GUIDEBOOK

The inception, growth and demise of a pelagic carbonate platform: Jurassic and Lower Cretaceous of the Krížna Nappe in the Western Tatra Mountains

Guide to field trip A6 • 21–22 June 2015

Renata Jach, Jacek Grabowski, Andrzej Gaździcki, Alfred Uchman

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The inception, growth and demise of a pelagic carbonate platform: Jurassic and Lower Cretaceous of the Krížna Nappe in the Western Tatra Mountains

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Route (Figs. 1, 2): From Kraków we drive south by E77 to Rabka, then by road 958 to the village of Witów. In Chochołowska Valley (Dolina Chochołowska), to the exposures located near Huciska Glade (Polana Huciska). First-day hike starts from Huciska Glade, than leads through **Huciański Klin ridge** (**stops A6.1–A6.4**) and ends in **Lejowa Valley** (Dolina Lejowa; **stops A6.5, A6.6**). This all-day eastward traverse runs along unsigned trail through forest, with some fairly demanding hiking at altitudes 900–1300 m. The relatively short loop walk of the second day starts from Huciska Glade and leads up along **Długa Valley** (Dolina Długa) to the **Pośrednie ridge** (**stop A6.7**). The way back from the Pośrednie ridge is an easy downhill walk along **Kryta Valley** (Dolina Kryta; **stop A6.8.1–4**).

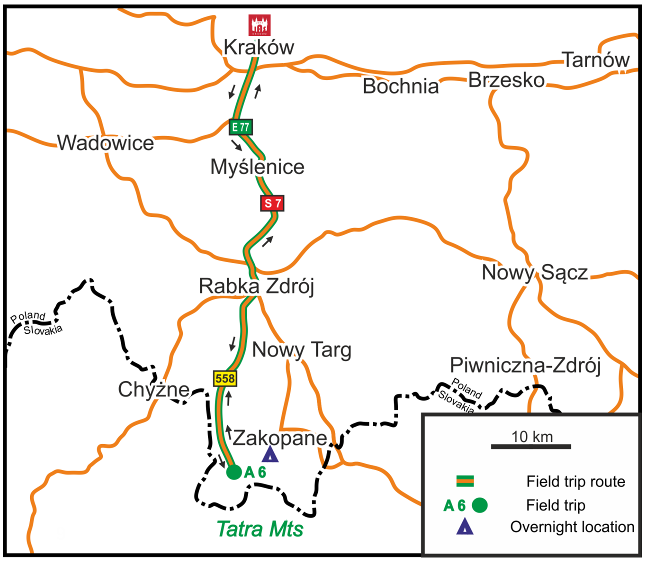

Fig. 1. Route map of field trip A6.

Introduction to the trip

The Tatra Mountains are the highest ridge of the Western Carpathians with the highest peak Gerlach (2655 m). The Tatra Mts were uplifted during Neogene and form a block which is asymmetrically tilted to the north and bounded from the south by a prominent fault (Fig. 3).

The block consists of a pre-Mesozoic crystalline core (granitoidic intrusions and older metamorphic rocks), ?Permian and Mesozoic autochthonous sedimentary cover, allochthonous Mesozoic sedimentary rocks, detached locally from their crystalline basement in form of nappes and smaller thrust sheets (Kotański, 1965,

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Guidebook is available online at www.ing.uj.edu.pl/ims2015

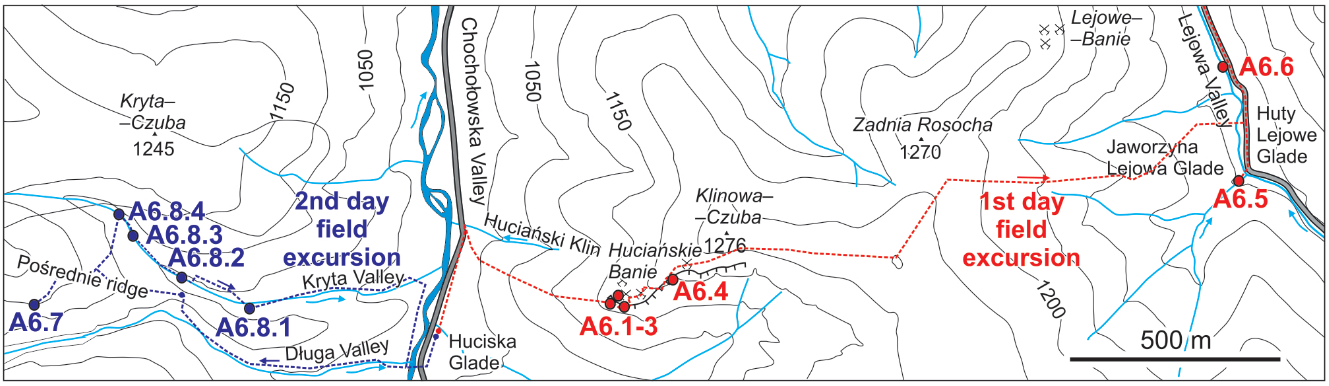

Fig. 2. Detailed itinerary of field trip A.6; A6.1–4 – Huciański Klin ridge; A6.5–6 – Huty Lejowe glade; stops:A6.7–8a–c – Długa Valley and Kryta Valley.

1971). The nappes were formed in Late Cretaceous. The Mesozoic sedimentary rocks are discordantly covered by rocks of the Central Carpathian Paleogene, which include Oligocene flysch, more than 2 000 m thick, in their upper part. The sedimentary rocks of the Tatra Mts dip generally to the north due to the Neogene uplift (Fig. 4; Bac-Moszaszwili *et al*., 1979). They are ascribed to three main tectonic-facies domains: Tatricum (High-Tatric autochthon and allochthon), Fatricum (Krížna Nappe = Lower Sub-Tatric Nappe) and Hronicum (Choč Nappe = Upper Sub-Tatric Nappe) on the basis of their characteristic facies successions and tectonic position.

The Krížna Nappe overlies the High-Tatric units and is covered by the Choč Nappe (Kotański, 1965). It comprises Lower Triassic to Lower Cretaceous deposits and is built of several thrust sheets and so called ‘partial nappes’. The Krížna Nappe belongs to the Fatricum Domain of the Central Carpathian block (Plašienka, 2012). During most of the Jurassic time, it was one of the domains located between the Alpine Tethys to the north and the Meliata Ocean to the south (Fig. 5; Thierry and Barrier, 2000; Schmid *et al*., 2008). As a consequence of its location, the succession studied displays a strong similarity to the Jurassic of other Tethyan basins.

During the Jurassic, the Fatricum Domain was bordered by the uplifted Tatricum Domain to the north and the Veporicum and Hronicum domains to the south (Csontos and Vörös, 2004; Plašienka, 2012). The Fatricum Domain was regarded as being an extensional basin during the Jurassic, located on thinned continental crust (e.g., Plašienka, 2012). As a result, the Jurassic successions of the Fatricum Domain are characterized by an almost

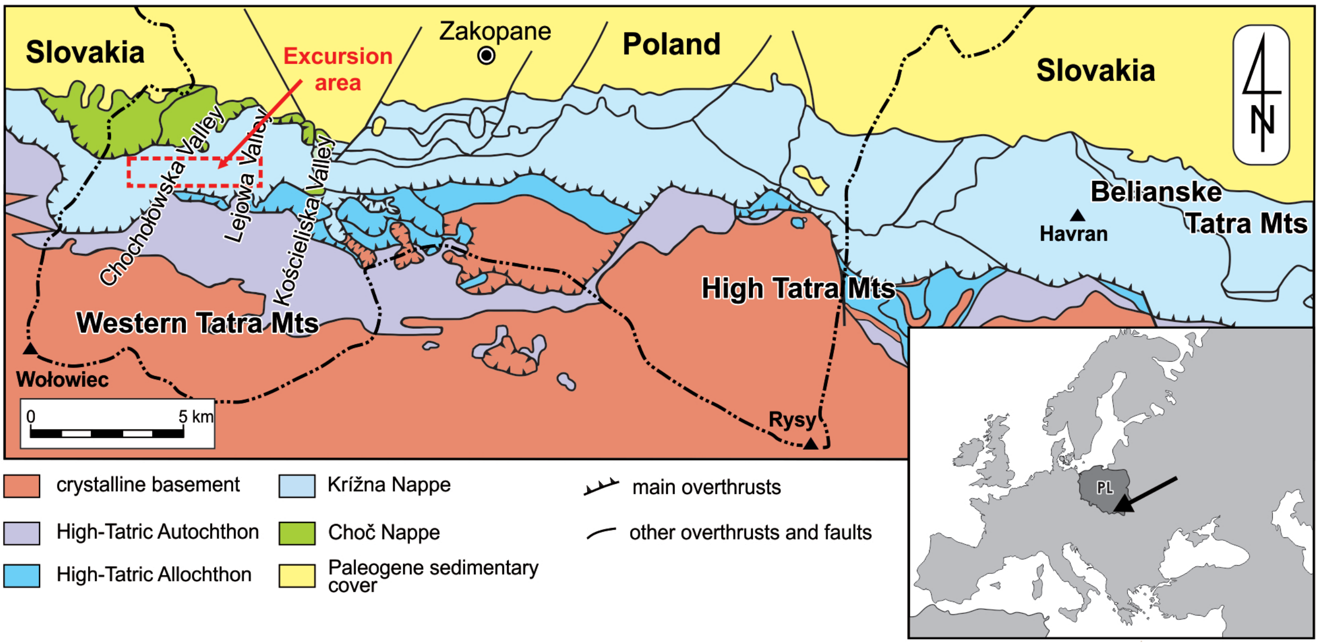

Fig. 3. Tectonic sketch map of the Tatra Mts showing location of the excursion area (after Bac-Moszaszwili *et al*,. 1979; modified)

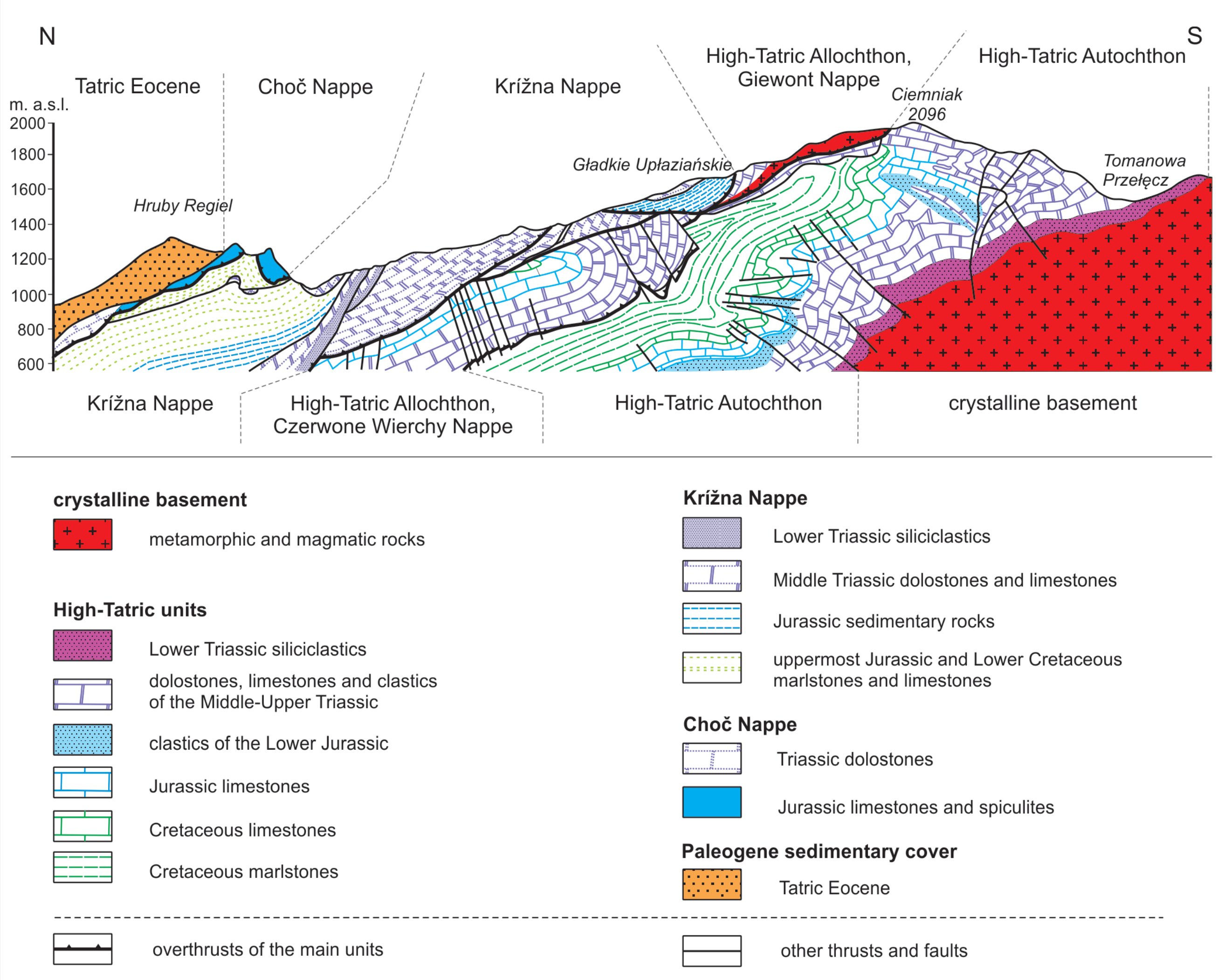

Fig. 4. Geological cross section through the Tatra Mts along the Dolina Kościeliska valley (after Kotański, 1965, changed)

continuous record of deepening, with a transition from littoral through hemipelagic to pelagic deep-sea sedimentation.

