**Dear HKT/Tethyan Friends,**

The **36th HKT Workshop** (***Himalaya-Karakorum-Tibet***) is organised by the AGH University of Krakow, Poland. Kraków is situated at the front of the Carpathian orogenic system of the ancient Tethys Ocean and is perfect place for meeting between European-Asian Alpine-Himalayan orogenic belt with long history of the Western and Eastern Tethyan Ocean Realms.

The 36th HKT Workshop concerns mainly on this first area and the meeting will be focused on Himalaya-Karakorum-Tibet regions but will be also concentrated (especially during field trip) on comparison between Asian (eastern) and European (western) parts of orogen and geotectonic events based on research of different branches of geology: stratigraphy, paleontology, sedimentology, paleogeography, structural geology, petrology, geochemistry, paleomagnetism, and geophysics. The 36th HKT Workshop is aimed to provide a forum for colleagues from America and Asian/European countries, to present their work and discuss their ideas covering various aspects of the development of the Eurasian Alpine history.

We are glad that we can meet personally during the 36th HKT Workshop after the pandemic time, which disrupted our idea of regular, annual meetings. Simultaneously we are under **IGCP 710 Project** logo (***Western Tethys meets Eastern Tethys – geodynamical, paleoceanographical and paleobiogeographical events***), and we have now the chance to discuss face to face and go to the field together to touch Carpathian/"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 microcontinents 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. 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 the field trip guidebook ["Geotourism", 2024, 76–77(1–2)]. We hope it will be useful for both Alpine-Himalayan/Tethyan friends and geotourism enthusiasts.

**Enjoy Kraków during the stationary part of the 36th HKT Workshop and the Polish-Slovak-Czech Carpathians during a 5-day field trip!**

*Michał Krobicki*

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**Cover photo:** Dhaulagiri (8,167 m) and Tukuche (6,920 m) peaks in the Himalayas, view from Muktinath village, Nepal (photo by M. Krobicki)
**Back cover photo:** Route of the field trip during 36th HKT Workshop

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# Field trip – Outer Flysch Carpathians, Pieniny Klippen Belt (PKB), and Tatra Mountains

## 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, 1977b; 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 and 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, 1977b; 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.


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

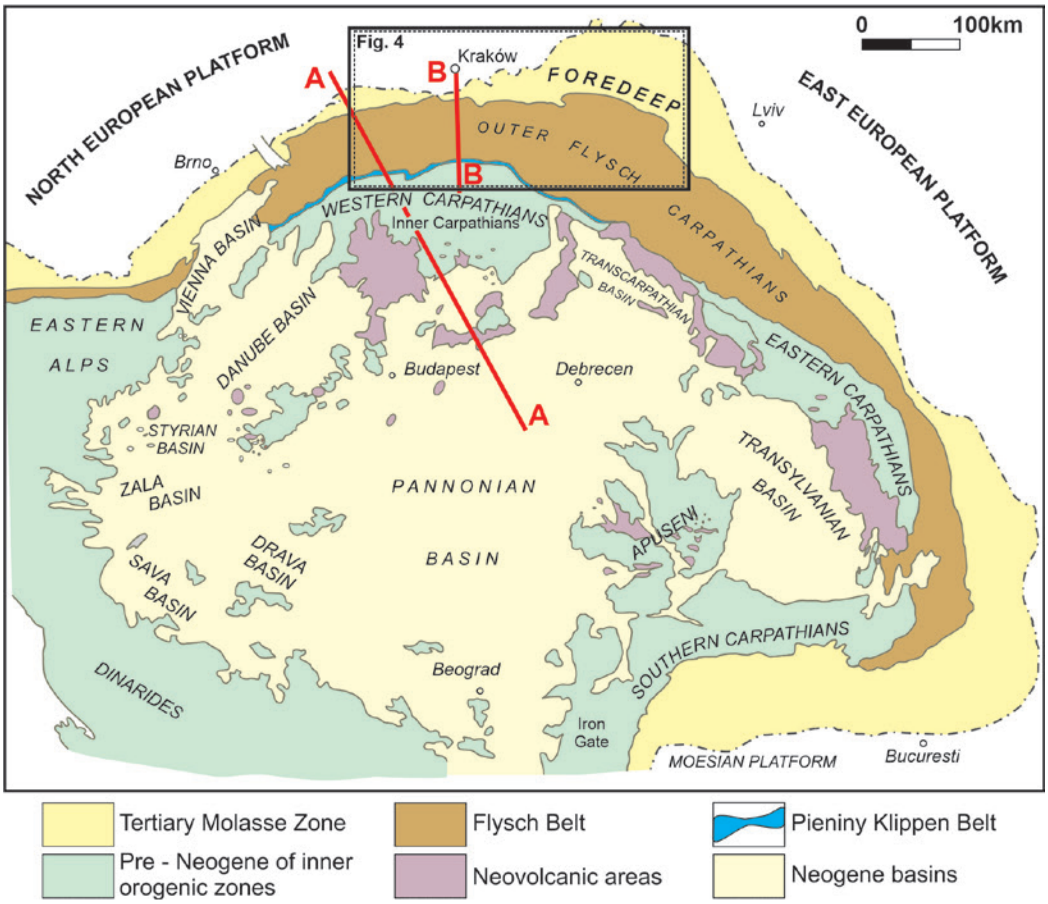


Fig. 2. Tectonic sketch map of the Alpine-Carpathian-Pannonian-Dinaride basin system (modified after Plašienka *et al*., 2000). A-A and B-B – localization of cross-sections (see Fig. 3)

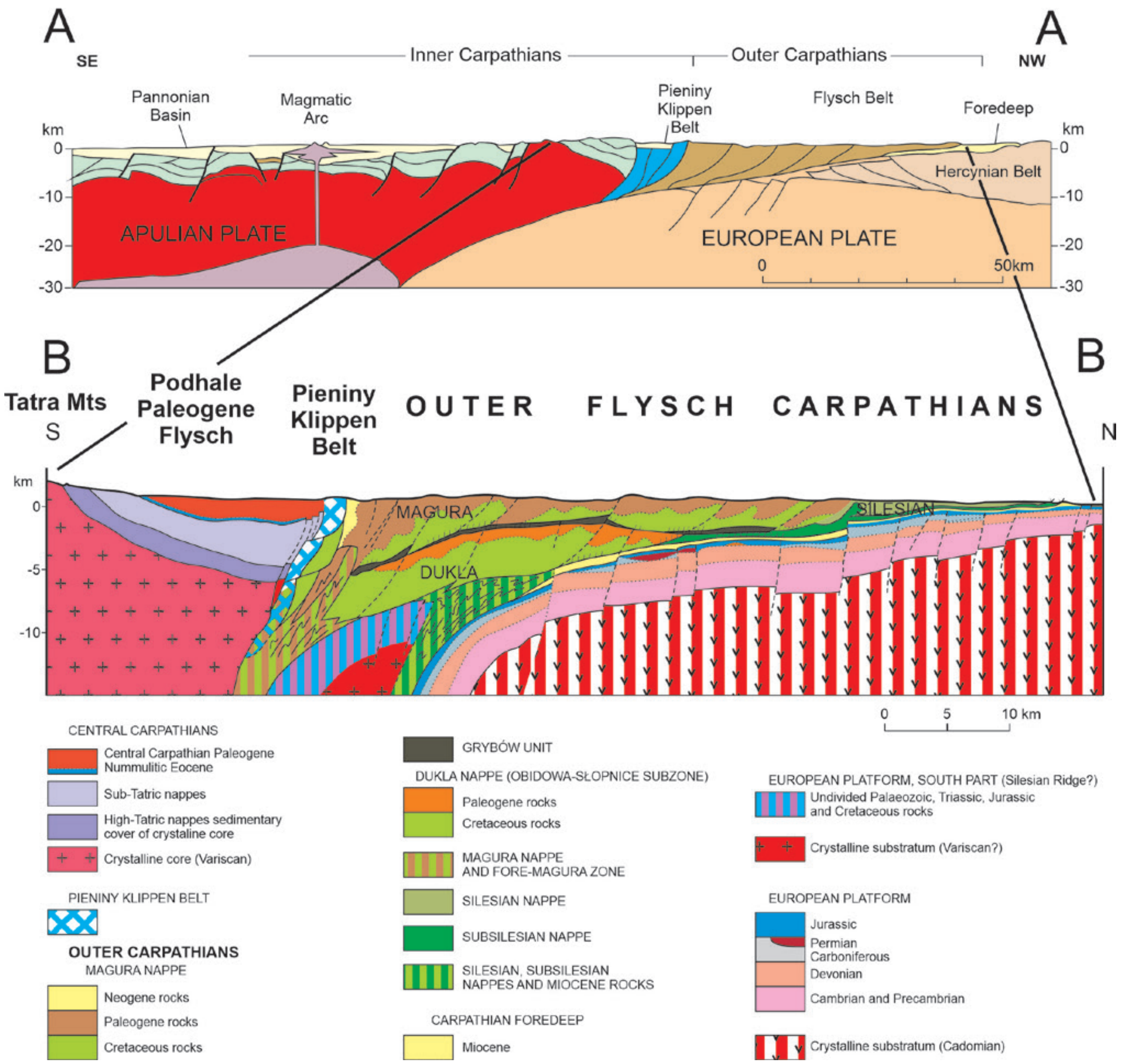


Fig. 3. Generalized cross-section across Carpathian-Pannonian region (Picha, 1996) (upper) and generalized cross-section across Polish Carpathians (after Golonka *et al*., 2005) (lower)

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 Mesozoic-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.

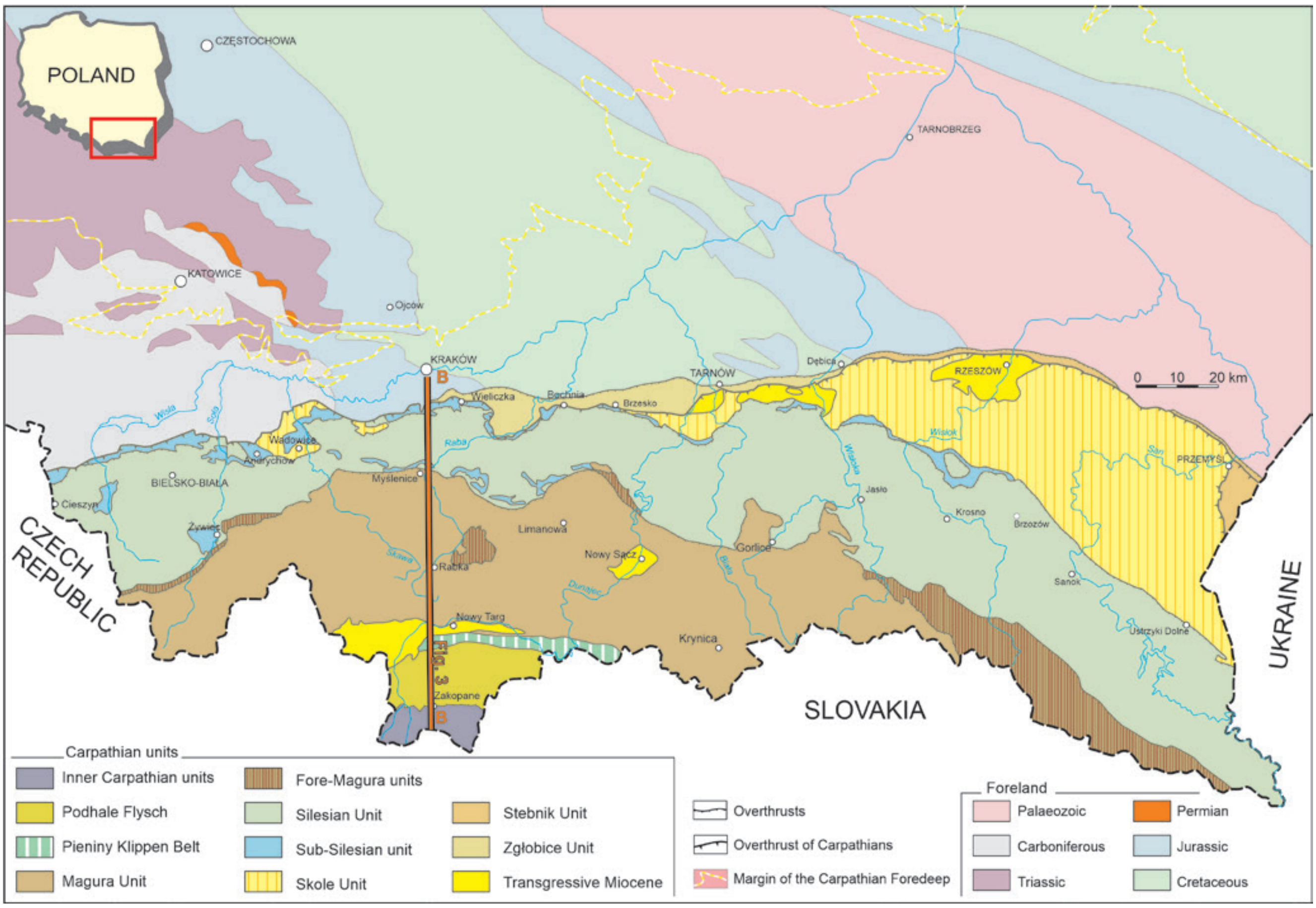


Fig. 4. Geological map of the Polish Carpathians and Foreland (after Żytko *et al*., 1989; simplified)

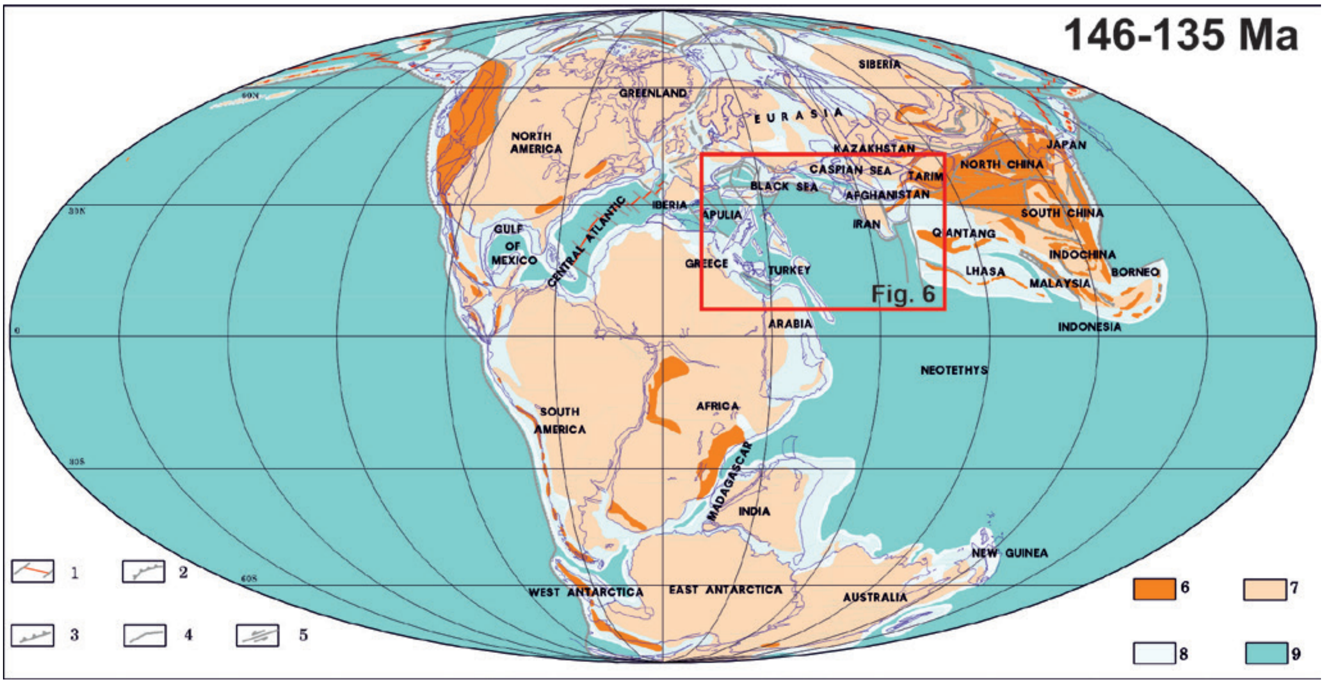


Fig. 5. Global plate tectonic map of latest Jurassic–earliest Cretaceous. Explanations: 1 – oceanic spreading center and transform faults; 2 – subduction zone; 3 – thrust fault; 4 – normal fault; 5 – transform fault; 6 – mountains; 7 – landmass; 8 – shallow sea and slope; 9 – deep ocean basin (from Golonka,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

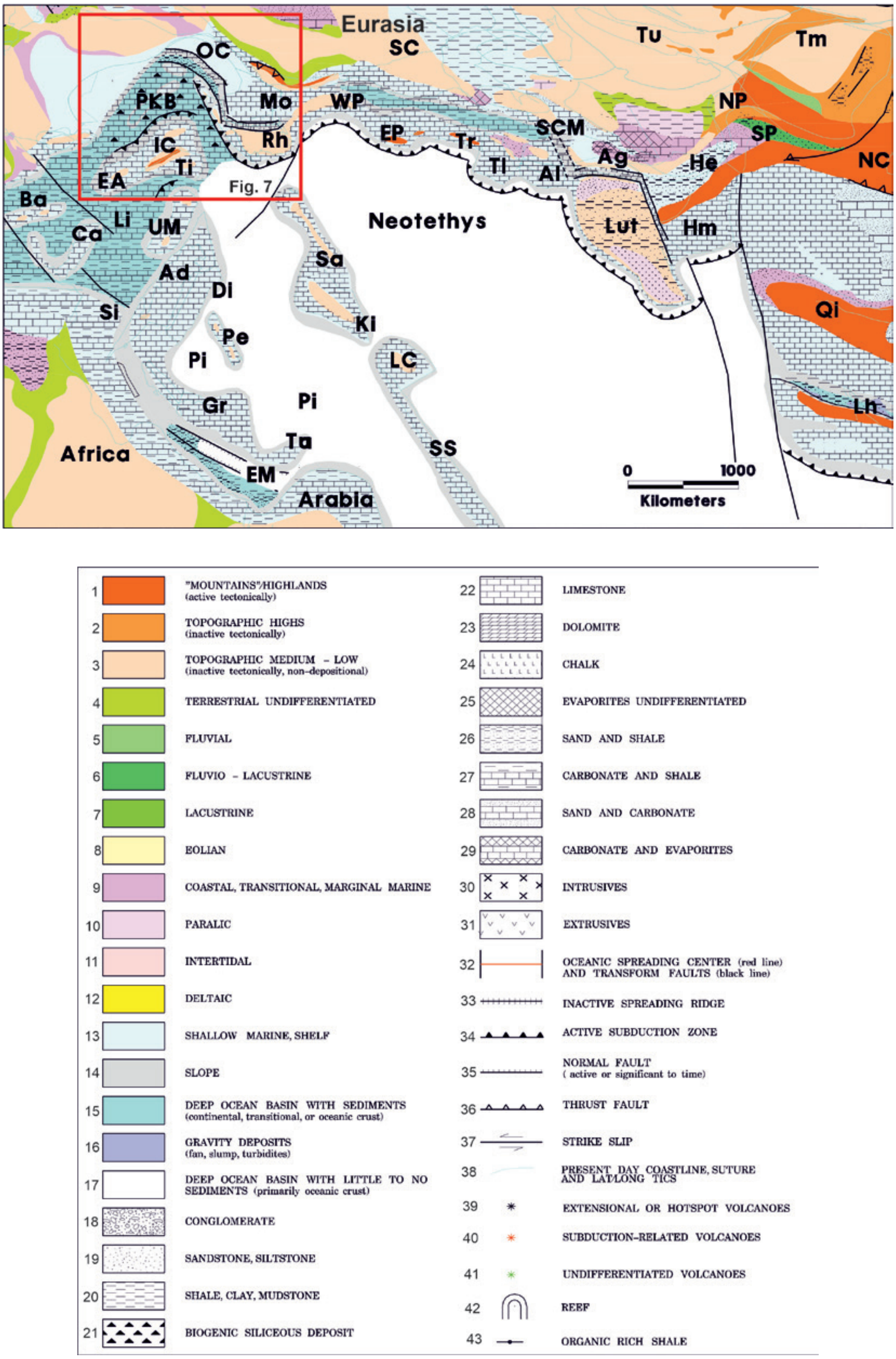


Fig. 6. Plate tectonic, palaeoenvironment and lithofacies map of the western Tethys, Central Atlantic and adjacent areas during latest Jurassic–earliest Cretaceous time (after Golonka, 2007a; modified). Abbreviations of oceans and plates names: Ad – Adria (Apulia); Ag – Aghdarband (southern Kopet Dagh); 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

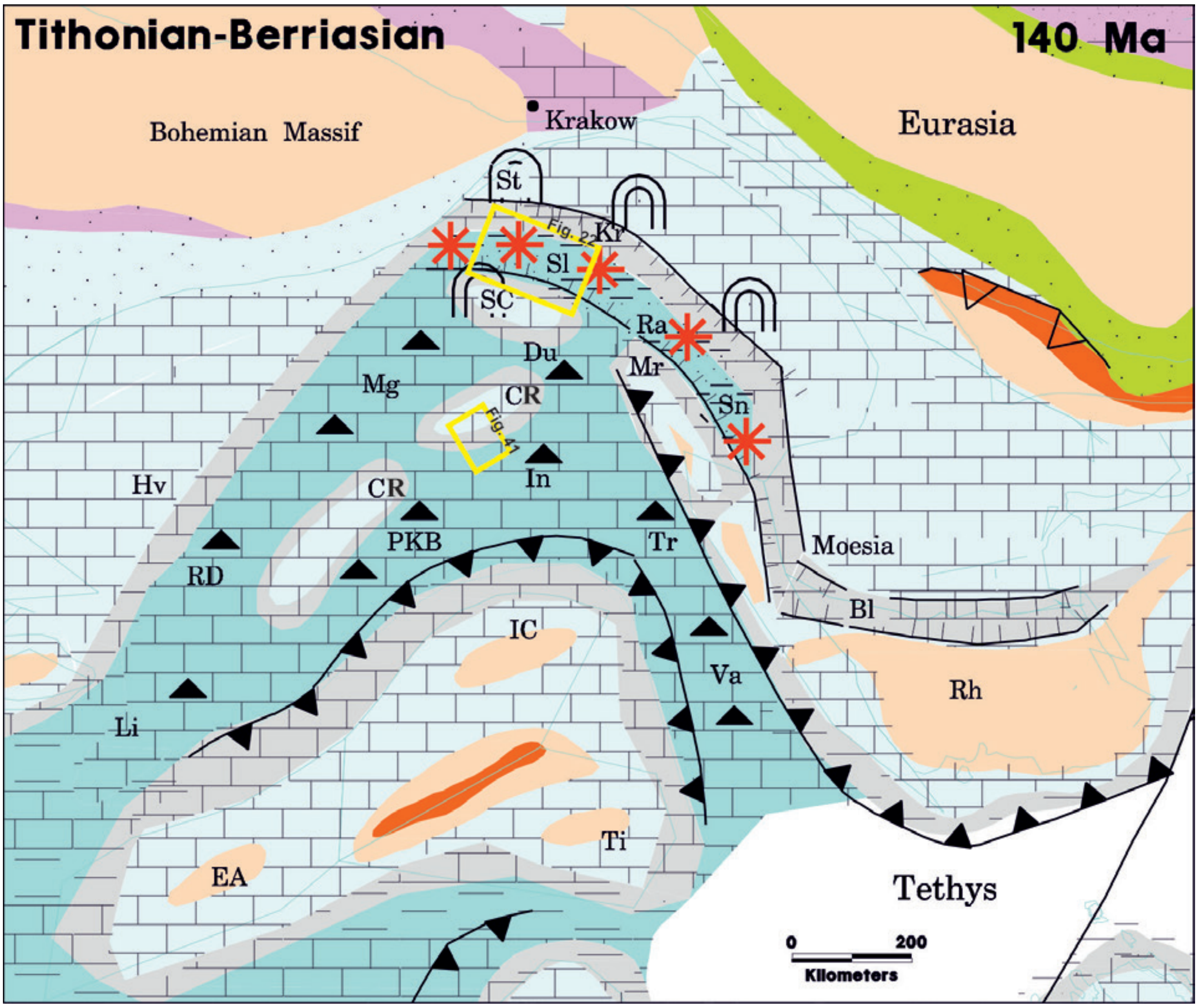


Fig. 7. Palaeoenvironment and lithofacies of the circum-Carpathian area during latest Jurassic–earliest Cretaceous; plates position at 140 Ma (modified from Golonka *et al*., 2006). Abbreviations: Bl – Balkan rift; Cr – Czorsztyn Ridge; Du – Dukla Basin; EA – Eastern Alps; Hv – Helvetic shelf; IC – Inner 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, 1977b; 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).

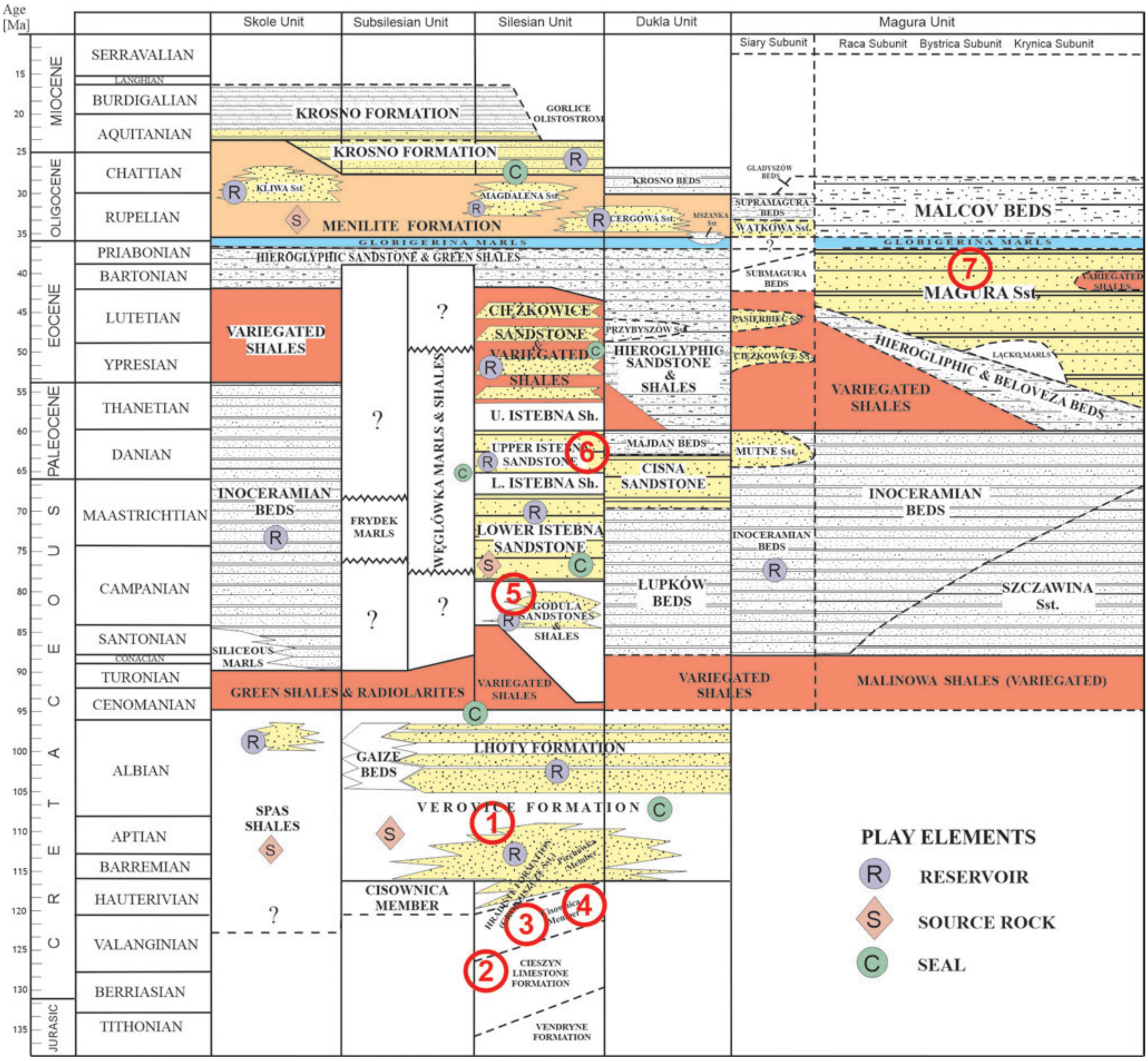


Fig. 8. Simplified lithostratigraphy of the Outer Polish Carpathians (after Koszarski *et al*., 1985; Dziadzio *et al*., 2001; modified) with the position of field trip stops

## 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, 1977b; 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, 1977b; 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).

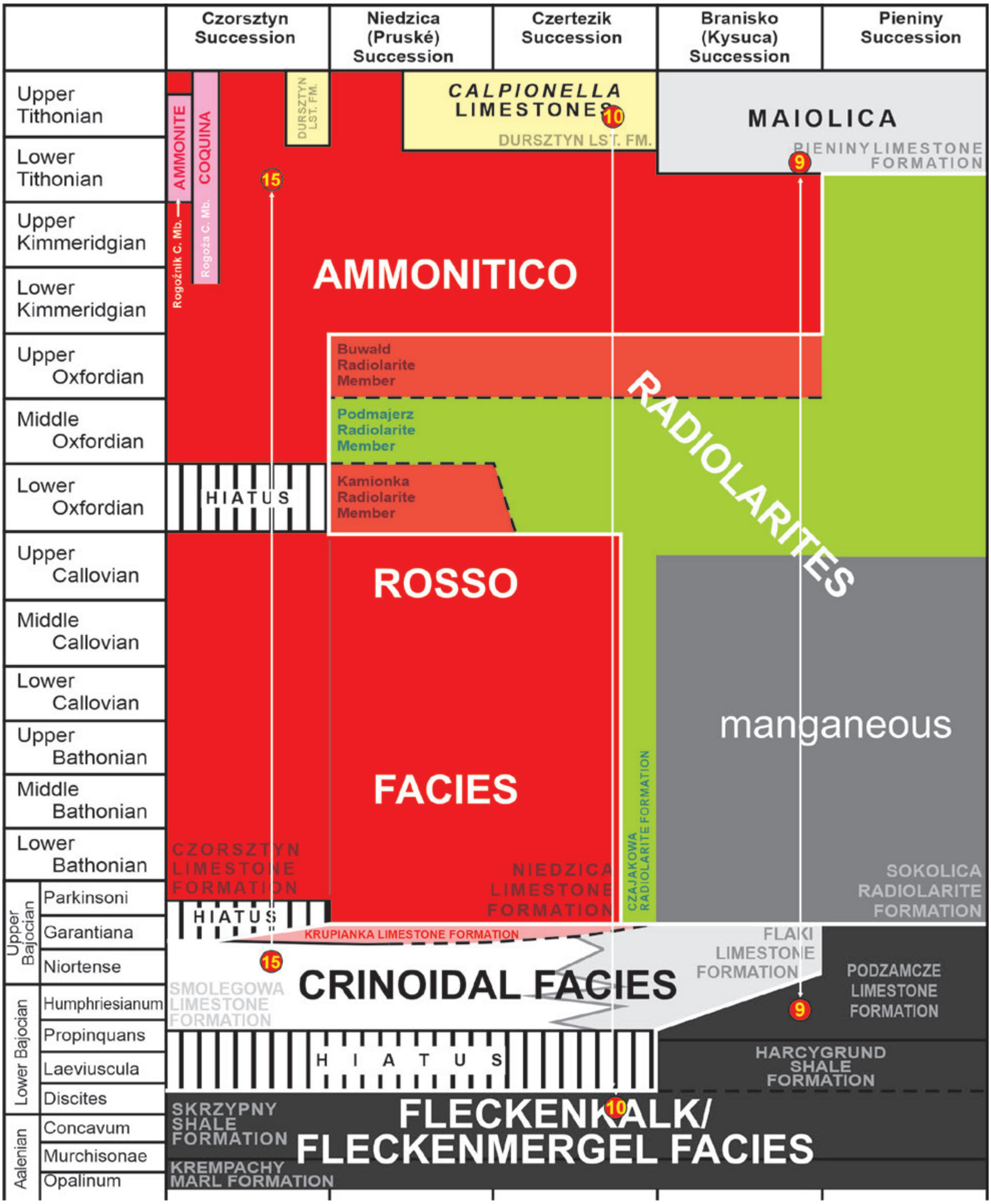


Fig. 9. Stratigraphical correlation between Jurassic lithofacies (lithostratigraphic units) of the Pieniny Klippen Belt successions (after Wierzbowski *et al*., 2004; supplemented by Krobicki & Wierzbowski, 2004)

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).

