ABSTRACT VOLUME & FIELD TRIP GUIDEBOOK

6th International Symposium of the International Geoscience Programme (IGCP) Project-589

Development of the Asian Tethyan Realm: Genesis, Process and Outcomes

Western Tethys meets Eastern Tethys

Kraków (Poland), 29 September—5 October 2017

ORGANIZERS

United Nations Educational, Scientific and Cultural Organization

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SCIENTIFIC COMMITTEE

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Editor: Michał KROBICKI

Accepted to print: 18.09.2017
prof. dr hab. eng. Przemysław BORKOWSKI, Deputy Director of Polish Geological Institute – NRI

Cover design by: Brygida GRODZICKA, on the basis of graphical concept by Joanna KACZMARZYK
Layout & typesetting: Agnieszka BYLINIAK, Magda CICHOWSKA, Brygida GRODZICKA

Editorial Office: Polish Geological Institute – National Research Institute,
4, Rakowiecka St., 00-975 Warsaw, Poland

Print: Mdruk Sp. z o.o., Sp. kom., 82, Jagiellońska St., 03-301 Warsaw, Poland

Circulation: 100 copies

ISBN 978-83-7863-770-7

The contents of abstracts are the sole responsibility of the authors

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ABSTRACTS

RECONSTRUCTION OF THE TECTONIC TERRANES IN THE MYANMAR TERRITORY

Hla Hla AUNG

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Myanmar represents an evolving continent of two crustal formation history consisting of the Burma plate and the Indochina plate. The Burma plate consists of three distinct lithotectonic entities of: 1) a continental fragment; 2) a subduction-related accreted complex and 3) a coastal area. Eastern Myanmar that is western continuation of Indochina plate is composed of two tectonostratigraphic terranes. Biostratigraphic correlation between the known distribution of dominant Mesozoic representatives of *Monotis*, *Halobia*, and *Daonella* fauna and microfossil assemblages of Triassic age from Myanmar (Figs 1, 2) are made with those from neighboring countries of SE Asia to get estimated palaeogeographic position and reconstruction of tectonic terranes for Myanmar (Fig. 3). These terranes may have originated in Gondwana in Paleozoic. Each tectonic terrane has been separated by three suture zones of the Paleo-, Meso- and Neo-Tethys. They are: 1) The Than Lwin Belt in the easternmost part of Myanmar, which is a tectonic linkage between Inthanon Zone of West Thailand and Changning-Menglian belt (West Yunnan); 2) Shan Boundary Belt of Meso-Tethys suture in the western edge of Indochina plate; 3) Rakhine Western Ranges (Rakhine-Andaman-Nicobar belt) of Neo-Tethys suture at the westernmost part of Myanmar. These accretionary episodes which ended in Early Tertiary, have been followed by post-accretionary deformation of strike-slip faulting of the Sagaing Fault in Myanmar; West Andaman Fault and Sumatra Fault System in Sumatra; spreading in Andaman back-arc basin.

* *Helobia salinarum* Bronn in Shweminbon Formation composed of mudstones, shales and sandstones * Lociality – east of Kalaw, Southern Shan State * Age – Middle-Upper Triassic	* *Halabia* and *Monotis*, *Monotis*-like shell and *Avicular girthicans* Britten of Himalayan type in hard black limestones * Locality – Kayah State * Age – Middle-Upper Triassic
* *Pseudomonotis* in Napeng Formation composed mainly of variegated shales, indurated sandy marls and micritic limestone together with *Aviculopecten* * Locality – Theinni and Lashio area of Nothern Shan State * Age – Norian	* *Halobia comota*, *Halobia lonelli* and *Monotis*, together with forminifera in Thanbaya Formation, composed of dark grey, calcareous mudstones and micritic limestones * Locality – from India – Myanmar frontier southward through Kalay, Gengaw, Mindon to Hainggyi islands * Age – Upper Triassic
* *Daonella indica* Britten * *Halobia salinarum* Broon in sandstones, mudstones and shales sometimes slaty or calcareous, Shweminbon Formation * Locality – Lwekyaw village near Than Lwin river (eastern Shan State) * Age – Upper Triassic	* *Daonella indica* Britten * Locality – Southeastern Yunnan * Age – Middle Triassic * *Monotis salinaria* Bronn in shales * Locality – Himalaya region * Age – Upper Triassic

Fig. 1. The known distribution of dominant Mesozoic representatives of *Monotis*, *Halobia*, and *Daonella* fauna in Shan Massif terrane and correlation with those of neighboring terranes

AGE (my)	PLIOCENE EPOCH		Benthic Foram / Planktonic Foram	Rakhine Coastal Terrane	Rakhine Western Ranges	Central Burma Basin Terrane	Neighbouring Countries Major Palaeo-biogeographic Provinces
10 15 20	MIOCENE	Early Middle Late	Planktonic Foram		*Catposydrax strainforth* *Globorotalia fobsifobl* (Pyawbwe Fm)		Archipelago Series Andaman-Nicobar Ridge Telisa Fomation Sumatra (Indonesia)
25 30 35	OLIGOCENE	Early Middle Late		*Globorotalia kugieria* *Globorotalia opima* *Globiger cuperoensis* (Kalabon Fomation)			Barail Series Bangladesh
25 30 35	OLIGOCENE	Early Middle Late	*Acarinina pentacamerata-Planorotalites palmerae* zone 1. *Acarinina soldadoensis angulosa* *Morozorella aragonensis* zone.2 Lr. Eocene, Leungshe Fm.		*Nummulites atacicus* (Tabyin Fm.) (Larger Foram) *Nummulites atacicus* (Laungshe Fm.) (Larger Foram)		Khirar Stage India Laki Stage India Patala Shale N. Parkistan Zhepure Fm. (Tingri Region) Tibet Indonesia

Fig. 2. Biostratigraphic correlation of micro- and macro-forminifera of Myanmar with major palaeogeographic provinces of SE Asia

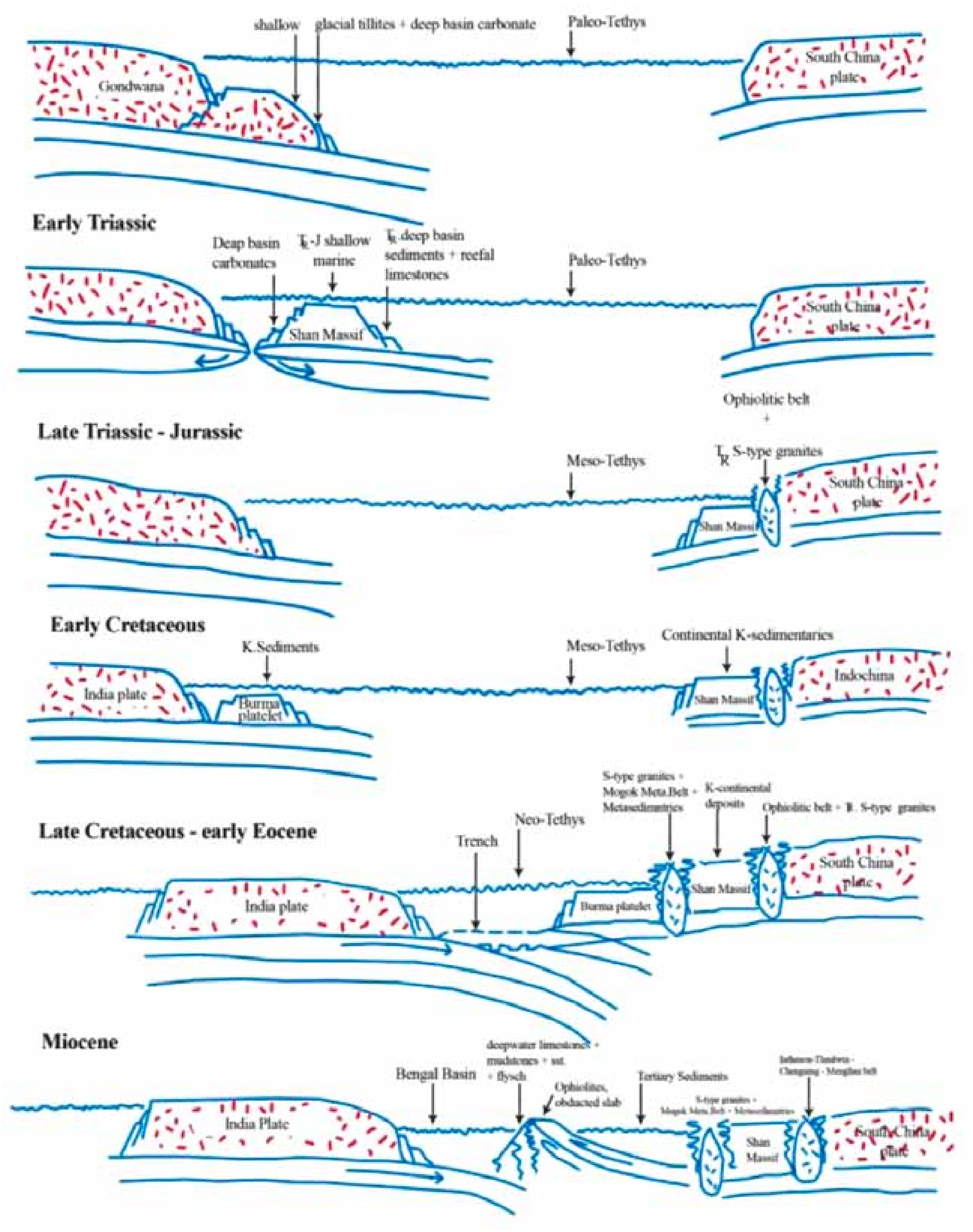

Fig. 3. The sequential evolution cross-sections are derived from available data of the published literatures

PECULIARITY OF ROCK UNIT EXPOSED EAST OF THE KHAO LUAK FORMATION, CENTRAL THAILAND

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This work attempts to review the established stratigraphic units exposed particularly in east of the Khao Luak Formation along the western margin of the Indochina Block of Thailand. We have incorporated new data which were obtained through our field mapping during April–May 2017. We have divided the rocks in the study areas into 4 units: A, B, C, and D units in ascending order. The unit A are turbidites with inverted well-bedded alternations of tuffaceous sandstone and shale. The strata include possibly Late Carboniferous to Early Permian ammonoids (Agathiceras). In some areas, various-sized of fractured and deformed limestone blocks within highly weathered shale have been observed. The unit A can be compared to Ps or Late Carboniferous to Early Permian Khao Luak Formation. The unit B consists of well-bedded to massive fossiliferous limestones which include Early to Middle Permian fusulines. This unit is equivalent to a part of Permian Saraburi Group or Ps-1 (DMR, 2007; Ueno & Charoentitirat, 2011). The unit C is composed of conglomeratic limestone with some volcanic clasts, formed as high ridge mountain in NNE–SSW direction. It is mapped as Trhl unit or Triassic Huai Hin Lat Formation (DMR, 2007; Chonglakmani, 2011). And, the unit D is widely distributed in the eastern part of the study area. Although, this unit was not well studied, it is mapped as Jpk or Jurassic Phu Kradung Formation (DMR, 2007). The unit D consists mainly of agglomerate with reddish brown siltstone and sandstone matrices. Its clasts contain limestone, basalt, andesite, granite, quartzite, silicified mudstone of the older rocks. This unit is not formed a high mountain but undulating terrane. Many limestone clasts yielding Late Carboniferous and Early Permian fusulines were found in this unit. Some strata within unit D show graded-bedding, cross-lamination, trough cross bedding etc. and some show angular clasts with no grading or any sedimentary structure. Many older rocks with varied in grain size and composition of clasts from one place to another have been observed. This may be because of the unit D being deposited in rifted basins on an eroded surface of older rocks during Late Triassic before a period of non-marine Khorat Group was initiated through the rest of the Mesozoic time. We think that the unit D should be excluded from the Phu Kradung Formation which is one formation belonging to the Khorat Group. And it may be one member of the Huai Hin Lat Formation which is considered as pre-Khorat Group. It is possible that the unit C and D might be the same or equivalent units. However, more detail study needs to be done.

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SIMULTANEOUS LATE DEVONIAN ARC-COLLISION IN NORTH QIANGTANG AND RIFTING IN SOUTHERN LHASA: IMPLICATION FOR THE EARLY EVOLUTION OF EASTERN PALEO-TETHYS OCEAN

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Compared with west Paleo-Tethys Ocean, little is known about the early evolution of the east Paleo-Tethys Ocean. During past years, a few Late Devonian–Early Carboniferous magmatic rocks were discovered in southern Lhasa terrane during past years, but their tectonic setting is speculative and it is unclear about its relationship to the Paleo-Tethys Ocean. New SIMS zircon U–Pb dating reveals a 370–365 Ma arc, which was intruded by the 360 Ma A-type granites, in the central Qiangtang, Tibet. Based on the newly discovered ~367 Ma ophiolites located west of Ganma Co, the 370–365 Ma volcanic arc is proposed to be the earliest intro-oceanic arc in east Paleo-Tethys Ocean, and accreted to the Northern Qiangtang terrane during 365–360 Ma. This initial subduction arc is located on the north margin of the eastern Paleo-Tethys Ocean, and it was formed earlier than the 360–350 Ma magmatism in southern Lhasa, located on the south margin of the eastern Paleo-Tethys Ocean. But the arc-continent collision in Northern Qiangtang occurred at 365–360 Ma is only slightly or coeval with the 360–345 Ma magmatism in Southern Lhasa. Thus, the arc-continent collision in the north side of east Paleo-Tethys Ocean probably induced the contemporary magmatism in Southern Lhasa terrane, south side of east Paleo-Tethys Ocean.

CARPATHIANS ARC *VERSUS* ITS FORELAND – HYDRODYNAMIC REGIME A FRONT OF OROGENIC WEDGE (CASE STUDY FROM THE SOUTHERN POLAND)

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The Carpathians form a great arc of mountains, which stretches more than 1300 km from the Vienna Forest to the Iron Gate on the Danube. On the west the Carpathians are linked with the eastern Alps and on the east pass into the Balkan chain and traditionally are subdivided into the West and East Carpathians (Mahel, 1974). The West Carpathians consists of an older range known as the Inner or Central Carpathians and the younger one, known as the Outer or Flysch Carpathians (Mahel, 1974; Książkiewicz, 1977; Ślączka & Kaminski, 1998). The Inner Carpathians are regarded as a prolongation of the Northern Calcareous Alps, and formed part of the Apulia plate in regional sense that is a promontory of the African plate (Picha, 1996).

The Miocene tectonic phase has been dominant tectonic phase that affected the Outer Carpathian and tectonic movements took place during the collision between the overriding Alcapa plate and the North European plate (Cieszkowski, 2003). The boundary between these plates is the Pieniny Klippen Belt. As a result of the intense Neogene orogeny the sediments filling the Outer Carpathian basins were folded and detached from their substrate and several uprooted nappes were created that reflect the original configuration of the basins. Outer Carpathian nappes, thrust one upon another, all together overthrust onto the North European Platform. The Carpathian Foredeep has been formed in front of steeply northward advancing nappes (Golonka & Lewandowski, 2003) and filled by the marine Miocene sediments which cover the area of the entire foreland of the Flysch Carpathians in southern Poland. The foreland consists of basement constructed by Precambrian, Paleozoic and Mesozoic platform formations which were covered by Neogene sediments filling the so-called Carpathian Foredeep and therefore this foredeep is deeply thrust over their foreland.

The thickness of these sediments varies from a few hundred m in the western and northern parts to over 3000 m in the eastern part of the Carpathian foreland. They consist mainly of clays, sands and evaporites (Garlicki, 1979). The allochthonous unit in southern part of foredeep is also called the Zgłobice unit (Kotlarczyk, 1985) or Badenian folds (Książkiewicz, 1972) and in more northern position is so-called unfolded Miocene. These tectonic features can be observed in numerous cross-sections through the marginal zone of the Miocene in front of the Carpathian thrust belt in Poland.

Hydrodynamically analyzed region on this foreland is nearby the Busko Zdrój Spa (about 75 km far of the northern margin of the Carpathian front) which is situated in the Miechów Trough (= Nida Synclinorium) which belongs to southernmost part of the Szczecin–Łódź–Miechów Synclinorium as Alpine (Laramide) tectonic units of extra-Carpathian Poland. By this reason whole area has been uplifted and Miechów Trough originated with intensive tectonic dislocations (with almost W–E orientation mainly – parallel to the Carpathian front) and reorganization (origin of tilted blocks) due to partially erosion of Mesozoic deposits. Therefore location of the same horizones (in this case – Cenomanian deposits) may occur in complitely different position, even up to some hundred m (e.g. OB–I i OB–II wells).

The Cretaceous deposits overlie unconformably the Jurassic rocks (mainly Kimmeridgian) and are represented by the Upper Albian–Lower Maastrichtian, and in the southern part of the Miechów Trough they are covered by the Miocene deposits of the Carpathian Foredeep. Palaeogeographically, this part of the so-called Polish Basin (= German-Polish Cretaceous Basin), during the early Late Cretaceous times (the latest Albian?–Cenomanian) has been covered by an epicontinental sea as a result of the rapid transgression which mainly deposited glauconitic-rich conglomerates and sandstones. These rocks are the lowermost part of the Cenomanian–Santonian siliciclastic-carbonate sequence which is overlying by carbonate sequence of the Campanian–Lower Maastrichtian strata. Usually the thickness of the lower sequence is less than 300 m, whereas the upper one is about 450 m.

Recently, in the vicinity of Busko-Zdrój, have been drilled 6 new boreholes (OB–I – OB–VI), which were partially cored both in the so-called mid-Cretaceous (Cenomanian–Turonian) and Upper Cretaceous (Coniacian–Campanian) strata. From the hydrogeological point of view the most important horizon is connected with Cenomanian clastic deposits.

The thickness of these deposits in analyzed boreholes are variable and never exceed 20–25 m, but according to former drills may reach even 100 m. The overlying strata, sometimes with full record of transitional sequence, are more and more carbonatic and pelagic up to the open-marine limestones with cherts of Turonian in age. After such unification of facies started very thick sequence of spotty-limestones/marly limestones and opoka-limestones/marls full in trace fossils and spectacular syn-sedimentary slump structures as intensive gravitational mass-movements on the sea-floor, which were connected with syn-sedimentary tectonic reorganization and movements which presumably reflect origin of the early Late Cretaceous tilted blocks (see Barbacki, 2004) on the Peri-Tethys region as effect of extensional regime on the Tethyan ocean margin.

In Busko-Zdrój area main hydrogeological regime is connected with Neogene tectonic reorganization (Oszczypko, 1981) when the Carpathian arc was originated and moved northwardly. Hydrodynamic system reflects regional relationships between Carpathians on one side (in south) and foreland on the other side (in north) and its regional tectonic reaction of covering by huge thickness of Carpathians nappes. Under such big pushing effect geostatic and geodynamic regimes changed rapidly. Deep-located sedimentological water (during so-called “elisional” episode – sensu Oszczypko, 1981 – comp. Różkowski & Rudzińska-Zapaśnik, 1983) under high-pressure conditions moved up (“un-gravitationally”) from south to north contrary to gravitational water movement which moved down from north to south.