The Jurassic deposits of the Krížna Nappe in the Tatra Mts represent a generally deeper sea succession of the Zliechov type (Michalík, 2007; Plašienka, 2012; Jach *et al.*, 2014). The oldest Jurassic deposits are represented by dark shelf carbonates and shales, which refer to sea transgression onto emerged lands (Figs 6–8; Gaździcki *et al.*, 1979; Michalík, 2007). These Rhaetian–Hettangian deposits have diachronic lower and upper boundary. The Early Jurassic basin featured wide basins and narrow horsts, characteristic of the first phase of extension (Plašienka, 2012). The second phase of extension took place during the latest Early Jurassic and resulted in extensional tilting of blocks, which formed a horst-and-graben topography (Wieczorek, 1990; Gradziński *et al.*, 2004; Jach, 2005; Plašienka, 2012). The formation of horsts (e.g., the excursion area) and grabens (in the High Tatra and the Belianske Tatry Mountains) led to distinct facies changes (Guzik, 1959; Jach, 2007). The subsiding basins were filled with *Fleckenmergel*-type sediments (bioturbated "spotted" limestones and marls, ranging from Sinemurian to Bajocian; Wieczorek, 1995; Iwańczuk *et al.*, 2013). During the Late Pliensbachian–?Aalenian, the horsts acted as submarine highs with sedimentation of spiculites on their slopes (Jach, 2002a) and neritic crinoidal sedimentation (e.g., crinoidal tempestites; Jach, 2005), replaced by condensed pelagic carbonates (Gradziński *et al.*, 2004) that were deposited on a pelagic carbonate platform (*sensu* Santantonio, 1993). During the Middle Jurassic, significant topographic relief was still present and controlled facies distribution. The deposition of carbonate sediments terminated with the onset of a uniform radiolarite sedimentation during the Middle Jurassic (Jach *et al.*, 2012, 2014). Generally, the Lower–Upper Jurassic sediments of the Fatricum Domain record a gradual deepening and transition

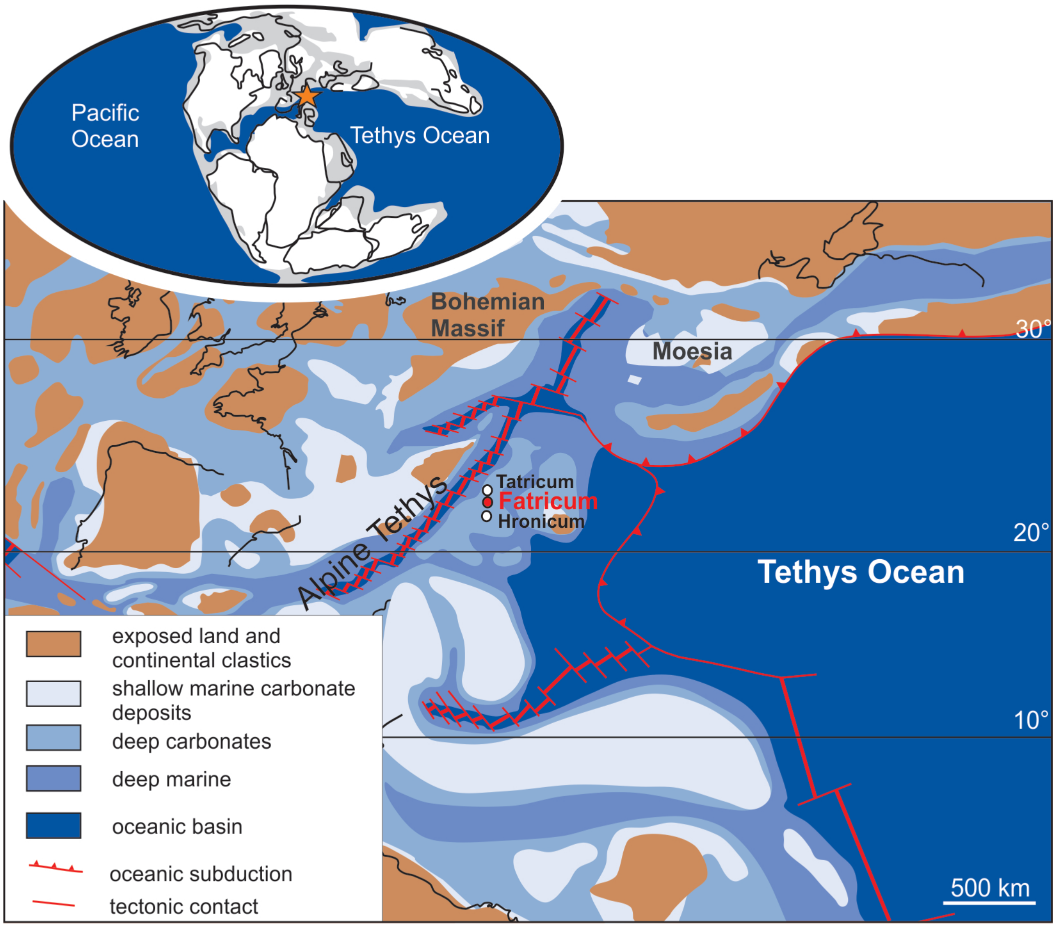

Fig. 5. General palaeogeographic position of the Fatricum Domain during the Callovian (after Thierry and Barrier, 2000; simplified)

from hemipelagic to pelagic deep sea environment. A major recovery of carbonate sedimentation started in the Late Tithonian and Early Berriasian, when the Maiolica-like deep sea limestones and marls were deposited (Pszczółkowski, 1996; Grabowski *et al*., 2013). Widespread marl and marly limestone sedimentation started in the Late Berriasian and continued to the Aptian, with intercalations of turbiditic sandstones and calcareous fluxoturbidites in the Valanginian and Hauterivian (Pszczółkowski, 2003a).

The main aim of this excursion is to present general depositional history of the studied part of the Krížna Unit and to discuss factors controlling their sedimentary environment.

Stop descriptions

Attention: All the localities are in the protected area of the Tatra National Park. Please do not hammer the rocks.

A6.1 Huciański Klin ridge – Late Pliensbachian spiculites

(49°15′37″ N, 19°49′17″ E)

Leader: Renata Jach

The section presented during the first part of the excursion is located on the forested southern slope of the Huciański Klin ridge. The oldest are 16–20 m thick spiculites which overlie basinal “spotted” limestones and marls of the Sinemurian–Lower Pliensbachian not exposed in the presented locality (about 150 m thick; see stop A6.6). Spiculites are well-exposed in crags a few metre high. Late Pliensbachian age is ascribed to the spiculites on the basis of their position in the section. They are dark, very hard and well-bedded, with bed thickness from a few to 30 cm. Siliceous sponge spicules, which belong to Hexactinellida and Demospongia (mainly Tetractinellida) with loose skeletons, are the major components of spiculites (Fig. 9A–C; Jach, 2002a). A gradual substitution of Hexactinellida by Demospongia is observed

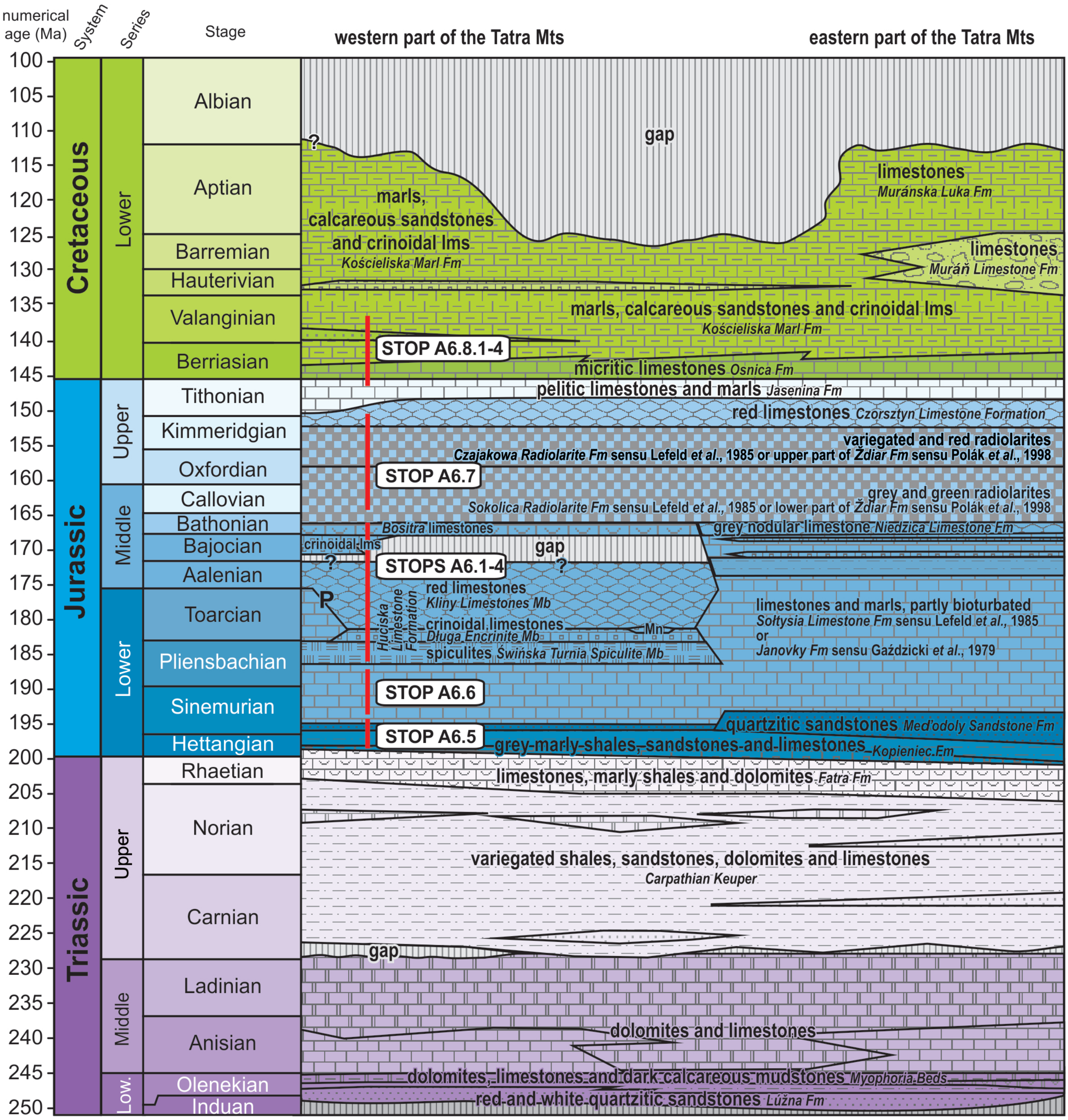

Fig. 6. Simplified litostratigraphic scheme of the Fatricum domain succession in the Tatra Mts (partly after Lefeld *et al.* (1985); modified)

upward the spiculite section. Crinoidal ossicles, benthic foraminifera and detrital quartz grains have also been found in very small quantities.

Spiculites are interbedded with some crinoidal wackestones, packstones and grainstones, which form beds up to 20 cm thick. Limestone beds thicken upwards and show a general trend of grain coarsening, accompanied by an increase in their textural maturity from wackestones to grainstones (Jach, 2005). Low-angle cross-bedding, normal grading and erosional bases observed in crinoidal interbeds indicate that they are laid down in higher energy conditions than the intervening spiculites. This is attributed to the currents generated by storms that were capable of sweeping crinoidal material from shallow, elevated parts of the basin to the area inhabited by siliceous sponges. As such, they represent event beds deposited below storm wave base on the slopes of an elevation. The gradual replacement of Hexactinellida by Demospongia and grain coarsening and thickening of crinoidal intercalations indicate a shallowing trend

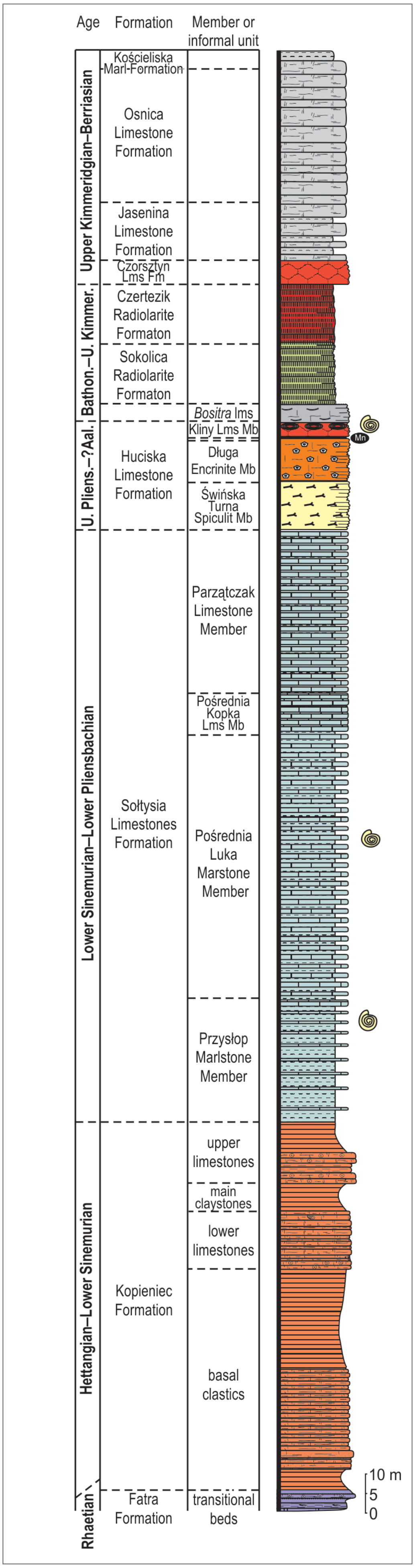

Fig. 7. Lithostratigraphic log of the Lower Jurassic–Early Cretaceous of the Krížna Nappe in the Western Tatra Mts (after Lefeld et al 1985; modified)

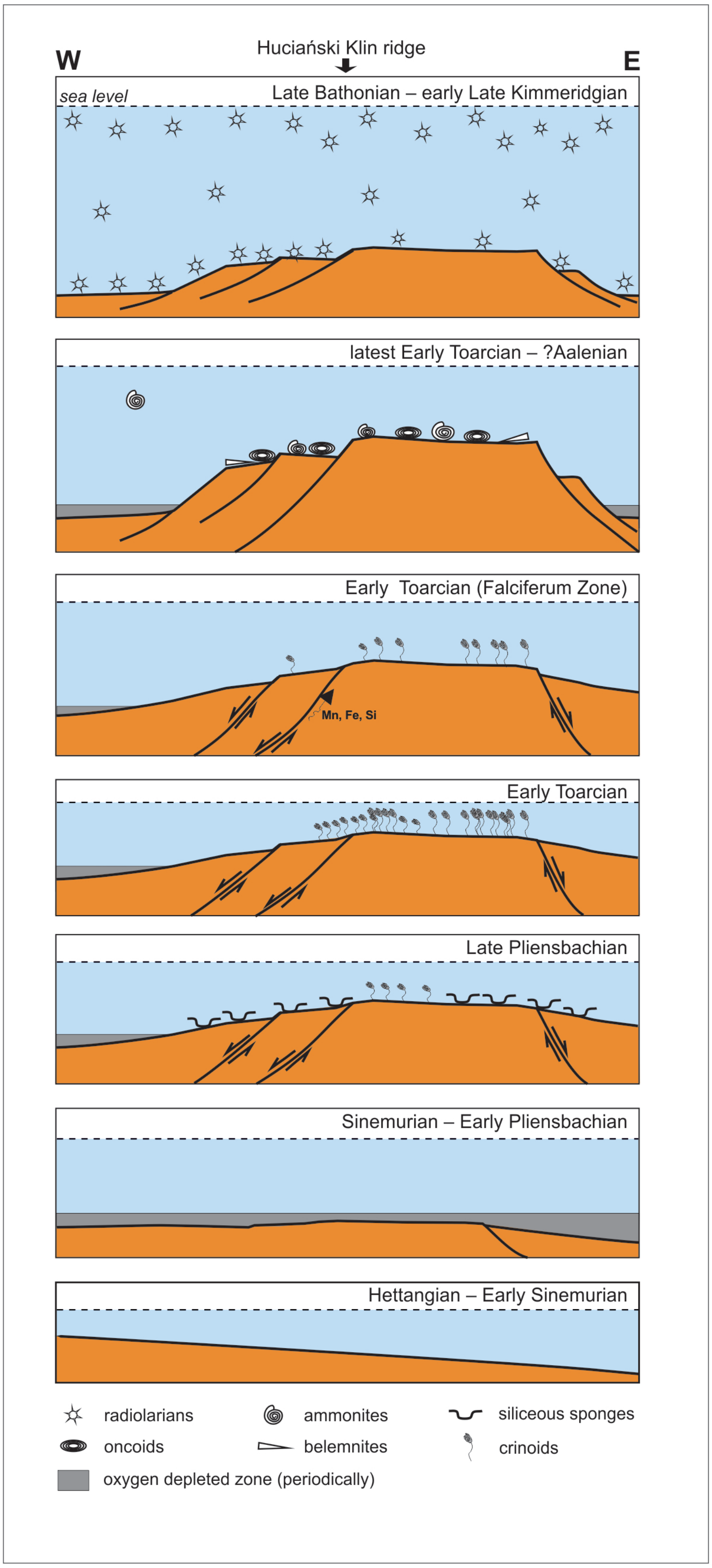

Fig. 8. Early–Middle Jurassic evolution of the studied part of the Fatricum Basin (after Jach, 2003; modified)