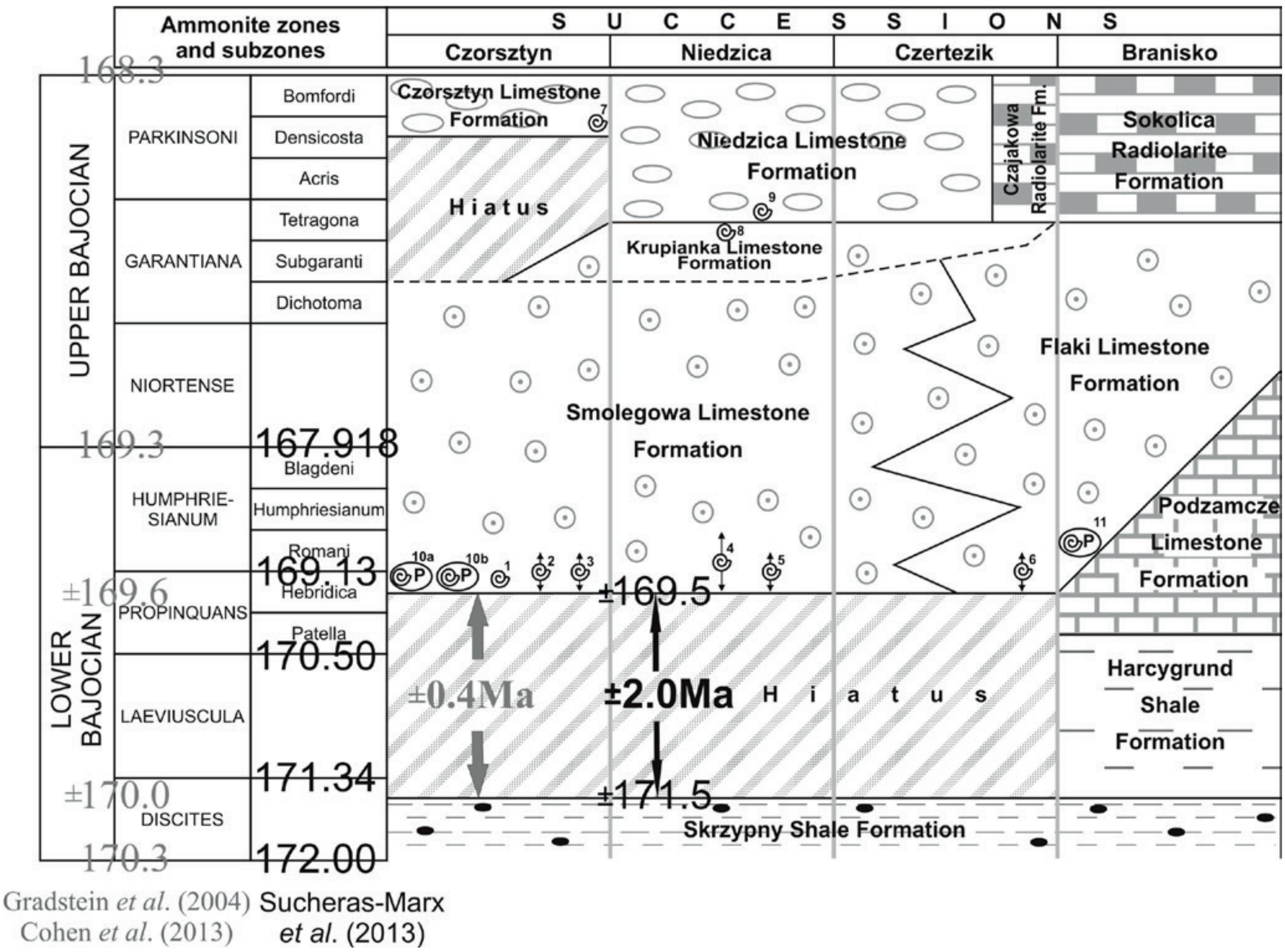


Fig. 10. Lithostratigraphical scheme of the klippen successions of the Pieniny Klippen Belt (after Krobicki & Wierzbowski, 2004, slightly modified; lithostratigraphical units after Birkenmajer, 1977) with data of duration of the Early Bajocian. Numeration indicates outcrops with ammonites (described 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).

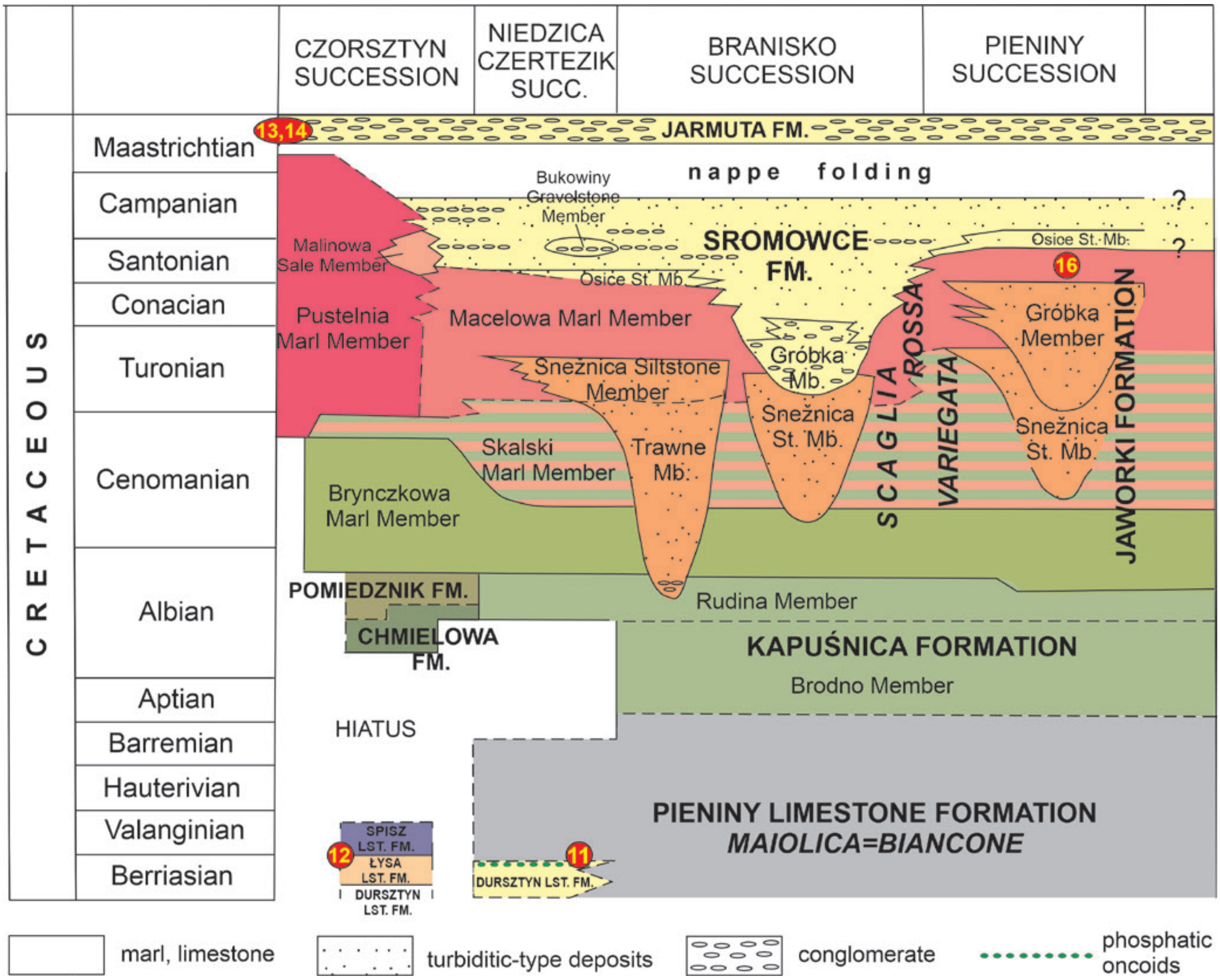


Fig. 11. Detailed stratigraphic table of the Cretaceous rocks of the Czorsztyn, Niedzica (Pruské), Czertezik, Branisko (Kysuca) and Pieniny successions in the Pieniny Klippen Belt in Poland (from Birkenmajer & Jednorowska, 1987, simplified) with locations of field trip stops

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).

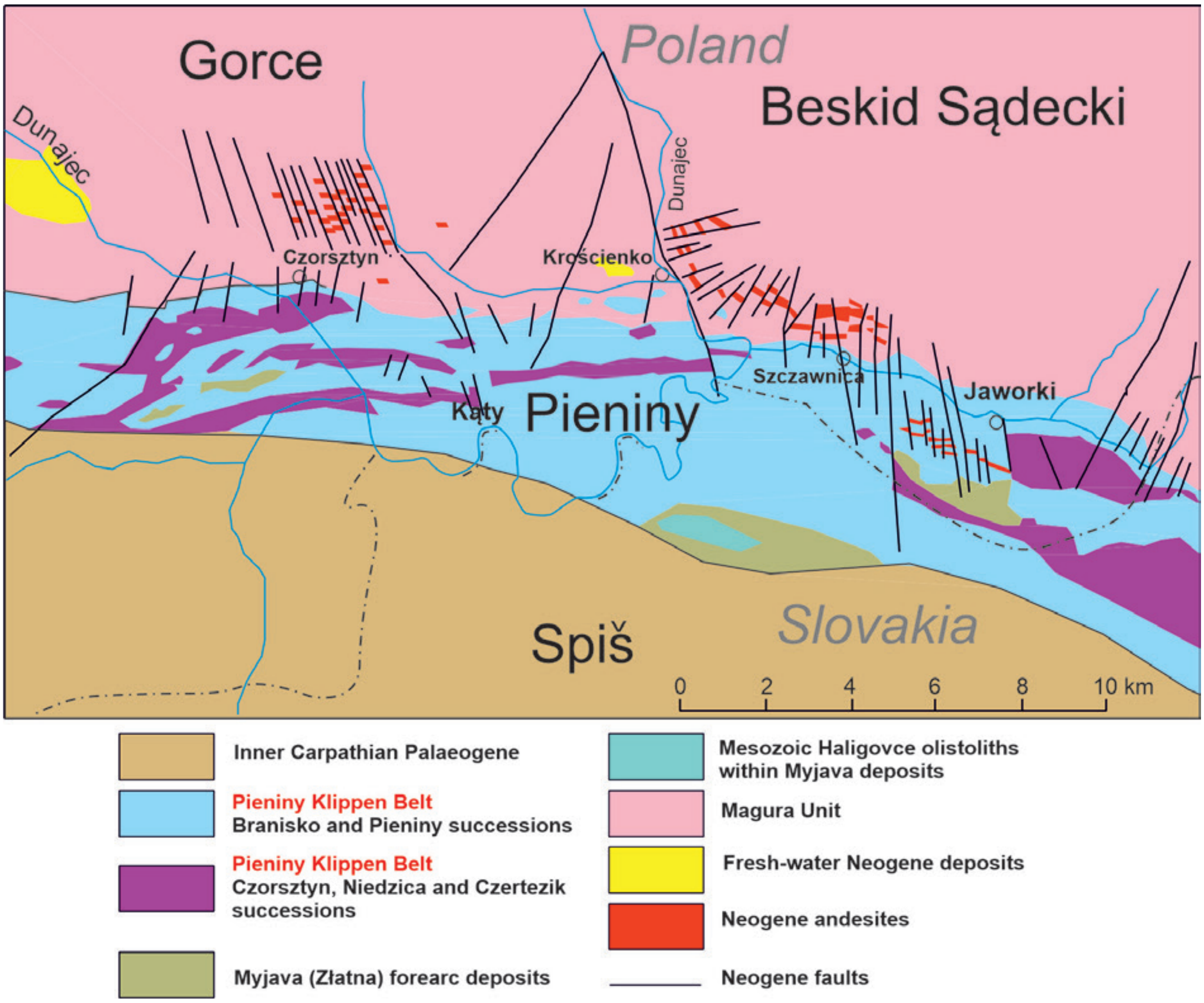


Fig. 12. Geological sketch of the Pieniny Klippen Belt (Polish sector) and surrounding regions (after Birkenmajer, 1979 – simplified)

## DESCRIPTION OF THE TRIP

### *Passage: Kraków – Rzyki – Leszna Górna – Wisła – Szczawnica*

(*Jan Golonka*, *Michał Krobicki*)

The field trip starts in AGH University of Krakow 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 10th century and remodeled several times. The most important remodeling was done by Queen Bona and her team of Italian architects in 16th 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.

***Authors' editorial note***

*The material presented in this description is based upon the publications printed in: "Geotourism" journal (2023, vol. 20, iss. 3–4), and other materials of international geological excursions (e.g., IGCP 589'2017) and "Przegląd Geologiczny" ("Geological Review"), cited partly in extenso with permission of editors.*

## Stop 1 – Rzyki village – Lower Cretaceous dark grey/black heterolithic deposits (Veřovice Formation) (Figs 8, 13, 14–18)

(*Piotr Łapcik*)

The Zagórnik-Rzyki section is situated south of Andrychów in the western part of the Silesian Nappe. The section commences where the uppermost part of the Veřovice Formation (ca. 24 m thick) is cropping out along the incised Wieprzówka stream (starting point GPS coordinates: N 49° 50′ 00.6″; E 19° 22′ 16.9″; ±6 m). The strata exhibit a gentle southerly dip and are characterised by minor tectonic disruptions, although the overall stratigraphic order remains intact. The Zagórnik-Rzyki section is an excellent representation of Lower Cretaceous deposits in deep water setting, which remains a permanent point of geological field trips in the Polish Outer Carpathians (e.g., Cieszkowski *et al.*, 2001, 2003; Uchman & Cieszkowski, 2008; Uchman *et al.*, 2022).

The Veřovice Formation is dominated by dark grey and black mudstones and siltstones. These non-calcareous, partially siliceous lithologies frequently incorporate heterolithic sandstone-mudstone facies, irregularly interbedded with thin to medium beds of fine-grained sandstone. Heterolithic deposits show complex but hierarchical internal structure organised into nine heterolithic divsions A1–A9 with varied tractional depositional structures (Łapcik, 2023). This sediments are interpreted as a spectrum of waxing, waning and pulsating waxing to waning turbulent flow and transient turbulent flow deposits with possible hyperpycnal flow origin (Łapcik, 2023). The cooccurring sandstone beds exhibit planar parallel-lamination and/or ripple cross-lamination as a part of classic Bouma $T_b$ and $T_c$ sequences. Ferruginous concretions appears throughout the section. Exposed Veřovice Formation include alternating mud-dominated successions and intervals with increased proportion of sand and heterolithic deposits, each a several metres thick. Previously these deposits were considered as basin plain facies that were periodically influenced by bottom currents (e.g., Uchman & Cieszkowski, 2008; Waśkowska *et al.*, 2009), however recent detailed sedimentological studies revealed their proximal depositional setting with a relatively dynamic seafloor environment on the flank of mud-dominated depositional lobe (Łapcik, 2023). Non-calcareous nature of the sediments along with analyses of foraminiferal assemblage suggest deposition within the lower bathyal zone (Książkiewicz, 1975; Szydło, 1997).

The uppermost part of the Veřovice Formation in the Zagórnik-Rzyki section was dated with dinocyst *Cerbia tabulata* to the Lower Albian stage (Gedl, 2001, 2003). The dark mudstones represent organic-rich deposits that accumulated plant detritus sourced from neighbouring terrestrial source. High phytoplankton productivity also likely contributed to the organic content, with TOC values reaching approximately 4% (Strzeboński *et al.*, 2009; Pavluš & Skupien, 2014; Wójcik-Tabol & Ślączka, 2015). Deposits of the Veřovice Formation were claimed as associated with the Lower Cretaceous oceanic anoxic events and perhaps related with early Aptian Selli Event (OAE 1a) or the early Albian Paquier Event (OAE 1b). Nevertheless, macroscopically it is challenging to distinguish the episodes of oxygen depletion and episodes of improved oxygenation that separates the OAEs. Ichnological data partially supports the interpretation of anoxic conditions in the Veřovice Formation (Uchman, 2001b, 2004) and most deposits lack trace fossils and exhibit primary lamination. However, bioturbated intervals occur repetitively indicating at least poor delivery of oxygen, perhaps with the subsequent flow events (Uchman *et al.*, 2022). Localised occurrences of bioturbated horizons, less than 1 centimetre thick, are observed at the top of some sedimentary sequences. These horizons display a slightly lighter colour and contain a limited assemblage of ichnotaxa dominated by *Phycosiphon incertum*, *Chondrites intricatus* and *Ch. targionii*, *Planolites* isp., *Palaeophycus* isp and rarely *Thalassinoides* isp. Additionally, bivalve burrows identified as *Protovirgularia pennata* and *P. obliterata* are locally abundant and concentrated within a zone a few centimetres below the *Chondrites* occurrences (e.g., Uchman *et al.*, 2022). Presence of *Protovirgularia* in the lowest tier, distinctly below the *Chondrites* assemblage, suggests these burrows were produced by chemosymbiotic bivalves capable of burrowing in anoxic sediments (Uchman *et al.*, 2022). Similar examples include the solemyacid bivalve *Solemya* (Seilacher, 1990) and specific lucinid and thyasirid bivalves (Powell *et al.*, 1998) is occasionally present as well. The Early Cretaceous anoxic events likely hindered the colonization of the deep-sea floor by the irregular echinoids responsible for generating *Scolicia* trace fossils (Tchoumatchenco & Uchman, 2001). Notably, *Scolicia* is absent within the Veřovice Formation and older Carpathian formations. Consequently, the trace fossil assemblage at Zagórnik-Rzyki is likely influenced by global events associated with the Cretaceous anoxic episodes.

A transitional zone to the Lhoty Beds is marked with the first appearance of thin-bedded, green-grey, bioturbated, non-calcareous mudstone shales. Succession of transitional beds, 5.5 m to 6 m thick, appears approximately 160 m upstream from the designated starting point. The frequency of green-grey mudstones and the overall contribution of sandstone beds progressively increase up the section. Notably, sideritic concretions are present at intervals of roughly 1.5 m both above and below the first green-grey layer.

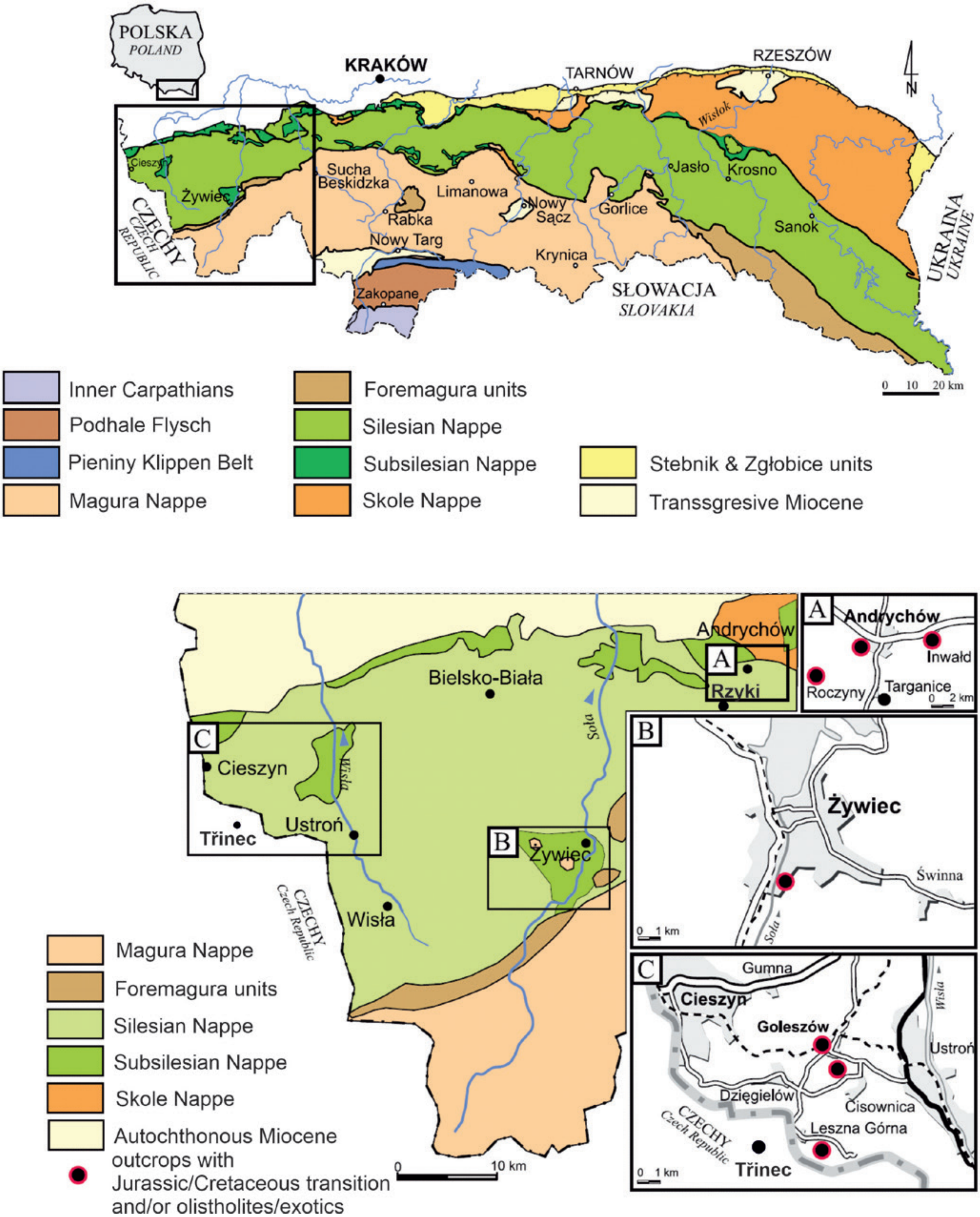


Fig. 13. Geological sketch between Żywiec and Cieszyn with locations of field trip stops (after Żytko *et al*., 1989; simplified)

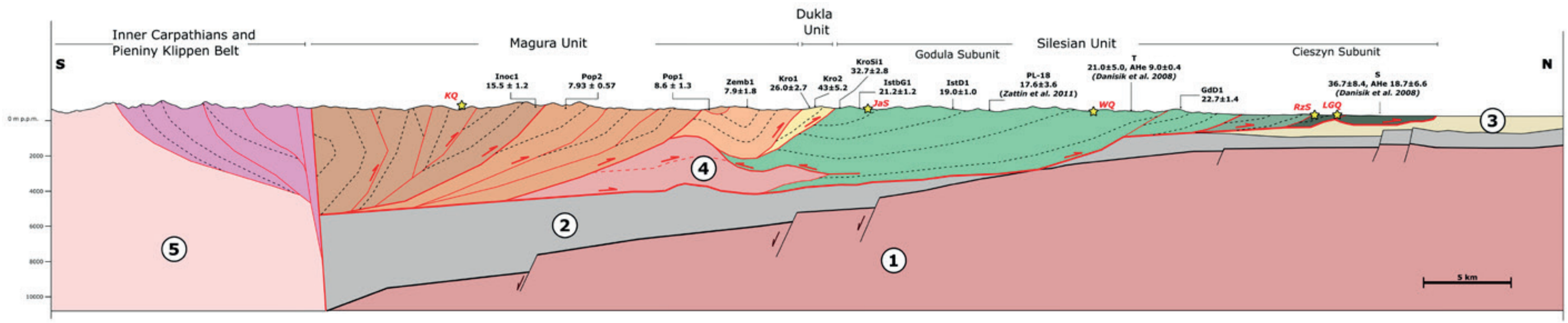


Fig. 14. Regional geological cross-section showing the structure of the western Polish Carpathians with apatite fission track ages: 1 – the crystalline basement of the North European Platform; 2 – the autochthonous sedimentary cover of the North European Platform; 3 – foredeep; 4 – Dukla Unit; 5 – Inner Carpathians. LGQ – Leszna Górna Quarry; RzS – Rzyki Stream; WQ – Wisła quarry; JaS – Janoszka Straem; KQ – Klubina qaurry (based on Barmuta *et al*., 2023; simplified)

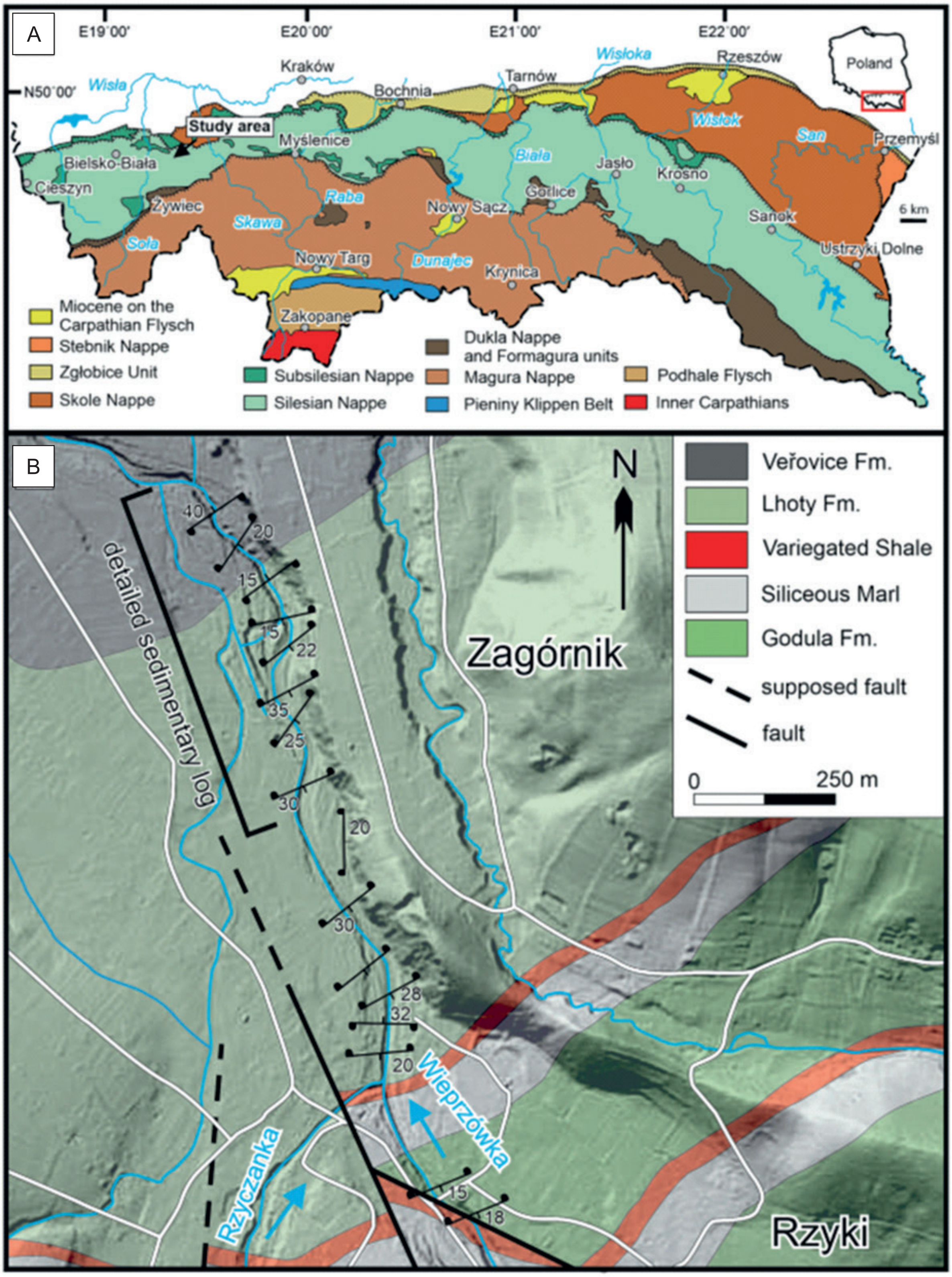


Fig. 15. Location map: A – general context of the Silesian Nappe in the Polish Carpathians; B – study area in the vicinity of Zagórnik and Rzyki villages (after Łapcik, 2023)

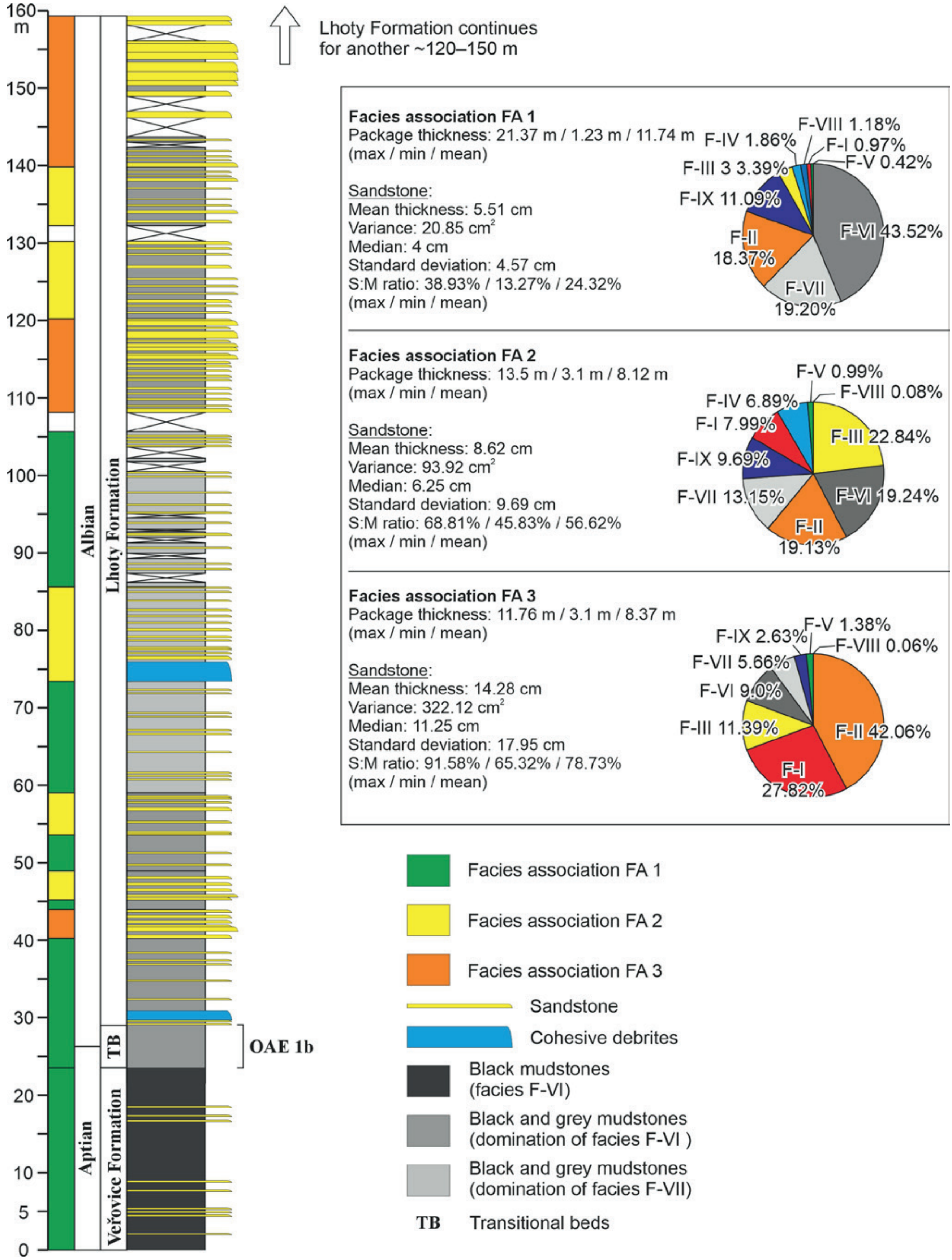


Fig. 16. General sedimentary log of the Zagórnik-Rzyki studied section with distinguished packages of particular facies associations FA1–FA3 and statistical characteristics for each of them. OAE 1b – interval with the carbon isotope anomaly referred to as equivalent to the anoxic oceanic event (Wójcik-Tabol & Ślączka, 2015) (after Łapcik, 2023)

