer and Zapfe, 1969). These are the youngest Miocene deposits known in the Styrian Basin. Subsequent basin inversion caused considerable erosion.

1.4. Pliocene and Quaternary

A phase of basaltic volcanism started during the Pliocene and continued until the Early Pleistocene (Balogh et al., 1994). Aside from lava extrusions (Klöch, Stradner Kogel), which locally covers "Prebasaltic Gravels" (Winkler-Hermaden, 1957), and intrusions (Steinberg, Stein), phreatomagmatic volcanisms produced pyroclastic rocks and formed diatrems, which became filled with fine-clastic maar lake deposits (Burgfeld/Fehring; Pöschl, 1991; Fritz, 1996).

Fluvial gravels ("Postbasaltic Gravels") and residual soils partly cover these volcanic rocks and are interpreted as preglacial deposits. Quaternary erosion formed terraces, alluvial cones and landslides shaping the present-day landscape of southern Styria (Winkler-Hermaden, 1957; Flügel and Neubauer, 1984; Ebner and Sachsenhofer, 1991).

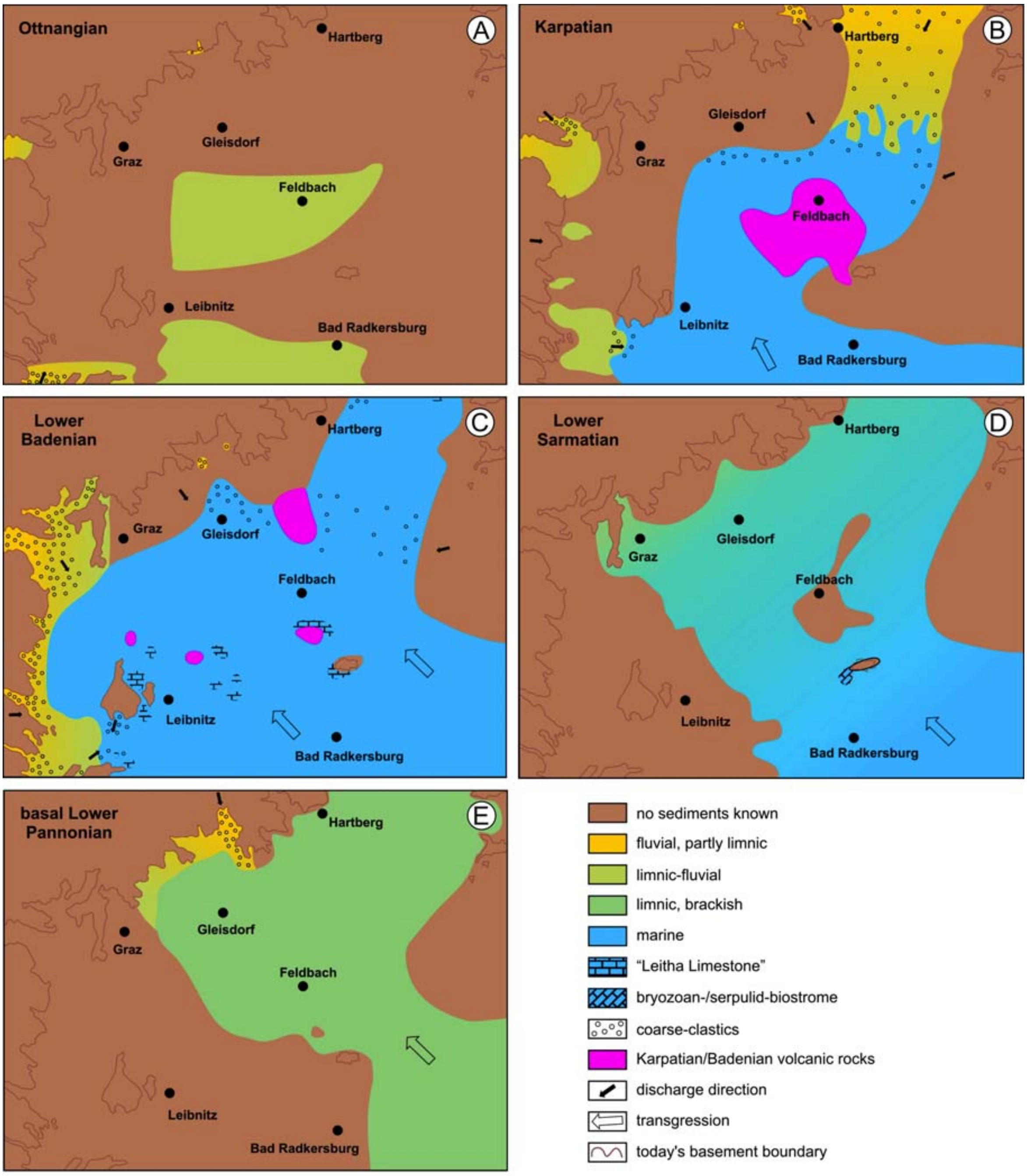

Figure 3: Palaeogeographic evolution of the Styrian Basin (Gross et al., 2007). (A) Ottnangian; (B) Karpatian; (C) Lower Badenian; (D) Lower Sarmatian; (E) basal Lower Pannonian.