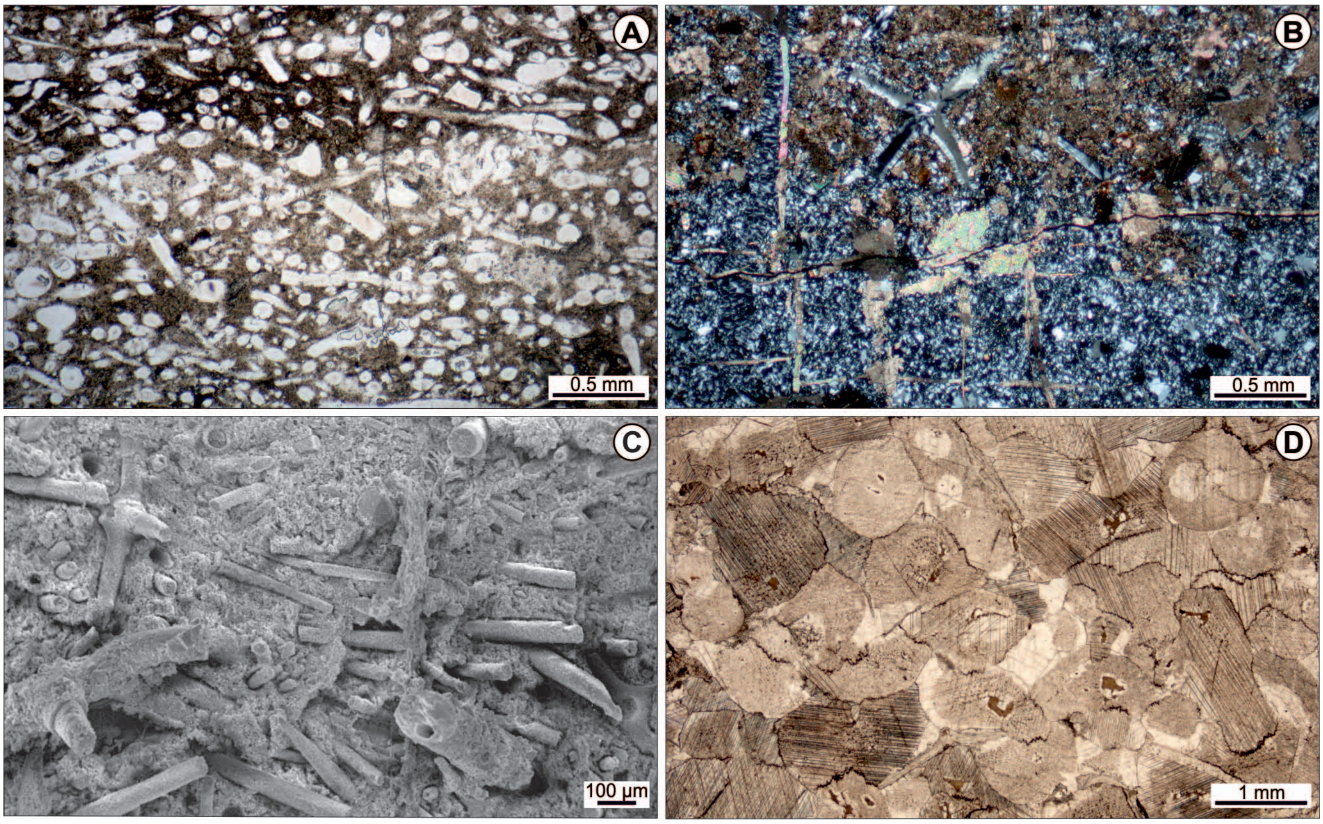

Fig. 9. Spiculites and crinoidal limestones. **A)** Spiculite in thin section. Plane-polarised light; **B)** Spiculite in thin section. Cross-polarised light; **C)** Densely packed siliceous sponge spicules. SEM image of HF-etched surface; **D)** Crinoidal grainstone with syntaxial cements. Thin section, plane-polarised light.

during spiculite deposition (Jach, 2002a). Most probably this trend was related to local changes of seafloor topography caused by synsedimentary faulting in the Fatricum Basin (Jach, 2005). This notion is supported by the occurrence of packages displaying chaotic bedding within the spiculites in the upper part of the spiculite succession. They are interpreted as submarine slumps.

A6.2 Huciański Klin ridge – Early Toarcian crinoidal limestones

(49°15′36″ N, 19°49′21″ E)

Leader: Renata Jach

White, grey and slightly pinkish crinoidal limestones (grainstones) overlie spiculites. They crop out directly over spiculites in rock crags. The crinoidal grainstones, about 12 m in thickness, display irregular bedding with subtle cross-bedding, graded bedding and erosional bed amalgamation. Crinoid stem plates (columnals) are predominant components whereas cirri and arm fragments are less common (Fig. 9D; Głuchowski, 1987). Crinoid assemblage is dominated by *Isocrinus* sp., *Balanocrinus* sp. while Cyrtocrinida occur less commonly (Świńska Turnia crag; Głuchowski, 1987). Fragments of echinoid spines, mollusc and brachiopod shells, ostracods, bryozoans, belemnite rostra, and benthic foraminifera occur subordinately (Jach, 2005). An indistinct trend in coarsening of the crinoidal grains from 0.5 mm to 1–1.2 mm is observed upward the sections.

A lack of stratigraphically diagnostic fossils hindered the precise age determination of the crinoidal limestones. They are ascribed to the Lower Toarcian since they are overlaid by red limestones of the latest Early–Late Toarcian (Lefeld *et al.*, 1985; Myczyński and Lefeld, 2003; Myczyński and Jach, 2009; see stop A6.4). Chemostratigraphic data from crinoidal limestone indicate that in their uppermost part a significant $\delta^{13}C$ positive excursion occurs ($\delta^{13}C \sim 3.6‰$; Krajewski *et al.*, 2001). According to Jenkyns (2003), pronounced positive $\delta^{13}C$ excursion is associated with Early Toarcian Tenuicostatum and Falciferum zones; the excursion is interrupted with the negative shift in the early Exaratum subzone of the Falciferum Zone. As the isotope excursion occurs in the upper part of the crinoidal grainstones, it can be assumed that this part of the section refers most probably to the Tenuicostatum Zone of the Early Toarcian.

All the above described sedimentological characteristics point to a multiple reworking and winnowing of crinoidal material by storm-induced oscillatory currents from elevated parts of the basin. Thus, the sedimentation of the crinoidal grainstones took place between the storm and fair-weather wave bases. The succession of the crinoidal intercalations in the spiculites and the overlying crinoidal grainstones show a vertical transition from distal to proximal tempestites, which is a record of a progressive shallowing upward trend (Jach, 2005).

A6.3 Huciański Klin ridge – Early Toarcian manganese deposits

(49°15′37″ N, 19°49′20″ E)

Leader: Renata Jach

The Mn-bearing deposits of the Krížna Unit in the Western Tatra Mts occur locally between the crinoidal tempestites (Lower Toarcian) and the pelagic red limestones (Lower Toarcian–?Aalenian). They crop out exclusively between the Chochołowska and Lejowa valleys where they were mined as Mn ore in the 19th century (Jach, 2002b). The Mn-bearing deposits form a lens which is up to 2 m thick and stretches at the distance of a few hundred metres at the Huciański Klin ridge. The old shafts provide the only accessible outcrops of these rocks there. The longest of them is 41 m long (Jach, 2002b). The Falciferum Zone of the Early Toarcian may be estimated for the Mn-bearing deposits on the basis of their position in the section (Krajewski *et al.*, 2001; Myczyński and Lefeld, 2003; Myczyński and Jach, 2009).

For the purpose of simplicity, the Mn-bearing deposits are subdivided in three parts: rocks underlying the Mn-rich bed (30–70 cm thick), the Mn-rich bed (up to 110 cm thick), and rocks overlying the Mn-rich bed (up to 40 cm thick; Jach and Dudek, 2005; Fig. 10).

Generally rocks underlying the Mn-rich bed consist of: (i) a Fe-rich layer, (ii) an X-ray amorphous Mn-oxide layer, (iii) shales, and (iv) a massive jasper bed. The 15 cm thick Fe-rich layer comprises up to ~30 wt% Fe (mainly hematite). The overlying 10 cm thick Mn-oxide layer is composed mainly of X-ray amorphous Mn oxide (up to 46 wt% Mn). Some 2–15 cm thick shales occur between the rocks described above and the main Mn-rich bed. Clay minerals comprise a complex suite dominated by mixed-layer clays with chlorite, smectite, and vermiculite layers (Dudek and Jach, 2006). The Fe-rich layer, and the Mn-oxide layer underwent locally silicification which resulted in the development of a massive jasper bed, which is 10 to 50 cm thick (Figs 10, 11A, B).

The Mn-rich bed, with sharp and well defined top and bottom, ranges in thickness from 35 cm to 1.1 m. It comprises clearly defined, lenticular zones composed either of Mn carbonates or of Mn silicates (Fig. 11C–E). The carbonate zones are purple-red, whereas the silicate zones are usually black.

The Mn-carbonate zones are built of calcite and Mn-calcite, with various Ca/Mn ratios whereas Mn sili-

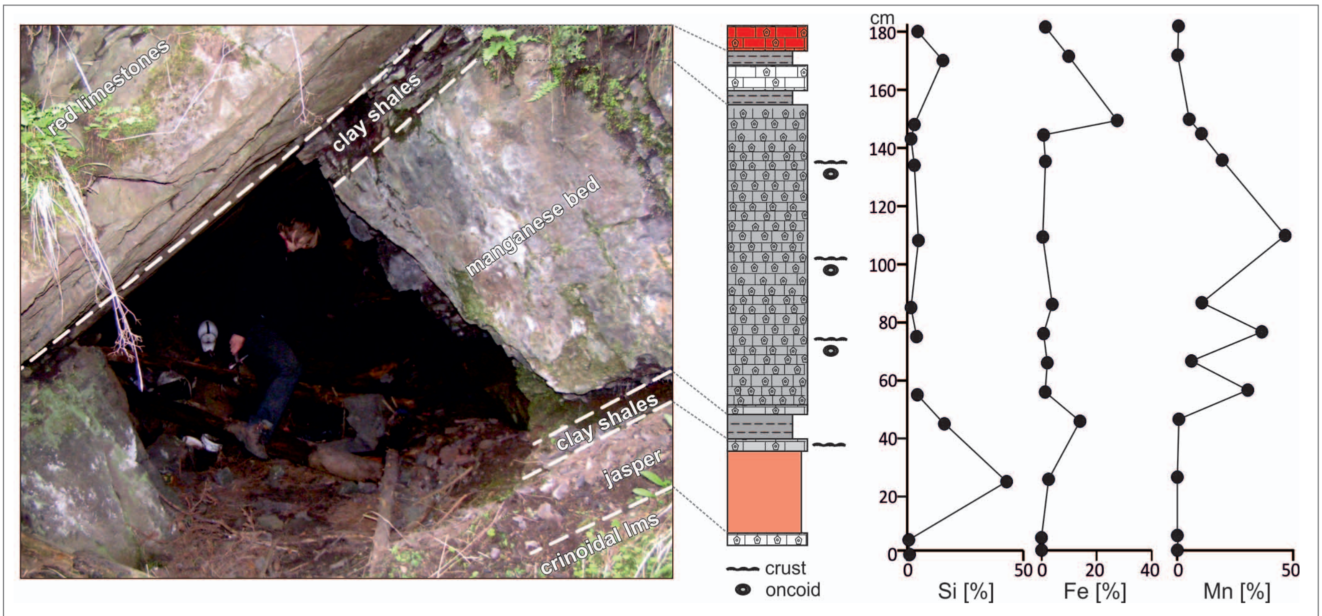

Fig. 10. Manganese-bearing deposits cropping out in a shaft entrance. Lithological section with general geochemical characteristic (Si, Fe and Mn).

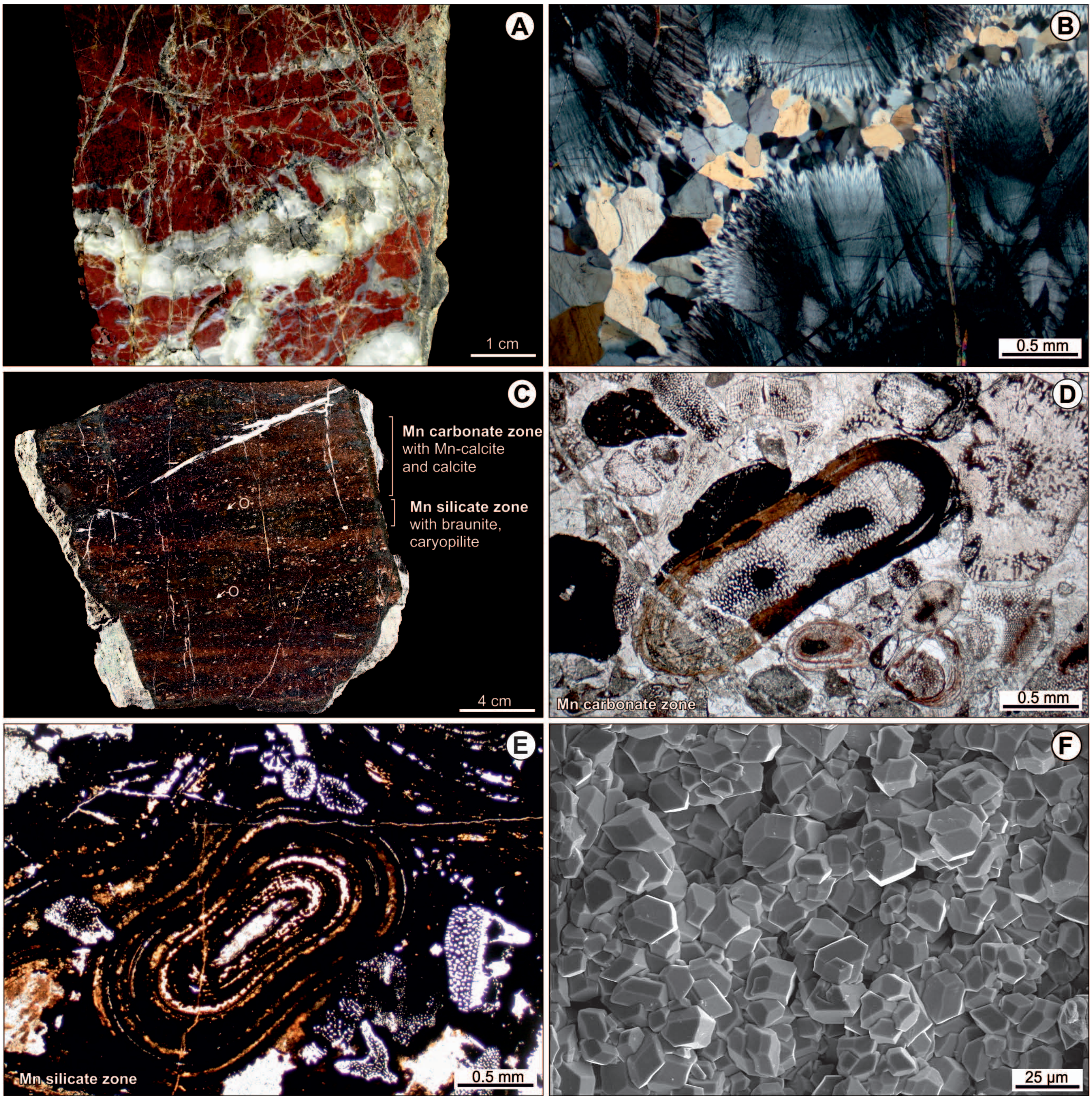

Fig. 11. Manganese-bearing deposits **A)** Jasper, polished slab; **B)** Internal structure of jasper bed, fibrous chalcedony and blocky quartz. Thin section, cross-polarised light; **C)** Mn-bearing bed shows subtle stratification underlined by occurrence of carbonate and silicate zones. Polished slab; **D)** Carbonate zone. Manganese oncoide within echinoderm grainstone, mainly echinoid-crinoidal, with syntaxial cements built of calcite and manganese calcite. Thin section, plane polarised light; **E)** Silicate zone with abundant microbial structures, mainly oncoids. Thin section, plane polarised light; **F)** Silicate zone. Authigenic idiomorphic crystals of braunite, SEM image of HCl-etched surface.