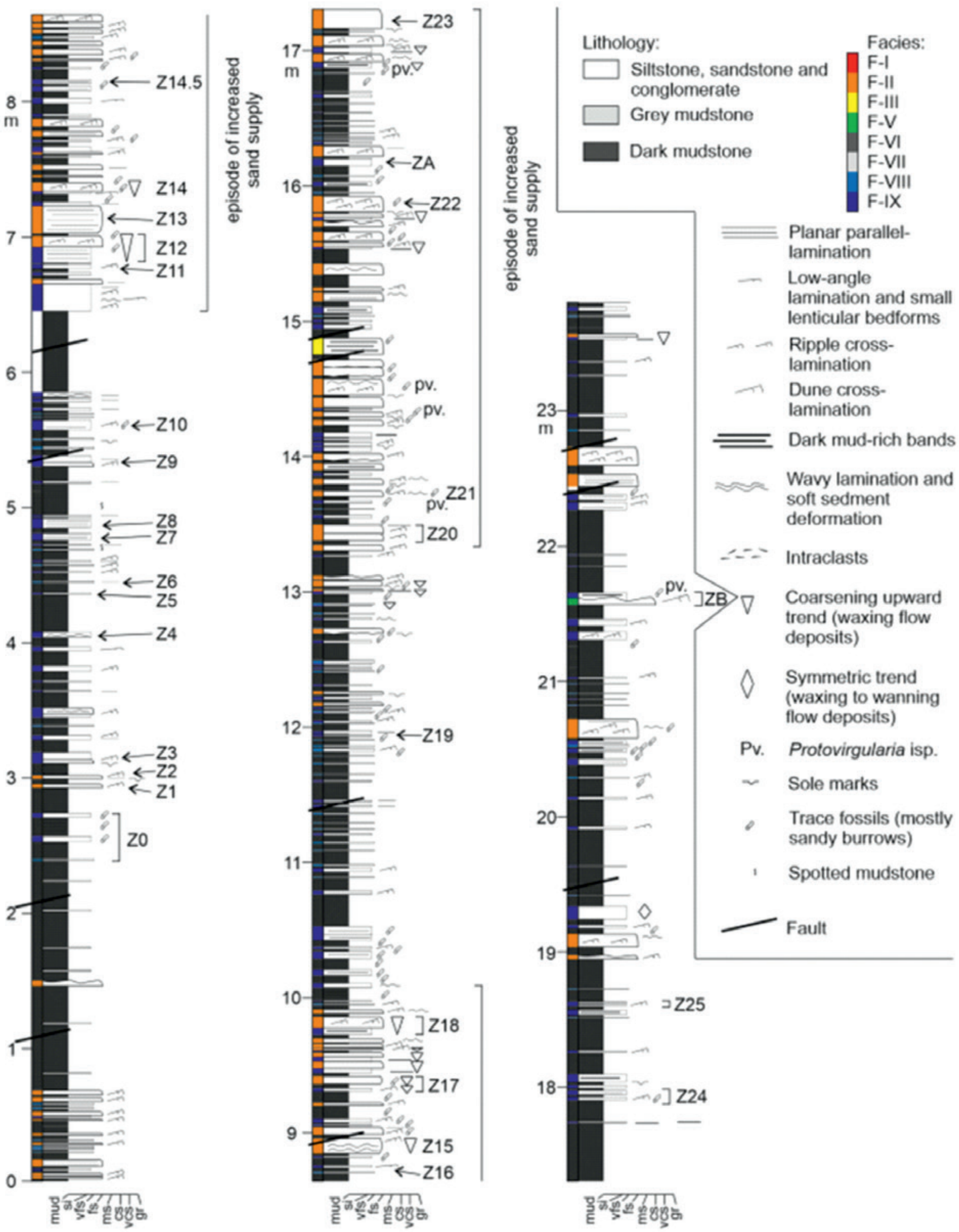


Fig. 17. Representative log of the lobe flank dominated by mud (facies association FA1) with marked samples positions (after Łapcik, 2023)

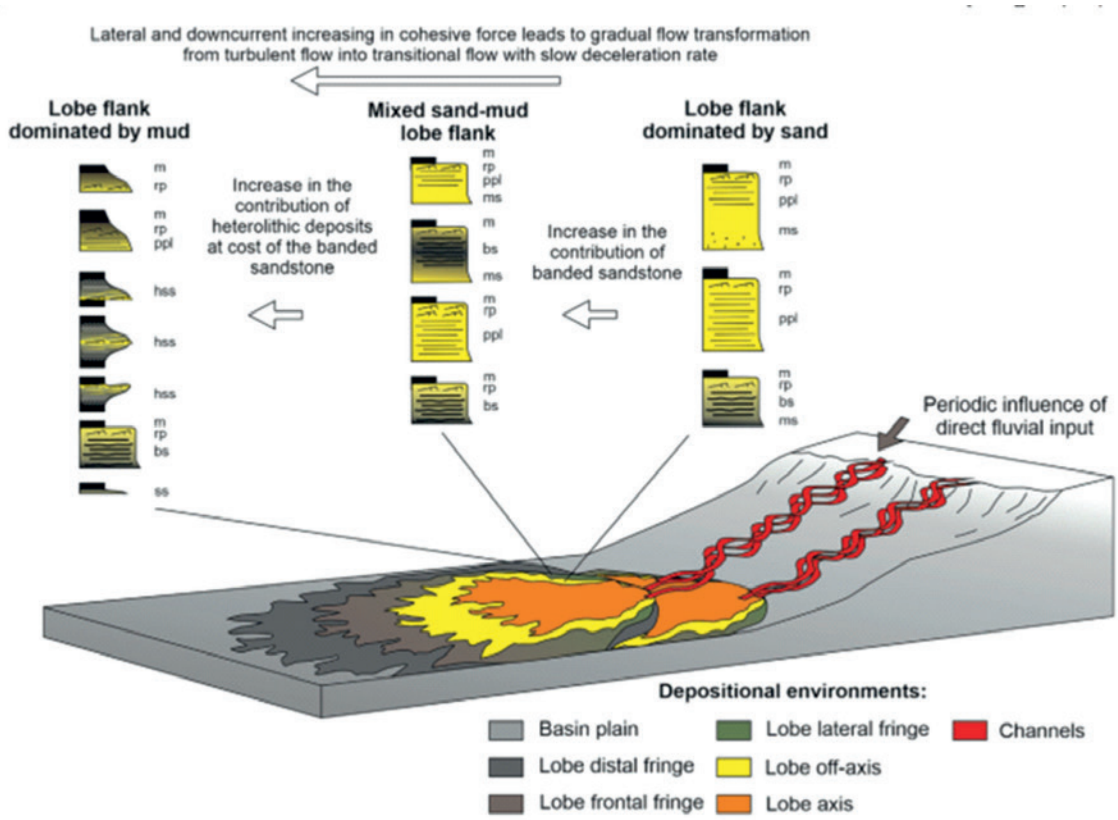


Fig. 18. Studied deposits of the Veřovice Fm and Lhoty Fm in the context of a submarine fan with a general trend of evolution of the most commonly occurring bed types: m – massive mudstone; rp – current ripples; ppl – plane-parallel lamination; ms – massive sandstone; bs – banded sandstone; ss – thin siltstone; hss – sedimentary structures of heterolithic deposits (after Łapcik, 2023)

The ichnofossil *Phycosiphon incertum* is abundant within this interval. This characteristic section represents the initial phase of oceanic circulation recovery and enhanced oxygenation of the seafloor environment (Uchman & Cieszkowski, 2008; Strzeboński *et al.*, 2009; Wójcik-Tabol & Ślączka, 2015). The improved ventilation following deposition of the Veřovice Formation may potentially be linked to the late Aptian decrease in global temperatures (Scott, 1995; Price *et al.*, 1998), which could have invigorated oceanic circulation patterns and it is consistent with the increasing diversity and abundance of trace fossils (Uchman *et al.*, 2022). Wójcik-Tabol & Ślączka (2015) proposed the presence of deposits linked to OAE 1b at the transition to the overlying Lhoty Beds, based on a carbon isotope anomaly (indicated by low δ13C values).

The Lhoty Formation begins with a sequence of intercalated thin-bedded sandstones, black and greenish-grey mudstones, and heterolithic sandstone-mudstone. These non-calcareous quartz arenites exhibit a colour spectrum ranging from white and grey to yellow and rusty orange. The total thickness of this basal interval reach ca. 75 m. Up the section, the contribution and thickness of sandstone beds steadily increase within a subsequent 55 m of section dominated by medium to thick beds. The Lhoty Formation continues for at least 120–150 m and passes into the Cenomanian Barnasiówka Radiolarian Shale Fromation (e.g., Bąk K. *et al.*, 2001; Cieszkowski *et al.*, 2003). Throughout the entire section of the Lhoty Formation proportion between black and greenish-grey mudstones vary and either may dominate.

Dinocyst analyses indicate a late Albian age for a layer situated 12 meters above the base of the Lhoty Formation, while the 55-meter thick upper part of the succession is dated on the Albian–Middle Cenomanian (Cieszkowski, 2001, 2003). Based on foraminiferal assemblages, deposition of the Lhoty Formation is inferred to have occurred within an upper bathyal setting (Książkiewicz, 1975). However, the presence of non-calcareous hemipelagites at the pinnacle of turbidite-hemipelagite rhythms suggests deposition below the local lysocline.

Depositional setting of the Lhoty Formation was interpreted as varied depositional lobe flank sub-environments in relatively proximal area. This is supported with the rare occurrence of lag facies, abundant banded sandstone facies, complex waxing-to-waning flow structure recorded in the heterolithic facies and erosional features that occur even in the thin heterolithc deposits. Three facies associations represents a gradual shift closer towards the main distributary paths, from lobe distal fringe, through lobe fringe and to lobe off-axis, which is represented by increasing sand content and bed thickness (Łapcik, 2023). The lobe deposits of Zagórnik-Rzyki section lack typical signature of progradational trends or compensational stacking, hence more random distribution of thick-bedded sandstone is related with partial confinement of the depositional area, which hinders typical stacking of lobe elements (Łapcik, 2023).

While evidence suggests at least partially anoxic conditions on the seafloor during the deposition of the Lhoty Formation, recent research has proposed correlations of two such events to OAE 1c (Wójcik-Tabol & Ślączka, 2015), and OAE 1d (Górny *et al*., 2022). The majority of green-grey shales within the Lhoty Formation exhibit trace fossils such as *Planolites*, *Chondrites*, and *Thalassinoides* against a thoroughly bioturbated background. This particular ichnofabric signifies a generally enhanced level of oxygenation compared to the underlying Veřovice Formation. While *Phycosiphon incertum* is still present in the lower portions of the Lhoty Formation, its occurrence becomes less frequent in the middle and upper sections. Additionally, *Protovirgularia* is observed very rarely. Bioturbation extends upwards to encompass the entirety of the turbiditic sandstone beds, with *Thalassinoides* being abundant in certain layers. Questionable specimens of *Zoophycos* isp. and *Nereites* isp. have been identified within the uppermost section. The soles of turbidites within the Lhoty Formation exhibit a paucity of trace fossils, with those present primarily consisting of semi-reliefs of *Thalassinoides* and *Planolites*. The presence of "*Arthrophycus*" *tenuis* is noted in specific beds, while graphoglyptids are entirely absent. Notably, this location marks the first documented occurrence of *Scolicia* within the Flysch Carpathians (Książkiewicz, 1970, 1977a). In rare instances, dark grey or black shales are observed underlying turbidite layers, and these lack any evidence of trace fossils or bioturbation activity. While this suggests the occurrence of brief anoxic events, the overall oxygenation of the Lhoty Formation sediments was significantly more prevalent compared to the underlying Veřovice Formation.

## Stop 2 – Leszna Górna quarry – carbonate flysch (lowermost Cretaceous, Berriasian) (Figs 8, 13, 14, 19, 20)

(*Jan Barmuta*, *Krzysztof Starzec*, *Justyna Kowal-Kasprzyk*, *Lothar Ratschbacher*, *Michał Krobicki*)

The Leszna Górna quarry is located in the northern part of the Silesian Unit, i.e. the Cieszyn Subunit (Fig. 14), in the folded zone of the Kopiniec – Jelenice thrust-sheet. In the quarry, the Upper Tithonian(?) – Berriasian rocks named the Cieszyn Limestone Formation, are excavated, which are the oldest rocks incorporated into the thrust-and-fold belt.

The unique character of the Cieszyn Limestone Formation, also known as Lower and Upper Cieszyn limestones, is related to the origin of the Proto-Silesian Basin and the uplift of its shoulders, so-called the Silesian and the Baška-Inwałd ridges, which were the source area for these rocks. The ridges were covered by shallow-water, Štramberk type carbonates (Słomka, 1986, 2001), which were eroded due to the constant uplifting, most likely linked with the Neo-Cimmerian tectonic phase (Golonka *et al*., 2003). The detrital material produced by this process was transported to the deeper, axial part of the basin (Słomka, 1986; Leszczyński & Malik, 1996). The Cieszyn Limestone Formation is thus interpreted as deep-water allodapic limestones deposited from calciturbiditic and calcifluxoturbiditic currents. The cessation of the Cieszyn Limestone sedimentation is marked by the transition from calcareous to siliciclastic deep water sediments, which indicates that the erosion reached the crystalline basement as evidenced by the presence of exotics of metamorphic and igneous rocks (Słomka, 2001).

Within the outcrop, two different facies can be distinguished: the lower part of the outcrop (Upper Tithonian?) consisting rhythmic, thin-to-medium bedded sequence, corresponds to the outer fan facies, while the upper part, where a thickening-upward sequence can be observed, is interpreted as a lobe facies. Both facies associations are abundant with sedimentological features and structures (Fig. 19), including a full Bouma sequence, bioturbations, or evidence of underwater slumping and landslides (Malik, 1994). The diversified kinds of clasts, i.e. ooids, bioclasts (broken fragments of brachiopods, echinoids, bivalves), fragments of microbial mounds, or coral-algal reefs, found within the Cieszyn Limestone Formation document the diverse environment of the source area (Matyszkiewicz & Słomka, 2004). This area is interpreted as narrow carbonate platforms with diverse kinds of reefs (Hoffmann *et al*., 2021). In the inner part of the platform coral-microbial patch-reefs developed, and foraminiferal-algal as well as peloidal-bioclastic limestones were deposited. On the high-energy platform margin ooid grainstones and poorly sorted, detrital limestones were formed. The latter were also deposited in the peri-reefal zone. Microencruster-microbial-cement buildups developed on the upper slope of the platform, and microbial and microbial-sponge buildups in a deeper setting. Pelitic limestones with calpionellids were deposited in a deeper part of the platform slope and in a basinal area.

In conclusion, such type of allochthonous sequence represents rare example of calcareous turbiditic system in the Mesozoic fossil record (e.g., Payros & Pujalte, 2008 with literature therein). 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 (e.g., Price, 1977; Wright & Wilson, 1984; Cooper, 1989, 1990; Wright, 2004; Brookfield *et al*., 2006). On the other hand, olistoliths/exotics of coral-algal reefs or microbial-sponge mud mounds well documented a wide range of shallow-water carbonate platform facies in source areas (Waśkowska *et al*., 2008).

In the quarry, different meso- and macro-tectonic structures can be identified and interpreted in the context of the Outer Carpathian evolution (Fig. 20). The structural observations of folds' limbs, stress inversion results and published data (Koprianiuk, 2007) allowed for the identification of two phases of shortening pointing to the NNW (older) and N (younger), documenting the counterclockwise rotation of this part of Outer Carpathians, also evidenced by the palaeomagnetic data and joint analysis (Mastella & Konon, 2002; Grabowski *et al*., 2006). The same shortening directions can be traced in all tectonic units in this part of the Outer Carpathians (see the Klubina quarry). The published thermochronologic data from the Cieszyn Subunit from the adjacent Czech Outer Carpathians, suggest that the thrust-related exhumation was initiated in the latest Eocene (36,7 Ma) and lasted until the Miocene (18,2 Ma) (Danišík *et al*., 2008).

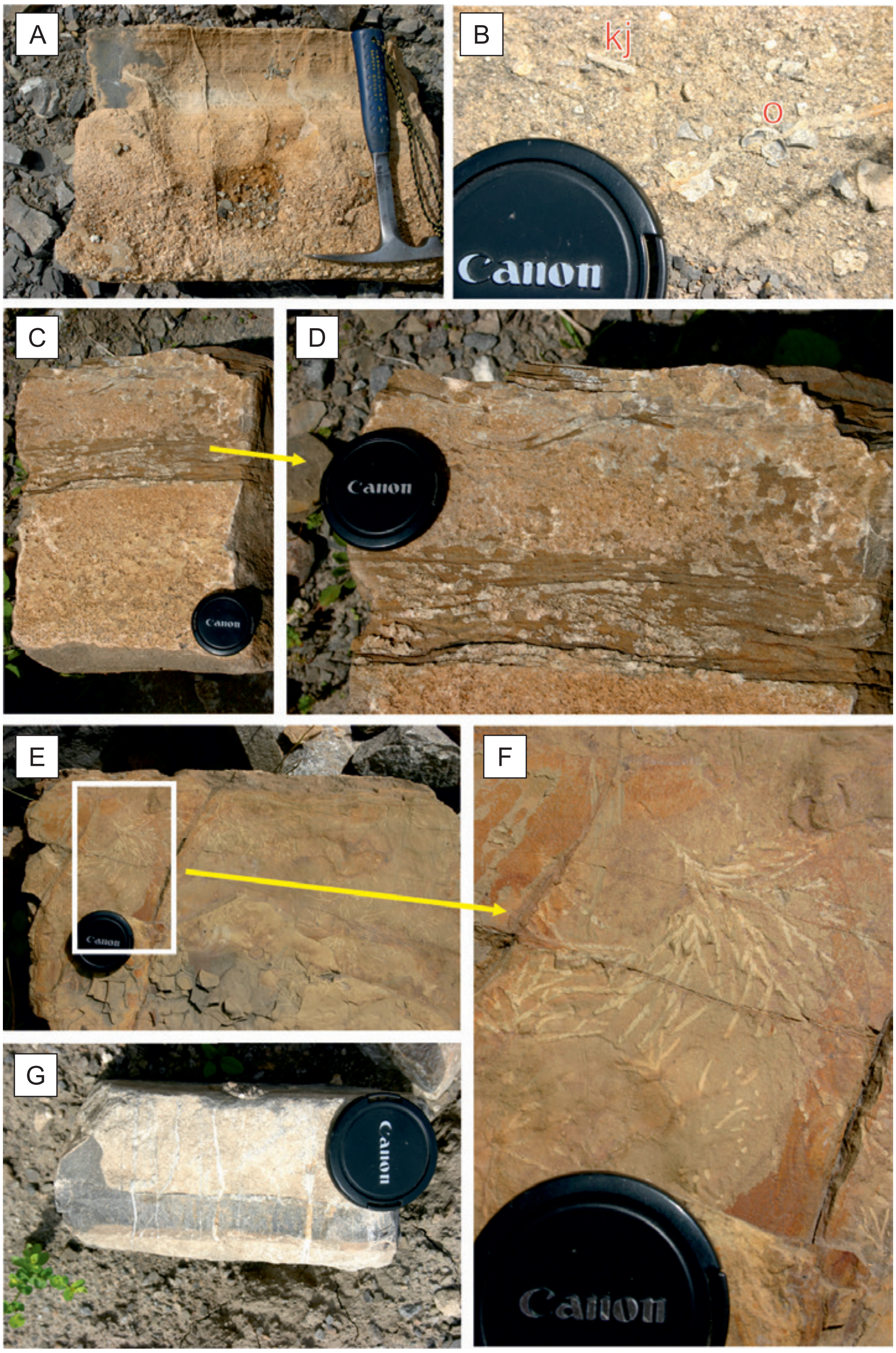


Fig. 19. Organodetrital limestones with gradational fractionation (A, C) sometimes with identifiable fossils at the base beds (B: o – *Nanogyra* sp. oyster; kj – echinoid spine) and convolution in the top (C, D) and trace fossils in more marly parts of bed (E, F – *Chondrites* isp.), and rarely with cherts (G) (after Waśkowska-Oliwa *et al*., 2008)


Fig. 20. Selected tectonic structures in the Leszna Górna quarry: A – a fold structure proving NNW shortening; B – hinge zone of the fold in picture A, with well-visible dextral shear zones (C) (hammer for a scale); D – conjugate set of normal faults formed as a result of fold axis-parallel extension.

## Stop 3 – Dolní Líštná near Třinec (Moravian part of the Czech Republic) – rare fossil polychelid lobsters in turbiditic palaeoenvironments (Lower Cretaceous, Valanginian) (Figs 8, 13)

(*Michał Krobicki*)

In the Dolní Líštná surroundings the Hradiště Formation is dominant, especially in its lower part, and was known earlier as the Upper Těšín (Cieszyn) Shales (= Oberen Teschener Schiefer of Uhlig 1902 or Oberen Těšín-Schichten of Vašíček 1975; see also Menčík *et al*., 1983) which belonged recently to the Cisownica Shale Member of the Hradiště Formation (Golonka *et al*., 2008) (Fig. 8).

The extremally sporadic benthic macrofossils of the Silesian Basin are Polychelidan lobsters which were indentified here in Hradiště Formation. They are one of the rare groups of decapod crustaceans which were first discovered as fossils long before being identified in extant deep-sea environments. As for other decapods, their fossil record is highly incomplete. Only three fossil Polychelidae have been identified to date. Species *Woodwardicheles neocomiensis* (Woodward, 1881) derived from the Valanginian (Early Cretaceous) of the Outer Carpathians (Dolní Líštná near Třinec, Moravian part of the Czech Republic). Analysis specimen is probably autochthonous or parautochthonous to turbiditic palaeoenvironments and corresponds to typical Polychelidae which inhabited deep water regime (Audo et al., 2018).

# Stop 4 – Baška village – part of the Hradiště Formation with so-called teschenite volcanics (including pillov lavas) (Figs 8, 21)

(*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 (Figs. 21, 22), 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.

Fig. 21. Teschenitic pillow lavas on the left bank of the Ostravice River near Baška village (after Skupien & Vašíček, 2008)

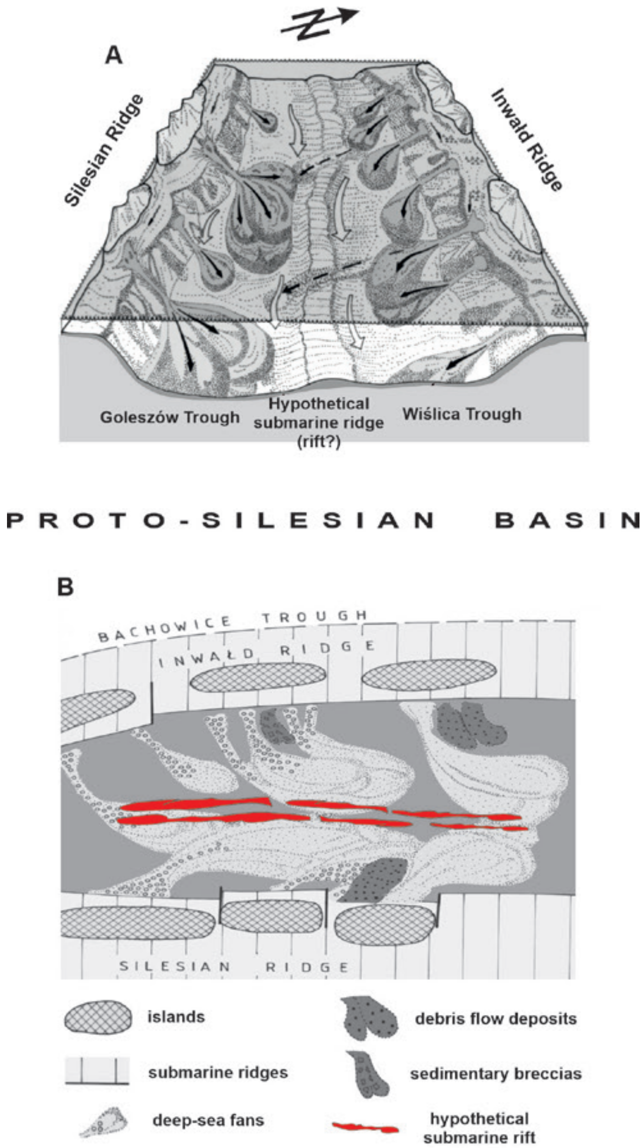


Fig. 22. Palaeogeographical blockdiagram of sedimentation of the oldest flysch deposits in the Proto-Silesian Basin (Jurassic/Cretaceous transition – Tithonian/Berriasian) (A) and its hypothetical palaeogeographical sketch (B) (after Słomka 1986a; slightly modified)

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.

## Stop 5 – Wisła quarry – siliciclastic flysch (upper Cretaceous) (Figs 8, 13, 23–26)

(*Krzysztof Starzec*, *Jan Barmuta*, *Lothar Ratschbacher*, *Saeideh Asal Seyedi*)

The Wisła quarry is located in the central part of the Silesian Unit, i.e., the Godula Subunit (Fig. 23) that is thrust over the Cieszyn Subunit described in the Stop 2. In the quarry, the Upper Cretaceous rocks of the Godula Formation are excavated, which represent the onset of the synorogenic sedimentation in the Outer Carpathian realm (Picha *et al*., 2006).

The Godula Formation represents the middle part of the sedimentary succession of the Silesian Unit (Fig. 24). In the western part of the Carpathians, in the Polish Silesian Beskid and Czech Moravian-Silesian Beskids, this formation constitutes the thickest part of the succession exceeding 3,000 m. However, to the east, its thickness decreases significantly until it is completely wedged out, moreover, the trend of decreasing thickness also occurs in the longitudinal direction, i.e., in most external parts of the Silesian Unit thickness of the Godula Formation reaches only a few hundred meters (Słomka & Słomka, 2001). Słomka (1995) related the wedge-shaped geometry of the Godula lithosome to its depositional development, as deep-sea turbidite fans and slope aprons, that were supplied from the southern Silesian ridge.

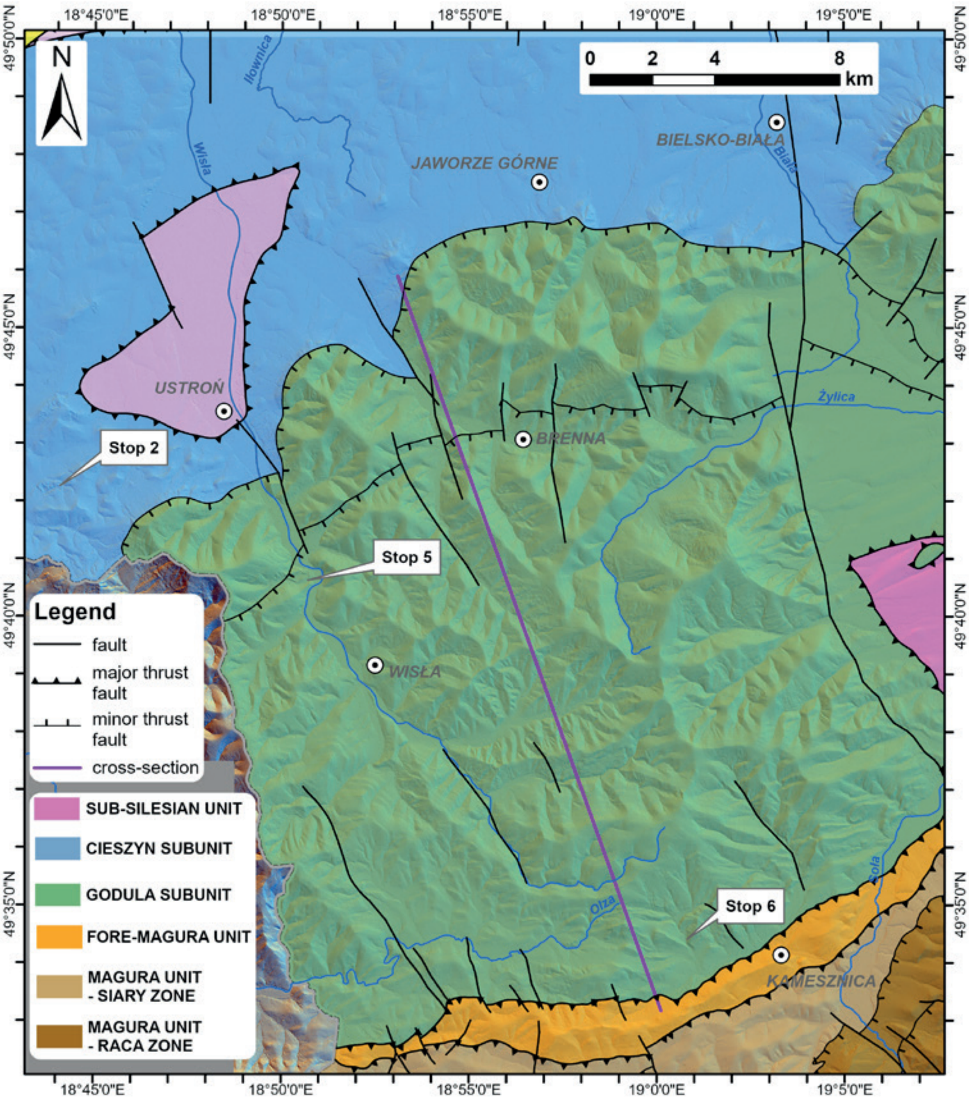


Fig. 23. Simplified geological map of the north-western part of the Polish Carpathians showing location of stops 2, 5 and 6 within the Silesian Unit. Line indicates cross-section in the Fig. 27B

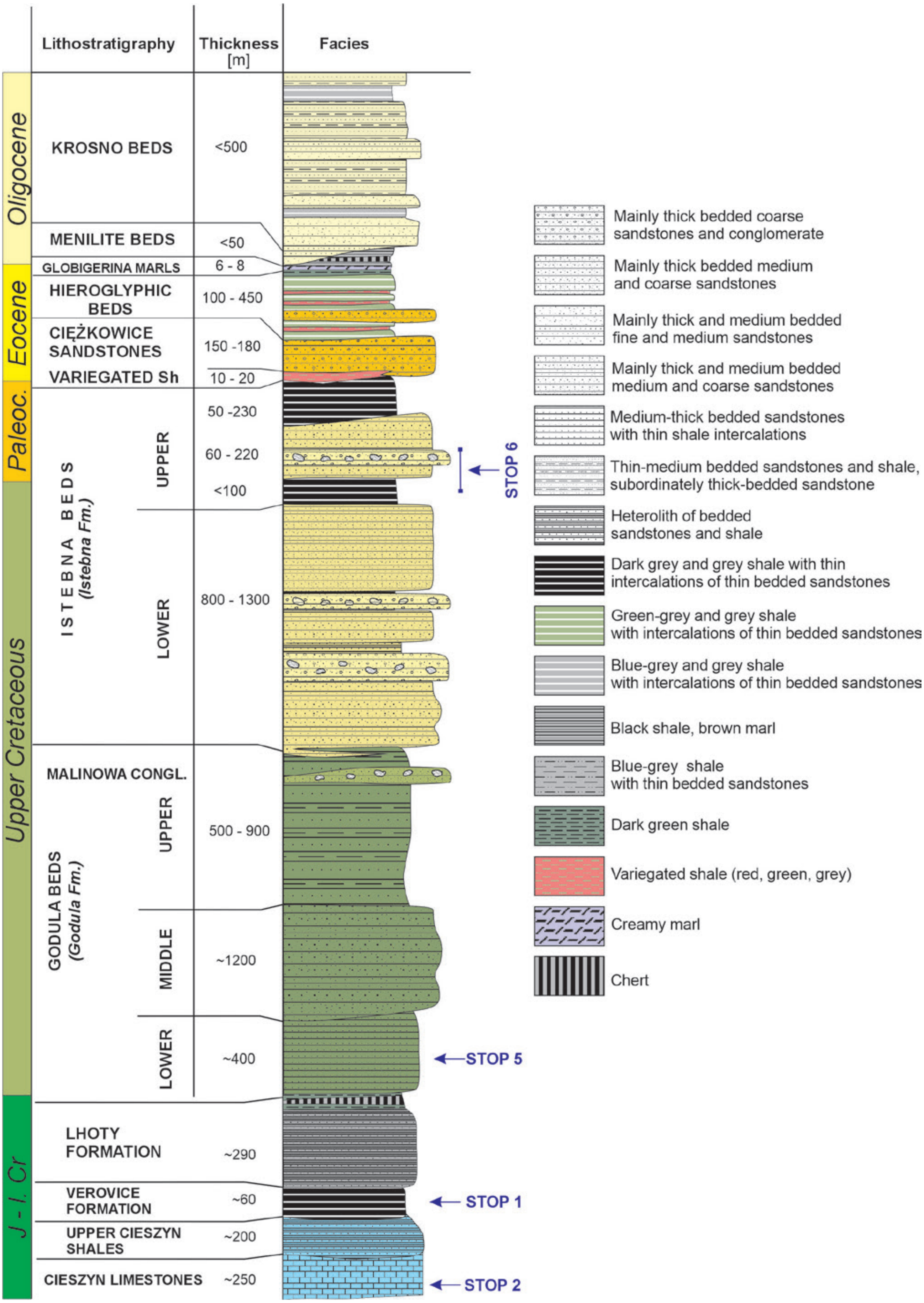


Fig. 24. Scheme of lithostratigraphic units for the Upper Jurassic-Oligocene Silesian Succession in the Silesian Beskid

Picha *et al*. (2006) linked the initiation of the Godula Formation sedimentation with the sudden rise of the Silesian ridge that was a result of compressional stresses associated with the accelerated subduction of the Penninic–Vahic ocean and the early collision of the Inner Carpathians with the fragmented margins of Europe in the early Late Cretaceous. The flysch deposits of the Godula Formation mark the beginning of the foreland depositional regime in the Outer Carpathian basins (Picha *et al*., 2006).