Description of Stops

Site 1: Brickyard Wagna

Topic: Styrian Unconformity
Locality: Abandoned brickyard at Aflenz an der Sulm near Wagna; the brickyard is located immediately south of the bridge over the river Sulm; 15°32'50''E, 46°45'12''N
Lithostratigraphy: Kreuzkrumpel Formation ("Steirischer Schlier"), Weissenegg Formation
Biostratigraphy: Calcareous nanoplankton zones NN4 – NN5
Age: Middle Karpatian/late Early Miocene (Kreuzkrumpel Formation), Early Badenian/early Middle Miocene (Weissenegg Formation)

Description: The section (Fig. 4) is 80 m thick. The lower part of the section (ca. 60 m) exposes dark-grey, silty shales of the Kreuzkrumpel Formation, which are dipping 20–25° towards SE. Thin beds of turbiditic sandstone are intercalated in the lower part of this succession. In the upper part, a small channel with rounded crystalline pebble is incised into shales. The fine-siliciclastics of the Kreuzkrumpel Formation are terminated by an erosional unconformity covered a 2 m thick deposit of marls and silts with lithoclasts composed of "Steirischer Schlier" and crystalline pebbles ("Geröllmergel"). Above the erosive surface the dip angle changes from 20° to 5° (Fig. 4). The "Geröllmergel" is overlain by mixed siliciclastic-carbonate deposits of the Weissenegg Formation, which include further unconformities. A more detailed description of the section is given by Latal and Piller (2003), Gross et al. (2007), Spezzaferri et al. (2009), and Hohenegger et al. (2009, 2014).
Microfaunas (foraminifera) and -floras (calcareous nannoplankton, dinnoflagellate cysts) of Brickyard Wagna were studied by Rögl et al. (2002), Spezzaferri et al. (2002, 2004, 2009), and Solimann and Piller (2007).

Interpretation: For the Kreuzkrumpel Formation a Karpatian age is indicated by the presence of the planktonic foraminifera *Globigerina ottnangiensis* and *Globigerinoides bisphericus* as well as the benthic foraminifera *Uvigerina graciliformis* and *Pappina primiformis* (Spezzaferri et al., 2004; Hohenegger et al., 2009). The benthos/plankton ratio suggests inner shelf environments (maximum water depth 50 m) what is, however, inconsistent with specific taxa like *Spirorutilis carinatus*, *Budashevaella* spp., *Gaudryinopsis beregoviensis*, *Karrerulina* spp., *Bathysiphon* spp. indicating depths between 200 and 350 m (Spezzaferri et al., 2004, Hohenegger, 2005). One explanation for this discrepance may be a depletion of planktonic taxa due to certain oceanographic conditions (e.g., carbonate undersaturation, corrosive bottom waters; Spezzaferri et al., 2002, 2004). Planktonic foraminifera and calcareous nannoplankton also indicate a cool climate as well as a high productivity of the surface waters, which may have been related to the increased volcanic activity during the Karpatian, and (Spezzaferri et al., 2004, 2009). Consistently, the dominating heterotrophic taxa of dinoflagellates (*Lejeunecysta*, *Selenopemphix* and *Sumatradinium*) point to nutrient rich waters (Solimann and Piller, 2007).
By the occurrence of *Praeorbulina sicana*, *P. glomerosa,* and *Orbulina suturalis* an Early Badenian age is indicated for the succession above the angular unconformity (Hohenegger et al., 2009, 2014). This surface as well as the directly overlying "Geröllmergel" documents an erosional hiatus between the Karpatian and Badenian sediments, which has been related to the "Styrian Tectonic Phase" (indicated by tilting of the Karpartian sediments) and the global sea level fall at the Lower/Middle Miocene boundary. According to Rögl et al. (2007)

this hiatus spans a time interval of about 400 ka. During the Early Badenian, shallow water conditions established. Shallow marine benthic foraminifera faunas from siliciclastic facies above the Styrian Unconformity are dominated by low salinity tolerant species (i.e., *Ammonia* spp., *Elphidium* spp.) pointing to episodes of increased riverine input (Spezzaferri et al., 2009), while more normal marine conditions led to shallow marine carbonate deposition. The occurrence of reef corals in the carbonate facies further reflects warming of mid-latitudes during the Middle Miocene Climate Optimum.

Figure 4: Field aspect of the Styrian Unconformity at Wagna brickyard locality.

Site 2: Quarry Retznei

Topic:	Shallow marine carbonates of the Weissenegg Formation
Locality:	“Old Quarry” and “Quarry Rosenberg” (Larfage-Perlmooser Concrete AG) in the area of Retznei near Ehrenhausen; 15°33’34.9’’E, 46°44’41N
Lithostratigraphy:	Weissenegg Formation
Biostratigraphy:	Regional foraminiferal zones: Lagenidae Zone; calcareous nanoplankton zone NN5
Age:	Early Badenian; a tuff horizon on top of the carbonates was dated to 14.39 ± 0.12 Ma

Description: Rosenberg quarry exposes a ca. 25 m thick carbonate complex of clay-rich coral- and coralline algal-limestones over 600 m in the NW–SE and 200 m in the NE–SW direction (Reuter et al., 2012). It rests on siltstones and fine-grained sandstones with intercalated conglomeratic channel fills representing the “Geröllmergel” (e.g., Kollmann, 1965; Friebe, 1988). The upper surface of the “Geröllmergel” (level A) exhibits an erosive relief of 6.5 m altitude difference throughout the outcrop. The above following carbonate succession is heterogeneous, vertically as well as laterally. The limestones are generally impure and interrupted by discontinuity surfaces (levels A, B, F) and distinct tuffite layers (levels C, D, E), which are excellent correlation horizons (Reuter and Piller, 2011; Reuter et al., 2012). According to these reference levels, strata geometries (Fig. 5) and the dominant skeletal and non-skeletal components 4 depositional units were defined for the limestones. The distribution of facies is illustrated in Fig. 6.

