IUGS UNESCO IGCP
Project 710
Western Tethys meets Eastern Tethys

United Nations
Educational, Scientific and
Cultural Organization

Dear Tethyan Friends,

We are glad that we can meet personally during the 2nd Symposium of the IGCP 710 Project, after the pandemic time, which disrupted our idea of regular, annual meetings. After the 1st virtual meeting during the autumn of 2021, we have now the chance to discuss face to face and go to the field together to touch „Tethyan” rocks for a better understanding of what happened hundreds/decades of millions years ago in our lovely ancient ocean. As you know, through your knowledge and experience, the Tethyan Ocean history, both in its western and eastern parts, is fascinating, but enigmatic from time to time, to say the least.

Generally, the geological history of the Tethys Ocean is broadly established. Yet many details are still unknown and many major questions remain, related to geotectonics, palaeogeography, palaeoceanography and palaeobiogeography. Improved understanding of the Mesozoic-Cenozoic ocean/climate history is based on accurate reconstruction of the distribution of continents and ocean basins and on opening and closing of seaways along the Tethys. There is little or no agreement about the number or size of separate basins, nor on their space-time relationships. Moreover, there is no consensus on the number and location of former micro-continents and on their incorporation into the present-day Eurasian-Mountain Belt. Geologists studying individual parts of these belts have been educated within different geological systems and adhere to different geological paradigms. Correlation between Western and Eastern Tethys is difficult, not only because of the large distances involved, but also because they are separated by the area of the huge Himalayan collision within which much of the pre-Paleogene tectonostratigraphic information has been lost. The aim of this IGCP project is to bring together geologists from the western and eastern parts of the former Tethys (Morocco/Iberia–SE Asia) to establish a common framework and a common tectonostratigraphic concept (latest Paleozoic–Mesozoic with emphasis on Permian–Jurassic).

On the one hand, UNESCO forms a special umbrella for the IGCP Projects, and on the other hand, it has been very active in supporting the ideas of “geoparks” and “geotourism” for years. For this reason, we decided to use an international magazine – *Geotourism* – to print our materials, both abstracts and a field trip guidebook. We hope it will be useful for both Tethyan friends and geotourism enthusiasts.

Enjoy Krakow during the stationary part of the Symposium and the Polish-Slovak-Czech Carpathians during a 5-day field trip!!

Michał Krobicki

IUGS
International Union of Geological Sciences

Ministerstwo
Edukacji i Nauki

Wydział Geologii, Geofizyki
i Ochrony Środowiska

Geo TOURISM
GEOTURYSTYKA
vol. 20, 3–4 (74–75) • 2023

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GEOTURYSTYKA

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Head of Publishing of AGH University Press: Jan Sas
Linguistic Corrector: Marta Golonka
Technical Editor: Monika Filipek
Cover Designer: Pracownia Kreatywna Bezliku
Typesetting and Desktop Publishing: Wydawnictwo JAK

The electronic version of the journal is the primary one.
eISSN 2353-3641
ISSN 1731-0830
DOI: https://doi.org/10.7494/geotour
Circulation: 150 copies

"Geotourism/Geoturystyka" is a scientific magazine published by AGH University of Science and Technology (Krakow).
In the years 2004–2009 the magazine was published in cooperation with Stanislaw Staszic Scientific Association (Krakow).

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www.wydawnictwo.agh.edu.pl

DOI: https://doi.org/10.7494/geotour.2023.74-75.5

Field trip –
Outer Flysch Carpathians and Pieniny Klippen Belt (PKB)

The position of the West Carpathians in the Alpine-Carpathian fold-and-thrust belt

(*Jan Golonka, Michał Krobicki*)

The Polish and Slovak West Carpathians form the northern part of the great arc of mountains, which stretch more than 1,300 km from the Vienna Forest to the Iron Gate on the Danube. Traditionally the Carpathians are subdivided into an older range known as the Inner Carpathians and the younger ones, known as the Outer Carpathians. From the point of view of the plate tectonic evolution of the basins the following major elements could be distinguished in the Outer Carpathians and the adjacent part of the Inner Carpathians (Golonka *et al*., 2005b; Ślączka *et al*., 2006):

- *Inner Carpathian Terrane* – continental plate built of the continental crust of Hercynian (Variscan) age and Mesozoic-Cenozoic sedimentary cover. The Inner Carpathians form a prolongation of the Northern Calcareous Alps, and are related to the Apulia plate (in a regional sense (Picha, 1996). The uppermost Paleozoic–Mesozoic continental and shallow marine sedimentary sequences of this plate are folded and thrust into a series of nappes. They are divided into the Tatric, Veporic and Gemeric nappes that are the prolongation of the Lower, Middle and Upper Austroalpine nappes respectively.
- *North European Platform* – large continental plate amalgamated during Precambrian-Paleozoic time. Proterozoic, Vendian (Cadomian), Early Paleozoic (Caledonian), Late Paleozoic (Hercynian) fragments could be distinguished within the folded and metamorphosed basement of this plate. Beneath of the Outer Carpathians the sedimentary cover consist of the autochthonous Upper Paleozoic, Mesozoic and Cenozoic sequences covered by the allochtonous Jurassic- Neogene rocks. The autochthonous Jurassic rocks within *North European Platform* are represented by mainly paltform facies. These allochtonous rocks are uprooted and overthrust onto the southern part of the North European Platform at a distance of at least 60–100 km (Książkiewicz, 1977; Oszczypko & Ślączka, 1985). They form stack of nappes and thrust-sheets arranged in several tectonic units. In Poland these allochtonous mainly flysch units are being regarded as Flysch Carpathians. Along the frontal Carpathian thrust, a narrow zone of folded Miocene deposits was developed.
- *Penninic realm* is a part of the Alpine Tethys (e.g., Birkenmajer, 1986; Săndulescu, 1988; Oszczypko, 1992; Plašienka, 1999, 2002; Stampfli, 2001; Golonka *et al.*, 2005b), which developed as a basin during Jurassic time between Inner Carpathian- Eastern Alpine terrane and North European Platform. In the western part it contains the ophiolitic sequences indicating the truly oceanic crust. In the eastern part the ophiolitic sequences are known only as pebbles in flysch, the basement of the Penninic realm was partly formed by the attenuated crust. In Poland, Slovakia and Ukraine the Penninic realm is represented by the sedimentary sequences of Jurassic, Cretaceous, Paleogene and Miocene age belonging to the Pieniny Klippen Belt (PKB) and the Magura Unit (Golonka *et al*., 2003). Some of these sequences are recently located in the suture zone between Inner Carpathian terrane forming the PKB, other sequences are involved in the allochtonous units covering the North European platform (Magura Nappe) or accreted to the Inner Carpathian terrane. Because of the evolutionary connotations of the Penninic realm, the PKB could be also regarded as belonging to the Outer Carpathians (e.g. Książkiewicz, 1977; Picha, 1996). The Czorsztyn submerged ridge was a part of the Penninic realm dividing the oceanic basin into two subbasins. The southern subbasin and the ridge traditionally constitute the Pieniny domain. Its sequences are involve in the PKB – strongly tectonized structure is about 800 km long and 1–20 km wide, which stretches from Vienna on the West to the Poiana Botizei (Maramures, NE Romania) on the East. The largest part of the northern subbasin form the Magura Unit, traditionally belonging to the Outer Carpathians. The PKB is separated from the Magura Nappe by the Miocene sub-vertical strike-slip fault (e.g. Birkenmajer, 1986, 1988). The Jurassic rocks of the Penninic realm are represented by basinal, slope and ridge facies.

The Polish Carpathians form the northern part of the Carpathians (Figs 1, 2). The Carpathian overthrust forms the northern boundary. The southern goes along the Poland-Slovakia national border. The Outer Carpathians are built of a stack of nappes and thrust-sheets showing different lithostratigraphies and tectonic structures. The Outer Carpathians nappes were thrust over each other and onto the North European Platform and its Paleocene-Miocene cover (Figs 3, 4). The present authors provided a systematic arrangement of the lithostratigraphic units according to their occurrence within the original basins and other sedimentary areas.

Fig. 1. Sketch of Alpine geology in Europe (after Picha, 1996; modified)

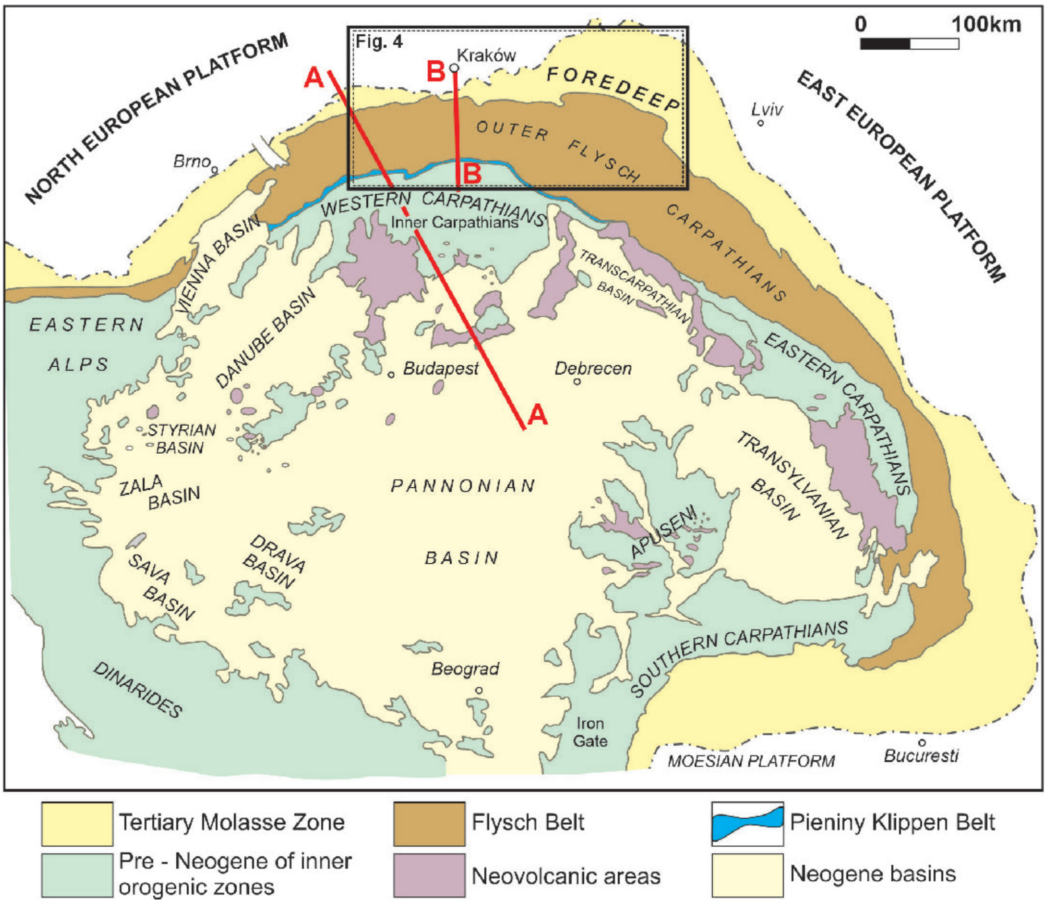


Fig. 2. Tectonic sketch map of the Alpine-Carpathian-Pannonian-Dinaride basin system (modified after Plašienka *et al*., 2000). A-A and B-B – localization of cross-sections (see Fig. 3)

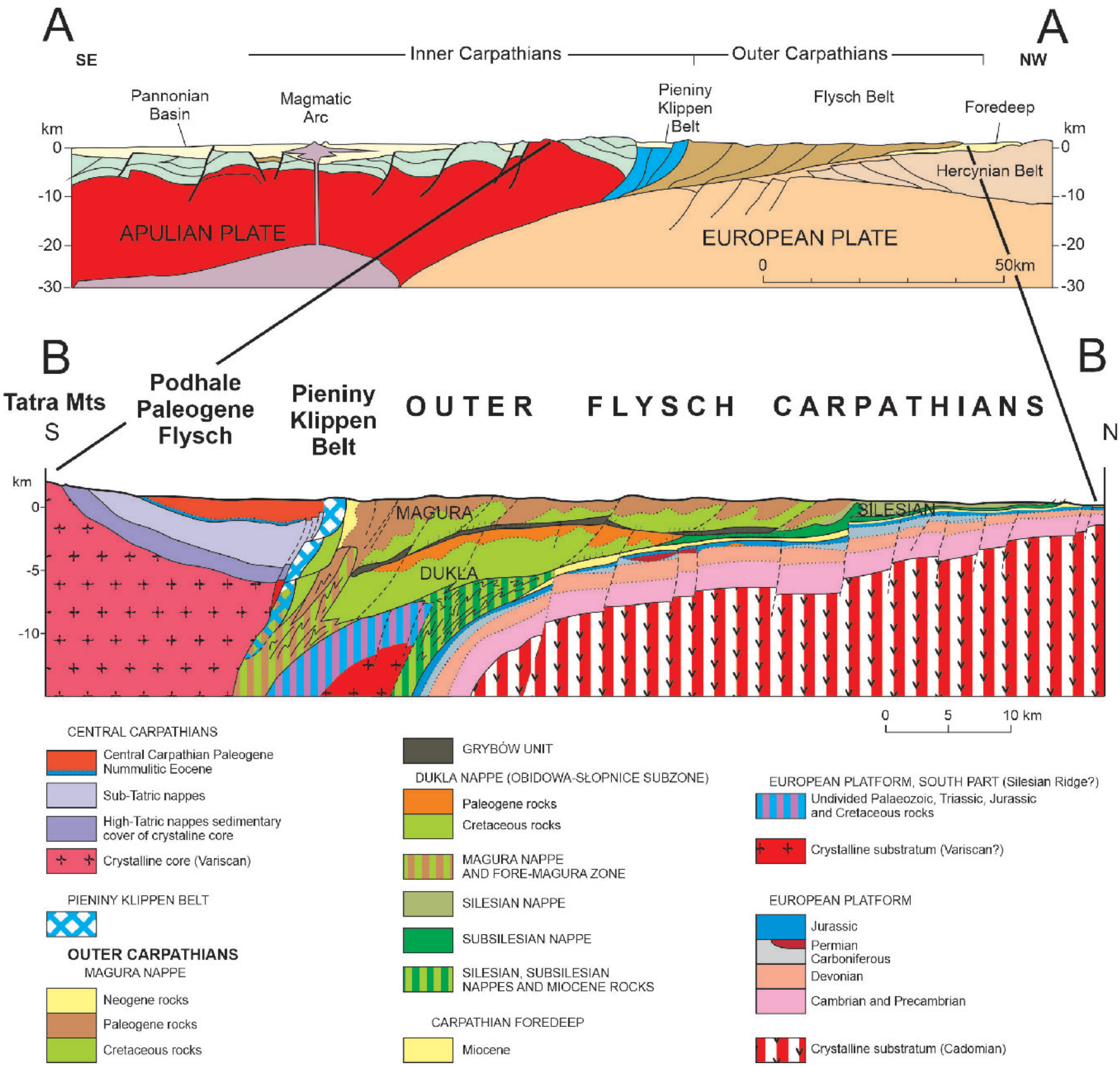


Fig. 3. Generalized cross-section across Carpathian-Pannonian region (Picha, 1996) (upper) and generalized cross-section across Polish Carpathians (after Golonka *et al*., 2005) (lower)

This guide focuses also on the plate tectonic elements important to understanding the geology of the Polish Carpathians. The Inner Carpathian Terrane is a continental plate built of continental crust of Hercynian (Variscan) age and a Meso-zoic-Cenozoic sedimentary cover. The uppermost Mesozoic sedimentary sequences of this plate are folded and thrust into a series of nappes. The large continental plate, amalgamated during Precambrian and Paleozoic times, is known as the North European Platform. Proterozoic, Vendian (Cadomian), Caledonian, and Variscan fragments occur within the platform. The southern part of the North European Platform, adjacent to the Alpine Tethys is known as Peri-Tethys.

The Alpine Tethys constitutes important palaeogeographic elements of the future Outer Carpathians, developed as an oceanic basin during the Jurassic as a result of the break-up of Pangea (some palaeogeographical sketches from global trough regional to local scales are given for example for Jurassic/Cretaceous transition times – Figs 5–7). The Czorsztyn submerged ridge (Pieniny Klippen Basin) was a part of the Alpine Tethys dividing the oceanic basin into two sub-basins. The southern sub-basin and the ridge are traditionally taken to constitute the Pieniny domain. Its sequences are involved in the PKB – a strongly tectonized structure about 600 km long and 1–20 km wide, which stretches from Vienna in the west to the Poiana Botizii (Maramures, NE Romania) in the east. The largest part of the northern sub-basin forms the Magura Unit, traditionally taken as belonging to the Outer Carpathians.

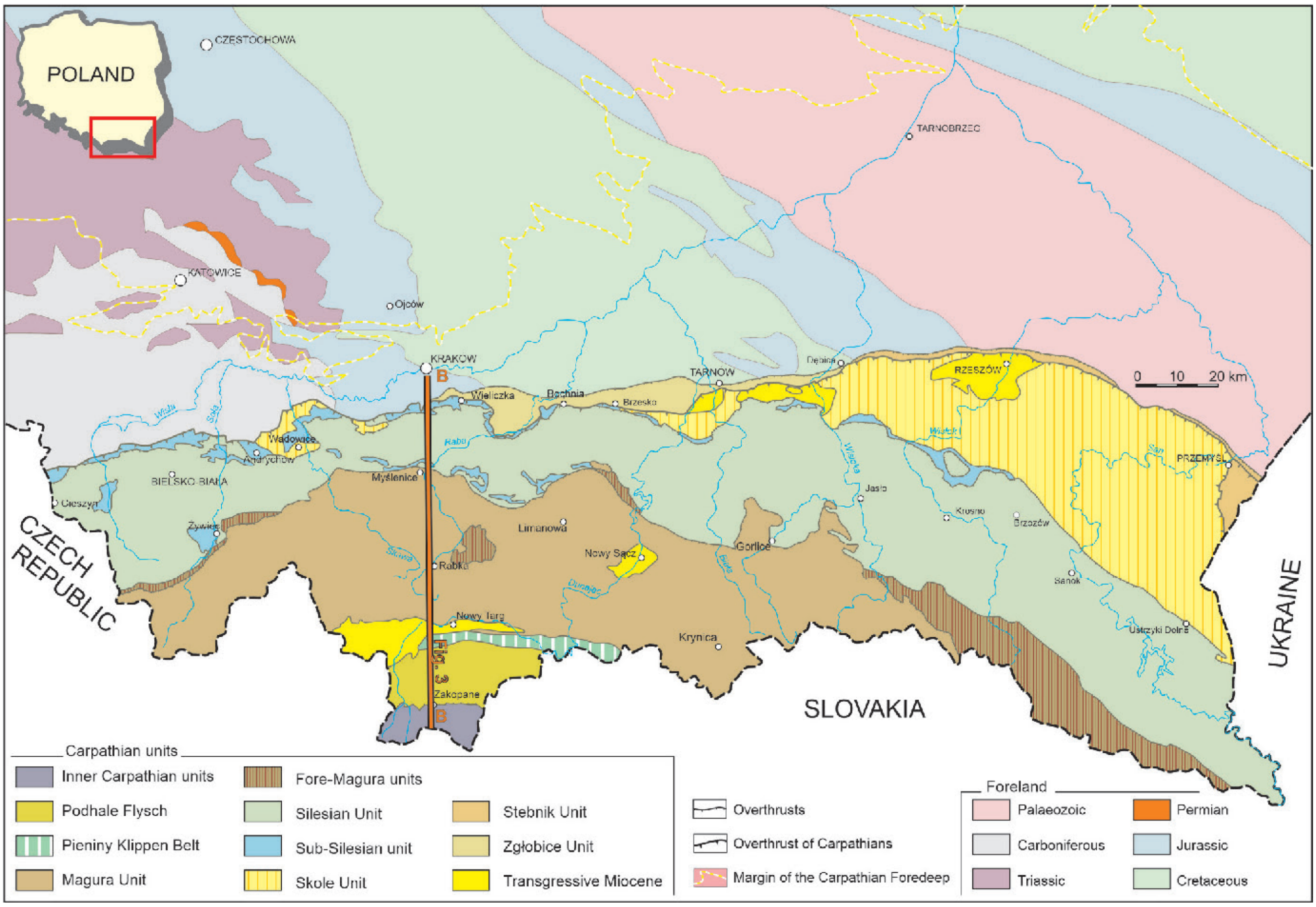


Fig. 4. Geological map of the Polish Carpathians and Foreland (after Żytko *et al*., 1989; simplified)

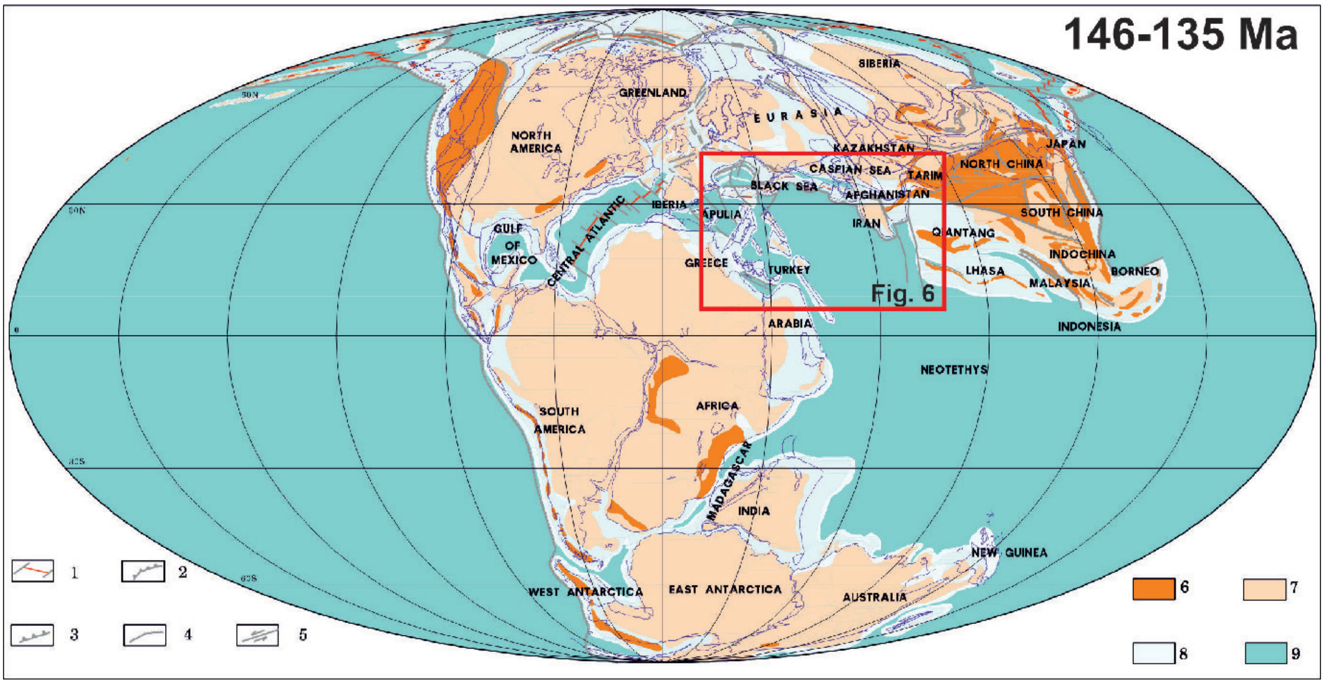


Fig. 5. Global plate tectonic map of latest Jurassic–earliest Cretaceous. Explanations: 1 – oceanic spreading center and transform faults; 2 – subduction zone; 3 – thrust fault; 4 – normal fault; 5 – transform fault; 6 – mountains; 7 – landmass; 8 – shallow sea and slope; 9 – deep ocean basin (from Golonka, 2000; modified)

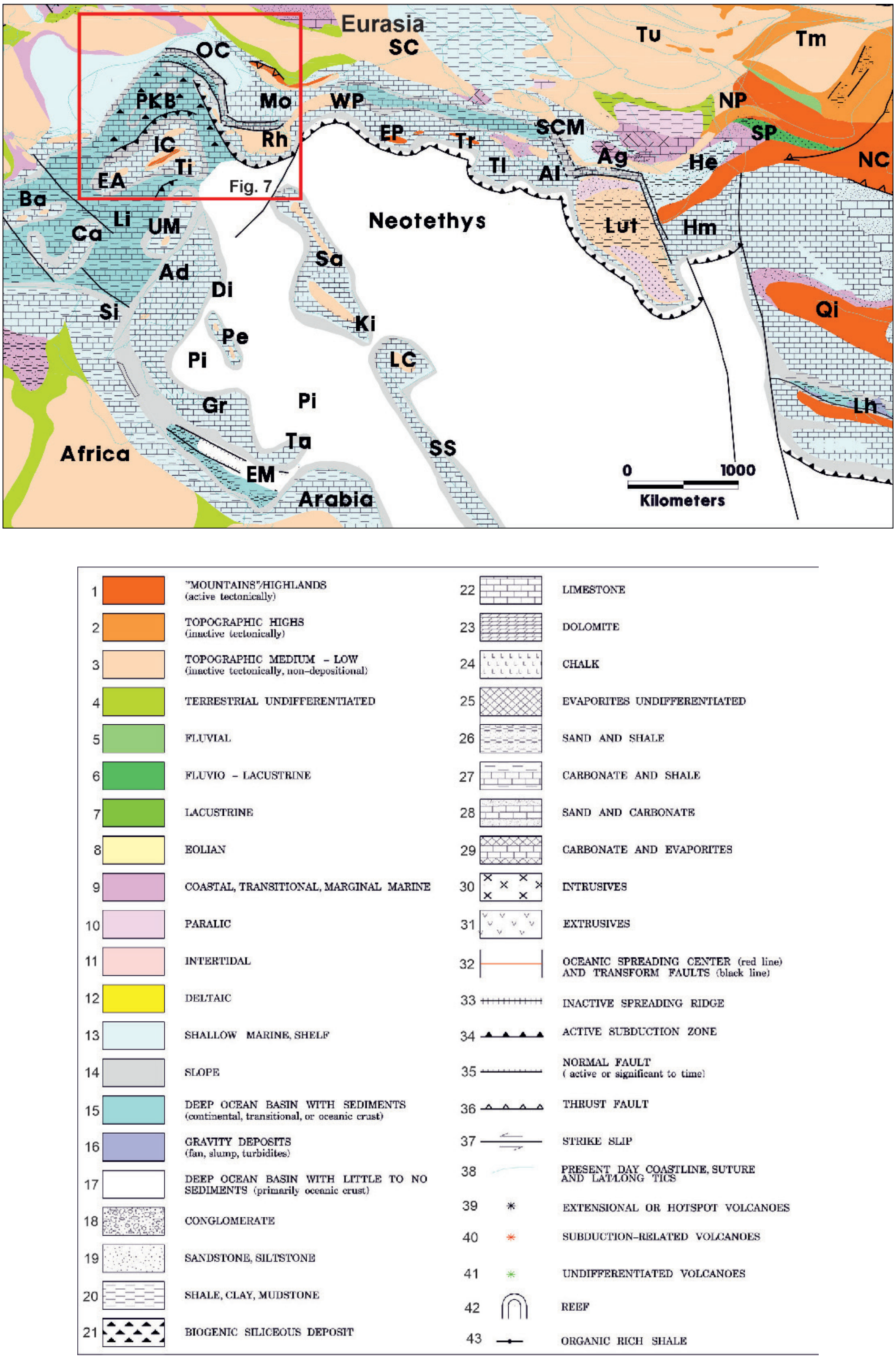


Fig. 6. Plate tectonic, palaeoenvironment and lithofacies map of the western Tethys, Central Atlantic and adjacent areas during latest Jurassic–earliest Cretaceous time (after Golonka, 2007a; modified). Abbreviations of oceans and plates names: Ad – Adria (Apulia); Ag – Aghdarband (southern Kopet Dagh); Al – Alborz; Ba – Balearic; Ca – Calabria-Campania; Di – Dinarides; EA –Eastern Alps; EM – Eastern Mediterranean; EP – Eastern Pontides; Gr – Greece; He – Heart; Hm – Helmand; IC – Inner Carpathians; Ki – Kirsehir; LC – Lesser Caucasus; Lh – Lhasa; Li – Ligurian (Piemont) Ocean; Mo – Moesia; NC – North China; NP – North Pamir; OC – Outer Carpathians; PB – Pieniny Klippen Belt Basin; Pe – Pelagonian plate; Pi – Pindos Ocean; Qi – Qiangtang; Rh – Rhodopes; Sa – Sakarya; SC – Scythian; SCM – South Caspian microcontinent; Sl – Sicily; SP – South Pamir; SS – Sanandaj-Sirjan; Ta – Taurus terrane; Ti – Tisa; Tl – Talysh; Tm – Tarim; Tr – Transcaucasus; Tu – Turan; UM – Umbria-Marche; WP – Western Pontides

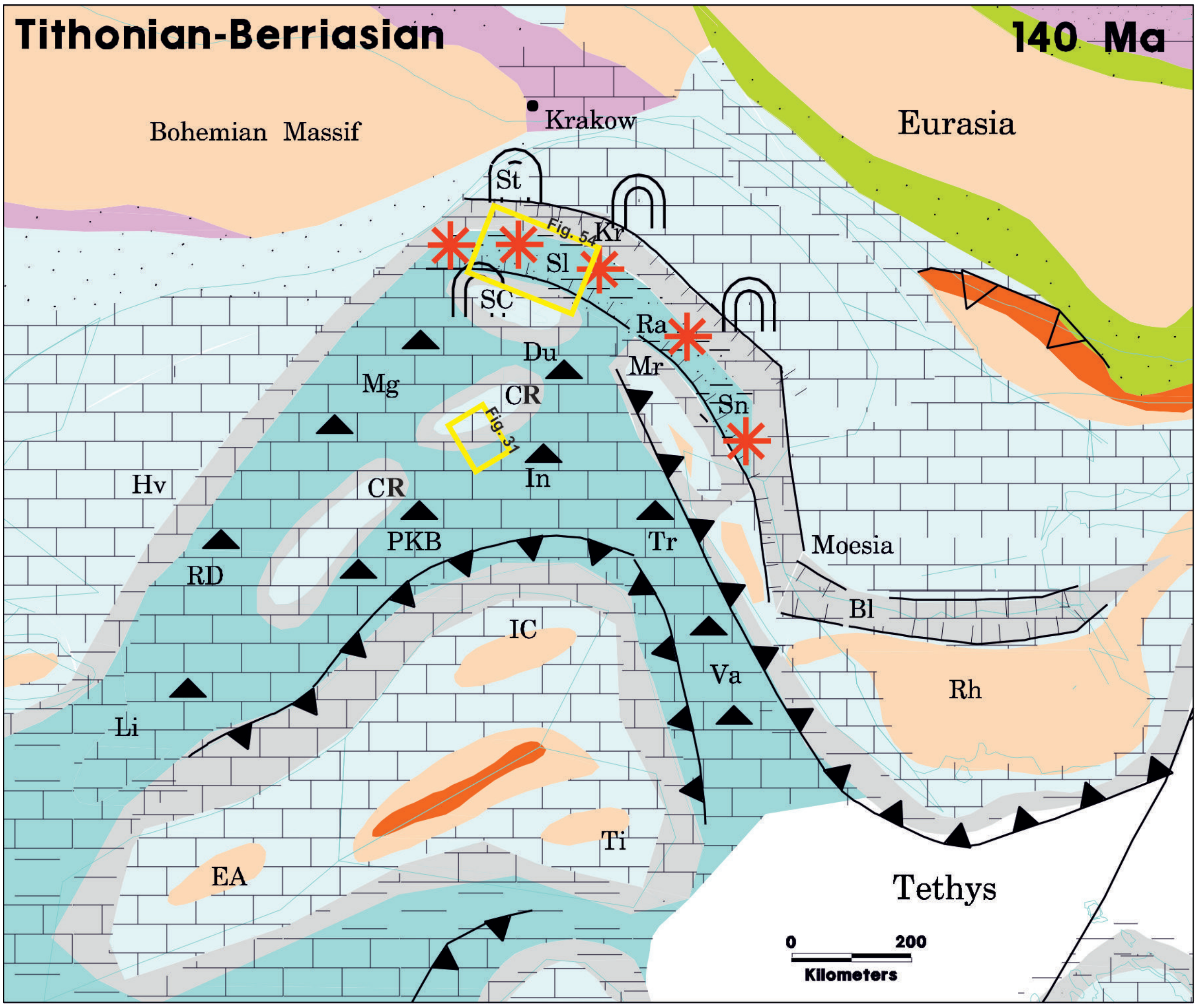


Fig. 7. Palaeoenvironment and lithofacies of the circum-Carpathian area during latest Jurassic–earliest Cretaceous; plates position at 140 Ma (modified from Golonka *et al*., 2006). Abbreviations: Bl – Balkan rift; Cr – Czorsztyn Ridge; Du – Dukla Basin; EA – Eastern Alps; Hv – Helvetic shelf; IC – Inner Carpathians; In – Inačovce-Kričevo zone; Kr – Kruhel Klippe; Li – Ligurian Ocean; Mg – Magura Basin; Mr – Marmarosh Massif; PKB – Pieniny Klippen Belt Basin; Ra – Rakhiv Basin; RD – Rheno Danubian Basin; Rh – Rhodopes; SC – Silesian Ridge (Cordillera); Sl – Silesian Basin; Sn – Sinaia Basin; St – Štramberk Klippe; Ti – Tisa plate; Tr – Transilvanian Ocean; Va – Vardar Ocean. Explanations of colors and symbols – see Fig. 6

Outer Flysch Carpathians

The Outer Flysch Carpathians are built up of a stack of nappes and thrust sheets spreading along the Carpathians, built mainly of up to six kilometers thick continual flysch sequences, representing the time span from Jurassic to Early Miocene. All the Outer Carpathian nappes are overthrust onto the southern part of the North European platform covered by the autochthonous Miocene deposits of the Carpathian Foredeep on the distance of 70 km at least (Książkiewicz, 1977; Pescatore & Ślączka, 1984) (Fig. 4). Boreholes and seismic data indicate that the distance of the Carpathian overthrust was at least 60 km. During overthrusting movement the northern Carpathians nappes became uprooted from the basement and only their basinal parts were preserved. The succession of the nappes from the lowest to the highest is as follows (concordant to our trip – from north to south): Skole (Skiba) Nappe (mainly easternmost part of Carpathians), Subsilesian Nappe, Silesian Nappe, Fore-Magura group of nappes and Magura Nappe (Fig. 8). We discuss here main units only.

The **Subsilesian Nappe** underlies tectonically the Silesian Nappe. In the western sector of the West Carpathians both nappes are thrust over the Miocene molasse of Carpathian Foredeep and in the eastern sector they are thrust over the Skole Nappe. This nappe consists Upper Cretaceous–Paleogene flysch deposits.

The **Silesian Nappe** occupies central part of the Outer Carpathians, pinching out below the most internal nappes. Sedimentary facies of the Silesian Nappe represent continuos succession of deposits of age interval from Late Jurassic to Early Miocene. The oldest sediments of the Silesian are known only in Moravia and Silesia areas in the

Western Carpathians. They are represented by the Kimmeridgan-Lower Tithonian dark grey, calcareous mudstones (Lower Cieszyn Shales) which begin euxinic cycle that lasted without major interruption till Albian. The Silesian and Subsilesian basins have been connected during their sedimentation period.

The **Magura Nappe** is an innermost and largest tectonic unit of the Western Carpathians thrust over the various unit of the Fore-Magura group of nappes and of the Silesian Nappe. To the south it is in the tectonic contact with the PKB that separates it from the Inner Carpathians. The oldest Jurassic-Lower Cretaceous rocks are only found in this part of the Magura basin which was incorporated into the PKB (i.e. the Grajcarek Unit) (Birkenmajer, 1977).