cates (braunite, caryopilite) occur in minor amounts (Korczyńska-Oszacka, 1978; Jach and Dudek, 2005). Pure rhodochrosite forms small lenses up to 6 cm long in the upper part of the bed (Korczyńska-Oszacka, 1979). Conversely, the silicate zones are dominated by braunite, forming frequently idiomorphic crystals (Fig. 11F), and caryopilite. Calcite and subordinate amounts of dolomite, apatite, and barite have been identified as non-manganese minerals. The chemistry of the carbonate and silicate zones differs mainly in the Mn content: ~5 wt% Mn in the former ones in contrast to more than ~50 wt% Mn in the latter ones (Jach and Dudek, 2005). The Mn-rich bed comprises low concentrations of transitional elements (Co + Ni + Cu < 0.01 wt%). On a ternary diagram of Co+Ni+Cu, versus Fe and Mn, our samples plot in the area of Mn deposits of hydrothermal and diagenetic origins. In contrast, it is enriched in Ba (up to 4500 ppm). Total REE (8 elements) contents range from 83 to 151 ppm which may indicate that the deposits studied may have formed by both hydrothermal and hydrogenetic processes.

The Mn-rich bed, especially the silicate zones, abounds in microbial structures: crusts and oncoids (Fig. 11C, E; Jach and Dudek, 2005). Crusts cover and bind bioclasts. Oncoids, 3–30 mm across, are usually elongated, rarely isometric. The nuclei of the oncoids are usually composed of bio- and lithoclasts, less frequently of barite crystals (cf. Krajewski and Myszka, 1958). Their cortex is composed of concentric laminae of Mn-silicates and Mn-calcite. Microbial structures are accompanied by bioclasts, namely fragments of echinoids and crinoids. Tests of benthic foraminifera, shells of molluscs, ostracods, holothurian sclerites, and bryozoan fragments are less common. This assemblage of fauna and microbial structures occurs exclusively within the Mn-rich sequence; it is not found in the overlying or underlying deposits or in deposits that are lateral equivalents of the Mn-rich sequence.

Rocks overlying the Mn-rich bed are composed of two layers of shale separated by a layer of bioclastic limestone. The agglutinating foraminifera, represented almost exclusively by *Recurvoides*, occur in the upper shale (Tyszka *et al*., 2010). The clay fraction of the shales is dominated by illite and illite-rich illite-smectite mixed-layer clays (Dudek and Jach, 2006).

The Mn-bearing deposits are interpreted as formed in several stages controlled by a pulse-like activity of a shallow submarine exhalative vent (Krajewski and Myszka, 1958; Jach and Dudek, 2005). It is proved by sedimentological, mineralogical and geochemical evidence, coupled with the occurrence of specific microbial structures and a peculiar fauna assemblage (Jach and Dudek, 2005). The limited lateral extent of the Mn-bearing sequence may be an effect of low efficiency of the vent and/or of the seafloor topography. The specific faunal assemblage dominated by deposit feeders was intimately associated with trophic conditions which persisted near the vent orifice. During cessation in the vent activity, the sediments would have been exposed to and would have interacted with sea-water. Other early diagenetic processes included silicification and reduction of Mn oxides during suboxic diagenesis (Krajewski *et al*., 2001). Hydrothermal activity was the most likely source of colloidal silica. Idiomorphic shapes of braunite crystals embedded in Mn-calcite indicate that silicification preceded crystallization of Mn carbonates. The process of silicification resulted in the formation of Mn silicates in the Mn-rich bed, the massive jasper bed in the lower part of the sequence, and possibly also numerous silica lenses in the underlying crinoidal grainstones. The following stage of diagenesis – precipitation of Mn carbonate cements in the Mn-rich bed, was controlled by abundance of organic matter in the sediments. The latter is interpreted as genetically related to the microbial productivity *in situ*.

The position of the Mn-bearing deposits over the crinoidal tempestites and below the red pelagic limestone indicates that they were laid down at neritic depths. Moreover, it implies that Mn-bearing deposits mark a substantial change from relatively shallow (tempestites) to deeper water (red pelagic limestones) sedimentation milieu. This change most probably resulted from synsedimentary tectonic activity. It is in line with expelling of fluids by the vent. It was associated with extensional faults which provided channelways for geofluid migration upward (Jach and Dudek, 2005). The internal facies variation of the Mn-bearing sequence may be explained by changes in bottom-water chemistry, geofluid temperature or by lateral migration of the vent orifices.

A6.4 Huciański Klin ridge – Lower Toarcian-?Aalenian red limestones and marlstones and Bathonian *Bositra* limestones

(49°15′37″ N, 19°49′24″ E)

Leaders: Alfred Uchman, Renata Jach

The Mn-bearing sequence is covered by red limestones and marlstones, about 4 m thick, which in turn, are overlain by thin-shelled bivalve-bearing limestones (hereafter called *Bositra* limestones). The latter are up to 3 m thick (Figs 12, 13A; Jach, 2007). The ammonites found in the red limestones indicate the upper part of Lower Toarcian–Upper Toarcian (Serpentinum–Pseudoradiosa zones; Myczyński and Jach, 2009). The outcrop of red limestones is located at the entrance to one of the Mn adits.

The red limestones and marlstones are characterized by concentrations of pelagic fauna remains (Fig. 13B), such as ammonites, belemnites and fish teeth, which are a common feature of the Jurassic pelagic limestones deposited on elevated settings (pelagic carbonate platforms *sensu* Sanantonio, 1993). Six microfacies have been distinguished within red nodular limestones: crinoidal-ostracod packstone, crinoidal packstone, crinoidal wackestone, marly mudstone, *Bositra* packstone and *Bositra*-crinoidal packstone (Gradziński *et al*., 2004). Finning-upward trend (from packstones to wackestones), is accompanied by increasing upward abundance of microborings.

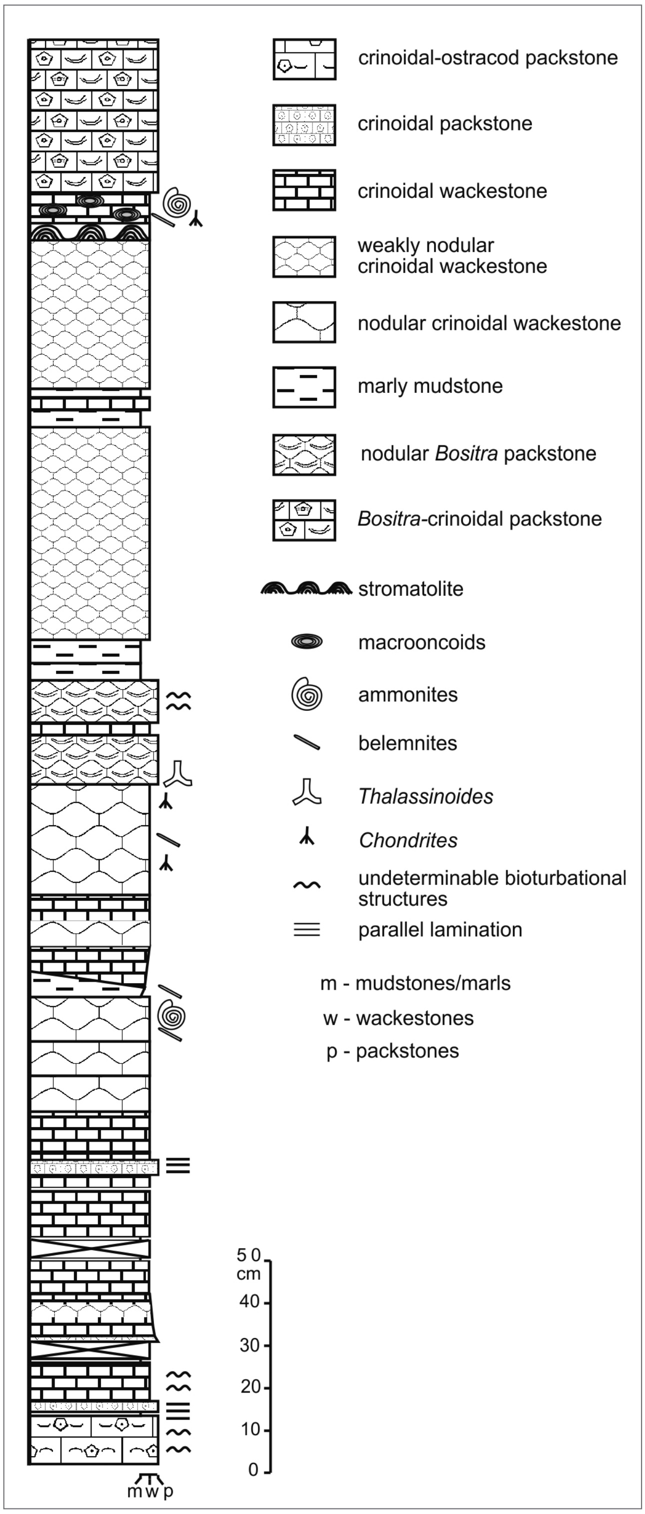

Fig. 12. Lithological section of the red limestones in the Huciański Klin site (Gradziński *et al*., 2004)

Crinoid and echinoid fragments are most abundant within microfossil assemblage. Fish teeth, benthic foraminifera, smooth-walled ostracods, juvenile ammonites are less common (Gradziński *et al*., 2004). Echinoid fragments dominate in the lower part of the red nodular limestone section and decrease upwards. Conversely, fish teeth are relatively rare in the lower part of the section and abundant in the upper part.

The upper part of red limestones and marlstones displays several features, such as concentration of nektonic fauna remains, occurrence of stromatolites and oncoids, and abundance of microborings typical of condensed section deposited on submarine elevations (Fig. 13A–D; Jenkyns, 1971; Bernoulli and Jenkyns, 1974). The features mentioned above collectively indicate a low depositional rate.

Microbial-foraminiferal oncoids, which have been recognized in the upper part of the red deposits, are the most peculiar feature of the studied section (Fig. 13C, D; Gradziński *et al*., 2004). The oncoids are up to 10 cm across and mostly display discoidal shape. Intraclasts or internal moulds of ammonites have acted as oncoid nuclei whereas oncoid cortices are composed of dark red laminae (mainly of iron hydrooxides/oxides), encrusting foraminifera (Fig. 13D; *Nubecularia* aff. *mazoviensis*, *Dolosella*, and agglutinated *Tolypammina*) and calcite cements. The dark red laminae dominate the oncoidal cortices and show a reticulate ultrastructure which is interpreted as mineralized biofilm, visible under SEM. Mineralized microbial bodies, globular and filamentous in shape, also built of iron hydrooxides/oxides, are abundant within the laminae. The association of foraminifera with microbes is supposed to be an adaptation of foraminifera to oligotrophic condition on the sea floor. It is very probable that biofilms served as food source for encrusting foraminifera. Formation of oncoids was possible under periodic water agitation (Gradziński *et al*., 2004).

The red limestones locally display nodular structure and contain a few discontinuity surfaces in the studied section (Gradziński *et al*., 2004). Some of such surfaces are burrowed with *Thalassinoides*, which indicates colonization of a firmground by crustaceans. Another discontinuity surface, which occurs in the uppermost part of the section, is manifested by concentration of internal moulds of ammonites and some fragments

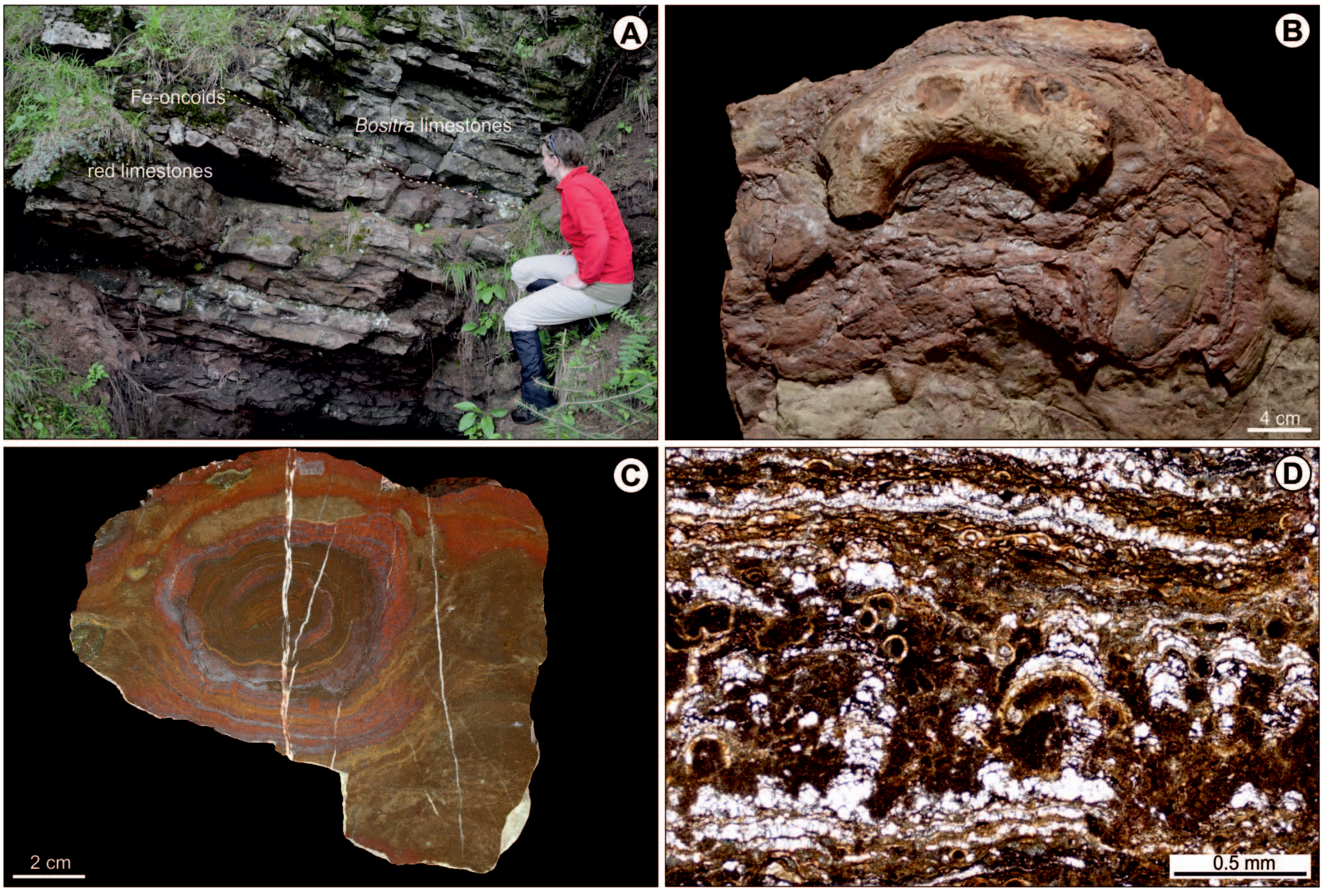

Fig. 13. Red limestones. **A)** Section of the red limestones and *Bositra* limestones at the entrance to one of the adits; **B)** Omission surface with ferruginous stromatolite encrusting ammonite mould and shell (*Lytoceras* sp.). Weathered surface; reprinted from Myczyński and Jach (2009); **C)** Cross-section through a microbial-foraminiferal ferrugineous macrooncoids. Polished slab; **D)** Cortex of oncoid built of encrusting foraminifers; thin section, plane-polarised light.

of ammonite shells covered with a stromatolite (Fig. 13B). There is a concentration of glaucony grains and crusts in the topmost part of the section, just at the contact of the red limestones and the *Bositra* limestones.