The Godula Formation is a typical turbidite sequence of alternating sandstone and shale layers with an admixture of conglomerates (Słomka, 1995; Maceček, 2021). The proportion of these lithological components is variable which was a basis for lithological subdivisions of this lithostratigraphic unit into three informal members (Fig. 25): the lower member is predominantly composed of thin-bedded, fine- to medium-grained sandstones and shales with a thick Ostravice Sandstones Member at the base (Cieszkowski *et al*., 2016); that is followed by a facies dominated by medium- to coarse-grained glauconitic sandstones and conglomerates in the middle part of the Godula Formation, whereas its most upper part is also thin-bedded succession, although it contains lenticular bodies of coarse-grained sandstones and conglomerates of the Malinowska Skała Conglomerate (Burtanówna *et al*., 1937; Słomka, 1995; Maceček, 2021).

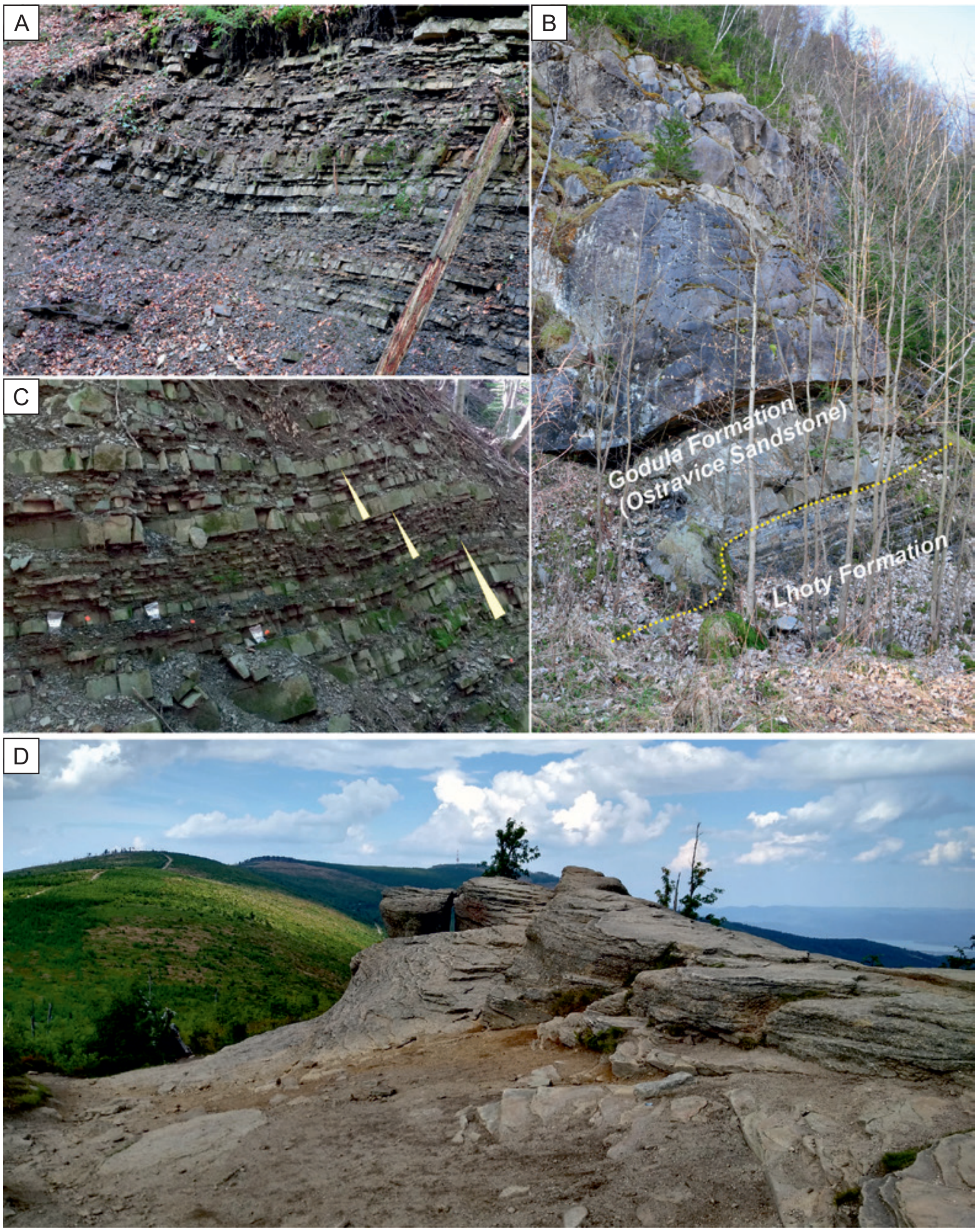


Fig. 25. Variable facies within the Godula Formation: A – thin-bedded sandstone-shale couplets of the upper Godula member; B – the lowermost part of the Godula Formation, i.e. Ostravice Sandstone Member, at the contact with the underlying Lhoty Formation; C – thin-bedded packet of sandstone-shale layers showing thinning upwards sequences; D – thick-bedded conglomerates of the Malinowska Skała member, forming the top of the Skrzyczne range

Rocks outcropped in the Wisła quarry are dominated by thin- to medium-bedded sandstones representing the lower part of the Godula Formation, however, they display significant facies variation. Facies of alternating sandstone and mudstone dominates. It is composed of medium- to fine-grained sandstones that gradually passes into mudstones (shales) (Fig. 26). Sandstone thickness ranges 10 cm to 30 cm, top and bottom layer contacts are flat. They are usually parallel laminated, subordinately cross-lamination occurs in the layers. Mudstones usually are non-calcareous and grey to green-grey. This facies is interbedded by packets of sandstone-dominated facies that comprises mainly medium-grained sandstones, often conglomeratic at the base, occurring in medium- to very thick-bedded layers (Fig. 26). There is a gradual transition between these two facies and they are arranged in thinning upward sequences. The first facies is interpreted as the result of deposition of sandy-clay material from gravity currents of different densities, whereas the second facies represents the deposition from high-density turbidity currents (Słomka, 1995; Maceček, 2021).

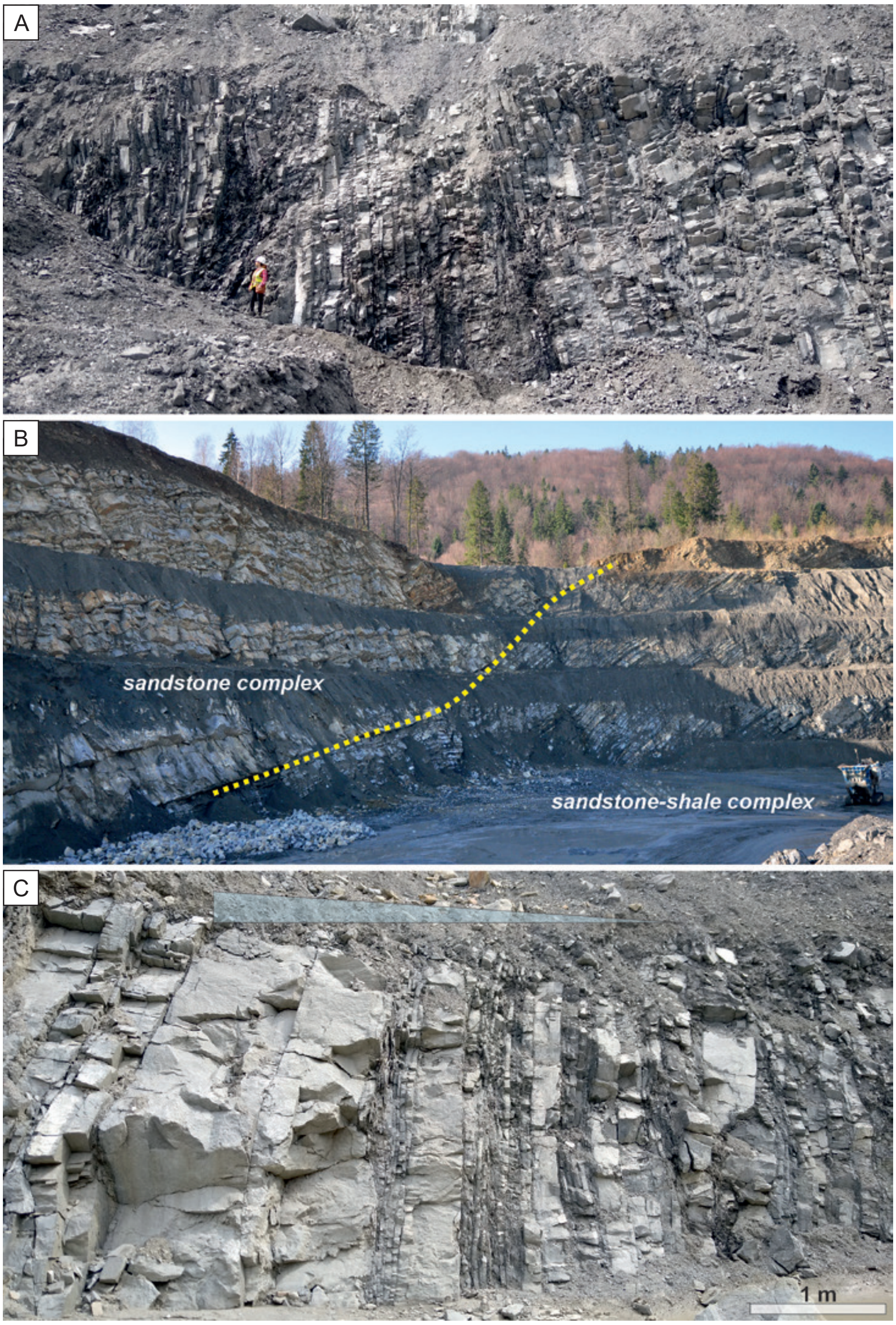


Fig. 26. Wisła quarry: A – sandstone dominated facies of thin-bedded layers; B – border (yellow line) between lower member of the Godula Formation composed of thin-bedded sandstone and shales couplets and overlying sandstone complex representing middle member of the Formation; C – sequence of thinning upward layers within the sandstone-shale complex

In general, Słomka (1995) and Maceček (2021) interpret that the high- to low-density turbidity currents were mainly responsible for the deposition of the Godula succession, subordinately the deposition took place from sandy debris flows and sporadically sediments originated from traction currents and dispersed suspensions. The facies of the Godula Formation are characteristic for deep-water turbidite fans and they represent a spectrum of sub-environments of this type of sedimentation, i.e., channel sediments, depositional lobes, inter-channel sediments, fan margins and non-channelized apron sediments (Słomka, 1995; Maceček, 2021). Thus, they are characterized by high lateral variation that reflects the dynamic depositional environment.

The Wisła quarry is located at the frontal zone of the Godula Subunit. In general view, the Silesian Unit in the westernmost part of the Outer Carpathians exhibits simple, almost homoclinal character (Fig. 27). Namely, the Silesian Unit structure starts with the Jurassic – Lower Cretaceous formations exposed in the northern marginal zone of the Silesian Unit (the Cieszyn Subunit) and continues through the thickest middle part of the Late Cretaceous to Paleocene deposits (the Godula Subunit including the Godula and Istebna formations), to the uppermost complex of Eocene and Oligocene deposits. Strata generally dip gently towards the south at ca. 20° (Fig. 27B). When compared to the imbricated structure of the Cieszyn Subunit, the Godula Subunit is composed of more competent formations and formed only two thrust sheets. Duplication of the lower part of the Godula sequence permits the identification of two thrust sheets within the northern part of this Subunit, of which the northern one is comprised solely of the Godula Formation.

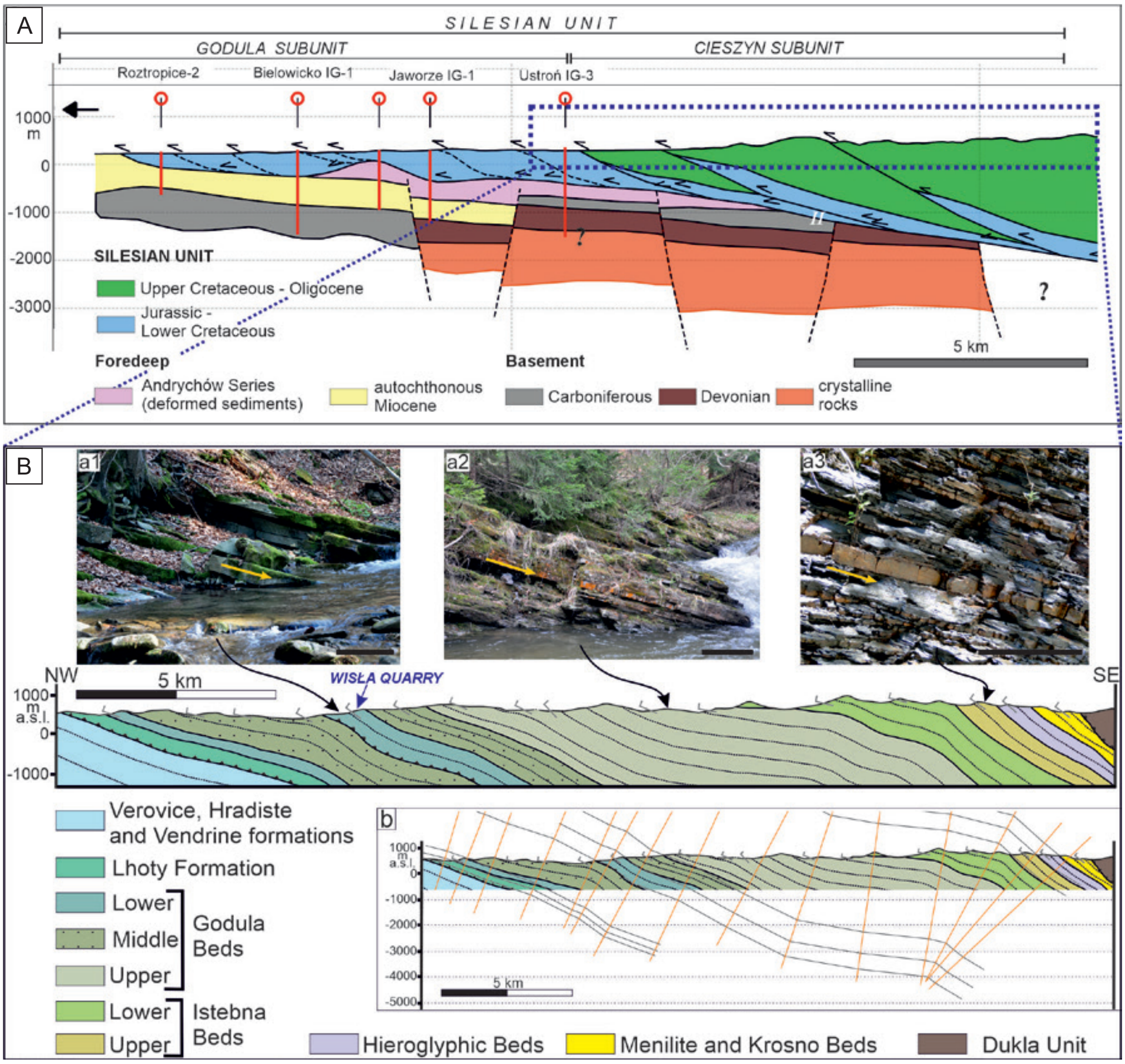


Fig. 27. Cross-sections of the Silesian Unit: A – geological structure of the unit based on surface and borehole data; B – detailed cross-section through the Godula Subunit extrapolated from the surface data, an outcrop-scale homoclinal structure of different parts of the subunit is shown in the photos: (a1) sequence of thin- to medium-bedded sandstones typical of the lower Godula member, (a2) sequence of mainly thin- and medium-bedded sandstone–shale interbeddings characteristic of the upper Godula member, (a3) thin-bedded black shales intercalated with sandstones and siderites typical of shaly interval of the Upper Istebna Beds; (b) geometry of strata approximated with the use of the Move Software by projection from dip data with the kink-band method

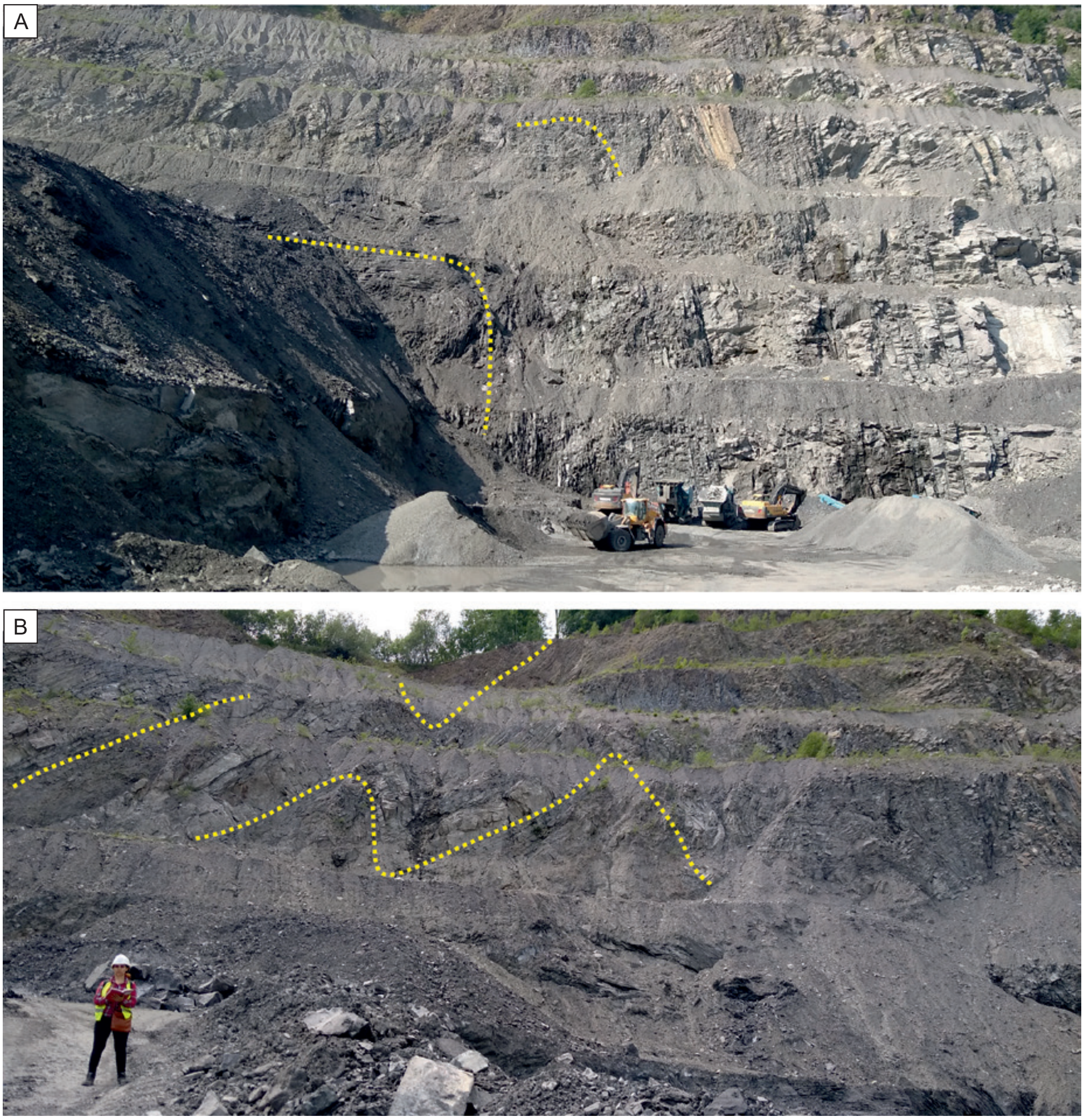


Fig. 28. Tectonic deformations formed at the front of the southern thrust sheet within the Godula Subunit: A – lower part of the Wisła Quarry; B – upper levels of the Quarry

The southern one represents the main part of the Subunit, including its complete succession from the lowermost part of the Godula Formation to the Oligocene in age Krosno Beds, and on the entire length it preserves undisturbed stratigraphic succession. Within the Wisła Quarry meso- and macrotectonic structures are observable that characterized the front of the southern thrust sheet (Fig. 28). For the Godula Subunit, a general decreasing character of the AFT age toward the north can be inferred. It is a typical AFT age pattern related to exhumation due to thrusting (Almendral *et al.*, 2015), which supports the map-scale picture of the southern part of the Godula Subunit as a single block rotated due to reverse faulting.

## Stop 6 – Janoszka stream in Kamesznica village sandy-to-gravelly debris flow deposits (Upper Cretaceous–Paleocene) (Figs 8, 13, 27–30)

(*Krzysztof Starzec*)

The Stop is located in the Kamesznica village along the Janoszka stream. The sequence of the Upper Cretaceous–Paleocene Istebna Formation is exposed along the stream

(Fig. 29). The rocks outcropped in this area belongs to the southern part of the Silesian Unit. They strike latitudinally about 2,5 km north of the thrust front of the structurally higher Fore-Magura Unit.

This formation was first described as the Istebna Beds by Burtan (1936) and characterized by Burtanówna *et al*. (1937). The Istebna Formation is spread from the Moravian-Silesian Beskids in the Czech and Slovak Republic through the Silesian Beskid, Mały Beskid, up to the Bieszczady Mountains area in the eastern part of the Polish Carpathians (see Żytko *et al*., 1989). This formation is best developed in the Silesian Beskid, reaching around 2000 m in thickness (e.g., Burtanówna *et al*., 1937; Unrug, 1963, 1968). The lower 1,500 m of this Formation belongs to the Lower Istebna Beds (represented by the Lower Istebna Sandstones), and over 400 m represents the Upper Istebna Beds that are subdivided into three members: the Lower Istebna Shale, Upper Istebna Sandstone and Upper Istebna Shale (Burtanówna *et al*., 1937; Burtan *et al*., 1956; Burtan, 1972, 1973). These subdivisions are clear foremost in the stratotype area of the Istebna Beds in the Silesian Beskid (Strzeboński, 2015), whereas more to the east of the Outer Carpathians, the Lower Istebna Shale and the Upper Istebna Sandstone pinch out locally, so the remaining subdivisions are merged and indistinguishable (Książkiewicz, 1951; Strzeboński, *et al*., 2017). In the Moravian-Silesian Beskids, in turn, a few to several principal lithological horizons are distinguished within the Istebna Formation, which are hard to correlate with the above-mentioned subdivisions from other areas (Strzeboński & Uchman, 2015).

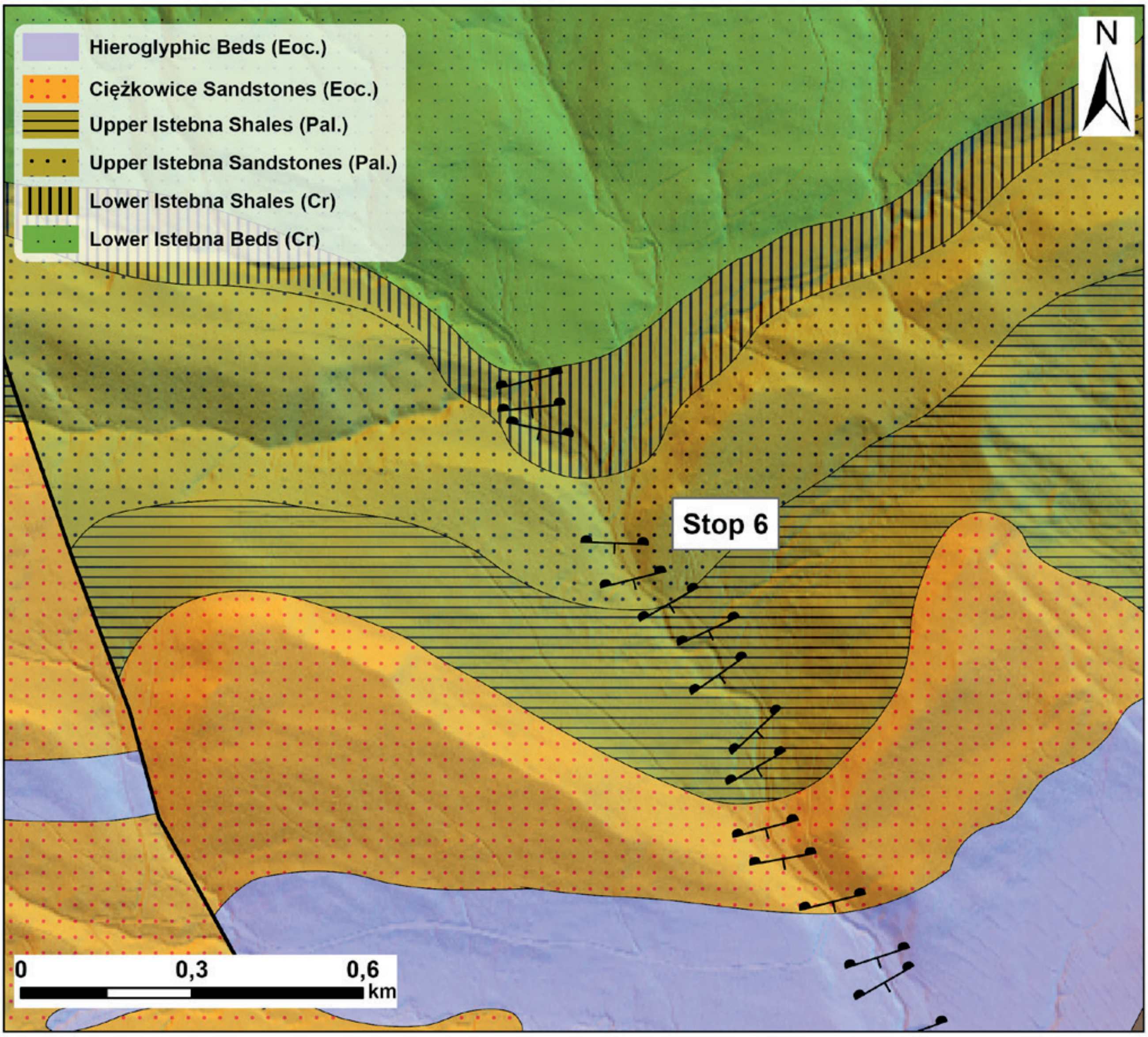


Fig. 29. Detailed geological map of the Kamesznica area. Stop 6 includes outcrops along the Janoszka stream

The Lower Istebna Beds is a sand-dominated lithostratigraphic unit. It consists mostly of thick-bedded, amalgamated, coarse-grained sandstones and granule/fine-pebble conglomerates with rare interbeds of shales or thin-bedded fine-grained sandstone-shale packets. They form the largest and highest part of the Silesian Beskid range.

The tripartite lithostratigraphical unit of the Upper Istebna Beds occupies a much smaller area. Its middle part is built of similar sandstones and conglomerates as in the lower member, and it is sandwiched between the lower and upper shale members represented by dark grey to black, very thin bedded mudstones, abundant in sphaerosiderites (Fig. 30). The sandstone complex reaches about 100 m in thickness, while the cumulative thickness of the mudstones complexes is about 300 m (Starzec *et al*., 2017), although their thickness, and as a consequence, the width of the outcrops, varies along the strike. This particularly holds true for the sandstone complex. The layers of both the lower and upper members of the Istebna Formation lie concordantly, being tilted to the south at an angle of about 22° (Fig. 29).

The continuous sequence of the Upper Istebna Beds can be traced in the Janoszka riverbed and banks. The sequence includes contact between the shale and sandstone/conglomerate members in which a distinctive and abrupt change of facies is visible, i.e., the complex of black sandy shales, with only occasional intercalations of thin sandstone layers, is overlain by very thick-bedded, coarse-grained sandstones and conglomerates (Fig. 30). The latter contain pebbles and even boulders up to 70 cm in diameter. The sandstone-conglomerate layers are characterized mostly by massive or graded structure. The sedimentary structures are especially well visible within the rocky sandstone landforms that are scattered throughout the area. Within the tors south of the Stoczek, i.e., a hamlet of the Istebna village, the sedimentary sequence begins with a thick conglomerate layer that reveals normal grading (Fig. 31). Quartz and different crystalline lithoclasts form the main body of the conglomerate, with maximum size reaching about 4–6 cm. In the middle part of the layer shale clasts and more often caverns after these clasts occur (Fig. 31). The caverns are elongated, usually disc-shaped, and sometimes contain only remnants of the material that once filled them. Shale clasts are interpreted to have formed when already lithified shaly deposits were torn from the underlying beds by gravity currents. The fragments of shales were subsequently, deposited with coarse-grained material (Stadnik & Waśkowska, 2015). The conglomerate layer reveals a sharp top surface markinga distinctive border with another conglomerate. Within the lower part of the latter one, a thirty-cm thick interval with reverse grading occurs, passing to the top into normal grading. The normal graded interval starts with granules, reaching about 1 cm to 6 cm in diameter, and changes to the top of the layer into medium to fine gravel. At about 1.3 m above the bottom of the layer, the large crystalline boulders are enclosed. The biggest boulder reaches 70 cm. They have different shapes from angular to well rounded.

The conglomerate is followed by a sandstone layer of ca. 1.7 m thickness. This sandstone reveals rather poorly defined normal grading with dispersed fine gravel-sized grains. Within some parts of the layer a through cross-stratification can be recognized, the origin of which is unclear. This kind of structure has been interpreted as a sedimentary structure (Ślączka & Thompson, 1981; Dziadzio *et al*., 2006), but recently Leszczyński *et al*. (2015) described it as a tensile fracture. Moreover, the sandstone contains isolated, lenticular gravel pockets (Fig. 32), up to 80 cm in length and 20 cm in thickness that are irregularly distributed within the layer. The top contact of the sandstone with the overlaying conglomerate is sharp, but highly uneven. It is a result of the loaded erosional base of the conglomerate with large load-flame structures. The lithology and sedimentary features of deposits correspond to the sequence described by Lowe (1982) (Fig. 31). Within individual layers, the R2-R3-S1 or R3-S1 divisions can be recognized, i.e., very coarse and coarse-grained conglomerate with reverse grading (R2), upwards with normal grading (R3), turning into to coarse-grained sandstone with lenses of pebbles (S1).

Siliciclastic material of these deposits was accumulated in the Silesian Basin within a slope-apron depositional environment, mainly by sediment gravity-driven processes (Fig. 33). Initially, the material was chaotically redeposited from edge of a shelf-margin by mass-wasting processes (slides, slumps) and mass flows (avalanche, different types of debris flows) in the form of elongated tongues, which formed covers of a proximal slope apron. During further downslope redeposition, these slumps or debris flows were frozen or partly transformed into turbulent fluid-sediment gravity flows, including turbidity currents (e.g., Unrug, 1963; Strzeboński, 2015, 2022). Later, structural changes of these sediments and final accumulation could took place under the influence of in-situ liquidization, bottom current reworking with tractional deposition and hemipelagic sedimentation also occurred in the background.

The palaeotransport directions in the upper Istebna sandstones indicate the south-to-north direction of the material's supply, i.e., the hypothetical Silesian Ridge (cordillera) was the source area for the crystalline clasts. The Silesian Ridge represented an elevated fragment of the sea floor developed along the south edge of the North European Platform. Initially, during Jurassic and Cretaceous times, the Silesian Ridge separated the Magura and Severinic-Moldavidic (Protosilesian) basins. It formed a barrier between the Silesian and Fore-Magura basins after the Late Cretaceous geotectonic reorganization (e.g., Golonka *et al*., 2005, 2006a, 2006b, 2013). It was later destroyed in the process of expansion of the Carpathian accretionary prism and now is known only from olistolits and exotics.

The crystalline blocks represent magmatic, metamorphic, and rarely sedimentary rocks with domination of varied white-grey gneisses.