The Outer Carpathian rift had developed with the beginning of calcareous flysch sedimentation (so-called Cieszyn beds). The Western Carpathian Silesian Basin probably extended in the Eastern Carpathian (Sinaia or "black flysch") as well as to the Southern Carpathian Severin zone (Săndulescu, 1988). The remnants of carbonate platforms (Olszewska & Wieczorek, 2001) with reefs (Štramberk-type limestones) along the margin of Silesian Basin were results of the fragmentation of the European platform in this area. The Silesian Ridge (= exotic cordillera) separated the Silesian and Magura basins (Golonka *et al.*, 2000). During the late Tithonian and Early Cretaceous opening of the western part of the Silesian basin alkaline magma (teschenites association rocks) intruded the flysch deposits (Lucińska-Anczkiewicz *et al.*, 2002).

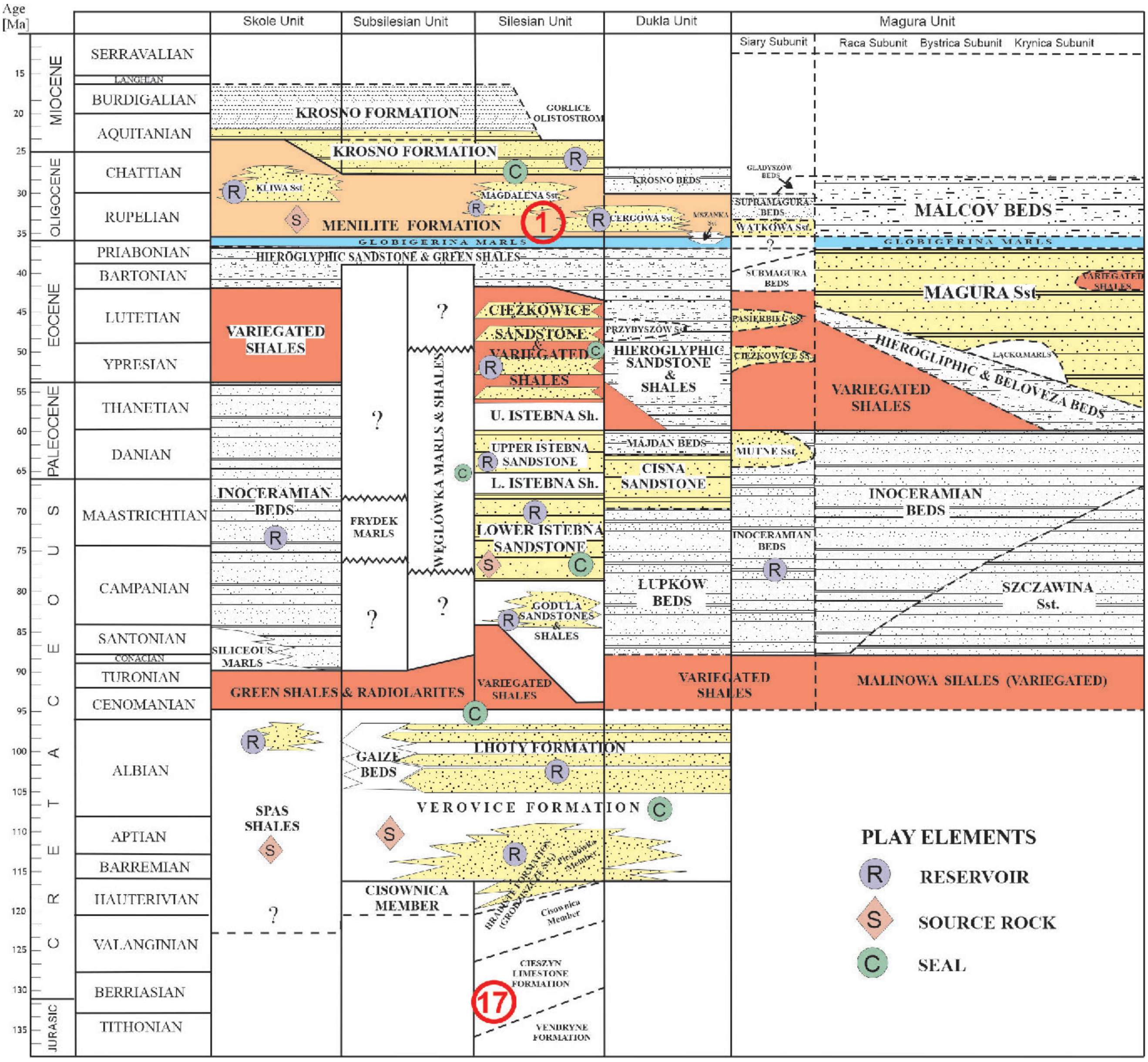


Fig. 8. Simplified lithostratigraphy of the Outer Polish Carpathians (after Koszarski *et al.*, 1985; Dziadzio *et al.*, 2001; modified) with the position of field trip stops

General history of the PKB

(*Michał Krobicki, Jan Golonka*)

The Northern Carpathians are subdivided into an older range known as the Inner Carpathians and the younger one, known as the Outer or Flysch Carpathians. The PKB is situated at the boundary of these two ranges (Figs 2–4). The Inner Carpathians nappes contact along a Tertiary strike-slip boundary with PKB.

The relationship between PKB and the Magura Nappe changes along the PKB strike. In the Vah and Orava valleys these two units are divided by the Miocene sub-vertical strike-slip fault and both units are involved in the complex flower structure. Present day confines of the PKB are strictly tectonic. They may be characterized as a (sub)vertical faults and shear zones, along which a strong reduction of space of the original sedimentary basins took place. The NE-SW striking faults accompanying the PKB have the character of lateral slips. It is indicated by the presence of flower structures on the contact zone of the Magura Unit and the PKB, or by the structural asymmetry of the Inner Carpathian Paleogene Basin.

The tectonic character of the Polish section of PKB is mixed. Both the strike slip and thrust components occur here (e.g. Książkiewicz, 1977; Golonka & Rączkowski, 1984; Birkenmajer, 1986; Ratschbacher *et al*., 1991; Jurewicz, 1994, 1997; Nemčok & Nemčok, 1994). In general the subvertically arranged Jurassic-Lower Cretaceous basinal facies display the tectonics of the diapir character originated in the strike-slip zone between two plates. The ridge facies are often uprooted and display thrust or even nappe character. The Niedzica Succession is thrust over the Czorsztyn Succession, while the Czorsztyn Succession is displaced and thrust over the Grajcarek Unit (e.g., Książkiewicz, 1977; Golonka & Rączkowski, 1984; Jurewicz, 1997). The Grajcarek Unit is often thrust over the Krynica Sub-Unit of the Magura Nappe. The Upper Cretaceous–Paleogene flysch sequences of the Złatne Furrow (Golonka & Sikora, 1981) are often thrust over the various slope and ridge sequences. In the East Slovakian section of the PKB, the back-thrusts of the Magura Nappe onto PKB, as well as PKB onto the Central Carpathian Paleogene, are commonly accepted (see Nemčok, 1990; Lexa *et al*., 2000). The PKB tectonic components of different age, strike-slip, thrust as well as toe-thrusts and olistostromes mixed together, are giving the present-day melange character of the PKB, where individual tectonic units are hard to distinguish.

The PKB is composed of several successions of mainly deep and shallower-water limestones, covering a time span from the Early Jurassic to Late Cretaceous (Andrusov, 1938, 1959; Andrusov *et al*., 1973; Birkenmajer, 1958b, 1977, 1986, 1988; Mišík, 1994; Golonka & Krobicki, 2001, 2004). This strongly tectonized structure is about 600 km long and 1–20 km wide, stretching from Vienna to the West, to Romania to the East (Fig. 2).

During the Jurassic and Cretaceous within the Pieniny Klippen Basin the submarine Czorsztyn Ridge (= “pelagic swell” of Mišík, 1994, mainly so-called Czorsztyn Succession) and surrounding zones formed an elongated structure with domination of pelagic type of sedimentation (Birkenmajer, 1977, 1986; Mišík, 1994; Michalík & Reháková, 1995; Aubrecht *et al*., 1997; Plašienka, 1999; Wierzbowski *et al*., 1999; Golonka & Krobicki, 2001; 2004). The Pieniny Klippen Basin trends SW to NE (e.g. Aubrecht & Túnyi, 2001; see discussion in Golonka & Krobicki, 2001, 2004) (Fig. 7). Its deepest part shows the presence of deep water Jurassic-Early Cretaceous deposits (pelagic limestones and radiolarites) of Złatna Unit (Sikora, 1971; Golonka & Sikora, 1981; Golonka & Krobicki, 2002) later described also as Ultra-Pieniny Succession (Birkenmajer, 1988; Birkenmajer *et al*., 1990) or Vahicum (e.g. Plašienka, 1999). Somewhat shallower sedimentary zones known as the Pieniny, Branisko (Kysuca) successions have been located close to central furrow. Transitional slope sequences between basinal units and ridge units are known as Czertezik and Niedzica successions (Podbiel and Pruské successions in Slovakia) near the northern (Czorsztyn) Ridge, and Haligovce-Nižná successions near the southern so-called Exotic Andrusov Ridge (Birkenmajer, 1977, 1986, 1988; Aubrecht *et al*., 1997; Wierzbowski *et al*., 2004). The strongly condensed Jurassic-Early Cretaceous pelagic cherty limestones (Maiolica-type facies) and radiolarites of the Grajcarek Unit were also deposited in northwestern Magura Basin.

Palinspastic reconstruction of the PKB Basin indicates occurrence of submarine ridge during the whole Jurassic and Cretaceous times. This so-called Czorsztyn Ridge, an elongated structure, subdivided Pieniny and Magura basins within the Carpathian part of the northernmost Tethyan Ocean (Figs 5–7) (comp. Golonka, 2004, 2007a, 2007b with references cited therein). Its SW-NE orientation and location within the Tethyan Ocean is interpreted by means of palaeomagnetic data, relationship of sedimentary sequences and palaeoclimate (see discussion in Golonka & Krobicki, 2001, see also Aubrecht & Túnyi, 2001; Lewandowski *et al*., 2005; Grabowski *et al*., 2008). The basins divided by the Czorsztyn Ridge were dominated by a pelagic type of sedimentation. The deepest part of the PKB Basin is well documented by deep water Jurassic-Early Cretaceous deposits (radiolarites and pelagic *Maiolica*-type cherty limestones) (Birkenmajer, 1979, 1986; Golonka & Sikora, 1981; Golonka & Krobicki, 2004; Krobicki *et al*., 2006) of the so-called Branisko and Pieniny successions. The transitional, shallower sequences, which primary occupied slopes between deepest basinal units and the Czorsztyn Ridge are known as Czertezik and Niedzica successions, and the shallowest zone is Czorsztyn Succession which primary occupied SE slope of the Czorsztyn Ridge (Birkenmajer, 1986; Golonka & Krobicki, 2004; Krobicki & Golonka, 2006).

The **oldest Jurassic** rocks known only from the Ukrainian and Slovakian part of the PKB (Krobicki *et al*., 2003; Schlögl *et al*., 2004; Wierzbowski *et al*., 2012, 2021) consist of different type of *Gresten*-like clastic sediments with intercalations of *Gresten*-like dark/black fossiliferous limestones with brachiopods and grypheoids (?Hettangian-?Sinemurian) (Schlögl *et al*., 2004 with literature). However, Pliensbachian-Lower Bajocian *Bositra* (“*Posidonia*”) black shales with spherosiderites (Skrzypny Shale Formation in local, formal nomenclature, see Birkenmajer, 1977) as well as dark marls and spotty limestones of widespread Tethyan *Fleckenkalk*/*Fleckenmergel* facies, indicate the oxygen-depleted conditions (Birkenmajer, 1986; Tyszka, 1994, 2001) (Fig. 9).

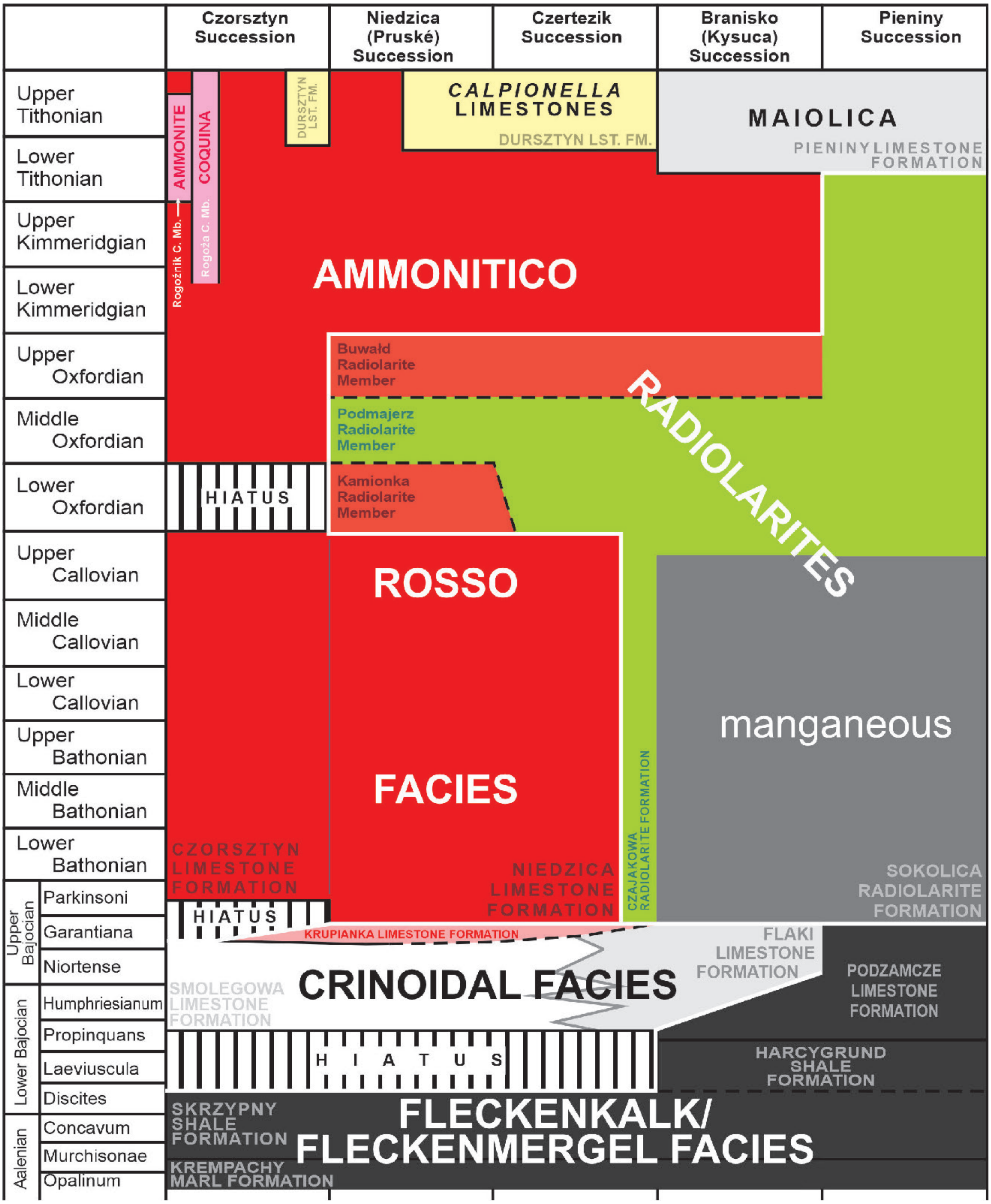


Fig. 9. Stratigraphical correlation between Jurassic lithofacies (lithostratigraphic units) of the Pieniny Klippen Belt successions (after Wierzbowski *et al*., 2004; supplemented by Krobicki & Wierzbowski, 2004)

One of the most rapidly change of sedimentation/palaeoenvironments within the PKB basins took place during late Early Bajocian when well-oxygenated multicoloured crinoidal limestones replaced dark and black sedimentation.

The origin of the above mentioned Czorsztyn Ridge was connected with this Bajocian postrift geotectonic reorganization (Golonka *et al.*, 2003; Krobicki, 2006, 2009).

One of the most important geotectonic element within Western Carpathians basins was the Czorsztyn Ridge (Swell), which originated during the Middle Jurassic (Early Bajocian) time. Palaeogeographicaly it has been the main object which separated, between the Middle Jurassic to the Late Cretaceous times, two large Carpathians basins, the Magura Basin on NW side and the Pieniny Basin on SE side. Therefore, the precise dating of its origin and first uplift is crucial for recognition of its geodynamic significance. Drastic change of sedimentation from dark/black shales of oxygen-poor environments (latest Pliensbachian–earliest Bajocian) to white/light grey crinoidal limestones of well oxygenated regimes, which presently directly overlie shales, were separated by significance stratigraphical hiatus (Fig. 10). It was biostratigraphicaly perfectly dated by ammonites collected from the basal part of crinoidal limestones in several outcrops of the Polish part of the PKB (Krobicki & Wierzbowski, 2004).

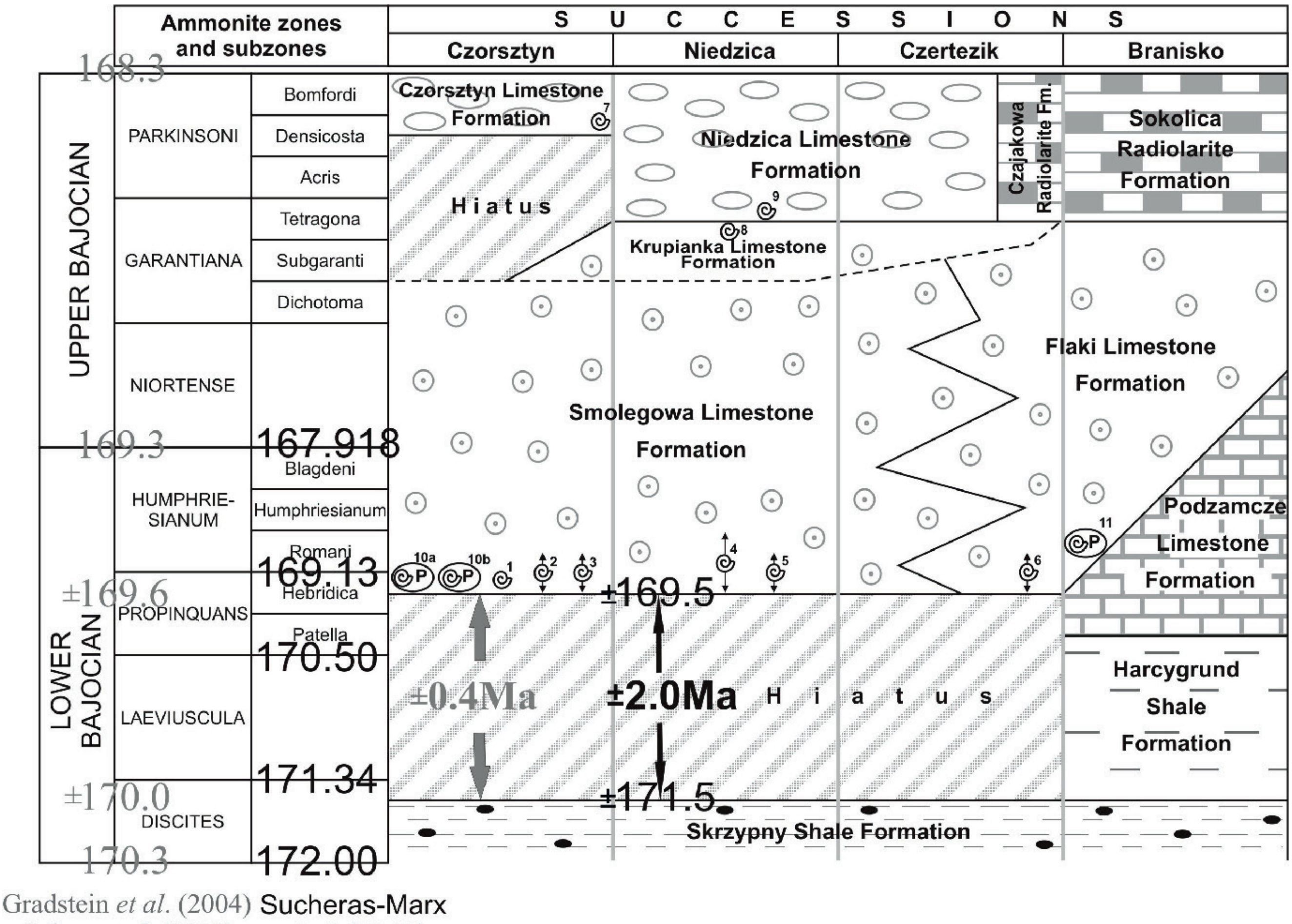


Fig. 10. Lithostratigraphical scheme of the klippen successions of the Pieniny Klippen Belt (after Krobicki & Wierzbowski, 2004, slightly modified; lithostratigraphical units after Birkenmajer, 1977) with data of duration of the Early Bajocian. Numeration indicates outcrops with ammonites (described in Krobicki & Wierzbowski, 2004): 1, 2, 7 – Czorsztyn-Sobótka; 3 – Krupianka; 4, 8, 9 – Niedzica-Podmajerz; 5 – Czajakowa Skała; 6 – Wysokie Skałki; ammonites in phosphatic concretions: 10a – Falsztyn; 10b – Czorsztyn-Sobótka; 11 – Flaki. Lithology of lithostratigraphical units: Skrzypny Shale Formation – black shales with spherosiderites; Harcygrund Shale Formation – dark spotty shales; Podzamcze Limestone Formation – dark spotty limestones; Smolegowa Limestone Formation – white crinoidal limestones; Flaki Limestone Formation – grey crinoidal limestones; Krupianka Limestone Formation – red crinoidal limestones; Czorsztyn and Niedzica Limestone formations – red nodular limestones; Czajakowa Radiolarite Formation – green radiolarites; Sokolica Radiolarite Formation – black manganeous radiolarites. Chronostratigraphic data: grey Times New Roman – Gradstein *et al*. (2004) and Cohen *et al*. (2013); black arial – Sucheras-Marx *et al*. (2013)

When we try to estimate absolute time of this uplift event (= origin of the Czorsztyn Ridge) we can use two proposed scales of duration of the Early Bajocian. First one is described and illustrated by Gradstein *et al*. (2004) and Cohen *et al*. (2013), which suggest 2 Ma for the whole Bajocian, and by this reason the hiatus has about 0.4 Ma only. Second idea is based on estimation of the duration of this sub-stage, based on a cyclostratigraphic analysis of the carbonate content from the Chaudon–Norante section (Subalpine Basin, France) (Sucheras-Marx *et al*., 2013), which indicates that the Early Bajocian only lasted c. 4.082 Ma. Using these authors calibration of duration of the Early Bajocian ammonite zones (the Discites zone lasted 0.66 Ma, the Laeviuscula zone 0.84 Ma, the Propinquans zone 1.37 Ma, and the Humphriesianum zone 1.22 Ma) we can conclude that the hiatus, which corresponds with time necessary for origin/uplift of the Czorsztyn Ridge, is about 2 Ma. From geotectonical processes point of view such calculation is more probably (Krobicki, 2018) (Fig. 10).

The central Atlantic (Withjack *et al*., 1998) and Alpine Tethys went into a drifting stage during the **Middle Jurassic**. The oldest oceanic crust in the Ligurian-Piemont Ocean was dated as late as the Middle Jurassic in the southern Apennines and in the Western Alps (see Ricou, 1996 and literature cited therein). Bajocian oceanic spreading of the Alpine Tethys documented by isotopic methods (Bill *et al*., 2001) fit well with the Pieniny data (Winkler & Ślączka, 1994), which well correspond to the supposed opening of the Ligurian-Penninic Ocean. Crinoidal limestones were developed in more elevated parts of the Pieniny Klippen Basin (Czorsztyn, Niedzica and Czertezik successions), and were redistributed to deeper-water Branisko Succession as the grey crinoidal cherty limestones. Sedimentation of still younger (since latest Bajocian) red nodular *Ammonitico Rosso*-type limestones was effect of Meso-Cimmerian vertical movements which subsided Czorsztyn Ridge and produced tectonically differentiated blocks as well as accompanied by the formation of neptunian dykes and scarp-breccias (e.g. Birkenmajer, 1986; Aubrecht *et al*., 1997; Wierzbowski *et al*., 1999; Aubrecht, 2001; Aubrecht & **Túnyi, 2001;** Krobicki, 2006; Krobicki & Golonka, 2006).

The **Late Jurassic** (Oxfordian-Kimmeridgian) history of the PKB reflects strongest facial differentiation within sedimentary basin where mixed siliceous-carbonate sedimentation took place. The formation of limestones of the *Ammonitico-rosso* type was mostly related with existence of elevated part of sea bottom (Czorsztyn Ridge and its slopes), whereas deposition of radiolarites (Birkenmajer, 1977, 1986; Mišík, 1999) took place in deeper parts of the bordering basins. The main phase of this facial differentiation took place later, mainly during Oxfordian times when the greatest deepening effect is indicated by widespread Oxfordian radiolarites which occur in the all basinal successions, whereas the shallowest zone (Czorsztyn Succession) is completely devoid of siliceous intercalations at that time. Oxfordian radiolarites are typical for transitional (Niedzica and Czertezik) successions and strictly basinal parts of the basin (Branisko and Pieniny successions). Similar compositions of facies are well known in several European Alpine regions (e.g. Betic Cordillera, Southern Alps, Apennine, Karavanke, and Ionian Zone). These regions, together with PKB basins formed the so-called Alpine Tethys (Golonka, 2004).

During the **latest Jurassic–Early Cretaceous** (Tithonian-Berriasian), the Czorsztyn Succession included hemipelagic to pelagic organogenic carbonate deposits of medium depth, for example white and creamy *Calpionella*-bearing limestones. Several tectonic horsts and grabens were formed, rejuvenating some older, Eo- and Meso-Cimmerian faults (Birkenmajer, 1986; Krobicki, 1996a). Such features resulted from the intensive Neo-Cimmerian tectonic movements and are documented by facies diversification, hardgrounds and condensed beds with ferromanganese-rich crusts and/or nodules, sedimentary-stratigraphic hiatuses, sedimentary breccias and/or neptunian dykes (Birkenmajer, 1958a, 1975, 1986; Michalík & Reháková, 1995; Krobicki, 1996a; Aubrecht *et al*., 1997; Krobicki & Słomka, 1999; Golonka & Krobicki, 2002; Plašienka, 2002; Golonka *et al*., 2003; Krobicki *et al*., 2006). In the same time within deeper successions (mainly Branisko and Pieniny ones) cherty limestone of *Maiolica*-type (=*Biancone*) facies deposited. It is one of the famous, widespread Tethyan facies well known both from the Alpine and the Apennine regions (Pszczółkowski, 1987; Wieczorek, 1988). In whole western Tethys this facies originated mainly in deep basins (above CCD but above ACD levels) but also on submarine elevations or drowned platforms and around the Jurassic/Cretaceous boundary reflects the greatest facies unification in this ocean (e.g., Winterer & Bosellini, 1981; Wieczorek, 1988).

Late Cretaceous pelagic deposits with the youngest part developed as *Scaglia Rossa* pelagic, foraminiferal, multicoloured green/variegated/red marl deposits (= *Couches Rouge* = *Capas Rojas*) deposited during the latest, third episode of evolution of the Pieniny Klippen Basin (Birkenmajer, 1986, 1988; Bąk K., 2000), when unification of sedimentary facies took place within all successions (Albian-Coniacian). Still younger are flysch and/or flyschoidal facies (Santonian-Campanian) (i.a. Birkenmajer, 1986; Mišík, 1994; Aubrecht *et al*., 1997; Birkenmajer & Jednorowska, 1983a, 1984, 1987a, 1987b; Gasiński, 1991; Birkenmajer & Gasiński, 1992; Bąk, K., 1998; Bąk M., 1999). During this syn-orogenic stage of the development of the PKB Basin these flyschoidal deposits developed as submarine turbiditic wedges, fans and canyon fills (Rawdański, 1978; Birkenmajer, 1986) with several episodes of debris flows with numerous exotic pebbles took place (Late Albian-Early Campanian) (Fig. 11).

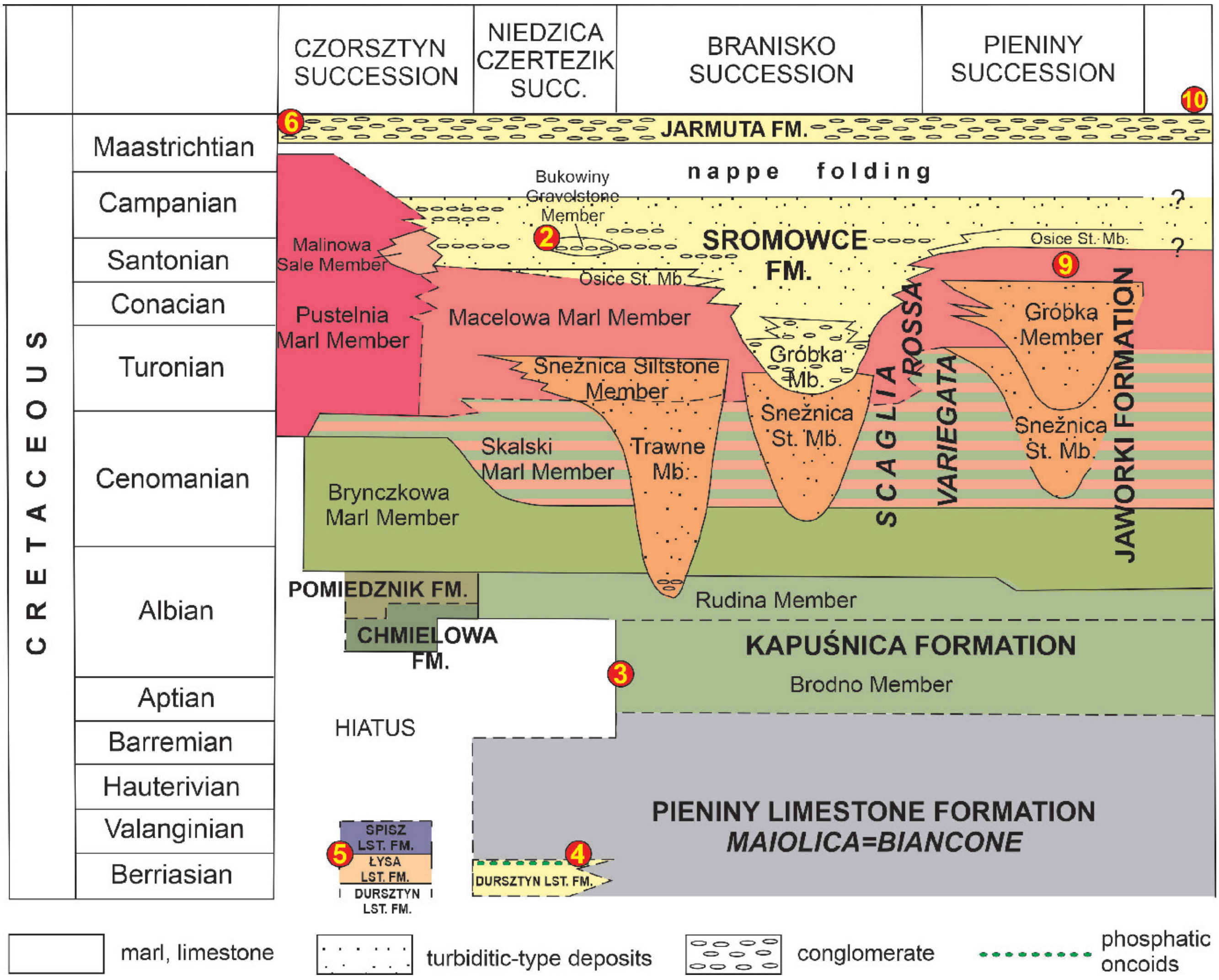


Fig. 11. Detailed stratigraphic table of the Cretaceous rocks of the Czorsztyn, Niedzica (Pruské), Czertezik, Branisko (Kysuca) and Pieniny successions in the Pieniny Klippen Belt in Poland (from Birkenmajer & Jednorowska, 1987, simplified) with locations of field trip stops

The Pieniny Klippen Basin was closed at the **Cretaceous/Paleocene** transition, as effect of strong Late Cretaceous (Subhercynian and Laramian) thrust-folding (Birkenmajer, 1977, 1986, 1988). From south to north folding of the successive nappes, built by Jurassic-Cretaceous deposits of early mentioned sedimentary successions, took place. Simultaneously with this Laramian nappe folding the uppermost Cretaceous (Maastrichtian) fresh-water and marine molasse with exotic material was deposited and Paleocene flysch was continuation of this sedimentary event. They covered with unconformity several klippen nappes folded earlier and this so-called Klippen Mantle was refolded together with them somewhat later.

The second tectonic episode was connected with strong Savian and Styrian (**Early** and **Middle Miocene** respectively) compression, when the Cretaceous nappes, the Klippen Mantle and the new Paleogene deposits were refolded together (Birkenmajer, 1986) and originated system of transverse strike-slip faults. Good visible effect of several tectonic phases of folding and deformations within PKB is geomorphologic view of tectonically isolated klippes of Jurassic and Cretaceous hard rocks surrounding by softer shales, marls and flysch deposits.

The last important event in the PKB was **Middle Miocene (Sarmatian)** volcanism represented by calc-alkaline andesite dykes and sills which cut mainly Paleogene flysch rocks of the Outer Carpathians (Magura nappe) (Małkowski, 1958; Birkenmajer, 1979, 1986, 1988) recently precise dating radiometrically (Birkenmajer & Pécskay, 1999, 2000). They formed so-called Pieniny Andesitic Line (PAL) (Fig. 12).

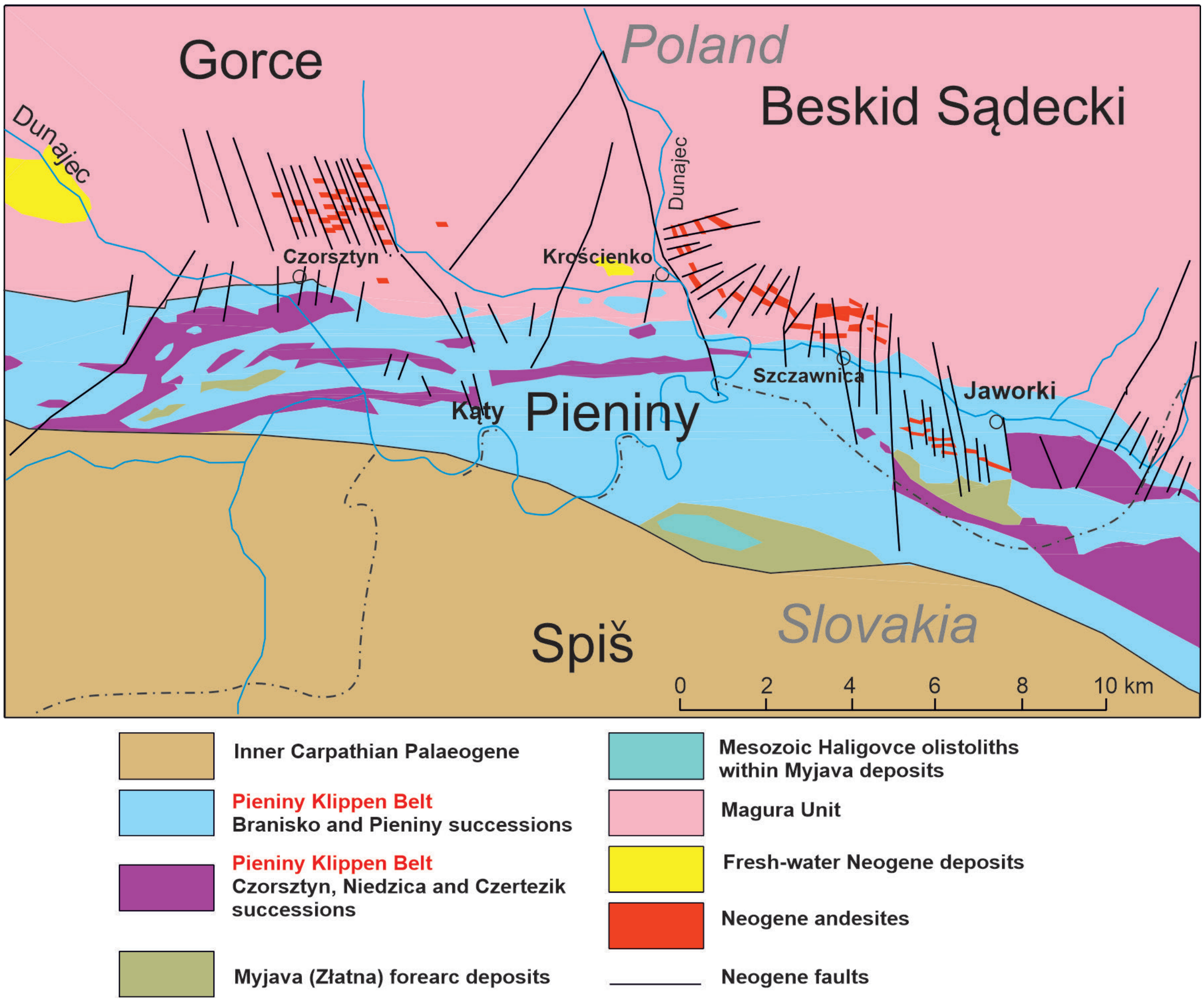


Fig. 12. Geological sketch of the Pieniny Klippen Belt (Polish sector) and surrounding regions (after Birkenmajer, 1979 – simplified)

DESCRIPTION OF THE TRIP

Passage: Kraków – Skrzydlna – Szczawnica

(*Jan Golonka, Michał Krobicki*)

The field trip starts in Krakow AGH Science and Technology University parking lot and leads southward to the Carpathians. In Krakow and its vicinity the Mesozoic rocks of the North European Plate are exposed. The platform is dissected by numerous faults into several horsts and grabens. The grabens are filled with the Miocene Molasse deposits, while horsts elevate the Upper Jurassic rocks. These rocks are represented mainly by Oxfordian cyanobacterial-sponge buildups with associated nodular, chalky and micritic limestones (Matyszkiewicz, 1997). Passing the bridge on Wisła River we can observe the hill of Wawel with the Polish Royal Castle on the top. The Royal Castle was built in X century and remodeled several times. The most important remodeling was done by Queen Bona and her team of Italian architects in XVI century giving the castle its Renaissance character. The Wawel hill is built by the white-weathering Upper Oxfordian massive limestones. These limestones are horst elevated and shaped by karst phenomena. Following southwards the road crosses the Carpathian Foredeep filled with Miocene molasse deposits. Springs of hydrosulphuric mineral waters are connected with the Miocene deposits (Cieszkowski & Ślączka, 2001). These mineral waters are being utilized at spas Mateczny and Swoszowice located within Krakow City limits. After a few kilometers the route passes over the frontal thrust-faults of the Outer Carpathian flysch belt.