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LATE TRIASSIC GEOTECTONIC EVENTS – EUROPEAN WESTERN TETHYS *VERSUS* SOUTH–EAST ASIAN EASTERN TETHYS

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Late Triassic was a time of collisions, now known as Early Cimmerian and Indosinian orogenies. Blocks of the Cimmerian provenance and Eurasia (Sengör & Natalin, 1996) were involved in these collisions with the southern margin of Eurasia (Golonka *et al.*, 2006; Golonka, 2007; Richards, 2015). This series of collisions closed the Paleo-Tethys Ocean. The closure happened earlier in the Alpine-Carpathian-Mediterranean area (Western Tethys), later in the Eastern Europe-Central Asia and latest in the South-East Asia (Eastern Tethys). Microplates now included in the Alpine-Carpathian systems formed the marginal part of Europe. Subduction developed south of this zone. Late Triassic collisional events occurred in the Moesia-Rhodopes areas (Golonka, 2007; Okay & Nikishin, 2015; Petrík *et al.*, 2016). Almost in the same time Alborz and South Caspian Microcontinent collided with the Scythian platform in Eastern Europe, and the other Iranian plates (including the large Lut block), collided with the Turan platform (Golonka, 2007, Wilmsen *et al.*, 2009; Okay & Nikishin, 2015; Zanchi *et al.*, 2016). The relationship between Panthalassa terranes and Cimmerian plates was previously postulated and mapped (Golonka, 2007). The Panthalassa terranes bearing reef complexes were also mentioned by Flügel (2002). According to Payberness *et al.* (2016) the Western Panthalassa reefs from Japan corresponds with those of the Tethys Ocean during the Late Triassic. On the other hand, Indochina SE Asian plates and Qiangtang were sutured to South China and therefore the Paleo-Tethys between Qiangtang and Eurasia was closed during Late Triassic times (Metcalfe, 2013; Zhai *et al.*, 2013, Zhu *et al.*, 2013; Song *et al.*, 2015; Wu *et al.*, 2016). The eastern Cimmerian plates were involved in the Indosinian orogeny. This name was derived from Indochina, the region where orogeny was noted one hundred years ago (e.g. Deprat, 1914; Fromaget, 1952). The major unconformity was observed in Northwest Vietnam. The deformed Lower – lowermost Upper Triassic (up to Carnian) marine metamorphosed rocks arranged into nappes and thrusts are covered by Upper Triassic continental red conglomerates ('terraines rouges", see Deprat, 1914; also Golonka *et al.*, 2006). According to Lepvrier *et al.* (2004; see also Lepvrier & Maluski, 2008 and references therein), the main metamorphic event occurred 250–240 Ma. The Late Triassic unconformity and 225–205 Ma postorogenic plutonism was noted by Faure *et al.* (2014) who also suggested that the geodynamic evolution Jinshajiang and Ailaoshan belts in China correlate with Vietnamese events marking the same Indosinian Orogeny. It was related to the closure of Paleo-Tethys Ocean along Raub-Bentong, Sra Kaeo and Nan-Uttaradit suture between Sibumasu and Indochina and Ailaoshan suture between Sibumasu and South China (Metcalfe, 2013; Golonka *et al.*, 2006 and references therein). One of the best examples of the Late Triassic orogenic event occurs in Thailand/Myanmar trans-border zone. The Triassic-Jurassic succession in the Mae Sot area (northern Thailand), belongs to the ShanThai terrane. This zone contains rocks of Triassic cherts (radiolarites), carbonates and flysch (turbiditic) facies, which indicate both pelagic condition and synorogenic deposits. From a palaeogeographic point of view, the Shan-Thai block was a remnant of Paleo-Tethys Ocean (Meesook & Sha, 2010), which occupied a wide realm between Cimmerian Continent and Eurasian plate during Late Paleozoic–Early Mesozoic times. On the other hand, the Late Triassic Indosinian orogenic event has been connected with docking and amalgamation of Indoburma, Shan-Thai (Sibumasu) and Indochina terranes, which constitute recently the main part of SE Asia, well dated by the so-called "base-conglomerate" of the youngest Triassic bed (or the oldest Jurassic in age) characterized by limestone and chert pebbles-bearing conglomerate, which is significant for the understanding of the tectonic evolution of the Shanthai terrane (Ishida *et al.*, 2006; Meesook & Sha, 2010) and determine the age of the Indosinian (Shanthai = Mae Sariang) orogeny. Such synorogenic turbidites and postorogenic molasses were associated with this Indosinian orogeny. The comparison of main orogenic events in the South East Asia regions and their main pulses indicates diachronic, multi-stages movements of Indosinian Orogeny [for example – Early Triassic and Carnian/Norian transition orogenic time in Vietnam (Lepvrier *et al.*, 2004) and late Middle Triassic–early Late Triassic (so-called second Indosinian event; Hahn, 1984; Lepvrier & Maluski, 2008, see also – Cai *et al.*, 2017) and close to Triassic/Jurassic transition in Thailand], firstly terranes docking to Asian plate in the East and later in the West, progressively (in recent coordinates). Additionally, the Late Triassic volcanogenic-sedimentary event in Myanmar correlates presumable with such synorogenic processes, which are represented by the Late Triassic flysch deposits with basaltic pillow lavas [Shweminbon Group (Upper Triassic – Lower Jurassic turbidites), formerly part of Loi-an Gp.; Bawgyo Group (Upper Triassic) and their equivalents; Upper Triassic turbidites (Thanbaya/Pane Chaung group/formations; Bannert *et al.*, 2011; Win Swe, 2012; Cai *et al.*, 2017)].

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PALAEOGEOGRAPHY OF THE TETHYAN ORGANIC-RICH ROCKS IN POLAND

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The deposits of the Outer Carpathian nappes contain organic-rich rocks, which can serve as a target for exploration of unconventional hydrocarbons. They were deposited in the basins belonging to the Tethys realm during Jurassic–Early Cretaceous and to the Paratethys realm during Oligocene times. The palaeogeographic reconstructions can serve as a tool recognizing conditions favorable for organic productivity. The palaeogeography is also the input for palaeoclimating computer modelling, which indicates upwelling zones responsible for the organic-richness (Golonka *et al.*, 2009b).

The Alpine Tethys developed during break-up of Pangea in the Early-Middle Jurassic times. It contained two basins – Magura and Złatne divided by the Czorsztyn Ridge. The Złatne basin included several successions belonging now to the Pieniny Klippen Belt in Poland, Slovakia and Ukraine. The organic-rich sediments were deposited during the early stage of the Alpine Tethys development, before the origin of the Czorsztyn Ridge, when the basin was relatively narrow. The Aalenian-Lower Bajocian dark shales of the Skrzypny Shale Formation belong to the oldest organic-rich sediments in the Pieniny Klippen Belt in Poland. The Rock-Eval pyrolytical analysis revealed that are indeed enriched in the organic matter, but the character of kerogen and TOC content are not good enough and this subject requires farther investigations (Golonka *et al.*, 2009a).The oldest organic-rich deposits within the Magura Basin belong to the Upper Cretaceous-Paleocene Ropianka Formation. The palaeogeographic position and palaeoenvironment assessment indicates that the Siary Zone within the Magura Basin fulfills certain conditions for organic productivity and preservation during the Paleocene times. The results of the Rock-Eval pyrolysis on samples collected from the Świątkowa Member of Ropianka Formation with TOC up to 2% of TOC (Waśkowska *et al.*, 2017) seem most promising.

The Severinic-Moldavidic (Proto-Silesian Basin) constitutes the northern part of the Western Tethys. It developed as back-arc basin within the North European Platform north of the Alpine Tethys. The sedimentary cover is represented by several sequences of the Late Jurassic-Early Miocene age belonging recently to various tectonic units in Poland. The rifting process was accompanied by a volcanic activity, which persisted up to the end of Hauterivian. The Late Jurassic–Early Cretaceous deposition of the Proto-Silesian Basin led to the origin of several formations. The Vendryne, Cieszyn, Hradište (divided into Cisownica and Piechówka members) Veřovice and Lhoty formations belong now to the Silesian and Subsilesian nappes. The Spas Formation with Bełwin and Kuźmina members represents the Skole-Skyba Basin with Borislav–Pokutian Zone. The anoxic black clayey and siliceous shales with lenses of siderites and thin silicified sandstones dominate in these formations. According to Golonka *et al.* (2008) the TOC value of Veřovice Formation is 4%. The value of TOC is up to 11.86% on the Spas Formation rich in organic matter (Kotarba & Koltun, 2006; Ślączka *et al.*, 2014). The thin-bedded flysch deposits with dark mudstones of the Hradište Formation contain 2.5% TOC (Golonka *et al.*, 2008).

The Oligocene Menilite Formation were deposited in the Krosno Basin, which belonged to the Paratethys. This basin originated in front of the advanced accretionary prism. The high organic productivity was caused by restricted conditions as well as by symmetric and circular upwelling (Krobicki *et al.*, 2012). The actual geochemical characteristic of the Menilite Formation provided by Kotarba & Koltun (2006), shows the Total Organic Carbon (TOC) content ranges from 0.18 to 17.25% (mean 4.48%) in the Oligocene deposits occurring now within the Silesian Nappe. Type II, algal oil-prone kerogen dominates throughout the formation, types I and III are not so frequent.

This research has been financially supported by statutory founds of AGH University of Science and Technology (No. 11.11.140.005) and National Centre for Research and Development grant no. BG2/ShaleCarp/14.

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SANDSTONE PROVENANCE AND DETRITAL ZIRCON U–Pb AGES OF THE PERMIAN–TRIASSIC BACK-ARC BASIN SEDIMENTS IN SA KAEO–CHANTHABURI ZONE, EASTERN THAILAND

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The Sa Kaeo–Chanthaburi Zone, developed along the western margin of the Indochina Block, is interpreted as peculiar geotectonic and tectono-sedimentary strata related to the infilling of the Nan Back-arc Basin (Ueno & Charoentitirat, 2011). We investigated sandstone provenance and detrital zircon U–Pb ages from the Nan Back-arc Basin sediments for understanding of tectonic evolution of back-arc basin. According to these previous studies (Hada *et al.*, 1999; Sone *et al.*, 2012), and our new observations, we recognized two geological unit within the Sa Kaeo–Chanthaburi Zone; the Thung Kabin Mélange, the Pong Nam Ron Formation in the present study. The Thung Kabin Mélange is characterized by intensively foliated mélange including basaltic rocks, Permian chert, Permian limestone, and metamorphic rocks within serpentinite and argillaceous matirixes. The Pong Nam Ron Formation is regards as a sedimentary cover on the mélange units, composed of thick massive and bedded sandstone, interbedded sandstone and mudstone, and conglomerate.

Sandstones of the Thung Kabin Mélange and the Pong Nam Ron Formation are feldspathic to lithic greywacke, commonly ill-sorting. Our geochemical analysis suggests that source of both sandstones from the Thung Kabin Mélange and the Pong Nam Ron Formation were strongly influenced to basalt–andesite field in provenance, presenting very similar character. In addition, based on the La–Th–Sc ternary diagram (Bhatia & Crook, 1986), most of both sandstones plotted within the "oceanic island arc" field closed to "basalt–andesite" field. We also determined detrital zircon U–Pb age from two sandstones of the Thung Kabin Mélange and three sandstones of the Pong Nam Ron Formation. The youngest single-grain (YSG) zircon U–Pb ages range from 224 to 235 Ma, whereas the youngest cluster U–Pb ages (YC1s) are between 251 and 262 Ma. In addition, we found several stratigraphic key successions which are originally a conformity relationship through basalt and chert to covered sandstones. These results suggest that sandstones have no significant difference between the Thung Kabin Mélange and the Pong Nam Ron Formation, regarded as originally same sediments.

We also described strike-slip faults within the Thung Kabin Mélange, based on slickenlines and slickensteps observed on representative fault vein along NW–SE to NWN–SES direction. This strike-slip tectonics in the mélange corresponds to that of the Klaeng Fault Zone (e.g., Kanjanapayont *et al.*, 2013). The Thung Kabin Mélange is possibly reinterpreted as "strike-slip mélange", including of oceanic rocks (basalt, Permian limestone and chert) originated from back-arc basin, and covered sediments (Triassic Pong Nam Ron Formation).

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GEOLOGIC SIGNIFICANCE OF THE ZAGROS MOUNTAINS TO SPREAD OF *HOMO SAPIENS*

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The Zagros Mountains of Iran formed along the convergent boundary between the colliding Arabia and Eurasia plates during late Miocene to early Pliocene time. The range consists almost entirely of limestone. Recently, the Zagros Mountains have yielded key evidence of the expansion from Africa of *Homo sapiens*, which originated in East Africa 200,000 to 100,000 years ago. There are two main routes from Africa to Eurasia, a northern route from the Sinai Peninsula to the Levant and a southern route around the Arabian Peninsula. Because recent research in Iran has documented Palaeolithic remains from before 50,000 years ago at Arsenjan, northeast of Shiraz, the southern Zagros Mountains have received attention for their role in the southern route of early human migration. The objective of this study was to determine what geological factors in the Zagros Mountains brought benefits to the first humans coming out of Africa.

Humans taking the southern route around the Arabian Peninsula were able to cross the Strait of Hormuz during the lowered sea level of a glacial stage. One thing that brought them to that place may have been the presence of salt domes west of the Strait of Hormuz.

The salt domes are volcano-like structures a few to dozen kilometers in diameter, which arose in this region as Cambrian evaporite layers, intruding upward as diapirs, penetrated and uplifted the overlying rocks to form domed structures. Where the rising salt breaches these domes, it may flow downhill in “salt glaciers”. Salt diapirism is the result of the mobile, lower-density salt exerting buoyant force against the denser overlying rocks. The presence of salt domes must have ensured a ready supply of this indispensable mineral to ancient humans. The great abundance of salt domes in the Zagros Mountains must have lent confidence to these early people as they made their way into and across the range.

Although salt domes may have initially attracted humans into the Zagros Mountains, something else induced them to cross the range. Traversing the range is not especially difficult as its structure largely consists of whalebacks, structural landforms in which high parallel ridges as long as several dozen kilometers are surrounded by flat areas. From the air, these ridges resemble a group of whales in the ocean. Each whaleback is an isolated anticlinal ridge with its ends plunging downward or cut by faults in the Simple Folded Belt of the Zagros Mountains. The surrounding flat areas are synclinal basins filled with clastics shed from the surrounding ridges. Early humans could have taken routes through the whalebacks along these flats and reached the High Zagros Belt without much difficulty. What attracted them to the High Zagros Belt may have been its large exposures of Lower Cretaceous radiolarite, in conjunction with the Upper Cretaceous limestone terrain that afforded numerous caves to serve as their dwellings. Thus, it was the association of limestone and radiolarite that made the High Zagros Belt a major foothold for early humans in the Zagros Mountains. Here came to be a land replete with limestone caves and widespread radiolarite that welcomed ancient *Homo sapiens*.

AGE DURATION OF THE TETHYS IN THE CHANGNING-MENGLIAN BELT IN WESTERN YUNNAN, SW CHINA: A REVIEW

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The Changning-Menglian Belt in western Yunnan, China (Fig. 1) has been interpreted to be a major suture scar of Tethys in eastern Tethyan realm (Zhong, 1998). One critical concern of this belt is how long the Tethys existed in this particular region. The present study aims to summarize multiple lines of evidence to delineate the temporal duration of Tethys in the Changning-Menglian Belt. The most notable testimony of the Tethys in this belt is dissembled ophiolites, whose isotopic ages previously fell into the spectrum from Devonian to Permian (Duan *et al.*, 2006; Jian *et al.*, 2009). Recent advances further unraveled the existence of Ordovician–Silurian ophiolites, e.g. zircon ages of 443 Ma and 470 Ma (Wang *et al.*, 2013; Liu *et al.*, 2017). Besides, radiolarites have been utilized as remnants of the vanished Tethys in this belt and intermittently span an age range from Early Devonian until Middle Trias-

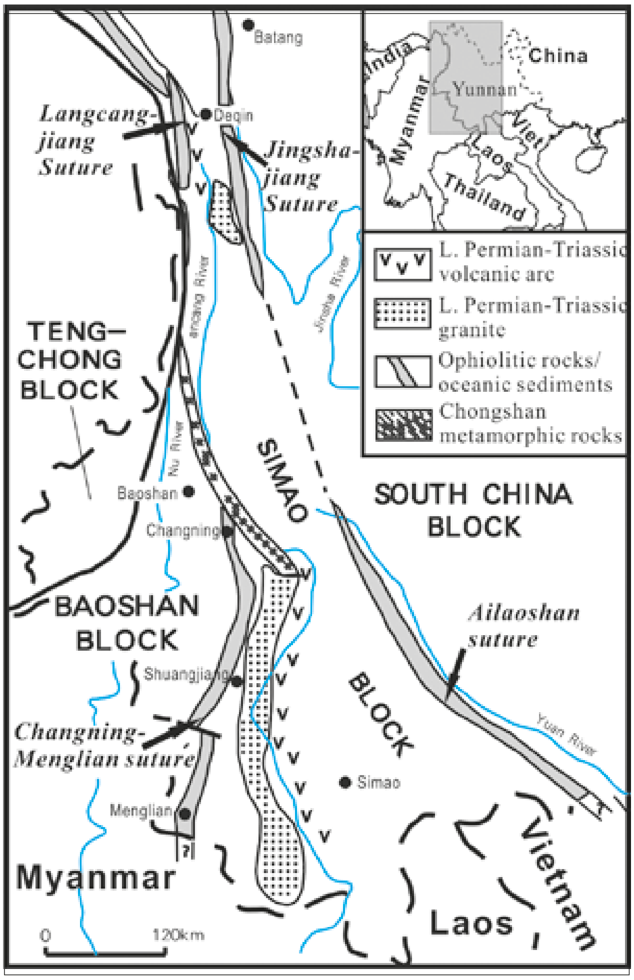

Fig. 1. Tectonic subdivision of the western Yunnan, China showing the distribution of the Changning-Menglian Belt

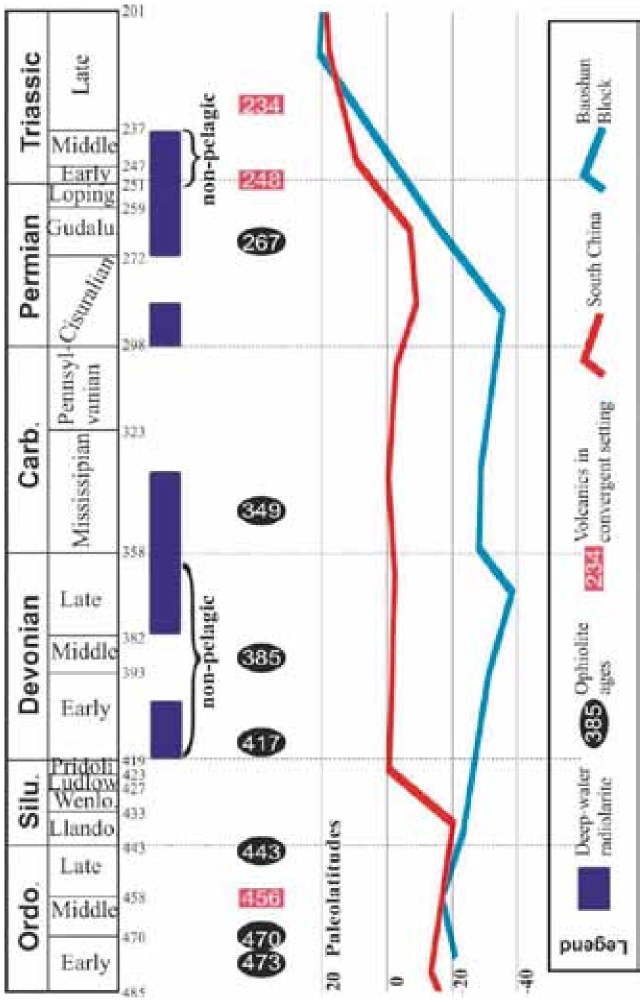

Fig. 2. Multiple lines of evidence for the age duration of the Tethys represented by the Changning-Menglian Belt in western Yunnan, China

sic (Fig. 2) (Liu *et al.*, 1991; Duan *et al.*, 2012). One caveat concerning these radiolarites is that a minority of them are unlikely pelagic deposits, as suggested by geochemical signatures (Ding & Zhong, 1995; Ito *et al.*, 2016). In addition, the presence of Tethys in this belt is also suggested by the sharp contrast between the Baoshan Block with Gondwana affinity to the west and South China of Cathaysian affinity to the east in terms of sedimentary sequences and fossil taxa, especially during the Late Paleozoic (Wang *et al.*, 1996).

On the other hand, convergent episodes of the Tethys in the Changning-Menglian Belt can be deduced from following geological observations. Arc-like volcanics indicative of subduction have been reported from the eastern boundary of this belt and dated to be 456 Ma and 248 Ma (Peng *et al.*, 2008; Nie, 2016). Furthermore, the ophiolites in this belt at places are unconformably overlain by Late Triassic conglomerates (GSTY, 2008). The Lincang granitic batholith flanking this belt on its eastern edge shows geochemical features of post-collision type and was dated to be 234 Ma (Wang *et al.*, 2015).

In addition, the palaeomagnetic analysis revealed that the palaeolatitudes of the Baoshan Block and South China diverged in Early Silurian and converged in Late Triassic (Fig. 2) (Li *et al.*, 2005). Taking all these into account, the Tethys in the Changning-Menglian Belt probably experienced an "accordion-like" tectonic evolution: existed at least from Early Ordovician through Late Permian, contracted in Late Ordovician and eventually terminated in Late Triassic.

Financial support from National Science Foundation of China (41630104) and China Geological Survey (12120011020000150010) are gratefully acknowledged.