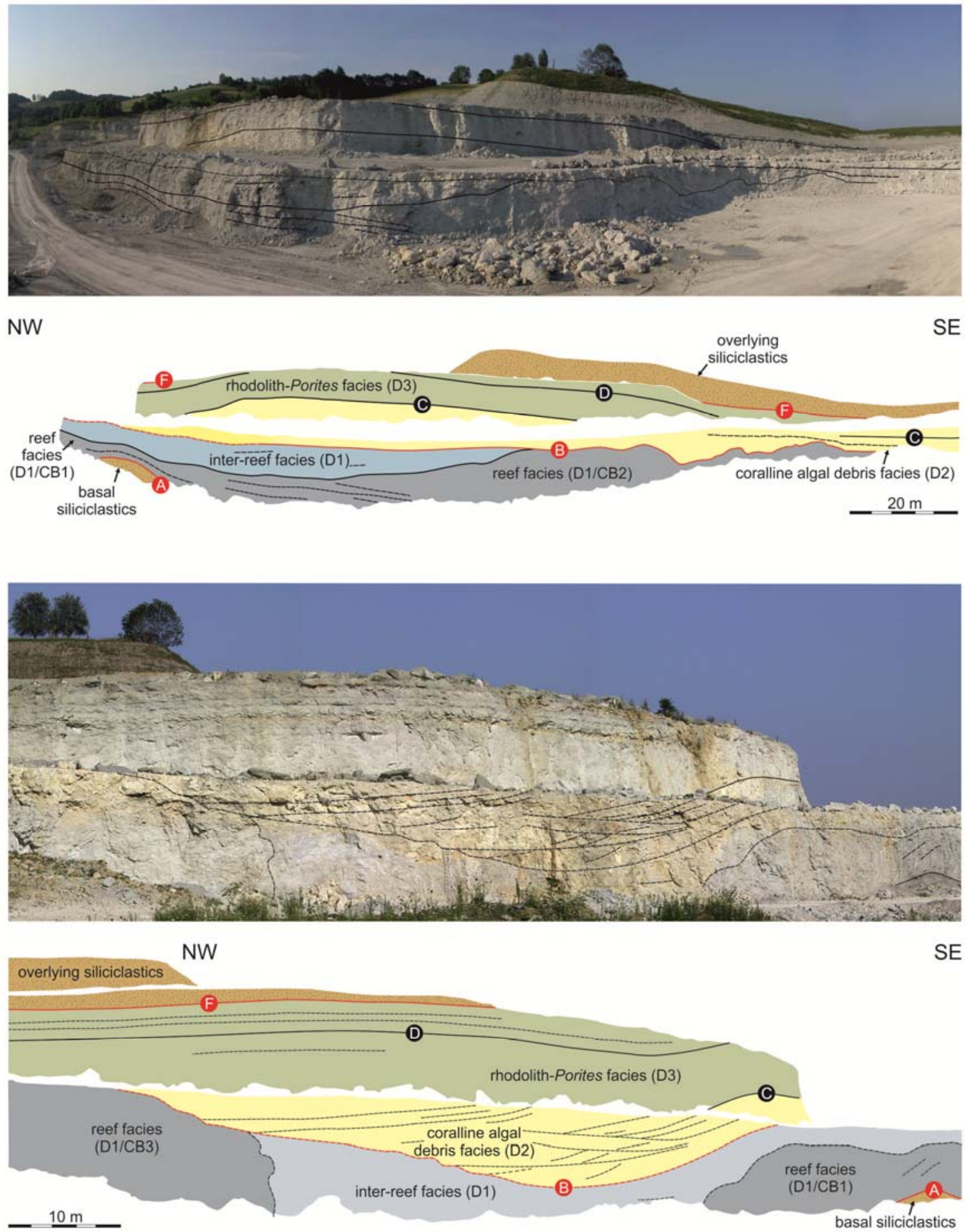

Figure 5: Field photographs and line diagrams demonstrating facies distributions and geometric relationships in the Rosenberg quarry; D1–4 = depositional units.