Stop 1 – Skrzydlna quarry – olistostrome-bearing succession of the Menilite Beds (Eocene-Oligocene) (Figs 13–26)

(*Marek Wendorff, Krzysztof Starzec, Aneta Siemińska*)

Skrzydlna is a village located about 50 km south-east of Kraków (Fig. 13). In the southern part of the village, near the local road between Skrzydlna and Kasina Mała, there is a quarry in which an almost 200-meter thick succession of the Menilite Beds (Oligocene) is exposed. The Menilite Beds is a unit widespread in the Carpathians (Fig. 14), rich in organic material, and composed mainly of dark brown shales, cherts and siliceous marls with sandstone interbeds. The total thickness of the complex varies from about 100 m in the southern part of their occurrence to maximum 550 m in the northern part (Kuśmierek, 1995). Greater thickness of this unit in some areas are mainly due to the intercalations of sandstone bodies, which sometimes may constitute half of the total thickness of the Menilite Beds. The relatively high content of organic matter, reaching 20% (e.g. Curtis *et al*., 2004), makes this unit the most important source rock of hydrocarbons in the entire Carpathian area. On the other hand, the interbedded thick sandstone complexes tend to be important reservoirs. The Polish Carpathians are one of the oldest oil provinces in the world, where the exploitation of crude oil dates back to the mid-nineteenth century.

The Skrzydlna quarry is located in an area with a complex structural structure. The Menilite Beds outcropping in the quarry together with the overlying Krosno Beds (beyond the quarry's perimeter) form a narrow zone of steeply dipping strata. The geological context of this zone is interpreted in various ways (e.g. Burtan, 1974; Polak, 1999; Cieszkowski *et al*., 2012; Jankowski & Margielewski, 2012). According to the authors of this chapter, this zone belongs to one of the Fore-Magura tectonic units, i.e. the Dukla or Grybów unit (Fig. 15). In the area of Skrzydlna, the Fore-Magura unit extends NW-SE and is located at the front of the Magura nappe, the margin of which here forms a tectonic bay. Further to the SE, this unit is continuing under the cover of the Magura Nappe, as evidenced by drilling data in the area of Słopnice. Still further to SE, the presence of this unit is marked by outcrops in several tectonic windows within the Magura Nappe (Fig. 2). The quarry at Skrzydlna is actively exploited, therefore some details of the unveiling change over time; this guide describes the situation observed in 2017–2019.

The structural development of the Fore-Magura Unit at Skrzydlna is related to the Lanckorona-Żegocina Zone, bordering the Fore-Magura Unit from the north. The origin of this zone has been variously interpreted (e.g. Skoczylas--Ciszewska, 1960; Książkiewicz, 1972; Burtan, 1978; Polak, 1999;

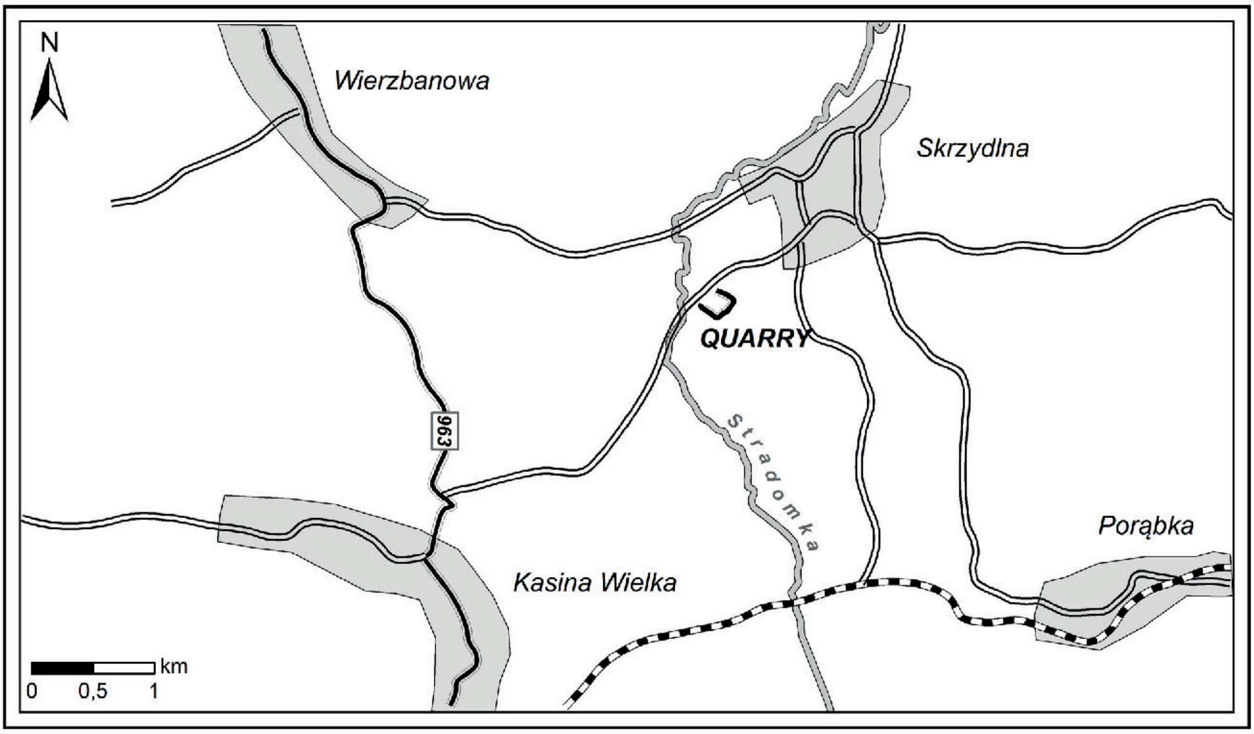


Fig. 13. Location of the Skrzydlna quarry

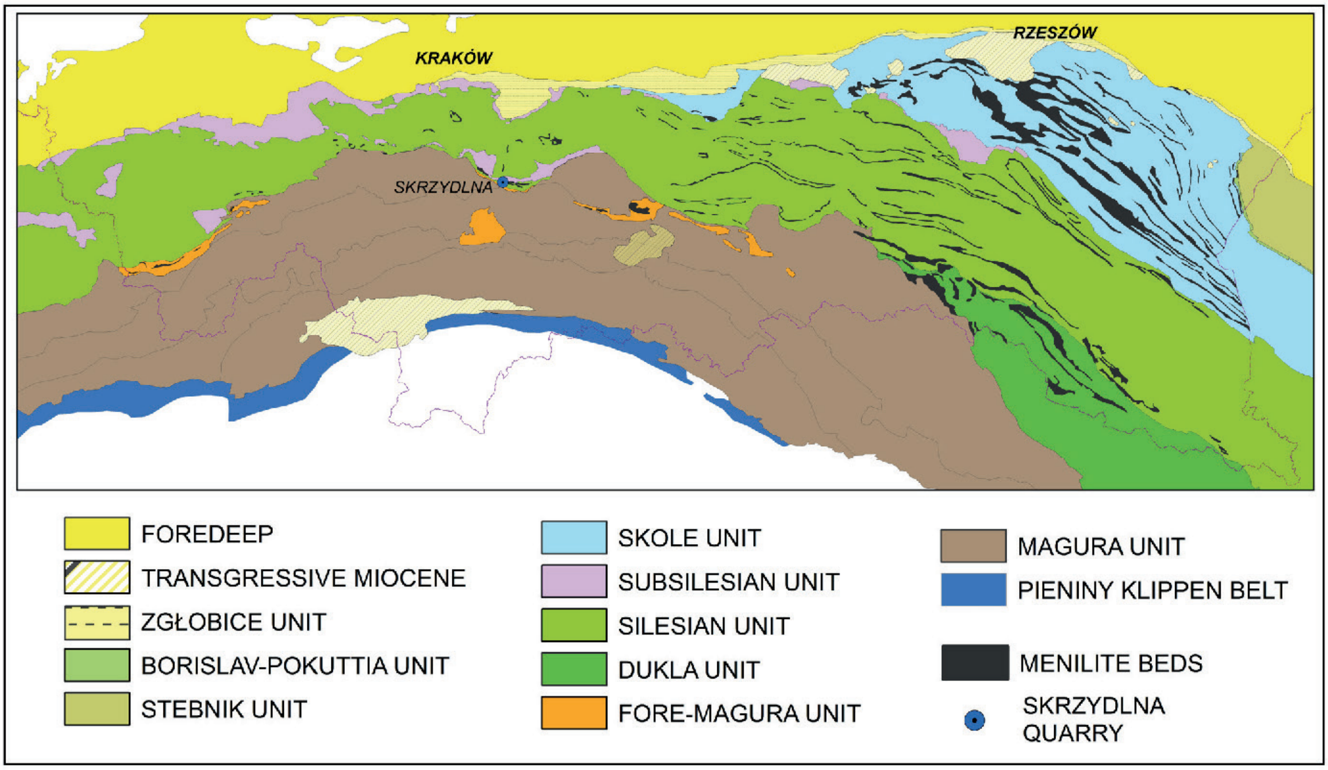


Fig. 14. Occurrence of the Menilite Beds within the Polish segment of the Carpathians

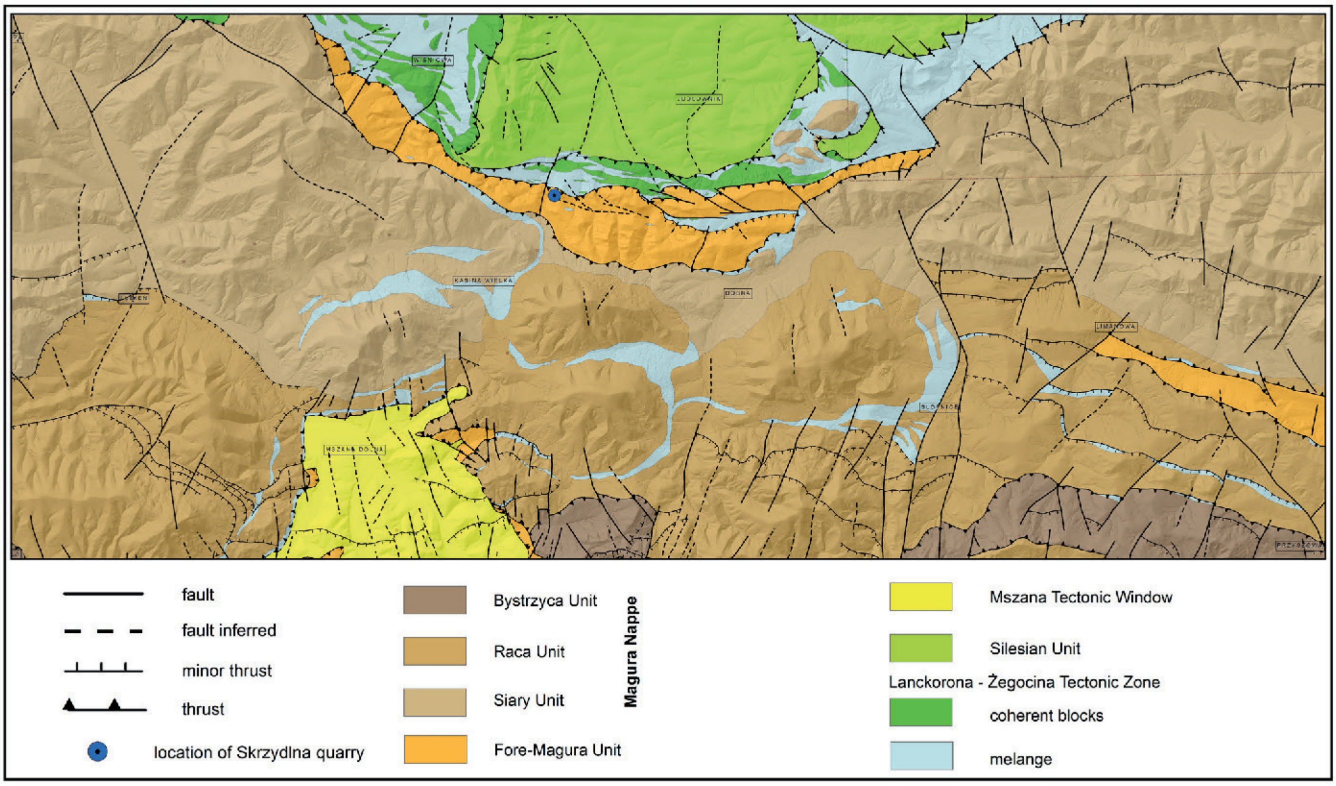


Fig. 15. Regional tectonic sketch with the location of the Skrzydlna quarry (based on Żytko *et al*., 1989, modified)

Jugowiec-Nazarkiewicz & Jankowski, 2001). Detailed mapping studies revealed that it is a zone of tectonic mélange (Fig. 16), composed of Early Cretaceous elements of provenance from the Silesian Tectonic Unit and formations considered typical of the Sub-Silesian series (mainly Węglowiec and Frydek marls, sandstones from Rajbrot, Szydłowiec and Czerwin). The cartographic image shows a series of blocks composed of continuous fragments of stratigraphic successions, which are most often embedded and isolated within a muddy-sandy matrix, rarely in direct contact with each other. According to Golonka *et al*. (2011) the melange in the Lanckorona-Żegocina zone was formed as a result of diapiric migration of less competent rocks along a fault zone. Jugowiec-Nazarkiewicz & Jankowski (2001) and Jankowski (2015) recognize that this is the result of a multi-stage deformation process involving thrusting, followed by strike-slip movements parallel to the strike of the fold structures, and at the final stage – the formation of normal faults. According to the authors of this part of the guide, following the Skrzydlna overthrust origin in the compressional phase of the formation of the Carpathian accretionary prism, this unit was subjected to the influence of strike-slip movements along the Lanckorona-Żegocina zone. The effect of this movement is the arrangement of tectonic elements of the unit, i.e. the presence of fault zones containing tectonic melange and parallel to the strike of the Menilite and Krosno Beds, as well as a steep bedding observed in the quarry, forming a floral structure (Fig. 17). The quarry is actively mined, so some details of the outcrop change over time; this guide describes the situation observed in 2017–2019.

The succession exposed at Skrzydlna is a heterogeneous association of several facies complexes reflecting radical changes in tectonically-controlled sedimentation in the early Oligocene (Polak, 2000; Cieszkowski, 2006; Wendorff *et al*., 2015; Siemińska *et al*., 2020). The oldest part of the exposed succession is a complex of anoxic/dysoxic brown menilite shales, containing a thick layer of sandstone orange in colour and underlying whitish weathered marls and limestones (Fig. 18). Above rests a complex dominated by orange conglomerates (olistostrome succession), evolving upwards into a thinning/fining turbidite sequence (FU-'fining-upwards'; Fig. 19). The orientation of the quarry wall is approximately perpendicular to the strike of beds. These dip steeply and rest in the stratigraphically normal position on the left side of the exposure, gradually changing the orientation to the reversed position in the stratigraphically youngest part of the sequence on the right hand side of the exposure (Fig. 20). Today, following rapid progression of the mining work, the recently exposed face does not show this pattern and the steep bedding is more uniform instead.

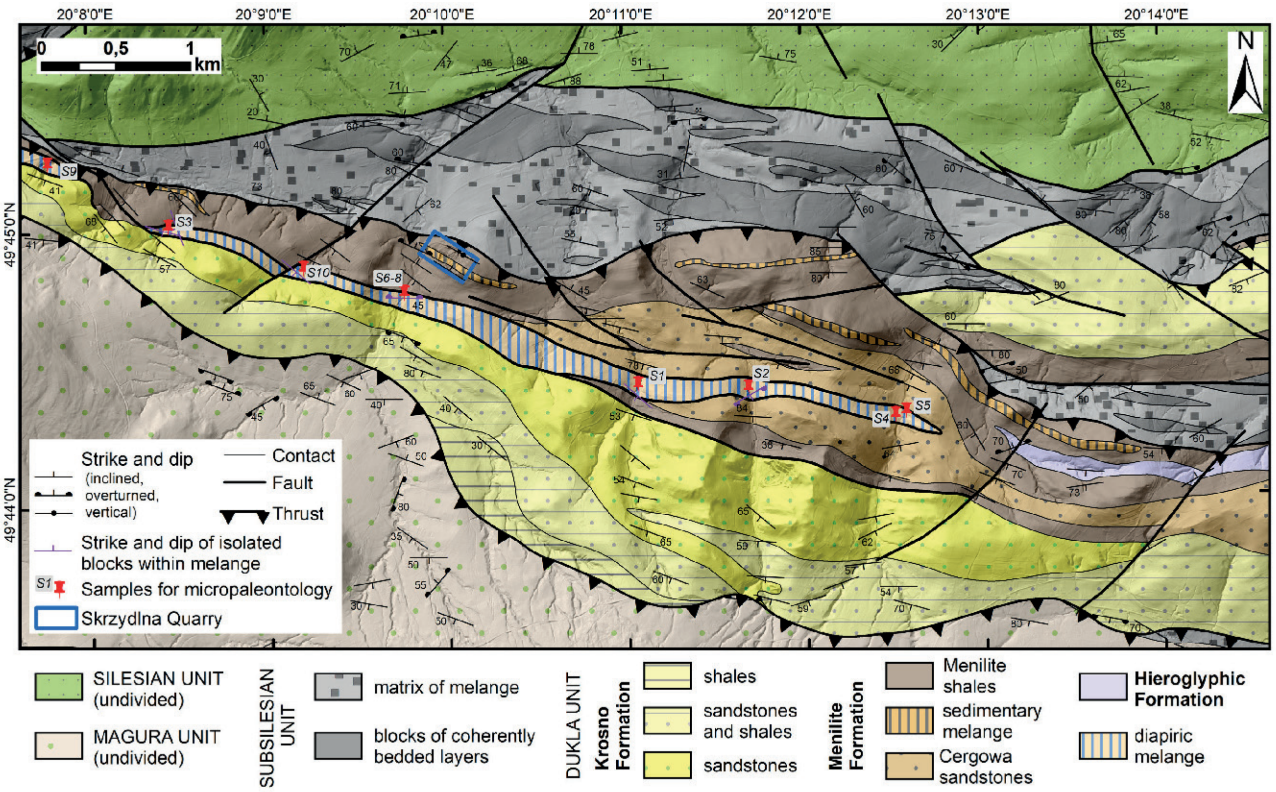


Fig. 16. Geological map of Skrzydlna region

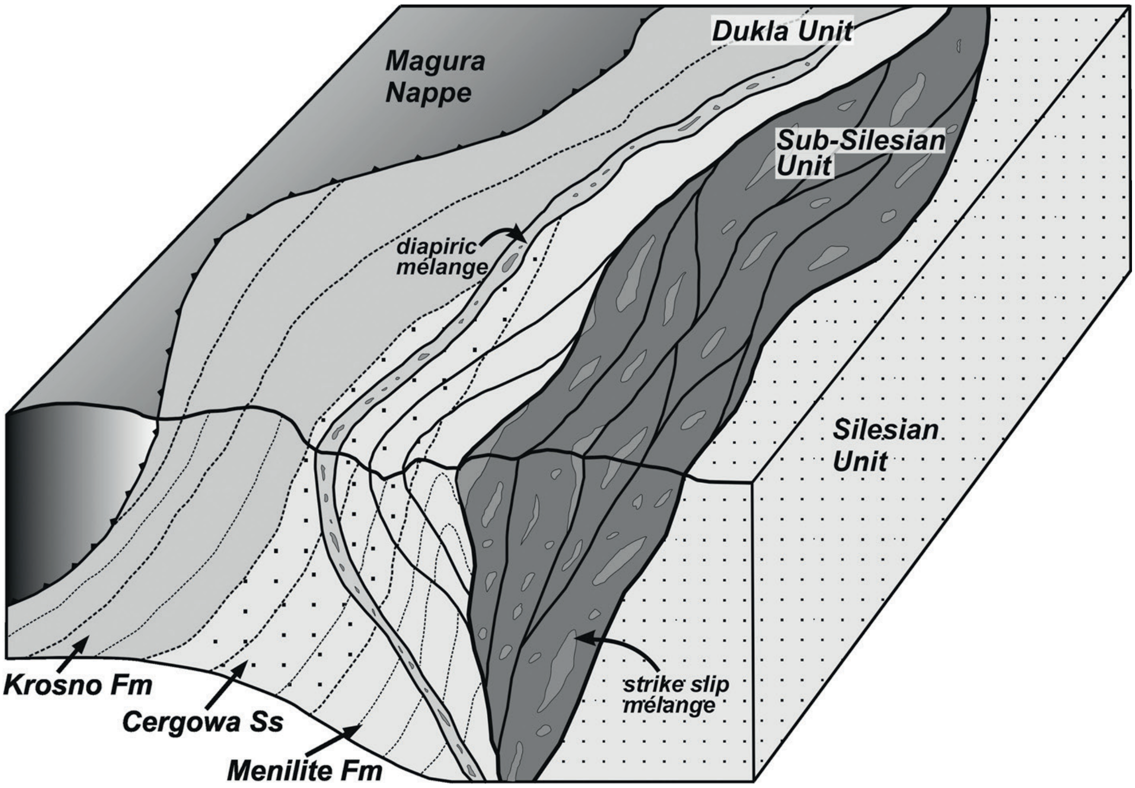
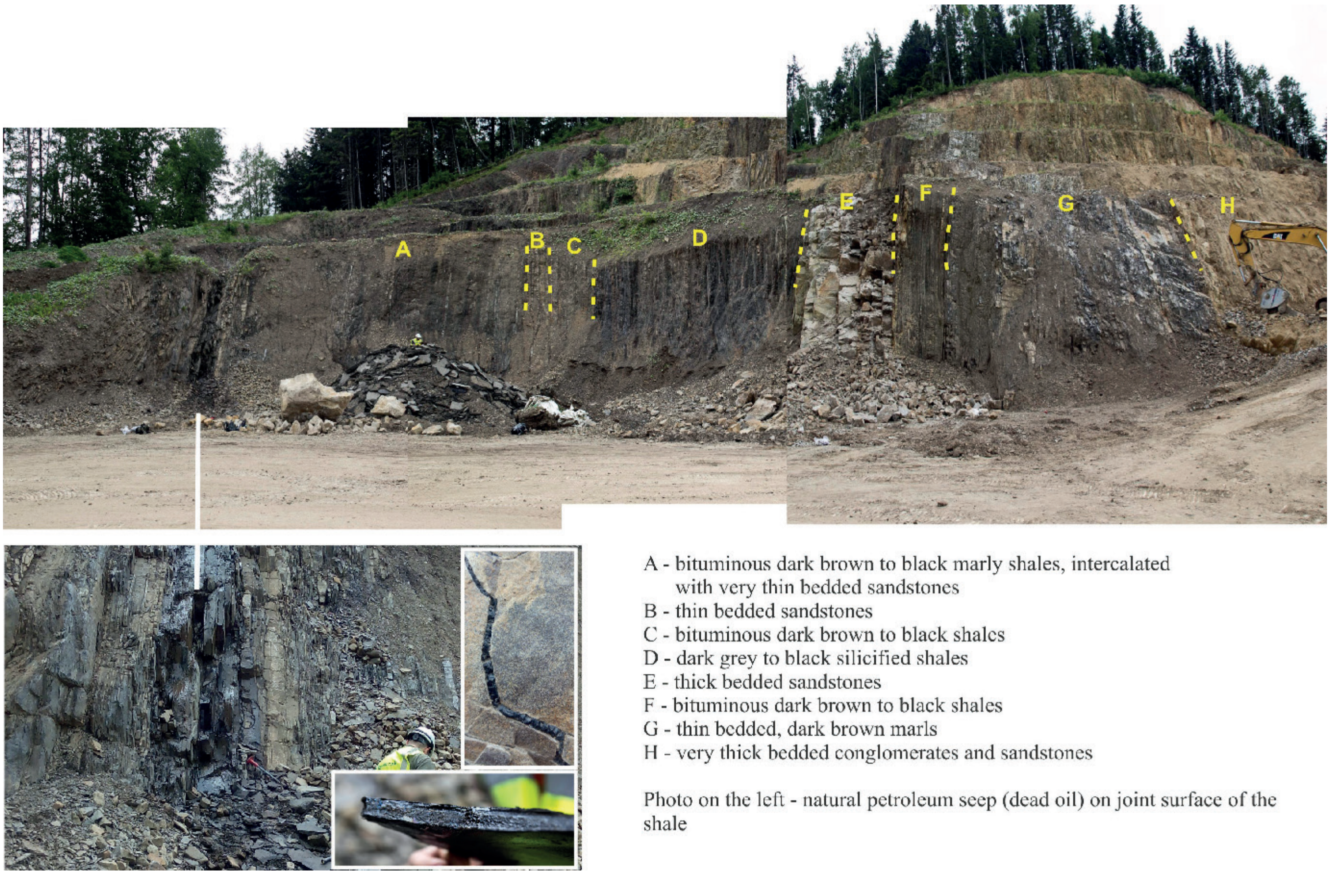


Fig. 17. Model of geological structure of Skrzydlna region



A - bituminous dark brown to black marly shales, intercalated with very thin bedded sandstones
B - thin bedded sandstones
C - bituminous dark brown to black shales
D - dark grey to black silicified shales
E - thick bedded sandstones
F - bituminous dark brown to black shales
G - thin bedded, dark brown marls
H - very thick bedded conglomerates and sandstones

Photo on the left - natural petroleum seep (dead oil) on joint surface of the shale

Fig. 18. General view of the NE part of the Skrzydlna quarry with lower part of the Menilite succession dominated by dark shales

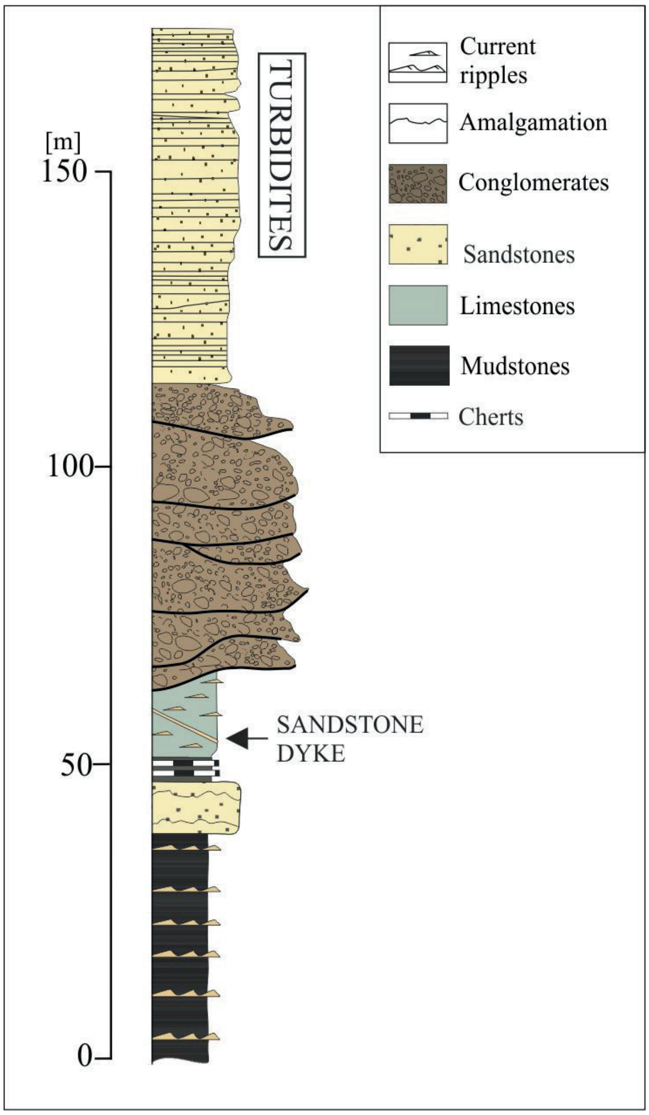


Fig. 19. Schematic section of the lithological succession at Skrzydlna quarry

Considering the succession age, the calcareous nannoplankton assemblages found in marl beds in the lower part of the exposed succession typified by shales analysed by K. Žecova, represent the NP 21 zone (Late Eocene). The youngest species, indicating the NP 22 zone (early Oligocene), were found in the limestone layer within the turbidites in the highest part of the open pit (Siemińska *et al.*, 2017).

The complex of menilite shales and siliceous shales located in the stratigraphically oldest part of the exposed succession (Fig. 6) contains rare, very thin interbeds of turbidite sandstones with current ripplemarks (Bouma Tcde; Tce) and cherts, and is intersected by several thin sandstone dykes. Above, there is a layer of fine- and medium-grained sandstone, 9 m thick, composed of massive, amalgamated intervals, devoid of mudstone interbeds, with a sharp erosive lower boundary. Following the next interval of the menilite shales with cherts, there is a 10-metre-long complex of thin layers of pelitic limestones and marls, locally silicified and containing isolated cross-laminated lenses of current ripplemarks of medium-grained quartz arenite.

The menilite shales complex represents a very low-energy deposition in a deep, calm, anoxic, tectonically inactive basin, interrupted by an episode of sandy material deposition from dense, rapidly decelerating gravity currents. On the other hand, sandstone dikes in menilite shales are interpreted here as the effects of liquefaction of unlithified sand layers caused by seismic events. We interpret the combination of these two phenomena as the earliest signals of uplift initiated in the source area (Fig. 21), accompanied by seismic shocks (Siemińska *et al.*, 2018; Jankowski & Wysocka, 2019). A massive sandstone interval is interpreted as the result of deposition out of high-density turbidite within a distribution channel, eroded and extending into the open menilite basin. The nature of this interval may suggest a series of long-lasting hyperpycnal flows. Features of the overlying limestone complex suggest hemipelagic deposition periodically interrupted by traction currents reworking a limited amount of sandy material, as indicated by isolated starved current ripplemarks. This unit also injected with sandstone dikes that represent three generations. We suggest that two intersecting thin veins and one 30 cm thick injection together with a massive layer of 9 m thick sandstone are the earliest signs of the approaching uplift responsible for the origin of the overlying coarse clastic complex.

The overlying olistostrome sequence consists of amalgamated layers deposited by debris flows, in which a rich sandy matrix with extra-basinal pebbles supports single blocks (olistoliths) over 0.5 meters long. The olistostrome basal surface is uneven, erosionally incised within the quarry by min. 10 m into the underlying carbonate complex (Fig. 22), and contains casts of drag marks of large objects extending NW-SE (Fig. 23). The intrabasinal rock fragments are plastically deformed elements of broken sandstone layers and clasts of black menilite shales. In 2017, among the extrabasinal olistoliths, in the middle of the complex's thickness, there was a block of Jurassic limestone measuring 10 m × 2.5 m. Two layers of black shales mark breaks in the sedimentation of the high-energy facies. The conglomerate layers are accompanied by irregularly appearing layers of massive ('structureless') sandstones geometry of which is determined by uneven contact with the conglomerate. Most often, the overlying conglomerate rests on the sandstone erosionally, while the bottom of the overlying sandstone layer adapts to the irregular surface of the underlying conglomerate (Fig. 24). Correlation between sections outcropping in the quarry benches suggests a kind of rough, indistinct organisation of facies: the coarser-grained conglomerate bodies filling lower parts of erosional channels, followed by and passing laterally to finer conglomerates and some pebbly sandstones. The facies distinguished in this complex are illustrated and summarized in Figure 25. In the upper part of the olistostrome sequence, the texture of the sandy matrix of conglomerates irregularly becomes finer and the structure better ordered.