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EXTREME GREENHOUSE CONDITIONS: PALAEOENVIRONMENTAL RECONSTRUCTIONS AT THE TRIASSIC–JURASSIC BOUNDARY FROM THE SOUTHWESTERN MARGIN OF THE NEOTETHYS IN THE SALT RANGE, PAKISTAN

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The Triassic was generally dominated by warm-semiarid to arid palaeoclimate. However, palaeoclimatic reconstructions at the Triassic-Jurassic boundary, representing a mass extinction event, indicate a prominent sea-level fall and a change from warm-arid to warm-humid conditions in the Tethyan realm. These events are linked to the Central Atlantic Magmatic Province (CAMP) activity and Pangaea breakup and indicate a demise of the platform carbonates and onset of terrestrial sedimentation across the European basins. In the Tethyan Salt Range of Pakistan, a lithological transition from Late Triassic dolomites through green-black mudstones and thin evaporites (Kingriali Formation) to the overlying Lower Jurassic quartzose sandstones, mudstones, laterites and conglomerates/pebbly sandstones (Datta Formation) provides information on the palaeoenvironmental evolution and sea-level fluctuation of the area. This lithological transition provides sections comparable to the European basins. Field observations and petrographic studies reveal the onset of fluvio-deltaic sedimentation (Datta Formation) on top of the platform carbonates (Kingriali Formation). Bulk rock geochemistry of the succession indicates an increasing upward trend in the intensity of chemical weathering from Rhaetian to Hettangian. The Chemical Index of Alteration (CIA) displays an increasing trend from the Upper Triassic mudstones (CIA 75–80) through the overlying sandstones/mudstones-laterites to the overlying quartz rich sandstones and mudstones (CIA 90–97). Clay mineralogy indicates high illite to kaolinite ratios in the upper part of the Kingriali Formation (Rhaetian). The kaolinite content, a reflection of the advanced stage of chemical weathering and hence greenhouse conditions, increases up-section in the overlying sandstone-mudstone succession (Hettangian). The overlying laterite-bauxite horizons lack illite/smectite and are rich in kaolinite, boehmite and haematite. The overall results for the succession reveal deposition of the fluvio-deltaic, Lower Jurassic, Datta Formation under warm-humid palaeoclimate, probably extreme greenhouse conditions following a comparatively warm-semiarid/arid palaeoclimate during the Late Triassic.

PALEOZOIC AND MESOZOIC BACK-ARC BASIN CHERT OF THE PALEO-TETHYS IN THAILAND (PRELIMINARY REPORT)

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Geologically, SE Asia consists of the several continental blocks derived from Gondwana Supercontinent during the Paleozoic to Mesozoic era. Geological history of the formation of SE Asia contains various tectonic settings such as the dispersion of supercontinent, the subduction of oceanic crust, the formation of the Arc, and the collision of continents after the closure of huge main oceans (e.g., Metcalfe, 1999, 2013). Radiolarian chert has been understood as an indicator of pelagic and "deep-sea" environments of the huge ocean (e.g., Isozaki *et al.*, 1990; Matsuda & Isozaki, 1991). However, deposition of bedded chert on the continental shelf and continental margins have been reported (e.g., Kamata *et al.*, 2009, 2014). These occurrence of the bedded chert shows that the sedimentation of the bedded chert may not necessarily occur integrally in pelagic and "deep-sea" site on basaltic abyssal planes. For implication of geotectonic history the SE Aisa, we should clarify the sedimentary environments of these sediments precisely not only the main ocean basin but also associated smaller basins such as fore-arc and back-arc basins.

In Thailand, the Nan-Uttaradit Suture Zone and Sa Kaeo Suture Zone, which contain ultramafic, mafic rocks with Permian to Triassic radiolarian chert, have long been considered to represent the closure zone of the main Paleo-Tethyan ocean between the Sibumasu and Indochina continental Blocks (e.g., Bunopas, 1981; Metcalfe, 1999). In the last decade, however, accumulation of geological information have led to the new interpretation of the geotectonic divisions of the mainland Thailand (e.g., Ueno & Charoentitirat, 2011), and the new geological significance of these suture zones as a closure zone of back-arc basin situated between the Sukhothai Zone (arc system) and the Indochina continental block (e.g., Ueno & Hisada, 1999; Ueno, 2002).

Early to Late Permian (Asselian to Wuchianpingian) with Middle Triassic (Anisian) radiolarians have been reported from the bedded chert distributed in these suture zones (e.g., Hada *et al.*, 1999; Kamata *et al.*, 2003; Saesaengseerung *et al.*, 2008). The Permian bedded chert occurs with ultramafic and mafic rocks such as serpentinite and gabbro, and basaltic rocks. Geochemical features of the ultramafic rocks of the Nan-Uttaradit Suture Zone indicate arc or back-arc origin (Barr & Macdonald, 1987). X-ray diffractometer of the Permian chert shows that the Permian chert particularly contain a lot of clay minerals such as chlorite, illite, smectite, and kaoline more than Permian bedded chert of the Fan chert distributed in the Inthanon Zone, northern Thailand. Geochemical feature of the REE profiles suggest that sedimentary basin of the Permian chert was influenced both of hydrothermal activity and terrigenous influx. Take tectonic outlines of the Indochina continental blocks and the Sukhothai Zone into account, sedimentation interval of the chert (ca. 57 m.y.), lithology, and geochemical features indicate that the sedimentary basin of the Permian bedded chert distributed in the sutures should be not the main Paleo-Tethys but another marginal sea such as the back-arc basin.

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GEOCHEMISTRY, PETROGENESIS AND TECTONIC SETTING OF THE PLAGIOGRANITES IN THE YESILOVA OPHIOLITES BETWEEN YESILOVA AND TEFENNI (BURDUR, SW TURKEY): NEW TRACE-ELEMENT CONSTRAINTS

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The plagiogranites of the Yesilova ophiolite is located in the Alpine zone, as a major part of the Lycian nappes of Western Taurus Mountains, and formed at the Southern branch of Neotethys. The plagiogranites are formed as lens and pockets within gabbro. It contains plagioclase, quartz and amphibole in a hipidiomorphic granular texture. The samples have high SiO_2 (62.1–79, av. 73.8 wt%), Na_2O (1.67–5.76 wt%), and low Al_2O_3 (10.4–14.9 wt%), K_2O (0.13–1.12 wt%) and TiO_2 (0.18–1.1 wt%) contents. They are predominantly tonalite, and minor trondjhemite and granodiorite in composition, with tholeiitic affinity. SiO_2 increases with decreasing Al_2O_3, TiO_2, CaO, Fe_2O_3, MnO, TiO_2, P_2O_5, Co, Ga, Y, suggesting fractional crystallisation of plagioclase, amphibole (± pyroxene), ilmenite and apatite. In general, plagiogranite shows almost a flat pattern in MORB normalized trace element spider diagram, with positive K and Nb, and negative P and Ti anomies. In chondrite-normalized REE diagram, the samples are also characterized by almost a flat pattern, with a variably developed negative Eu anomaly. The samples are plotted on "Mantle fractionates", and "volcanic zone of within plate" in geotectonic setting diagrams.

Based on geochemical characteristics of the 15 samples, it has been sugested that the plagiogranites differ from the well-known ophiolite plagiogranites with having relatively higher Ta, Nb; and lower Ba and Ce contents as well as higher K_2O/Rb ratio. Accordingly, the samples plot onto the field of "within-plate granites" in various geotectonic diagram. Moderate K_2O content and A/CNK (molar) ratios in the felsic rocks show similarity to that of continental trondhjemites.

EMEISHAN VOLCANISM IN NORTHERN VIETNAM – GEOTECTONIC SETTING AND PERMIAN/TRIASSIC MASS EXTINCTION EVENT

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Geotectonic evolution of Indochina block and their connection with Late Permian Emeishan magmatic event in whole Late Paleozoic geodynamic context is still object of high controversy. Plate tectonic reorganization of this part of SE Asia have been connected with break-up of Gondwana and rapid northward movements of Sibumasu, Lhasa and Qiangtang terranes as an eastern end of Cimmerian Continent. Their palaeopositions from rifting episode during latest Paleozoic time and drifting through Paleo-Tethys, and consequence of opening of Neotethyan Ocean, is also matter of discussion (see Golonka *et al.*, 2006 and references therein). The South China plate includes southern part of China and northeastern fragment of Vietnam. It is separated from North China, by Quingling-Dabie suture, from Indochina by Song Ma suture, from Sibumasu terrane by Ailao Shan suture, from Songpan-Ganzi accretionary complex by Longmenshan suture (Nie *et al.*, 1990; Metcalfe, 2002).

During the Permian time the northern Chinese plate was sutured to the Amurian (Mongolia) terranes (Nie *et al.*, 1990). The south dipping subduction, which existed prior to this collision along the North China plate, jumped south forming the north-dipping subduction along the northern coast of the Paleo-Tethys. At this time the Neotethys Ocean was already opened and widening (Sengör & Natalin, 1996). This ocean had Arabia, Greater India and Australia on one side, and Lut–Qiangtang–Sibumasu on the other. The Neotethyan passive margin is marked by the volcanism. The Emeishan Basalts in southwest China (Ali *et al.*, 2005) belong to the Late Permian worldwide volcanic episodes. Emeishan large igneous province (ELIP) basalts occupy a rhombic shape with an area of 250 000 km^2 (Xu *et al.*, 2001), or more (even up to ~700 000 km^2 – Li *et al.*, 2017) with a total thickness ranging from several hundred m to ~5 km.

The most intensive volcanism effect in northern Vietnam occurs between Song Ma and Song Da suture zones. In this region volcanic rocks of the very thick, up to 1000 m, sequence of the Camthuy Formation (including several outcrops with pillow lava flows, peridotite dykes, and serpentinite bodies – Trung *et al.*, 2006) construct narrow (only to 20 km width) and long (up to 250 km) belt. For example they occur in Bac Yen, Son La, Tuan Giao and Quynh Hani villages, between Vanien village and state China border. There is probably SE branch of magmatic region of ELIP (Chung *et al.*, 1998), which has been moved to present position during Oligocene-Miocene time as tectonic extrusion in SE direction, along Ailao Shan – Red River Fault Zone about 600 km in distance (e.g., Xiao *et al.*, 2003; Ali *et al.*, 2005). Radiometric data indicate 251–253 Ma age of this volcanic event using Ar–Ar and 259–260 Ma using U–Pb method (259 Ma) (Lo Ching-Hua *et al.*, 2002; Li *et al.*, 2017). Both methods indicate age close to Permian/Triassic boundary. Petrologic-geochemical data indicate features similar to famous Emeishan volcanism known from China and Laos (Ali *et al.*, 2005), and Siberian basaltic traps. The South China longitudinal position is quite speculative, so the Emeishan traps fit quite well the present hotspots (Golonka & Bocharova, 2000). The question remains: is this mantle plume related to the continental rifting activity and back-arc opening of basin in Indochina and South China? From the geodynamical point of view this event was perhaps related to the plate reorganization and mantle plume activity. Geochemical and bio-lithostratigraphic data provide strong evidence supporting climatic change and biological extinction (Ali *et al.*, 2005). The cause of the largest known mass extinction event, at the end of the Permian period, is still a matter of hot debate, with many rapid and catastrophic as well as gradual scenarios having been extensively considered, including impact with an asteroid or comet (see e.g. Erwin, 1993; Isozaki, 1997; Wignall & Twichett, 2002). Palaeoclimatic changes in this time, as effect of both geotectonic revolution and simultaneous origin of LIP provinces (including ELIP) have presumable been main reason for P/T global mass extinction (Wignall, 2001; Zhou *et al.*, 2002; Ali *et al.*, 2005; Wignall *et al.*, 2009; He *et al.*, 2010; Tian *et al.*, 2016; Li *et al.*, 2017). The change of palaeocirculation within Tethyan-Panthalassa Ocean and suggested superanoxia event during this time (Wignall & Hallam, 1992; Isozaki, 1997; Kakuwa & Matsumoto, 2006; Schneider *et al.*, 2006) have been results of mentioned interactions.

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GEOLOGICAL RECORD OF COSMIC IMPACTS ON TERRESTRIAL PROCESSES

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Earth rock cycles are controlled by extraterrestrial forces – from the planetary motion vector, the gravity of the moon and nearby planets, to fluctuations of the Sun energy input. Their recognition supports clarification of both the astronomical time scale and more detailed knowledge on Earth forming processes and the life on it in the past (Michalík, 2016).

The topmost stone of the package – the lithosphere, produced by a constant interplay of internal and external forces acting on its surface, recorded billion years long history of our planet. The record of the geological time is represented by thick sedimentary sequences. As the stone archive contains many other data often unknown to us and even a large part of them is missing, the nature offers instead of a solid chronicle, a lot of decayed, incomplete pages, often in mixed order. Their detailed study revealed periodic repetitions of individual texture and structural composition types of sediments. Thus, the formation of sedimentary sequences involved periodic influence of environmental factors.

Complex interpretation of astronomical factors has been given by a Serbian mathematician Milutin Milankovič (1941). It justifies impacts on the planet's surface through changing physical mechanisms governing the energy transition through the atmosphere. The most important astronomical mechanisms that stimulate insulation changes are represented by secular eccentricity, precession and inclination changes of the Earth's rotation axis. The eccentricity of Earth circulation over time changes in response to gravitational pull of other planets. Changes are quasi-periodic, with different amplitude and events in different time slices. Although already from the beginning his theory met skepticism of practical geologists, they later accepted the term "Milankovitch cycles" (with a frequency of 104 to 105 years) and its use as a tool for the interpretation of orbital periods.

The sun energy is the dominant factor in Earth's climate system. Its intensity fluctuates in short-term (11 years Schwabe- and 22 years Hale-) solar cycles that modulate ten- or/and millennial climate changes. Formation of microlaminated sediment has been controlled by the solar activity which ruled rate of microbial growth, biomass production, rainfall, temperature, redox, etc., leading to variations in elements (Ca, Fe, Br) concentrations during deposition and early diagenesis.

Rotation of the Earth's axis is changing in precessional cycles with a periodicity of 23.7 (25.8, respectively) ka (kilo-annum, thousands of years). Thus, the axis describes a cone with a top angle of 23.44°. If this angle changes to 54°, average tropical temperatures will not change dramatically, but the climate in polar regions gets warmer. The actual precession changes were defined already in Hipparchos sky map (150 y B.C). Copernicus (1543) correlated these changes with changing position of the Earth rotational axis. Change of its inclination to the ecliptics (from 23.5° by about 3°) at intervals of about 41 (39.7 to 56.3) ka causes obliquity cycles.

Eccentric cycles – periods of 100– and 405 ka cause the interference of terrestrial planets with the Venus and the Jupiter. Eccentricity variation affects the distance of the Earth from the Sun: the amount of solar radiation reaching the Earth is inversely proportional to the square of the distance of the Sun and the Earth. Thus, astronomically induced signal is recorded in continental and marine sedimentary systems in cycles that correspond to climatic changes lasting hundreds of millions of years.

Depending on the degree of sensitivity to global climate turnover in a given period of time, each component is capable to cause significant climate changes. Global seasonality is low in times of low obliquity (when the planet›s rotational axis is almost perpendicular to the ecliptic plane). At this stage, called as the climatic optimum, long-periodicity eccentric cycles are recorded in sediments. Conversely, when bevelling the Earth's axis, temperature gradients and increasing seasonality are marked. Seasons contrast is rising, polar regions became cooler. If at least one continent is in a polar position, equatorial oceanic current is blocked and seaway for circumpolar currents opens, atmosphere is depleted of greenhouse gases, solar activity is reduced starting the albedo effect, a way to glaciation is open. At this time, called as the climate minimum, short-term periodicities record is dominating.

Global oceanic/atmospheric currents system that controls humidity, rainfall and temperature is the main control element of the climatic regime. Atmospheric current system consists of Hadley, Ferrel and polar jet cells. Where circulation given by these three cells collides with the Earth›s surface, constant atmospheric currents – the trade winds will stabilize. Atmospheric upwelling is characterized by higher humidity caused by adiabatic phenomena that control air saturation. Regional atmospheric circulation is triggered by different heat capacity of land and sea-level. Changed position of circulating cells during seasons leads to seasonal flow changes (monsoons). Sea winds bring moisture that condenses in lifting flow over the coast. Therefore, windward slopes tend to be more humid and the climate of eastern and western coasts can be diametrically different.

Hedley and Ferrel cells interface runs between 15° and 35° latitude, where hot deserts with high evaporation form during climate minimum (when the Earth's axis is oblique to the ecliptics). Monsoon belts are few moving during seasons, limiting the input of precipitations into the area. Hadley/Ferrel downwelling zone moves to the poles and desert zone shifts to 35–40 during climate maximum (reduction of the axis tilt). The monsoon cycle is more efficient and humidity is rising in the zone between 10°–35°. Upwelling arms of both Ferrel and Polar cells shift to 70°, producing the non-glacial cycles (Miall, 1997). De Boer and Smith (1994) stressed the effectivity of precessional cycles at low latitudes, modulated by Earth's orbit eccentricity. They shift caloric equator and borders of climatic zones. Orbital changes affect relative length of seasons, winter and summer contrasts, and monsoon intensity in mid-latitudes. At high latitudes, the effect of changing obliquity is more evident.

Milankovič major claim to fame was to demonstrate the phase difference, on the basis of laborious and lengthy time slices calculations (today they can be relatively easy to master by standard computer). Calculation of orbital changes (based on interactions between sediment, climate and stimulated orbital insolation) are now combined with the geometry measuring cycles, sedimentological assays, mineralogy of clays, analyzing the isotopes of carbon and oxygen monitoring and spectrum of organic residues. Exceptionally well-preserved series of astronomical signals gave rise to a method known as «cyclostratigraphy» and build the model the dynamics of the solar system in a continuous sequence called as «astronomical time scale».

The use of cyclostratigraphy and compilation of astronomical time scale unprecedently specified dating of geological boundaries, especially of the Cenozoic and Mesozoic era (from the existing uncertainty of ~0.5 million of years to ~40 to ~20,000 years). Of course, the most complete resolution was achieved for the Cenozoic period with an accuracy of 0.02 my. The resolution towards the older parts of the rock record: the Oligocene sediments (prior to 30 Ma) decreased to 0.04 ka and Eocene–Paleocene strata (50–60 Ma ago) to 100,000 ka. So far, development of dating is limited by uncertainties related to the diffusion of the solar system (Hinnow & Ogg, 2007). Tolerance of the Mesozoic astronomical time scale is currently about 0.4 to 0.5 mil. years, that is roughly equivalent to accuracy of the biostratigraphic method. The cyclostratigraphy of Paleozoic sediments, over 250 million years provides sufficient data, but relating astronomical time scale is hampered by the lack of astrodynamical model of this period, which is quite different (eg different duration of precession and obliquity). A development of settled astronomical time tends towards improved distinctiveness of the geological time scale at least an order of magnitude. As noted by Hinnow and Ogg (2007), the solution of detailed chronically thorny problems associated with lithospheric plate tectonics, global geochemical cycles, paleoclimate, sea level changes, or biotic processes depends on it.

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THE CARPATHIANS AND THE BANDA ARCS: A 17-YEAR UPDATE

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At a meeting of IGCP 460 in Covasna in Romania in 2000 and at a workshop in the same year of the Europrobe PANCARDI project in Dubrovnik, Croatia, attention was drawn to some striking geological parallels between Eastern Indonesia and the Carpathian-Pannonia region. This theme has been intermittently revisited since that time, and there has recently been an upsurge of interest in the Banda Sea as an extensional basin. A recent paper has attracted considerable publicity by identifying the Weber Basin, in the extreme east of the Banda Sea, as the possible site of the world's longest exposed fault plane. During the same period there have been considerable, but independent, advances in our understanding of the Carpathians and the Pannonian Basin, linked in some cases to work in the Alboran, Tyrrhenian and Aegean extensional zones of the Mediterranean.