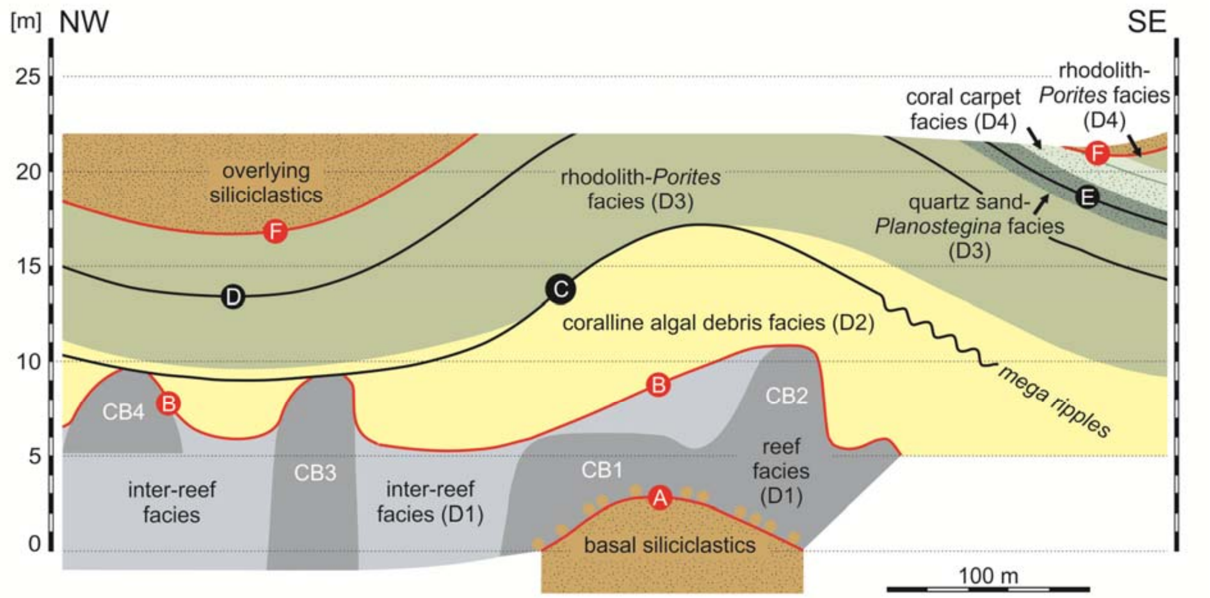

Figure 6: Schematic sketch summarizing the facies architecture of the Retznei carbonate body (Reuter et al., 2012); D1–4 = depositional units.

Depositional unit 1

The first depositional unit is characterized by 4 coral buildups (CB1–4). They range in lateral extent from 30 to 100 m, achieve heights of up to 9 m. These buildups are formed of decimetre-sized (up to 1 m), massive corals, which are in life position and in lateral contact (framestone). Two taxa dominate the coral fauna, out of which *Tarbellastraea reussiana* is the most abundant one, followed by *Porites*. *Montastraea* and *Mussismilia* contribute with minor amounts to the frameworks as well as branching in situ *Porites* colonies of 30 cm height. The original topography of the patch reefs is reflected by intercalated marl layers, which dip toward the buildup's margins and trace their outer shape. Typically they are discontinuous in the center of the buildups. Rarely, isolated reddish-brown weathered pyroclasts up to 3 cm long were also found within the coral frameworks. Characteristically for many coral colonies their surface is black stained (Fig. 7E) and intensively bored by bivalves and clionid sponges as well as encrusted by crustose coralline algae and balanids.

The coral buildups are surrounded by coarse-grained coralline algal-dominated skeletal limestones (pack-, float-, rudstones) with variable quantities of coral debris, rhodoliths, bryozoan colonies, foraminifers, and mollusks. A laterally symmetrical succession of biotic associations was observed in the depression between CB1 and CB3. These coral buildups are rimmed by rhodolith floatstones and rudstones. The transition from the reef facies to the rhodolith belt is gradual and takes place on less than 1 m distance. The rhodoliths become laterally replaced by celleporiform bryozoan nodules and then grade into a *Planostegina* facies. The latter is characterized by large (1–3 cm) individuals of *Planostegina giganteoformis*, which occurs locally in rock-forming quantities. Characteristically, many skeletal components of the inter-reef facies between CB1 and CB3 are stained black. Depositional unit 1 is terminated by a erosive surface (level B).

Depositional unit 2

The second depositional unit onlaps against level B. It comprises well sorted and winnowed cross-bedded coralline algal debris grainstones with large foresets. A distinct surface (level

C) is intercalated with the coralline algal debris facies, following the topographic high formed by patch reefs CB1 and CB2, and correlating with a megaripple field in the eastern part of the outcrop (Fig. 7D). This surface is covered with a few centimeters of soft, dark gray to gray-greenish marl with idiomorphic biotite crystals. Immediately above the marl of level C clusters of oysters and Isognomon occur, as well as abundant *Clypeaster campanulatus* coronas and in situ *Pinna*.