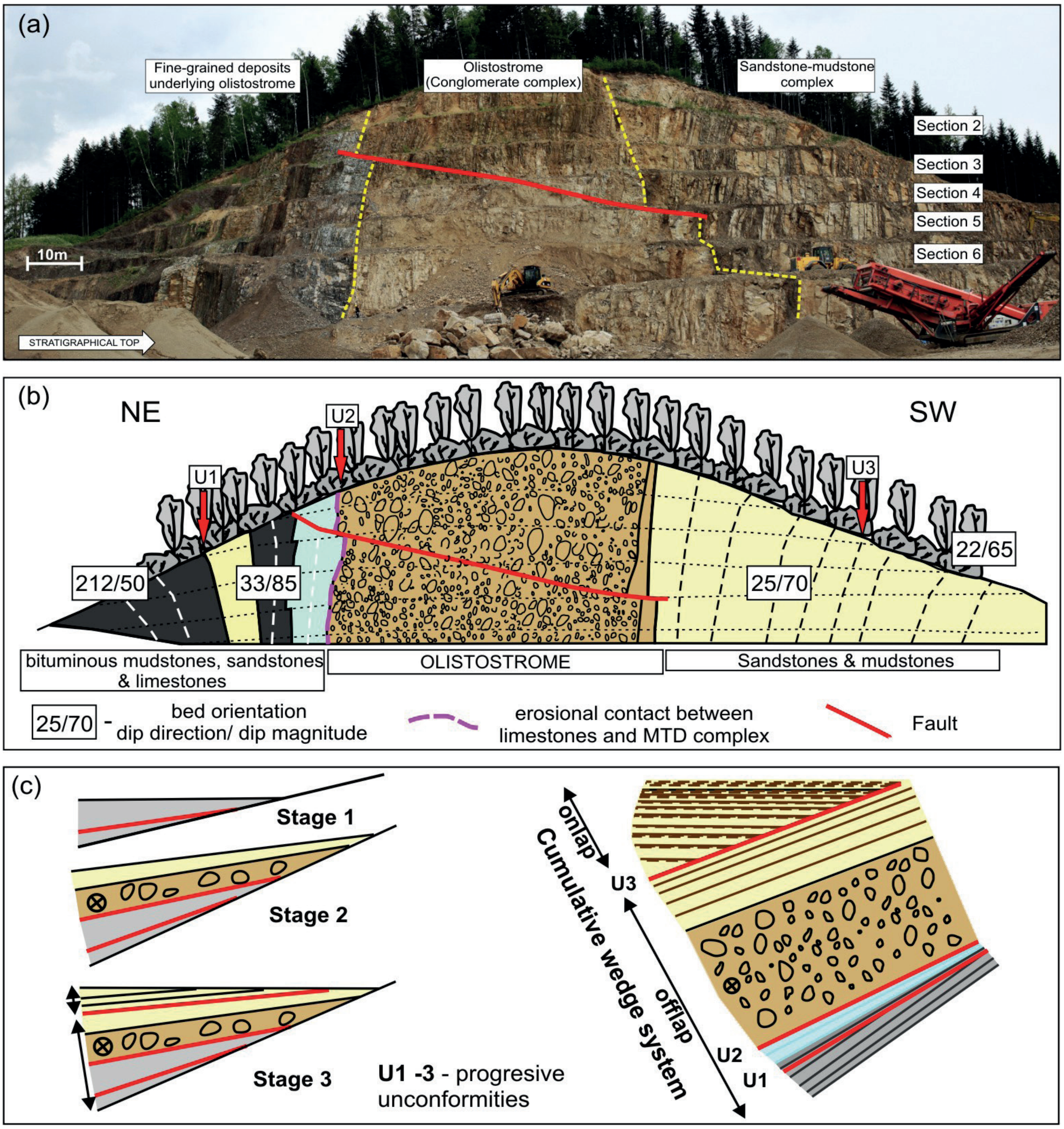


Fig. 20. General view (a) of the main lithological complexes exposed in the Skrzydlna quarry face and the sketch of the main lithological intervals in the quarry face (b). Dotted lines denote mining benches 2–6 labelled as Sections 2–6 in photograph (a); U1–U3 indicate unconformable contacts. Note gradual changes of bed attitude. Progressive unconformities related to synsedimentary rotation of the succession; geometry and stratigraphic relations based upon thirty bed attitude readings (modified from Siemińska *et al*., 2020) (c)

The transition from the olistostrome succession to the overlying turbidite complex is marked by several hybrid strata, each consisting of turbidite sandstone followed by linked debrite (i.e muddy sandstone rich in mudstone clasts). Some of the other layers have the characteristics of continuous/long-term gravity flows: hyperpycnally-fed turbidites. The overlying sequence, deposited mainly by short-lived classic surge-type turbidites, consists of three fining-upwards cycles (F-U). Two of them begin with very thick, amalgamated layers of massive sandstone deposited by high-density turbidity currents that fill channels cut 2.5 m into the underlying strata. Bouma sequences appear above, representing 'normal' density to dilute turbidites, with Ta-e rhythms fining and thinning to Tce in the uppermost thin-bedded sandstone and shale assemblages. Turbidite sandstones are quartz arenites with frequent admixture of carbonized plant remains. Overall, this 60 m thick turbidite succession forms an upward thinning megasequence composed of smaller upward fining/thinning cycles.

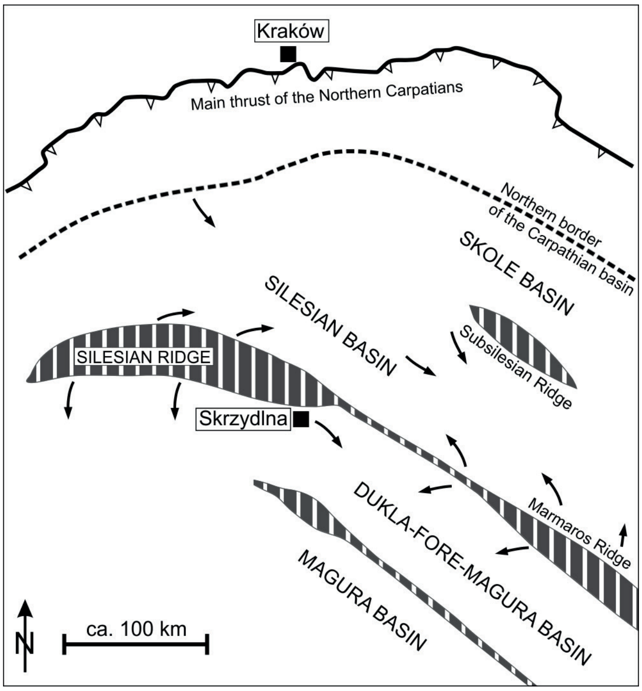


Fig. 21. Palinspastic sketch map showing arrangement of sedimentary basins, source areas (ridges) and main palaeotransport directions during sedimentation of the Menilite Formation (modified from Winkler & Ślączka, 1992)

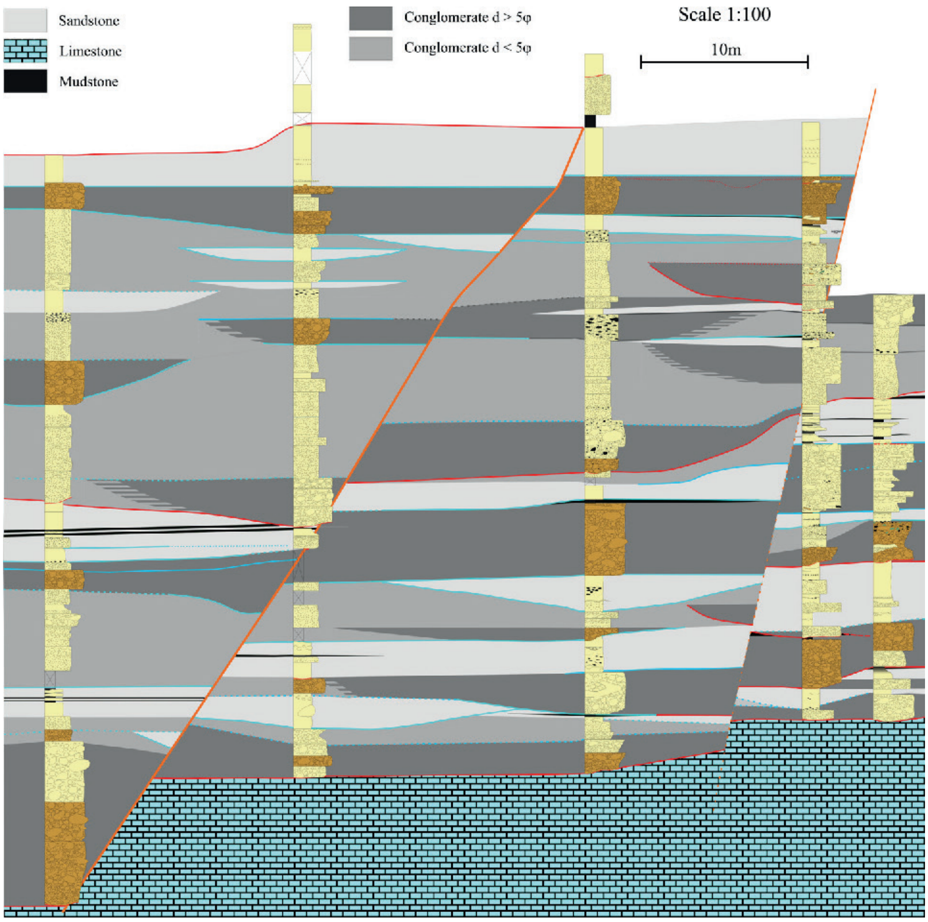


Fig. 22. Correlation of detailed sedimentological logs of Skrzydlna quarry outcrop within five exploitation levels as documented in years 2017–2018. Two main stages of olistostrome deposition are indicated by broad erosional surfaces marked red (modified from Siemińska *et al*., 2018, 2020 and interpretation by Wendorff *et al*., 2015)

Extreme facies contrasts between the fine-grained, anoxic basin sediments underlying the olistostrome and the suddenly appearing olistostrome complex deposited by gravity mass flows delivering very coarse-grained unsorted materials and oxidized water to the basin, suggest rapid uplift and very intense erosion of the source area (Wendorff *et al*., 2015; Siemińska *et al*., 2018). The abundance of dark shale intraclasts indicates the involvement of the slope and/or basin bottom paved with dark menilite-type sediments, the proximal part of which has been transformed into a slope bordering an uplifted block – the source of detrital material. Vertical changes in the conglomerate lithology of the olistostrome complex are a reflection of three main phases of deposition (Fig. 10). The beginning of each phase is marked by an erosive surface on the scale of the entire exposure (Fig. 6). The broad lithological diversity, age distribution and shapes of extraclasts in the olistostrome imply a complex structure of the source area, which included Mesozoic rocks. This suggests that the topography of the source zone was very diverse and changed over time. Very well rounded boulders, some equant and prolate, suggest long-term rolling of large blocks on the abrasion platform. The huge angular block of Jurassic limestone reflects collapse of a cliff face and the subsequent block sliding. Slumped fragments of sandstone layers with much better sorted than the conglomerate matrix imply landslides and slides of slabs – gravity-induced dismemberment of sandstone layers initially deposited in the proximal/upper slope zone. All these features suggest that the olistostrome succession originated as a result of mass transport of debris flows (MTD) locally transformed into high density turbidity currents, landslides, sliding of blocks and slabs of dismembered beds. This stage was succeeded by various turbulent sediment flows, from high to low density turbidites in the later phase of deposition (Siemińska *et al*., 2020). The occurrences of dunes at the turbidite deposition stage indicates periods of long-lasting flows redepositing sands by traction, most probably at the mouths of the distributary channels.

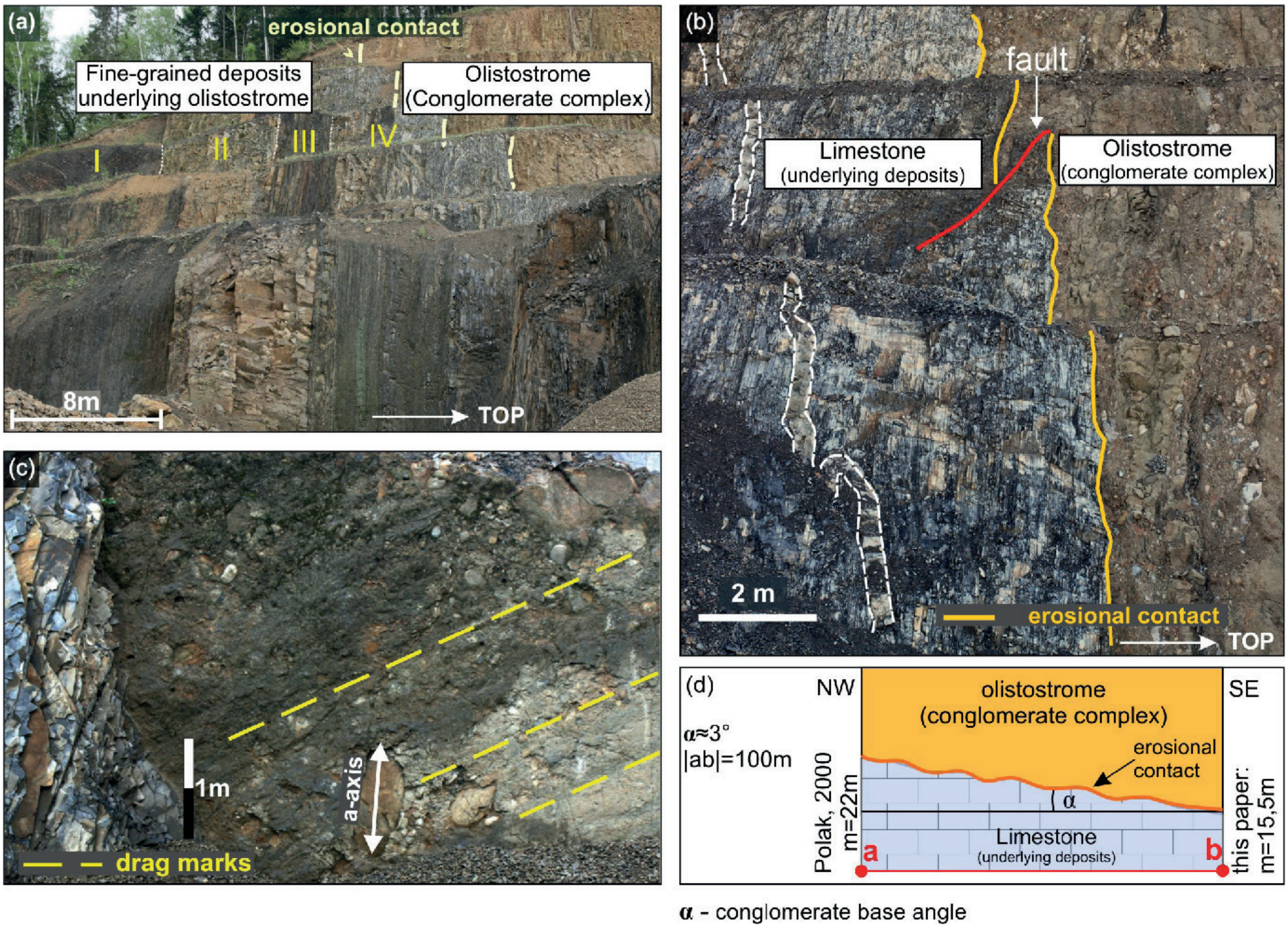


Fig. 23. General view (a) of deposits underlying olistostrome: I – calcareous and siliceous shales, II – amalgamated sandstone, III – siliceous shales, IV – limestone. Close-up view of erosional contact between limestone complex, containing sandstone dyke shown with dashed outline, and overlying olistostrome (b). Grooves at the basal erosional surface of olistostrome complex (c). Orientation of long axis of the elongated clast suggests rolling during the last stage of transport. Difference in limestone thickness measured by Polak (2000) and the present authors (in 2016) in section sub-parallel to the palaeocurrents reflects inclination of the erosional base of the olistostrome relative to the underlying limestone bedding (d).

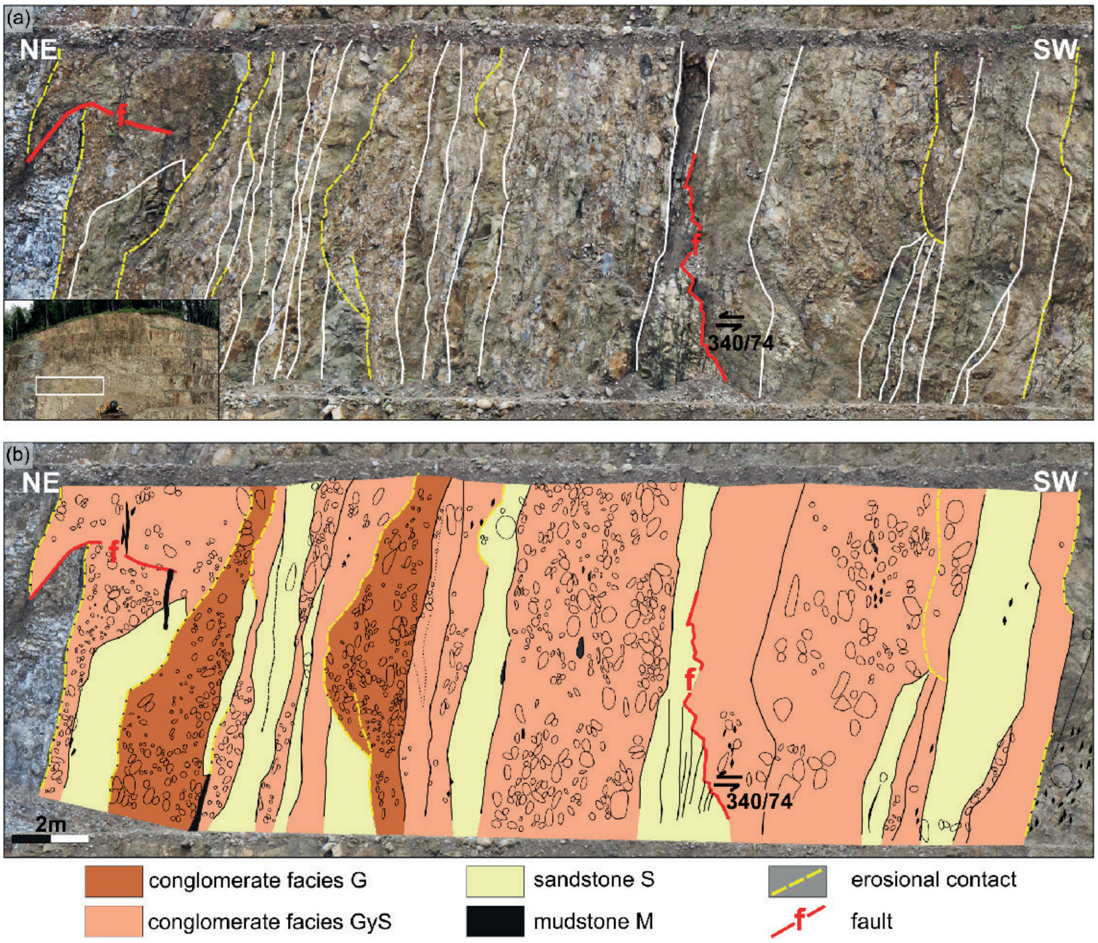


Fig. 24. The main characteristics of the conglomerate complex are very high variations of bed thickness and sedimentary features, even within beds. These are shown in the photograph (a) and emphasised in drawing (b) by subdivision into facies. Note that lateral and vertical changes occur even at a distance of some tens of centimetres. Inset in (a) indicates position of the photograph in the quarry face

From the point of view of basin infilling processes, the sudden appearance of the olistostrome in a low-energy, oxygen-poor menilite basin, the features of the olistostrome basal surface, and the presence of the menilite shales intraclasts suggest that this thick clastic complex deposited by debris flows fills a channel eroded in the slope of the uplifted source zone (Fig. 26). The trend of changes in the bedding dip direction observed in 2017–2019, in the absence of clear traces of slip and shear on the bedding surfaces, suggests syntectonic deposition on the substratum of a rotated block (Fig. 20), similar to the stratigraphic relationship and the formation of progressive unconformities documented for the first time in the Pyrenees (Riba, 1976). Lateral and vertical facies changes in the olistostrome complex reflect significant hydrodynamic variations within and between successive flows. The first-order F-U sequence composed of second-order F-U sequences within the overlying turbidite succession implies the formation of a retrograding submarine fan (Fig. 26; Wendorff *et al.*, 2015). Age-wise, the olistostrome and fan succession were deposited during the climate-controlled (cooling) global sea level fall in the Eocene and Oligocene. Therefore, its genesis can be associated with a magnification of influence of the tectonic uplift of the source zone by the simultaneous marine regression, i.e. amplification of uplift of the alimentation area. In addition, the observed nannoplankton assemblages may reflect eustatic sea level fluctuations and temporarily low environmental salinity, which could be related to tectonic movements, climate change and the gradual isolation of the Paratethys in the late Eocene and Oligocene (Švábenická *et al.*, 2007; Siemińska *et al.*, 2017; Pszonka *et al.*, 2023).

Authors' editorial note:

The material presented in this description of the Skrzydlna quarry succession is partly based upon the following publications: Siemińska et al. (2017, 2018, 2020); Wendorff et al. (2015), and also includes unpublished results of the first author. This is edited and modified English version of the Guide to Skrzydlna previously prepared in Polish for the 88th Meeting of the Polish Geological Society in May 2023.

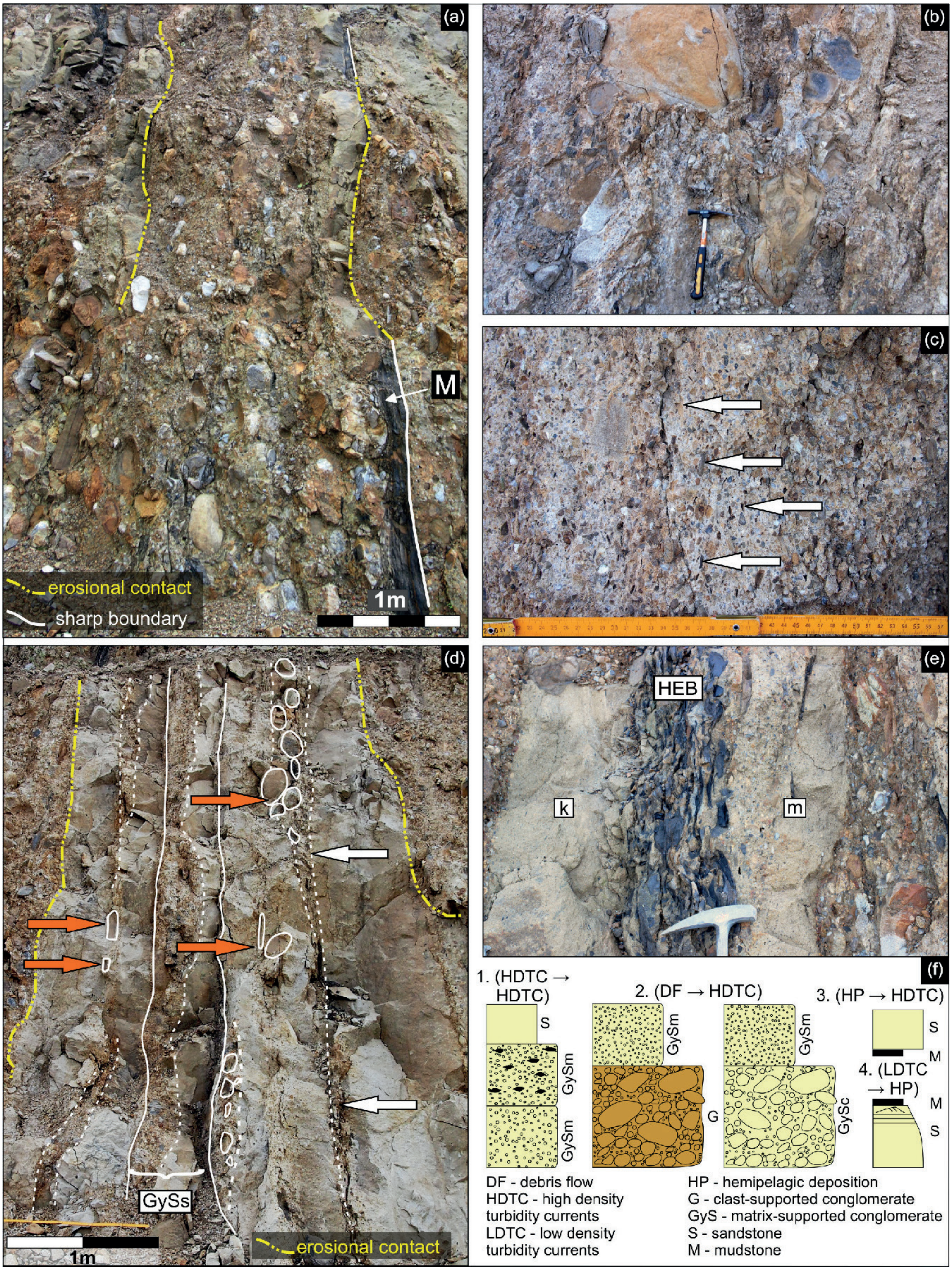


Fig. 25. Examples of olistostrome lithofacies exposed in Skrzydlna quarry. Conglomerates: very coarse-grained with blocks/olistoliths (a), with clasts of black menilite mudstone (b), and medium-grained with clasts oriented parallel (c). Sandstones (d) with conglomerate pockets, and hybrid beds with clasts of black mudstone (e). Subfacies of conglomerates and their predominant transitions (f)

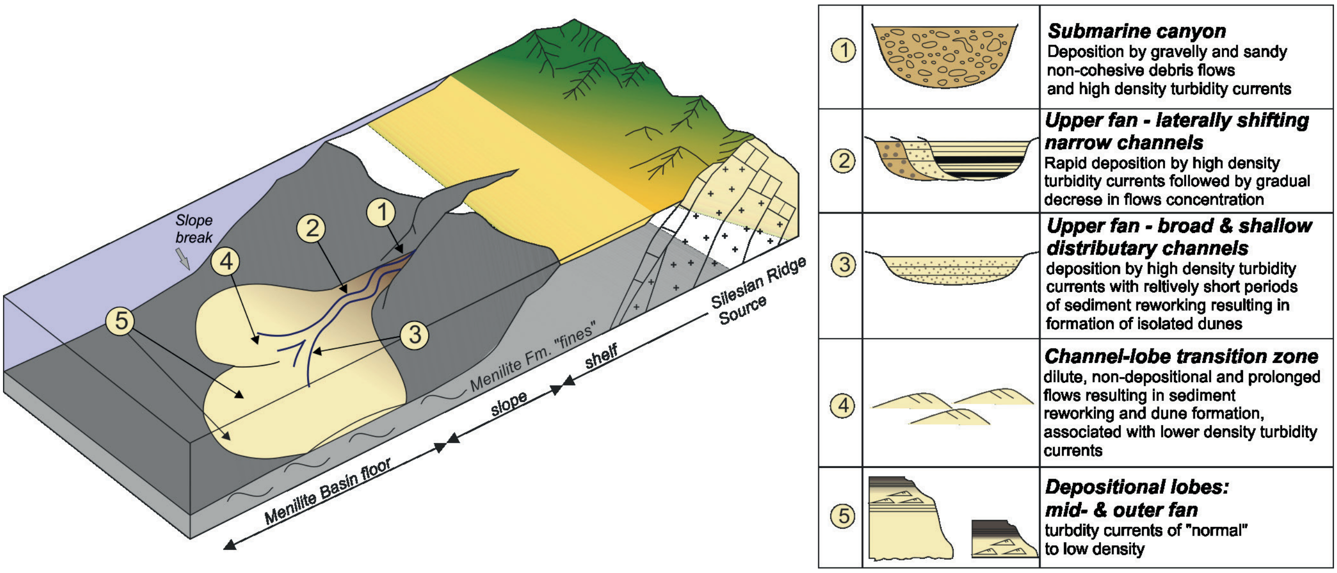


Fig. 26. Model of evolution of the Skrzydlna olistostrome to turbidite megasequence, interpreted in terms of Walther's Law of Facies as grading from proximal submarine canyon fill to distal small turbidite fan. Note five associations of facies, their palaeotopographic expression, and selected sedimentary features, deposition zones and the main controlling processes (1–5). Interpretation by Siemińska *et al.* (2018, 2020) and Wendorff *et al.* (2015)

Stop 2 –
Skalski stream near Jaworki village (Late Cretaceous deposits with exotics) (Figs 27, 28)

(*Michał Krobicki, Barbara Olszewska*)

The Upper Cretaceous and Paleogene exotic-bearing gravelstones of the PKB are linked with flysch/flyschoidal sequences of the Sromowce Formation and the Jarmuta Formation, respectively (Birkenmajer, 1977, 1979, 1986). These exotic rocks are useful for reconstructing the basement and sedimentary cover of source areas (Birkenmajer, 1988). The present outcrop is located in the Skalski stream (below the lower station of ski lift – Birkenmajer, 1977, 1979, 1988; Birkenmajer & Lefeld, 1969; Radwański, 1978; Birkenmajer & Wieser, 1990; Birkenmajer *et al.*, 1990). Lower Cretaceous Urgonian-type exotic rocks often occur within such gravelstones (Bukowiny Gravelstone Member – Birkenmajer & Lefeld, 1969) which represents a submarine slump, in the middle of the Sromowce Formation of the Niedzica Succession (Birkenmajer, 1977, 1979; Birkenmajer & Jednorowska, 1987b) (Fig. 28).

The Urgonian (named after the village Orgon, east of Tarascon, France) is a characteristic shallow-water carbonate facies that accumulated along the Tethys northern shelf from the Barremian to the Late Albian. The facies encloses hard, light-coloured limestones with foraminifers and pachyodonts, marls with *Orbitolina* (foraminifers) and transitional sediments – detrital or siliceous limestones (Foury, 1968). Characteristic fossils of the facies are bivalves (rudists), corals, hydrozoans, bryozoans, small and large foraminifera and algae. The origin of the Urgonian facies is connected with the Barremian rearrangement of the world ocean (Renard, 1986). The Barremian regression uncovered large parts of the shelves on which abundant shallow-water communities proliferated until the mid-Aptian transgression caused, locally, their emersion and destruction (Scott, 1995). In the Inner Carpathians, the Urgonian facies is represented in the Hightatric units of the Tatra Mts (Lefeld, 1974, 1988; Masse & Uchman, 1997) and the Manín Unit of the Váh valley (Andrusov, 1953; Mišík, 1990). In the PKB, the Urgonian-like facies occurs in the Haligovce Nappe (Birkenmajer, 1959; Haligovce Limestone Formation – Birkenmajer, 1977) and as exotic pebbles in the Upper Cretaceous Sromowce Formation and the Upper Cretaceous–Paleocene Jarmuta Formation (Birkenmajer & Lefeld, 1969; Birkenmajer, 1970, 1977, 1979, 1986; Birkenmajer & Wieser, 1990). In the Outer Carpathians, the Urgonian-type limestones occur exclusively as exotic pebbles in younger deposits (Birkenmajer, 1970, 1973, 1977; Birkenmajer & Lefeld, 1969; Oszczypko, 1975; Burtan *et al.*, 1984). Micropalaeontological investigations of these limestones were so far limited to determination of orbitolinids, without full documentation of other foraminifers typical for the Urgonian facies (except for remarks by Mišík, 1990; Krobicki & Olszewska, 2004). Foraminiferal assemblages of the Early Cretaceous Urgonian-type limestones contain many stratigraphically significant species of

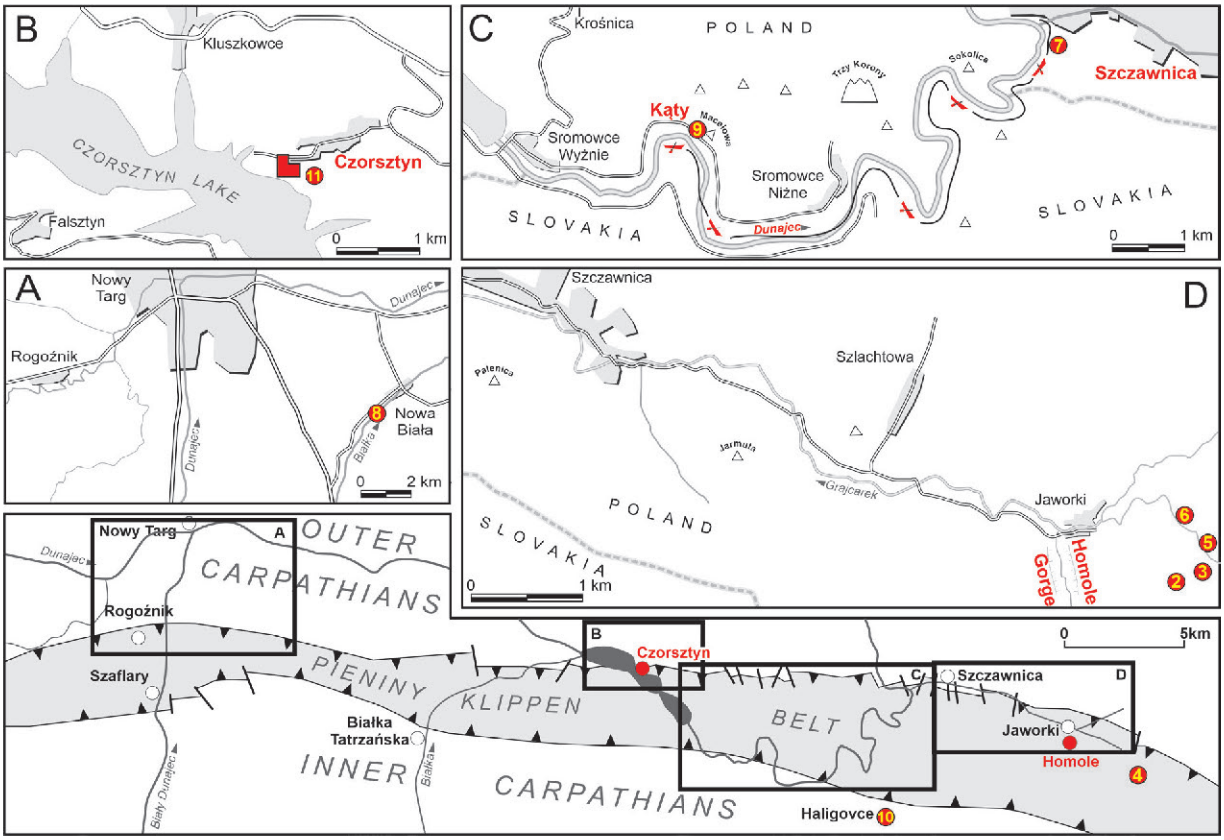


Fig. 27. Polish part of the Pieniny Klippen Belt and locations of visited outcrops – stop points

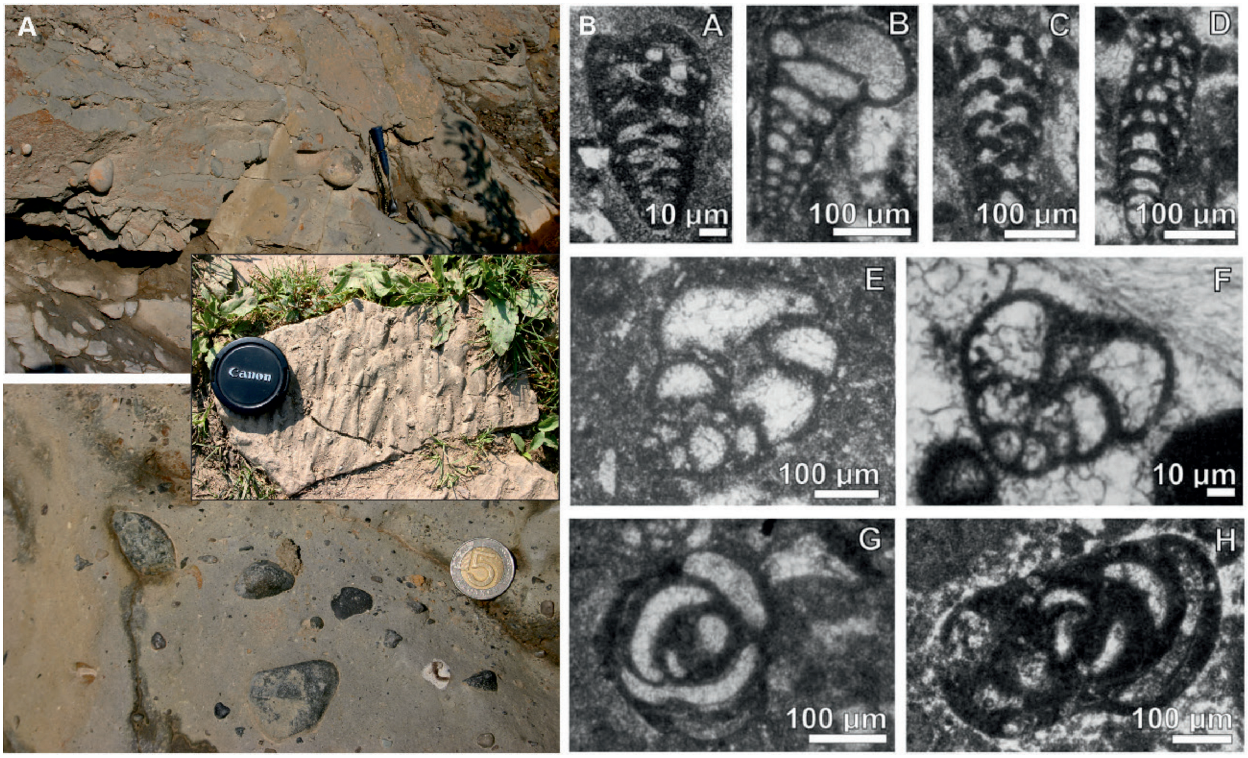


Fig. 28. View of the Bukowiny Gravelstone Member and Sromowce Formation in Skalski stream (A) and foraminifers of the Urgonian limestones (B): A – *Praechrysalidina infracretacea* Luperto Sinni (vertical section); B – *Praedorothia praeoxycona* (Moullade) (longitudinal section); C, D – ?*Pseudotextulariella scarsellai* (de Castro) (longitudinal sections); E – *Pfenderina* cf. *janae* Neagu (longitudinal section); F – *Pfenderina aureliae* Neagu (vertical section); G – *Pseudonummuloculina aurigerica* Calvez (oblique section); H – *Pseudonummuloculina aurigerica* Calvez (sub-axial section) (after Krobicki & Olszewska, 2022, modified)

both larger and smaller foraminifera. They cover the time span of Hauterivian–Albian, predominantly the Barremian to Early Aptian. Sporadic planktonic organisms, such as planktonic foraminifera (*Hedbergella*, *Heterohelix*, *Favusella*), tintinnids (*Colomiella* cf. *recta* Bonet) and calcareous dinocysts ("Calcispheres") [*Crustocadosina semiradiata* (Wanner)] have also been observed. Tintinnids (*Colomiella* cf. *recta* Bonet) and planktonic foraminifera [*Heterohelix* aff. *moremani* (Cushman)] suggest that the age of some assemblages may be younger than the earliest or even Late Albian (Reháková & Michalík, 1997; Nederbragt *et al*., 1998).