Overlying Bositra limestones are composed of Bositra-crinoidal packstones, and Bositra packstones/grainstones in this section (Jach, 2007). The latter dominate in the uppermost 3 m of the section. They were laid down in a relatively high-energy setting, which controlled the good sorting of these deposits. However, the domination of the Bositra bivalves seems to have resulted from some ecological factors, such as eutrophication of the water column. The dissolution of non-calcitic bioclasts is also suggested. Bositra-limestones lack of index fossils; their age is inferred as Lower Bathonian, based on superposition (Jach et al., 2014).

The red nodular limestones and the overlying *Bositra* limestones were formed during gradual deepening of the basin, which caused lowering of the depositional rate and, hence, condensation (Wieczorek, 2001). The deposition of *Bositra* limestones marks the first stage of unification of facies, which probably took place during Lower Bathonian. This process was later manifested by the deposition of radiolarites in the whole Fatricum Basin (Ožvoldová, 1997; Polák *et al*., 1998; Jach *et al*., 2012, 2014). Thus, the deposition of *Bositra* limestones reflected the intermediate stage in the basin evolution leading to formation of radiolarites (Lefeld, 1974; Jach, 2007).

A6.5 Lejowa Valley – Lower Jurassic mixed siliciclastic/carbonate deposits

(49°15′46″ N, 19°50′39″ E)

Leader: Andrzej Gaździcki

The uppermost Triassic and lowermost Jurassic strata of the Fatricum (Krížna Nappe) in the Tatra Mts are represented by marly shales and quartz sandstones with limestone intercalations (Uhlig, 1897; Goetel, 1917; Gaździcki, 1974, 1975, 1983, 2003, 2014; Uchman, 1991; Krobicki and Uchman, 1993; Hołda, 2002). They are assigned to the Fatra and Kopieniec formations (Gaździcki *et al*., 1979; Lefeld *et al*., 1985).

The Lower Liassic (= Grestener Schichten of Uhlig, 1897; Kopieniec Schichten of Goetel, 1917; Kopieniec Formation of Gaździcki *et al*., 1979) strata crop out

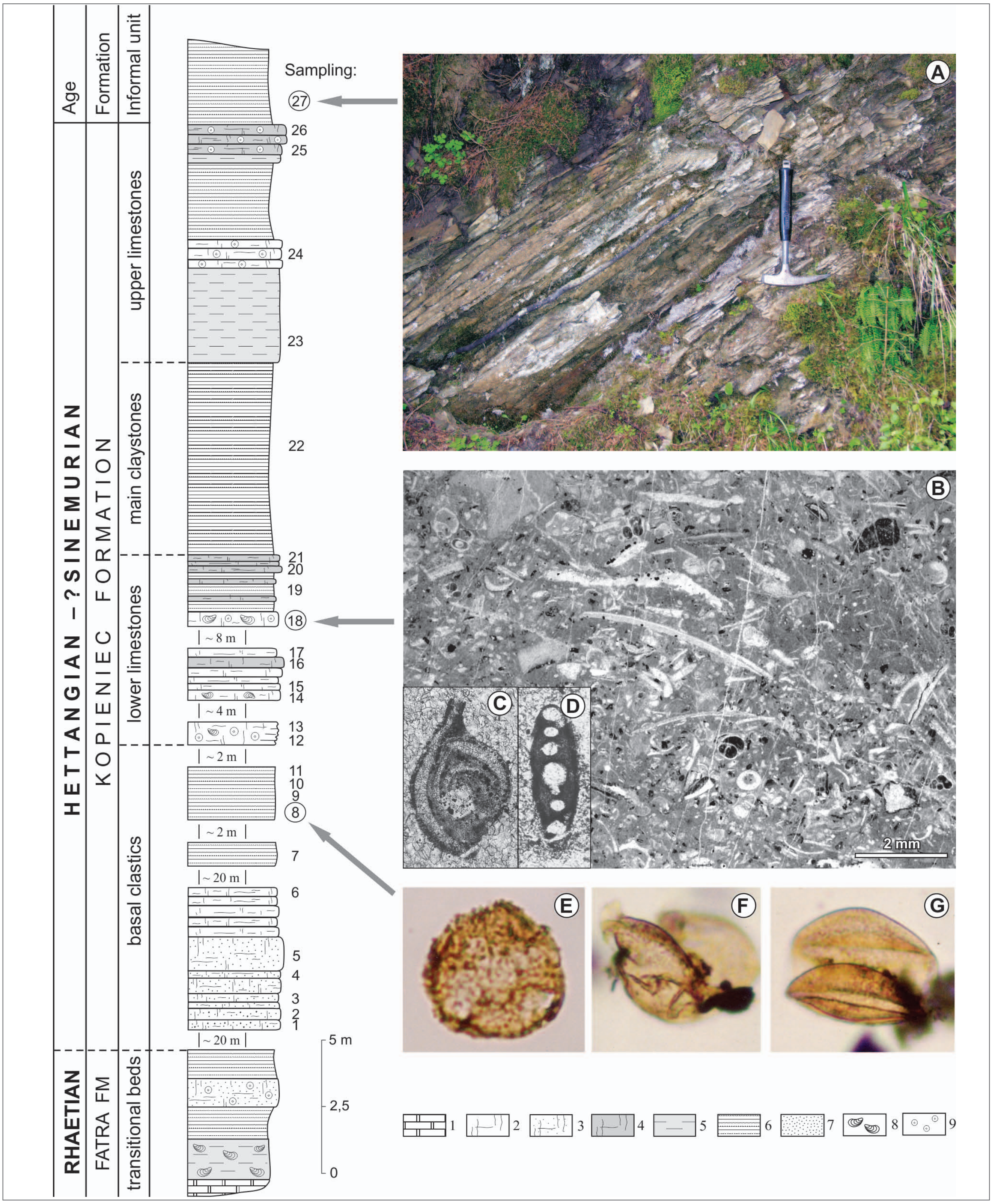

Fig. 14. Section of the uppermost Fatra and Kopieniec formations in Lejowa Valley. The section shows lithology and distribution of some important biota components. 1 – dolomites, 2 – limestones, 3 - sandy limestones, 4 – marly limestones, 5 – marls, 6 – shales, 7 – sandstones, 8 – bivalve shells, 9 – crinoids (trochites). **A)** Exposure of marly shales in the upper part of the formation (hammer for scale); **B)** Bivalve-crinoid-gastropod-foraminifer biopelmicrite microfacies; **C, D)** Benthic foraminifers *Ophthalmidium leischneri*; x 110; **E-G)** Palynomorphs: **E)** *Baculatisporites comaumensis*, x 500; **F)** *Schismatosporites ovalis*, x 300; **G)** *Quadreaqualina anellaeformis*, x 300.

over a large part of the Sub-Tatric (Krížna) Unit along the northern slopes of the Tatra Mts (Guzik *et al.*, 1975, Bac-Moszaszwili *et al.*, 1979; Gaździcki, 2014, figs 4.4.2, 5.1.1).

In the Lejowa Valley, the Kopieniec Formation is well exposed on the northeastern slopes of Wierch Spalenisko Mt. The presented section (Fig. 14) was traced along the ravine (creek) from Wierch Spalenisko Mt. to Huty Lejowe Glade at the altitude 1040–985 m a.s.l. and rests on the transitional beds of the Rhaetian Fatra Fm.

The Kopieniec Formation (up to 100 m thick) is subdivided into the following informal litostratigraphical units of the member rank: basal clastics, lower limestones, main claystones, and upper limestones (Fig. 14, see also Gaździcki *et al.*, 1979). Quartz sandstones, slightly calcareous with clayey-limonitic matrix and represented in form of distinct layers in the basal part of the formation are the main lithological type here and belong to the "basal clastics" of the formation (Gaździcki, 2014, fig. 5.1.6). In the upper part of the unit, an assemblage of palynomorphs: *Baculatisporites comaumensis*, *Schismatosporites ovalis*, *Quadreaqualina anellaeformis* as well as *Concavisporites*, *Dictyophyllites* and *Leptolepidites* was found in the brown-gray laminated marly shales (Gaździcki *et al.*, 2006, fig. 1) see also Fig. 14E–G). The next unit — "lower limestones" — contains of dark-grey sandy organodetrital limestones. These are bivalve-crinoid-gastropod-foraminifera biopelmicrites with *Pycnoporidium*? encrustacions and envelopes. The spores *Globochaete* and *Eotrix* are common. Among the foraminifera, *Ophthalmidium leischneri* (Fig. 14B–D), *Planiinvoluta, Nodosaria, Lenticulina* and post-Triassic involutinids predominate (Gaździcki 2014, figs 5.1.8). Brown-gray shales (claystones) with marly intercalations prevail in the "main claystones". The upper unit of the Kopieniec Formation — "upper limestones" — comprises a sequence of brown-gray marls, shales and dark-gray organodetrital limestones (mostly crinoid-ostracod-brachiopod biointrapelsparite). The limestone intercalations contain numerous benthonic foraminifera: *Ophthalmidium leischneri, O. walfordi, Involutina liassica, I. turgida, I. farinacciae* and nodosarids (Gaździcki, 2014, figs 5.1.10–11). The uppermost part of the Kopieniec Formation section is terminated by brown-gray marly shales with marly intercalations (Fig. 14A).

The clastic-carbonate shallow-marine Late Triassic/Early Jurassic successions of the Tatra Mts contain biostratigraphicaly important microfossils, mostly foraminifera, ostracodes, and conodonts, as well as palynomorphs and coprolites (Błaszyk and Gaździcki, 1982; Gaździcki, 1974, 1975, 1977, 1978, 1983, 2014; Fijałkowska and Uchman, 1993). They have been used to erect local zonations and they may also be of prime importance for regional biostratigraphic correlations and palaeogeographic reconstruction. The recognized evolutionary lineages and the rapid rates of evolutionary changes of representatives of the benthonic foraminifera families Involutinidae and Ammodiscidae and the subfamily Ophthalmidiinae permit an establishment of relatively high resolution zonation. A sequence of two foraminifera zones: *Glomospirella friedli* and *Triasina hantkeni* Zone (aassemblage zone, Rhaetian) and *Ophthalmidium leischneri* and *O. walfordi* (assemblage zone, Hettangian–?Sinemurian) is recognized (Gaździcki, 1983). The foraminifera biostratigraphical zonation of the late Triassic and Early Jurassic in the Tatra Mts shows that the boundaries of the litostratigraphical units, i.e. the Fatra- and Kopieniec formations in the Fatricum Domain are diachronous (Fig. 15, see also Gaździcki and Iwanow, 1976). It may coincide with the Rhaetian-Hettangian boundary, as was widely assumed, or pass through the Rhaetian.

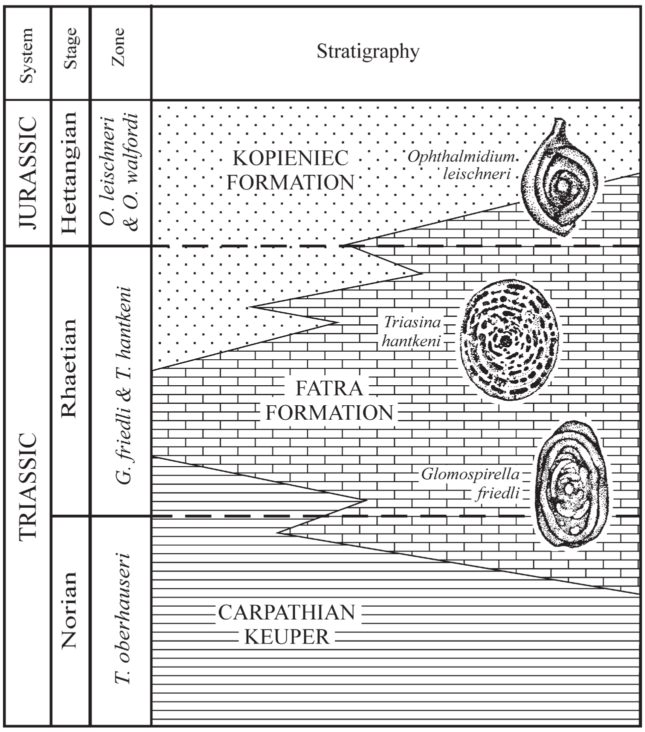

Fig. 15. Diachronism of the Fatra and Kopieniec formations in the Tatra Mts.

The Kopieniec Formation of the Tatra Mts comprises shallow-marine clastic (with trace fossils which indicate the *Cruziana* ichnofacies see Uchman, 1991) and mixed clastic-carbonate sediments (of tempestitic origin) deposited in the photic zone (Gaździcki, 2014). The character of these deposits reflects some general changes and especially epeiric movements at the turn of the Triassic and Jurassic. The sedimentary sequence and floral and faunal assemblages of the Kopieniec Formation are almost identical to those from contemporaneous strata of the Tethys realm. On the other hand, it is possible to note some similarity to contemporaneous deposits of the epicontinental basins in the north-western Europe. This was already noted by Goetel (1917), who emphasized a marked resemblance to the sandstones with *Cardinia* from the Tatra Mts and the Lower Liassic of Swabia in sedimentary environment and faunal community.

A6.6 Huty Lejowe Glade, Lejowa Valley – Late Sinemurian-Early Pliensbachian spotted limestones and marls

(49°15′55″ N, 19°50′37″ E)

Leaders: Alfred Uchman, Renata Jach

Bedded grey bioturbated limestones and marly limestones with marl intercalations are characteristic Early Jurassic, and locally Middle Jurassic, facies in the Krížna Unit. This hemipelagic facies is an equivalent of the Allgäu Formation (*Fleckenmergel/Fleckenkalk*) from the Northern Calcareous Alps (e.g., Gawlick *et al.*, 2009). It is called the spotted limestones and marls in the Carpathians.