There are, inevitably, significant differences. The Banda Arcs orogen is nearly twice as long as the arc of the Carpathians, and encloses a considerably larger basin. Moreover, extension in Pannonia has not, so far as is known, gone beyond the stretching continental crust, whereas much of the Banda Sea is oceanic. These differences are not, however, so great as to rule out the action of essentially similar tectonic processes in the two areas. At the PANCARDI workshop it was noted that large-magnitude extension had formed metamorphic core complexes in the SE Pannonian Basin, that extensional collapse of the Pannonian region had involved two microplates that were juxtaposed in the Early Miocene, that gravity highs are associated with the deep sub-basins beneath the Pannonian plain, that seismic data and of gravity modelling had indicated the lack of a significant crustal root to the western Carpathians and that whereas the Transylvanian basin is cold, with a heat flow less than 50 mW.m^{-2}, the heat flow in the Pannonian Basin, is generally high, averaging about 90 mW.m^{-2}. All these features were identified as having

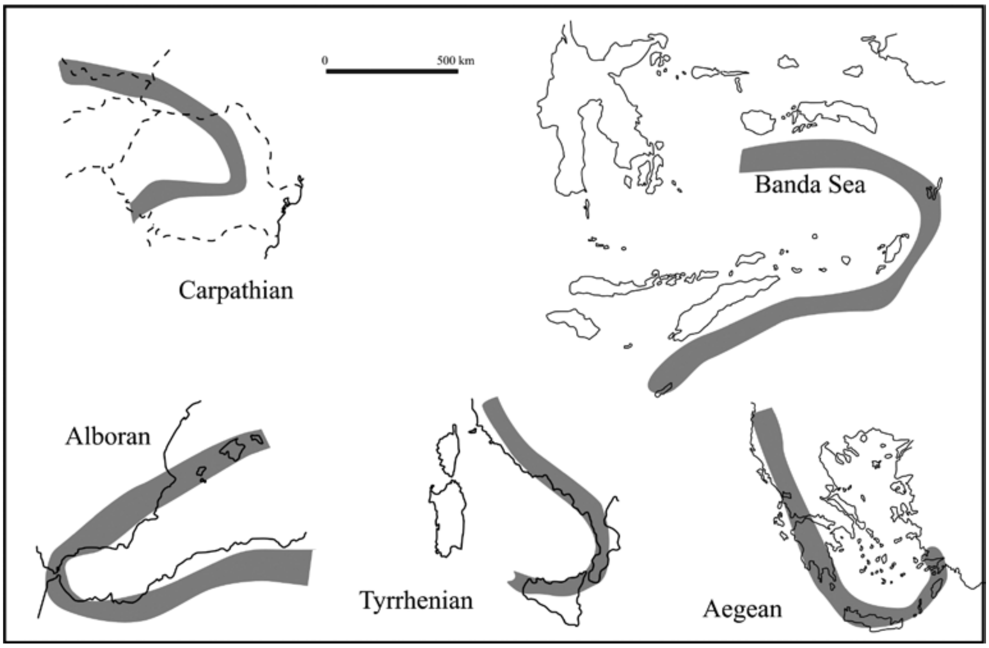

Fig. 1. The Carpathian and Banda Arcs, together with the principal Mediterranean oroclines. All figures are to the same scale

analogues in eastern Indonesia. In the light of the similarities, topics for research were identified in the two areas. For Pannonia it was suggested that the analog approach provided explanations for:

1. The extreme curvature of not only the Carpathians but of other parts of the Alpine chain in terms of interactions between currents in the asthenosphere (‘mantle winds’) and deep rigid keels of continental lithosphere or subducted oceanic lithosphere.
2. The low heat flow in the Transylvanian basin as due to the presence of cold lithosphere at shallow depths, in partial analogy with the situation in the Weber Basin.
3. The break in the Alpine chain at the Vienna Basin, which might have a partial analogue in the North Banda Basin, where extension has fragmented the formerly unitary Buton-Buru-Banda orogen.
4. The “Inselberg” basement exposures in the Pannonian Basin, which analogies with the Banda Ridges, which are demonstrably surrounded by oceanic crust, suggest are not necessarily reliable guides to the nature of the basement in the deeper parts of the basin.

For the Banda Arc it was noted that:

1. That palaeomagnetic results from Pannonia showed that large rotations of blocks several hundreds of kilometres across can occur during extension. Such rotations are needed in the Banda Arc to explain the ENE–WSW orientation of magnetic anomalies in the South Banda Basin when E–W extension would have produced N–S oriented anomalies.
2. That cessation of volcanic activity north of Timor and south of Buru/Seram might be a consequence of the eastwards migration of active subduction, rather than an effect of collision, as is commonly supposed.
3. That the initial extensional drive was probably fueled by the collapse of a small collision orogeny.

The advances in understanding that have been achieved since these suggestions were made are reviewed. It is notable that while the importance of asthenospheric flows in the Mediterranean region is now taken almost for granted, their role in controlling crustal tectonics the Banda Sea has yet to be widely accepted, and the role of orogenic collapse in both the Alpine and Eastern Indonesia collision belts remains controversial.

SINISTRAL SUBDUCTION ALONG THE EASTERN MARGIN OF THE ASIAN CONTINENT IN LATE CAMPANIAN

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Palaeomagnetic studies and hotspot track analyses show that the Pacific Plate existed as an independent plate from about Late Cretaceous, and moved with a dominant sinistral sense of obliquity with respect to the eastern margin of the Asian continent where the proto-Japan arc was situated. Shear directions deduced from the asymmetric fabrics of the tectonic mélanges in the accretionary complex provide important information in examining accretionary kinematics and estimating ancient oceanic plate motions. In this study, we carried out structural analysis of the tectonic mélange of the Late Campanian strata in the Hanazono Accretionary Complex of the Shimanto Belt of the Southwest Japan to better understand the accretionary tectonics related to the oceanic plate motion. The results indicate that mélange fabrics show a sinistral sense of shear both at outcrop and microscopic scales. In addition, restored shear directions in the tectonic mélange are good correspond to the direction of the sinistral oblique subduction of the Pacific Plate. Thus, it is highly possible that the Pacific Plate subducted sinistrally along the eastern margin of Asia in Late Campanian. Previous study indicates the dextral shearing had been occurred before Early Campanian, and therefore changing from dextral to sinistral subduction occurred at about 76 Ma. This study provides a useful information to better understand the accretionary tectonics along the eastern margin of the Asian continent during the Late Cretaceous.

Sr ISOTOPE VARIATIONS IN THE CARNIAN–NORIAN SUCCESSION AT PIZZO MONDELLO (SICANI MOUNTAINS, SICILY): CONSTRAINT ON THE TIMING OF UPLIFT OF THE CIMMERIAN MOUNTAINS

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The Late Triassic–Early Jurassic Cimmerian orogeny produced the final closure of the Palaeo-Tethys Ocean and the accretion of Gondwana-derived microcontinents (Cimmerian continent) to southern Eurasia. Determination of the timing of Cimmerian orogeny is fundamental to assessment of large-scale climate changes driven by the uplift of the Cimmerian Mountains (e.g., Hornung & Brandner, 2005). In order to constrain the timing of uplift of the Cimmerian Mountains, the stratigraphic variations of $^{87}Sr/^{86}Sr$ in the Upper Triassic limestone succession (~430 m in thickness) in Sicily were examined. The Pizzo Mondello section studied here crops out continuously on an eastern slope of the Pizzo Mondello in the Sicani Mountains, western Sicily. This section mainly consists of a pelagic carbonate sequence of the Scillato Formation, and ranges in age from Late Carnian (Tuvalian) to Rhaetian (Muttoni *et al.*, 2004; Mazza *et al.*, 2012). The Scillato Formation represents a deep-water pelagic facies deposited along the Sicanian Basin in the western Tethys Ocean.

We selected fine-grained limestone samples from both the microfacies of lime-mudstone and wackestone to approximate the primary $^{87}Sr/^{86}Sr$ signature of the limestone beds. The $^{87}Sr/^{86}Sr$ values are relatively constant in the Late Carnian (Tuvalian) and Early Norian (Lacian). However, the remarkable rise in $^{87}Sr/^{86}Sr$ occurred across the Early–Middle Norian (Lacian–Alaunian) transition. Variations in $^{87}Sr/^{86}Sr$ values show an increasing trend in $^{87}Sr/^{86}Sr$ from 0.7077 at the base of Lacian to 0.7080 in the Late Norian (Sevatian). These $^{87}Sr/^{86}Sr$ isotope values of samples from the Pizzo Mondello section are consistent with Late Triassic values reported previously from the Carnian–Norian interval in Tethyan sections (Korte *et al.*, 2003). The seawater $^{87}Sr/^{86}Sr$ ratio reflects a balance between input of radiogenic material, relatively high in $^{87}Sr/^{86}Sr$, weathered from continents and non-radiogenic material of low $^{87}Sr/^{86}Sr$ introduced by hydrothermal activity. Korte *et al.* (2003) suggested that the rise in the $^{87}Sr/^{86}Sr$ values from the Middle Carnian to the Late Norian coincide with the Cimmerian orogeny. Our new $^{87}Sr/^{86}Sr$ data from the Pizzo Mondello section reveal a comparable trend, with a sharp increase in $^{87}Sr/^{86}Sr$ within the Alaunian, suggesting the rapid uplift and erosion in the Cimmerian Mountains at this time.

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GEOCHEMISTRY, TECTONIC SETTING AND Nd–Sr ISOTOPIC SIGNATURES OF THE LATE PROTEROZOIC MAFIC MAGMATISM, POSHT-E-BADAM BLOCK, CENTRAL IRAN

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Major, trace and rare earth elements, along with Nd–Sr isotopes, were determined for Late Precambrian–Cambrian mafic magmatism including volcanic, plutonic plus greenstone assemblages and mafic dyke swarms of the metallogenic belt of Posht-e-Badam Block of Central Iran in order to gain insights on the nature of mafic magmas across the Late Proterozoic. These mafic rocks reveal that the Low-Ti samples are Fe–Ti, whereas, the High-Ti alkaline mafic rocks show much more Fe–Ti enrichment occur across a large surface area of the Posht-e-Badam Complex. This geographic separation, along with the numerous inheritance ages of greenstone lithologies that replicate those of older volcanic-plutonic assemblages, consistent with these belts being the remnants of an extensive autochthonous mafic volcanic cover sequence. An increase of Th/Nb ratios in light REE-enriched basalts reflects a change in the nature of coeval crustal contaminants. The isotopically-enriched character of these magmas after ~533 Ma implies more extensive contamination by a felsic crust affected by regional partial melting, succeeding the amalgamation of the terranes. Late Proterozoic mafic dyke swarms follow the stabilization of the Proterozoic crust. The presence of all mafic rocks and dyke types within the older volcano-sedimentary terrane reflects the heterogeneous nature of its lithospheric mantle. All contain a crustal component in the form of an increase in La/Yb with Zr/Nb ratios, and negative εNd values. In contrast, voluminous alkaline mafic sills and dykes of the Central Iran Terrane display an increase in La/Yb with a decrease in Zr/Nb ratios, and positive εNd values indicative of an alkaline mantle component similar to that of Proterozoic alkaline rocks across the Posht-e-Badam Block. The Lower Cambrian time correlation suggests that the composition of the mantle was modified by a pervasive metasomatic associated with the emplacement of the mafic sills and dykes. The secular variation of Proterozoic mafic rocks of this area shows that their composition has been controlled by the nature of the crust and lithosphere they intruded. These geochemical characteristics are possibly imposed upon the sub-continental lithosphere at the time of continent formation but, in the case of younger CFB suites, this may have been modified to a lesser extent by following additions of asthenospheric mantle material.

FEATURES OF CAUCASIAN-ARABIAN SEGMENT OF THE ALPINE-HIMALAYAN CONTINENTAL COLLISIONAL ZONE

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The Caucasian-Arabian segment of the huge Cenozoic Alpine-Himalayan collision belt consists of two domains: the EW-striking Greater Caucasus (GC) at the north and the Caucasian-Arabian Syntaxis (CAS) at the south. GS, in essence, represent the southern margin of the Eurasian plate uplifted over the Main Caucasian Fault (MCF), which is a part of the Kopetdag-Caucasian-Trans-European megafault, whereas CAS includes tectonic domains of the Lesser Caucasus and Eastern Anatolia. It is characterized by large NS-trending positive isostatic anomaly, which suggests presence of a mantle plume head beneath it (Sharkov *et al.*, 2015).

The Alpine (Late Cenozoic) structure of the Caucasus formed by NS horizontal compression generated by interaction of two plates: the Arabian indenter and the Eurasian Plate, which resulted in appearance of transitional vast fold-thrust zone and in transverse shortening of the CAS to 400 km, mainly at the expense of the territory south of the MCF, i.e. Arabian plate (Leonov, 2007).

There is no any subduction zone beneath the GC (Sharkov *et al.*, 2015). However, some north-directed seismic zones occurred beneath transitional fold-thrust zone between Arabian plate (Bitlis-Zagros Fault) and the MCF, which trace up to 50–70 km depth and look like possibly peculiar "embryonic" subduction zones. We assume that mention above shortening occurred due: 1) the tectonic "diffluence" of crustal material apart from the Arabian indenter, in front of the East European Craton, and 2) involving another part of material in descending flows beneath fold-thrust foreland.

The CAS includes the Neogene-Quaternary volcanic belt, which is extended from Eastern Anatolia to the Lesser Caucasus and farther to the Greater Caucasus (Keskin *et al.*, 2013; Sharkov *et al.*, 2015 and references herein). The belt is dominated by two types of volcanic rocks: 1) extensive plateau basalts possessing geochemical characteristics of plume-related rocks, and 2) calc-alkaline and shoshonite-latite volcanics, which origin we considered with "embryonic" subduction zones. Isotopic-geochemical data evidence that both types of magmatism contain some admixture of material of each other (Lebedev *et al.*, 2011; Chugaev *et al.*, 2013). From geophysical (isostasy and seismicity) and petrological (two types of magmatism) data follow that modern deep-seated structure beneath the CAS is area of an active interaction of plume head with descending currents of crustal material.

The Late Cenozoic CAS's basaltic magmatism is occurred at the north termination of the belt of close in time typical plume-related basaltic plateaus of Arabian Plate to the south from the CAS known as Afro-Arabian LIP (Ernst, 2014). Judging on geophysical and petrological data, the north continuation of the Afro-Arabian mantle plume "dive" under the Main Caucasus Fault. It was led to lifting of the Greater Caucasus and provide appearance of the Quaternary Elbrus and Kazbek volcanoes. U–Pb isotopic data evidence about presence of material of the Afar Hot Spot in the Elbrus's volcanics (Chugaev *et al.*, 2013).

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TECTONOBIOGEOGRAPHY AND THE FINAL CLOSURE OF THE PALAEO-ASIAN OCEAN IN EASTERN ASIA

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The Central Asian Orogenic Belt (CAOB) is one of the world's largest and longest-lived tectonic collages, stretching from the Urals and much of central Asia in the west–northwest, then swung eastward to connect with the Tianshan–Mongolian (Solonker)-Hingan orogens in East Asia. The precursor palae-ocean basins to CAOB are generally (and, indeed, loosely) known as the Palaeo-Asian Ocean. Though a large body of research has been published on CAOB, many fundamental questions remain unresolved. Perhaps, one of the most debated questions is the one that concerns the timing of the final closure of the Palaeo-Asian Ocean. Much of the previous and ongoing approach to resolving this problem has been based on the study of ophiolites and associated rock suites thought to represent past subduction zones and oceanic crust. However, not only are there numerous ophiolite belts distributed within CAOB, there also exists a diverse range of views on their ages and geochemical affinities, rendering the establishment of a spatio-temporally consistent and coherent tectonic model for the evolution of the Palaeo-Asian Ocean literally impossible.

This paper attempts to approach the challenge from a completely different perspective, by using the evolutionary history of the Palaeo-Asian Ocean's biogeography as a tool to infer the timing of its final closure and manner of closure. In other words, I will demonstrate how tectonobiogeography can be employed to constrain tectonic models. In this context, tectonobiogeography is referred to as a study of the dynamic relationship, both in space and through time, between biotas and the configuration and motion of lithospheric plates. The role of other important biogeographic determinants such as climate and ocean currents is also implied in this kind of study since these factors are in part co-variants of lithospheric dynamics.

More specifically, this paper focuses on the Permian fossil record in the eastern part of the CAOB that extends from NW China through northern China and SE Mongolia, and eastward to NE China, Russian Far East and parts of Japan. The study explores how the Permian biotas in this region changed their biogeographical identity over time. Of considerable interest is the recognition of two biogeographically mixed Permian biotas found throughout this region: a mixed marine fauna with both palaeoequatorial warm-water and temperate cool-water species, generally followed by a mixed flora characterized by co-occurring Cathaysian (tropical) and Angaran (temperate) floristic elements. Further, we have found that the mixed Angaran -Cathaysian floras show a highly consistent spatial and temporal pattern in that their age of mixing decreased (younged) progressively eastward, from Early Permian in NW China, through mid-Permian in northern China, eastward to Late Permian in NE China. The mechanism leading to the genesis of this succession of biogeographical changes across the Palaeo-Asian Ocean during the Permian may be multi-faceted, but the most likely and parsimonious interpretation appears to lie in plate tectonics and associated palaeogeographic/palaeo-oceanographic dynamics. Under this model, the mixed mid-Permian marine faunas with both warm- and cool-water species are interpreted to suggest the presence of a shallow seaway (not a large ocean) between North China and Siberia that was enlarged and open to the east, and subject to concomitant influence of both warm- and cold-water ocean currents. The mid-Permian marine facies and mixed faunas are generally overlain by mixed Angaran–Cathaysian floras, indicating the cessation of marine deposition and thus the closure of the Palaeo-Asian Ocean. Furthermore, the eastward progressive younging of the floristic mixing is considered to suggest that the collision of North China with Siberia proceeded in a manner much like closing a pair of scissors in that the closure of the Palaeo-Asian Ocean proceeded gradually and progressively from west to the east.

IN-SITU LA–ICP–MS TRACE ELEMENTAL ANALYSES OF MODERN LARGER BENTHIC FORAMINIFERAL TESTS

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The trace element composition of foraminifera test is a well-used proxy to explore the modern ocean and reconstruct the ancient ocean environment. Bulk analysis and individual analysis are the two methods to measure the trace element concentration in foraminiferal test, and both have their own advantages. The former uses solution of a mount of foraminiferal sample and therefore could avoid the possible heterogeneity among different individuals. And individual analysis employs either electron microprobe or laser ablation technic to acquire sample in situ and is suitable for the rare sample or microfossil. To discover the potential elemental heterogeneity among individual tests of the symbiont-bearing benthic foraminifera and differences between two measuring methods, the trace element composition in the tests of four species, *Amphistegina lobifera*, *Heterostegina depressa*, *Laevipeneroplis malayensis* and *Archaias angulatus*, were measured using laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS). Further the results were compared with that of bulk analysis using ICP-MS. Mg, Na, Sr, Si, B and Fe were found as minor elements for the four targeted species with their proportions to Ca more than 200 µmol/mol and show no significant difference among individuals of same species. Li, Cr, Mn and Ba are trace elements with the ratios to Ca less than 50 µmol/mol, and variability does exist within individual tests of same species. Mg/Ca, Sr/Ca and Li/Ca of *Amphistegina lobifera* and *Heterostegina depressa* tests show no significant difference between two methods of bulk analysis and individual analysis.

SEDIMENTARY FEATURES AND NANNOSTRATIGRAPHY OF OLISTOSTROME-BEARING SEQUENCE AT SKRZYDLNA QUARRY (OLIGOCENE) – PRELIMINARY RESULTS

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Olistostromes in the Silesian tectonic unit reflect all evolution stages of a flysch basin: from rifting to orogenic to postorogenic phases (Cieszkowski *et al.*, 2012) The investigated olistostrome is located in the central part of the Polish Outer Carpathians near the southern margin of the Silesian Nappe central sector, occurs within the Oligocene-age Menilite Formation, is dominated by extremely coarse-grained conglomerates and extends over a distance of several kilometres along strike near Skrzydlna village.

The stratigraphic section begins with dark, calcareous mudstones with thin intercalations of turbidite sandstones (Tc); siliceous beds appear higher in the succession. Thick amalgamated sandstone beds, dark mudstones with cherts and several m thick complex of limestone (lithotype of so-called Dynów Marls) rests above. The olistostrome is erosionally incised in the limestone complex and consist of conglomerates, sandstones and thin mudstone beds, which form mud drapes marking breaks in coarse-clastic sedimentation. Fragments of black mudstone derived from erosion of the underlying reduced Menilite Formation strata can be regularly or chaotically distributed within the olistostrome succession. A F–U sequence of turbidite sandstones with subordinate mudstone intercalations overlie the olistostrome.

The characteristic features of the olistostrome-bearing sequence are: (i) clastic dykes with different mineralogical composition, which occur within the deposits underlying the olistostrome; (ii) rapid changes in olistostrome composition: from very coarse grained conglomerates with boulders and contorted slumped sandstone beds to fine grained deposits; (iii) erosional channels: numerous within the olistostrome succession and a few in the overlying some turbidite sequence – which suggest variable dynamics of deposition environment. Extreme facies contrast between the underlying, fine-grained dark-coloured deposits and the succeeding light-coloured olistostrome complex suggest that the originally anoxic Menilite Basin, was temporarily fed by oxidised waters supplied by coarse-clastic mass-gravity flows.