Additionally, magmatic/sub-volcanic exotic pebbles occur here, which are very well rounded. The exotics are represented by granitic and andesitic-type rocks (mainly andesite, basaltic andesite, basaltic trachyandesite, trachyandesite and rhyolitic pebbles, and rare dacite, tephrite, trachybasaltic and basaltic pebbles). Radiometric dating of these exotics (by U-Pb SHRIMP 206Pb/238U method) are following: 266.0 Ma ±1.6 Ma, 266.4 Ma ±1.8 Ma, 268.8 Ma ±1.9 Ma and 269.7 Ma ±1.8 Ma (Middle Permian; Guadalupian) (Poprawa *et al*., 2013; Krobicki *et al*., 2018). Additionally, such features suggest origin of these exotics from the same source area (Inner Carpathians the most probably) and have been connected with the Middle Permian oceanic crust subduction and origin of magmatic arc mentioned above, presumable connected with southern margin of Laurusia with subduction of oceanic crust of the Palaeotethys (proto-Vardar Ocean?). In conclusion, results of SHRIMP data and geochemical character of investigated exotics excluded their Barremian-Albian age of subduction and the Early Cretaceous age of oceanic crust (Birkenmajer, 1986, 1988), and existence of so-called Exotic Andrusov Ridge as well, postulated earlier in several papers (e.g., Birkenmajer, 1977, 1986, 1988).

Stop 3 – Bereśnik hill near Jaworki village ("mid"-Cretaceous pelagic shell beds, Late Cretaceous deposits and inversion structure) (Fig. 29)

(*Michał Krobicki*)

Pelagic deposits of the Albian/Cenomanian transition occur mainly in the Kapuśnica and Jaworki formations. The Kapuśnica Formation (Upper Aptian-Albian) is represented by dark-grey shales, grey-blue and green marly shales with intercalations of light-grey pelitic spotty (also cherty) limestones, rare fine-grained turbidite sandstone intercalations, and a few layers of black radiolarian shale. Other fossils, i.e. belemnites, ammonites and pelecypods, are very rare (see Kokoszyńska & Birkenmajer, 1956). The shell beds described here occur in the Niedzica Succession and represent the late Upper Albian through early Lower Cenomanian (foraminiferal palaeobathymetric association B1 of Gasiński (Gasiński 1991; Birkenmajer & Gasiński, 1992). The Albian-Cenomanian shell beds discussed are exposed in a right tributary of the Skalski Stream, about 1.5 km southeast of the Jaworki village (Krobicki, 1995) (Fig. 29). A section 7.5 m thick is seen in the steep bank of the stream. The section exposes mostly green and green-grey hard marls, spotty marly limestones and shales (cherry red-grey at some places), with subordinate intercalations of green spotty pelitic limestones and shell beds from a few to 30 cm thick. Some beds are technically boudinaged. The shell beds occur as layers and lenses varying in thickness from about 0.5 cm to 20 cm. They are built almost exclusively of small thin-shelled bivalves of the genus *Aucellina*. The shells are dismembered and severly crushed, many are also deformed by compaction, as is shown by broken shells whose fragments remain in place. Unbroken shells are preserved in the lower parts of some shell beds. Abundant and very well preserved mainly planktonie foraminiferal tests are present together with the bivalve shells, other fossils are represented by small pieces of indeterminable bivalve shells with spines and fragments of echinodenns. The shell beds are nearly monospecific (paucispecific sensu Kidwell *et al*., 1986). Belemnites and single shells of the bivalve *Aucellina* sp. Are very rare in the accompanying marls and marly limestones which show the presence of abundant trace fossils. They are represented mainly by *Chondrites* and by *Zoophycos* and *Planolites*-like burrows in marls and pelitic limestones. Sedimentary features of the described deposits are indicative of deep water pelagic deposition. Soft-substratum conditions are suggested by body fossils (*Aucellina* shells) and trace fossils (deposit-feeders dominating), the latter indicative of low energy of the bottom water. The foraminiferal assemblages are characteristic of middle slope (Gasiński, 1991; Birkenmajer & Gasiński, 1992). The dominant sedimentary features of the Albian-Cenomanian pelagic marls and limestones, their foraminifers and the macrofauna, all indicate low energy of their depositional environment In contrast, the skeletal accumulations represent highenergy events. Wide occurrence of shell beds in different carbonate or mixed clastic/carbonate shallow-marine sediments, both ancient (e.g. Kreisa & Bambach, 1982; Aigner, 1985; Fürsich & Oschmann, 1986; Eyles & Lagoe, 1989; Johnson, 1989) and modem (e.g. Gagan *et al*., 1988; Davies *et al*., 1989) is well known. Most of the shell beds formed above the storm wavebase, and skeletal accumulations formed below it are rare (Kidwell *et al*., 1986: fig. 5). Pelagic bivalve turbidites occurring widely in the Alpine-Mediterranean region are an exception (e.g. Bernoulli & Jenkyns, 1974).

In the upper part of the Bereśnik hill we can see the Late Cretaceous *Scaglia Rossa* deposits again overlying by flysch-type of the Sromowce Formation with exotics in local inversion structure.

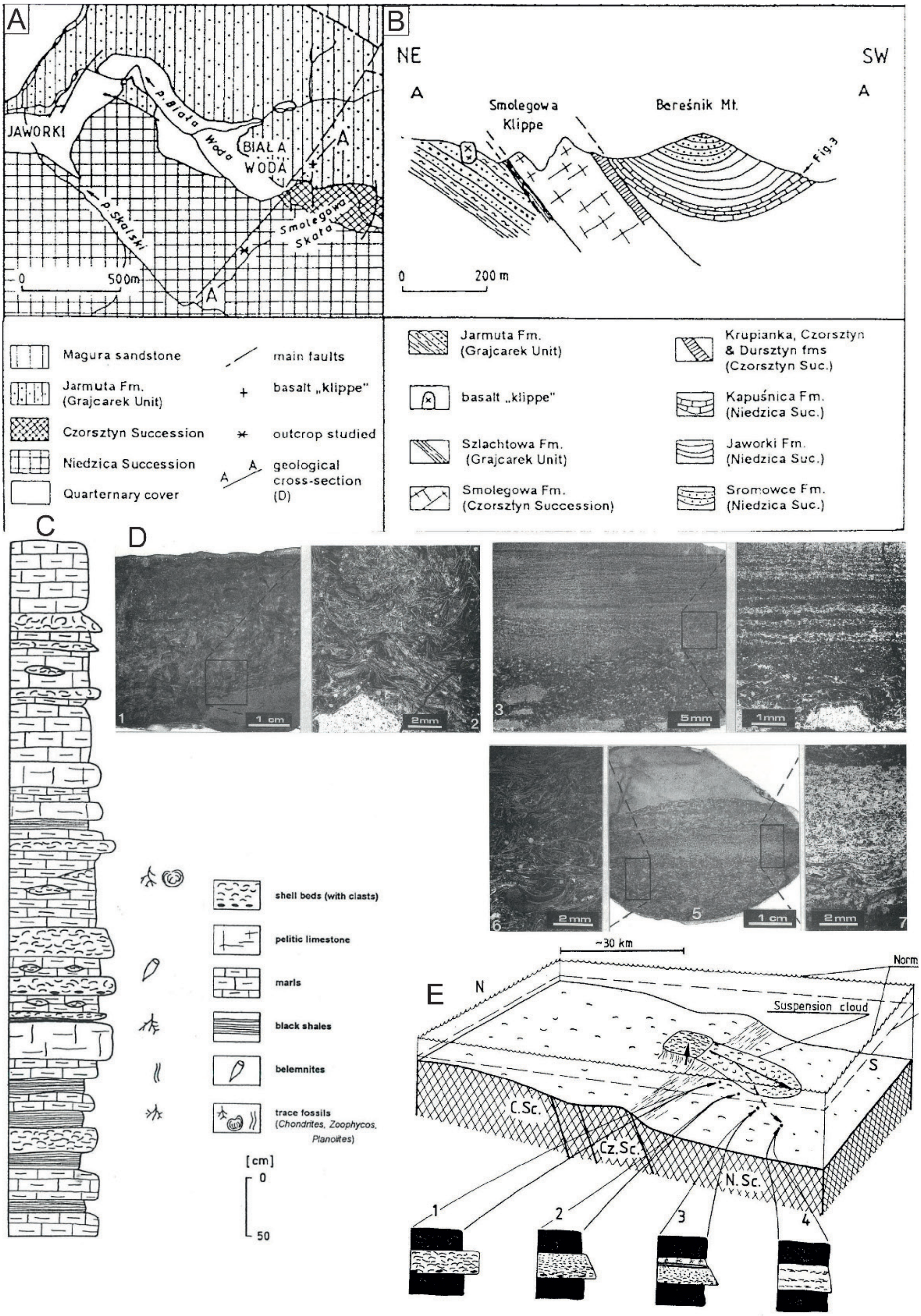


Fig. 29. Position of the outcrop with mid-Cretaceous shell beds: A – geological sketch-map of the Pieniny Klippen Belt in the vicinity of Jaworki (after Birkenmajer, 1979 simplified); B – geological cross-section at Biała Woda valley (after Krobicki, 1995); C – *Aucellina* shell beds-bearing exposure of the Kapuśnica Formation attributary of the Skalski Stream; D – different types of *Aucellina* shell beds in pelagic Albian sediments, Kapuśnica Formation of the Niedzica Succession; E – a generalised model for the origin of the four types (1–4) of storm-generated shell beds. C.Sc. – Czorsztyn Succession; Cz.Sc. – Czertezik Succession; N.Sc. – Niedzica Succession. Type of shell beds: 1 – densely packed, homogenous shell beds, in some cases with clasts, sharply delimited from marl or pelitic limestone beds; 2 – graded shell beds with clasts at base; 3 – sequences starting from small clasts and shell lag at base, through graded, laminated to bioturbated at top; 4 – thin shell accumulations in silty marls and limestones (after Krobicki, 1995, 2022a; Krobicki & Olszewska, 2022)

Stop 4 –
Biała Woda valley (near Brysztan Klippe) and Berriasian phosphatic structures (Figs 30, 31)

(*Michał Krobicki*)

The present stop includes data from Niedzica Succession situated in the eastern part of the Polish sector of the PKB. The Dursztyn Limestone Formation, subdivided into the Korowa Lime stone Member and the Sobótka Limestone Member (Birkenmajer, 1977), represents the Tithonian/Berriasian boundary strata of the Niedzica Succession. East of Jaworki (Fig. 27) the Niedzica Succession occurs as a large sheet (nappe) thrust over the Czorsztyn Succession. The out crops studied expose white, massive, micritic lime stone which grades upwards into thin-bedded, micritic limestone rich in bioclasts (crinoids, brachiopods, ammonites, small gastropods) of the Sobótka Limestone Member. At the top part of this member, a 10–20 cm thick layer composed of green micritic limestone, rich in phosphorite occurs.

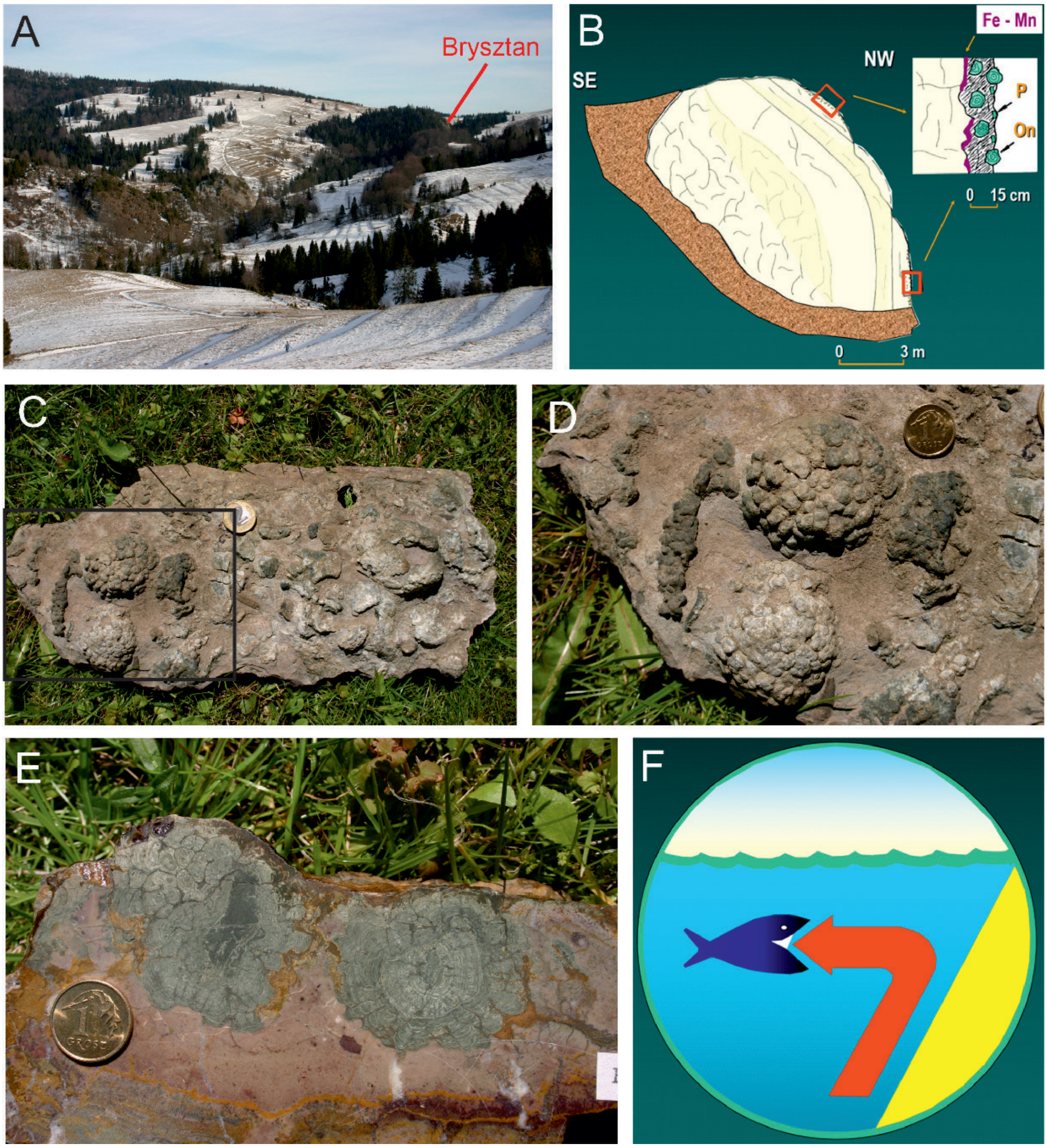


Fig. 30. Location of the Biała Woda-Brysztan section (A); outcrop of the Dursztyn Limestone Formation, Sobótka Limestone Member (Berriasian), Niedzica Succession (B): in enlargement – top of last bed (after Krobicki, 1993, 1994): Fe-Mn – Fe-Mn crusts, P – phosphorites, On – phosphatic macrooncoids. On the photos C–E: phosphatic macrooncoids, and F – upwelling logo (after Krobicki, 2022b; modified)

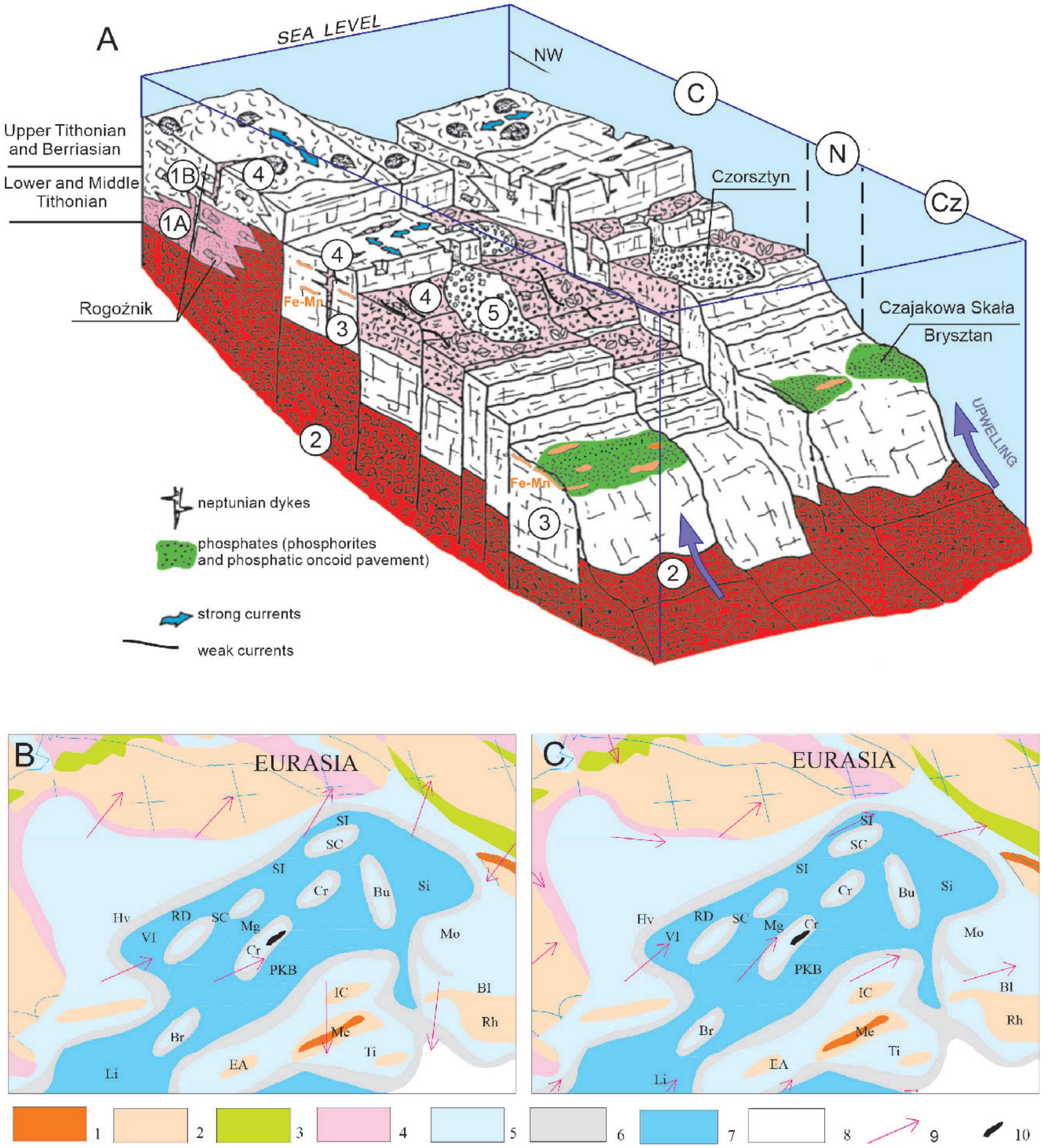


Fig. 31. Model of sedimentation on the intraoceanic Czorsztyn pelagic swell in Berriasian (A) with effects of pronounced Neo-Cimmerian tectonic movements and upwelling currents (after Krobicki, 1996, modified): 1 – Rogoźnik Coquina Member (Dursztyn Limestone Formation: A – sparitic coquina, B – micritic coquina); 2 – Czorsztyn Limestone Formation (*Ammonitico Rosso* facies); 3 – Sobótka Limestone Member (Dursztyn Limestone Formation); 4, 5 – Łysa Limestone Formation (4 – Harbatowa Limestone Member, 5 – Walentowa Breccia Member); successions: C – Czorsztyn, Cz – Czertezik, N – Niedzica. Palaeoenvironments, wind direction and upwelling zones of the Carpathian area during Tithonian-Berriasian time (palaeogeography after Golonka *et al*., 2000, modified): summer Northern Hemisphere (C), andwinter Northern Hemisphere (B): 1 – mountains/highlands (active tectonically), 2 – topographic medium-low (inactive tectonically, non-deposit), 3 – terrestrial undifferentiated, 4 – coastal, transitional, marginal marine, 5 – shallow marine, shelf, 6 – slope, 7 – deep ocean basin with sediments (continental, transitional, or oceanic crust), 8 – deep ocean basin with little to no sediments (primarily oceanic crust), 9 – wind directions, 10 – upwelling zone; abbreviations of oceans and plates names: Bl – Balcans, Br – Briançonnais terrane, Bu – Bucovinian terrane, Cr – Czorsztyn Ridge, EA – Eastern Alps, Hv – Helvetic zone, IC – Inner Carpathians, Li – Ligurian (Piemont) Ocean, Me – Meliata suture, Mg – Magura Basin, Mo – Moesia Plate, PKB – Pieniny Klippen Belt Basin, RD – Rheno-Danubian Basin, Rh – Rhodopes, SC – Silesian Ridge (cordillera), Si – Siret, Sl – Silesian Basin, Ti – Tisa Plate, Vl – Valais Trough

At this level large (8–10 cm across), phosphatic macrooncoids form an oncolitic pavement (see Krobicki, 1993, 1994, 1996a, 2022). On bedding surfaces, large ammonites (up to 30 cm in diameter) are visible. The rock is strongly fractured; Fe-Mn crusts cover the irregular surfaces of the sedimentary discontinuities (Fig. 30). The rock record from PKB shows that upwelling happened in the earliest Cretaceous time. The inception of upwelling may be associated with the time of plate tectonic reorganisation. Tectonic movements generated shallow platforms and islands along the NE–SW trending ridges between the main part of Tethys and the Eurasian Platform. Palaeoclimate modelling (Fig. 31) suggests that the Jurassic-Cretaceous prevailing wind directions in the Circum-Carpathian Tethys area were north-north-east, parallel to the ridges.

Upwelling may have been in duced at the south-eastern margin of the ridges. This type of oceanic circulation has been recorded in a specific association of deposits. These are, given the extremely high biological productivity associated with upwelling, mainly biogenic rocks with high contents of organic matter, silica, and phosphates in different forms, and deposits with elevated contents of some trace elements. The coincidence of these factors in a given palaeogeographical situation might help to re construct the palaeoceanographical conditions of a specific type of upwelling circulation. Upwelling areas are marine regions, in which masses of cold sea waters, rich in nutrients, are lifted from ocean depths to ward more shallow zones, situated most often along the continent margins. Such a nutrient-rich water circulation facilitated growth of zoo- and phytoplankton. Organic production at the lowest trophic level might have been very high, as it caused flourishing growth of benthic organisms along several hundreds of the kilometres-long, intraoceanic Czorsztyn pelagic swell. At the same time, at the Tithonian-Berriasian boundary, a microplankton (mainly calpionellids) explosion took place in the sedimentary basins of the Western Carpathians, triggered by palaeogeographic changes related to Neo-Cimmerian tectonics (Reháková & Michalík, 1994). The presence of phosphate-rich deposits (phosphorites and microbial phosphate structures macrooncoids) in the Berriasian deposits of the Niedzica Succession, which in a palinspastic reconstruction represents a shelf-edge/slope boundary, supports this idea (Figs 7, 31).

Stop 5 –
Biała Woda valley (waterfall) – Berriasian crinoidal limestones and synsedimentary breccia and Valanginian crinoidal limestones (Fig. 32)

(*Michał Krobicki, Andrzej Wierzbowski*)

The deposits of the uppermost Jurassic–lowermost Cretaceous of the Czorsztyn Succession are represented by: the Dursztyn Limestone Formation, the Łysa Limestone Formation (including the Harbatowa Limestone Member, the Walentowa Breccia Member and the Kosarzyska Limestone Member), and the Spisz Limestone Formation (Fig. 11). The age of the Spisz Limestone Formationwas poorly known some years ago. The study by Wierzbowski (1994), Krobicki (1996b) and Krobicki & Wierzbowski (1996) have shown that the lowermost part of the Spisz Limestone Formation reveals in many sections signs of stratigraphic condensation, containing ammonites characteristic of the Early Valanginian and, locally even of the early Late Valanginian age.

The sections studied in the Biała Woda Valley are located in its western (left) slope, at the waterfall (see Fig. 22). The sequence ofthe deposits corresponds strictly to that exposed at the waterfall, in the eastern (right) slope of the valley (see Birkenmajer, 1963, 1977). The oldest deposit in the section A is a grey and grey-brown breccia consisting of angular limestone clasts some millimeters in diameter. The ciasts are calcareous mudstones with abundant calpionellids, mostly *Calpionella alpine* Lorenz, and *Globochaete*. This breccia bed is exposed at water level of the stream. The discussed part of the section is now referred to the Walentowa Breccia Member of the Łysa Limestone Formation (denoted as WBM in Fig. 32A; comp. Birkenmajer, 1977). The overlying beds numbered 1–4 in section A, and 3–4 in section B (Fig. 32), 3.65 m thick, consist of brown, red-brown, and red-violet-brown crinoid-brachiopod limestones. They are packstones and grainstones with abundant crinoid ossicles (up to 5 mm in diameter), shells of brachiopods and, less commonly, tests of benthic foraminifers. The rare ammonites, brachiopods, and calpionellids, such as *Calpionellopsis* sp., *Tintinnopsella carpathica* (Murgeanu and Filipescu) and *Remaniella* sp., are indicative of Late Berriasian. These deposits were originally attributed to the lower part of the Spisz Limestone Formation by Birkenmajer (1963, 1977). In fact, they are almost identical in their lithological development and stratigraphic position to those of the Kosarzyska Limestone Member of the Łysa Limestone Formation (Krobicki & Wierzbowski, 1996; comp. Wierzbowski & Remane, 1992; Wierzbowski, 1994). This sections in the Biała Woda Valley yielded brachiopods of the lowermost Cretaceous. The brachiopods were collected bed by bed, starting from crinoid-brachiopod limestones of the Walentowa Breccia Member of the Łysa Limestone Formation and proceeding up to the middle part of the Spisz Limestone Formation (bed 5).

On the other hand, brachiopods are infrequent in the Spisz Limestone Formation (Krobicki, 1994, 1995, 1996b). The dominating brachiopod species in assemblages from crinoid limestones of the Spisz Limestone Formation are the same as those occuring in the Tithonian and/or Berriasian strata (Barczyk, 1972a, 1972b, 1979a, 1979b, 1991; Krobicki, 1994). The Valanginian brachiopod assemblages are useful for ecostratigraphy. Comparison of the Valanginian and Late Berriasian assemblages from the PKB shows that they are also useful in palaeoenvironmental reconstructions.

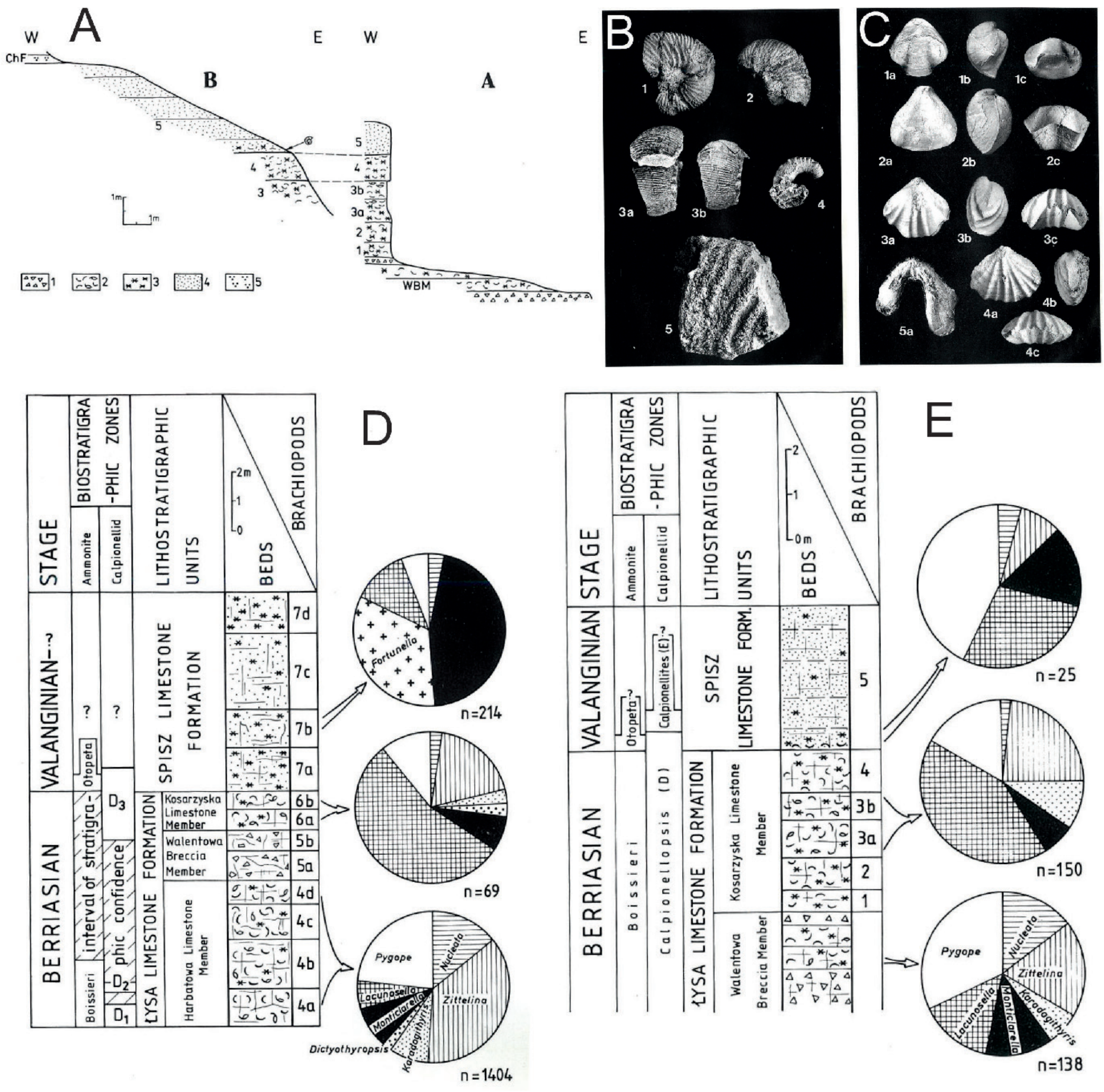


Fig. 32. Geological sections in the Biała Woda Valley at waterfall (A) (after Krobicki & Wierzbowski, 1996). Lithostratigraphy: Łysa Limestone Formation (WBM – Walentowa Breccia Member, beds 1-4 Kosarzyska Limestone Member); Spisz Limestone Formation (bed 5); Chmielowa Formation (ChF). Lithology: 1 – breccias; 2 – brachiopod limestones; 3 – crinoid calcarenites; 4 – fine-grained limestones consisting mostly of crinoid and shell debris; 5 – micritic limestones, *Hedbergella* microfacies; ammonite finds indicated; B, C – ammonites (B) and brachiopods (C) of the Berriasian and Valanginian age (B: 1, 2 – *Jeanihieuloyiles* sp.; Upper Valanginian, Spisz Limestone Fm., Korowa Klippe; 3 – *Olcostephanus* sp.; Lower Valanginian. Spisz Limestone Fm., Biała Woda Valley, section B. lowermost part of bed 5; 4 – *Rodighieroites* sp., Upper Valanginian, Spisz Limestone Fm., Korowa Klippe; 5 – ?*Dicostella* sp., Upper Valanginian, Spisz Limestone Fm., Korowa Klippe; C: 1 – *Zittelina pinguicula* (Zittel); Upper Berriasian, Walentowa Breccia Member of the Łysa Limestone Fm., section A; 2 – *Zittelina wahlenbergi* (Zejszner); Upper Berriasian, Kosarzyska Limestone Member of the Łysa Limestone Fm., section A, bed 2; 3 – *Lacunosella hoheneggeri* (Suess); Upper Berriasian, Kosarzyska Limestone Member of the Łysa Limestone Fm., section A, bed 3a; 4 – *L. zeuschneri* (Zittel); Lower Valanginian, Spisz Limestone Fm., section B, lowermost part of bed 5; 5 – *Pygope janitor* (Pictet); Upper Berriasian, Walentowa Breccia Member of the Łysa Limestone Fm., section A, bed 3b); D, E – comparison between brachiopod assemblages in Berriasian-Valanginian strata (D – Czorsztyn-Sobótka Klippe and E – Biała Woda Valley). Lithostratigraphic units after Birkenmajer (1977); stratigraphy and numbering of beds in Czorsztyn-Sobótka after Wierzbowski & Remane (1992); brachiopod pie charts after Krobicki (1994, 1996) and Krobicki & Wierzbowski (1996); composition after Krobicki & Wierzbowski (2022)

Stop 6 – Biała Woda valley – “mid”-Cretaceous basaltic olistolith (Figs 29, 33, 34)

(*Nestor Oszczypko, Michał Krobicki, Dorota Salata*)

A block of basalt a few meters in diameter has long been known from the Biała Woda valley (Horwitz & Rabowski, 1929; Kamieński, 1931; Birkenmajer, 1958b, 1979). This is an olistolith occurring within conglomerates of the Jarmuta Formation (Maastrichthian–Paleocene) belonging to the Grajcarek Succession (Birkenmajer & Wieser, 1990 and references therein). The radiometric age (K-Ar) of this basalt was determined as 140 Ma ±8 Ma, an age which corresponds to the boundary between the Jurassic and Cretaceous (Birkenmajer & Wieser, 1990). More recent radiometric dating by Birkenmajer & Pécskay (2000) for both columnar and platy-jointed varieties of the basalt gave ages of 110 Ma ±4.2 Ma and 120.3 Ma ±4.5 Ma respectively, equivalent to the Barremian-Albian interval. The basalt has geochemical features of intraplate alkali basalts (Birkenmajer & Lorenc, 2008) and geochemically resembles two olistoliths in the Proč Formation in eastern Slovakia (Spišiak & Sýkora, 2009; Oszczypko *et al*., 2012). The Early Cretaceous volcanism at the northern edge of the PKB was probably related to the opening of the Magura Basin, although this theory is still under discussion (Oszczypko & Oszczypko-Clowes, 2009). Traditionally an Early/Middle Jurassic age, coeval with opening of the Ligurian–Penninic Ocean, has been accepted (see Birkenmajer, 1986; Oszczypko, 1992, 1999; Golonka *et al*., 2000, 2003; Oszczypko *et al*., 2012) (Fig. 34).