An about 150 m-thick series of partly spotty calcilutites and calcisiltites (calcimudstone and wackestones) interbedded with darker marls is exposed along the Lejowy Potok stream near Huty Lejowe Glade (Polana Huty Lejowe) as well as in the lower part of a gully running from slopes of Pośrednia Kopka Kościeliska Mt. to Huty Lejowe Glade. According to Lefeld *et al.* (1985), the spotted limestones and marls, distinguished as the Sołtysia Marl Formation, are Early Sinemurian – Early Pliensbachian in age. Their Sinemurian age is based on the ammonites *Arnioceras falcaries* and *Arnioceras ceratitoides* of the Bucklandi Zone (Early Sinemurian) and *Echioceras raricostatum* and *Echioceras raricostatoides* of the Late Sinemurian Raricostatum Zone (e.g., Gaździcki and Wieczorek, 1984; Uchman and Myczyński, 2006). The same deposits are ascribed as the Janovky Formation in Slovakia (Gaździcki *et al.*, 1979).

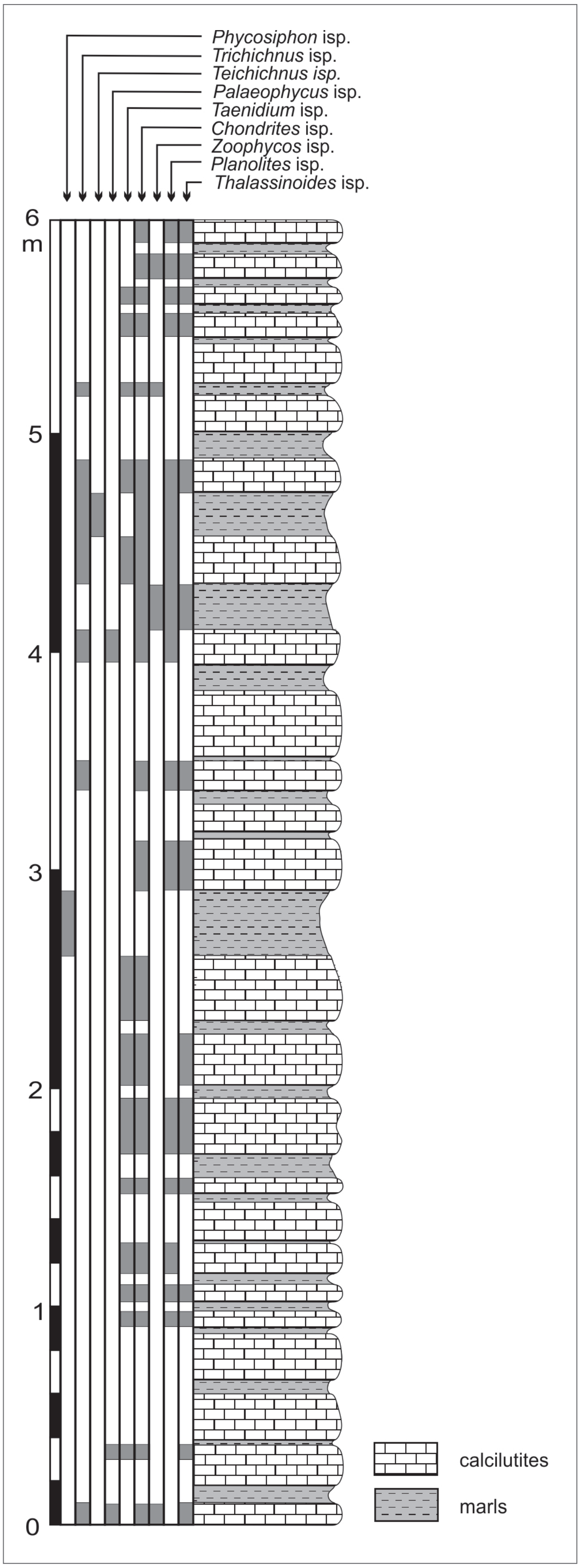

Fig. 16. A fragment of the section of the Sołtysia Marl Formation.

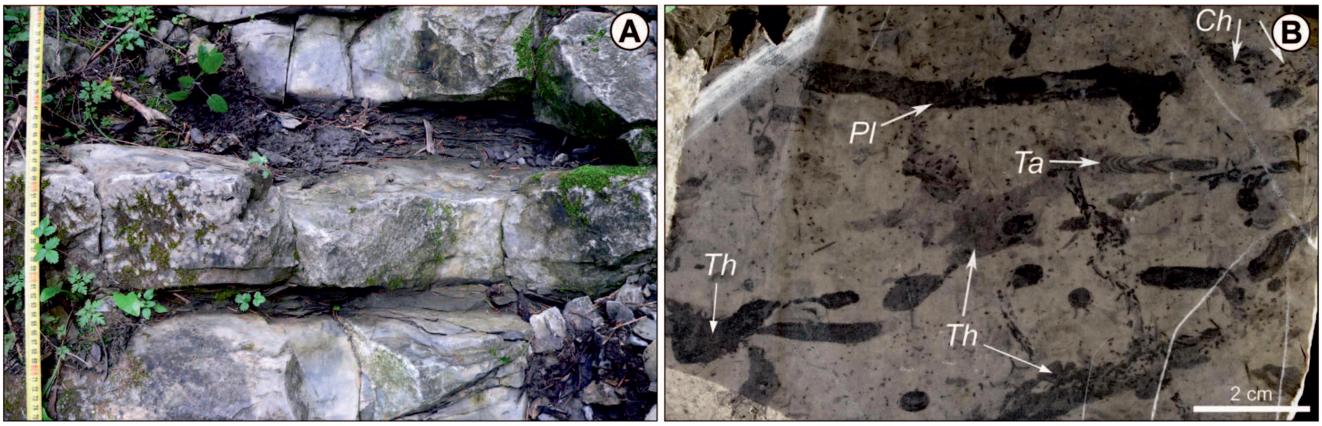

Fig. 17. Trace fossils from the Sołtysia Marl Formation, Lejowa Valley. **A.** Vertical cross section. *Th* – *Thalassinoides*; *Ch* – *Chondrites*; *Pl* – *Planolites*; *Ta* – *Taenidium*. **B.** Horizontal section. *Th* – *Thalassinoides*; *Pl* – *Planolites*; *Pa* – *Palaeophycus*; *Tr* – *Trichichnus*.

Proportions of the thickness of limestone beds to marl interbeds are variable (Figs 16, 17). For this reason spotted limestones and marls are subdivided into four subfacies:

- scarce in trace fossils, dark grey marls, interbedded with dark limestones, with average ratio of marl/limestone bed thickness 4:1 to 1:1. This subfacies occurs in the lower part of the section and refers to the Przysłop Marlstone Member (~ 30 m thick);
- bioturbated limestones, interbedded with dark marls, with marl/limestone bed thickness ratio 2:1 to 1:20. These deposits are ascribed to the Pośrednia Hala Marlstone Member;
- light grey limestones, subordinately interbeded with thin marls, with rare bioturbational structures. They are distinguished as the Pośrednia Kopka Limestone Member;
- dark grey limestones regularly interbeded with thin marls, which become strongly silicified toward the top. Trace fossils are scarce. This facies refers to the Parzątczak Limestone Member.

The deposits of the Sołtysia Marl Formation contain relatively common tests of benthic foraminifers, radiolarians, ostracods, sponge spicules, whereas macrofossils such as bivalve shells, ammonites, nautiloids and belemnite rostra are rare (e.g., Gaździcki *et al*., 1979).

Trace fossils are observed in cross sections as variable dark spots visible against lighter, totally bioturbated background (Fig. 17B). *Planolites*, *Chondrites* and *Thalassinoides* are common. *Zoophycos*, *Teichichnus*, *Taenidium*, *Phycosiphon* (formerly *Anconichnus*) and *Palaeophycus* occur subordinately. The trace fossils were described by Wieczorek (1995), who supposed that their tracemakers were controlled by bathymetry, oxygenation and trophic changes. Wieczorek (1995) distinguished between two main phases of bioturbation related to oxygenation changes. During the first phase, the totally bioturbated background was produced in well-oxygenated sediment. The fodinichnia-dominated ichnoassemblage containing *Chondrites*, *Planolites* and *Zoophycos* is typical of the second phase, when oxygenation of the sea floor dropped.

All the trace fossils display typical cross-cutting relationships. *Planolites* and *Thalassinoides* are cross cut by *Zoophycos, Chondrites* and *Trichichnus*, and *Zoophycos* is cross cut by *Chondrites* and *Trichichnus*. This order can be related to vertical partitioning of burrows in sediment (tiering) and their gradual shift following sediment accumulation or to sequential colonization in time by changing burrowing communities.

Gaździcki *et al.* (1979) suggested a depths from deeper neritic to bathyal zones for the discussed deposits, and Wieczorek (1984) interpreted them as basinal but shallower sediments deposited at the beginning of basin deepening. The general vertical trend in trace fossils suggests changes in nutrient supply and oxygenation of sediments. Also geochemical data, such as the values of the V/(V+Ni) ratio, comprised between 0.55 and 0.65, additionally imply changes in oxygen content from oxic to strongly dysoxic conditions. It is not excluded that the changes of calcium carbonate content are related to some climatic oscillations. The limestone-marl alternations probably reflect periodic delivery of siliciclastic material from adjacent lands. Such a process seems to be climatically dependent probably due to Milankovitch cyclicity (see Mattioli, 1997).

A6.7 Długa Valley, Pośrednie ridge – Middle–Upper Jurassic radiolarites

(49°15′37″ N, 19°48′04″ E)

Leaders: Renata Jach, Alfred Uchman, Nevenka Djerić[1], Špela Goričan[2], Daniela Reháková[3]

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The Middle–Upper Jurassic radiolarites crop out in the rock cliff located in the Długa Valley along the southern slope of the Pośrednie ridge (Fig. 2). The radiolarite-bearing succession comprises the following facies: (1) grey spotted radiolarites, (2) green radiolarites, (3) variegated radiolarites, and (4) red radiolarites (Figs 18; 19A–C). They are overlain by red limestones. The radiolarian-bearing deposits are ~30 m thick.

The radiolarites, on the basis of $\delta^{13}C$, radiolarians and calcareous dinoflagellata are of the Late Bathonian–early Late Kimmeridgian age (Fig. 18; Unitary Association Zones 7–11; Moluccana Zone; Polák *et al.*, 1998; Jach *et al.*, 2012, 2014). The broad negative excursion recorded

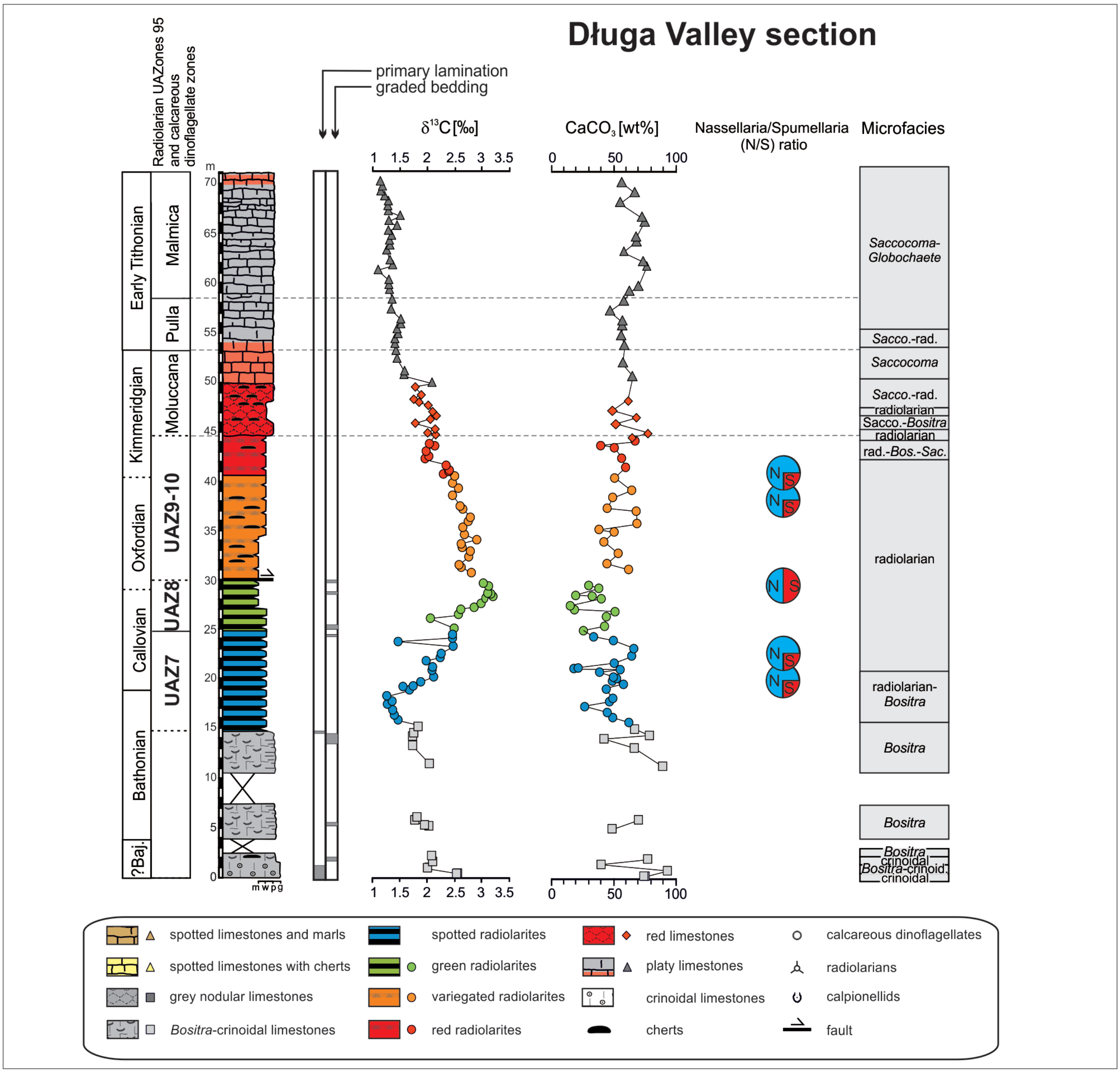


Fig. 18. Długa Valley section. Lithology, radiolarian and calcareous dinoflagellate biostratigraphy and results for carbon isotope measurements, $CaCO_3$ content and microfacies analysis (after Jach *et al.*, 2014; modified).