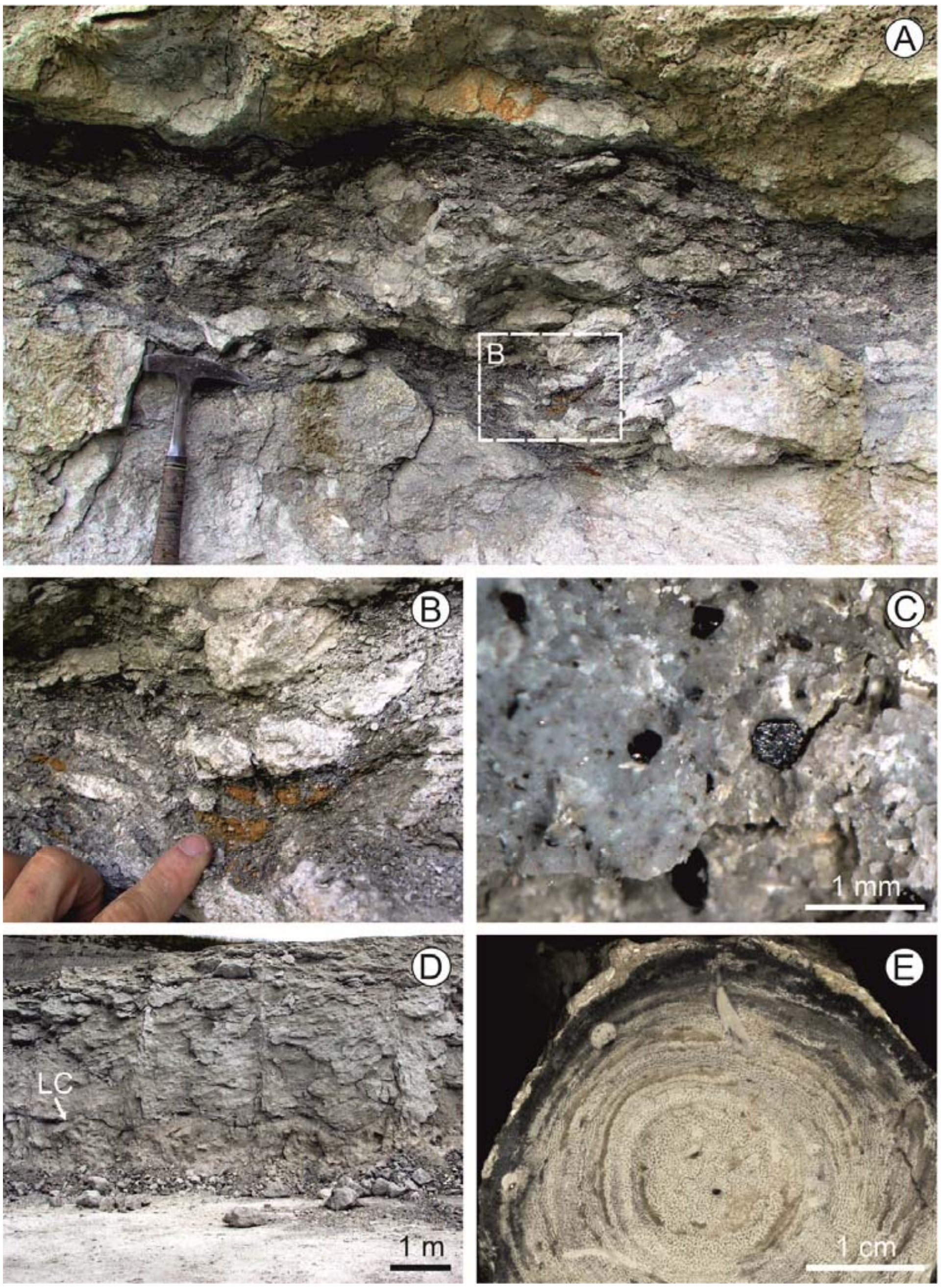

Figure 7: Volcanic ash deposits (Reuter and Piller, 2011). (A) Argillaceous tuffite intercalated within the rhodolith-*Porites* facies (level D, the white box locates Fig. 8B); (B–C) Evidence for a pyroclastic origin are friable, red oxidized pyroclasts (B) and idiomorphic biotite crystals (C); (D) Megaripple field buried beneath the tuffite of level C (LC). The detailed preserved topography points to event sedimentation; (E) Columnar *Porites* branch from the margin of patch reef CB1. This coral branch was found in life position and exhibits a circumferential black-stained rim due to the infiltration of fine volcanic ash at the surface

Depositional unit 3

The coralline algal debris facies is topped by an 8.5 m thick unit of thick-bedded marly limestones characterized by variable amounts of rhodoliths and non-framework forming platy *Porites* colonies. Bedding is caused by increasing clay content towards the top of each bed. A between 5 cm and 40 cm thick, distinct layer of soft, dark gray to gray-greenish marl with a sharp irregular lower surface is intercalated with the rhodolith-*Porites* facies (level D). This fine-grained deposit contains isolated idiomorphic biotites (Fig. 7C) as well as up to 15-cm long dark gray to greenish-gray friable volcaniclasts with idiomorphic biotite crystals and oxidized pyroclasts of reddish-brown color (Fig. 7A, B). Locally, the pyroclasts make up ca. 30% of the sediment. Similar to level C, patches of in situ oysters and frequent *Clypeaster campanulatus* coronas are found on the upper surface of level D.

The rhodolith-*Porites* facies grades upsection gradually into well-sorted, bioclastic coralline algal-*Planostegina* limestones with quartz sand. Many bioclasts are stained black. A distinct 15-cm-thick smeary marl horizon with sharp irregular bottom surface (level E) is intercalated in the quartz sand-*Planostegina* facies. Similar to the marl deposits of levels C and D, it contains idiomorphic biotite platelets and large (10 cm) dark gray to greenish-gray biotite-rich pyroclasts. Directly above level E, the amount of *Planostegina* debris increases and a concentration of *Clypeaster campanulatus* coronas occurs.

Depositional unit 4

This depositional unit starts with a 2 m thick succession of two coral carpets composed of phaceloid corals (lower carpet) and flat plate-like *Leptoseris* (upper carpet; Fig. 8). The coral carpet facies is covered by a ca. 5 m thick unit of argillaceous rhodolith limestone with scattered platy *Porites* (rhodolith-*Porites* facies). An erosive surface (level F) truncates the carbonate succession.

Above follows a 35 m thick unit of sandstones and siltstones. Two pyroclastic layers containing idiomorphic biotites and zircons, unaltered feldspar phenocrysts, and bentonites are interbedded within the siliciclastics (Hauser 1951; Bojar et al. 2004; Handler et al. 2006; Hohenegger et al. 2009).