In order to assess the age and sedimentary environment changes nannoplankton analysis was performed. Calcareous nannofossil assemblages observed in all samples were poorly diversified and represented by badly preserved coccoliths. For deposits underlying the olistostrome the nannoplankton zone NP 21 (Late Eocene) was determined. The species preferring temperate environment were found in this assemblage, however cold-climate favouring species e.g. *Reticulofenestra lockeri* Müller, *Isthmolithus recurvus* Deflandre, *Lanternithus minutus* Stradner were relatively often preserved. The youngest species were found in samples from upper part of the succession, i.e. in deposits overlying the olistostrome, which indicate zone NP 22 (Early Oligocene age). Aside from cold-preferring species also warm-water species for e.g. *Helicosphaera compacta* Haq, *Helicosphaera* cf. *recta* Haq, *Pontosphaera multipora* (Kamptner) Roth were recognized in these samples. Additionally *Reticulofenestra ornata* Müller which was found in limestones within the turbidite complex represents an evidence of extreme ecological conditions, being probably related to very low salinity conditions and/or high concentration of nutrients (Krhovský *et al.*, 1992). Quite frequent presence of Mesozoic species provide the evidence of redeposition.

Both sedimentary and nannostratigraphical observation show extreme changes during deposition. Extreme facies contrasts between the underlying, fine-grained deposits, which settled in anoxic environment, and the succeeding olistostrome complex deposited by mass gravity flows delivering coarse debris and oxidised waters suggest rapid uplift of the source area. Abundance of dark shale intraclasts indicates involvement of the basin plain, the proximal part of which was transformed into slope adjacent to the elevated source. Lateral and vertical facies distribution in olistostrome complex suggest significant hydrodynamic changes within and between successive flows. The absence of typical Late Eocene to Early Oligocene species (zones NP18–NP23) in Menilite Formation confirms unique conditions of sedimentation of the succession exposed in the area of Skrzydlna. The alternation of time intervals and mainly the presence of relatively large amount of redeposited species, may reflect eustatic fluctuation of sea-level and temporally low salinity of the environment, which may be related to climatic changes and gradual isolation of Paratethys in Late Eocene and during Oligocene (Švábenická *et al.*, 2007).

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CHANGES IN DEPOSITIONAL ENVIRONMENTS CAUSED BY AN INTERACTION BETWEEN RIVER MEANDERING AND MARINE INCURSION IN PAK NAM PRAN AREA, PRAN BURI, PRACHUAP KHIRI KHAN, SOUTHERN THAILAND

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The Sirinart Rajini Ecosystem Learning Center in Pak Nam Pran was chosen to be the study area. South of the Khlong Khoi meander bend was chosen as a location for collecting soil samples. The 370 cm long sedimentary core provided seventy-four soil samples with five centimeters of each interval. Soil analyses had been made by using lithology, geochronology, chemical compositions, pH and palynology. The sedimentary profile can be divided into four zones on the basis of palynology including *Micrasterias* Zone, *Ilex* Zone, *Cyclotella striata* Zone and Rhizophoraceae Zone respectively from the bottom to the top. The *Micrasterias* Zone is consisted of very dark alkaline clay depth from 370 to 335 cm with abundant *Micrasterias* algal remains representing a back swamp on the floodplain with high silica sediments deposited around 1,000 years B.P. The upward layer is *Ilex* Zone depth from 335 to 292 cm consisting of alkaline clay with sporadic Ilex pollen. This about 1,000-year old zone is described as swampy area when the Ilex trees were invading. The overlying layer is *Cyclotella striata* Zone, depth from 292–55 cm, dominated by diatoms *Cyclotella striata* together with *Navicula* sp. and *Aulacoseira* cf. *granulata*. This zone is still alkaline with high silica representing the swampy area nearby the Pran Buri river bank. Water flowing into the area was from direct rainfalls, runoffs and from the river by annual overflows during the flood seasons. The uppermost layer, 55–0 cm deep, is Rhizophoraceae zone dominated by common to abundant rhizophoraceous pollen. This zone is very dark grey acidic clay with low silica and high chloride as well as plant roots and organic debris. This zone has been accumulated under a mangrove forest with seawater intrusions during the daily high tides. Depositional environments have been changed for the last 1,000 years mainly caused by river meandering course and marine incursions (Fig. 1).

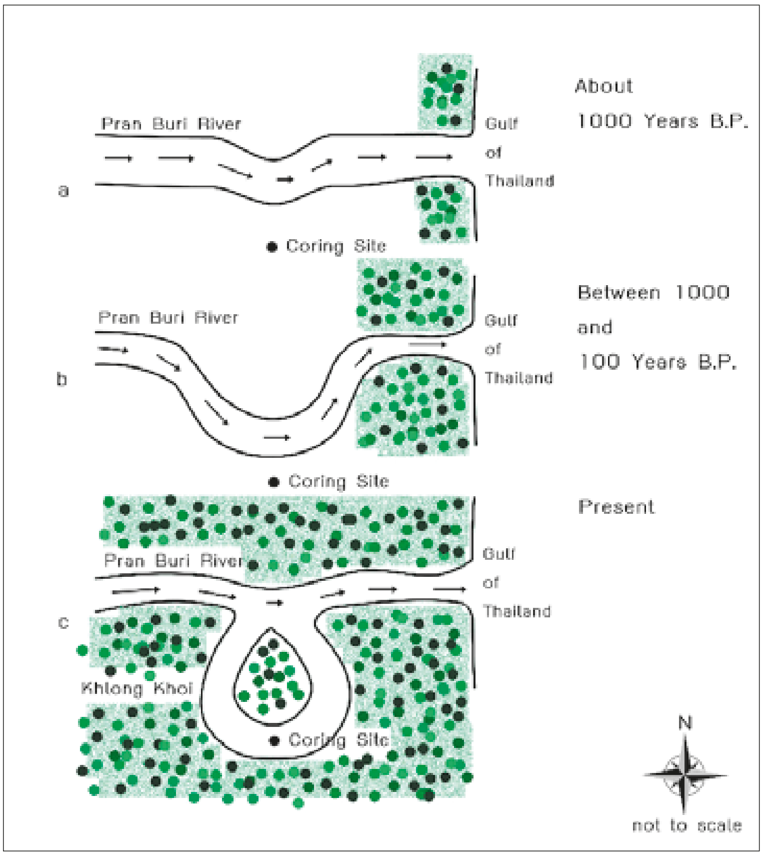

Fig. 1. Diagram showing development of Pran Buri river meandering course and mangrove forest invasion in Pak Nam Pran area in the last 1,000-year period

GEOLOGICAL STRUCTURE INDUCED BY COLLISION OF THE IZU-BONIN ARC; A CASE STUDY OF JURASSIC ACCRETIONARY COMPLEX IN THE JAPAN ALPS OF JAPAN

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In Central Japan, where the northern end of the Izu-Bonin Arc has been colliding with Honshu Arc since Middle Miocene. The collision formed the Southern Japan Alps where is most uplifting area of Japan, and also caused overturned strata and northwards bending of the geological structure. The characteristic structure is called "rolling up structure". However, the detailed and quantitative information of the geological structure has not been reported. Therefore, we carried out the detailed geological survey on the Jurassic accretionary complex in the Southern Japan Alps to better understand the geological structure induced by the collision. As the results, the geological structure in the study area can be divided into three types (named Type A to Type C from north to south) based on their strike and dip. Type A is characterized as overturned strata with N–S striking and gently dipping eastward. Type B show the overturned strata with N–S striking and steeply dipping eastward. Type C is characterized as NE–SW striking and steeply dipping westward. The strata of Type C is not overturned. Basically, the strata gradually change westward dipping to eastward dipping from south to north, and the structural change induced by the collision are gradually larger from south to north part of the Southern Japan Alps.

ACCURACY OF DETERMINING THE AGE OF DEPOSITION BY DETRITAL ZIRCON U–Pb DATING ON THE EASTERN MARGIN OF THE ASIAN CONTINENT DURING THE LATE CRETACEOUS TO PALEOGENE

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Detrital zircon U–Pb dating plays an important role in determining the age of strata. However, the zircon ages indicate not the depositional age but the crystallization age, and therefore it is necessary to evaluate the accuracy of determining the age of deposition using zircon age data.

We carried out U–Pb dating of detrital zircons from sandstone in the Cretaceous to Paleogene Shimanto accretionary complex in Kii Peninsula, Japan, with the aim of evaluating the accuracy of U–Pb zircon ages as indicators of the depositional age of sedimentary rocks by comparing between zircon ages and radiolarian ages within the same geological bodies.

As the results, the youngest peak ages are in good agreement with depositional ages inferred from radiolarian fossils in the Cretaceous, and the youngest peak ages become younger as tectono-structurally downward. Therefore, newly crystalized zircons were continuously supplied to the sediment by constant igneous activity during the Late Cretaceous, and the zircon ages provide remarkably useful information for determining the age of deposition in the Late Cretaceous. On the other hand, the youngest peak ages during the Paleogene are obviously older than the depositional ages inferred from radiolarian fossils. Basically, it is difficult to obtain the depositional ages by only using the zircon dating of the sandstone in the eastern margin of the Asian continent in the Paleogene.

DEPOSITIONAL PROCESS OF CHERT FRAGMENT DOMINANT SANDSTONE IN THE JURASSIC ACCRETIONARY COMPLEX IN JAPAN

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Sandstone and conglomerate that contain a large amount of chert lithic fragments have been reported from Paleozoic–Mesozoic accretionary complexes in the Japanese Islands. These sandstone and conglomerate were interpreted as redeposited material caused by sub-marine landslide in accretionary complexes (Saito & Tsukamoto, 1993), and understanding its depositional process will provide further information on shallower portion of accretionary complexes. This study reports sedimentary and deformation structure of chert lithic fragment dominant sandstone and conglomerate to estimate its depositional process.

The Jurassic accretionary complex, which belongs to the Kamiyoshida Unit of the Northern Chichibu Belt, in the Kanto Mountains are examined for this study. Geological age of shaly matrix in the Kamiyoshida Unit is estimated to late Early–early Late Jurassic (Sekine *et al.*, 2001). Two types of chert fragment dominant sandstone (or granule) was recognized in the study area. One type of chert sandstone (type 1 chert sandstone) are associated with faults. The foot wall is composed of angular sandstone or granule dominated by chert lithic fragments. Brecciated chert occur in the hanging wall, and the slickenside is observed in the lowermost part of the brecciated chert. The fault zone is 30 cm in thickness, and is composed of lensoidal shaped brecciated chert block and fault clay. The second (type 2 chert sandstone) occur as sand bed in the alternation of sandstone and shale. The frequency and thickness of sand beds are increase upward in the alternation of sandstone and shale. Grading and cross lamination division of Bouma Sequence are observed in the alternation of sandstone and shale. The amount of chert lithic fragment in the type 2 chert sandstone is lesser than that of the type 1 chert sandstone, and the type 2 chert sandstone also contain quartz and feldspar grains.

The sedimentary structure recognized in the type 2 chert sandstone show that the type 2 chert sandstone was transported by turbidity current. On the other hand, brecciation of type 1 chert sandstone is possibly caused by fault activity. Although the relationship between the types 1 and 2 chert sandstone is still unclear, this study suggest possibility of tectonic brecciation for origin of chert sandstone.

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PERMIAN ADAKITIC MAGMATISM AT THE SOUTHERN MARGIN OF SIBERIAN CONTINENT, MONGOLIA

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The Central Asian Orogenic Belt (CAOB) has been brought to attention these days in the tectonic process of Eurasian continent. Although subduction-related magmatism of the Selenge complex, which is the Permian volcano-plutonic rock series at the southern margin of "Siberian continent", is a key factor to understand development process of the CAOB, very few attempts have been made at its geochemistry. Here, geochemistry of the Khanui Group, which is a main volcanic facies of the Selenge complex, is described to discuss the Permian magmatism of the southern margin of Siberian continent.

Geochemistry of the Khanui Group. The SiO_2 concentration ranges from 53 to 63 wt.%. The samples are nearly identical to each other in their concentrations of the following elements: Al_2O_3 ca. 17 wt.%; CaO ca. 5 wt.%; Na_2O ca. 5 wt.%; Zr ca. 200 ppm; Nb ca. 7 ppm; Th ca. 9 ppm. The samples are decreased in TiO_2, MnO, MgO, P_2O_5, Co, Cr, Y, Ni, Zn and Sr, with an increase in SiO_2. The samples have high FeO*/MgO ratio (1.1–7.2, avg. 4.1), high Mg# (20–62, avg. 43), high Sr concentration (> 488 ppm, avg. 1458 ppm) and Sr/Y ratio (> 37, avg. 110), and moderate La/Yb ratio (7.3–33, avg. 24). The MORB-normalized multi element pattern for the samples shows an enrichment of LILE against HFSE with clear negative Nb anomaly, and Chondrite-normalized REE pattern shows a gradual HREE depletion. These features strongly suggest volcanic arc origin. Besides, the chemical characteristic of the Khanui rock is similar to that of the High-Mg andesite, especially of "adakite" (Fig. 1). The present samples, richer in Cr, Ni, Fe and Mg than the typical adakitic melt, are equivalent to bajaite/transitional adakite (Figs 1, 2).

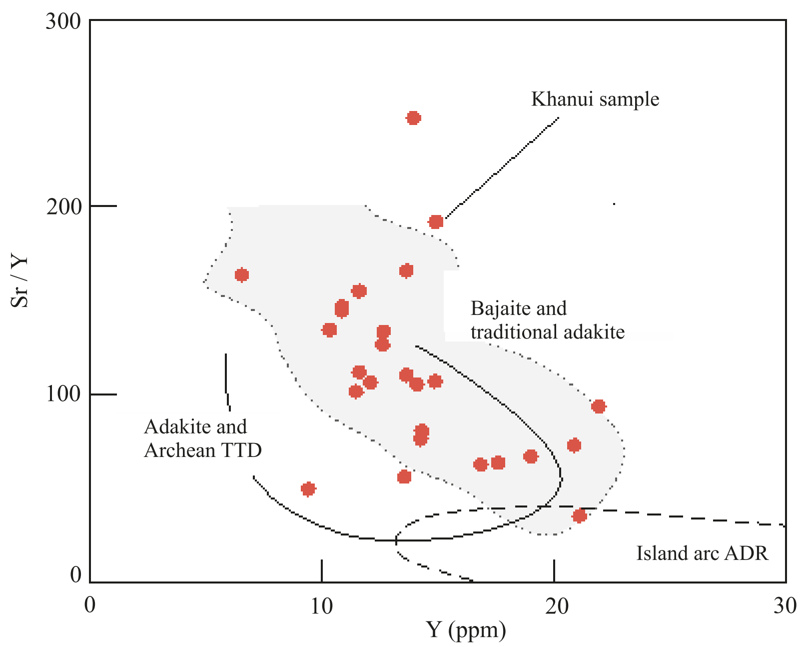

Fig. 1. Sr/Y–Y diagram (modified from Tsuchiya *et al.*, 2005)

Petrogenesis and tectonic implication. The adakitic initial magma is, in many cases, attributed to a partial melting of basaltic slab at subduction zone (e.g. Defant & Drummond, 1990). In addition, it is known that the interaction between the slab-melt and mantle peridotite causes great enrichment of Mg, Cr and Ni to form bajaitic magma (e.g. Tsuchiya *et al.*, 2005). The Khanui rock is plotted beside the mixing line between primitive mantle and slab-melt in the relationship between Ni and Cr (Fig. 2). Thus, the enrichment of Mg, Cr and Ni of the samples appears to be attributed to the interaction between the slab-melt and overlying-mantle peridotite. In other words, the Khanui rock gives evidence that a basaltic slab, namely subducted oceanic plate, had existed beneath the Siberian continent in Permian. The adakitic initial magma is generated by partial melting (10–20%) of basaltic slab at ca. 2.0 Gpa and 900–950°C in normal/steep subduction (< 30˚), and "partial melting of subducted young/hot (< 40 m.y. old) oceanic plate" was proposed as a practical model to explain the adakitic magma generation (Defant & Drummond, 1990; Sen & Dunn, 1994; Rapp & Watson, 1995). However, the Khanui case, that the subducted slab was estimated as more than 50 million years old (Kurihara *et al.*, 2009), seems not to be applicable to this model. That is, the subducted slab is inferred to have been too old/cold to cause slab-melting. On the other hand, Gutcher *et al.* (2000) demonstrated that the "flat subduction" could cause the melting of old/cold slab. Although further discussion is required, the initial magma of the Khanui rock might have been generated in connection with the Late Paleozoic flat subduction of the Mongol-Okhotsk oceanic plate beneath the Siberian continent.

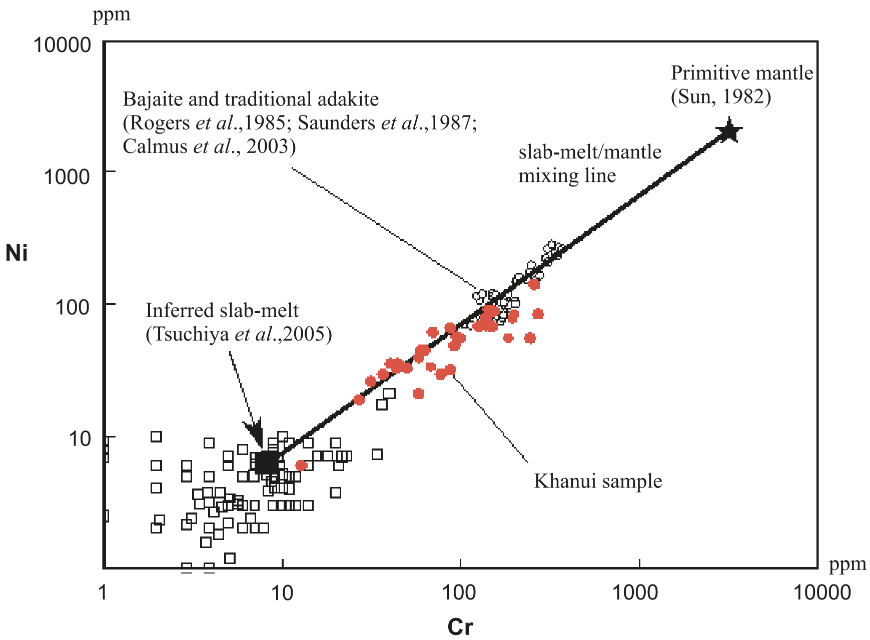

Fig. 2. Ni–Cr relationships. Open squares are adakitic rocks in the Kitakami Mts, NE Japan

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WHAT HAVE WE DONE FOR THE GEOLOGY OF SOUTHEAST ASIA?: OUR FOOTMARKS IN THE LAST TWO IGCP PROJECTS ON THE EASTERN/ASIAN TETHYS

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In the last two IGCP projects related to the development of Eastern/Asian Tethys, namely IGCP-516 (Geological Anatomy of East and South Asia: Paleogeography and Paleoenvironment in Eastern Tethys) and IGCP-589 (Development of the Asian Tethyan Realm: Genesis, Process and Outcomes), we had a number of contributions for the Paleozoic–Mesozoic geology and geotectonics of Southeast Asia. They were presented mainly as "fresh and hot" outcomes from our two overseas research projects, which were financially supported by the Japan Society for the Promotion of Science (JSPS) and were luckily running almost in parallel with the two IGCP projects. In every annual symposium of these two IGCPs, our research group always constituted a major part of "regular delegates", and in these symposia we have made contributions in various geological disciplines such as the stratigraphy, palaeontology, sedimentology, palaeogeography, petrography, geochemistry, geochronology, and tectonics of Thailand, Laos, Yunnan (China), Myanmar, Cambodia, and Indonesia. At closing these two successive and successful IGCPs over twelve years, we will try to look back our "footmarks" and make a short summary of our activities and results, in this presentation.

Totally more than 50 contributions, relevant to the above-mentioned subjects, were made by us in the last 11 symposia of IGCP-516 and IGCP-589. They include many different issues on the geology of Southeast Asia, of which the following are the major topics being focused. Ultimately our research aims at clarifying the geotectonic subdivision and evolution of mainland Southeast Asia, by combining all these lines of information obtained based fundamentally on fieldwork.