Fig. 33. View of basaltic olistolith in Biała Woda Valley

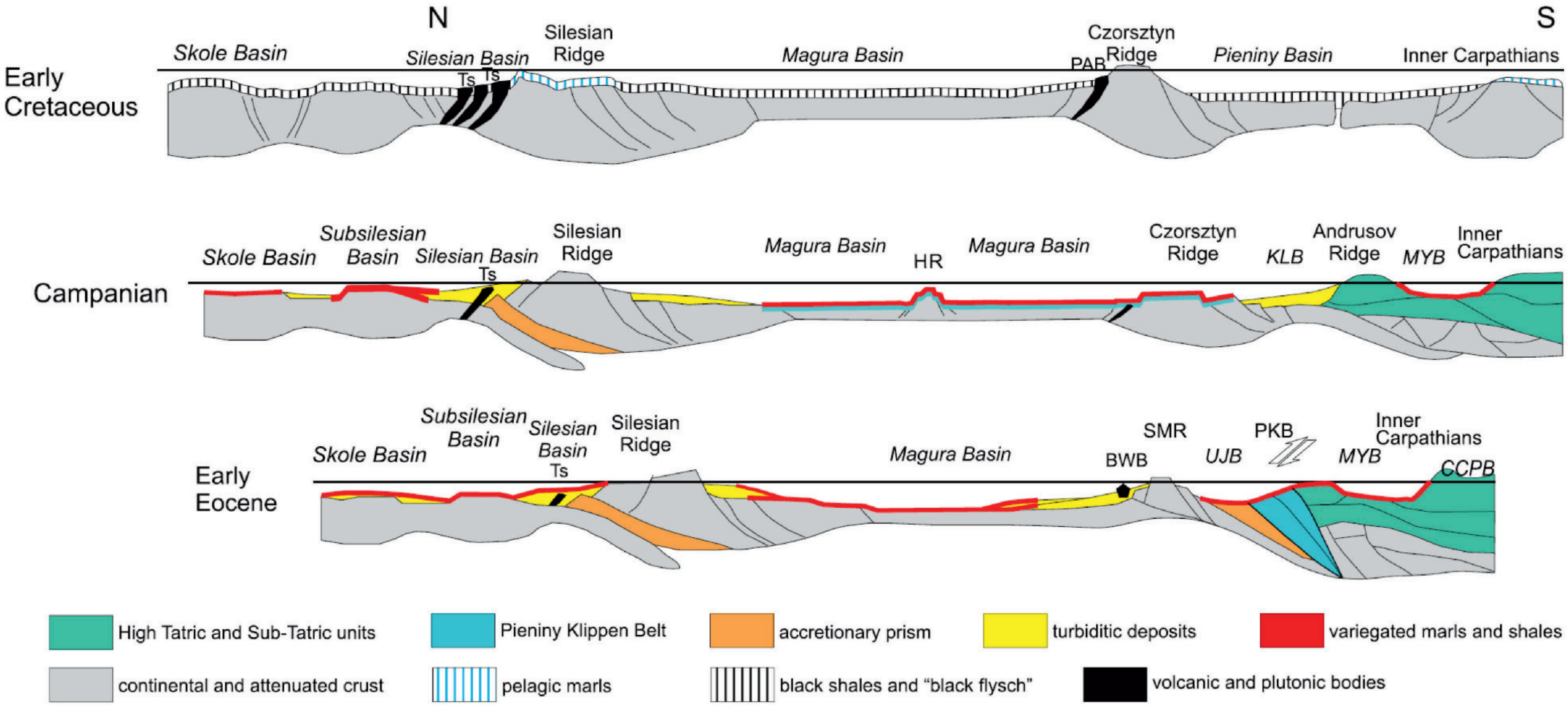


Fig. 34. Early Cretaceous–Early Eocene palinspastic evolutionary model for the Magura Basin, not to scale (based on Oszczypko, 2006, modified and Oszczypko *et al.*, 2012). CCPB – Central Carpathian Paleogene Basin, HR – Hluk Ridge, UJB – Ujak Basin, KLB – Klapa Basin, MYB – Myjava Basin, PKB – Pieniny Klippen Belt, SMR – South Magura Ridge, Ts – teschenites, PAB – Pieniny alkaline basalts, BWB – Biała Woda basaltic block

Stop 7 – Szczawnica ("Orlica") – history of the discovery of nappes in the Carpathians (Figs 35–37)

(*Michał Krobicki*)

As a representative of the young French school of alpine tectonics, 33 years old Maurice Lugeon took part in a seven-day geological field trip to the Pieniny and Tatra Mountains (from 11 to 18 July 1903) as a part of the 9th International Geological Congress organized in Vienna. The trip was led by the famous Victor Uhlig – an Austrian professor of geology at the Vienna University, author of synthetic studies of the Tatra and Pieniny geology (Uhlig 1890a, 1890b, 1891, 1897, 1903; Sokołowski, 1954b). In these two people, different concepts of origin of tectonic structures within Tatra and PKB clashed. V. Uhlig proposed adopting the autochthonism of fold structures, both for the Tatra Mountains and for the PKB (Limanowski, 1904, 1905; Świderski, 1923; Sokołowski, 1954a, 1954b), while M. Lugeon preferred their nappe style of tectonic structure (Lugeon, 1902a, 1903), even though he had never been to Poland before the aforementioned trip! This alpine geologist, mapping complex structure of the Alps of the Swiss-French borderland, he was a staunch supporter new, nappe interpretation of their structure. Relying only on the perfect geological maps of V. Uhlig, he came to the conclusion about a similar, as the Alps, tectonic style of the Polish Carpathians, including the Tatra and the Pieniny mountains (Limanowski, 1905). Already on the first day of this field trip (August 11, 1903), passing from Nowy Targ to Czorsztyn village and seeing isolated klippen of the Pieniny Mountains in the landscape very similar to Chablais region in the French Alps that was the object his doctoral thesis in 1895. In 1902 year, he published a note in which he presented tectonic analogies between the geology of the Alps and the Carpathians (Lugeon, 1902a, 1902b), and then extended this thesis in more detail the following year (Lugeon, 1903). The most likely in the vicinity of today's PTTK hostel "Orlica" in Szczawnica, in August 12, 1903, decisive observations and discussions took place about his suppositions as to the nappe genesis also of this part of the Alpine orogen (Krobicki, 2022c).

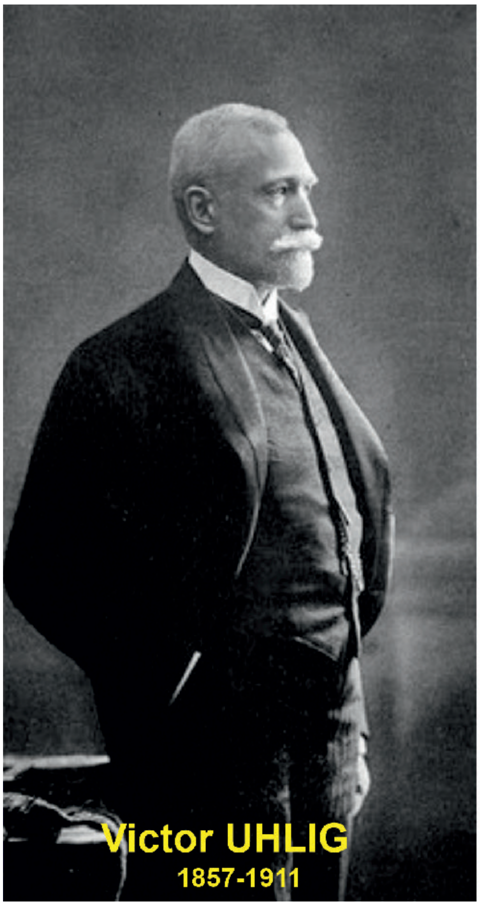
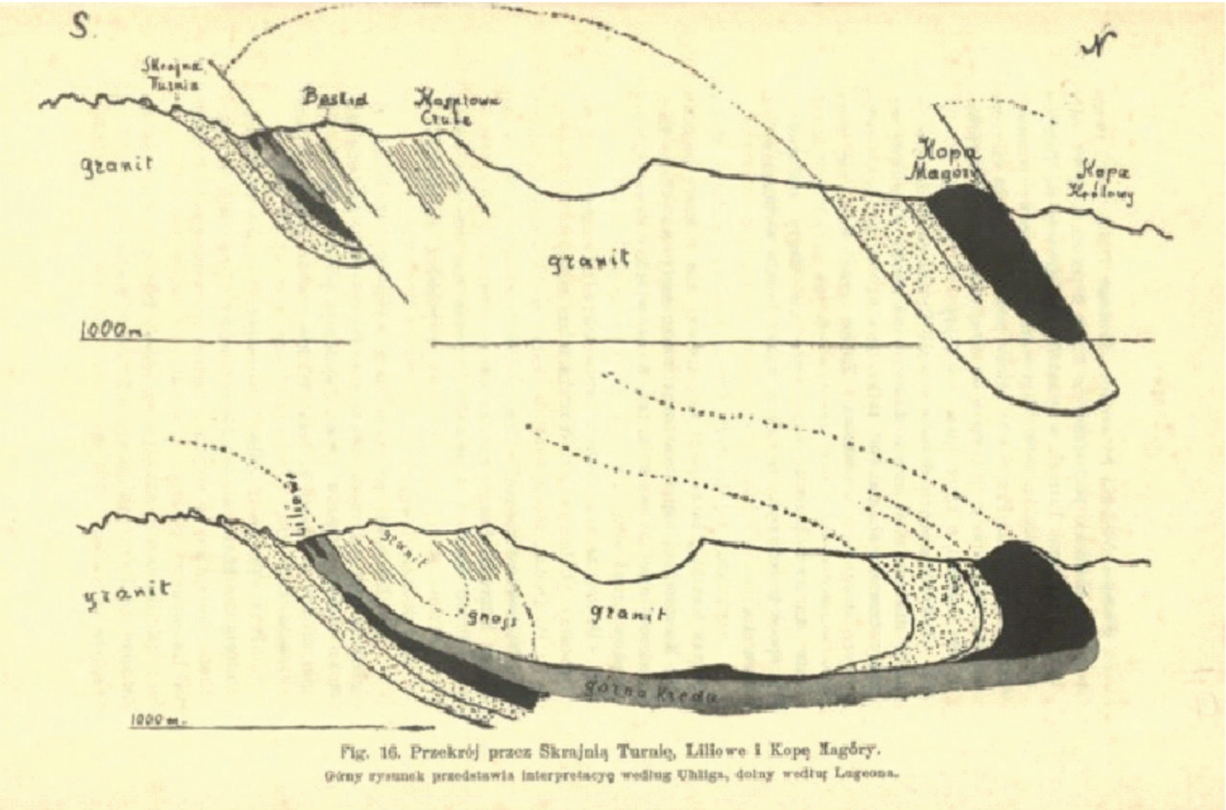


Fig. 35. Two prominent scientists – specialists of the Pieniny Klippen Belt geology



Organizator

Współorganizatorzy

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UNIWERSYTET JAGIELLOŃSKI W KRAKOWIE

Instytut Gospodarki Surowcami Mineralnymi i Energią Polskiej Akademii Nauk

88. Zjazd Naukowy Polskiego Towarzystwa Geologicznego

18-20 maja 2023 r., Dobczyce

Perspektywy badań geologicznych, hydrogeologicznych i poszukiwań złóż w Małopolsce w 120-lecie teorii płaszczowinowej w Karpatach

Viktor Uhlig

Maurice Lugeon

Partnerzy:

TERMY BUKOVINA

PSG

DOBCZYCE

www.ptgeol.pl/2023/01/16/88-zjazd-naukowy-ptgeol/

Fig. 36. Different interpretations of tectonic position of the Tatra Mts structures: upper cross section – Uhlig's idea and lower cross section – Lugeon's idea; lower position – circular of the 88th Polish Geological Society Meeting organised in 120 anniversary of the 9th International Geological Congress in Vienna (1903)

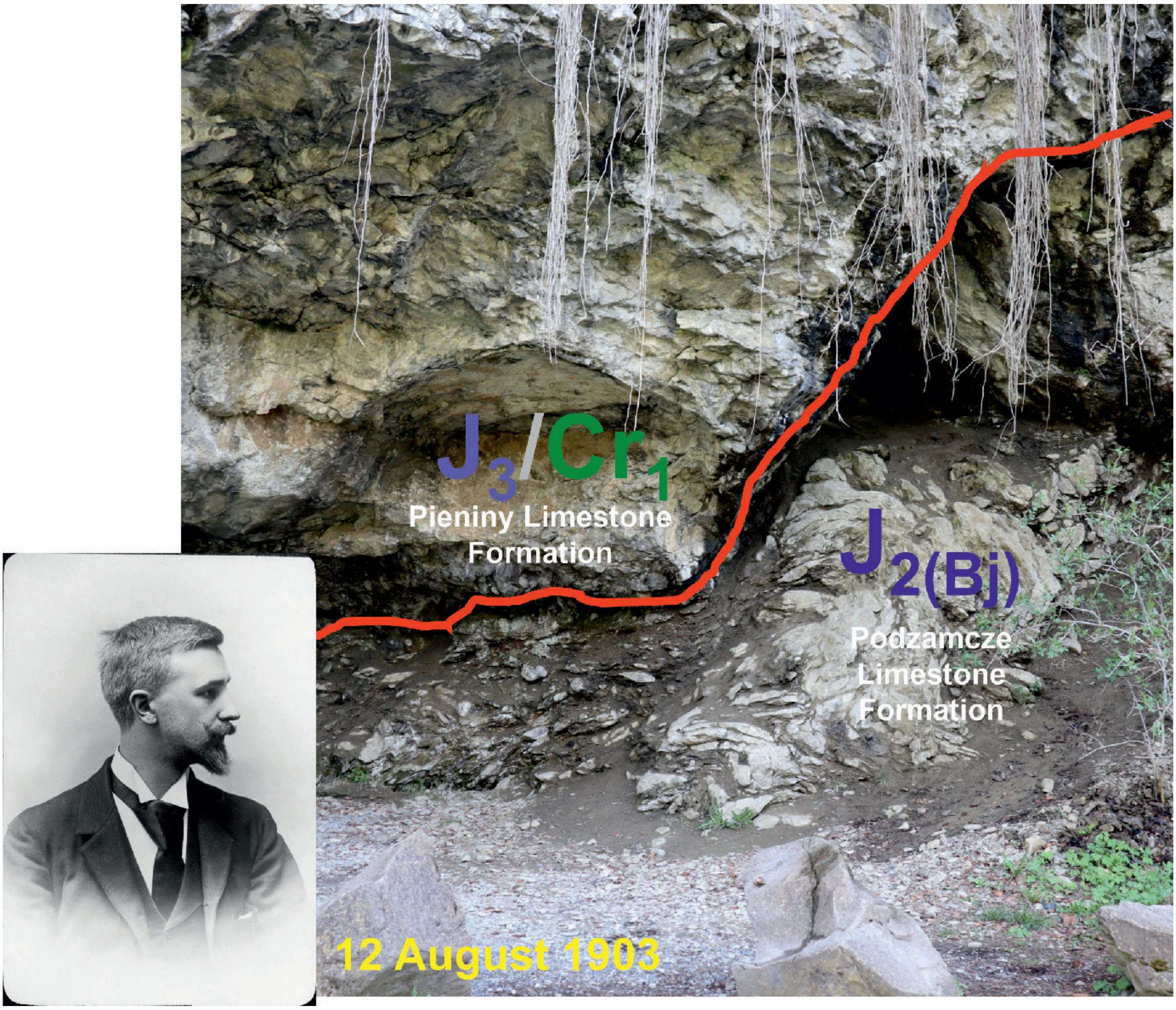

Fig. 37. Zyblikiewicz's cave in Szczawnica Niżna, along so-called "Pieniny road" along Dunajec River nearby of Orlica hostel – 120 anniversary of Maurice Lugeon visit of this place during International Geological Congress (Vienna'1903) – tectonic contact between two nappes?/thrust sheets?

Stop 8 – Flaki range (Jurassic deposits of the Branisko Succession) (Figs 38, 39)

(*Michał Krobicki, Jarosław Tyszka, Alfred Uchman*)

At road cutting through the Flaki Range we can see an outcrop of the Branisko Succession developed as: grey crinoid-cherty limestones and overlying greenisch micritic limestones and green chamosite-bearing marls (Flaki Limestone Formation), black-brown manganiferous and green radiolarites of ?Bathonian-Callovian-Oxfordian age (Sokolica Radiolarite and Czajakowa Radiolarite formations) (Birkenmajer, 1977) (Fig. 38). These rocks are surrounded by less resistant Upper Cretaceous marls and flysch siliciclastics belonging to different tectonic units of the PKB. At the road cut in the Flaki Range, the Branisko Succession crops out in tectonically overturned position. They are deep-water stratigraphic equivalent of shown earlier in the Czorsztyn Castle shallow-water facies of crinoidal and red nodular limestones of the Czorsztyn Succession (Myczyński, 1973; Birkenmajer, 1977, 1979, 1985). The Flaki Limestone Formation represents a condensed sequence of grey filament limestones, spiculites and green filament marls with ferruginous (chamositic) oncoids. The filament limestones and marls consist of pelagic bivalve *Bositra* shells.

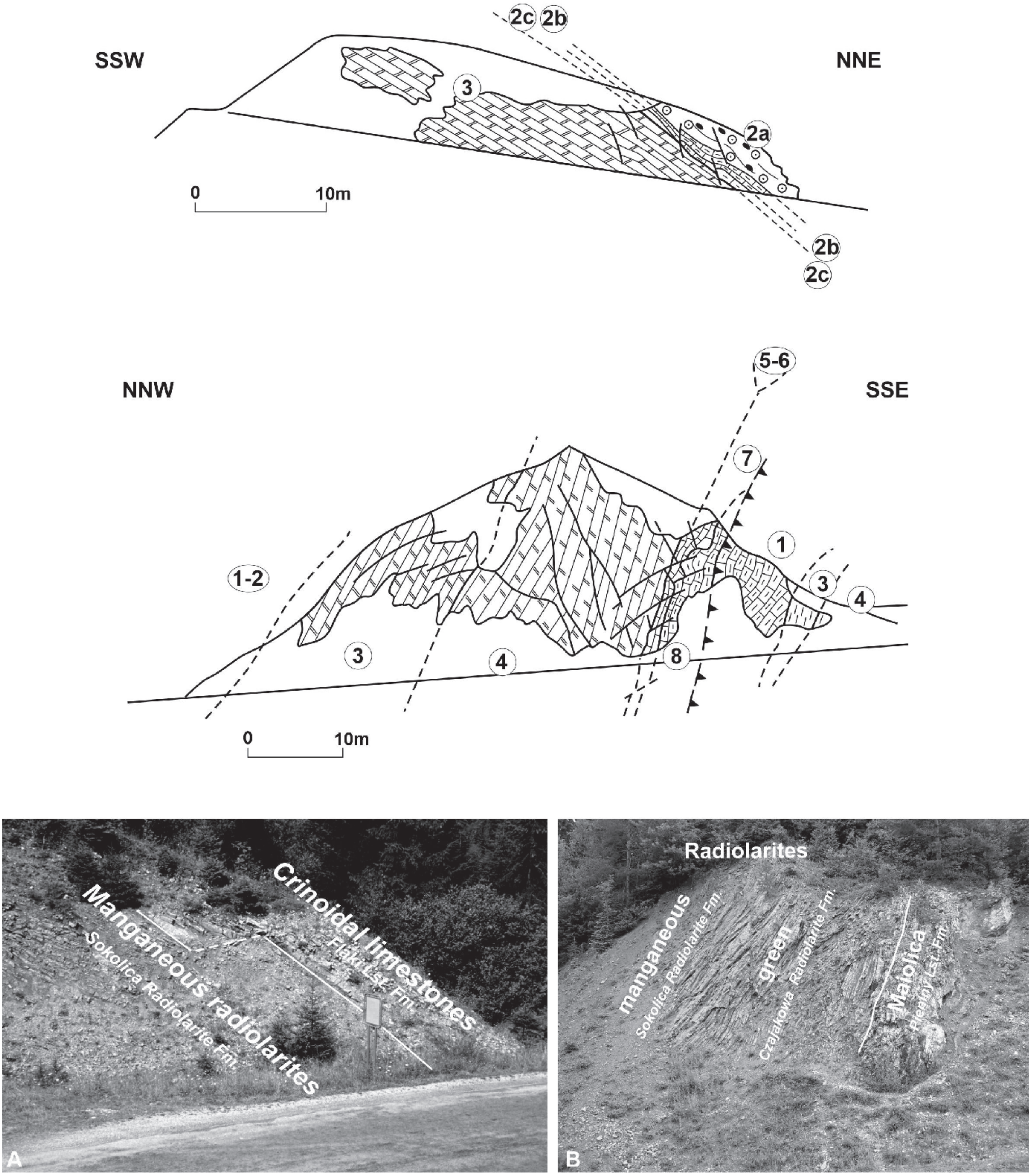

Fig. 38. View of the Flaki Range sections; Branisko Succession (lower part: A – western side; B – eastern side) and general sketch of studied sections (upper part) (after Birkenmajer *et al*., 1985). Lithostratigraphical units: 1 – Podzamcze Limestone Fm.; 2 – Flaki Limestone formations (grey crinoidal limestones with cherts in upper part (2a) and grey-green limestones (2b) and marls with chamosite concretions (2c); 3 – Sokolica Radiolarite Fm. (grey-black manganiferous spotty radiolarites); 4 – Podmajerz Radiolarite Mbr of the Czajakowa Radiolarite Fm. (green radiolarites); 5–6 – Czajakowa Radiolarite Fm. (Buwałd Radiolarite Mbr – red radiolarites) and Czorsztyn Limestone Fm. (Upszar Limestone Mbr – white nodular limestones) exposed upslope further east; 7 – Pieniny Limestone Fm. (micritic limestones with cherts of the maiolica-type facies) (strongly tectonically reduced); 8 – Kapuśnica Fm. (greenish spotty marls/limestones) (after Krobicki *et al*., 2006)

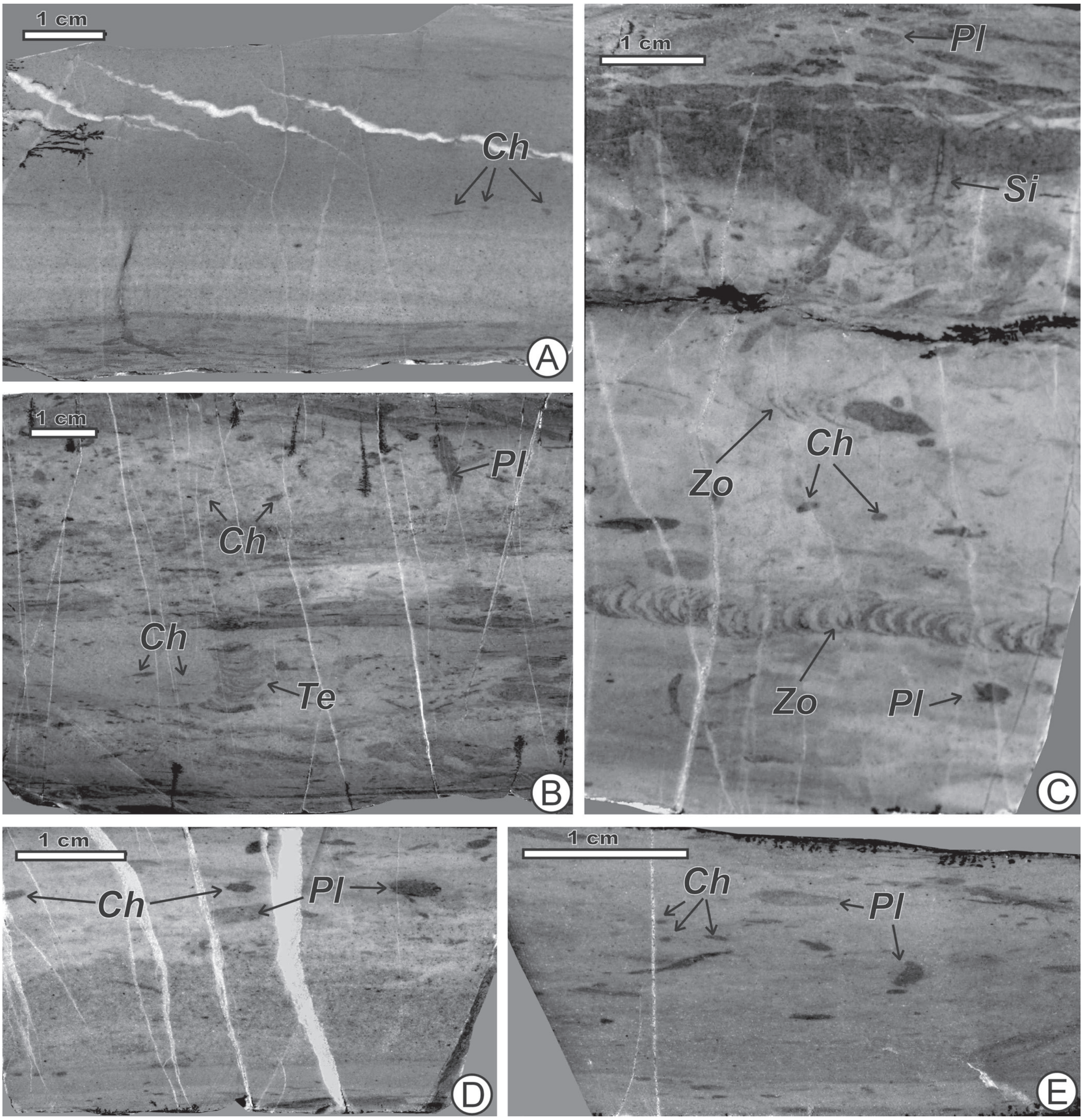

Fig. 39. Trace fossils within Sokolica Radiolarite Fm. in the Flaki Range, Branisko Succession: *Ch* – *Chondrites*; *Pl* – *Planolites*; *Si* – *Siphonichnus*; *Te* – *Teichichnus*; *Zo* – *Zoophycos* (after Krobicki *et al*., 2006)

In several radiolarite beds of Middle Jurassic manganiferous radiolarites (Sokolica Radiolarite Formation), normal graded bedding is noted in layers. In the layers trace fossils are abundant (common *Planolites* and *Chondrites*, less frequent *Taenidium* and *Teichichnus*, rare *Siphonichnus* and *Zoophycos*) (Krobicki *et al*., 2006). They belong to ichnogenera produced in the deepest tiers in the sediment. The trace fossil assemblage is typical of deep-sea fine-grained sediments deposited in well-oxygenated sea floor. Very little ichnological data come from radiolarites, however lately Kakuwa (2004) presented their ichnofabric from the Triassic and Jurassic of Japan.

Stop 9 – Sromowce – Upper Cretaceous *Scaglia Rossa* with clastics (Figs 40, 41)

(*Michał Krobicki, Jan Golonka*)

One of the major attraction of the PKB region is the rafting through the Dunajec River Gorge (Golonka & Krobicki, 2007; see also Alexandrowicz & Alexandrowicz, 2004). The rafting trip on the Dunajec River, which starts at Sromowce Kąty harbour, takes geotourist through the Dunajec Gorge to Szczawnica. The Dunajec offers magnificent view of the cliffs sculptured in the Pieniny Mountains by the tectonic activity and river's erosion. It offers also the close view of the outcrops of Jurassic and Cretaceous rocks of the Pieniny Succession and complex tectonics of the PKB.

Strongly folded Jurassic-Cretaceous strata are visible along the road from Sromowce Wyżne to Sromowce Niżne, close to the Dunajec River, on the southern slope of Mt. Macelowa, where the Pieniny Succession rocks lie in an overturned position.

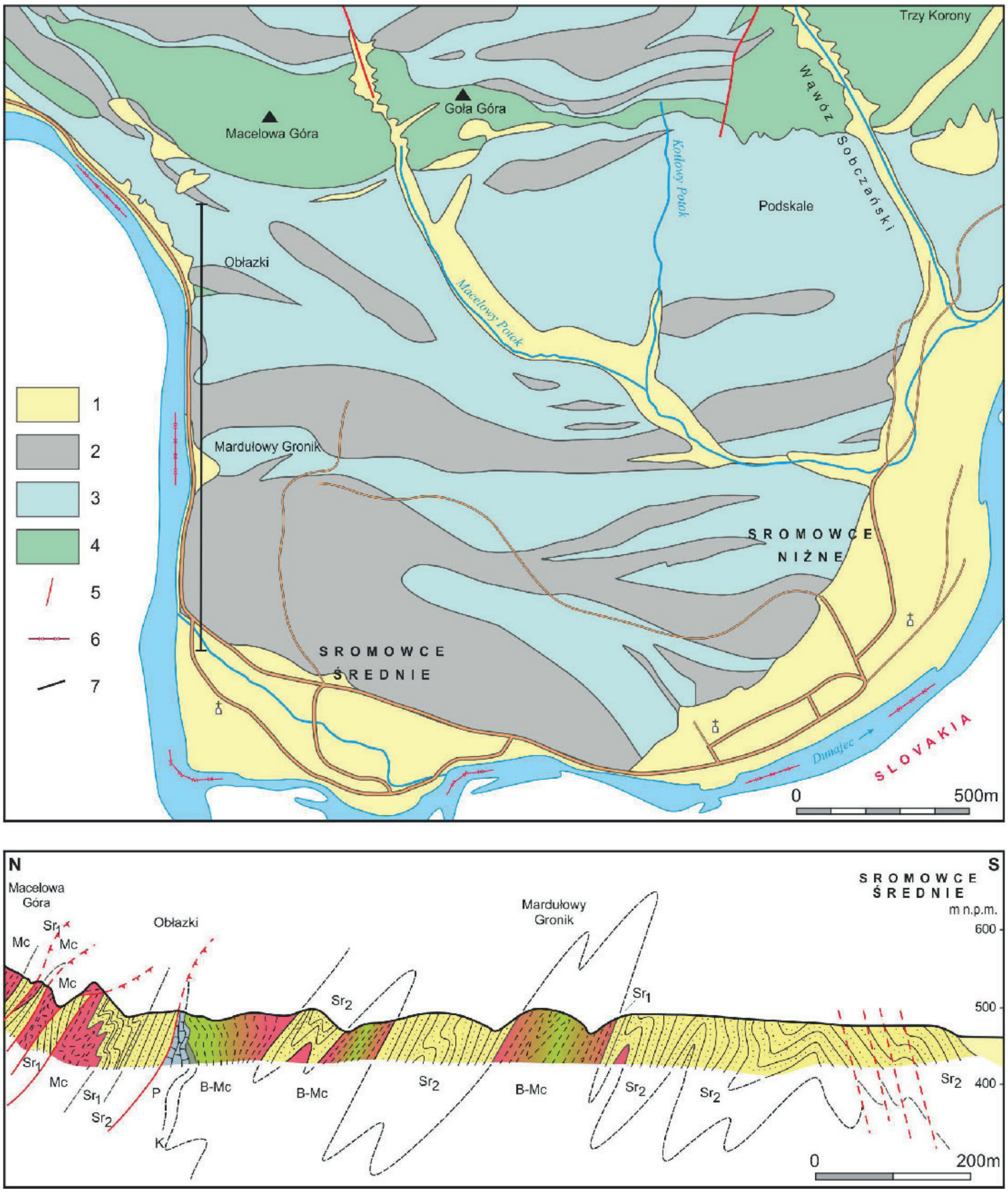

Fig. 40. Geological map of the vicinity of Sromowce (after Horwitz, 1963; Birkenmajer & Jednorowska, 1984, simplified) and geological cross-section: 1 – Quaternary; 2 – Sromowce Formation; 3 – Jaworki Formation, partly Kapuśnica Formation; 4 – Pieniny Limestone Formation, partly also Czajakowa Radiolarite Formation; 5 – faults; 6 – state border; 7 – geological cross-section (below); P – Pieniny Limestone Formation (grey cherty limestones); K – Kapuśnica Formation (green spotty marls); B-Mc – Jaworki Formation (Brynczkowa, Skalski and Macelowa Marl members – green, variegated and red marls respectively); Sr – Sromowce Formation (Sr1 – Osice Siltstone Member; Sr2 – flysch) (after Golonka *et al*., 2018)

Fig. 41. Aerial view of the central Pieniny Mts. and Dunajec River Gorge with points of photos: A – Upper Cretaceous red marls of the *Scaglia Rossa*-type facies of the Macelowa Marl Member of the Jaworki Formation (Macelowa Mt.); B – close to beginning of the rafting in Sromowce-Kąty harbor in the Pieniny Mts., boat full of tourists; C, D – Trzy Korony Mountain built of *Maiolica*-type well-bedded cherty limestones of the Pieniny Limestone Formation, usually strongly tectonically folded; E – Sokolica Mt. over the Dunajec River Gorge (after Krobicki & Golonka, 2008)

The oldest Oxfordian radiolarites occupy the topmost part of Mt. Macelowa (on its northern slope), gray cherty limestones of the *Maiolica* facies (Pieniny Limestone Formation) occupy the transitional position and in the lowest (topographically) position are the Late Cretaceous *Globotruncana*-bearing marls of the *Scaglia Rossa*-type (Birkenmajer, 1977; Bąk K., 1998, 2000). Figure 15 depicts the Birkenmajer & Jednorowska (1987a, 1987b) ideas about the Cretaceous lithostratigraphy of the Pieniny Mountains. Red marls and marly limestones of pelagic deposits with grayish intercalations of calcareous sandstones and siltstones of distal turbiditic origin predominate in this outcrop. This is the youngest part of the multicolored (green-variegated-red) globotruncanid marls of the so-called Macelowa Marl Member of the Jaworki Formation, with good foraminiferal Upper Cretaceous biozonation (*Dicarinella concavata – D. asymmetrica* foraminiferal zones of the Upper Coniacian-Santonian) (Bąk K., 1998, 2000). These deposits originated during the final episode of the evolution of the PKB, when the unification of sedimentary facies took place within all the successions. Widespread in the Late Cretaceous Tethyan Ocean, the *Scaglia Rossa*-type facies (= *Couches Rouge* = *Capas Rojas*) represented by the Jaworki Formation – which were widespread in the Late Cretaceous Tethyan Ocean – indicates open connections throughout the Northern Tethys.

Stop 10 – Haligovce/Lipnik – Paleocene reefs after C/P mass extinction event (Fig. 42)

(*Michał Krobicki, Jan Golonka*)

In the vicinity of the villages of Haligovce and Lipnik, within the Paleogene flysch, there are large blocks of Paleocene olistolithic limestones, the oldest deposits of the so-called of the Pieniny Paleogene (Scheibner, 1968; Potfaj, 2002; Krobicki *et al*., 2004; Buček & Köhler, 2017). Their palaeontological and microfacial analysis showed the presence of numerous corals, red algae and foraminifera as well as bryozoans, serpulids, fragments of bivalves, brachiopods, echinoderms, sponges, and sporadically sponge spicules. Analysis of small fragments of coral colonies revealed the presence of scleractinians (*Asirocoenia*, ?*Acropora*, *Goniopora*, *Actinacis*, *Rhizangia*, *Orbignygyra*, *Favites*, *Oculina*, and ?*Rabdophylliopsis*) (Krobicki *et al*., 2004). Corals are often coated by the red algae Corallinaecae (the most numerous) and Peyssonneliaceae (*Polystrata alba*), with detritus being the most common in the carbonate matrix. Foraminifera are represented by large forms of encrusting agglutinating foraminifera (e.g. *Haddonia* sp.) and calcarous. General taxonomic composition of the corals most closely resemble the Paleocene corals from Slovenia. According to the palaeogeographic evolution of this area in the Paleocene time, the PKB was closed as a result of the collision of the Central Carpathians terrains with the Czorsztyn Ridge (Birkenmajer, 1986, 1988). The terrains of Adria, the Eastern Alps and the Inner Carpathians continued their movement northward. The Paleocene subsidence of the Magura Basin was associated with the shift of the subduction zone north of the Czorsztyn Ridge. In the Paleocene, the Alcapa superterrain was formed by combining the blanks of the Eastern Alps, the Tisza, the Inner Carpathians and other small terrains. At the same time, the aforementioned Paleocene reefs (Mišík & Zelman, 1959; Andrusov, 1969; Scheibner, 1968; Samuel *et al*., 1972; Köhler *et al*., 1993; Buček & Köhler, 2017) were formed in the shallowest zones of the basin, today's isolated occurrences of which can be found from the Eastern Alps (near Kambühel near Ternitz in Austria, the stratotype of the so-called Kambühel limestones, Tollmann, 1976; Faupl *et al*., 1987; Tragelehn, 1996; Müller, 2004) through western Slovakia (Mišík & Zelman, 1959; Scheibner, 1968; Köhler *et al*., 1993) to vicinity of Haligovce (Scheibner, 1968; Potfaj, 2002; Krobicki *et al*., 2004; Buček & Köhler, 2017). Identical limestones have also been found to be exotic within the boundaries of the Strihov and Proč strata of Western Slovakia (Mišik *et al*., 1991a, 1991b).