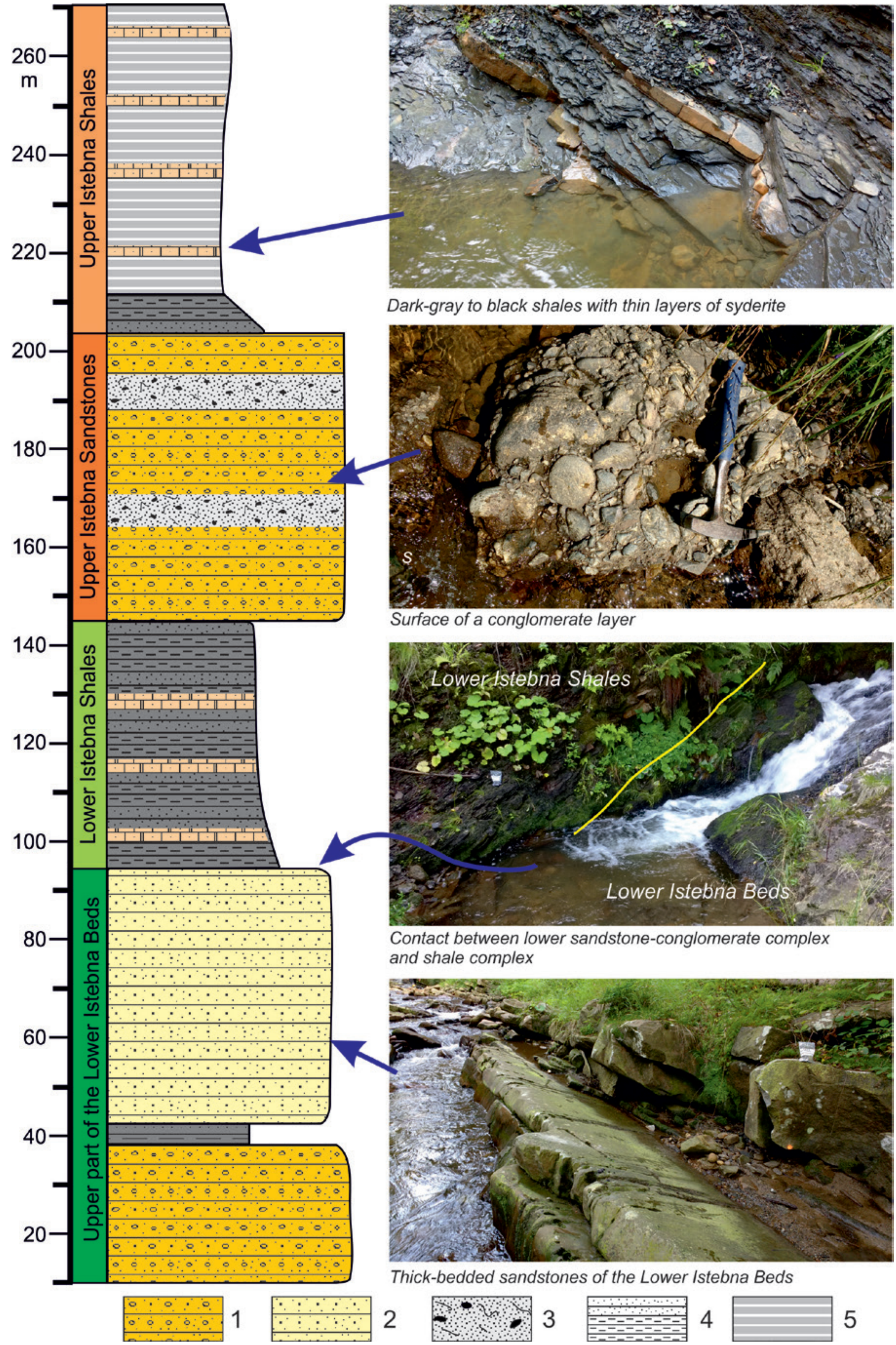


Fig. 30. Schematic lithological log of the upper part of the Istebna Formation. 1 – thick-bedded conglomerate and sandstone, 2 – thick-bedded, mainly coarse-grained sandstone, 3 – gravelly mudstone, 4 – sandy mudstone intercalated by thin-bedded sandstone and siderite, 5 – mudstone with thin-bedded siderite

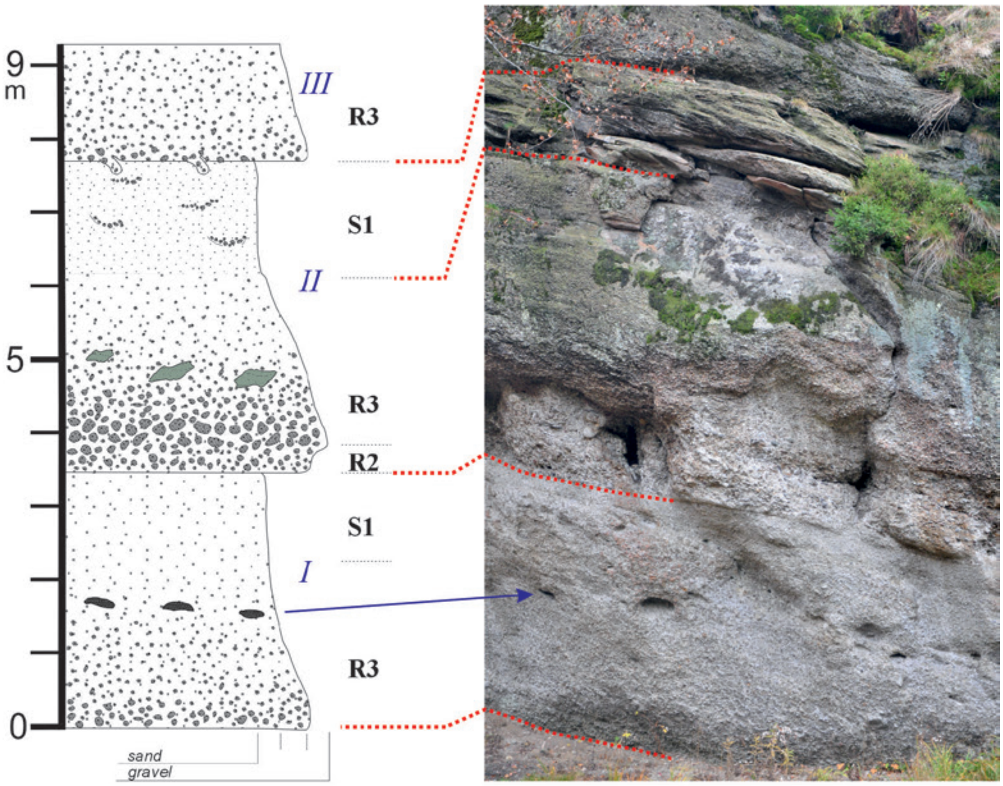


Fig. 31. Lithofacies profile of the part of the Upper Istebna Sandstones with the facies succession corresponding to Lowe's succession intervals; Bed I – very coarse and coarse-grained conglomerate displaying normal grading (R3) transiting to medium and fine-grained conglomerate (S1); Bed II – very coarse- and coarse-grained conglomerate with reverse grading (R2), upwards with normal grading (R3) and large gneiss blocks, transiting to coarse-grained sandstone with lenses of pebble; Bed III – coarse-grained conglomerate displaying normal grading (after Starzec *et al*., 2017)

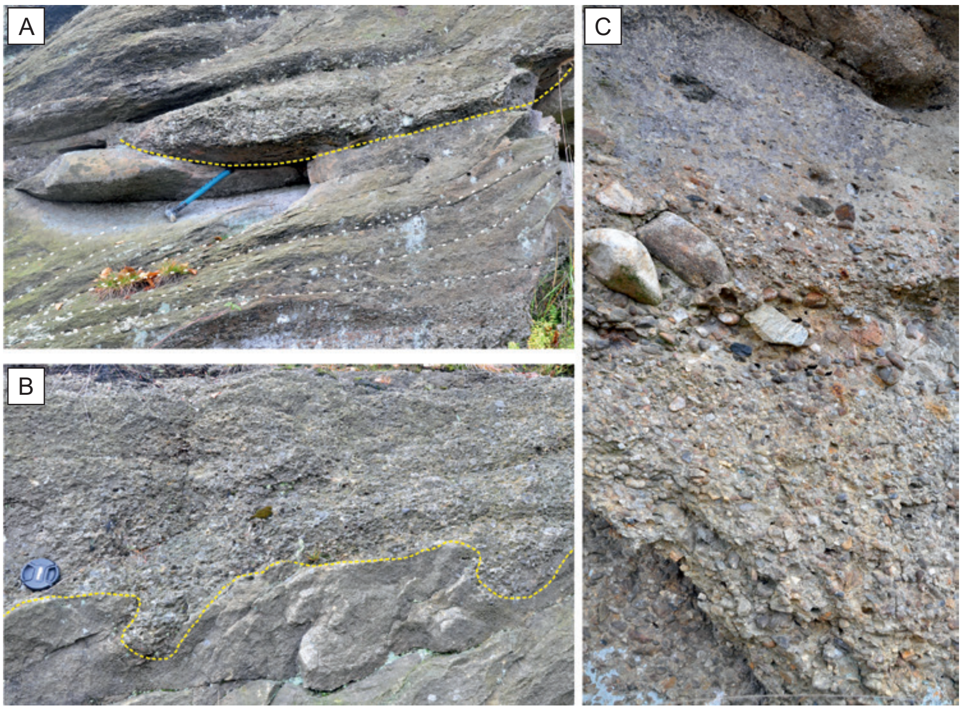


Fig. 32. Sedimentological features of the Istebna sandstones: A – fragment of sandstone bed with cross-bedded laminae marked by white lines and with lense-shaped conglomerates marked by yellow line laminae; B – loading deformations at the boundary of conglomerate and sandstone layers marked by yellow line; C – pebbles and cobbles of gneisses of various roundness constituting the basic components of conglomerate; photo K. Starzec

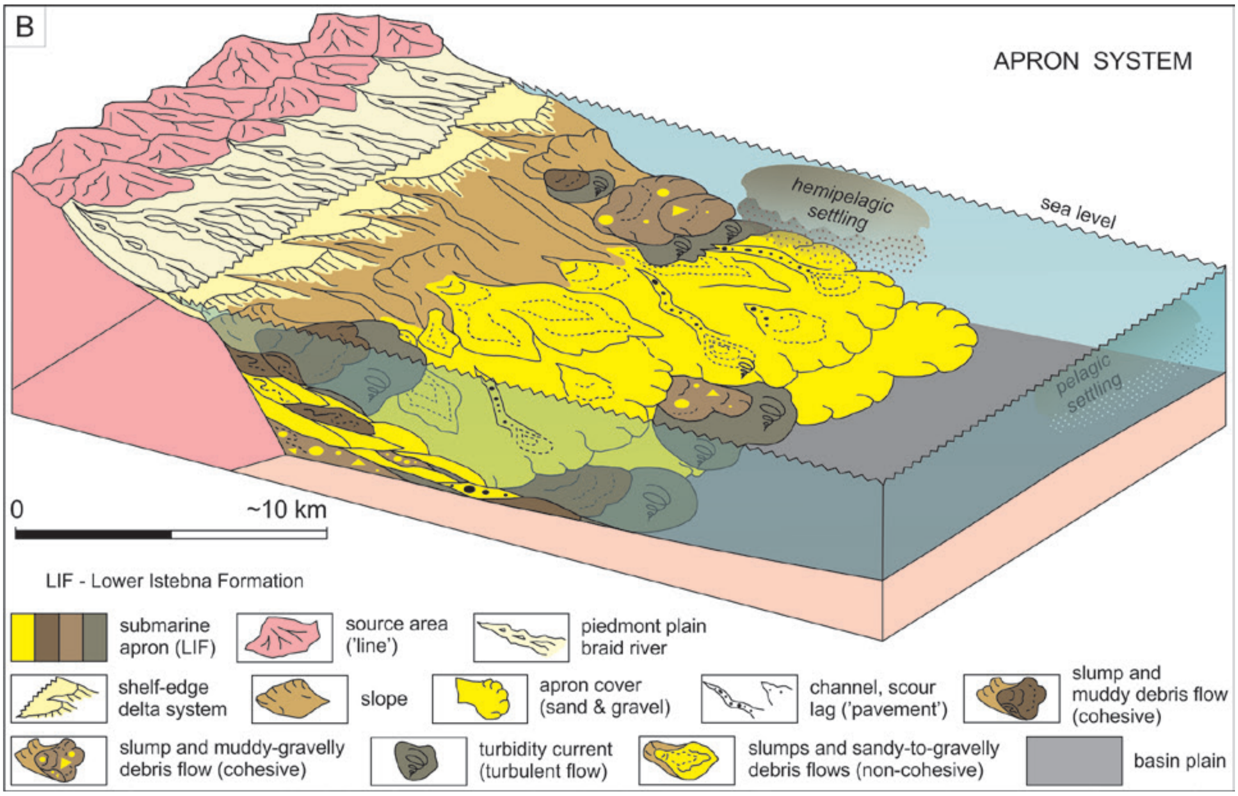


Fig. 33. Sedimentation model of the depositional system for the Istebna Formation – line-supplied slope resedimentation apron (after Strzeboński, 2022, with permission of author)

These gneisses together with schists, phyllites, pegmatites, milky or dark quartzites, pinkish granites, and sporadically dark limestones, represent the rocks inventory of the basement upon which the Carpathian basins, developed during Mesozoic and Cenozoic times. This, so far poorly known basement is customarily called the "Protocarpathians" (Gawęda & Golonka, 2011). The Protocarpathian exotic material plays a key role in palaeotectonic reconstructions. Data from the felsic crystalline clasts imply that the Silesian Ridge was an eastern prolongation of the Brunovistulia microcontinent (Gawęda *et al.*, 2019).

## Stop 7 – Klubina quarry – siliciclastic flysch (upper Eocene) (Figs 8, 13, 31–34)

(*Krzysztof Starzec*, *Jan Barmuta*, *Lothar Ratschbacher*, *Saeideh Asal Seyedi, Piotr Łapcik*)

The abandoned Klubina quarry is located in Slovak part of the Outer Carpathians, about 7 km south the Polish/Slovak border. This area belongs to the highest tectonic unit in the Western Carpathians, i.e., the Magura Unit (Nappe). The Unit forms a continuous belt along the Western Carpathian arc from the Vienna Forest in Austria to the Western Ukraine (Picha *et al.*, 2006). The sedimentary succession of the Magura Unit includes mostly flysch type deposits that evolved during the Late Cretaceous and Paleogene at the convergent stage of the Carpathian area development (Picha *et al.*, 2006).

Rocks exposed in the quarry represent the youngest (Late Eocene) stage of sedimentary infill of the Magura Basin. They are assigned to the Kýčera Member of the Zlín Formation of the Rača Subunit (Staňová *et al.*, 2009), i.e., one of the large thrust units within the Magura Unit. The Zlín Formation is regarded as an equivalent of the muscovite sandstone facies of the Magura Formation distinguished in the Polish part of the Magura Unit. The Kýčera sedimentary succession in the Quarry is a representative of the entire belt of the Magura Unit, along which the youngest synorogenic flysch deposits are more or less similarly developed. These deposits mark the last phase of the Magura Basin evolution. Rocks making up the sedimentary succession in the quarry form a sandstone-mudstone sequence, although with distinctive domination of the first lithology (Fig. 34). The sandstones occur mostly in thin to thick layers, very thick beds are also quite often present, some of them shows amalgamation structures. Most sandstone beds have flat or slightly wavy bases, very often with flute casts or tool marks.

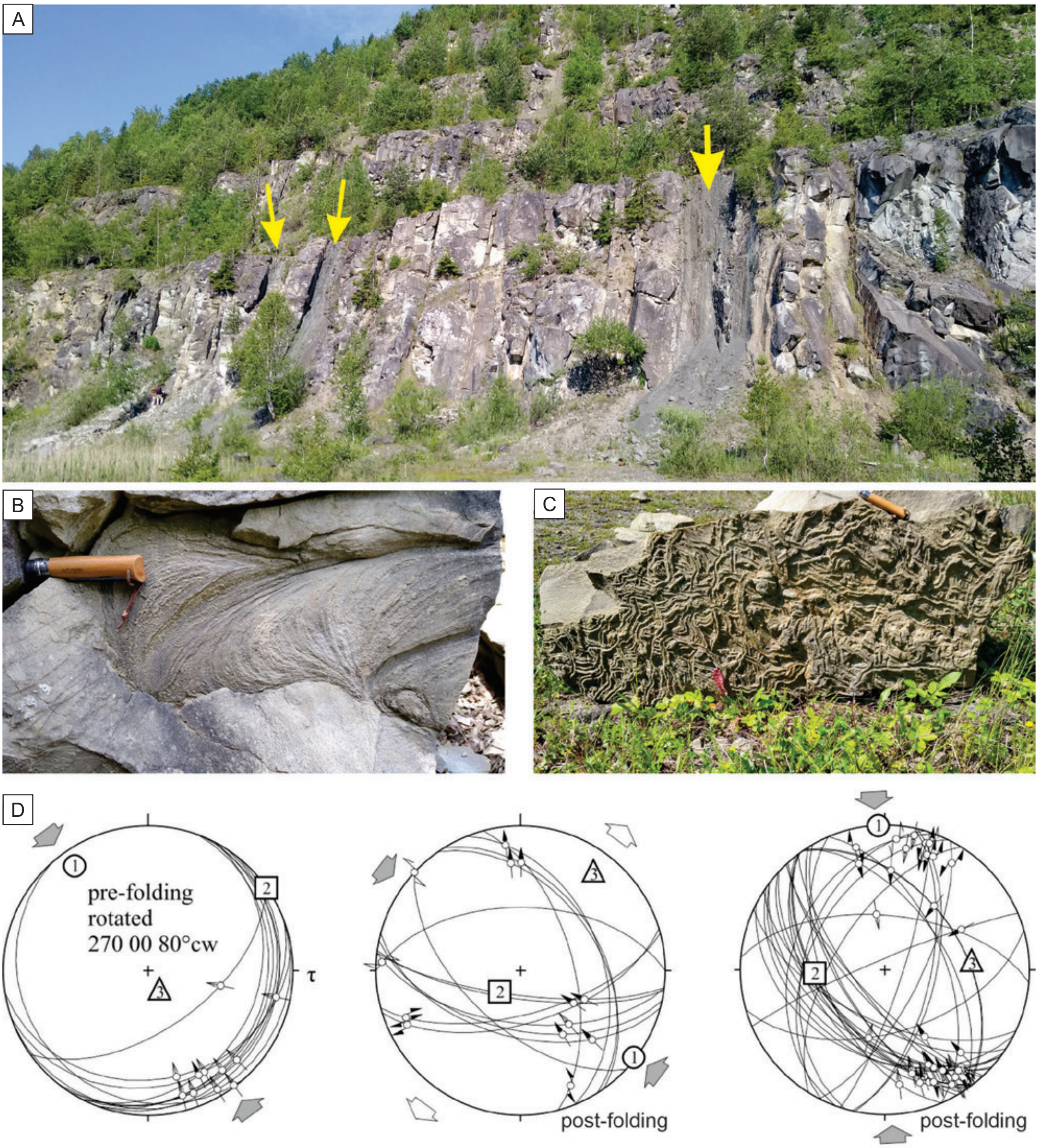


Fig. 34. General view of the Kýčera Member sequence in the Klubina quarry: A – bottom level of the quarry showing a sequence of turbidite deposits dominated by sandstone layers, thicker intervals of mudstone with very thin sandstone intercalations are marked with yellow arrows; B – *Zoophycos* trace fossil within the sandstone bed; C – sinusoidal trace fossils of *Scolicia* on the bottom surface of sandstone layer; D – Schmidt's lower hemisphere stereonet plots for faults measured in the quarry

They are characterized by massive structure or normal grading as well as parallel lamination. The Tb-d divisions of the Bouma sequence are visible within the beds, but more often they show only normal grading interval at the base and a thin laminated mudstone at the top. Trace fossils are quite abundant on layer surfaces or inside them, mostly *Zoophycos*, *Scolicia*, *Taphrhelminthopsis* can be encountered (Fig. 34B, C). Mudstone occur as thin interbreedings between sandstone layers or sometimes they form thicker intervals (Fig. 34). The proportion of mudstones and sandstones varies both vertically and laterally, thicker shale intervals occur in upper levels of the quarry. The Kýčera Member represents a depositional environment of the submarine turbiditic fan and is interpreted either as a channel system (or upper parts of lobes) alternating with interchannel areas (Starek & Pivko, 2001) or lobe system alternating with interlobe deposits (Staňová & Soták, 2007).

Performed structural observations in the Klubina quarry evidence of two reverse faulting phases (DF1 and DF2) with axes of maximum compression directed to NW (DF1) and N (DF2). These findings correspond with the results of Beidinger and Decker (2016), who described similar stress regimes in the adjacent Czech Outer Carpathians for the Oligocene – Miocene times (Fig. 34D). The rotation of the axis of maximum compression is interpreted as an effect of anticlockwise rotation of this segment of the Outer Carpathians.

## Stop 8 – Wżar Mount (Miocene andesites and panoramic view) (Figs 12, 35, 36)

(*Jan Golonka*, *Michał Krobicki*)

The most famous outcrop (artificial one – abandoned quarry) of the Middle Miocene volcanism of the Pieniny Mountains occur on the Wżar Mount, near Snozka pass, and is represented by two generation of intrusive dykes and sills. In half of the 20th 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-Palaeogene rocks of the PKB and Palaeogene flysch of the Magura Nappe of the Outer Flysch Carpathians. Andesites occur in the form of dykes and sills. At the Wżar Mount 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 Mount represents the westernmost occurrence of andesites in the Pieniny region. Amphibole-augite and/or augite-amphibole andesites dominate in the Mount 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).

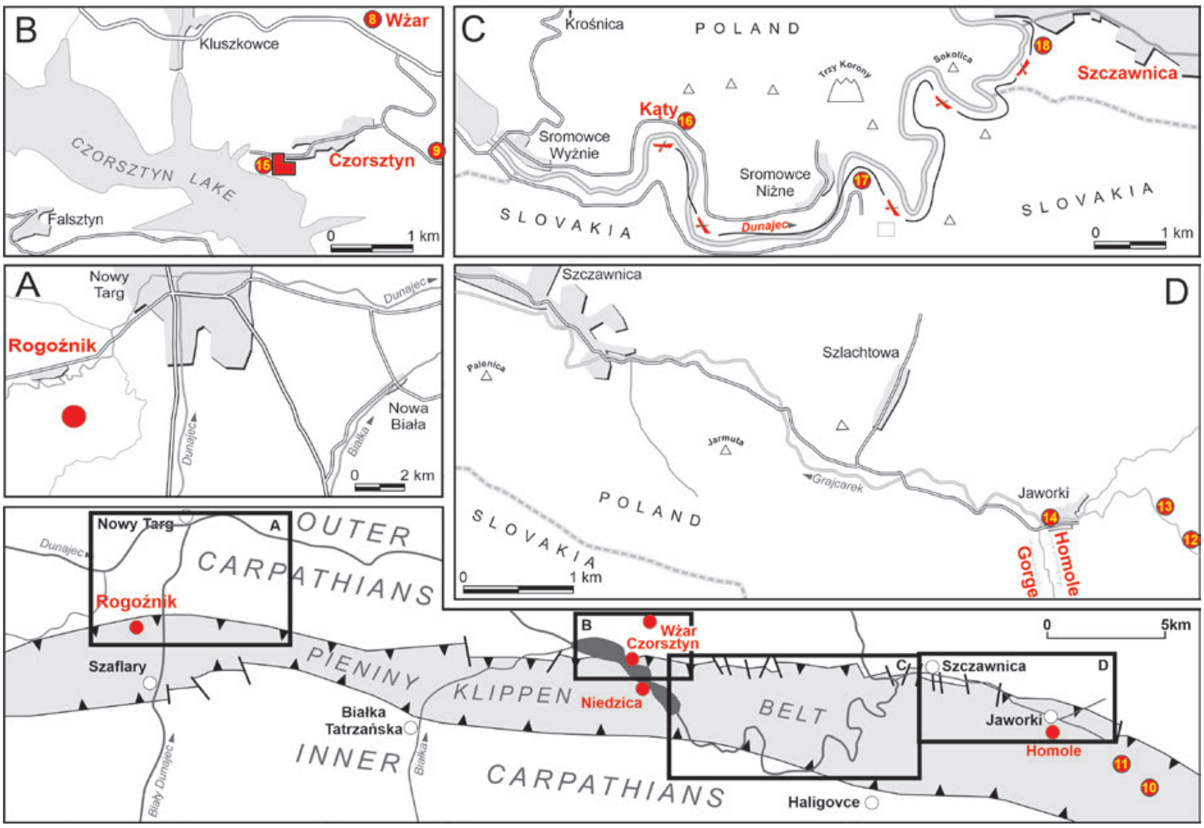


Fig. 35. Polish part of the Pieniny Klippen Belt and locations of visited outcrops – stop points

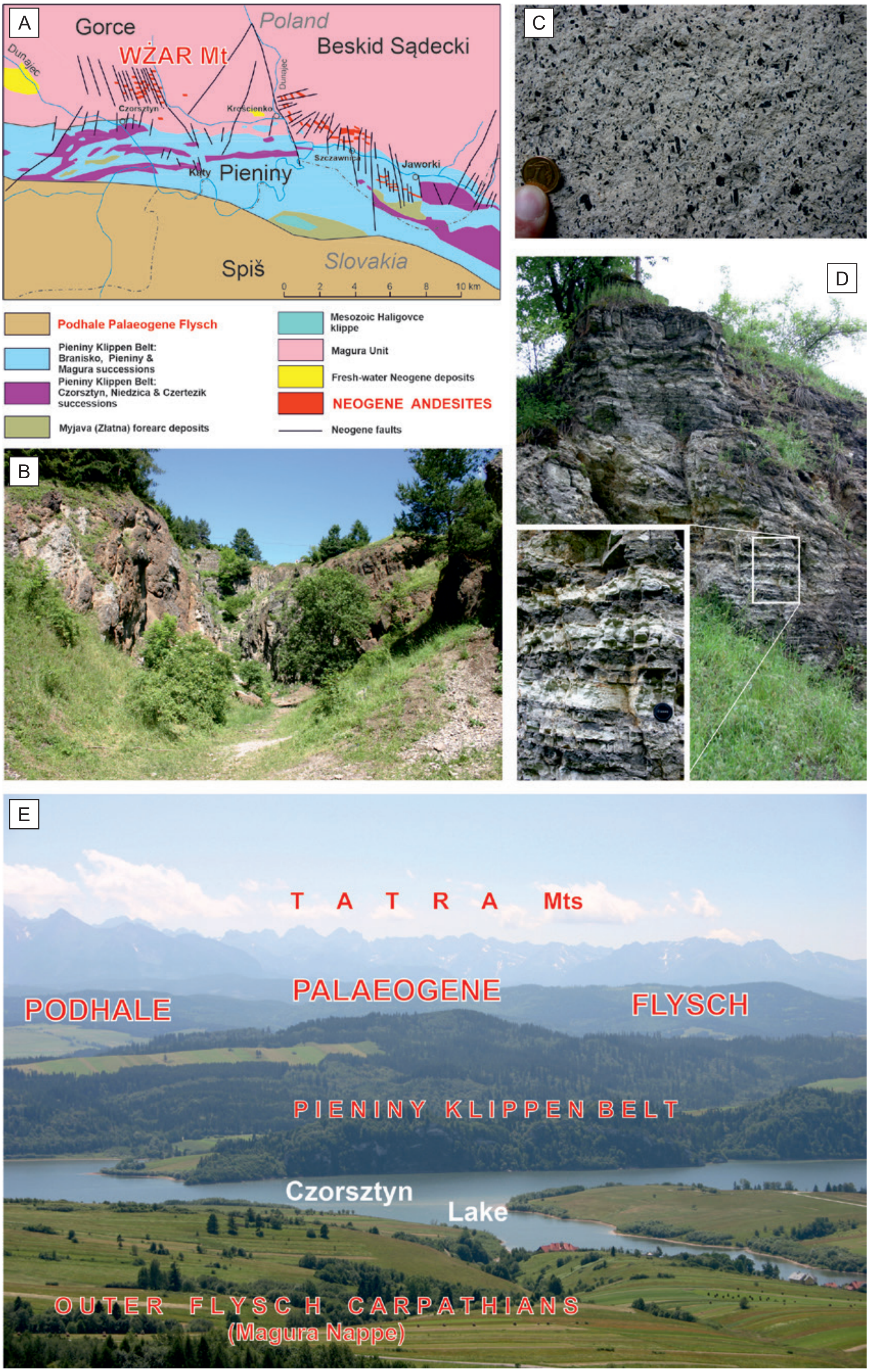


Fig. 36. Geological position of Miocene andesites of the so-called Pieniny Andesite Line: A – geological sketch of the Pieniny Klippen Belt (Polish sector) and surrounding regions (after Birkenmajer, 1979; simplified) with location of Wżar Mount; B – main entrance to abandoned quarry; C – andesites with 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 Mount

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 Palaeogene 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 Mount 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 15th century (Małkowski, 1958). 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).

## Stop 9 – Flaki range – Jurassic deposits of the Branisko Succession (Figs 35, 37, 38)

(*Michał Krobicki*)

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.

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.