Figure 8: A *Leptoseris* carpet from Retznei is the first evidence of this coral in the Central Paratethys. (A) Rudstone composed of thin *Leptoseris* plates. (B) Thin section of in situ corals. Bryozoan encrustations at the undersurface document their elevation above the seafloor

Interpretation: The carbonate development starts above coarse-grained siliciclastics of the "Geröllmergel", which were deposited in a fluvial-marine channel system above the Styrian Unconformity during a relative sea level lowstand (Friebe, 1990, 1993; Fenninger and Hubmann, 1997). Depositional unit 1 comprises patch reefs (reef facies) and flanking carbonate sands (inter-reef facies) that developed during a relative sea level rise. Remarkable is the lateral transition from the reef facies to the inter-reef facies. It is characterized by a succession of biotic associations: reef corals–rhodoliths–nodular celleporiform bryozoans–*Planostegina*. Similar rhodolith-rimmed patch reefs are found in the present-day Safaga Bay (Red Sea) at the transition from patch reefs to seagrass meadows (Piller and Rasser, 1996). Depositional unit 1 is terminated by a karst surface. The above following depositional unit 2 is represented by large-scale cross-bedded coralline algal sands (coralline algal debris facies) that suggest a submarine dune environment at the beginning of the next transgression. The vertical succession of the rhodolith-*Porites* facies to quartz sand-*Planostegina* facies in depositional unit 3 is interpreted as a deepening–shallowing trend. Renewed deepening is displayed as a transition from the coral carpets to the rhodolith-*Porites* facies in depositional unit 4. The carbonate succession ends with a karst surface. For a detailed facies interpretation the reader is referred to Reuter and Piller (2011) and Reuter et al. (2012). Subsequent suffocation by siliciclastics and drowning of the Retznei carbonate complex are documented for the overlying siliciclastic succession (Friebe, 1993; Gross et al., 2007; Hohenegger et al., 2009; Strahlhofer). Its biotic assemblages indicate always normal marine conditions (Gross et al., 2007; Hohenegger et al., 2009). Abundant plant remains point to a close hinterland, which acted as permanent source for siliciclastic supply. A generally increasing water depth can be reconstructed based on the dinocyst and benthic foraminiferal assemblages (Friebe, 1993; Gross et al., 2007; Hohenegger et al., 2009; Strahlhofer, 2013). Intercalated turbidites and slumps show a distinct topography. Synsedimentary volcanic activity is documented by tuff layers (Bojar et al., 2004; Handler et al., 2006). This shift from carbonate to siliciclastic sedimentation is interpreted as effect of accelerated basin subsidence and hinterland uplift owing to intensified tectonic activity (Friebe, 1993).

Basically, carbonate production was strongly influenced by terrigenous siliciclastic discharge (Friebe, 1990). Coarse-grained terrigenous fraction (>silt) was related to lowstands of relative sea level. Additionally, short-term disturbances of the shallow-marine carbonate factory were caused by volcaniclastic sedimentation events. These events produced distinct tuffite layers that mantle the former seafloor topography (Fig. 7D) and are characterized by the occurrence of idiomorphic biotite crystals (Fg. 7C), volcaniclasts (Fig. 7A, B) and bioclasts in argillaceous matrix (levels C, D, E; Reuter and Piller, 2011). Scattered black stained bioclasts in the sediments and black stained coral surfaces (Fig. 7E) in the patch reef facies point to further, probably thinner, volcaniclastic deposits, which were completely reworked soon after deposition. This shows that eruptive events must have occurred with higher frequency in the Middle Miocene Styrian Basin than the preserved volcanic ashbeds suggest and the submarine alteration of volcanic ashes must have been a permanent source for clay minerals in the isolated inner basin setting (Reuter and Piller, 2011).

Several sediment stress conditions caused species turnover in marginal coral reef communities, which existed in close proximity to their environmental limits, resulting in a unique succession of various low diverse coral assemblages (coral patch reefs, coral carpets, non-framework forming coral communities; Reuter et al., 2012). This finding reveals that the comparably high coral diversity on the Middle Styrian High (>10 coral genera) paradoxically rather result from local stress factors that adversely affect corals than to reflect a site with ideal living conditions. Important constraints for coral growth fabrics and faunal compositions were the amount and type of siliciclastic sediment supply and the water movement by washing out the fines and bringing them into suspension toward deeper water. The shifts of coral communities due to increasing siltation stress show the following general trends: 1. Decrease of coral diversity; 2. Replacement of suprastratal by constratal growth fabrics; 3. Replacement of massive by platy growth forms; and 4. Reduction of coral cover and colony sizes (Reuter et al., 2012).

Site 3: Clay pit Mataschen

Topic:	Limnic-deltaic sediments of Lake Pannonian
Locality:	Clay Pit Mataschen of the Lias Österreich Gmbh; 5.3 km SW Fehring; 15°57'29''E, 46°54'17''N
Lithostratigraphy:	Feldbach Formation (Eisengraben and Sieglegg members).
Biostratigraphy:	Regional mollusk zonation: *Mytilopsis ornithopsis*/*Melanopsis impressa* zone
Age:	Lower Pannonian (11.308–11.263 Ma)