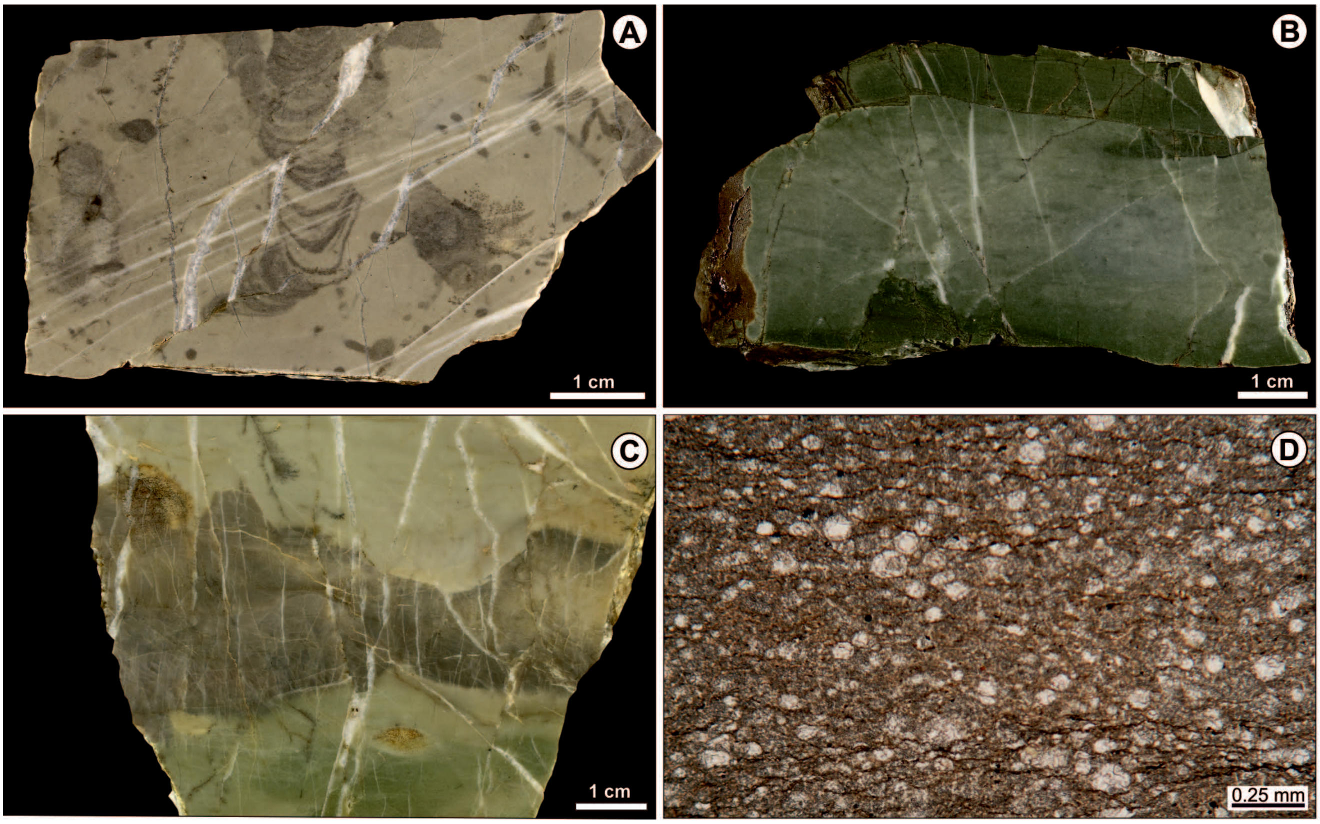


Fig. 19. Radiolarites. **A)** Bioturbated grey spotted radiolarites, polished slab; **B)** Green radiolarites. polished slab; **C)** Variagated radiolarites. polished slab; **D)** Green radiolarites, radiolarian wackestone-packstone. Thin section, plane polarised light.

in spotted radiolarites is Late Bathonian in age, whereas the pronounced positive $\delta^{13}C$ excursion detected in green radiolarites is referred to Late Callovian. It is worth mentioning that this excursion coincides with a distinct increase in radiolarian abundance and an extreme carbonate production crisis (Bartolini *et al.*, 1999; Morettini *et al.*, 2002). The variegated and red radiolarites and the overlying limestones display a pronounced decreasing $\delta^{13}C$ trend in the Oxfordian–Early Tithonian (Jach *et al.*, 2014).

The oldest radiolarites are grey and green, highly siliceous, thin- to medium-bedded, alternated with 0.2–2 cm thick siliceous shales (Fig. 19A, B). These chert-shale couplets are the most characteristic feature of the grey and green radiolarites. Average $CaCO_3$ contents in grey radiolarites and in green radiolarites are 36 wt% and 25 wt% , respectively (Fig. 18). The grey and green radiolarites show transition from *Bositra*-radiolarian to radiolarian microfacies, with calcified and extensively dissolved radiolarian tests (Fig. 19D). Grey spotted radiolarites are intensively bioturbated whereas green radiolarites locally show subtle microscopic lamination. The primary lamination can be referred to incidental rapid sedimentation marked by a subtle increase in grain size, or a short episode of anoxia.

The variegated and red radiolarites are calcareous, distinctly bedded, with bed thicknesses of about 10–15 cm (Fig. 19C). They have a higher content of $CaCO_3$, 50 wt% on average (Fig. 18). The common occurrence of massive chert nodules of various colours, mainly reddish or greyish, is typical of this facies. The very rare occurrence of thin shale intercalations is characteristic of the variegated radiolarites. The red radiolarites are calcareous, red and greyish red, and thin- to medium-bedded, with beds from 5 to 30 cm. The variegated and red radiolarite facies show upward transition from radiolarian to the *Bositra* and *Saccocoma* microfacies. They contain calcified or partially silicified radiolarians, sponge spicules, crinoids, planktonic foraminifera *Globuligerina* and cysts of calcareous dinoflagellates.

Almost all radiolarite and the associated deposits are bioturbated (Jach *et al.*, 2012). They contain the trace fossils *Chondrites*, *Planolites*, *Zoophycos*, *Teichichnus*, *Phycodes*, *Trichichnus*, *Phycosiphon* and *Thalassinoides* (Uchman and Jach, 2014). They belong to the *Zoophycos* ichnofacies, which characterizes deeper shelf – basin plain settings with pelagic and hemipelagic sedimentation. Generally, the abundance and diversity of trace fossils decrease up the succession. In the Upper Bathonian–Lower Callovian, the grey spotted radiolarites display typical spotty structures, that is relatively dense and diverse cross sections of trace fossils *Chondrites*, *Planolites*, *Thalassinoides*, and *Zoophycos* are common. Up the succession, in the green, variegated and red facies, the spots decrease in density, contrast and diversity, up to disappearance. The changes are not ideally linear, but fluctuations in these features do not discard the general trend. They are not related to grain size or lithology.

The changes of ichnological features in the studied interval are caused mainly by decrease in food content in the sediments (Jach *et al*., 2012; Uchman and Jach, 2014). With deepening of the basin and decreasing sedimentation rate associated with generally advancing flooding of epicontinental areas, less and less food was supplied to the basin from shallower areas and less and less of it was buried in sediment. In more eutrophic conditions (lower part of the interval), organisms penetrated deeply in the sediment, where distinct trace fossils were produced. A thicker layer of nutritional sediment gave an ecospace for a higher diversity of burrowing organisms. In more oligotrophic conditions (higher part of the succession), the organic matter was concentrated in the soupy sediment near the sediment-water interface, where preservation of distinct trace fossils was limited or impossible.

In the Długa Valley section, the Nassellaria/Spumellaria (N/S) ratio among radiolarians fluctuates, most probably in accordance with bioturbation intensity (Jach *et al*., 2015). It seems that the greater abundance of trace fossils coincides with the Nassellaria-dominated assemblage. It is thus possible that the observed pattern results from ecological requirements of these two groups of radiolarians. Spumellaria, which tend to be predominantly symbiont bearing, develop in more oligotrophic near-surface waters, whereas Nassellaria are non-symbiotic forms, which live in more eutrophic, deeper water column. Such a correlation may be explained by fluctuating input of nutrients from the neighboring lands, caused most probably by climate changes, for instance by enhanced continental weathering and runoff (cf. Baumgartner, 2013). An increased input of nutrients during humid climate leads to sea-water eutrophication, whereas decreased input leads to its oligotrophication.

The Late Bathonian–Kimmeridgian radiolarites are an evidence of deepening, with sedimentation taking place between the ACD and CCD or below the CCD. The middle Oxfordian–upper Kimmeridgian, variegated and red radiolarian-bearing facies, and finally red nodular and platy micritic limestones, record the recovery of carbonate sedimentation. It is evidenced by the middle Callovian – lower Oxfordian intervals characterised by drastically reduced $CaCO_3$ content, whereas an increase of carbonate content occurs in the middle Oxfordian–upper Kimmeridgian part of the section.

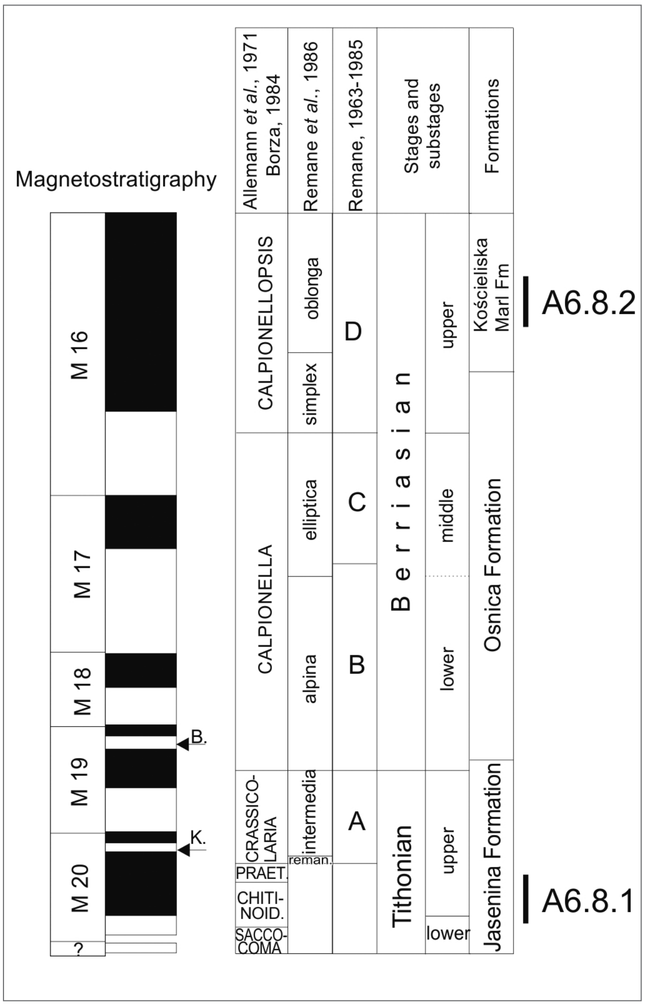
A6.8 Kryta Valley – Tithonian – Lower Valanginian limestones and marlstones: biostratigraphy, magnetostratigraphy, carbon isotope stratigraphy and paleoenvironmental changes

(49°15′41″ N, 19°48′ 20″ E)

Leaders: Jacek Grabowski, Andrzej Pszczółkowski

The Uppermost Jurassic (Tithonian) and Lower Cretaceous limestones and marlstones of the Križna Nappe crop out in the Kryta Valley and in neighbouring ridges between the Kryta and Długa valleys. Lithostratigraphically, this interval is divided into three units (Fig. 20). Shaly marls and olive grey, thinly-bedded micritic limestones occur in the upper part of the Jasenina Formation (Upper Tithonian to lowermost Berriasian). The light grey, calpionellid-bearing limestones of the Osnica Formation (Berriasian) overlie the Jasenina Formation deposits. The limestones are 25–37 m thick in various sections (Pszczółkowski, 1996) and pass gradually into the overlying strata of the Kościeliska Marl Formation (Lefeld *et al*., 1985). This formation, of Late Berriasian-Aptian age, comprises marls and limestones about 260 m thick (Lefeld *et al*., 1985; Pszczółkowski, 2003a; Kędzierski and Uchman, 1997).

Magneto- and biostratigraphic data were obtained from the Tithonian and Berriasian strata (Fig. 20; Grabowski, 2005; Grabowski and Pszczółkowski, 2006a), which enabled correlation of these sections with GPTS (Geomagnetic Polarity Time Scale), estimation of sedimentation rates and of palaeolatitudinal position of the area in the Berriasian (28°N, ± 4.5°). Detailed magnetic susceptibility (MS) and field gamma ray spectrometry (GRS), supported by geochemical analyses were used for reconstruction of palaeoenvironmental changes (Grabowski *et al*., 2013, see Fig. 21).



Stop A6.8.1

Platy, green to grey, micritic limestones of the Jasenina Formation, dipping to the north, are exposed along the road (Fig. 22).

The thickness of the section amounts to 6 m. In the lower part limestones prevail, while the upper part is dominated by marlstones. Limestones contain numerous remnants of crinoids (Saccocomidae) and chitinoidellids. The latter microfossils indicate the Boneti Subzone of the Chitinoidella Zone. The uppermost part of the section belongs to the Praetintinopsella Zone (Upper Tithonian). Bed K-1 is reversely magnetized while the bulk of the section, between beds K-2 and K-17, belongs to a normal magnetozone (interpreted as M20n). The sedimentation rate for the Boneti Subzone is estimated as 5.8–7 m/My (minimum value, without influence of compaction). In the neighbouring Pośrednie III section, the sedimentation rate calculated for entire magnetozone M20n is

Fig. 20. Litho-, bio- and magnetostratigraphic scheme of the Upper Tithonian and Berriasian of the Lower Sub-Tatric (Krížna Nappe) succession in the Western Tatra Mts (not to scale), after Grabowski and Pszczółkowski 2006b. B – Brodno magnetosubzone (M19n1r); K - Kysuca magnetosubzone (M20n1r).

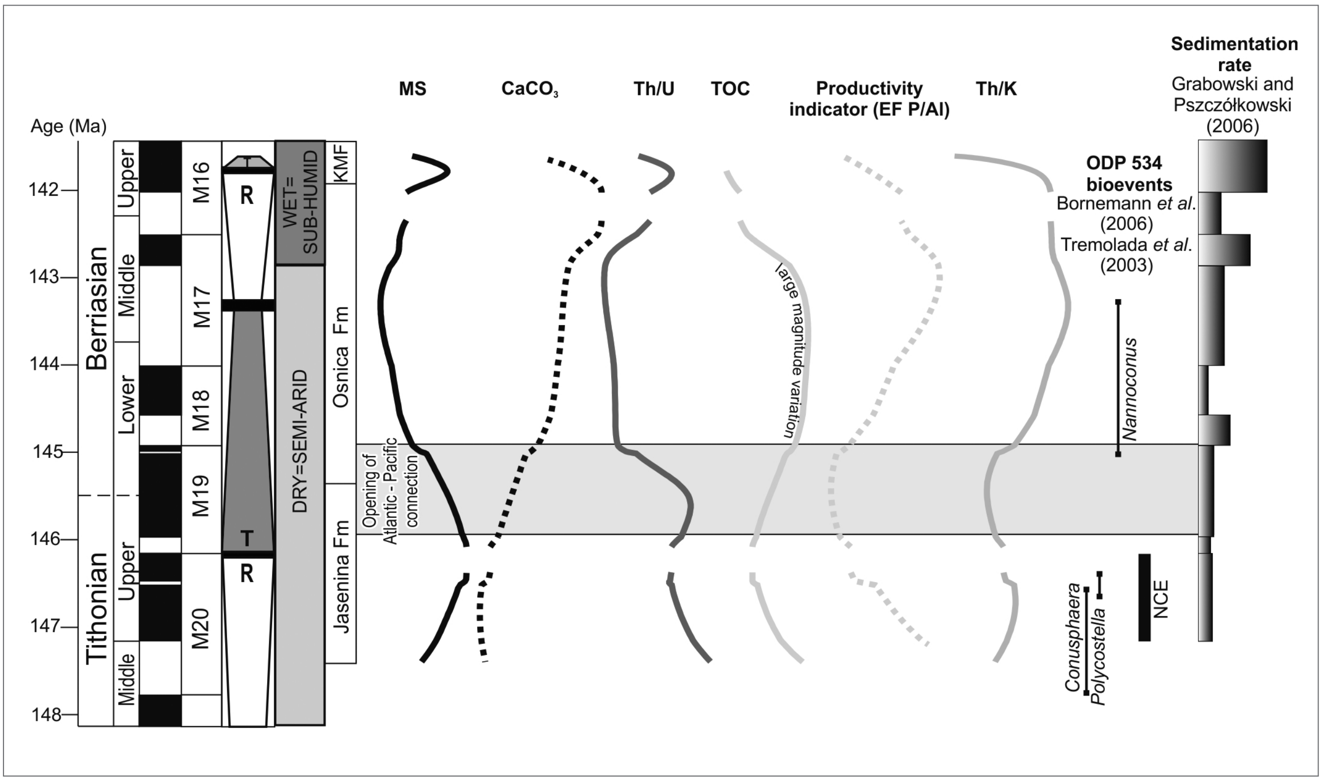


Fig. 21. Summary of magnetic and geochemical palaeoenvironmental trends at the Jurassic/Cretaceous boundary interval in the Pośrednie III section, Western Tatra Mts (after Grabowski *et al.*, 2013, slightly modified). The transgressive/regressive (T-R) cycles in the Tethyan domain, after Hardenbol *et al.*, (1998); paleoclimatic trends in ODP 534 after Bornemann *et al.*, 2003 and Tremolada *et al.*, 2006; general paleohumidity trends after Abbink *et al.* (2001) and Schnyder *et al.* (2006). NCE – Nannofossil calcification event.