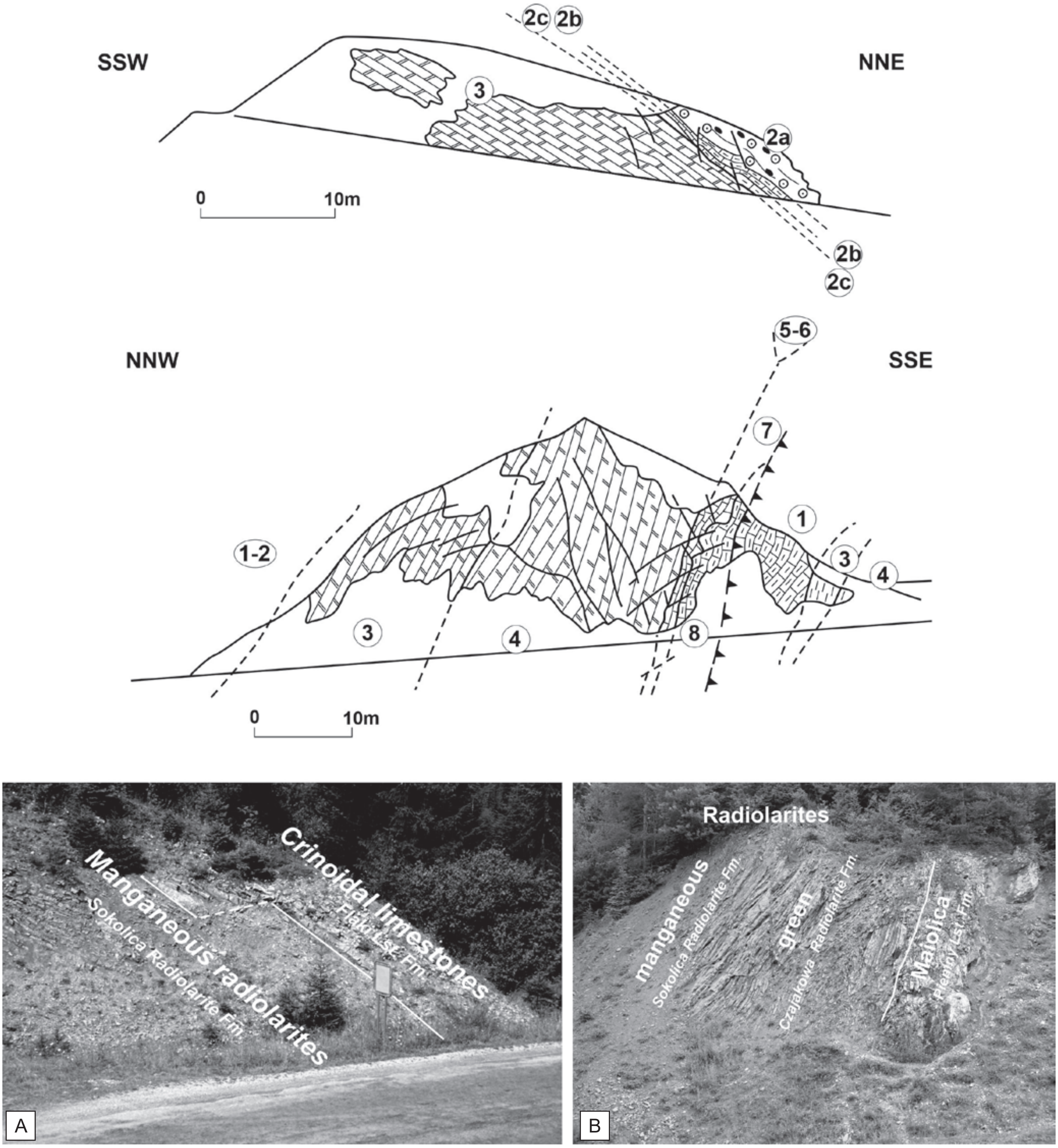


Fig. 37. View of the Flaki Range sections; Branisko Succession (lower part: A – western side; B – eastern side) and general sketch of studied sections (upper part) (after Birkenmajer *et al*., 1985). Lithostratigraphical units: 1 – Podzamcze Limestone Fm.; 2 – Flaki Limestone formations (grey crinoidal 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)

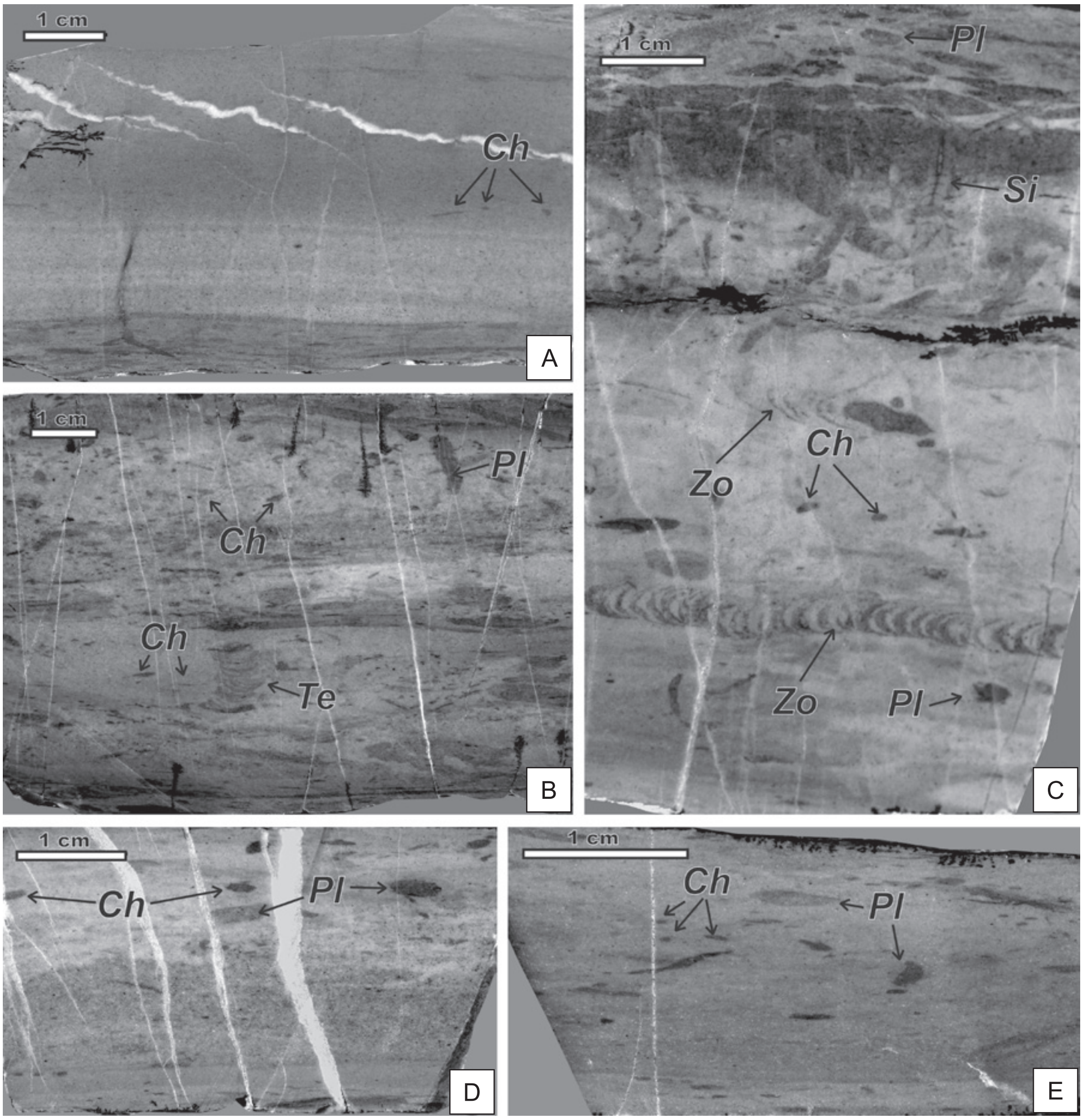


Fig. 38. Trace fossils within Sokolica Radiolarite Formation in the Flaki Range, Branisko Succession: Ch – *Chondrites*; Pl – *Planolites*; Si – *Siphonichnus*; Te – *Teichichnus*; Zo – *Zoophycos* (after Krobicki *et al*., 2006, 2023)

## Stop 10 – Litmanová village (Slovakia) – Jurassic-lowermost Cretaceous sequence of the Czertezik Succession of the Pieniny Klippen Belt (Figs 35, 39)

(*Andrzej Wierzbowski*, *Roman Aubrecht*, *Michał Krobicki*, *Bronisław Andrzej Matyja*, *Ján Schlögl*)

In the Litmanová village the best outcrop of the Czertezik Succession of the Pieniny Klippen Belt occurs in the eastern part of the Litmanovské Klippen. The oldest are black marly shales with discoidal spherosiderite concretions of the Skrzypny Shale Formation (see – Fig. 9). Younger deposits are grey coloured, fine-grained, well-bedded crinoidal limestones of the Smolegowa Limestone Formation which thickness is rather small, when usually has dozens meters. Near the base of this limetsones a thin condensed level marked by occurrence of small dark phosphatic nodules, abundant glauconite grains, and rich fauna of Early Bajocian ammonites (*Stephanoceras* (*Skirroceras*) sp.) and *Nonnolytoceras polyhelictum* (Boeckh), belemnites and brachiopods. The crinoidal limestone is a packstone with a marked admixture of detrital quartz grains. The topmost part of the bed is laminated – the laminae are wackestones and packstones of filamentous-crinoidal microfacies.

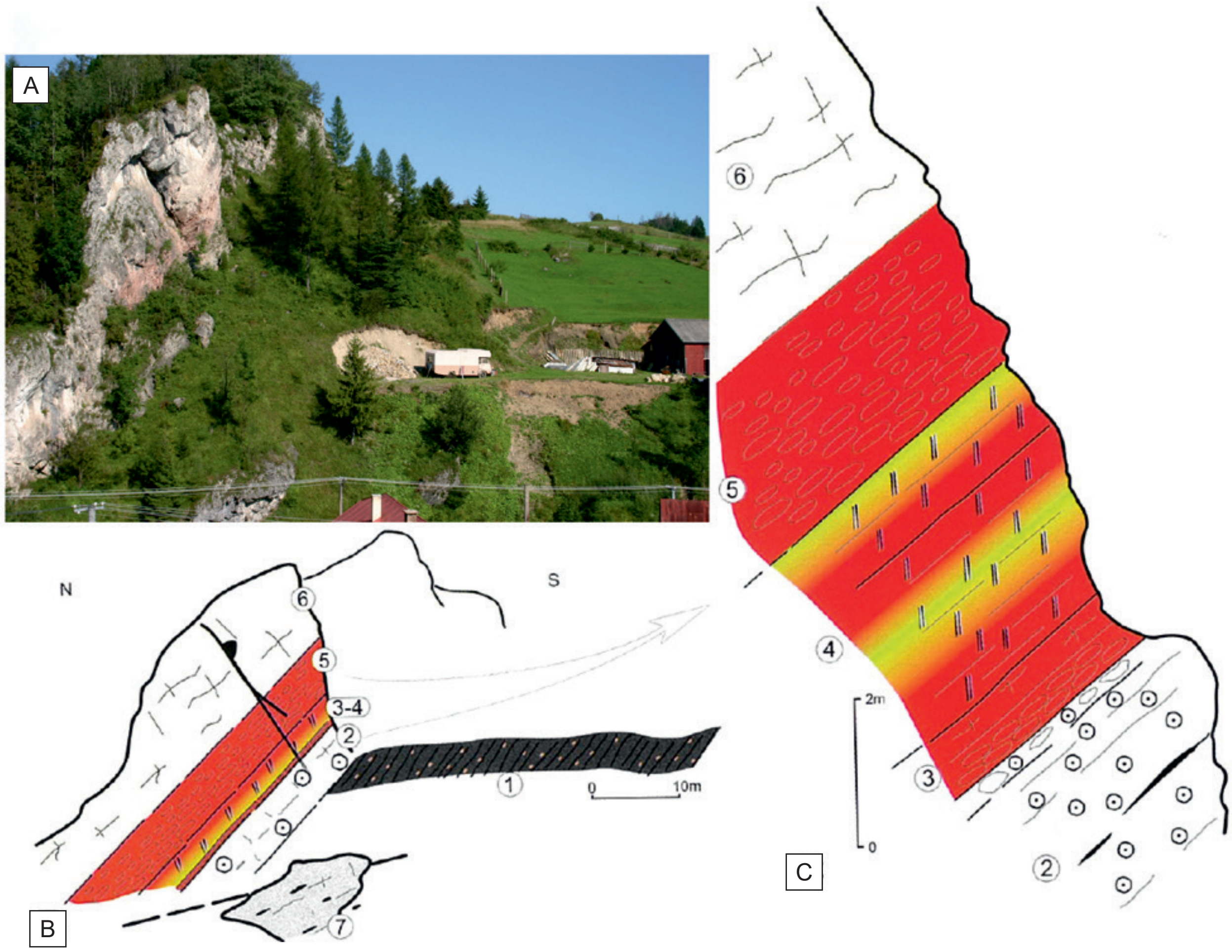


Fig. 39. Litmanovské Klippen section: A – general view of the klippen; B, C – geological cross-section sketches. Explanations: 1 – black shales with spherosiderites of the Skrzypny Shale Formation; 2 – grey crinoidal limestones of the Smolegowa Limestone Formation; 3 – *Ammonitico Rosso*-type cherry-red nodular limestones of the Niedzica Limestone Formation; 4 – green (lower part) and red (upper part) calcareous radiolarian cherts of the Czajakowa Radiolarite Formation (Podmajerz and Buwałd Radiolarite members, respectively); 5 – *Ammonitico Rosso*-type red nodular limestones of the Czorsztyn Limestone Formation; 6 – *Calpionella*-type white and cream micritic limestones of the Dursztyn Limestones Formation (Sobótka Limestone Member); 7 – *Maiolica*-type grey micritic limestones with cherts of the Pieniny Limestone Formation (after Wierzbowski et al., 2004)

The upper boundary of the crinoidal limestone unit is placed at the base of overlying of cherry-red coloured nodular limestone with scattered small crinoid fragments. The nodular limestone (Niedzica Limestone Formation) is wackestone to packstone rich in filaments. Overlying deposits are calcareous radiolarian cherts alternating with marly shales. The deposits are green-coloured in their lower part, and red-coloured in upper part and they correspond to the Czajakowa Radiolarite Formation (Ožvoldová *et al*., 2000; Wierzbowski *et al*., 2004). The younger deposits in the section are red nodular limestones of the Czorsztyn Limestone Formation of the *Ammonitico Rosso* facies which are wackestones, and packstones with abundant *Saccocoma* fragments (*Saccocoma* microfacies), and the youngest are massive, white-creamy *Calpionella*-bearing limestones of the Dursztyn Limestone Formation.

## Stop 11 – Biała Woda valley (near Brysztan Klippe) and Berriasian phosphatic structures (Figs 35, 40, 41)

(*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.

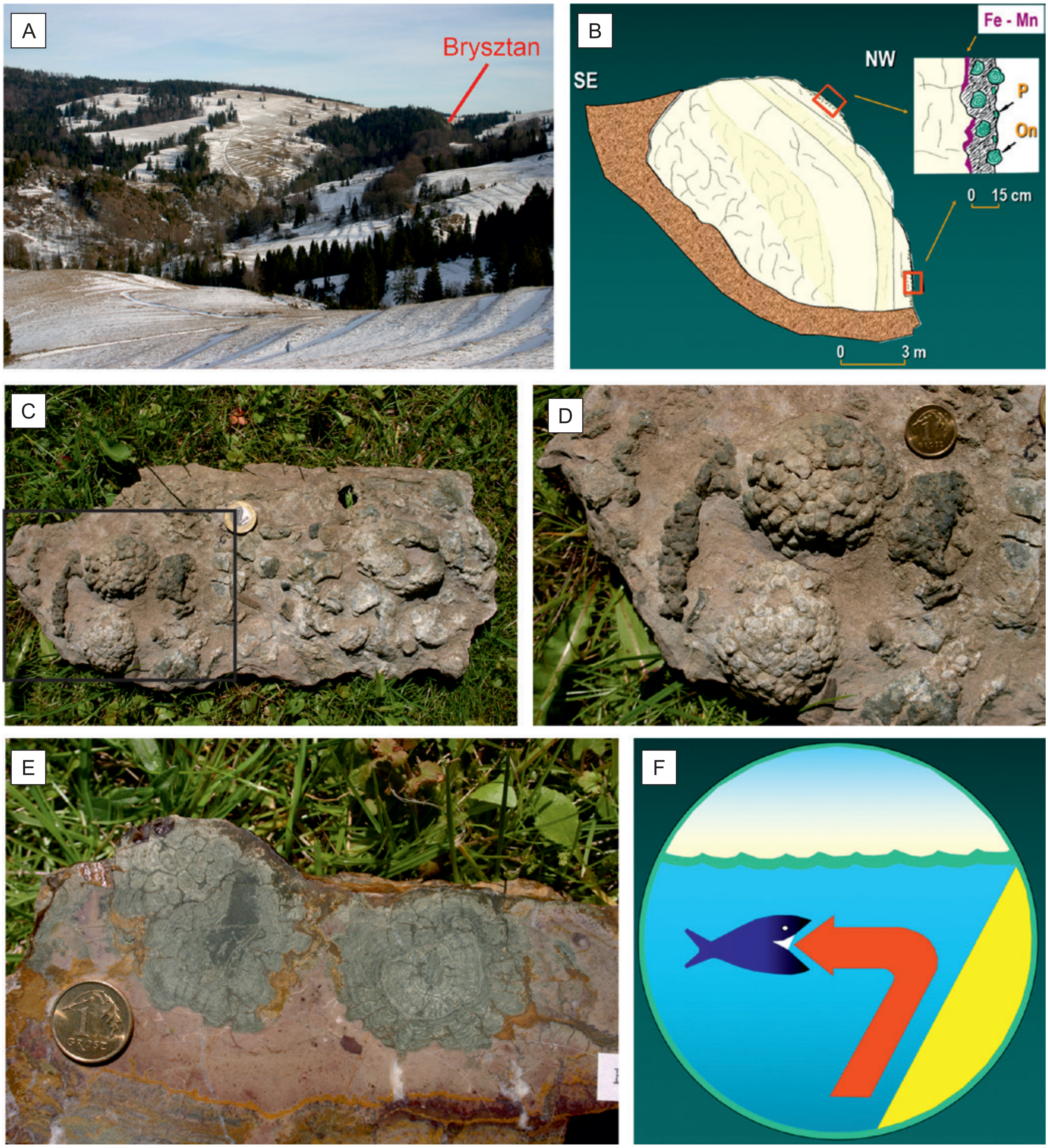


Fig. 40. Berriasian phosphatic structures in the Niedzica Succession: A – Location of the Biała Woda-Brysztan section; B – outcrop of the Dursztyn Limestone Formation, Sobótka Limestone Member (Berriasian), Niedzica Succession; in enlargement: top of last bed (after Krobicki, 1993, 1994): Fe-Mn – Fe-Mn crusts, P – phosphorites, On – phosphatic macrooncoids; C–E – phosphatic macrooncoids; F – upwelling logo (after Krobicki, 2023)

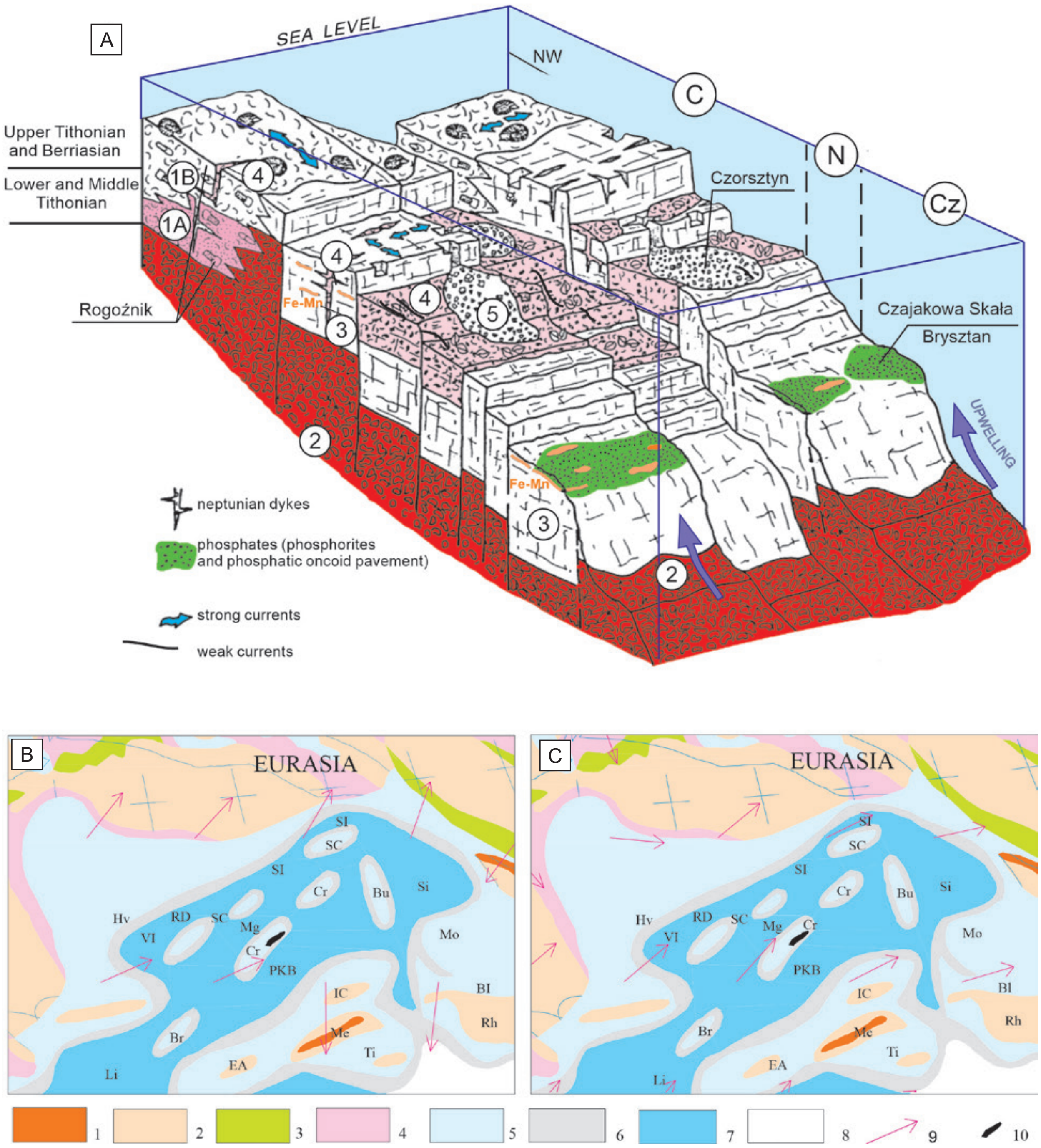


Fig. 41. The earliest Cretaceous block diagram of part of the Czorsztyn Ridge: A – Model of sedimentation on the intraoceanic Czorsztyn pelagic swell in Berriasian with effects of pronounced Neo-Cimmerian tectonic movements and upwelling currents (after Krobicki, 1996, modified): 1 – Rogoźnik Coquina 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; B, C – Palaeoenvironments, wind direction and upwelling zones of the Carpathian area during Tithonian-Berriasian time (palaeogeography after Golonka *et al*., 2000, modified): C – summer Northern Hemisphere, B – winter Northern Hemisphere: 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 (after Krobicki, 2023)

East of Jaworki (Fig. 35) 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. 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. 40). 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. 41) 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, 41).

## Stop 12 – Biała Woda valley (waterfall) – Berriasian crinoidal limestones, synsedimentary breccia and Valanginian crinoidal limestones (Figs 35, 42)

(*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. 42). 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. 42A; comp. Birkenmajer, 1977). The overlying beds numbered 1–4 in section A, and 3–4 in section B (Fig. 42), 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.

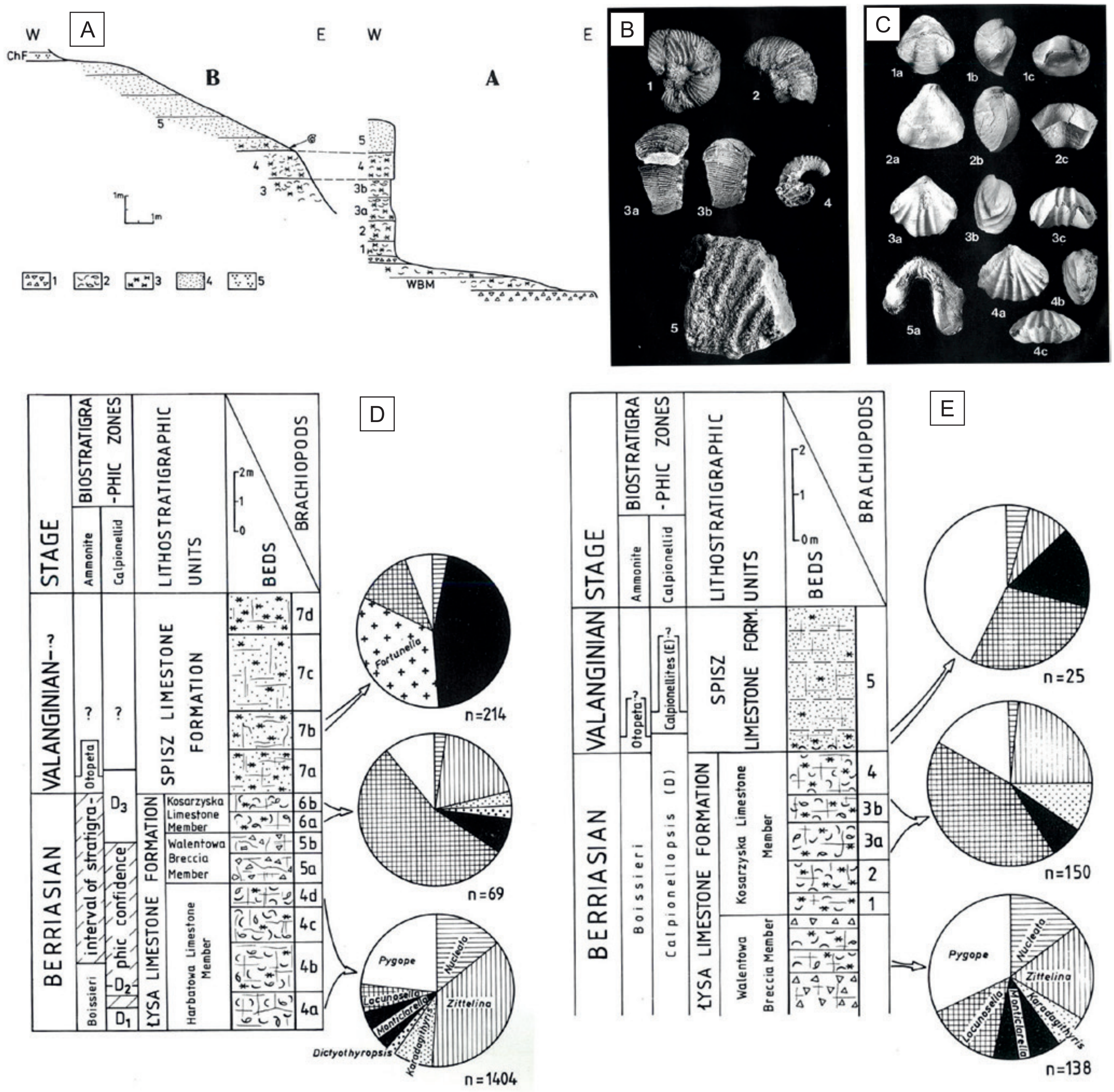

Fig. 42. Berriasian crinoidal limestones and synsedimentary breccia and Valanginian crinoidal limestones: A – Geological sections in the Biała Woda Valley at waterfall (after Krobicki & Wierzbowski, 1996). Lithostratigraphy: Łysa Limestone Formation (WBM – Walentowa Breccia Member, beds 1–4 Kosarzyska 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 (E – Czorsztyn-Sobótka Klippe and F – Biała Woda Valley). Lithostratygraphic 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) (after Krobicki, 2023)

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.

## Stop 13 – Biała Woda valley – "mid"-Cretaceous basaltic olistolith (Figs 35, 43, 44)

(*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–Palaeocene) 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. 44).

Fig. 43. View of basaltic olistolith in Biała Woda Valley (after Krobicki, 2023)

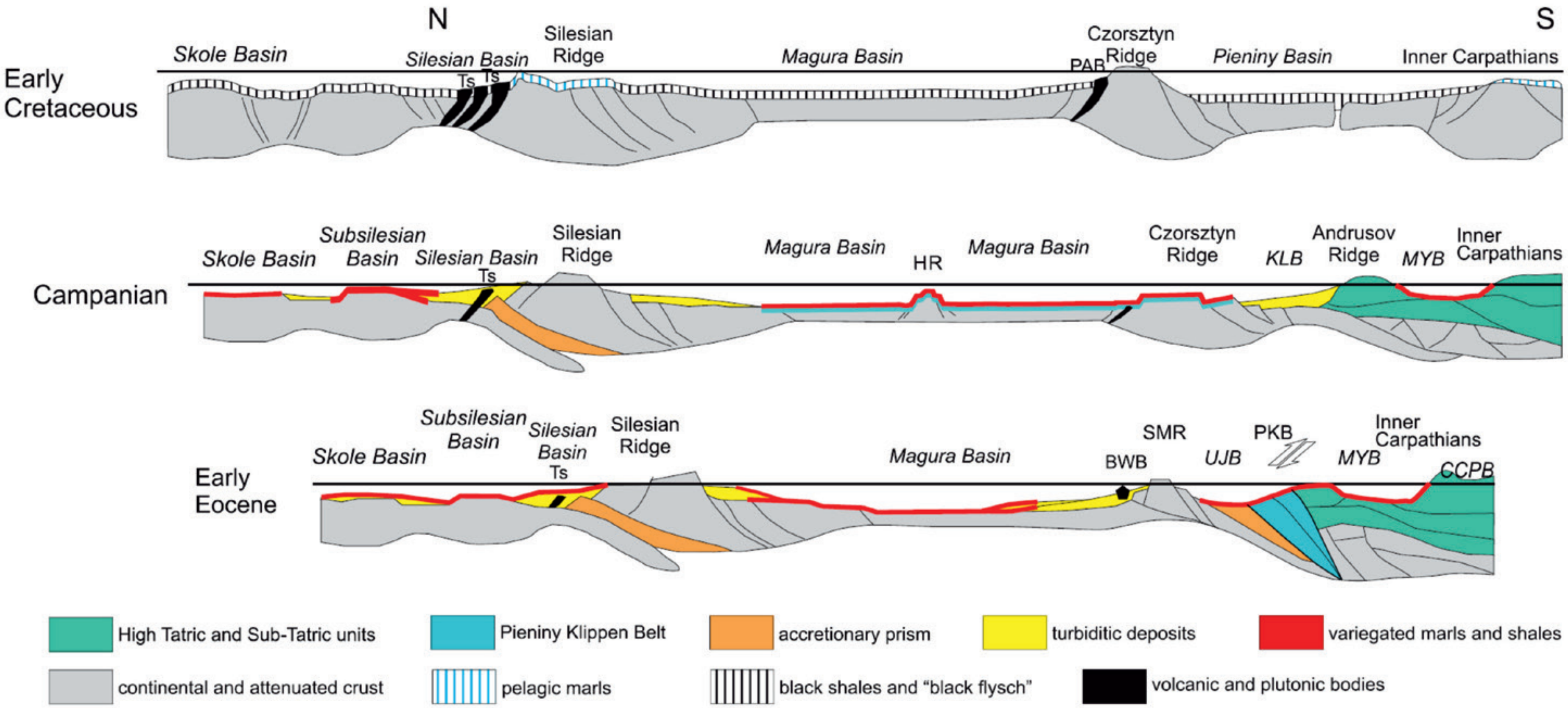


Fig. 44. Early Cretaceous–Early Eocene palinspastic evolutionary model for the Magura Basin, not to scale (based on Oszczypko, 2006, modified and Oszczypko *et al*., 2012). CCPB – Central Carpathian Paleogene Basin, HR – Hluk Ridge, UJB – Ujak Basin, KLB – Klapa Basin, MYB – Myjava Basin, PKB – Pieniny Klippen Belt, SMR – South Magura Ridge, Ts – teschenites, PAB – Pieniny alkaline basalts , BWB – Biała Woda basaltic block (after Krobicki, 2023)

## Stop 14 – Jaworki village, below church – Maastrichtian/Paleocene sedimentary breccia of the Jarmuta Formation (Figs 11, 35, 45)

(*Michał Krobicki*)

At Jaworki village (below church), along Grajcarek Stream, we can observe tectonic contact of the Czorsztyn Succession (Unit) and Grajcarek Unit (Magura Succession). The most interesting object here is Paleocene synsedimentary breccia of the Jarmuta Formation originated as post-Laramian sedimentation of the Mesozoic klippen fragments, presumable as effect of erosion of napped stractures – the Niedzica Unit over Czorsztyn Unit in the so-called Homole block on south (Fig. 45). Polimictic character of breccia reflects both erosion of the Czorsztyn Unit (Succession) elements and overlying tectonically the Niedzica Unit (Succession). Fragments are represented by clasts: white and redish crinoidal limestones (Smolegowa/Krupianka Limestone Formation), red nodular limestone of the *Ammonitico Rosso* facies (Czorsztyn and/or Niedzica Limestone formations), multicoloured radiolarites (Czajakowa Radiolarite Formation), chert limestones of the *Maiolica* facies (Pieniny Limestone Formation), yellowish sandstones of the Sromowce Formation etc. Mixture of different units clasts support mentioned idea of napping event just before formation of Jarmuta Formation during Maastrichtian/Paleocene time (Laramian phase of the Alpine orogenic events) (Birkenmajer, 1986, 1988).

## Stop 15 – Czorsztyn Castle Jurassic-Cretaceous deposits of the Czorsztyn Succession (Figs 9, 35, 46, 47)

(*Michał Krobicki*, *Andrzej Wierzbowski*, *Anna Kwietniak*)

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 19th 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* (*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.