Description: The 30 m thick section (Gross et al., 2011; Fig. 9) starts with a >1.5 m thick succession of laminated sandy clays and partly ripple-bedded fine-medium sands (Gross, 2004a). The top of these unit 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 (Fig. 10A) are regularly found at distances of c. 10–15 m throughout the outcrop (ca. 700×200 m). From 0.0 to 0.3 m the section is composed of densely packed, coaly plant fragments with clayey interlayers. They contain a low diverse, azonal plant assemblage and scattered vertebrate remains (beavers, dwarf hamster, pond turtles; Daxner-Höck, 2004; Gross, 2004b; Kovar-Eder, 2004; Meller and Hofmann, 2004). Upsection (0.3–0.8 m) follows a bed 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 (ca. 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. Up to these sandy beds lymnocardiid bivalves are present (frequently found in "butterfly" preservation). Fish skeletons occur associated between ca. 1.5–3 m; articulated specimens of the large dreissenid bivalve *Mytilopsis ornithopsis* were found rarely between ca. 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. Sandy beds are often rich in plant detritus and occasionally enclose diaspores (Meller and Hofmann, 2004). Close to the top (ca. 26.5 m), fine sandy silt layers yielded a highly diverse macroflora (Kovar-Eder and Hably, 2006). The top of the section 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.

Interpretation: By intergrating geophysical (gammay ray, magnetic susceptibility), geochemical (organic carbon, sulphur), sedimentological and palaeontological (mainly ostracods) data from the Mataschen section Gross et al. (2011) reconstructed 4 stages of lake evolution:

Stage 1 — before the rise of Lake Pannon (−1.5–0.0 m; duration: unknown)
The basal, sandy–silty layers were deposited in a fluvio-lacustrine freshwater environment as indicated by freshwater ostracods and lithology. Exploration drillings also document gravels and thin coal seams a few metres below, which indicate a highly variable wetland (Gross, 2004a).

Stage 2 — development of a taxodiacean-swamp (0.0–0.3m; duration: hundreds of years)
A rising groundwater table, which antedates a transgression of Lake Pannon, caused the establishment of a *Glyptostrobus*-swamp for a few centuries (Fig. 10C). Aside from the conifers, 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). Based on the palaeofloristic composition warm temperate to almost subtropical conditions are suggested.

Stage 3 — drowning of the swamp (0.3–0.8 m; duration: hundreds of years)
The swamp forest drowned within a few decades (>1 m water depth; Fig. 10B). Abundant plant remains (e.g., leaf litter, twigs), rare unionid bivalves, ostracods, insects, cyprinid fishes, amphibians and bird fossils indicate a close-by lakeside and almost freshwater conditions. The dreissenid bivalve *Mytilopsis neumayri*, which is mass-occurring between ca. 0.4 and 0.80 m, probably dwelled byssally attached to the submerged tree trunks (Fig. 10B) and refers to salinities around 2–3.5 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 *M. 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). 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.

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., *Cyprideis*, 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 palaeontological evidences as well as decreasing TOC/TS-ratios indicate the influx of saline waters (ca. 18 PSU) related to a transgression of Lake Pannon.