Stop 11 – Czorsztyn Castle (Jurassic-Cretaceous deposits of the Czorsztyn Succession) (Fig. 43)

(*Michał Krobicki*)

The Czorsztyn Castle klippen are one of the most famous geological site of the PKB with full sequence of Czorsztyn Succession from the Middle Jurassic up to Upper Cretaceous deposits, rich in invertebrate fossils such as: ammonites, brachiopods, crinoids, calpionellids, foraminifers, described and illustrated by numerous authors since beginning of the XIX century (e.g., S. Staszic, L. Zejszner, E. Suess, M. Neumayr, K. A. Zittel, V. Uhlig and others) (Uhlig, 1890a; Birkenmajer, 1963, 1977, 1979, 1983; Barczyk, 1972a, 1972b; Głuchowski, 1987; Krobicki, 1994, 1996b; Wierzbowski & Remane, 1992; Wierzbowski *et al*., 1999). Unfortunately, the water of present Czorsztyn lake covered the great part of this sequence (lowermost – lower part of the Middle Jurassic and upper part – Upper Cretaceous) and only Bajocian-Berriasian interval is available to study (partly by means of boat).

The Czorsztyn Castle klippen (more precisely – so-called Sobótka klippe) is a stratotype for the Czorsztyn Limestone Formation (red nodular limestone of the *Ammonitico Rosso* type facies; uppermost Bajocian-Tithonian in age) (Birkenmajer, 1977; see also Birkenmajer, 1963). In this section the oldest are grey crinoid limestones of the Smolegowa Limestone Formation. These are well-bedded grainstones and the youngest beds are cross-bedded well recorded shallow marine origin of these limestones. Gastropod trace fossils found in the base of these limestones supported such idea (Krobicki & Uchman, 2003). There follow thin-bedded reddish crinoid limestones of the Krupianka Limestone Formation with considerable amount of hematite-marly matrix. The ammonites are very rare and poorly preserved but brachiopods are rather common: *Capillirhynchia brentoniaca* (Oppel), *Septocrurella* ? *defluxa* (Oppel), *S. kaminskii* (Uhlig), *Linguithyris curviconcha* (Oppel), *Karadagella zorae* Tchorszhevsky et Radulović and *Zittelina* ? *beneckei* (Parona). The geological age of the crinoid units is Bajocian: the basal part of the Smolegowa Limestone Formation in the Sobótka Klippe of the Czorsztyn Castle klippes yielded ammonites – *Dorsetensia* (*Dorsetensia*, *Nannina*), *Pelekodites*, *Stephanoceras*

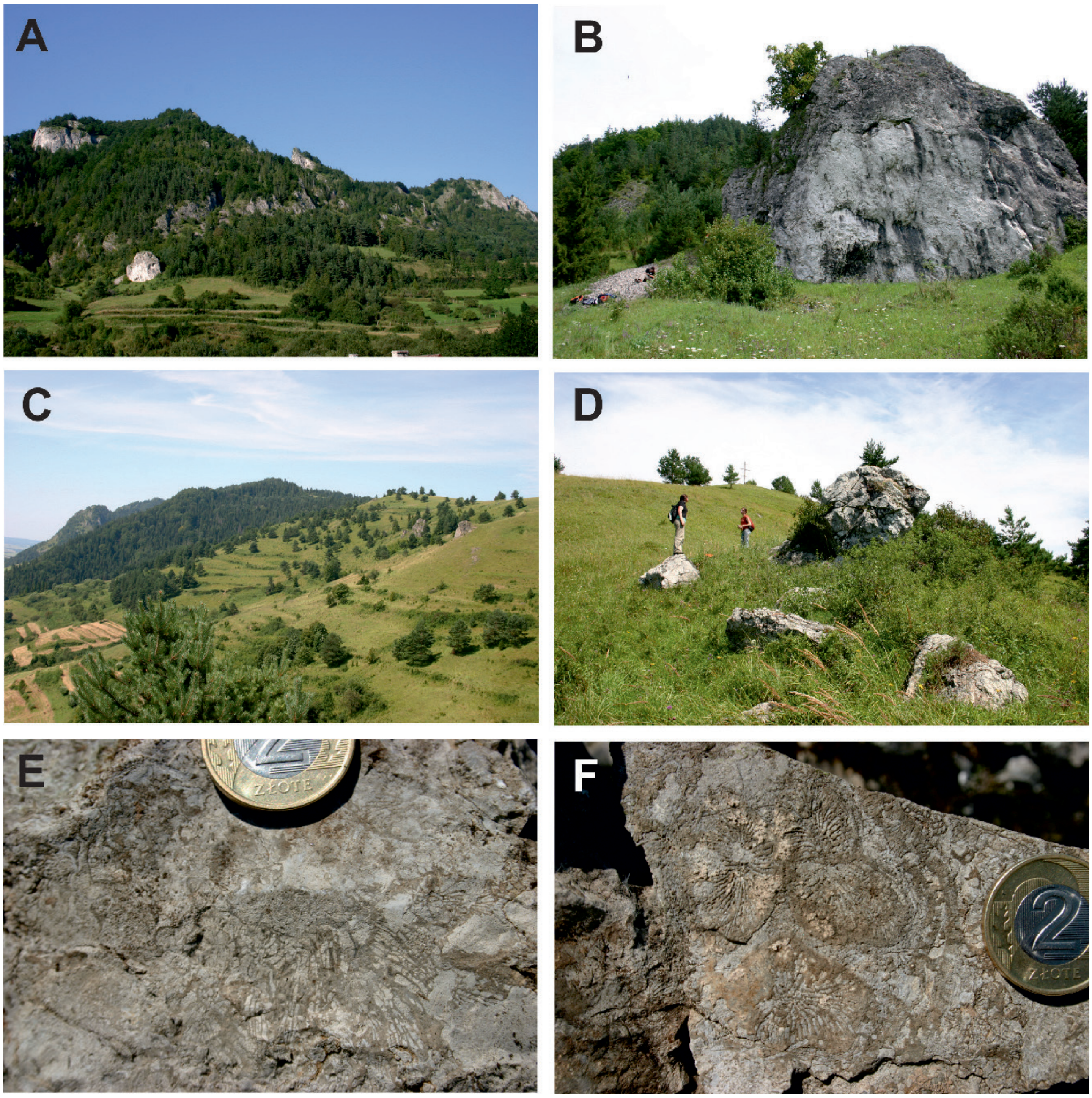

Fig. 42. View of Haligovce Klippen (A–C) and Lipnik Klippen (C, D) with olistoliths of Paleocene coral-bearing limestones (E, F) of the so-called Kambühel limestones (after Krobicki, 2022d)

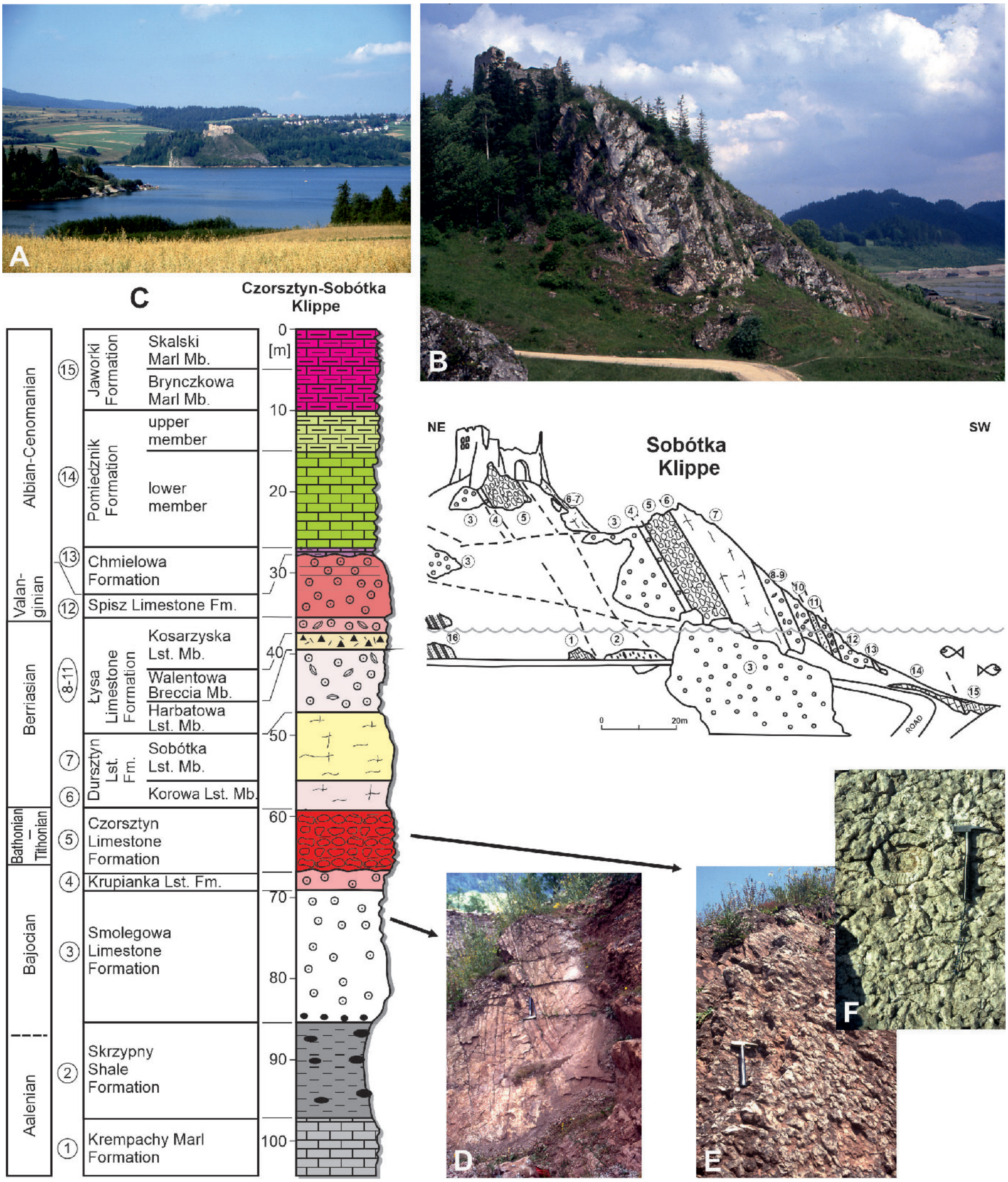

Fig. 43. Stratigraphical section of the Czorsztyn-Sobótka Klippe (A, B) with indication of position of the Walentowa Breccia Member of the Łysa Limestone Formation of the Czorsztyn Succession (C) (lithostratigraphy after Birkenmajer, 1977, slightly modiefied) (photo – state in 1992). Explanations of lithology: 1 – dark-grey/black marls/marly limestones; 2 – black spherosideritic shales; 3 – white cridoidal limestones (with phosphatic concretions in base – black dots); 4 – red/pink crinoidal limestones; 5 – red nodular limestones; 6 – pink micritic *Calpionella*-bearing limestones; 7 – creamy micritic *Calpionella*-bearing limestones; 8 – creamy brachiopodic- crinoidal limestones; 9 – limestone sedimentary breccia; 10 – pink-creamy brachiopodic-crinoidal limestones; 11 – cherry crinoidal limestones; 12 – violet-red marls; 13 – green marls, sometimes with cherts; 14 – green and variegated *Globotruncana*-bearing marls (formal litostratigraphical names of units – see Fig. 9) (after Krobicki *et al*., 2010)

(*Stephanoceras*, *Skirroceras*) which are indicative of the upper part of the Lower Bajocian (upper Propinquans Zone, and the Humphriesianum Zone) – see Krobicki & Wierzbowski (2004), whereas the nodular limestones directly overlying crinoid limestones of the Krupianka Limestone Formation in the Czorsztyn Castle Klippe section yielded ammonites of the uppermost Bajocian (Wierzbowski *et al.*, 1999; Krobicki *et al*., 2006). The upper surface of the topmost bed of the red crinoidal limestones is corroded and covered with ferro-manganese crust, very typical feature for this boundary surface, known from several other outcrops in the PKB, both in Polish, Slovakia and Ukrainian part of the region. The overlying nodular limestones correspond already to the Czorsztyn Limestone Formation (red nodular limestone). The lowermost part of the nodular limestones of the Czorsztyn Limestone Formation exposed in the Czorsztyn Castle Klippe yielded the rich ammonite faunas. These ammonites are indicative to uppermost Bajocian, Bathonian, and Callovian up to Oxfordian. The whole uppermost Bajocian up to uppermost Callovian and/or Oxfordian interval does not exceed 2.0 meters, therefore the oldest part of the Ammonitico Rosso type limestones (of the Czorsztyn Limestone Formation) represents very condense sequence (Wierzbowski *et al*., 1999). The whole Cretaceous strata were visible previously (from Berriasian limestones up to Santonian marls), including very characteristic synsedimentary limestone breccias of the so-called Walentowa Breccia Member of the Łysa Limestone Formation (Berrisian in age) which indicates the earliest Cretaceous (Neo-Cimmerian) tectonic movements in this part of the Tethys.

Stop 12 – Wżar Mt (Miocene andesites and panoramic view) (Fig. 44)

(*Jan Golonka, Michał Krobicki*)

The most famous outcrop (artificial one – abandoned quarry) of the Middle Miocene volcanism of the Pieniny Mts occur on the Wżar Mt, near Snozka pass, and is represented by two generation of intrusive dykes and sills. In half of the XX century several pioneer researches were done both geologically, mineralogically/petrographically and geophysically (e.g. Wojciechowski, 1950, 1955; Birkenmajer, 1956a, 1956b, 1958b; Kardymowicz, 1957; Małoszewski, 1957, 1958; Gajda, 1958; Kozłowski, 1958; Małkowski, 1958). The Neogene volcanic activity in Carpathian–Pannonian region was widespread. The Pieniny Andesite Line is an about 20 km long and 5 km wide zone, which cut both Mesozoic-Paleogene rocks of the PKB and Paleogene flysch of the Magura Nappe of the Outer Flysch Carpathians. Andesites occur in the form of dykes and sills. At the Wżar Mt two generations of andesitic dykes occur (Youssef, 1978). Numerous older dykes are sub-parallel to the longitudinal distribution of the PKB structure and younger are perpendicular to the first and are represented only by three dykes (Birkenmajer, 1962, 1979; Birkenmajer & Pécskay, 1999). Spatial distribution, temporal relationships, and geochemical evolution of magmas contribute to interpretation of the geodynamic development of this area (e.g., Birkenmajer, 1986; Kováč *et al*., 1998; Golonka *et al*., 2005a, 2005b).

The Wżar Mt represents the westernmost occurrence of andesites in the Pieniny region. Amphibole-augite and/or augite-amphibole andesites dominate in the Mt Wżar area. Numerous petrographical varieties were distinguished, based mainly on the composition of phenocryst assemblages (Michalik M. *et al*., 2004, 2005; Tokarski *et al*., 2006). The mainly Sarmatian age of first phase of andesite dykes from this quarry, which are parallel and subparallel with the northern boundary fault of the PKB, radiometrically determined as 12.5–12.8 Ma (K-Ar method) (Birkenmajer & Pécskay, 2000; Trua *et al*., 2006). The second, younger generation of dykes follows transversal faults, which cut the older generation (Birkenmajer, 1962) and is dated on 10.8-12.2 Ma (Birkenmajer & Pécskay, 2000; Birkenmajer, 2001). These calc-alkaline andesites interpreted by Birkenmajer (2001) as products of hybridization of primary mantle-derived magma over subducted slab of the North European Plate (Birkenmajer & Pécskay, 1999) connected with collision-related post-Savian tectonic, compression event. The newest results of andesitic rocks investigations indicate partial melting derived from an ancient metasomatized, sub-continental lithospheric mantle. Generation of the calc-alkaline magmas in the upper lithospheric mantle was effect of collision of the Alcapa block with southern margin of the European platform (Anczkiewicz & Anczkiewicz, 2016; see also Trua *et al*., 2006).

These andesitic rocks cut Upper Cretaceous and Paleogene flysch deposits of the autochthonous Magura Nappe (the Szczawnica, Zarzecze and Magura formations), which is the southernmost flysch tectonic unit of the Outer Carpathians – near northern strike-slip–type faults of the PKB. Near the entrance to this quarry contact metamorphism and hydrothermal activity within flysch sandstones are good visible (Birkenmajer, 1958b; Gajda, 1958; Małkowski, 1958; Michalik A., 1963; comp. Szeliga & Michalik, 2003). Two stages of magmatic activity resulted also in chemical variation in composition of surrounding sandstones (Pyrgies & Michalik, 1998). The similar Miocene volcanic activity is widespread within whole Carpathian-Pannonian region and can be use to geodynamic interpretation of syn-orogenic magmatic events of these regions (e.g., Kováč *et al*., 1997; Anczkiewicz & Anczkiewicz, 2016 with references cited therein).

Wżar Mt is one of the geological objects classified for the entry into the European network of GEOSITES (Alexandrowicz, 2006) and mining activity of prospecting and excavation of magmatic ore deposits connected with Pieniny andesites were known since beginning of the XV century (Małkowski, 1958).

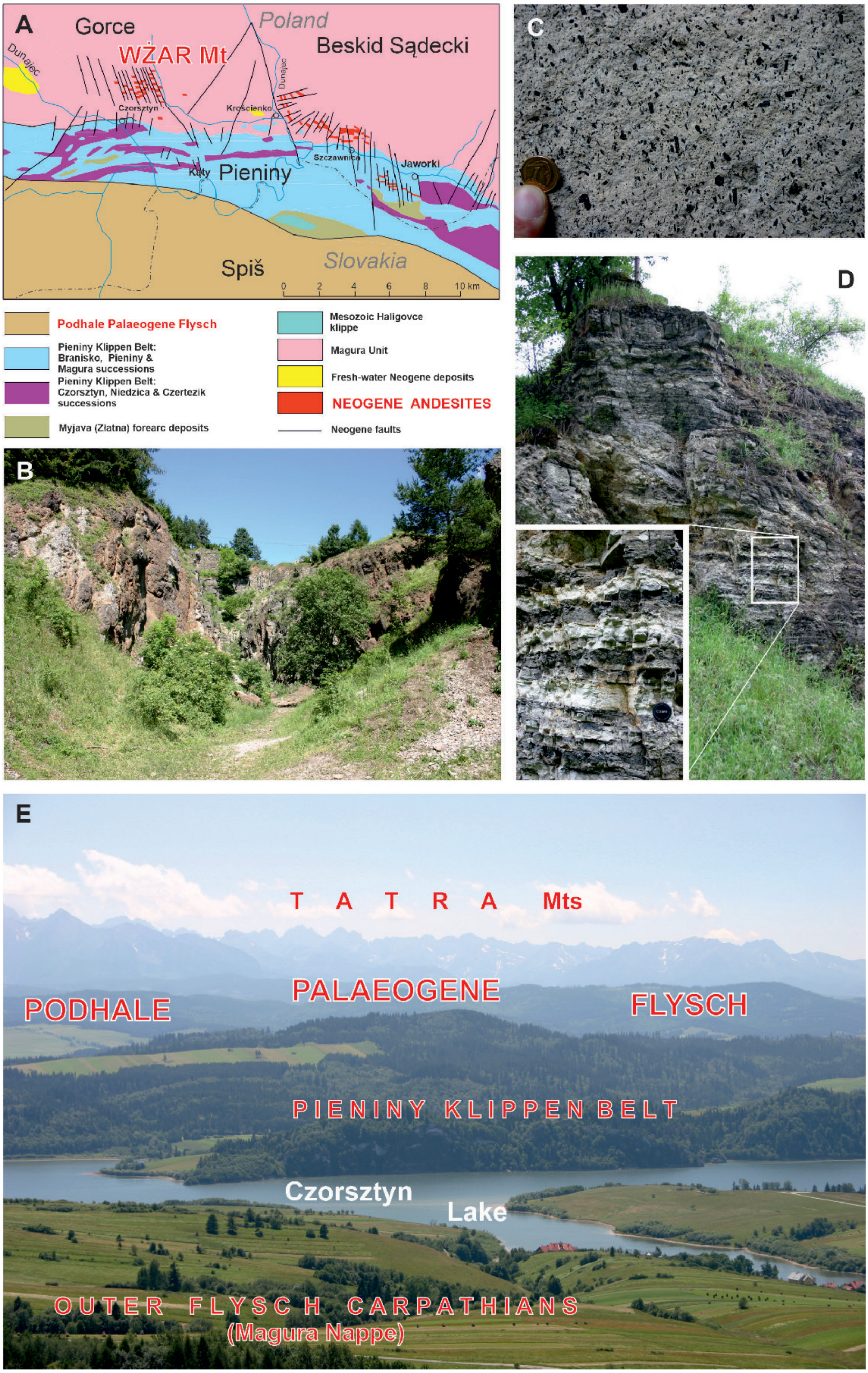

Fig. 44. Geological position of Miocene andesites of the so-called Pieniny Andesite Line: A – geological sketch of the Pieniny Klippen Belt (Polish sector) and surrounding regions (after Birkenmajer, 1979; simplified) with location of Wżar Mt; B – main entrance to abandoned quarry; C – andesites with piroxenes and amphibolites; D – thermally change of flysch deposits of the Magura Unit (Outer Flysch Carpathians) on the contact with andesites; E – general view of Inner Carpathians from topmost part of the Wżar Mt (after Krobicki & Golonka, 2008)

Finally, when looking southward, we can see perfect panorama of Tatra Mountains, Pieniny and Podhale trough with Czorsztyn Lake, and looking northward of Gorce Mountains are visible (see Golonka *et al*., 2005b).

From the Snozka Pass we are descending into Krośnica village across Magura Nappe and going uphill into Pieniny Mountains, which belong to the geological structure known as PKB. The Pieniny Mountains belong to the Polish Pieniny National Park (Pieniński Park Narodowy) and its Slovak equivalent Pieninský Narodný Park. The idea of the National Park was given by Władysław Szafer in 1921 after Poland gain her independence. The Park was established in 1932 in Poland and in 1967 in Slovakia (Kordováner *et al*., 2001b; Tłuczek, 2004). The Pieniny National Park area is 4,356 ha, 2,231 ha on the Polish (Kordováner *et al*., 2001a, 2001b; Tłuczek, 2004). One quarter of this area belongs to special nature sanctuaries, the most important ones are: Macelowa Góra, Trzy Korony, Pieniński Potok valley, Pieninki and Bystrzyk (Kordováner *et al*., 2001b; Tłuczek, 2004). 60% of the park area are forests mainly beech woods, the rest are meadows, agricultural areas and rocks. The Pieniny National Park fulfills its nature preservation role, conducting also scientific research, education and touristic activities (Kordováner *et al*., 2001b; Tłuczek, 2004; see also Museum of Pieniny National Park at Krościenko n/Dunajcem). From the Krościenko we are going to thw Szczawnica spa and farther east to the Jaworki village.

Stop 13 – Obłazowa Klippe – microfacies of the Czorsztyn Limestone Formation (Bathonian-Tithonian, Czorsztyn Succession) (Fig. 45)

(*Michał Krobicki, Magdalena Sidorczuk,
Andrzej Wierzbowski*)

The south-eastern part of the Obłazowa Klippe shows a fairly complete sequence of the Jurassic deposits of the Czorsztyn Succession (Birkenmajer, 1963, 1977; Wierzbowski *et al.*, 1999). The best section is exposed at a rock shelter in southernmost part of the klippe, and it shows the contact of the Czorsztyn Limestone Formation with underlying crinoid limestones.

The oldest are grey crinoid grainstones of the Smolegowa Limestone Formation attaining at least 25 m in thickness (Birkenmajer, 1963). The overlying pink to rusty coloured crinoid limestones with some admixture of heamatite-marly matrix, form a single bed about 0.10 – 0.15 m thick, which belongs already to the Krupianka Limestone Formation. The upper boundary of the crinoid limestones represents an omission surface coated with ferro-manganese crusts. Overlying this surface are nodular limestones of the Czorsztyn Limestone Formation. The ammonites collected from the lower part of bed 2 include *Procerites* (*Procerites*) *progracilis* Cox & Arkell, and *Procerites* (*Siemiradzkia*) sp., indicative of the Progracilis Zone – the lowest zone of the Middle Bathonian (Wierzbowski *et al.*, 1999). The nodular limestones are developed in two microfacies types: the filament microfacies occurring in lower and upper parts of the studied deposits of the Czorsztyn Limestone Formation, and the filament-juvenile gastropod microfacies found in the middle part of the deposits. Moreover, the filament – *Globuligerina* microfacies is recognized in the topmost part of the deposits studied – it still shows the presence of the filaments together with fairly common planktonic foraminifers of the genus *Globuligerina* (Wierzbowski *et al.*, 1999; Jaworska, 2000). The younger deposits represented by nodular limestones show the presence of the *Saccocoma* microfacies (Jaworska, 2000). The occurrence of *Saccocoma* microfacies in the Czorsztyn Succession is typical of the Kimmeridgian and Lower Tithonian (Myczyński & Wierzbowski, 1994; Wierzbowski, 1994).

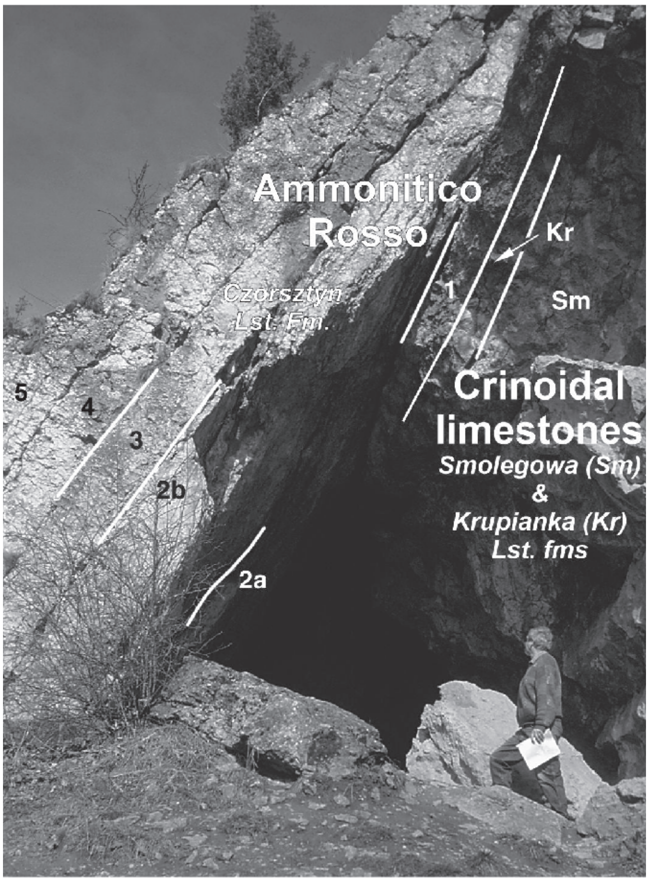

Fig. 45. Obłazowa Klippe: section studied; Sm – white crinoidal limestones of the Smolegowa Limestone Formation; Kr – red crinoidal limestones of the Krupianka Limestone Formation

Stop 14 – Halečková Klippe (quarry) – Bajocian to Oxfordian calcareous and radiolarite sequence of the Pieniny Succession (Figs 46, 47)

(*Roman Aubrecht, Michał Krobicki, Alfred Uchman*)

Halečková Klippe is situated SW of the small town of Trstená, near the main road connecting Trstená and Tvrdošín. In the active quarry, a part of the Pieniny Succession (deep-water sequence of the Pieniny Klippen Basin) is visible in tectonically overturned position. The uppermost part of the quarry shows by Fleckenkalk/Fleckenmergel-type facies of the Podzamcze Limestone Formation; the middle part displays by grey-brown, manganiferous radiolarites of the Sokolica Radiolarite Formation (uppermost Bajocian – Upper Callovian), whereas the lower parts of the quarry Czajakowa Radiolarite Formation (grey-greenish radiolarites of the Podmajerz Member, uppermost Oxfordian – Lower Tithonian). Radiolarian assemblages of the Sokolica and Czajakowa Radiolarite Formations have been studied by Ožvoldová (1992). The taxa are indicative of the Lower to Middle Callovian. The radiolarian assemblage from the lowermost part of the Czajakowa Radiolarite Formation contains radiolarian assemblage points to Upper Callovian-Lower Oxfordian (Unitary Association 5-6). The stratigraphically highest radiolarite beds of the Czajakowa Radiolarite Formation contain radiolarian fauna points to stratigrahic interval from upper Lower Oxfordian to Upper Oxfordian (Unitary Association 7-8).

In the Podzamcze Limestone Formation trace fossils are very abundant. They include *Chondrites*, *Planolites*, *Thalassinoides* and rare *Teichichnus*. They occur against totally bioturbated background, displaying distinct cross-cutting relationships. *Thalassinoides* and *Planolites* are cross cut by *Chondrites*. Totally bioturbated background points to total reworking of sediment near the sea floor and well oxygenated bottom conditions.

Normally graded bedding is common in some radiolarite beds of the Sokolica Radiolarite Formation. Trace fossils are abundant and well visible. *Chondrites* and *Planolites* are most frequent. *Thalassinoides* is less frequent. In the graded beds, density of bioturbational structures increases towards the top, where they occur against totally bioturbated background. The graded bedding and style of ichnofabric indicate that at least a part of beds was deposited by diluted density currents. Trace fossils *Planolites*, *Chondrites*, and rarely *Thalassinoides* occur in some beds and indicate well-oxygenated sediments.

In some beds, layers with normally graded bedding are present. The grain size decreases gradually towards the top and these layers were deposited probably by diluted density currents. This idea was suggested in case of radiolarites of the Niedzica Succession by Kwiatkowski (1981).

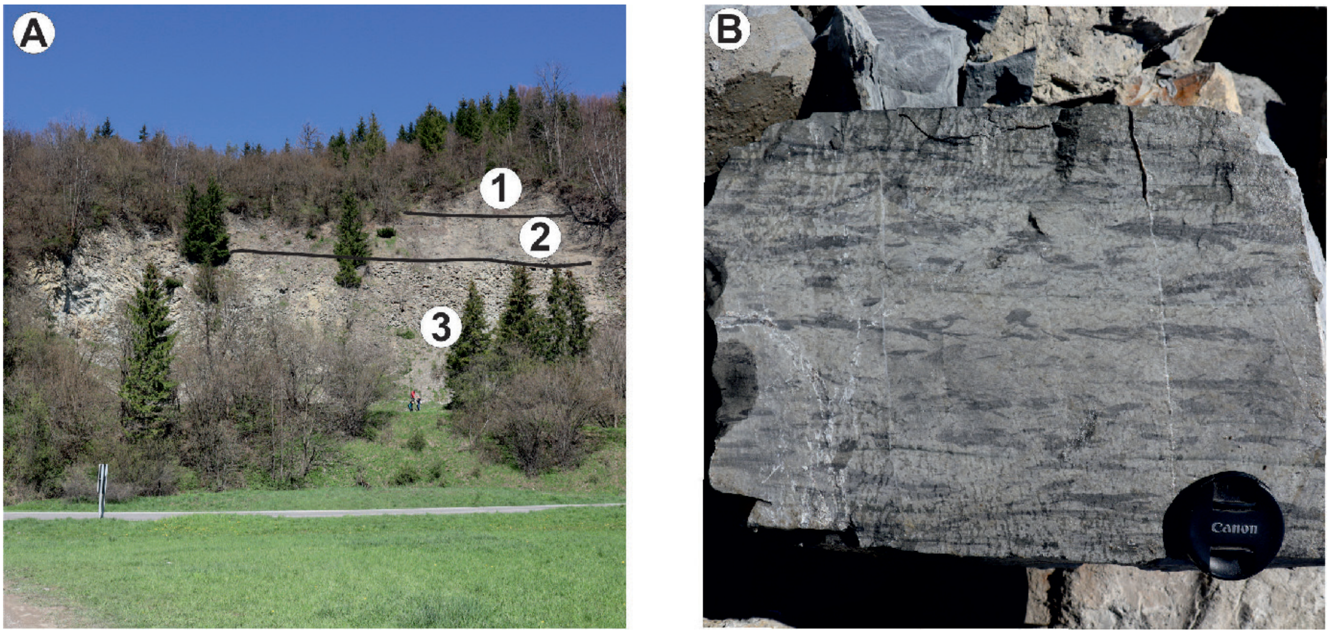

Fig. 46. View on the Halečková Klippe, with the individual lithostratigraphic units (A): 1 – Podzamcze Limestone Fm. (grey spotty limestones with trace fossils – B); 2 – Sokolica Radiolarite Fm. (dark/black manganiferous radiolarites); 3 – Podmajerz Radiolarite Mbr of the Czajakowa Radiolarite Fm. (green radiolarites)

The density of bioturbational structures and their size decreases from the Podzamcze Limestone Formation to the Sokolica Radiolarite Formation, and than to the Czajakowa Radiolarite Formation. This can be interpreted as an increase of environmental stress related to decreasing amount of food, which in turn is related to deepening of sea of the Pieniny Klippen Basin from Early to Late Jurassic, according to the general trend in this part of the Tethys.

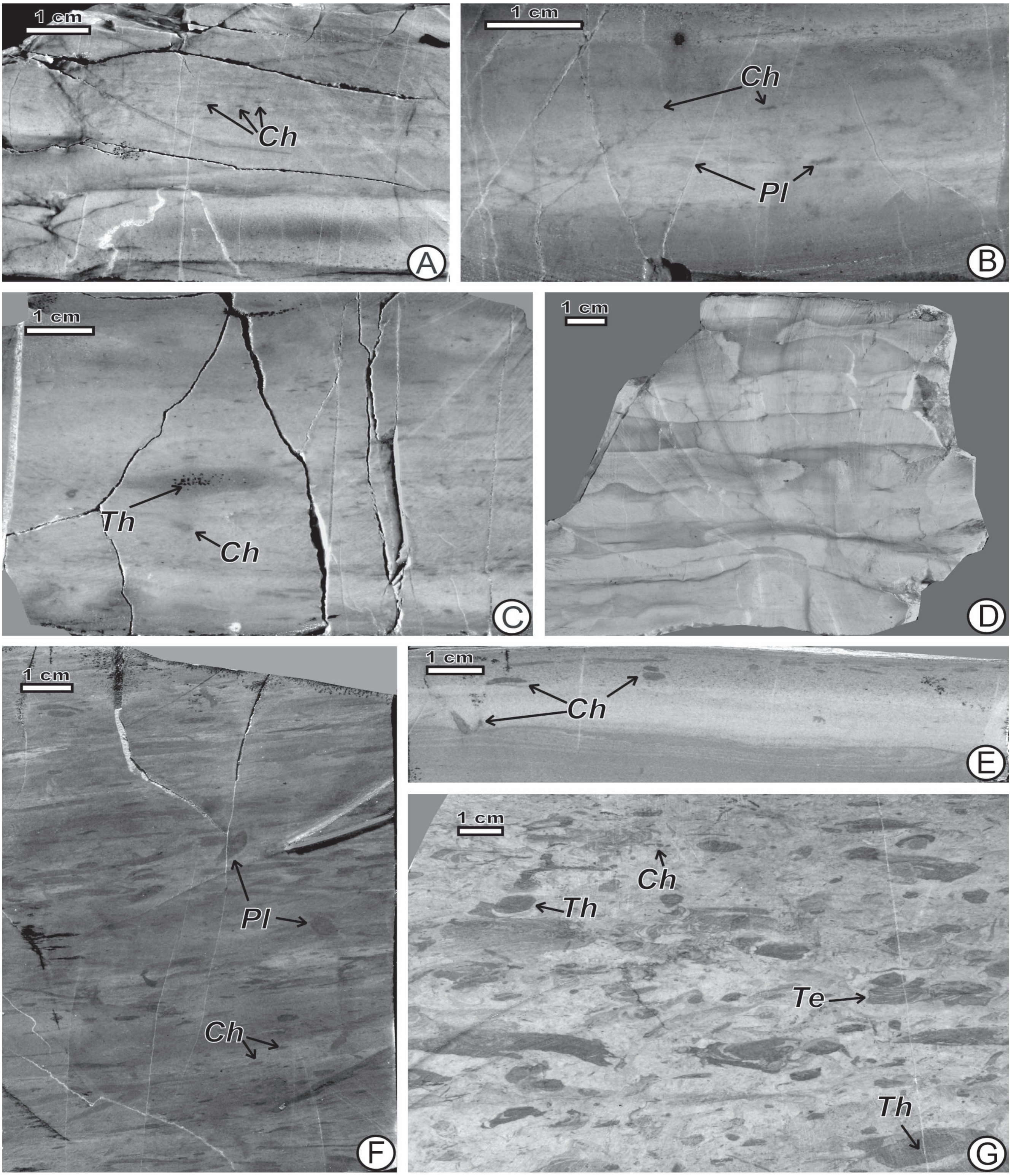

Fig. 47. Trace fossils within Podzamcze Limestone Fm. (G); Sokolica Radiolarite Fm. (F) and Czajakowa Radiolarite Fm. (A–E): *Ch – Chondrites*; *Pl – Planolites*; *Te – Teichichnus*; *Th – Thalassinoides* (after Aubrecht *et al*., 2006)

Stop 15 – Baška village – part of the Hradiště Formation with so-called teschenite volcanics (including pillow lavas) (Fig. 48)

(*Petr Skupien, Zdeněk Vašíček*)

Late Jurassic subsidence enhanced deposition in several troughs (from south to north the Magura-, the Silesian- and the Subsilesian basins) separated by ridges. During Early Cretaceous, black shale dysoxic sedimentation with local submarine clastic fans embraced almost all Outer Carpathian basins. Slow and uniform sedimentation of green and black shale took place during the Albian–Cenomanian, followed by sedimentation of red and variegated shale under well-oxygenated conditions in the Upper Cretaceous. Locally more than 6 km thick flysch deposits are typical of the Outer Carpathian sedimentary sequences.