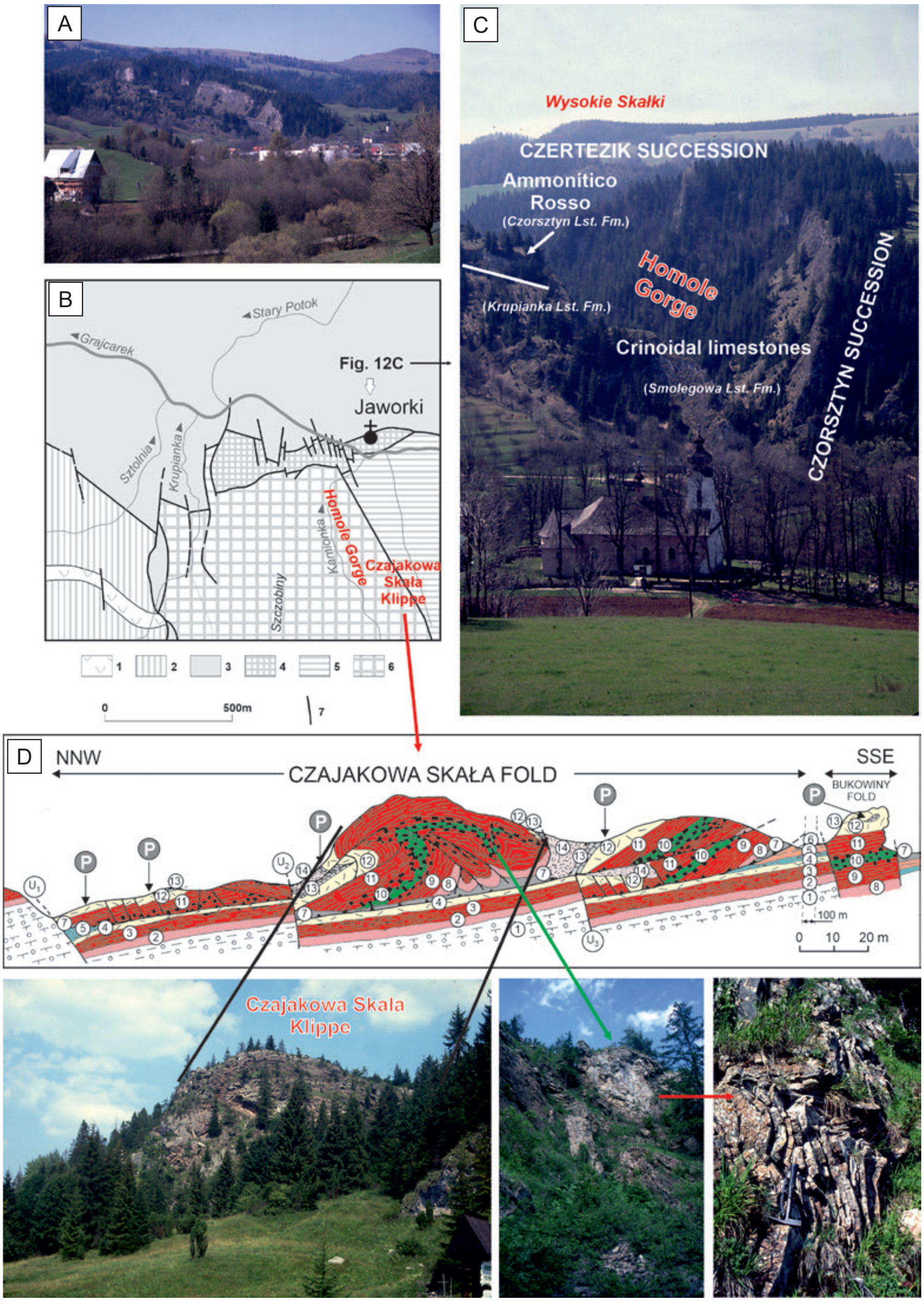


Fig. 45. General panoramic view (A) and geological sketch (B) of vicinity of Jaworki village with more detail view of entrance to Homole Gorge (C) and Czajakowa Skała Klippe and Bukowiny Fold (D) (geology after Birkenmajer, 1970; modified by Jurewicz, 1994). Explanations: B: tectonic sketch of the Homole block, northern part (after Birkenmajer, 1970, 1983); 1 – andesite intrusion (Middle Miocene: Sarmatian); 2 – autochthonous Magura-type Palaeogene; 3 – Grajcarek Unit (Magura Succession); 4 – Czorsztyn Unit (Czorsztyn Succession); 5 – Skalski Stream depression (Niedzica Nappe – Niedzica Succession); 6 – Homole block (Czorsztyn Unit with overthrust Niedzica and Branisko nappes); 7 – strike-slip faults; D: Czorsztyn Succession: 1 – Smolegowa Limestone Fm. (white crinoidal limestones); 2 – Krupianka Limestone Fm. (red crinoidal limestones); 3 – Czorsztyn Limestone Fm. (red nodular *Ammonitico Rosso*-type limestones); 4 – Dursztyn Limestone Fm. (pink and white *Callpionella* limestones); 5 – Pomiedznik Formation (marly limestones); 6 – Jaworki Formation (variegated marls); Niedzica Succession: 7 – Krempachy Marl and Skrzypny Shale formations (*Fleckenmergel*-type grey and black spotty marls/shales sometime with spherosiderite concretions – latter formation); 8 – Smolegowa Limestone Fm. (green-red crinoidal limestones); 9 – Niedzica Limestone Fm. (red nodular *Ammonitico Rosso*-type limestones); 10 – Czajakowa Radiolarite Fm. (red and green radiolarites); 11 – Czorsztyn Limestone Fm. (red nodular *Ammonitico Rosso*-type limestones); 12 – Dursztyn Limestone Formation (red and white *Callpionella* limestones); 13 – Pieniny Limestone Fm. (white and grey cherty *Maiolica*-type limestones); 14 – Kapuśnica Fm. (greenish spotty limestones). P – Location of phosphate deposits on the uppermost surface of the Sobótka Limestone Member of the Dursztyn Limestone Fm. (formal units after Birkenmajer, 1977) (after Krobicki & Golonka, 2008)

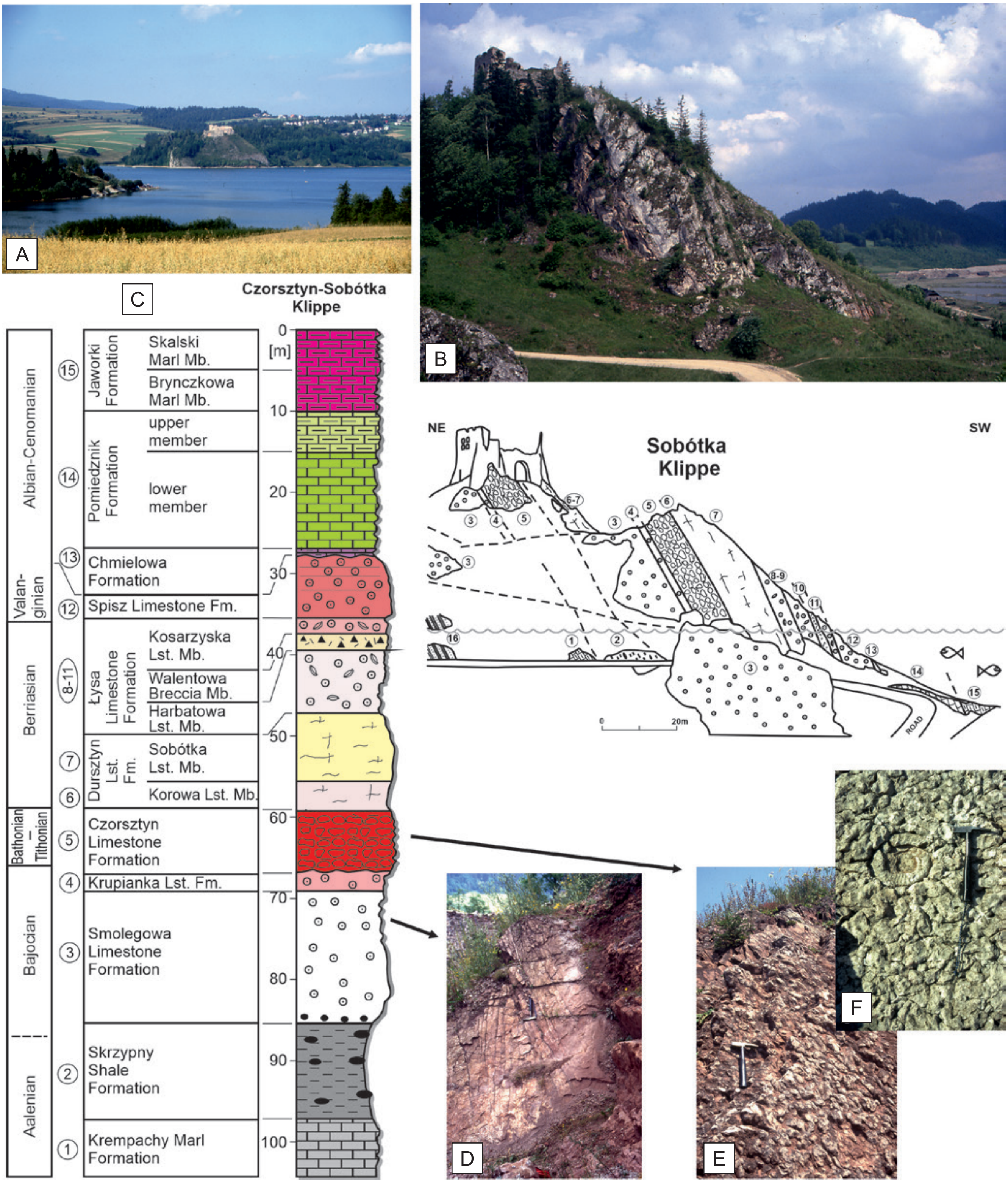


Fig. 46. Stratigraphical section of the Czorsztyn-Sobótka Klippe (A, B) with indication of position of the Walentowa Breccia Member of the Łysa Limestone Formation of the Czorsztyn Succession (C) (lithostratigraphy after Birkenmajer, 1977, slightly modified) (photo – state in 1992). Explanations of lithology: 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)

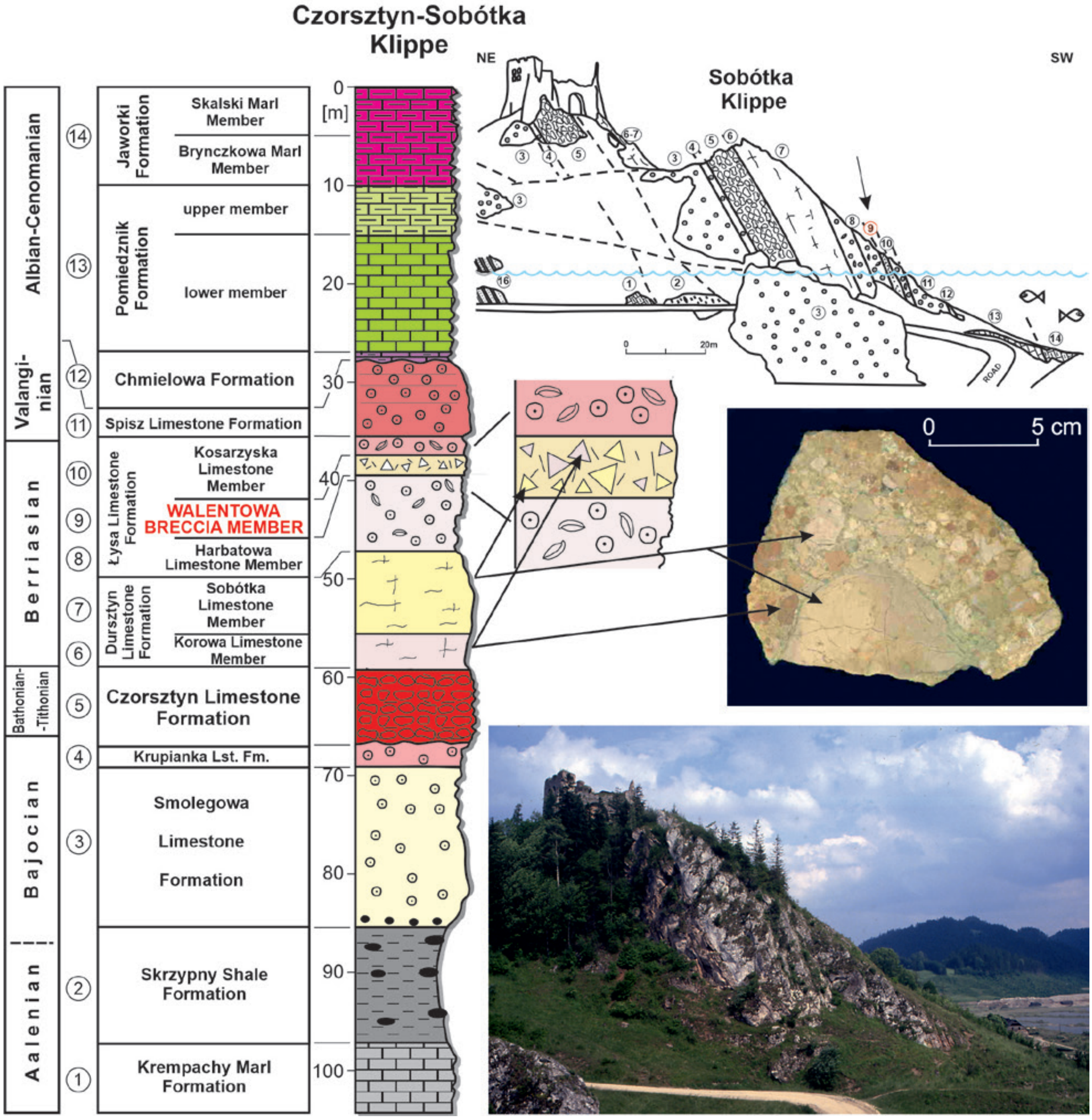


Fig. 47. Stratigraphical section of the Czorsztyn-Sobótka Klippe with indication of position of the Walentowa Breccia Member of the Łysa Limestone Formation of the Czorsztyn Succession (lithostratigraphy after Birkenmajer, 1977, slightly modiefied) (photo – state in 1992). Explanations of lithology as on Fig. 46.

The area of the Czorsztyn Lake is specific from the seismological point of view. As for the seismological background of Poland, which is considered to be a country of low seismicity, for which the return periods of earthquakes are long, the area of Podhale-Nowy Targ is the most active (Guterch, 2009). The Orawa-Nowy Targ Basin in the Carpathians is Poland's best seismologically recognized region, for which there is quite a good record of historical and instrumental seismicity. Most recently, a series of seismic events occurred on November 30, 2004, with the main earthquake reaching a considerable moment magnitude (Mw) of 4.4 with relatively high epicentral intensity, estimated to be at level 7 (EMS-98 scale) (Guterch, 2009).

This natural seismicity overlaps with anthropogenic seismicity (non-tectonic events), which is connected mainly with the existence of the water reservoir – the Czorsztyn Lake opened in 1997. As water reservoirs are often linked to increasing seismic activity (Wang & Manga, 2021), the Czorsztyn Lake is not an exception in this matter mainly swarm-like events were reported in the past (Białoń *et al*., 2015; Golonka *et al*., 2019).

The region is also a subject of vivid development of geothermal sites, where, for instance, in 2023–2024, the deepest geothermal well, designed to be 7 km, in Szaflary is currently being drilled (which has reached over 5 km (www1)). The factors mentioned above justify the need for seismological monitoring in the region, which will be further discussed in a short stop during the field trip.

## Stop 16 – Sromowce, Macelowa – Upper Cretaceous *Scaglia Rossa* with clastics (Figs 11, 35, 48, 49)

(*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 Mount Macelowa, where the Pieniny Succession rocks lie in an overturned position. The oldest Oxfordian radiolarites occupy the topmost part of Mount 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 17 – Sromowce-Szczawnica, Dunajec River rafting – uppermost Jurassic/Lower Cretaceous *Maiolica* limestones (Figs 9, 11, 35, 48)

(*Jan Golonka*, *Michał Krobicki*)

The rafting through the Dunajec River Gorge belongs to the major geotouristic attractions of Poland. It offers magnificent views of the cliffs sculptured in the Pieniny Mountains both by the tectonic activity and river erosion. It offers also the close view of outcrops of Jurassic and Cretaceous rocks as well as an insight in to the complex tectonics of the Pieniny Klippen Belt. Therefore, it serves as the site of the fieldtrips for the international conferences held in southern Poland. Additionally, in near future the Dunajec River Gorge could be the main object of trans-bordering PIENINY Geopark (Golonka & Krobicki, 2007). The trips on the Dunajec River have started in 19th century and are still very popular nowadays. The boats are run by licensed local guides living in the Pieniny Mountains villages and towns: Sromowce Wyżne and Niżne, Czorsztyn, Szczawnica and Krościenko. Originally, the boat starting point was in Czorsztyn, just beneath the castle hill. After the Dunajec River dams at Czorsztyn and Niedzica were constructed, the new harbor was built in Sromowce Kąty. Each Dunajec River boat carries 10 tourists and the pilot-guide. The boats are specially design in order to enable the easy maneuvering on the treacherous whitewaters of Dunajec. The route partly follows the state border of Poland and Slovakia. Slovaks also offer their own rafting trips although somewhat shorter route.

Present day confines of the Pieniny Klippen Belt are strictly tectonic reflecting its Paleogene-Neogene evolution when (sub)vertical faults and shear zones developed and a strong reduction of space of the original sedimentary basins took place. The NE–SW striking faults accompanying the Klippen Belt 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 Klippen Belt, or by the structural asymmetry of the Inner Carpathian Paleogene Basin. The tectonic character of the Polish section of the PKB is mixed. Both the strike slip and thrust components occur here (e.g., Książkiewicz, 1977; Golonka & Rączkowski, 1984; Birkenmajer, 1986; Nemčok & Nemčok, 1994; Jurewicz, 1997, 2005; Golonka *et al.*, 2005).

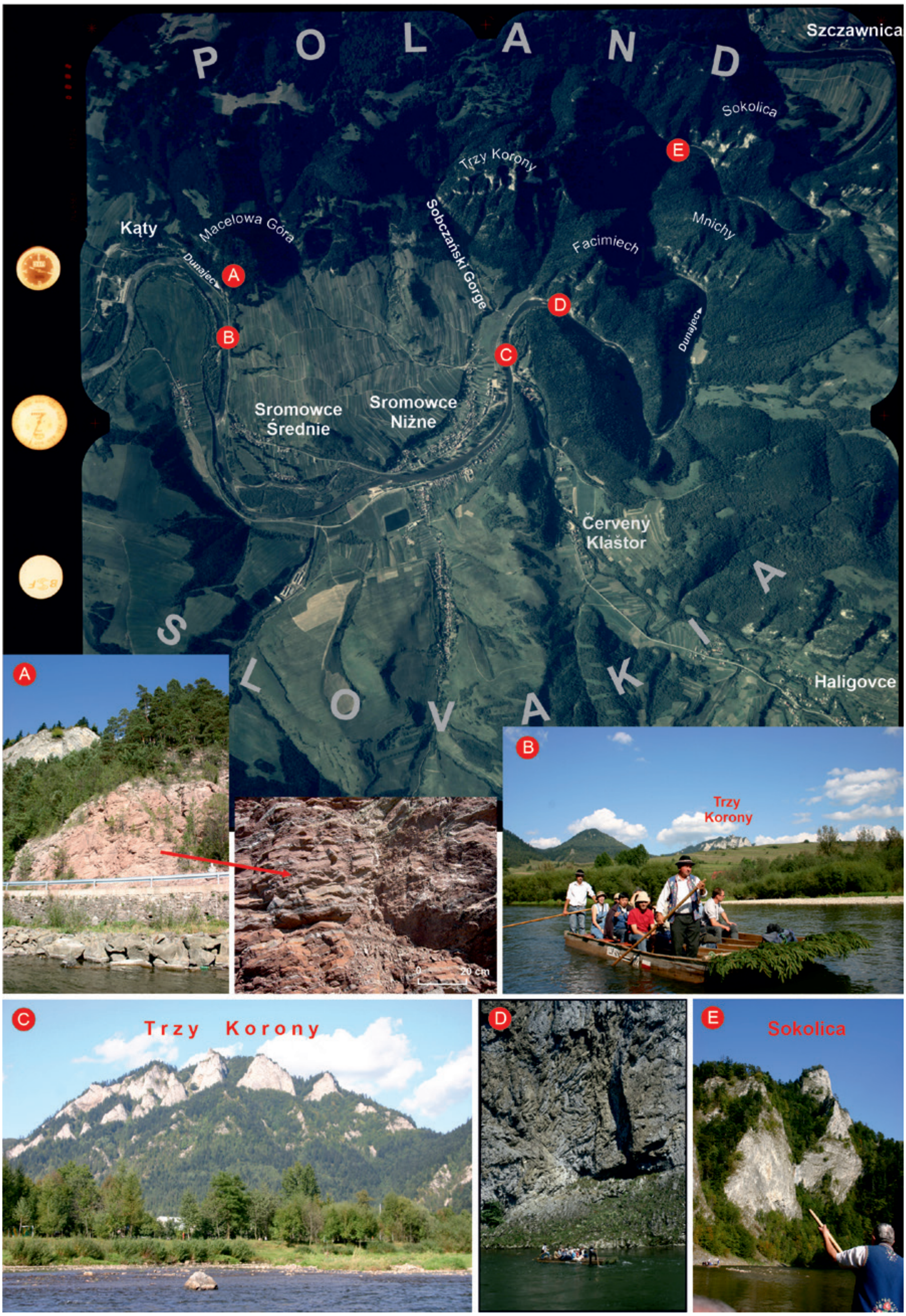


Fig. 48. Aerial view of the central Pieniny Mountains and Dunajec River Gorge with points of photos: A – Upper Cretaceous red marls of the *Scaglia Rossa*-type facies of the Macelowa Marl Member of the Jaworki Formation (Macelowa Mount); B – close to beginning of the rafting in Sromowce-Kąty harbor in the Pieniny Mts., boat full of tourists; C – Trzy Korony Mountain built of *Maiolica*-type well-bedded cherty limestones of the Pieniny Limestone Formation, usually strongly tectonically folded (D); E – Sokolica Mt. over the Dunajec River Gorge (after Krobicki & Golonka, 2008 and Krobicki, 2023)

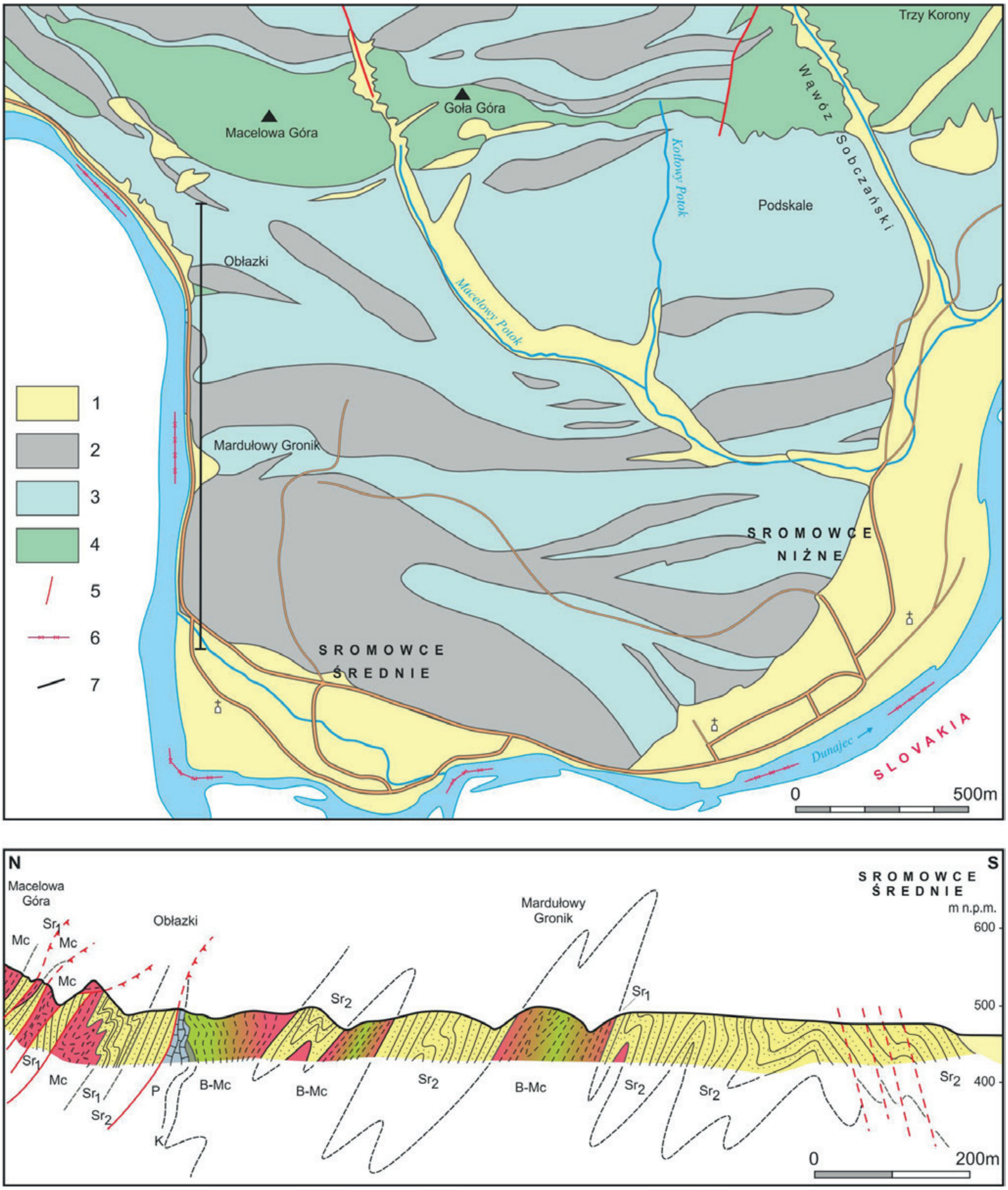


Fig. 49. Geological map of the vicinity of Sromowce (after Horwitz, 1963, Birkenmajer & Jednorowska, 1984, simplified) and geological cross-section: 1 – Quaternary; 2 – Sromowce Formation; 3 – Jaworki Formation, partly Kapuśnica Formation; 4 – Pieniny Limestone Formation, partly also Czajakowa Radiolarite 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 and Krobicki, 2023)

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 origin of the Gorge is related to the neotectonic movements during the Neogene time. Following the Serravalian formation of the Outer Carpathian fold-and-thrust belt, the plate boundary was covered during the Neogene by at least 600–900 m of sand, silt and clay, which were deposited in the Orava–Nowy Targ Depression east of the Gorge (Chrustek & Golonka, 2005). The Dunajec River valley reached the mature stage during the latest Miocene–Pliocene time. This stage is indicated by numerous meandering bends of the river. The vertical uplift of the Pieniny Mountains followed the meandering stage of the Dunajec River. Faulting and uplifting played a tremendous role during the Neogene tectonic evolution. Dense and regular fault net is one of the characteristic features of the Carpathians. Brittle, mainly strike-slips faults combined with other dynamic tectonic boundaries allowed the propagation of individual, detached blocks to the realm of the future Carpathian region (Golonka *et al.*, 2006). At least some of the faults were still active during the Quaternary (Baumgart-Kotarba, 1996, 2001; Zuchiewicz *et al.*, 2002). The studies on the 1995 earthquake (Baumgart-Kotarba, 2001 and references therein) show the good agreement of focal model with the trends of vertical crustal movements. The recent vertical movements in the area are up to +0.5 mm per year (Vanko, 1988; Vass, 1998). During the fault-related uplift the Dunajec River cut through the competent, Jurassic–Early Cretaceous cherty limestones, forming the magnificent cliffs of the Gorge. Most recently description of age and origin of the Dunajec River Gorge with the review of structural and geomorphological features of the Pieniny Mountains was published by Birkenmajer (2006, 2017).

The boat trip launches from Sromowce Kąty village within the Pieniny Klippen Belt (Fig. 48). The harbor is easily accessible by car or bus. Many travel companies offer the rafting trips combined with the coach transportation. The couches bring tourist to the starting point and pick them up at the final destinations in Szczawnica or Krościenko.

## Stop 18 – Szczawnica (“Orlica”) – history of the discovery of nappes in the Carpathians (Figs 35, 50–52)

(*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.

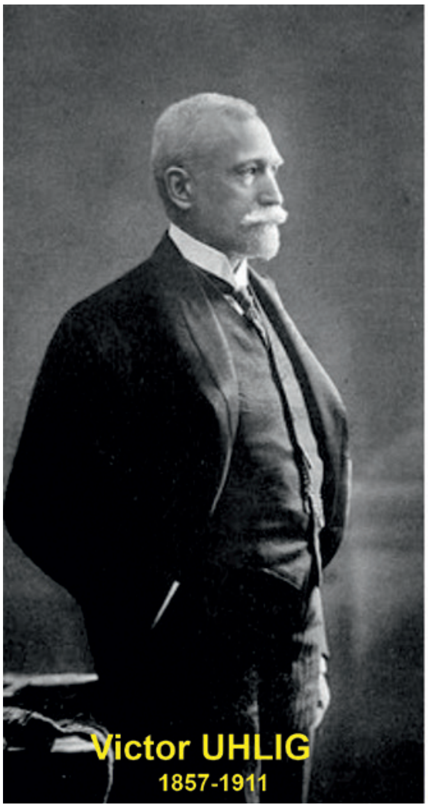


Fig. 50. Two prominent scientists – specialists of the Pieniny Klippen Belt geology (after Krobicki, 2023)

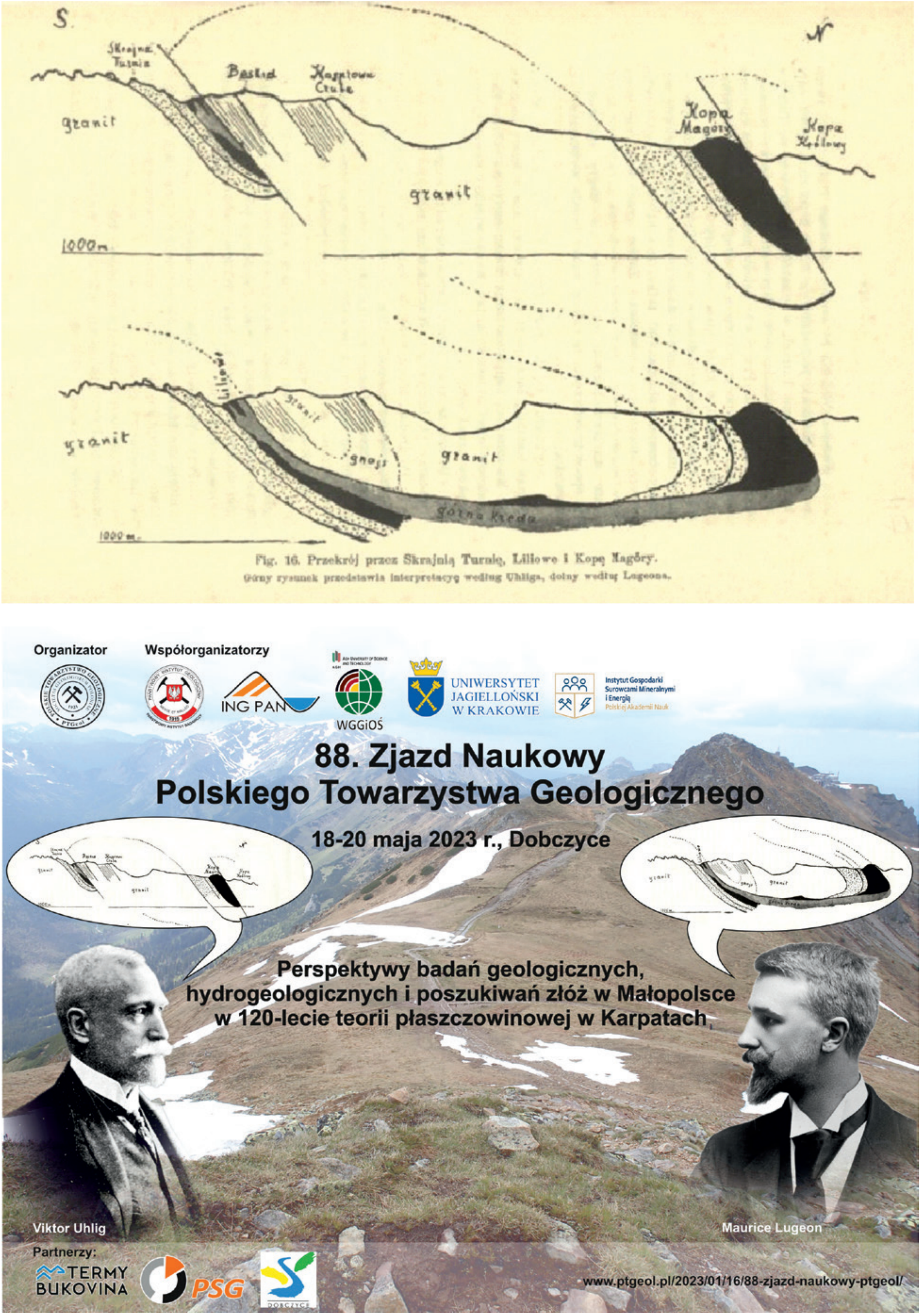


Fig. 51. Different interpretations of tectonic position of the Tatra Mts structures: upper cross section – Uhlig's idea and lower cross section – Lugeon's idea; lower position – circular of the 88th Polish Geological Society Meeting (after Krobicki, 2023)

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, 2022).