- Reconstruction of pelagic stratigraphy of Paleo-Tethyan seamount and ocean-floor sediments, found in the Inthanon Zone of Northern Thailand and the Changning–Menglian Belt of Yunnan
- Opening and closing processes of the Paleo-Tethys and its sequential environmental change
- Characterization of hemipelagic successions on Paleo-Tethyan margins of the Sibumasu Block and their emplacement process
- Chronology and palaeobiogeography of Permian faunas in the peri-Gondwanan Sibumasu Block
- Clarification of arc development of the Sukhothai Arc system formed by the subduction of a Paleo-Tethyan oceanic lithosphere
- Permian–Triassic stratigraphy and basin development of the Sukhothai Zone and its correlation to the south in mainland Thailand
- Stratigraphic and geotectonic characterization of Carboniferous–Triassic sedimentary successions of northern Laos and their linkage with those found in Northern and Northeast Thailand
- Geological nature of the Nan suture and petrochemical property of its ophiolitic rocks
- Geological features of the Sa Kaeo-Chanthaburi suture and the closure of the Nan back-arc basin
- Late Paleozoic basin development, palaeogeography, and tectonics along margin of the Indochina Block
- Origin of Tethys-type faunas from blocks located outward the peri-Gondwanan Sibumasu Block.

Then, our main results are summarized as follows although some of them are still ongoing research topics.

- Establishing entire biostratigraphy and time-frame of the seamount-type, Paleo-Tethyan Doi Chiang Dao Limestone in the Inthanon Zone and its correlation to the Banka Limestone of the Changning–Menglian Belt

- Clarification of detailed environmental change at the opening stage of the Paleo-Tethys
- Distinction of several different types of chert deposition in pelagic, hemipelagic, forearc, and back-arc settings, based on stratigraphic and geochemical characterization of siliceous rocks, which often have been jumbled together as "cherts"
- Elucidation of the development and closing process of the Paleo-Tethys, including the formation of accretionary prism and nappe based on clastic petrography and mélange kinematics in the Inthanon Zone and the crustal shortening along continental margin of the Sibumasu Block
- Understanding of the "two-storied" structure of the Inthanon Zone, which consists of a Sibumasu basement and tectonically overlying nappes (originally forming accretionary complexes along margin of the Sukhothai Arc) containing Paleo-Tethyan oceanic materials
- Establishing detailed chronologic framework of Permian–Triassic successions of the Sukhothai Zone and the volcanic-arc evolution based on their detrital zircon dating
- Clarifying the extension of the Sukhothai Zone in the Central Plain and the Klaeng Zone of Thailand
- Characterization of Northern Laos as the northern extension of the Sukhothai Zone, based on Permian–Triassic stratigraphic and fossil faunal similarities
- Distribution of Nan ophiolites and the new petrochemical constraints on their tectonic setting
- Documentation of gypsiferous deposits with a mid-Carboniferous evaporitic setting on a margin (shelf) of the Indochina Block
- Verification of large tectonic displacement of Cathaysian slivers along the outward of the Sibumasu Block, based on palaeobiogeographic characteristics of Permian faunas from the former areas.

It might sound like somewhat praising ourselves, but we rightly evaluate that we have been producing fairly healthy outcomes that substantially lead to the clarification of Paleozoic–Mesozoic geohistory of Southeast Asia. Overall our research could have contributed, in our own expertise, to better understandings of geotectonic subdivision of mainland Southeast Asia and evolution of the Eastern Tethys. Although IGCP–589 is finished at this last symposium in Kraków, an expected successor IGCP project will hopefully provide a new "haunt" of our colleagues for discussing current and new geological issues on the whole Tethys.

UPPER CRETACEOUS AND LOWER PALEOGENE SANDSTONE BODIES IN EXTERNAL PART OF THE MAGURA NAPPE (THE WESTERN TETHYS DEPOSITS)

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The Magura Basin developed as the partly isolated area of the Outer Carpathian Basins in the Western Tethys Ocean. It functioned from the Jurassic up to the Miocene times, and several stages of its development are recorded by the different types of turbiditic deposits. After the Miocene, when the orogenic movements took place, deposits of the Outer Carpathian Basin were folded, uplifted, moved into North and formed the assemblage of the nappes. Deposits of the Magura Basin were transformed into the Magura Nappe, which today extends mostly longitudinally from Czech and Slovak Republics in the West up to Ukraine in the East. The biggest part of the Magura Nappe is located within the Poland territory, where it is possible to study of the sedimentary history of the Magura Basin in its central part. The Magura Nappe is divided into 4 tectonic-facies zones (after Koszarski *et al.*, 1974). The most northern part of this nappe, called the Siary Zone, refers to external part of the basin.

In the Late Cretaceous and Early Paleogene, the thin-bedded turbiditic sedimentation of the Ropianka-type facies interrupted by hemipelagic deposition of complexes of variegated shales took place in the Magura Basin. Thick bodies build from sandstones with marginal amount of shales occur within these deposits. These sandstones are usually porous, mainly thick- to medium-bedded, fine- to very coarse grained and conglomeratic, generally massive with normal gradation. They are effect of the relatively short-term sedimentary events connected with delivery of big amount of clastic material to the external part of the Magura Basin. Within the Siary Zone in the Upper Cretaceous deposits, occurs one sandstone body – the Mutne sandstone; in the Early Paleogene occur the other two bodies – Żurawnica and Skawce sandstones.

The Mutne Sandstone up to 150 m in thickness is the Late Maastrichtian (uppermost part of Rzehakina inclusa foraminiferal zone) or the Late Maastrichtian–Early Paleocene in age (Cieszkowski *et al.*, 2007). It occurs in western and central part of the Magura Nappe, it was also noticed in the eastern part of the Beskid Niski range in Gorlice area (e.g. Kopciowski *et al.*, 1997a, b). Moreover, within the Upper Cretaceous deposits of the Siary Zone, the deposits with high share of the sandy deposits occur. Below the Mutne Sandstone occurs the Jaworzynka Sandstone. It is composed of the 100–150 m thick deposits of few to over a dozen m thick- and medium bedded sandstone complexes of characteristic arkosic sandstones with significant amount of muscovite and biotite divided by thin-bedded sandy-shaly deposits (Burtan, 1973). The Jaworzynka sandstone is Campanian-Maastrichtioan in age.

The Early Paleogene Żurawnica and Skawce sandstone complexes are lithostratigraphic units in rank of members within the Łabowa Shale Formation, which is developed as variegated shales. The Żurawnica Member up to 200 m in thickness is Late Paleocene in age (upper part of the Rzehakina fissistomata foraminiferal zone) (Cieszkowski & Waśkowska, 2011). It occurs in the eastern part of the Beskid Mały range and western part of the Beskid Makowski range and have relatively short geographic range, limited to the central part of the Magura Nappe. Younger body – the Skawce Sandstone Member up to 150 m in thickness is the Early Eocene in age (Saccamminoides carpathicus foraminiferal zone) (Cieszkowski & Waśkowska, 2011). It occurs in western and central part of the Magura Nappe, stretching from Żywiec up to Myślenice area.

The described above sandstone bodies can serves as such reservoir rocks. The Early Paleogene Żurawnica and Skawce sandstones look most promising. The variegated shales of the Łabowa Shale Formation can provide good seal, another component of the petroleum system. Natural gas seeps are observed in Beskid Makowski area. The source rocks are located probably within underlying facies-tectonic structures.

The deposits of the Outer Carpathian Basins were folded and thrust over each other and over the North European Platform during Miocene times (Alpine Orogeny), forming north-verging nappes. The Magura Nappe was thrust over the Foremagura group of nappes and the Silesian Nappe. The lower nappes contain organic-rich shales of the Oligocene Menilite Formation (Ślączka *et al.*, 2006). According to Kotarba & Koltun (2006) the Total Organic Carbon (TOC) content ranges from 0.18 to 17.25% (mean 4.48%) in the Menilite Formation deposits occurring now within the Silesian Nappe. These organic-rich shales can serve as the source rocks, components of the petroleum system. The numerous deep-reaching faults provide the migration paths. The hydrocarbon migrated to the potential reservoir rocks of the Magura Nappe. Additional probably source of hydrocarbons could be placed within the Magura Nappe. Between sandy deposits Late Cretaceous and Paleocene in age, the thin-bedded shaly-sandy deposits occur. The quarzitic sandstones are divided by complexes of muddy or marly shales green, grey and brown in color with average amount of TOC about 0,3–0,5%, but in the dark interbeddings the TOC reaches above 2.5%. Locally the dark-colored deposits prevail, that's situation is in Gorlice area where the black facies of the Ropianka Formation are distinguished as Świątkowa Member (*sensu* Jurkiewicz & Karnkowski, 1959).

This research has been financially supported by National Centre for Research and Development Blue Gas Program (BG2/ShaleCarp/14) and Statutory Founds of KGOiG WGGiOŚ AGH No. 11.11.140.005.

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SEDIMENTARY CHARACTERISTICS AND DETRITAL ZIRCON AGES OF THE NANDUAN AND LABA FORMATIONS IN THE CHANGNING-MENGLIAN BELT, WESTERN YUNNAN, CHINA

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The Changning-Menglian Belt, which extends in north-south direction in western Yunnan, China, has been considered to be a boundary between Cathaysia-affinity blocks to the east and Gondwana-affinity blocks to the west during the Paleozoic (Liu *et al.*, 1991; Jin, 1996; Jin *et al.*, 2003; Metcalfe, 2013). Late Paleozoic clastic sediments record information to understand the tectonic evolution of this belt, thus have received considerable attention (Jia, 1994; Cui *et al.*, 1998). However, the tectonic setting, sedimentary characteristics and provenance of these clastics remain controversial. The purpose of this study is to elucidate this issue via fieldwork, petrological and geochemical analysis, and detrital-zircon geochronology.

The studied clastic sequences are mainly distributed in the east part of the Changning-Menglian Belt and comprised of the Nanduan and Laba Formations. The Nanduan Formation is composed of thick-bedded quartzose sandstone, siltstone and mudstone with an age of (latest Devonian to) Carboniferous. The lower part of this formation adjacent to the Lincang Granite which is distributed to the east of this belt is metamorphosed to phyllite or schist. Grain size of the sandstone decreases and mudstone content increases from the lower part to the upper part. These sandstones are of high compositional maturity on the one hand and low textural maturity on the other hand. Framework grains are mainly quartz. Some of the quartz grains are well-rounded, while others are angular to subangular. Grain-size frequency diagrams exhibit double peaks. Thin sections show that dissolution, overgrowth and pressure solution of the quartz particles are developed in the sandstones of Nanduan Formation (Fig. 1). A transitional contact between upper part of Nanduan Formation and lower part of Laba Formation can be recognized at Banjiao village. The lower part of Laba Formation is comprised of purplish-red and yellowish-green siltstone and mudstone, and some fine-grained sandstone beds. Cherts are intercalated upwards in the Laba Formation and yield Permian to Triassic radiolaria. Limestone lenses also occur in the middle-upper part of Laba Formation. The overall sedimentary feature show that the Laba Formation was deposited in deeper water than Nanduan Formation was. In conjunction with these petrologic features, geochemical analysis of the clastic rocks in the Nanduan and Laba Formation shows that the detritus were derived from tectonic highlands of interior craton and deposited in passive margin environment.

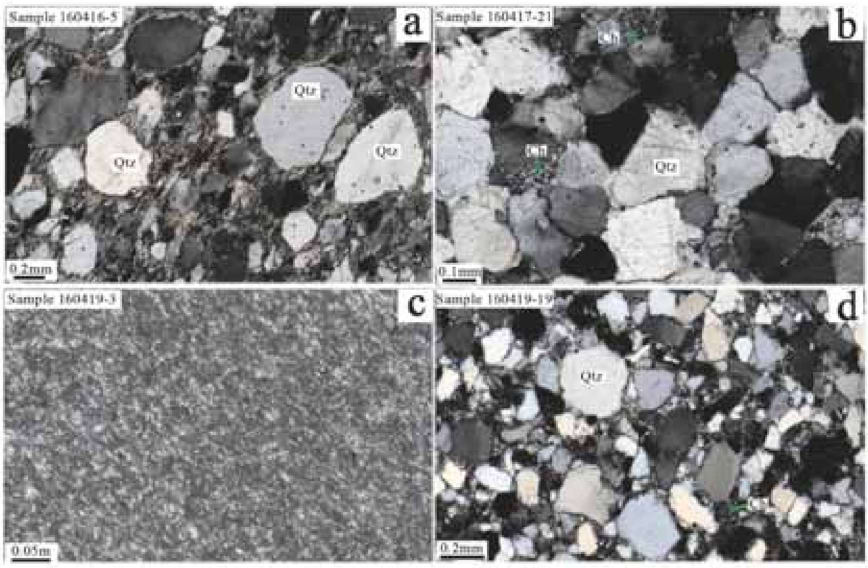

Fig. 1. Photomicrographs of sandstones from the Nanduan Formation Qtz-Quartz; Ch-chert;

Moreover, the detrital zircons from the Nanduan Formation were dated by LA-ICP-MS to range from 362 to 3685 Ma. Major peak ages at 554.5 Ma and 952.25 Ma could be further recognized (Fig. 2). The Grenvillian age (~950 Ma) and Pan-African age (~550 Ma) indicate that Changning-Menglian Belt probably shares similar provenance with Baoshan Block and is Gondwana-affinity. Detrital zircons from Juras-

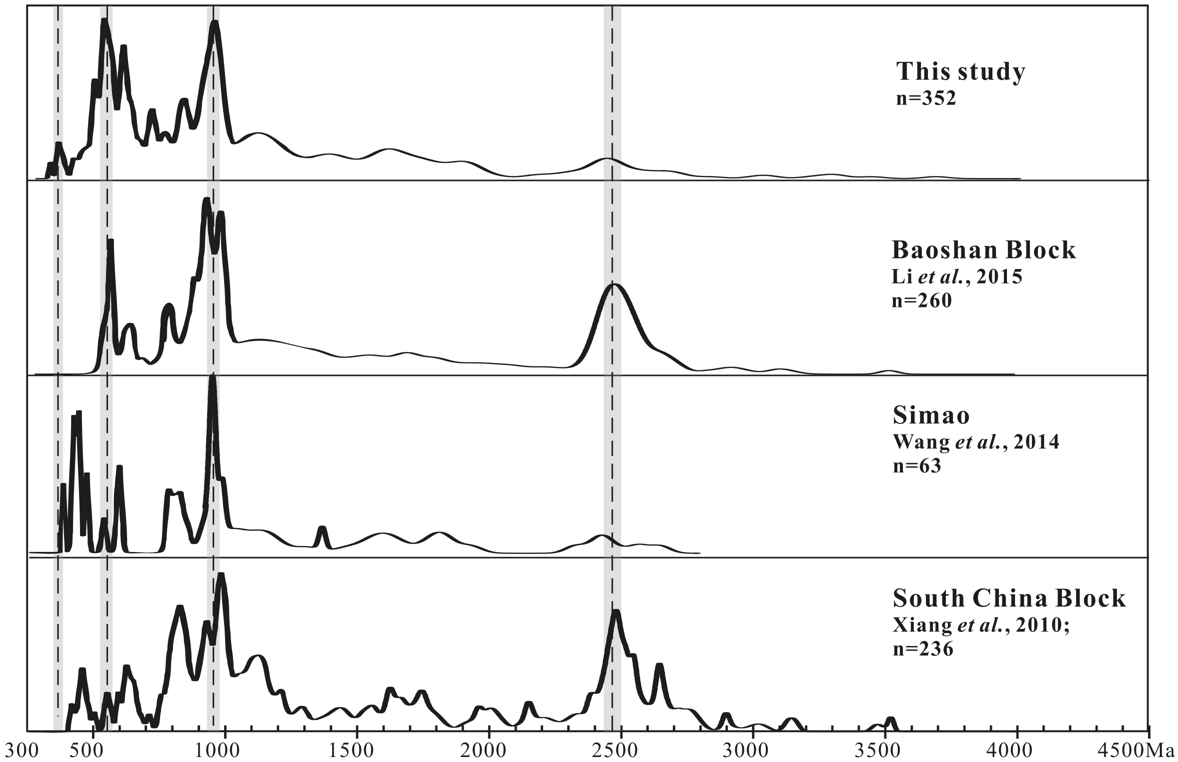

Fig. 2. Comparision of detrital zircon age distribution of the Changning-Menglian Belt, Baoshan Block, Simao Block and South China Block.

sic Huakaizuo Formation demonstrate with a peak age 220 Ma, hence indicate the detrital provenance from the Lincang granites which is considered a product of the closure of the Tethys represented by the Changning-Menglian Belt.

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KRAKÓW – URBAN – GEOLOGY

TRADITIONAL DECORATIVE AND BUILDING STONES OF KRAKÓW ARCHITECTURE

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Kraków is an excellent example of geological valorization. Architectural stones from the nearest vicinity of the town have been used for one thousand years of its history. They were applied as building and decoration stones, in sculptures, sepulchral art and pavements. These rock materials applied in the architecture of Kraków have shaped for centuries a characteristic stone landscape of the town. Unfortunately, within the span of the last years we face a dramatic change in the stony architectonic inventory of Kraków. Due to political and economic transformations, mass import of decorative stones has begun almost from all over the world. It must be stressed that such unfamiliar stones destroy and falsify the historic landscape of Kraków.

Kraków is situated at the junction of several geological units of a regional rank, differing in their structure. Initially, the stones were recovered from the nearest neighbourhood of the town, mainly from the Silesian-Kraków Monocline, the Carpathian Foredeep and from the Outer Carpathians, while few stone varieties from the Pieniny Klippen Belt (PKB) and Inner Carpathians. Other rocks were brought from more distant localities and other geological units: the Holy Cross Mts proper and their margin, the Ukrainian shield – today outside the Polish borders, from the Sudetes and the Fore-Sudetic Monocline. In the historic architecture of Kraków there are also stones imported from outside Poland, mostly from neighbouring countries, e.g. Hungarian marble. They were used for special purposes, for instance in interior designing or in sepulchral stonework.

To the most important rock raw materials, with the longest tradition of utilization in the architecture of Kraków, come from the Silesian-Kraków Monocline. Following the frequency of their occurrence in the architecture of Kraków we found: Jurassic and Devonian limestones, Triassic dolomites, Permian porphyry and Carboniferous "marble" (Figs 1, 2).

Upper Jurassic (Oxfordian) white limestone, commonly with epigenetic flints, belongs to the oldest and the most important rocks utilized in the architecture of the city, commonly applied in the form of irregular blocks and regular elements. These stones were used in erecting city fortifications, in defensive walls of the burgher Kraków and the royal Wawel (Stop 7), and also paving streets and surface of the Main Market Square, unfortunately removed in 1964. The apogee of the application took place in the Romanesque and Gothic times as small and big blocks mainly in constructing walls of sacral buildings e.g. the Wawel Cathedral, basilicas of St. Mary's, St. Catherine's, St. Trinity's, Corpus Christi, St. Andrew church (Fig. 1A) and also City Hall Tower.

Middle Triassic *Diplopora* **dolomite** has been used in the architecture of Kraków from the 14th century. It is the yellow-brownish rock, with a characteristic irregular porosity formed in the eogenetic stage of diagenesis. It is usually the rock with a micritic texture with stromatolite structures, flints, fragments of rock-forming algae *Diplopora* sp., trochites of crinoids, and also trace fossils. It is a building stone but also used for decorative purposes in slab facings, portals, window framings and floors. Beautiful examples of this dolomite stonework from the 17th century is the wall with gates bordering the Wawel cathedral (Fig. 1B) and the façade of the Church of St. Peter and St. Paul in Grodzka Street. This rock is also in the portal and fence of the building of the former clerical seminary at al. Mickiewicza 2 (Stop 4) and the façade of the National Museum facing the 3rd Maj Avenue (Stop 5). From this rock was also constructed in the beginning in the first half of the 20th century the largest "dolomitic investment" in Kraków the huge anti-flood hydrotechnic structures, situated along the Vistula River (Stop 8).