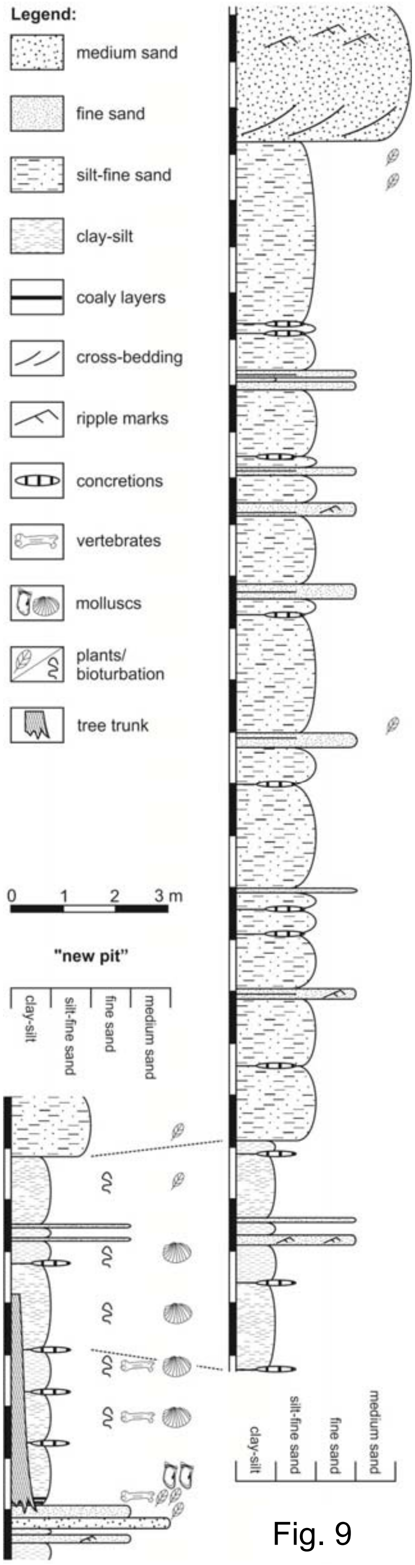

Fig. 9

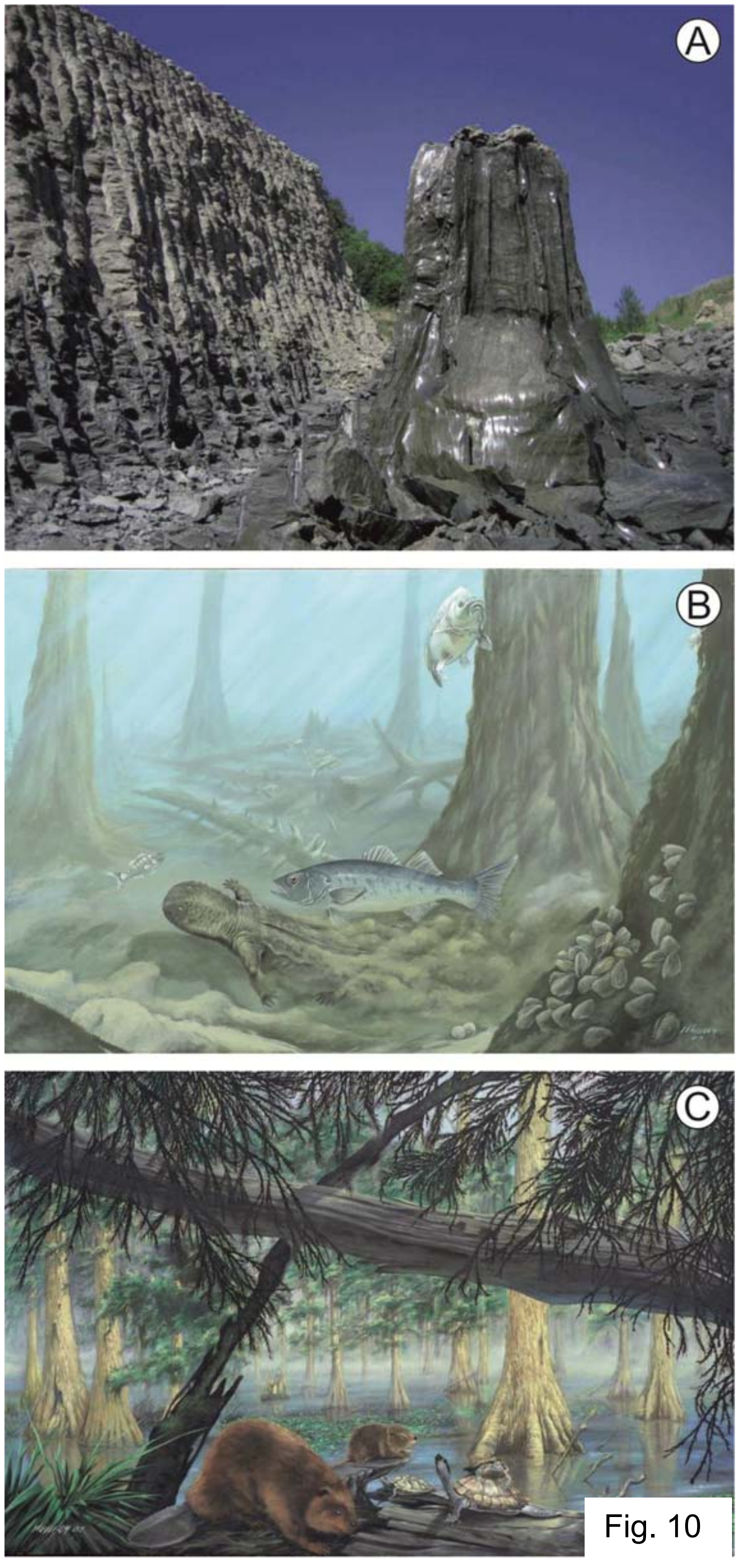

Fig. 10

Figure 9: Section Mataschen (Martin Gross).

Figure 10: Field aspect and paleoenvironmental reconstructions (postcards of the Universalmuseum Joanneum). (A) Fossil in situ tree trunk at the base of Mataschen section; (B) Stage 2: conifer swamp; (C) Stage 3: drowning of the swamp.

Overall, stage 4 is characterised by limited oxygenation of bottom waters. Rare benthic mollusc faunas, partly articulated fish skeletons and limited bioturbation support this assumption (Cziczer et al., 2008). Shallow burrowing, probably dysoxic-tolerant lymnocardiids are dominant and 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. The rare occurrence of intact carapaces suggests that ostracods avoided the low aerated sediment.
After a phase of highly fluctuating bottom water ventilation deepening of the environment (10–15 m water depth) established a meromictic system, which perturbed benthic life in the hypolimnion and favoured low susceptibility pyrite formation over greigite growth. Accordingly, the sediment became well laminated; ostracod- as well as bivalve-contents significantly declined. Vanishing of the trunk barrier (above 4 m in the section) due to burial or decay enabled coarser sediment (silt) to enter the system. Sand layers at 5.5 m and 6 m document occasional turbiditic events with hyperpycnical behaviour.

Stage 5 — delta progradation (7.5–30 m; duration: a few ten thousands of years)
The limnic phase is terminated by the progradation of a delta system. This is indicated the by the increasing silt content and higher abundance of turbiditic sands above 7.5 m in the section. Elevated TOC-values of the sand layers document the enhanced discharge of terrestrial plant material. Highly diverse, but fluvially transported leaf assemblages from layers close to the top (ca. 26.5 m) document the existence of nearby evergreen broad-leaved to mixed mesophytic forests (Kovar-Eder and Hably, 2006). Molluscs are completely missing in stage 5 and the ostracod content is notably reduced. Probably, the large amount of sand–silt in combination with a decrease in salinity disabled ostracods to colonize this environment.
In the uppermost part of the section (ca. 27–30 m), large scale cross-bedded and wave ripple-bedded, silty–sandy deposits indicate the change to a delta-front environment.

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