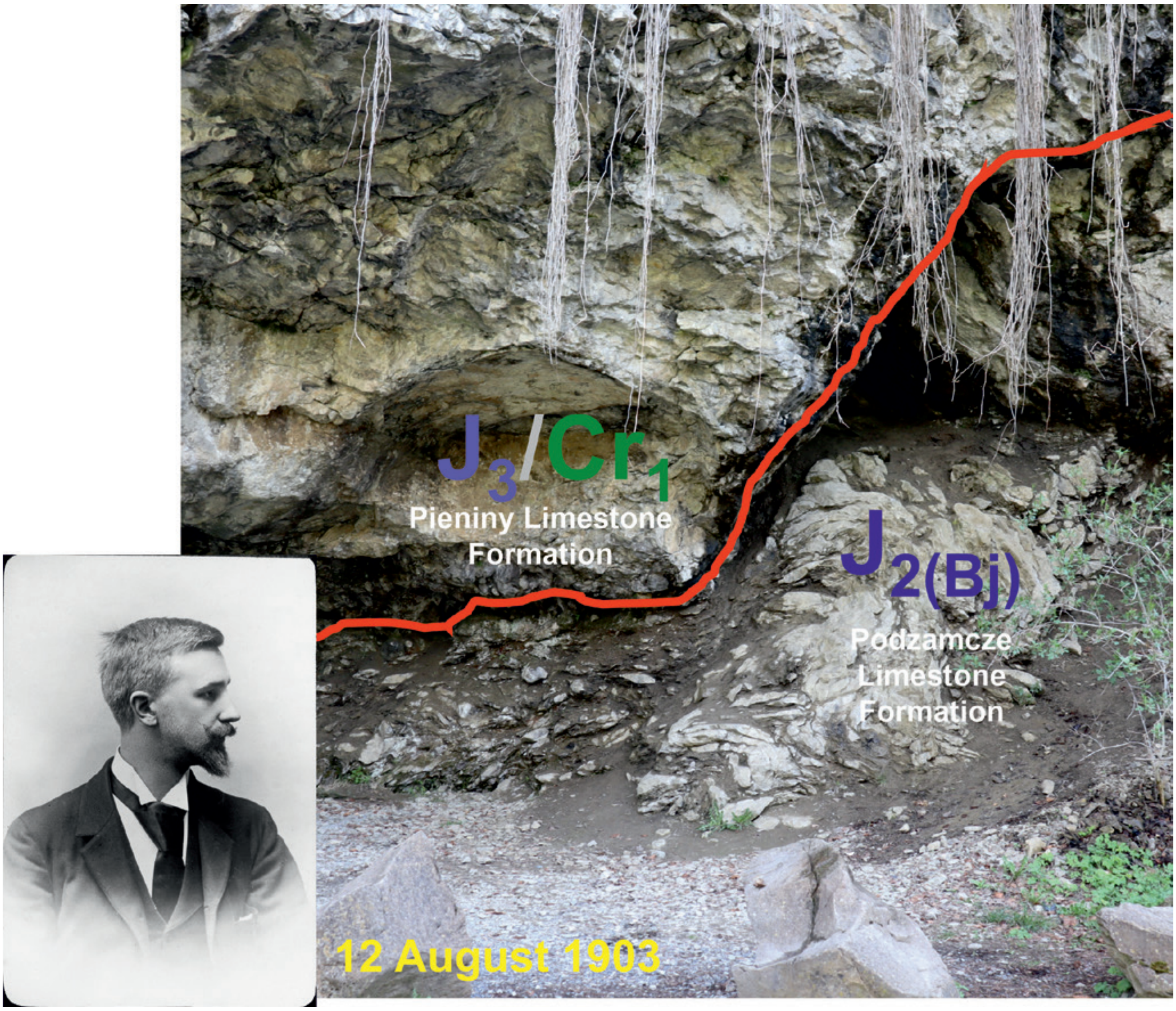


Fig. 52. Ziblikiewicz's cave in Szczawnica Niżna, along so-called Pieniny road nearby of Orlica hostel – tectonic contact between of nappe character; 120 anniversary of Maurice Lugeon visit of this place during International Geological Congress (Vienna'1903) with tectonic contact between two nappes?/thrust sheets? (after Krobicki, 2023)

## Tatra Mountains (Figs 53–57)

(*Michał Krobicki*, *Jan Golonka*)

The Inner Carpathian Paleozoic and Mesozoic rocks crop out in the Tatra Mountains. North of the Tatras, they are covered by the Central Carpathian Palaeogene (Fig. 53) and known only from boreholes and geophysical data (Golonka *et al*., 2005).

The Tatra Mountains are the highest mountain range of the Carpathians, located in their western part and drained by the Dunajec river with tributaries flowing to the Baltic Sea and Vah river with tributaries flowing to the Black Sea. The mountains extends 78 km along the Polish-Slovakian border. The total area of the range is 785 $km^2$, with 175 $km^2$ lying in Poland and the rest in Slovakia There are thirty-four summits with a prominence of at least 140 m in the range that reach over 2,000 m. Of these six reach 2,500 m. The highest peak in the range Gerlachovsky stit 2654.4 m, is in Slovakia.

The Tatra Mountains form a relatively high elevated asymmetric horst tilted northward, cut off from the south by a major Neogene-Quaternary normal fault and surrounded by sediments the Central Carpathian Palaeogene (Fig. 53). The crystalline core form central and southern part of the Tatra Mountains Western, northern and north eastern parts are covered by autochtonous Mesozoic rocks and several allochtonous thrust sheets and small nappes. All these units are discordantly covered with a post-nappe transgressive succession of the Central Carpathian Palaeogene Basin (Fig. 53). The crystalline basement consists of the granitoids and the metamorphic envelope (Figs 55, 56). Granitoids, represented by tonalities and granodiorites with subordinate amount of granites, originated during a continental collision between 360 Ma and 314 Ma (Poller *et al*., 2001). The metamorphic envelope, composed of two structural units is represented by migmatites, amphibolites, schists and gnessses.. Rocks of the Tatra Mountains were affected by Variscan (Palaeozoic and Alpine (Mesozoic-Cenozoic) tectonometamorphic events.

The Tatra Mountains Mesozoic sedimentary rocks were deposited within he Central Carpathian block which was bordered by Alpine Tethys to the North and Meliata Ocean to the South. The area between two oceans was divided into six paleogeographic domains, partially reflected by present-day tectonic units: Tatric, Fatric, Veporic, Gemeric, Hronic and Silicic (Andrusov *et al*., 1973). The Tatra Mountains rocks belongs to the Tatric, Fatric and Chronic domains. The Tatric domain is represented by the High-Tatric unit (Kotański, 1961), which includes the sedimentary cover of the crystalline core and the lower units of the overlying allochthon. The oldest, perhaps Permian conglomerates, crop out in a single locality (Koperšady; Fig. 57). The Lower Triassic is characterised by red bed sandstones, followed by mudstones, which are overlain by the Middle Triassic platform limestones and dolomites and Upper Triassic red beds of the Keuper type and coeval intertidal laminated dolomites, Rhaetian clastics, and shallow-marine fossiliferous limestones (Kotański, 1959, 1979). Lower Jurassic clastics and limestones occur in local troughs formed by block tectonics related to the rifting in the Western Tethys. The Middle Jurassic contains local crinoidal limestones, nodular limestones, commonly with stratigraphic gaps and condensations, stromatolites, and iron crusts, the Upper Jurassic pelagic limestones display locally nodular structures. Locally (Osobitá Mount), shallow-marine crinoidal limestones with volcanic rocks (limburgite) occur. Shallow-marine platform limestones of the Urgonian-type (Schrattenkalk) facies typify the Lower Cretaceous. The Lower-Middle Albian occurs only locally as condensed deposits with glauconite and phosphates, indicating drowning of carbonate platform. During the terminal basin development (Middle Albian-Cenomanian), marls with turbidites indicate a deepening of facies (Lefeld, 1985).

Fatric and Hronic domains are represented by the Krížna (Lower Sub-Tatric), Choč (Middle Sub-Tatric), and the (Strážov) (Upper Sub-Tatric) units occuring exclusively in thrust sheets, which overlie the High-Tatric units (Lefeld, 1999) (Fig. 54). The Krížna (Fatric) Triasic sequence contains the Lower Triassic red bed clastics, Middle Triassic platform dolomites, and the Carpathian Keuper and Rhaetian fossiliferous limestones (Kotański, 1959, 1979). The Jurassic facies are characterised by gradual deepening from shallow marine-clastics, through spotty limestones and marlstones (*Fleckenmergel* facies), spiculites and radiolarites, to nodular and *Maiolica* limestones. Basinal marlstones and limestones dominated in the Lower Cretaceous sediments in the western part of the Tatra Mountains, while in the eastern part (Belanské Tatry) massive carbonates occurred (Lefeld, 1985; Bac-Moszaszwili, 1993; Wieczorek, 2000).

The Choč units (Hronic) comprise typical Alpine Triassic facies, including the Hauptdolomit, the Rhaetian Kössen facies, and the Lower Jurassic encrinites, spiculites, and Hierlatz–type limestones (Grabowski, 1967; Kotański, 1973; Iwanow & Wieczorek, 1987; Uchman, 1993). The Upper Sub-Tatric units (also Hronic), represented only by two small thrust sheets, are typified by the basinal Middle Triassic Reifling Limestone, Partnach Marl, and the shallow-marine Upper Triassic Wetterstein facies. All of the allochthonous units were thrust northward in the Late Cretaceous (Kotański, 1986a, 1986b; Iwanow & Wieczorek, 1987; Jurewicz, 2002). The erosion of the Mesozoic units took place in the Late Cretaceous-Eocene times. A subsequent transgression took place in the Middle Eocene that resulted in the formation of conglomerates and limestones forming the basal member of the Podhale Palaeogene. Sediments of the calcareous Eocene are known from numerous natural exposures situated at the outlets of valleys draining the Tatra massif and from drillings made in the Podhale Basin (Golonka *et al.*, 2005). Directly on the transgressive deposits of the calcareous Eocene there occur stratigraphically younger strata of the Palaeogene, i.e. the Podhale flysch.

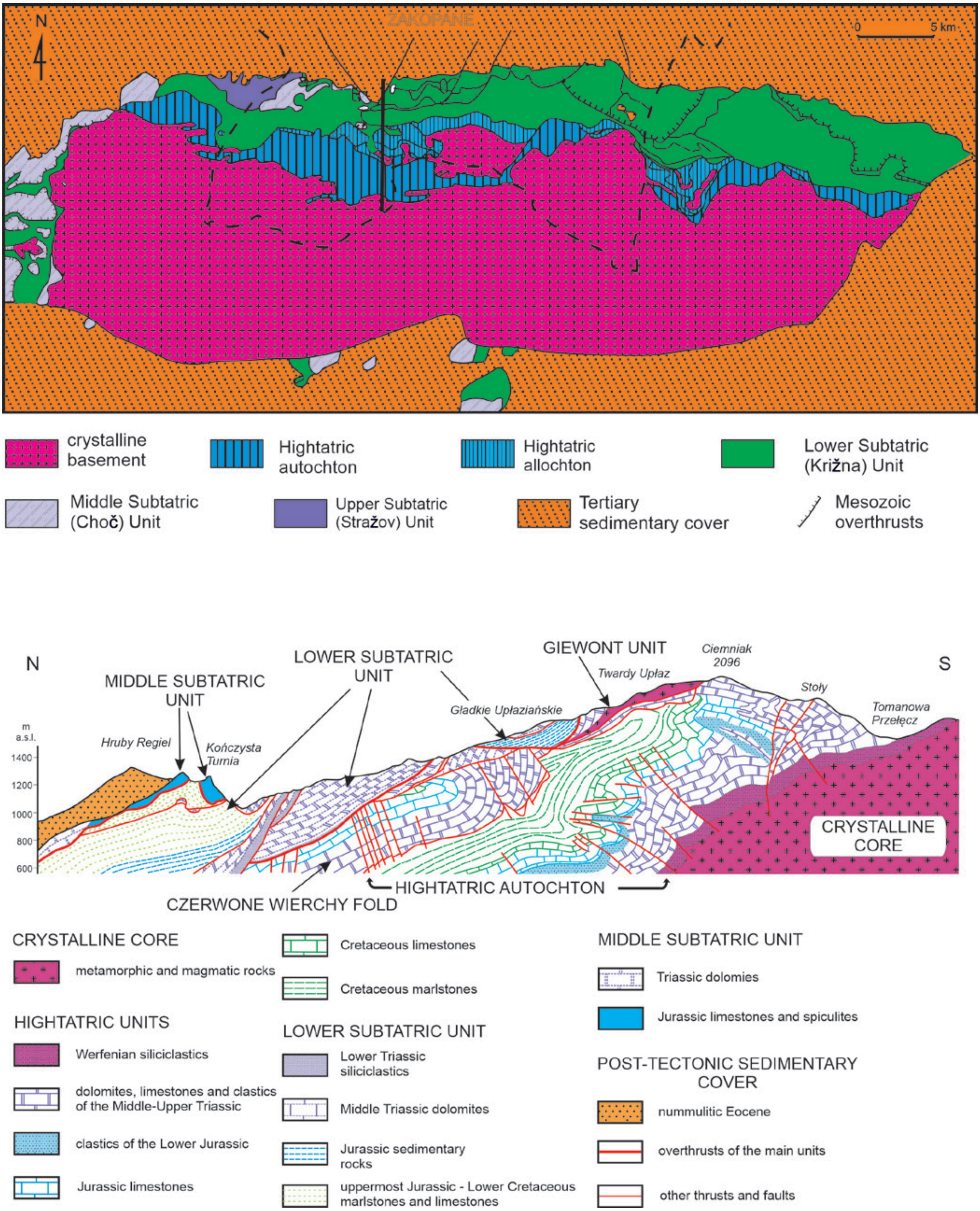


Fig. 53. Simplified geological map of the Tatra Mountains (based on Bac-Moszaszwili *et al.*, 1979) (upper part) and geological cross-section along the eastern slopes of the Kościeliska Valley (after Kotański, 1961; Uchman, 2006)

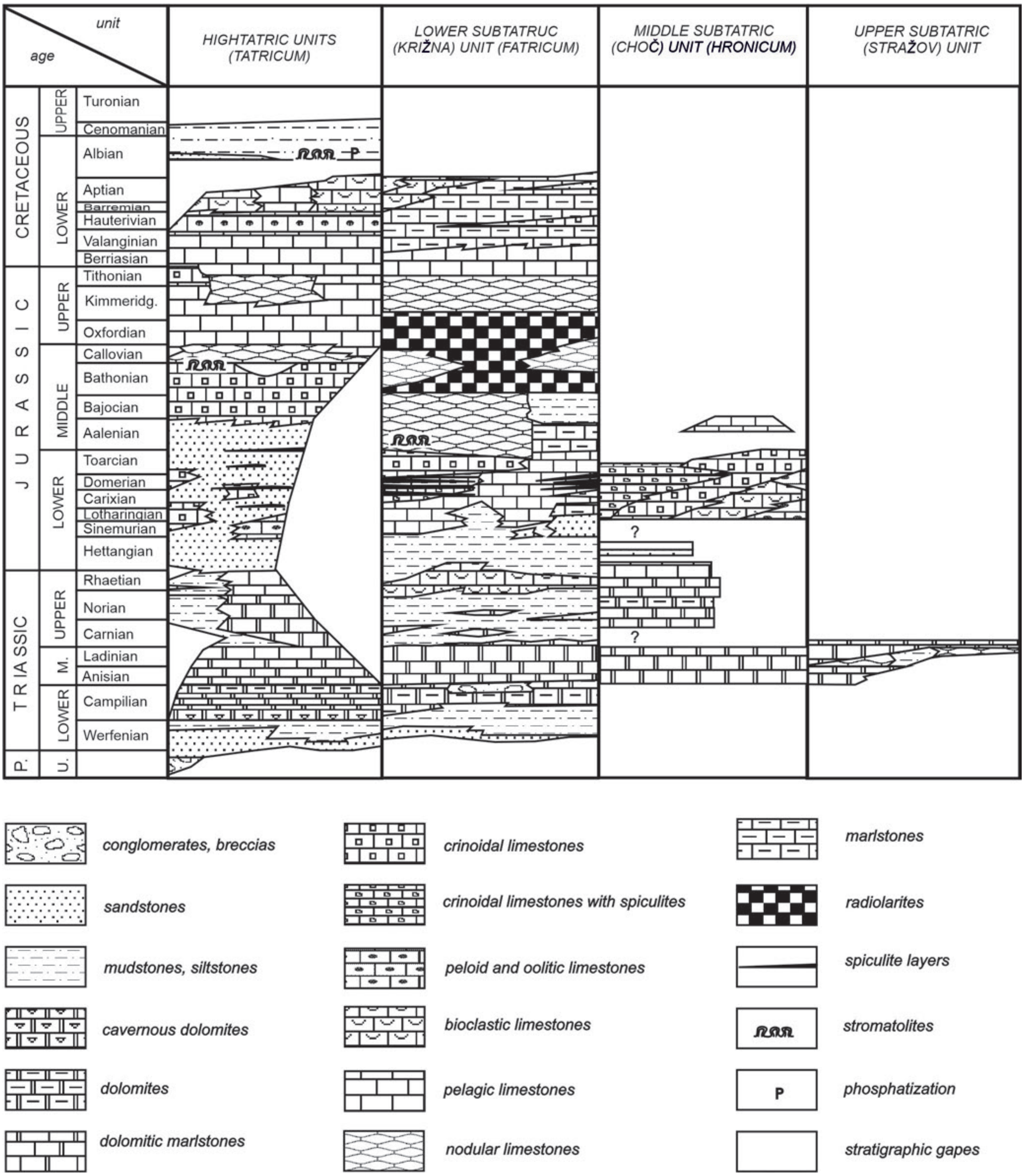


Fig. 54. Major Mesozoic facies of the Tatra Mountains (after Uchman, 2004; changed)

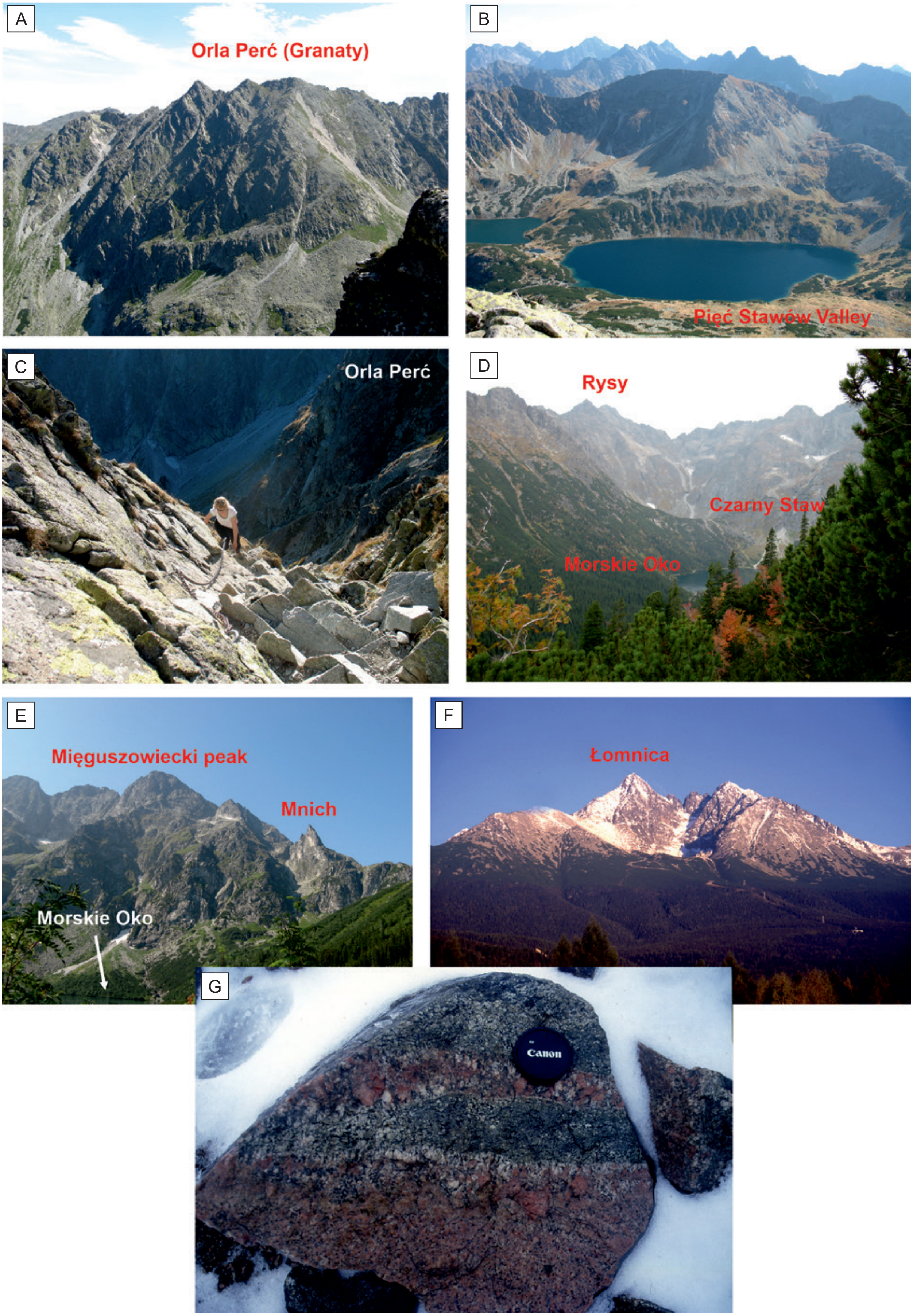


Fig. 55. Landscapes of the High Tatra Mountains with Hercynian granitoidic (and pegmatitic – G) rocks

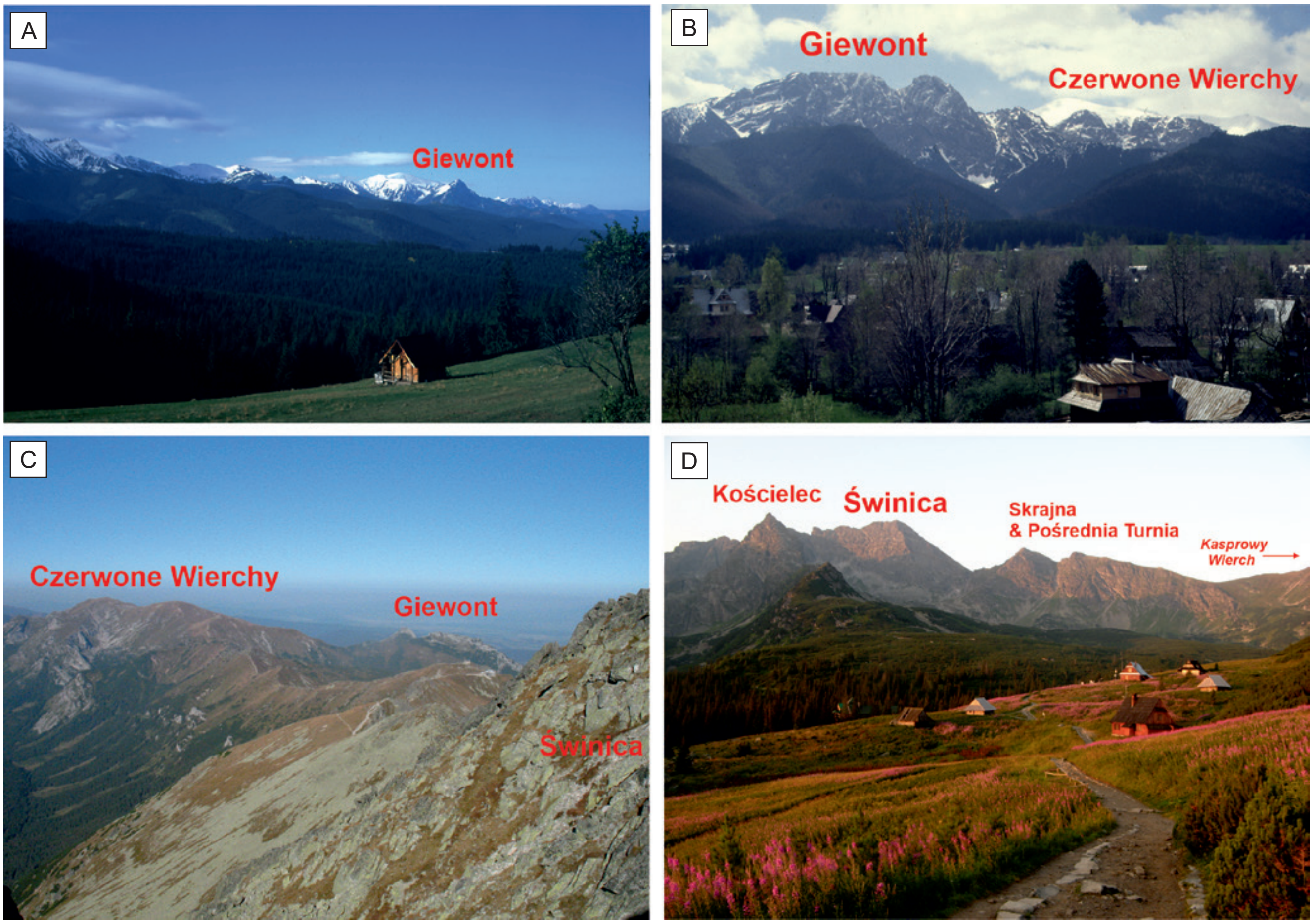


Fig. 56. Panoramic view of the Tatra Mts with most characteristic shape of the Giewont Mt (from north – A, B and from south – C) and the Gąsienicowa Hala area during sunrise (D), including Liliowe Pass (D)

The oldest part of this flysh – the Zakopane Formation (Lexa *et al.*, 2000) is well exposed in the streams in the marginal part of the Tatra Mountains. The uplift of the Tatras, dated using apatite fission tracks, took part probably during the Miocene (15–10 Ma) (Golonka *et al.*, 2005). Glaciation covered all higher areas of the High Tatras and parts of the Western Tatras. The Tatra Mountains (especially the High Tatras) are known to have undergone four glaciations. The most extensive transformations were caused by a glacier 100–230 m thick. Valleys were gouged by the glaciers into the characteristic U-shape. Hanging valleys were created in subsidiary valleys, he glacial erosion also sharpened the mountain ridges and formed deep cirques, with terminal moraines creating large numbers of glacial lakes after the ice had retreated. Material carried down by the glaciers to the foreland formed glacial cones, on one of which the Polish town of Zakopane now stands. The glaciers disappeared from the Tatras about 10,000 years ago. There is now no permanent lying snow on the mountains. The karst, whuch include karrens, abysses, vauclusive springs and limestone caves play important role in creating the Tatras sedimantary cover lanscape. Six caves are open to public in Poland, including Jaskinia Mroźna (the Frosty Cave) with electric ligh, however. the most interesting cave accessible to the public is Belianska jaskyna in Tatranska Kotlina in Slovakia.

Tatra Mountains belong today to Tatra National Pak and are well protected, by in the past, until 19th century the ores were mined in the Kościeliska and Chochołowska valleys. The old mining road, so-called Iron Road is now the popular walking trail connecting Kościeliska Valley with Kuźnice, the site of 19th century steel, today the site of lower station of Kasprowy Cable car.

## Stop 19 – Liliowe pass – Lower Triassic, nappe structures (Figs 54, 55)

(*Michał Krobicki*, *Jan Golonka*)

Using cable car from Kuźnice to Kasprowy Wierch (1987 m a.s.l.) we can go to the central part of the Tatra Mountains.

The peak of the Kasprowy Wierch is built by Hercynian (Carboniferous) crystalline rocks that form an isolated tectonic island (so-called "Goryczkowa hat" = Goryczkowa Crystalline Island) (Fig. 53) which overlying Mesozoic sedimentary rocks of the autochthonous Tatric domain. Liliowe Pass is very famous place in the Polish history of geology. Wieczorek (2000) described it as follow: "Here, during IX International Geological Congress (Wien, 1903) a discussion between Victor Uhlig – the author of Geological Map of the Tatra Mts and Maurice Lugeon – who had never been in the Tatra Mts before, took place. After this heated discussion the nappe conception of the Tatra Mts structure was accepted" (Wieczorek, 2000: 257) (comp. Fig. 51; Krobicki, 2022).

On this pass we can see autochthonous Lower Triassic red quarzitic sandstones of flood plain to lagoon deposits of the so-called Alpine *Werfen*-type facies. On the other hand, the whole tectono-structural position of the Tatra units is well visible (during good weather day!) and full context of connection between crystalline core of the High Tatra and sedimentary cover of the nappe structures.

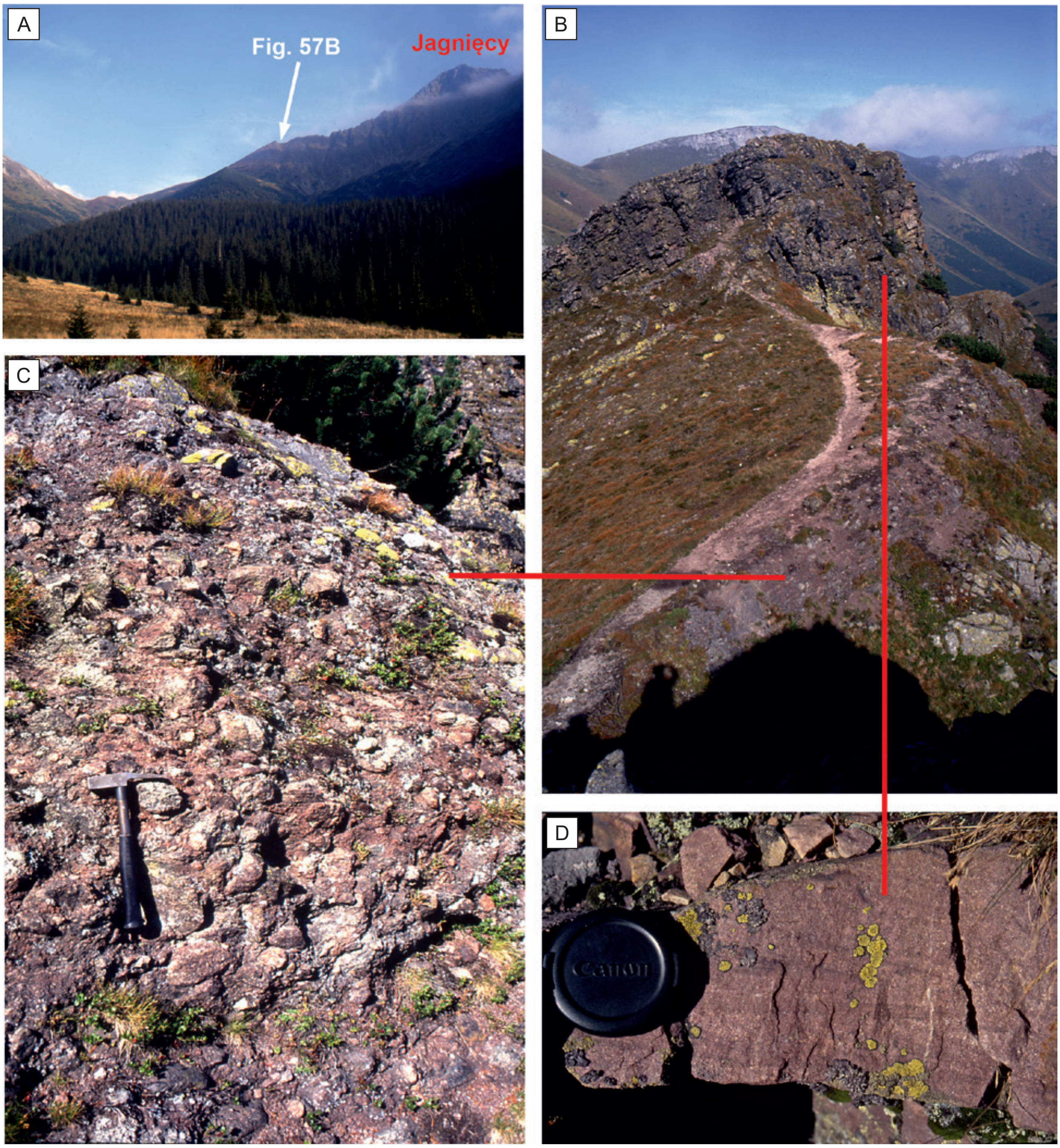


Fig. 57. Panoramic view of the Jagnięcy Mount (A) (Slovakian part of the Tatra Mountains) with granitoidic basement covered by the oldest sedimentary rocks in the Tatra Mountains – Verrucano facies (B, C) and the lowermost Triassic Verfen-type quartzic sandstones with cross-bedding structures (D)

## Stop 20 – Kraków vicinity – Middle-Upper Jurassic and Upper Cretaceous deposits

In Kraków 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 Italy/China/Germany(?) on the 37th HKT (*Himalaya–Karakorum–Tibet*) Workshop – Welcome…!

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