The Dębnik Devonian limestone, strongly lithified, usually described as a "marble", was quarried from at least 14th century. It occurs in a range of colour varieties, from black to grey; and very rare pink-, greenish- and even white-mottled formed by hydrothermal processes. Contains micritic texture, common wavy, bulging and nodular early-diagenetic structures and lighter in colour fossils,


A
B
C
D

Fig. 1. Kraków urban geology. **A** – Romanesque church of St. Andrew built of limestone Jurassic blocks (petit appareil) and Carpathian Istebna sandstone; the end of the 11th century; **B** – Gothic Wawel Cathedral from the 14th century. The walls built of the Jurassic limestone ashlars (grand appareil); the Baroque wall surrounding the cathedral and the gate constructed of the Triassic Diplopora dolomite date back to 1619; **C** – well-built of the black Dębnik Devonian limestone in the inner courtyard of Collegium Maius. The Collegium building dates back to the turn of the 14th century, the well comes from 1950s; **D** – portal of the St. Adalbert Church made of the black Dębnik Devonian limestone with marquetry of the Paczółtowice limestone. The church, about 1000 years old, is one of the oldest in Cracow; its late Baroque portal comes from the 18th century

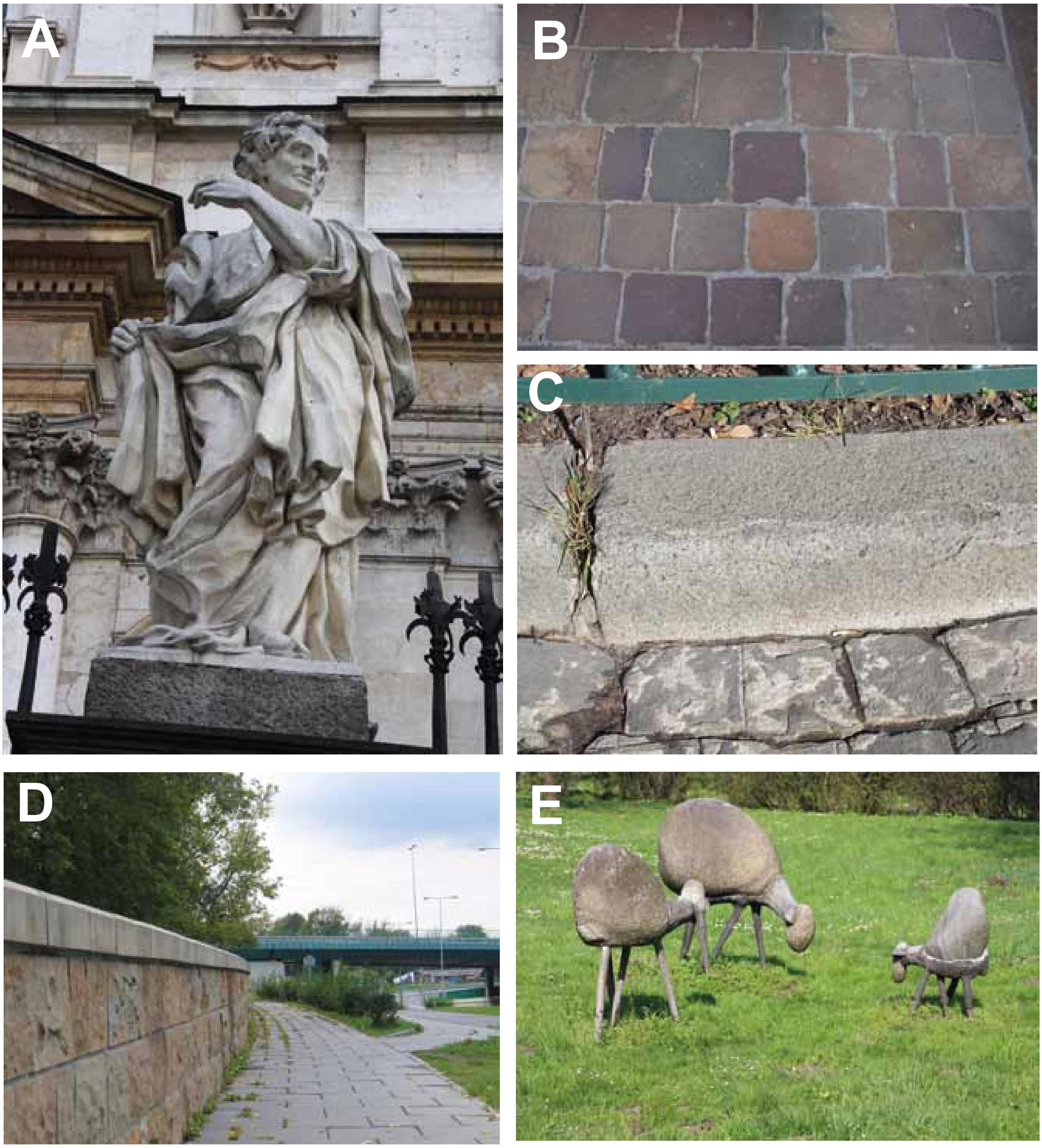

Fig. 2. Kraków urban geology. **A** – present-day replica of late-Baroque (beginning of the 17th century) sculpture of apostle, carved in the Pińczów Miocene limestone, placed at the front of the Church of St. Peter and St. Paul; **B** – sidewalk from the 19th century made of the Permian porphyry paving stone quarried in the Miękinia area; **C** – curbstone of the Miocene andesite along Oleandry Str; **D** – anti-flood hydrotechnic walls build with Istebna sandstone along the Vistula River; **E** – Tatra Carboniferous granite pebbles converted into the sculptures of sheep from 1960s “carved)” by Bolesław Chromy

mainly Hydrozoa, corals and trochites of crinoids. The Dębnik limestone owes its unusual popularity, extending over centuries in Poland and outside the country, due to its colour and capability of taking a beautiful "warm" mat polish (Fig. 1C). This black colour becomes unstable when is exposed to weather conditions. It was widely used in church interior architecture as: portals, floors, columns, stairs, and also altars, menses, baptismal fonts, memorial tablets and tombstones. A good example is the Kraków interior of St. Mary's Church and the Wawel cathedral, and outside the Poland is high altar of St. Stephen's cathedral in Vienna.

Lower Permian porphyries rocks intermediate between rhyolites and dacites, are utilized in Kraków since the middle of the 18th century as paving blocks, floor slabs, boles and curbs. The main violet variety comes from Miękinia whereas porphyries from Zalas counterpart changes from green-bluish to pale cherry-red to yellow-brownish. The rocks have porphyritic texture with phenocrystals of feldspars, biotite and quartz. Remnants of porphyries pavements can be seen in Small Market, at the front of the Wawel cathedral, on many streets (Fig. 2B) and in the courtyard of AGH US&T (Stop 1).

The Lower Carboniferous Paczółtowice "marble" are formed as grey-reddish limestone with veins of white calcite, whereas the other, called the "Polish onyx", as similar brecciated limestone rich in veins of pink and violet calcite. After polishing the Paczółtowice "marble" was mainly used as a stone marquetry component; in such a combination it has been the "inseparable companion" of the black Dębnik "marble". Such marquetries can be seen in the Baroque portals made of the black Dębnik limestone in St. Mary's Church and St. Adalbert's Church at the Main Market Square (Fig. 1D) and also on the façade of the Church of St. Peter and St. Paul in Grodzka Street.

The Carpathians and its Foredeep are another regions providing rocky resources characteristic of Kraków architecture. Belong to this group: granite from the Tatra Mts, andesite from the Pieniny Klippen Belt, varieties flysch sandstones from the Outer Carpathians, and Pińczów lithothamnium limestone from the Carpathian Foredeep.

The Lower Carboniferous granite of the Tatra Mts has minor but spectacular applications in the Kraków architecture. They occur mainly in the form of pebbles extracted from the rivers flowing from these mountains. About 2 millions of this pebbles decorate the walls of the Church of Our Lady Queen of Poland, called the Ark of Our Lord, which has been the first church of Nowa Huta built in communist times. Huge pebbles of appropriate naturally shapes have been used by a contemporary Kraków sculptor B. Chromy to create bodies stony animals (Fig. 2E); some of them are standing near the building of Agricultural University (Stop 2). Granite block with pegmatite veins was placed in 1860 at the top of the Tadeusz Kościuszko mound, but another big one with a weight of 26 tons, commemorates Pope St. John Paul II is standing close to the margin of the Kraków Błonie.

The Miocene andesite from the Pieniny Klippen Belt began quarrying from hypabyssal intrusions after the I World War, and was used as a building stone, pavement blocks and curb stones, also as tombstones and polished facing panels. In Kraków the andesite from the PKB was used mainly in 1930s as curb stones, exceptionally as paving stones. Polished with pedestrian's shoes, the andesite reveals characteristic porphyritic texture with grains of amphiboles, pyroxenes and plagioclase and rather numerous skialites and black xenoliths; the andesites are not mined at present. Such curb stones have been placed along the Ingardena Street (Fig. 2C) (Stop 3).

The Cretaceous–Tertiary sandstones from the Outer (i.e. Flysch) Carpathians were the first stone materials used in Kraków at the turn of the 10th and 11th centuries. Of the highest architectonic importance were polymictic and lithic Istebna and Godula sandstones, of the lesser the Ciężkowice, Krosno, Lgota and Magura sandstones. They were used in the oldest, Pre-Romanesque and Romanesque structures of the Wawel Hill, i.e. the so-called quadrilateral building and the Rotunda of Virgin Mary called as the Rotunda of St. Felix and St. Adauctus. Two Romanesque cathedrals had their walls, columns, column bases and capitels made of such sandstones, carefully worked to shape. In this time the Carpathian sandstone was a material complementary to the dominant limestone small blocks, used as cornerstones, portals and window frames. The best examples of such applications can be seen in the churches of St. Andrew, St. Adalbert and the Holiest Salvator and in St. Florian's Gate in Kraków fortifications. After a longer break, the Carpathian sandstone was used again on a larger scale in the Renaissance. The finest Renaissance chapels, the Sigismund and the Vasa chapels of the Wawel cathedral, were erected then of the greenish Godula sandstone and the yellowish Istebna sandstone, respectively.

The same sandstones were used again at the end of the 19th century, being more popular in the Kraków than they are today. Their numerous applications can be seen in burghers' houses, in industrial and railway constructions, road structures, in churches and fortifications. Currently, the Carpathian sandstones seldom represent building stones, being mainly used as a decorative material: as wall and floor slabs, either cut or broken to shape and with diversied surface texture. Good examples can be found in the façades of the Kraków-Balice Airport building, the new building of the Academy of Fine Arts and the building of the "Bagatela" theatre and new part of anti-flood hydrotechnic walls along the Vistula River (Fig. 2D) (Stop 6); all of them utilize the Istebna sandstone. Some parts of the façade of the Centre of Japanese Art and Technology in Konopnicka Street have been made of the grey-blue Krosno sandstone.

Miocene Pińczów limestone is a white phytogenic calcarenite or calcilutite with the calcite contact cement and a high porosity. It formed from detrite of the *Lithothamnium* sp. and fragmented calcareous skeletons of other marine organisms, with a subordinate admixture of quartz grains and clays on the northern border of Carpathian Foredeep. These limestones have found their application as building stones, facing slabs, in portals, cornices, etc., but first of all they represent an indispensable stone for sculpting, in which excellent sculptors executed their masterpieces, forming the cultural heritage of Kraków, spread all over the city. During dressing it does not take polish. The oldest, currently active region of its quarrying is situated in the Pińczów area. An excellent Renaissance sculptor and architect Jan Maria Padovano curved in this stone ornamental masks of the Cloth Hall attic. Regarded the most beautiful in Poland is the attic of the Boner family mansion in the Main Market Square, curved by Santi Gucci. The same stone was used earlier by Wit Stwosz to curve the figures of the Gethsemani chapel of St. Barbara's Church and a huge crucifix in St. Mary's Church. In the Wawel cathedral there are also 16th-century Renaissance canopies over the sarcophagi of kings Ladislaus Jagiełło (carved by Jan Cini of Siena) and Casimir the Great (carved by an unknown artist). A younger monument of twelve apostles standing at the front of St. Peter and St. Paul's Church in Grodzka Street (Fig. 2A), carved in 1715 by David Heel. Actually, Pińczów limestone is used mainly as outdoors facing slabs. Good example is the elevation of the House under Baszta on the Skawińska Street (Stop 9).

The stones used in the architecture of Kraków for a thousand years were quarried mostly from deposits situated not far from the town. Over hundreds of years they have shaped a unique architectonic image of the former capital of Poland. A significant role have been played by the described rock materials from Silesian-Kraków Monocline the Carpathian and the Carpathian Foredeep. Unfortunately, incompetence of the present town officials has resulted in an uncontrolled import of foreign stones, indiscriminately used in the strictly historical part of the city, in the places least suitable for such a purpose.

FIELD TRIP

THE POSITION OF THE WEST CARPATHIANS IN THE ALPINE-CARPATHIAN FOLD-AND-THRUST BELT

▸FIGS 1–23

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The Polish and Slovak West Carpathians form the northrn 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*., 2005; Ślączka *et al.*, 2005):

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. The nappes and the Hercynian basement are uncoformably covered by mid-Eocene/ Oligocene flysch and Early–Middle Miocene marine and terrestrial (continental) molasses. Another terrane with the Hercynian Basement known as Tisia–Dacia is amalgamated with the Inner Carpathian Terrane. According to Golonka *et al*. (2000) Inner Carpathian, Eastern Alps and Tisia-Dacia form Alcapa superterrane, according to *e.g.* Kováč *et al*. (1998) Alcapa and Tisa constitute different plates. The Jurassic rocks within the Inner Carpathian terrane are represented by platform and basinal deposits.

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* ar represented by mainly paltform facies.

These allochtonous rocks are uprooted and overthrust onto the southern part of the North European Platform at a distance of at least 60–100 km (Książkiewicz, 1977; Oszczypko & Ślączka, 1985). They form stack of nappes and thrust-sheets arranged in several tectonic units. In Poland these allochtonous mainly flysch units are being regarded as Flysch Carpathians. Along the frontal Carpathian thrust, a narrow zone of folded Miocene deposits was developed – Zgłobice Unit in the Polish Carpathians, Sambir Unit of the Ukrainian part of the Carpathians and the Subcarpathian Unit in Romania. The allochthonous Upper Jurassic rocks are represented by basinal, slope, rarely by ridge and platform facies.

Penninic realm is a part of the Alpine Tethys (*e.g.*, Birkenmajer, 1986; Săndulescu, 1988; Oszczypko, 1992; Plašienka, 1999, 2002; Stampfli, 2001; Golonka, 2005), which developed as a basin during Jurassic time between Inner Carpathian–Eastern Alpine terrane and North European Platform. In the western part it contains the ophiolitic sequences indicating the truly oceanic crust. In the eastern part the ophiolitic sequences are known only as pebbles in flysch, the basement of the Penninic realm was partly formed by the attenuated crust. In Poland, Slovakia and Ukraine the Penninic realm is represented by the sedimentary sequences of Jurassic, Cretaceous, Paleogene and Miocene age belonging to the Pieniny Klippen Belt 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 Pieniny Klippen Belt, 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 Pieniny Klippen Belt could be also regarded as belonging to the Outer Carpathians (*e.g.*, Książkiewicz, 1977; Picha, 1996). The Czorsztyn submerged ridge was a part of the Penninic realm dividing the oceanic basin into two subbasins. The southern subbasin and the ridge traditionally constitute the Pieniny domain. Its sequences are involve in the Pieniny Klippen Belt – 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 Pieniny Klippen Belt 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.

Transilvanian Domain formed perhaps the extension of the eastern Tethys between the European platform and *Tisia–Dacia terrane* (Săndulescu, 1988; Săndulescu & Visarion, 2000). In Romania it was truly oceanic realm as indicated by the existence of ophiolites. It developed during the Triassic and was closed during the Cretaceous time. In Poland, the presumed *Transilvanian Domain* sedimentary sequences are represented only by pebbles in the flysch of Pieniny Klippen Belt and Magura Unit. The Jurassic rocks in these pebbles represent the platform, slope and basinal facies.

Severin–Moldavidic realm (SM) (Balintoni, 1998), also known in Romania as Outer Dacides and Moldavides (Săndulescu, 1988) developed within the North European Platform as rift and/or back-arc basin. The SM basement is represented by the attenuated crust of the North European plate with perhaps incipient oceanic fragments. The sedimentary cover is represented by several sequences of Late Jurassic–Early Miocene age belonging recently to Dukla, Silesian, Sub-Silesian, Zdanice, Skole tectonic units in Poland and Czech Republic (Pescatore & Ślączka, 1984; Żytko *et al*., 1989; Ślączka & Kaminski, 1998; Stráník *et al*., 1993). The subbasins within the SM are divided by ridges and uplifted zones. The most prominent one is the Sub-Silesian Buldge between Silesian and Skole Sub-Basins. Another ridge was located between Dukla and Silesian Subbasins. The Sub-Silesian Bulge is in the West connected with the slope sequences between epicontinental part of the North European Platform and outer Carpathian basins (Silesian and Magura basins). The Severin-Moldavidic basinal realm ends in Moravia while slope sequences extend further westwards. The Severin part (Severinides, see *e.g.,* Balintoni, 1998, 2001 or Outer Dacides, Săndulescu, 1988) of the basin is represented in Ukraine by Kaminnyi Potik and Rahiv units. This basin was closed during the Cretaceous time. The Moldavic part (Moldavides, see *e.g.,* Săndulescu, 1988; Balintoni, 1998; Golonka *et al*., 2005) was closed during the Neogene time. The Jurassic rocks of this realm are known from the Silesian unit area. They are represented by basinal flysch facies.

The Getic-Marmarosh Ridge (Golonka *et al*., 2003, 2005) also known as Median Dacides (Săndulescu, 1988) constitutes a fragment of the North European Platform rifted away during the opening of the Severin–Moldavidic basins. It includes Precambrian, Early Paleozoic (Caledonian) granites and metamorphic rocks, Late Paleozoic (Variscan) metamorphic rocks as well as the late Paleozoic and Mesozoic sedimentary cover. The Getic-Marmaros Ridge separated the Severin-Moldavidic basin from the Transilvanian Basin. Westward the similar position has the Silesian Ridge (Săndulescu, 1988; Oszczypko, 1992), which separated Penninic–Magura realm and the Severin-Moldavidic realm. The eastern part of the Getic-Marmaros Ridge collided in latest Early–Late Cretaceous with the Tisia-Dacia forming several nappes. These nappes also included part of the Civcin-Severin-Moldavidic basins – Rahiv and Porkulets units. The western part of the ridge was reorganized during Late Cretaceous forming the basement of the uplifted Silesian Cordillera, Fore-Magura Basin, a ridge separated Magura Basin and Fore-Magura Basin as well as marginal part of Magura Basin. In Poland and Slovakia the Silesian and Fore-Magura riges sedimentary sequences are represented only by pebbles in the flysch of Pieniny Klippen Belt and Magura Unit. The Jurassic rocks in these pebbles represent mainly the the platform facies known as Štramberk limestones.

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

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

the North European Platform and its Miocene–Paleocene 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.

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 Peritethys.

The Alpine Tethys constitutes important paleogeographic 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 and for Oligocene times – Figs 5–10; then several palaeogeographical skechtes for circum-Carpathian area are shown on Figs 11–20). 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 Pieniny Klippen Belt – a strongly tectonized structure about 800 km long and 1–20 km wide, which stretches from Vienna in the west to the Poiana Botizii (Maramures, NE Romania) in the east (Fig. 2). The largest part of the northern sub-basin forms the Magura Unit, traditionally taken as belonging to the Outer Carpathians.

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.