The Silesian unit is a part of the flysch zone of the Outer Western Carpathians representing the complex of allochtonous nappes. Three subunits (facies) are preserved in the present-day structure of the Silesian Unit (Picha *et al.*, 2006), i.e. the Godula subunit (basinal setting), the Baška subunit (frontal slope setting) and |the Kelč subunit (continental slope setting).

Rocky bottom of the Ostravice River near Baška village exposes higher (Upper Barremian) part of the Hradiště Formation. Dark-grey marlstones and siltstones are penetrated and metamorphosed by intruding rocks of so-called teschenite association. A teschenite pyroxenite exposure more than 100 m long occurs in the river bed and both the banks of the Ostravice River. The exposure contains almost 2 m thick layers of dark grey calcareous claystones of the Těšín-Hradiště Formation, locally metamorphosed along the contact with teschenite. Fragments of ammonites and small gastropods occur in one of the claystone layers. *Partschiceras infundibulum* (d´Orb.) and *Costidiscus rakusi* Uhlig are the best-preserved ammonites. The latter species indicates deposits at the Early/Late Barremian boundary (Vašíček *et al.*, 2004). In the beds immediately underlying the igneous rocks, there is exposed a thrust plane separating the Silesian Nappe from structurally lower Subsilesian Nappe. Mandelstones with cavity diameters to 15 cm as maximum, filled by calcite, analcime and harmotome, occur locally near the contacts with the sediments.

Formation is teschenite association represents dikes, veins, lavas, pillow lavas, and pyroclastic rocks of the teschenite rift-related submarine alkalic, calc-alkalic, and basic volcanism. Šmíd & Menčík (in Menčík *et al*., 1983) distinguished three groups of volcanic rocks: picrites, teschenites, and monchiquites. Hovorka & Spišiak (1988) associated the teschenite volcanism with a short-term rifting of the continental crust. Dostal & Owen (1998) pointed to similarities of these rocks with basalts, basanites, and nephelinites derived from the upper mantle. The volcanic activity peaked during the deposition of the lower part of the Hradiště Formation in the Early Berriasian to Early Barremian time, although teschenite volcanic rocks are sporadically found also in the underlying Těšín Limestone and the Vendryně Formation.

Fig. 48. Teschenitic pillow lavas of on the left bank of the Ostravice River near Baška village (after Skupien & Vašíček, 2008)

Stop 16 – Štramberk, Kotouč quarry, Silesian Unit, Baška development, Hradiště Formation and Baška Formation (Upper Jurassic to Upper Cretaceous)

(*Petr Skupien, Zdeněk Vašíček,*
Justyna Kowal-Kasprzyk)

Sedimentation in the Baška Development (subunit), which was defined by Matějka & Roth (1949, 1955), is delimited in extent to a relatively small area of Palkovice Hills between Frýdek and Nový Jičín. Initial sedimentation is connected with the carbonate platform that was situated on the Baška Cordillera (elevation). The Štramberk Limestone, defined by Hohenegger (1849) was deposited there. Nevertheless, he did not state any type locality.

The limestones are white-grey, diversified in components, and rich in fossils, which were studied since XIX century. They contain the most diversified latest Jurassic–lowest Cretaceous coral assemblage in the world (Eliášová, 2008 and references therein). Ammonites were also extensively studied there (e.g., Houša, 1961; Vašíček & Skupien, 2016).

The Štramberk Limestone is mostly known as reefal limestones, but they were deposited in different settings of the carbonate platform, from lagoon with restricted circulation to forereef (Eliáš & Eliášová, 1984, 1986). According to these authors coral reefs occurred in a low-energy inner platform, as well as in deeper parts of the reef front. Hoffmann *et al.* (2021) based on exotics of the Štramberk-type

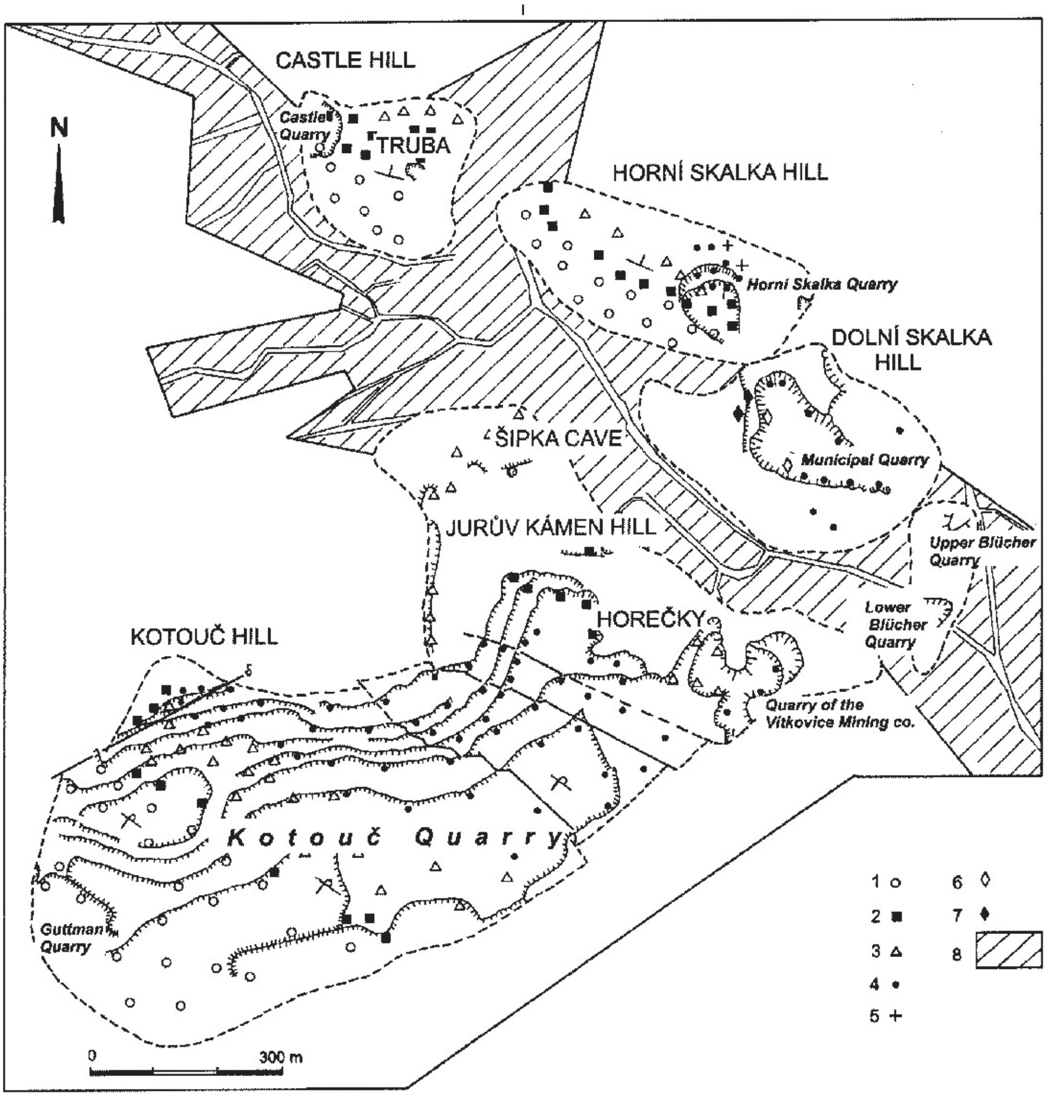

Fig. 49. A map of the main limestone bodies in the vicinity of Štramberk, with the positions of principal quarries and calpionellid zonation (from Houša, 1990, revised Houša & Vašíček, 2005). 1 – limestone without calpionellids; 2 – *Chitinoidella* Zone; 3 – *Crassicollaria* Zone; 4 – *Calpionella alpina* Subzone; 5 – *Remaniella ferasini* Subzone; 6 – *Calpionella elliptica* Subzone; 7 – *Calpionellopsis simplex* Subzone; 8 – town of Štramberk (after Skupien & Vašíček, 2008)

limestones dispersed in the flysch deposits of the Polish Outer Carpathians, paleoecology of corals, and some recently described facies of Štramberk Limestone (Hoffmann *et al.*, 2017) proposed a modified model of the latest Jurassic–earliest Cretaceous platforms, which were developed along the northern part of Western Tethys. According to this model coral-microbial patch-reefs grew in the inner carbonate platform, microencruster-microbial-cement buildups in the upper slope, and microbial as well as and microbial-sponge buildups in the deeper setting, while platform margin is represented by reef-derived detrital limestones and ooid limestones.

Opinions about time span occupied by the Štramberk Limestones differ in detail. Most frequently, the Tithonian is stated. According to Houša (1990) the sedimentation of these deposits began in the earliest Tithonian and ended in the lower Berriasian (calpionellid Remaniella Subzone of Calpionella Zone).

The Štramberk Limestone in a classical form occurs in several quarries (Fig. 49) in the immediate vicinity of the town of Štramberk in the shape of large carbonate blocks, smaller blocks, breccias and conglomerates. Opinions about the setting and position of the limestone in the Silesian Unit have been controversial up to now. Matějka & Roth (1955),

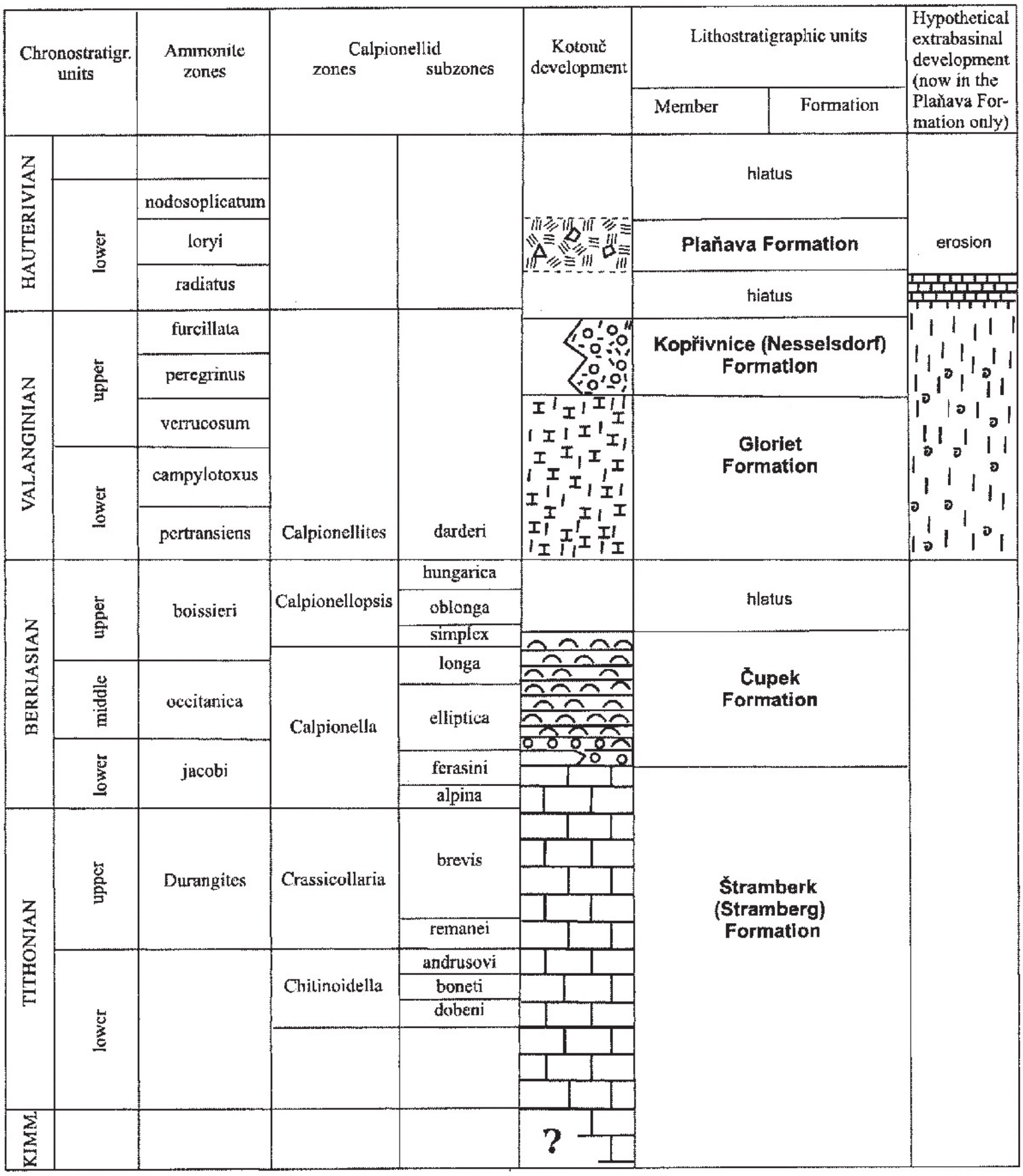

Fig. 50. Stratigraphic units in the region of Štramberk segment of the Baška elevation in the Tithonian and older part of the Early Cretaceous (after Houša & Vašíček, 2005; Skupien & Vašíček, 2008)

Frajová-Eliášová (1962), Houša (1976) and Menčík *et al.* (1983) interpret carbonate blocks as tectonic klippen separated from the carbonate platform in the course of Silesian Nappe overthrust. According to Eliáš & Stráník (1963), Eliáš (1979), Eliáš & Eliášová (1984, 1986), large and also smaller blocks were formed by the disintegration of the platform and redeposition of limestones into younger deposits at the foot of slope of the Baška ridge. None of the mentioned theories, however, explains completely the chaotic character of limestone-bearing deposits in the Štramberk area.

As stated by Picha *et al*. (2006) (in Picha & Golonka 2006), the truth lies somewhere in between both the opinions. The Štramberk carbonate platform rimmed by coral reefs was a source of clastics and debris. Gravitational slides and turbidite currents transported smaller and also larger blocks and fragments from the rim (edge) of platform as far as the foot of the adjacent basin. On the other hand, in the course of later tectonic transport, large tectonic pieces of carbonate platform were separated from softer, less competent rocks situated on the slopes of the platform. The result is a melange in which larger blocks from the carbonate platform have the characters of klippen. Smaller blocks and debris correspond to foot clastic sediments that developed especially in the Early Cretaceous and the early stage of Late Cretaceous.

Houša (1975, 1990) and also Houša in Houša & Vašíček (2004) proved that after the ending of sedimentation of Štramberk Limestone in the Štramberk area, the sedimentation of Early Cretaceous carbonates still continued intermittently (Fig. 50). This is proved by calpionellids and ammonites. Of these Lower Cretaceous carbonates, the Kopřivnice Limestone is the most famous. Suess (1858) described it under the name Kalke von Nesseldorf (German name of Kopřivnice). The type locality is the upper Blücher quarry between Štramberk and Kopřivnice. The Kopřivnice Limestone contains, in addition to abundant brachiopods and echinoderms, Upper Valanginian ammonites (Houša & Vašíček, 2004). Limestones of the Kopřivnice type are brown-red and red micrites, clayey micrites, biomicrites, intrabiomicrites, intramicrites etc.

Picha *et al*. (2006) included all local Cretaceous deposits and local lithostratigraphic units in the area of Štramberk under the name Kotouč Facies. In the original version, the Kotouč development was however defined, namely by Eliáš & Stráník (1963), as dark grey to black-grey pelitic deposits of variable sand content with layers of tilloid conglomerates, and others. At present, the Kotouč Facies represents all the above-mentioned (carbonate, pelitic, conglomerate) specific Cretaceous deposits linked prevailingly to the area of Štramberk.

From paleogeographic viewpoint, the block accumulations form a part of the succession of the continental rise facies of the Baška development below the hypothetical Baška cordillera. They include slumps, slides, fallen blocks (olistholiths), rarely also turbidites (especially proximal), the material of which comes – along with the allodapic limestones of the Baška Formation – from both the carbonate platform (Malm to Coniacian) and the reef complex (Malm/Lower Cretaceous boundary) on the Baška cordillera and its slopes.

Stop 17 – Leszna Górna quarry – carbonate flysch (lowermost Cretaceous, Berriasian) (Figs 51–54)

(*Michał Krobicki, Krzysztof Starzec,
Anna Waśkowska*)

Within the active quarry in Leszna Górna village, formations of the Cieszyn Limestone Formation (“Lower and Upper Cieszyn limestones”) are exposed, formed as medium- and coarse-rhythmic limestone flysch. At several exploitation levels of the quarry, both the lithofacies differentiation of these layers and the very visible phenomena of fold and fault tectonics can be observed. The total thickness of the deposits exposed here reaches about 120 meters. This flysch is mainly represented by calciturbidites and calcifluxoturbidites (cf. Malik, 1994), with a full range of sedimentary features typical of both the inner part of the fan (depositional lobes) and the outer fan of the model of submarine turbidite sedimentation (cf. Mutti & Ricci Lucchi, 1975; Stow, 1986; Mutti & Normark, 1987; Ghibaudo, 1992; Reading & Richards, 1994; Lowe, 1997; Shanmugam, 2000).

The main part of the formations exposed here is Berriasian in age, only the oldest parts of the lowest mining levels in the quarry may be of Upper Titonian. A striking feature of these formations is their facies formation corresponding to fine- and coarse-rhythmic calciturbidites, with stratigraphically higher beds characterized by a series of beds thickening upwards and thus corresponding to the facies ensemble of depositional lobes and their margins. The older, fine-rhythmic sequence would then correspond to the facies complex of the outer cone (Malik, 1994). In both complexes, many features of sedimentary sediments can be observed, indicating a very lively sys-sedimentary tectonics, expressed by a large number of reseded formations: landslides and undersea flows; large, large and small lithoclasts (especially clay-marly) in limestones; coarse biodetrital limestones with fractional graining, etc. Full Bouma sequences can be read in many calcareous beds, terminating in marly portions of intensely bioturbated sequences. A brief description of the most important features of the deposits exposed here was presented by Malik (1994), but it should be noted that due to the progress of exploitation over the last 15 years, at least a dozen to twenty meters of the older part of the formation have been exposed. By the way, this site is still waiting for detailed sedimentological profiling and, on this basis, a broader palaeoenvironmental and palaeogeographical interpretation (Waśkowska-Oliwa *et al*., 2008) is connected with destruction of shallow-sea carbonate platforms and the areas surrounding them and resedimented into the deeper parts of the basin.

Fig. 51. General view of Leszna Górna quarry (A) with intensive folded deposita of the Cieszyn Limestone Formation (B–E) (after Waśkowska-Oliwa *et al*., 2008)

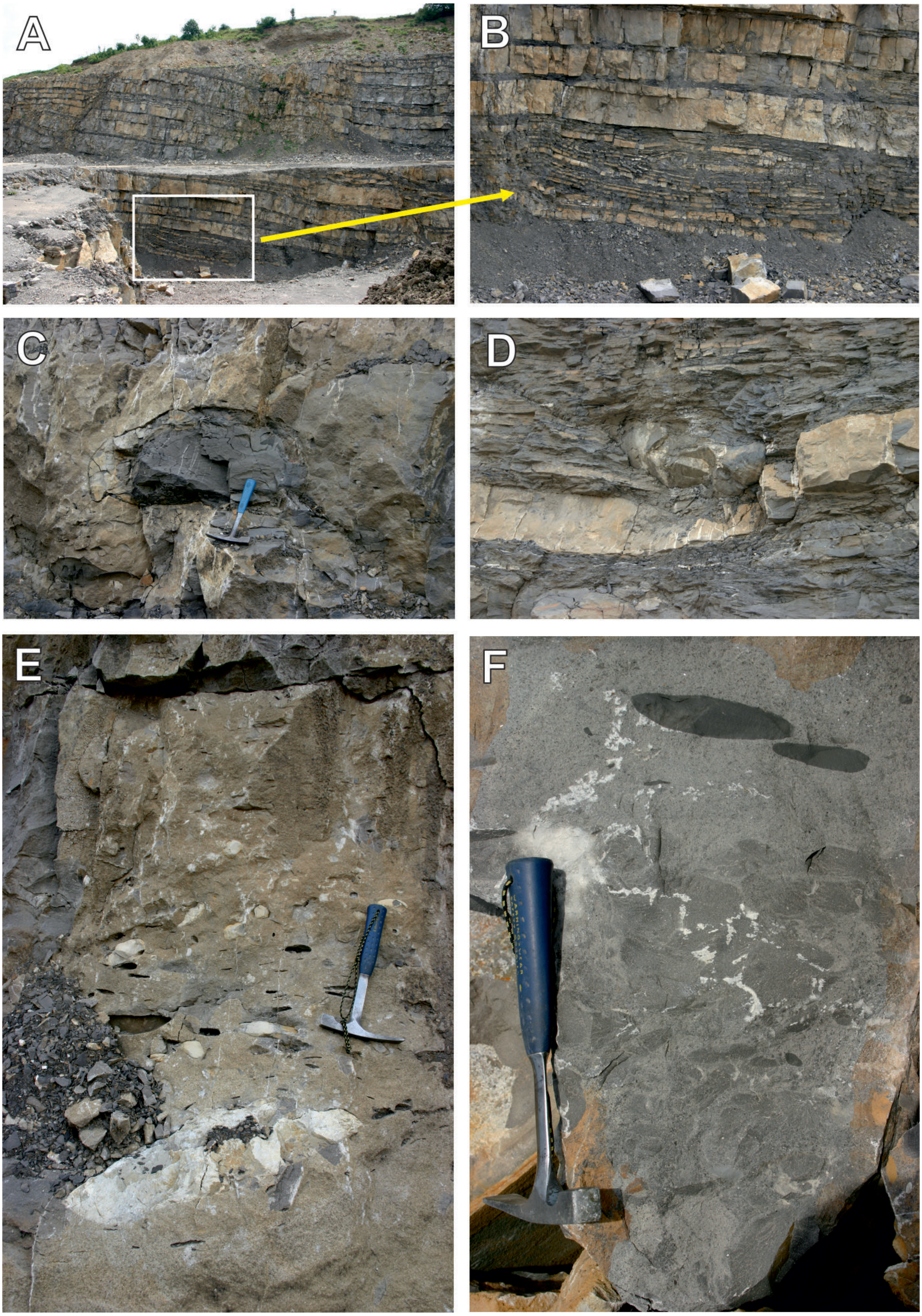

Fig. 52. Synsedimentary features of submarine mass movements in the Cieszyn Limestone Formation in Leszna Górna active quarry represented by: small submarine debris flow (A, B), huge resedimented shales and carbonate clasts (C, D) sometimes with fractionation in fine detritic limestones (E, F) (after Waśkowska-Oliwa *et al.*, 2008)

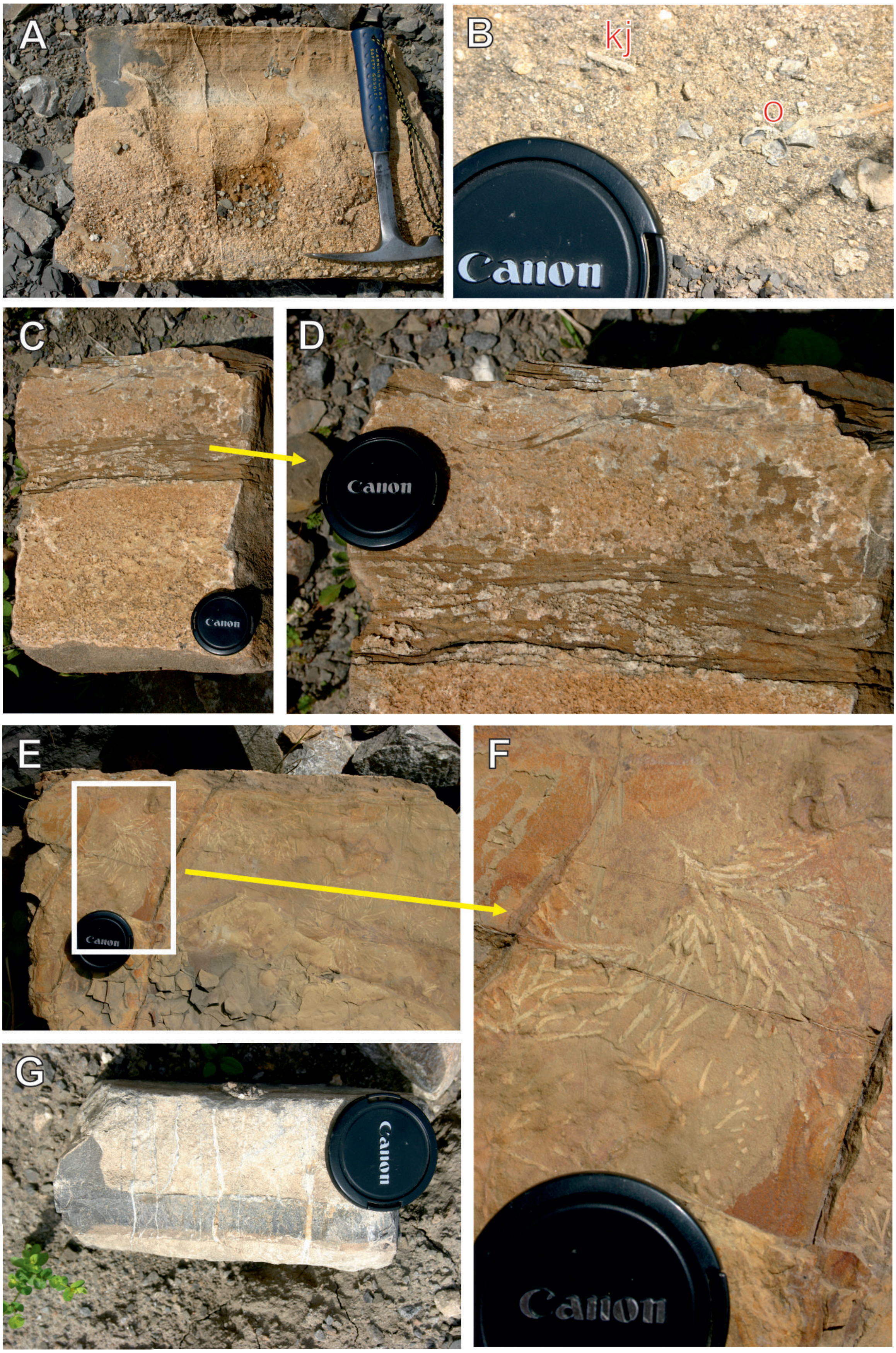

Fig. 53. Organodetrital limestones with gradational fractionation (A, C) sometimes with identifiable fossils at the base beds (B: o – *Nanogyra* sp. oyster; kj – echinoid spine) and convolution in the top (C, D) and trace fossils in more marly parts of bed (ER, F – *Chondrites* isp.), and rarely with cherts (G) (after Waśkowska-Oliwa *et al.*, 2008)

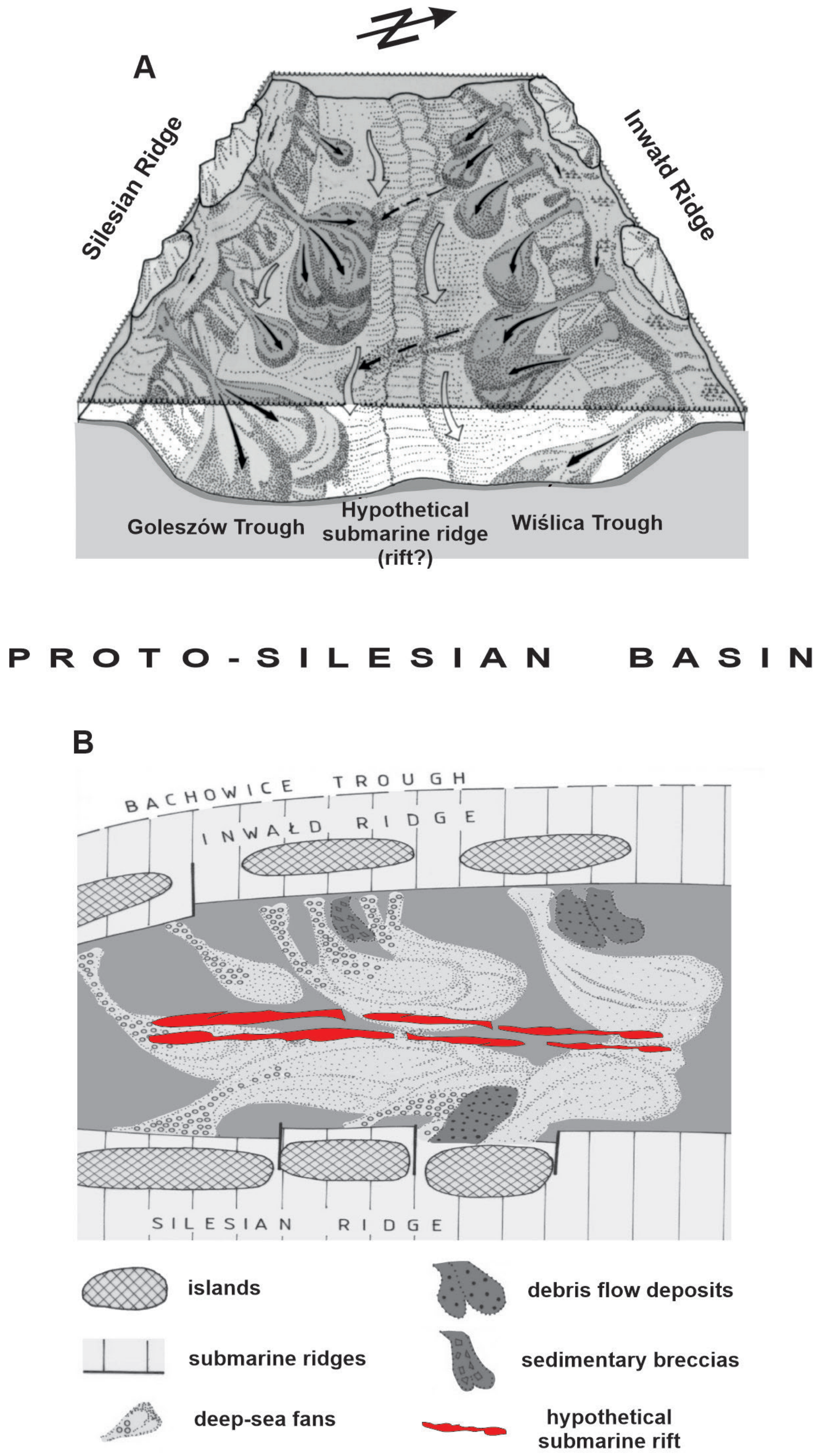

Fig. 54. Palaeogeographical blockdiagram of sedimentation of the oldest flysch deposits in the Proto-Silesian Basin (Jurassic/Cretaceous transition – Tithonian/Berriasian) (A) and its hypothetical palaeogeographical sketch (B) (after Słomka 1986a; slightly modified)

In turn, the consistent uplift of the source areas, which at that time were the Silesian and Baška-Inwałd ridges surrounding the proto-Silesian basin, led to the sedimentation of coarse-movement formations in the form of olistoliths, marl gravels, slides of limestone blocks of the Štramberk type (Nowak, 1973; Słomka, 1986b, 2001). This would not have been possible if not for the intense synsedimentary tectonic movements in this part of the Tethys associated with the Neo-Cimmerian movements of the Alpine orogeny (Krobicki, 1996a, Golonka *et al*., 2003; cf. Middleton & Hampton, 1973; Stampfli *et al*., 1998; Shanmugam, 2002; Dasgupta, 2003; Payros & Pujalte, 2008 with literature cited there).

The Beriasian episode of resedimentation in the Proto-Silesian basin was almost exclusively associated with the destruction of carbonate platforms and uplifted calcareous flysch formations of the Cieszyn Limestone Formation. In the Valanginian, the bedrock erosion occurred (the first exotics of igneous and metamorphic rocks) and the increasing amount of clastic material led to the replacement of calcareous flysch sedimentation by silicoclastic flysch (Słomka, 2001).

The limestone flysch of the Late Jurassic/Early Cretaceous of the Cieszyn Limestone Formation finds good counterparts in many flysch/flysch-like systems in Europe and beyond (Payros & Pujalte, 2008 with literature cited there) and it is striking that, as in the Carpathians (Matyszkiewicz & Słomka, 1994), in many cases, one of the basic grain components of these formations are ooids, as resedimented grains from shallow-sea carbonate platforms, which are the source areas for calcareous turbidite/fluxoturbidite systems (Price, 1977; Ruiz-Ortiz, 1983; Wright & Wilson, 1984; Eberli, 1987; Cooper, 1989, 1990; Zempolich & Erba, 1999; Wright, 2004; Robin *et al*., 2005; Brookfield *et al*., 2006). On the other hand, within the resedimented formations, fragments (as olistoliths/exotics) of organic structures such as coral-algal reefs or microbial-sponge mud mounds can be found (Karpaty – Matyszkiewicz & Słomka, 2004, other locations – Payros & Pujalte, 2008), which document a wide range of shallow-water facies of carbonate platforms in source areas.

Taking into account all the threads mentioned – from the reconstruction of shallow-water carbonate sedimentation environments in the source areas, through the synsedimentary Neo-Cimmerian tectonics catalysing the formation of various resedimented formations, to the isochronous beginnings of submarine volcanic activity (teschenite – Grabowski *et al*., 2003, 2004), they should be associated with the increasing geotectonic activity of the Silesian Cordillera (Poprawa *et al*., 2004, 2005). Originally, it was probably a fragment of the East European platform, detached from it as a result of a short-term, abortive rift, which probably did not lead to full oceanization of the Proto-Silesian basin (Narębski, 1990) (Fig. 54), despite its inclusion in a continuous rift system between the opening Atlantic in the west, through the Ligurian – in the center to the Chornohora-Sinaia branch of the Ukrainian-Romanian Carpathians in the east (Mišík, 1992; Golonka & Krobicki, 2004; Krobicki *et al*., 2005; Rogoziński & Krobicki, 2006 with the literature cited there).

Stop 18 – Krakow vicinity – Middle-Upper Jurassic and Upper Cretaceous deposits

In Krakow and its vicinity the Mesozoic rocks of the North European platform are exposed. The platform is dissected by numerous faults into several horsts and grabens. The grabens are filled with the Miocene Molasse deposits, while horsts elevate Jurassic (mainly Oxfordian) limestones which sometime are covered by the Upper Cretaceous deposits of the Peri-Tethys realm of the northern margin of the Tethys Ocean.

Outcrops? Rocks? Palaeoenvironments? – a surprise for you… – thanks for coming, interesting conversations and company. See you next year in Krakow again on the 36th HKT (Himalaya–Karakorum–Tibet) Workshop at AGH University of Krakow – Welcome…!

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