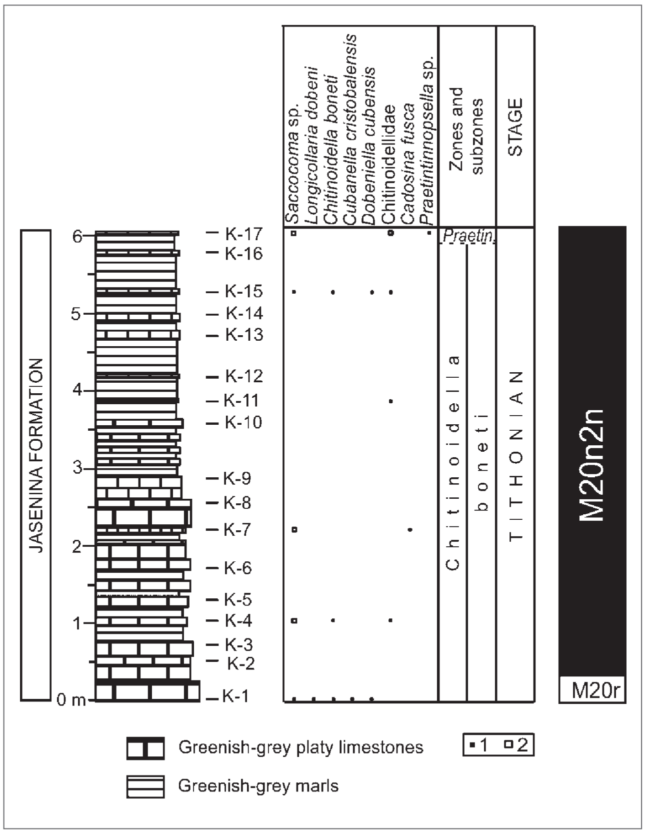


Fig. 22. Kryta section of the Jasenina Formation (Kryta Valley). *Praetin.* – *Praetintinopsella*. After Grabowski and Pszczółkowski (2006a). Taxon frequency: 1 – rare; 2 – common.

even lower (4.83 m/My). The distinct rhythm manifested by marlstone/limestone couplets might be related to Milankovich cyclicity (precession cycles of ca. 20 ky). The interval observed represents a local maximum of clastic input towards the basin. It is characterized by relatively high MS, high contents of Th, K and other lithogenic elements (Al, Ti, Zr, Rb etc.). High values of the Th/U ratio, abundant hematite horizons and rather low values of EF (enrichments factors) Cd/Al, Ni/Al and Mo/Al are evidences of well oxidized bottom waters. Decreasing values of productivity indicators (EF P/Al and Mn/Al) negatively correlate with terrigenous input (Fig. 21). The interval might be correlated to middle to a late Tithonian cooling phase evidenced by changes in calcareous phytoplankton (Tremolada *et al.*, 2006).

Stop A6.8.2.

We observe micritic limestones with *Calpionellopsis oblonga* (Cadisch) indicating the Oblonga Subzone (*sensu* Remane *et al.*, 1986) of the Upper Berriasian (Figs 20, 23).

They belong to the lowermost part of the Kościeliska Marl Formation. The gradual passage between the limestone-dominated Osnica and the marl-dominated

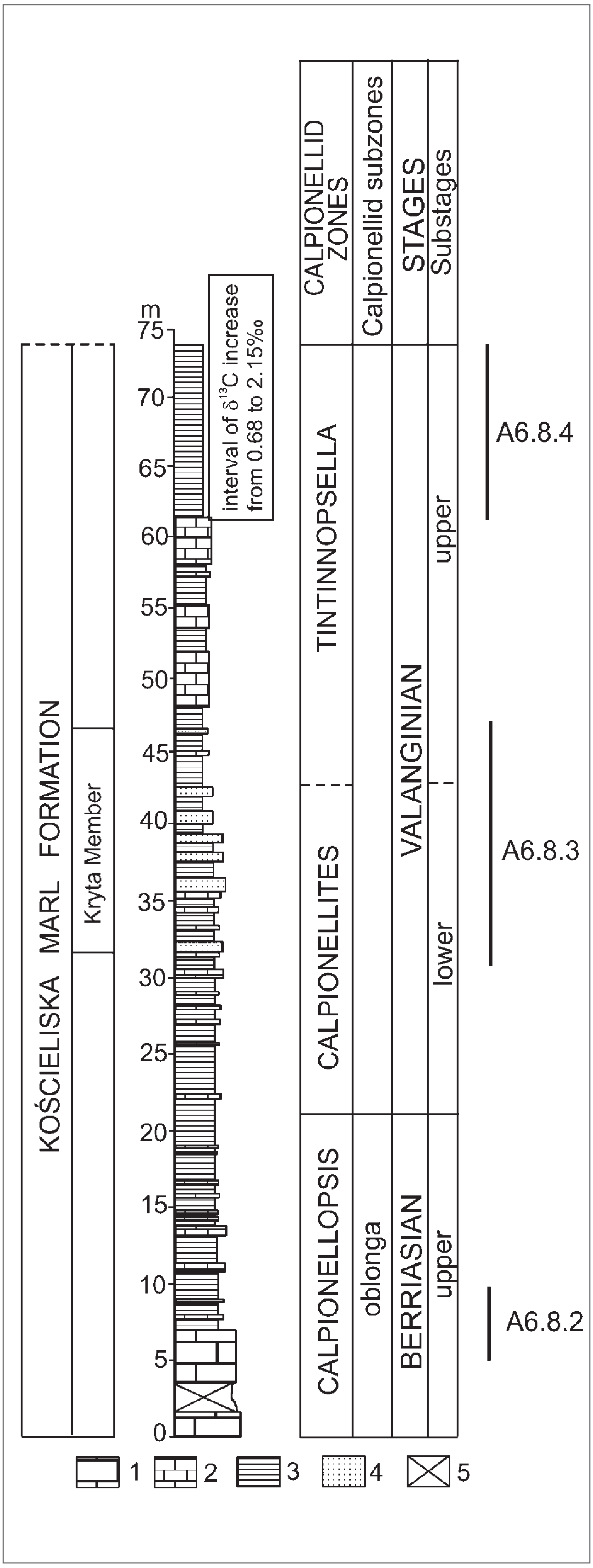


Fig. 23. Generalized stratigraphical section of the Kościeliska Marl Formation (lowermost part) in the Kryta Valley. After Grabowski and Pszczółkowski, 2006b, slightly modified. Calpionellid zones after Allemann *et al.* (1971) and Remane *et al.* (1986). Position of $\delta^{13}C$ event after Pszczółkowski (2003b) and Pszczółkowski *et al.* (2010). 1 – pelagic limestones; 2 –micritic and marly limestones; 3 –marlstones; 4 – sandstones (turbidites); 5 – covered intervals.

Kościeliska formations is covered in the Kryta creek. In the section localized in the adjacent Pośrednie ridge and section Rówienka (in the Lejowa Valley), the boundary between the two formations is situated in the lower part of the M16n magnetozone (close to the boundary between the Simplex and Oblonga subzones; see Fig. 20). The sedimentation rate of the Osnica Formation brackets between 10 and 17 m/My, and increases upward to 18–23 m/My in the lower part of the Kościeliska Marl Formation. In this section, the boundary between the Calpionellopsis and Calpionellites zones (i.e., the Berriasian-Valanginian boundary) falls in the magnetozone M14r. The Osnica Formation represents interval of carbonate sedimentation with limited influx of detrital material (Grabowski *et al.*, 2013). It is characterized by relatively low MS values, low contents of Th and K and other lithogenic elements (Fig. 21). The interval bears evidence of slightly oxygen-depleted conditions (low Th/U ratio, elevated values of EF Cd/Al, Ni/Al and Mo/Al) and enhanced productivity (higher EF P/Al and Mn/Al). The phenomena seem to be coeval with the warming period documented in the DSDP 534A core in the central Atlantic between magnetozones M18r and M17n (Tremolada *et al.*, 2006).

The transition between the Osnica and Kościeliska formations is manifested again by an increase in sedimentation rate, higher terrigenous influx, lower productivity and higher redox indices. The latter feature is independently supported by ichnofossil assemblages (Kędzierski and Uchman, 1997). The onset of terrigenous fraction delivery in the Late Berriasian might be regarded as a regional event within the Krížna (Fatric Domain) succession in the Slovak part of the Central Western Carpathians (e.g., Grabowski *et al.*, 2010). It can be followed as well, in similar stratigraphic position, in many sections of the Western Tethys: Northern Calcareous Alps, Western Balkan and Western Cuba (Grabowski and Sobień, 2015).

Stop A6.8.3

The Kryta Member of the Kościeliska Marl Formation is well exposed in the Kryta creek. This is the stratotype section of the member, containing marls interbedded with sandstones and marlstones, 15 m thick (Fig. 23).

The sandstones are mainly medium-grained lithic and arkosic arenites. However, hybridic arenites also occur (Świerczewska and Pszczółkowski, 1997). Magnetic separation revealed that Cr-spinels are an important component of the sandstones. The sandstones occur in similar stratigraphic position in the Oravice area, Slovakia, just 6–7 km west of our locality. After this locality, the Oravice event has been distinguished (Reháková, 2000; Pszczółkowski, 2003b). According to Reháková (2000), the Oravice event took place in the uppermost part of the Calpionellites Zone which was correlated with the ammonite Pertransiens Zone. Pszczółkowski (2003a) dated the sandstones as the upper part of the Calpionellites Zone

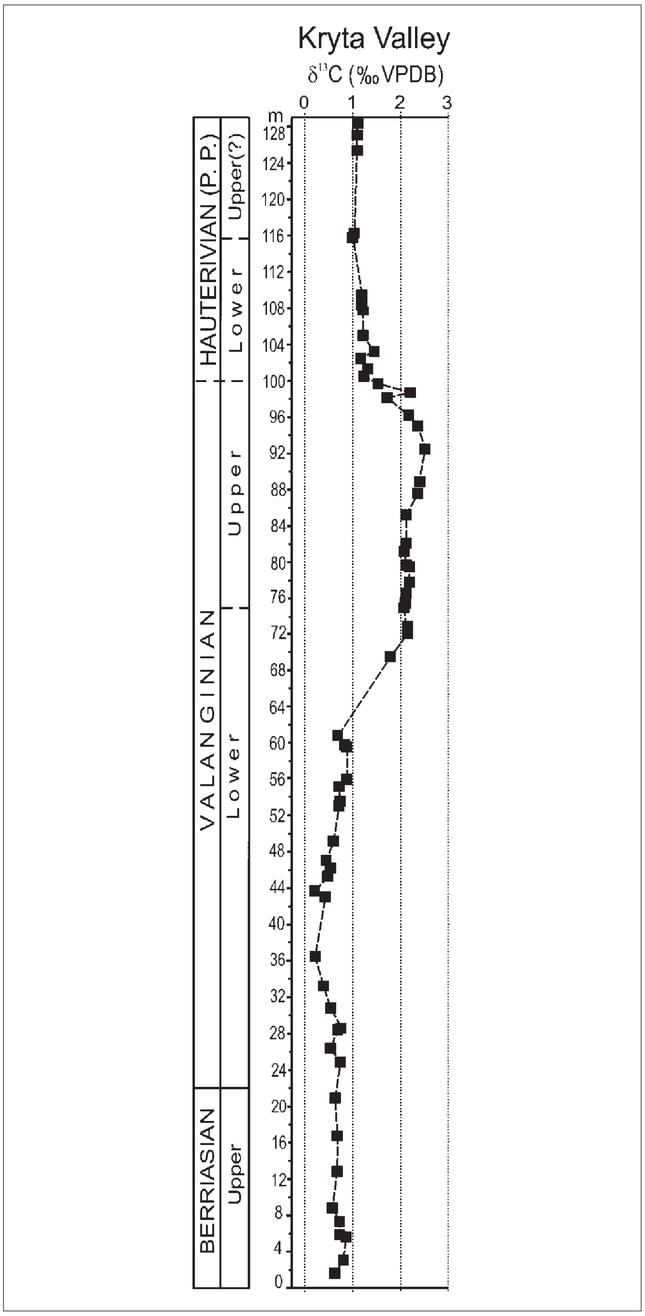


Fig. 24. $\delta^{13}C$ isotope curve within the uppermost Berriasian – Hauterivian interval of the Kryta section (after Pszczółkowski *et al.*, 2010).

and the lowermost Tintinopsella Zone (higher part of the Lower Valanginian).

Rocks of the Meliata suture zone situated to the (present day) east and/or south east of the Zliechov Basin (Slovakia) were probably the source of the clastic sediment (comp. Vašiček *et al*., 1994). The sedimentation rate within the Calpionellites Zone is estimated at 28–20m/My.

Stop A6.8.4

Within the upper Valanginian marls of the Kościeliska Marl Formation, a complete record of the $\delta^{13}C$ event was documented (see Fig. 24; Pszczółkowski, 2001; Pszczółkowski *et al*., 2010; see also Kuhn *et al*., 2005). The $\delta^{13}C$ values increase quickly from the values below 1‰ up to 2.15‰ close to the lower-upper Valanginian boundary. The event occurs in the interval of marly sedimentation and is not marked by any black shale deposition. This is similar to other Tethyan sections where anoxic sediments do not occur within the anomaly interval (e.g., Westermann *et al*., 2010). The integrated palaeoenvironmental study of the Kryta section is in progress, comprising detailed magnetic susceptibility logging and geochemical investigations.

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