The Carpathians form a great arc of mountains, which stretches more than 1 300 km from the Vienna Forest to the Iron Gate on the Danube. On the west the Carpathians are linked with the eastern Alps and on the east pass into the Balkan chain. Traditionally the Carpathians are subdivided into the West and East Carpathians (Mahel, 1974). The West Carpathians consists of an older range known as the Inner or Central Carpathians and the younger one, known as the Outer or Flysch Carpathians (Mahel, 1974; Książkiewicz, 1977; Ślączka, Kamiński, 1998). The Inner Carpathians are regarded as a prolongation of the Northern Calcareous Alps, and formed part of the Apulia plate in regional sense that is a promontory of the African plate

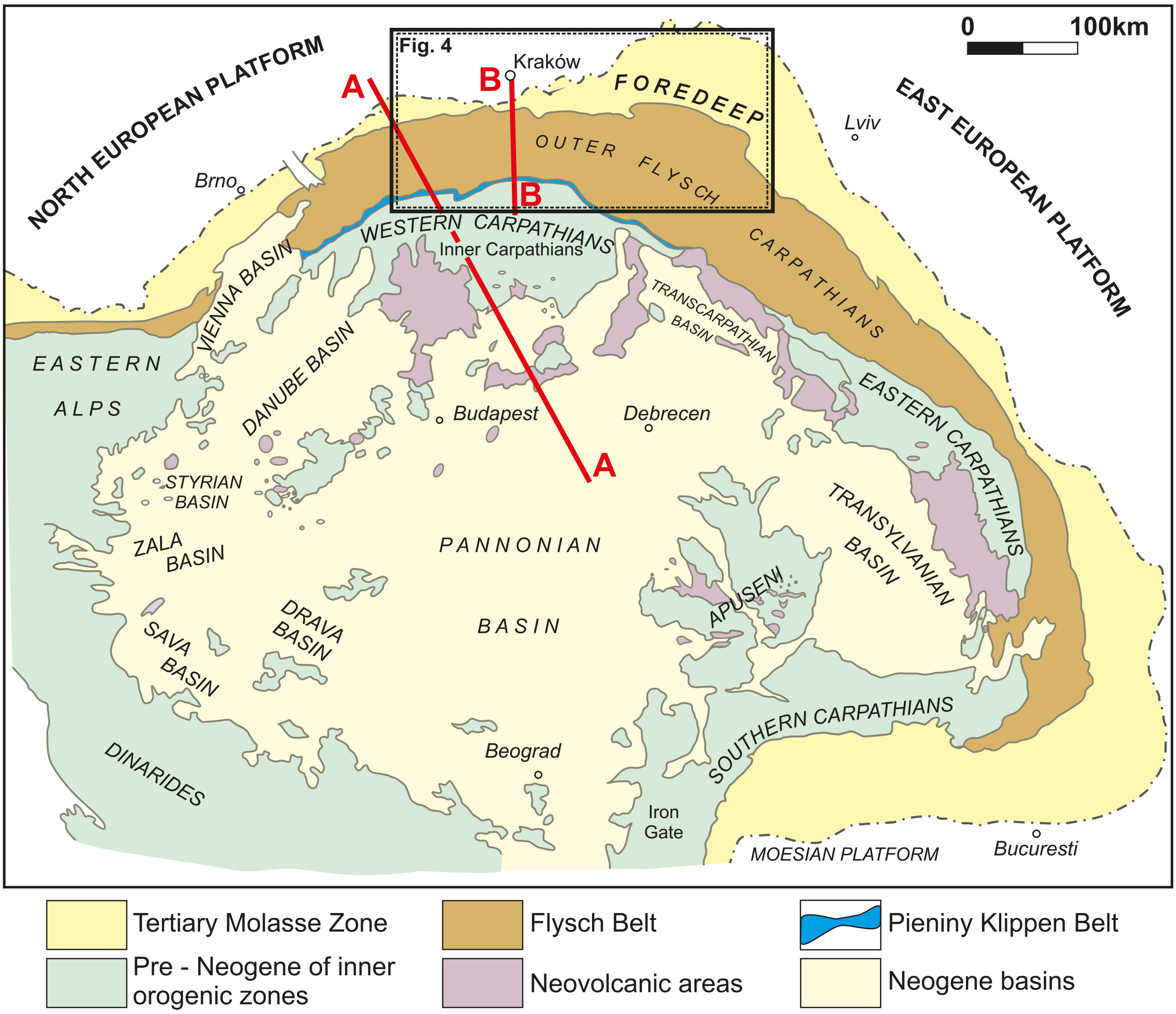

Fig. 2. Tectonic sketch map of the Alpine–Carpathian–Pannonian–Dinaride basin system (modified after Plašienka *et al.*, 2000). A–A and B–B – localization of cross-sections (see Fig. 3)

(Picha, 1996). Sedimentation in the Inner Carpathian was mainly calcareous, and took place from the Early Triassic to the mid Cretaceous. The Inner Carpathians were folded during the Late Cretaceous tectonic movements.

OUTER FLYSCH CARPATHIANS

The Outer Flysch Carpathians are built up of a stack of nappes and thrust sheets spreading along the Carpathians, built mainly of up to six kilometers thick continual flysch sequences, representing the time span from Jurassic to Early Miocene. All the Outer Carpathian nappes are overthrust onto the southern part of the North European platform covered by the autochthonous Miocene deposits of the Carpathian Foredeep on the distance of 70 km at least (Książkiewicz, 1977; Pescatore & Ślączka, 1984). 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. A narrow zone of folded Miocene deposits was developed along the frontal Carpathian thrust i.e. Zgłobice-Wieliczka Unit in the Northern Carpathians and its equivalent Subcarpathian (or Borislav or Sambor-Rozniatov Unit of the Ukrainian part) and Romanian parts of the Eastern Carpathians (Książkiewicz, 1977; Pescatore & Ślączka, 1984; Kruglov, 1989; Jankowski, 2015). 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), Sub-Silesian Nappe, Silesian Nappe, Fore-Magura group of nappes and Magura Nappe. We discuss here main units only.

The **Sub-Silesian 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 nap-

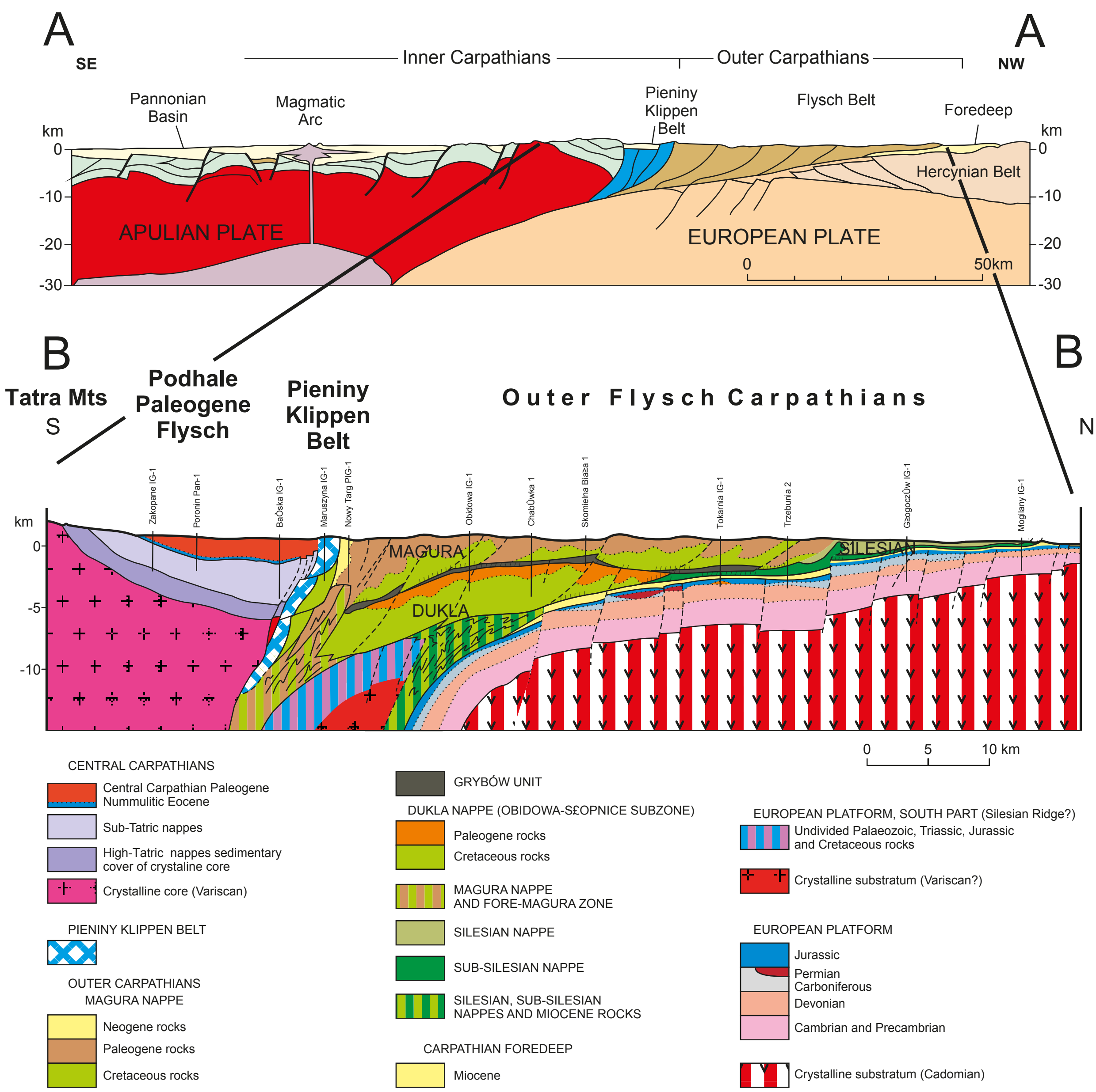

Fig. 3. Generalized cross-section across Carpathian-Pannonian region (Picha, 1996) (upper) and generalized cross-section across Polish Carpathians (after Golonka *et al.*, 2005) (lower)

pes. 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 (Vendryně Formation) which begin euxinic cycle that lasted without major interruption till Albian. The Silesian and Sub-Silesian 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 Pieniny Klippen Belt 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 Pieniny Klippen Basin (i.e., the Grajcarek Unit) (Birkenmajer, 1977).

The Outer Carpathian rift had developed with the beginning of calcareous flysch sedimentation (so-called Cieszyn Limestone Formation). 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*

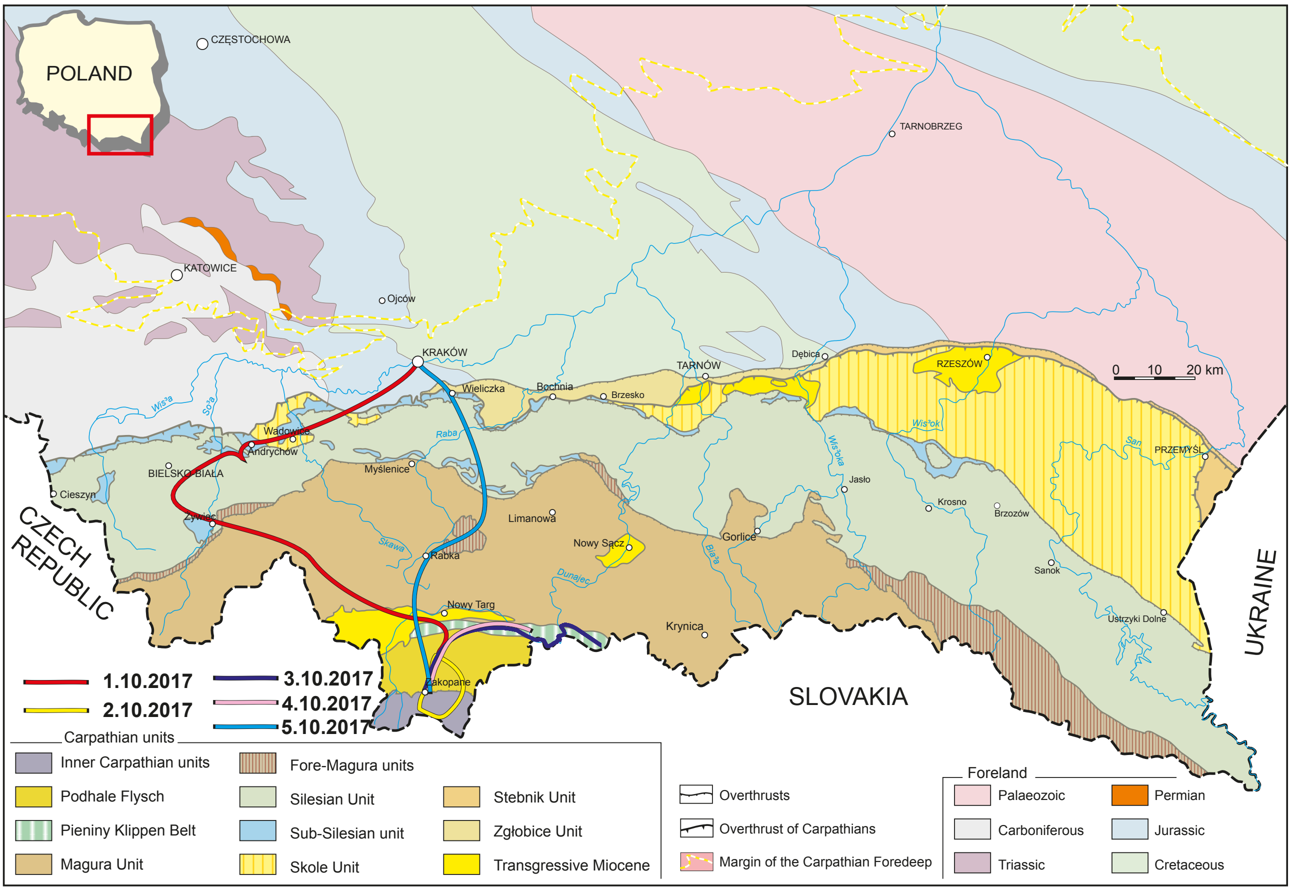

Fig. 4. Geological map of the Polish Carpathians and Foreland with route of the field trip (after Żytko *et al*., 1989; simplified)

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).

The **field trip starts in Kraków AGH University of Science and Technology** parking lot and leads southward to the Carpathians. In Kraków and its vicinity the Mesozoic rocks of the North European platform are exposed. The

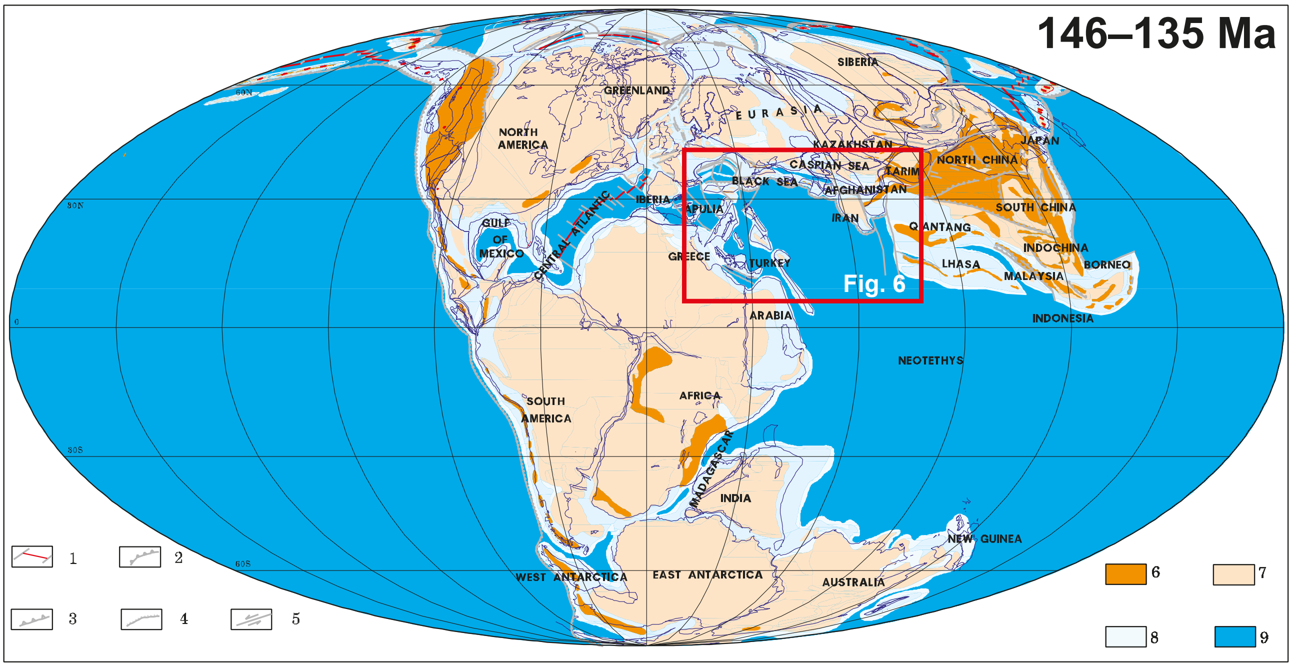

Fig. 5. Global plate tectonic map of latest Jurassic–earliest Cretaceous (from Golonka, 2000; modified)
Explanations: 1 – oceanic spreading center and transform faults; 2 – subduction zone; 3 – thrust fault; 4 – normal fault; 5 – transform fault; 6 – mountains; 7 – landmass; 8 – shallow sea and slope;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

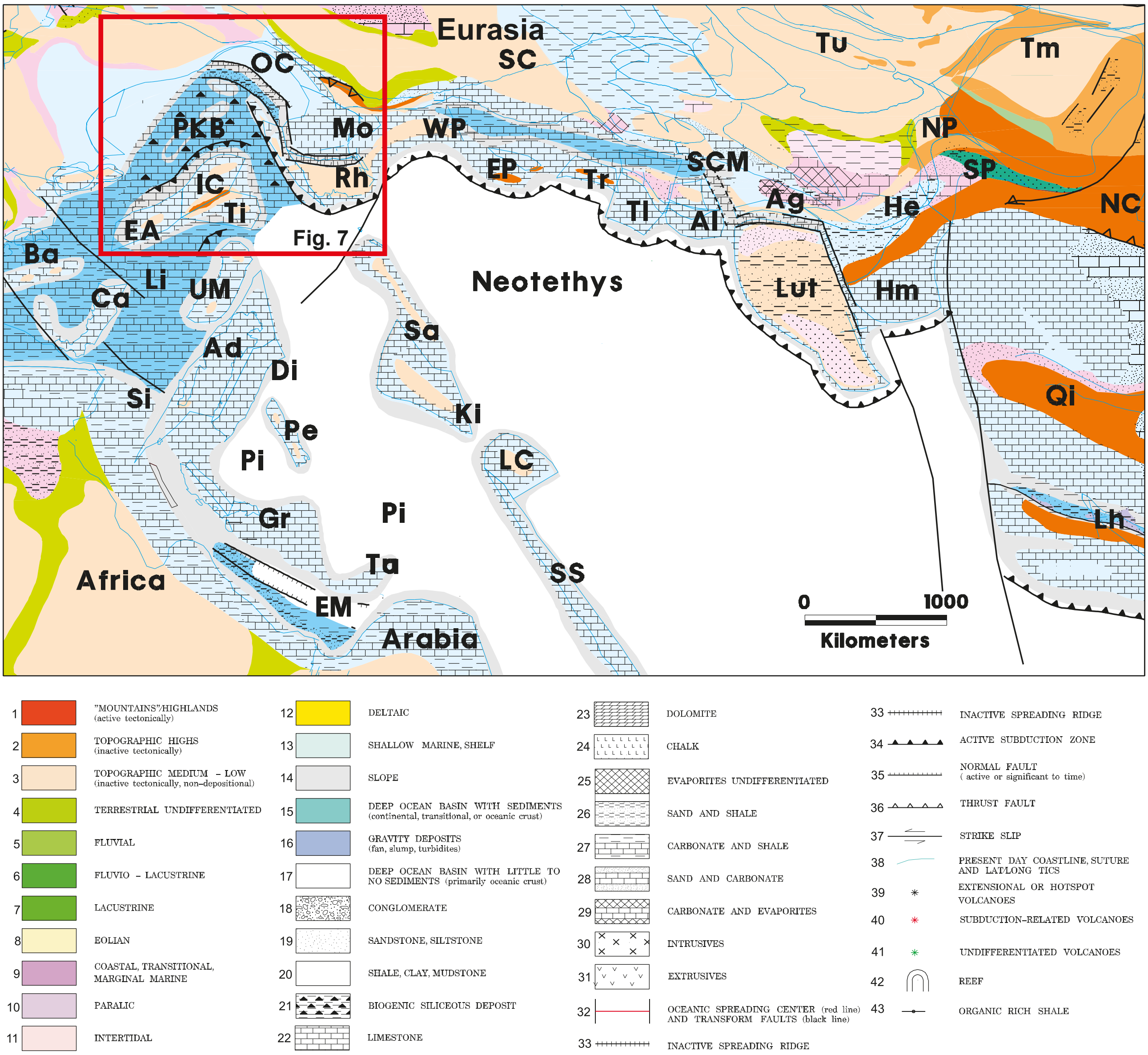

Fig. 6. Plate tectonic, palaeoenvironment and lithofacies map of the western Tethys, Central Atlantic and adjacent areas during latest Jurassic–earliest Cretaceous time (after Golonka, 2007a; modified)

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; PKB – 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

platform is dissected by numerous falts into several horsts and grabens. The grabens are filled with the Miocene Molasse deposits, while horsts elevate the Upper Jurassic rocks. This rocks are represented mainly by Oxfordian sponge buildups with associated nodular, chalky and micritic limestones. Passing the bridge no Wisła river we can observe the hill of Wawel with the Polish Royal Castle on the top. The Royal Castle was built in X century and remodeled several times. The most important remodeling was done by Queen Bona and her team of Italian architects in XVI century giving the castle its Renaissance character.

The Wawel hill is built by the white-weathering Upper Jurassic massive limestones. These limestones are horst elevated and shaped by karst phenomena. There is a cave inside the hill's rocks known as Smocza Jama. The name, which means Dragon Den, is derived from the Early Medieval legend about the dangerous dragon that terrorized the population of ancient Kraków. The dragon was killed by young shoemaker Scuba, who fed the beast with sheep filled with sulfur and other chemicals. Today this dragon is a symbol of Kraków and present in every souvenir boot. Shoemaker Skuba is rather forgotten, perhaps because

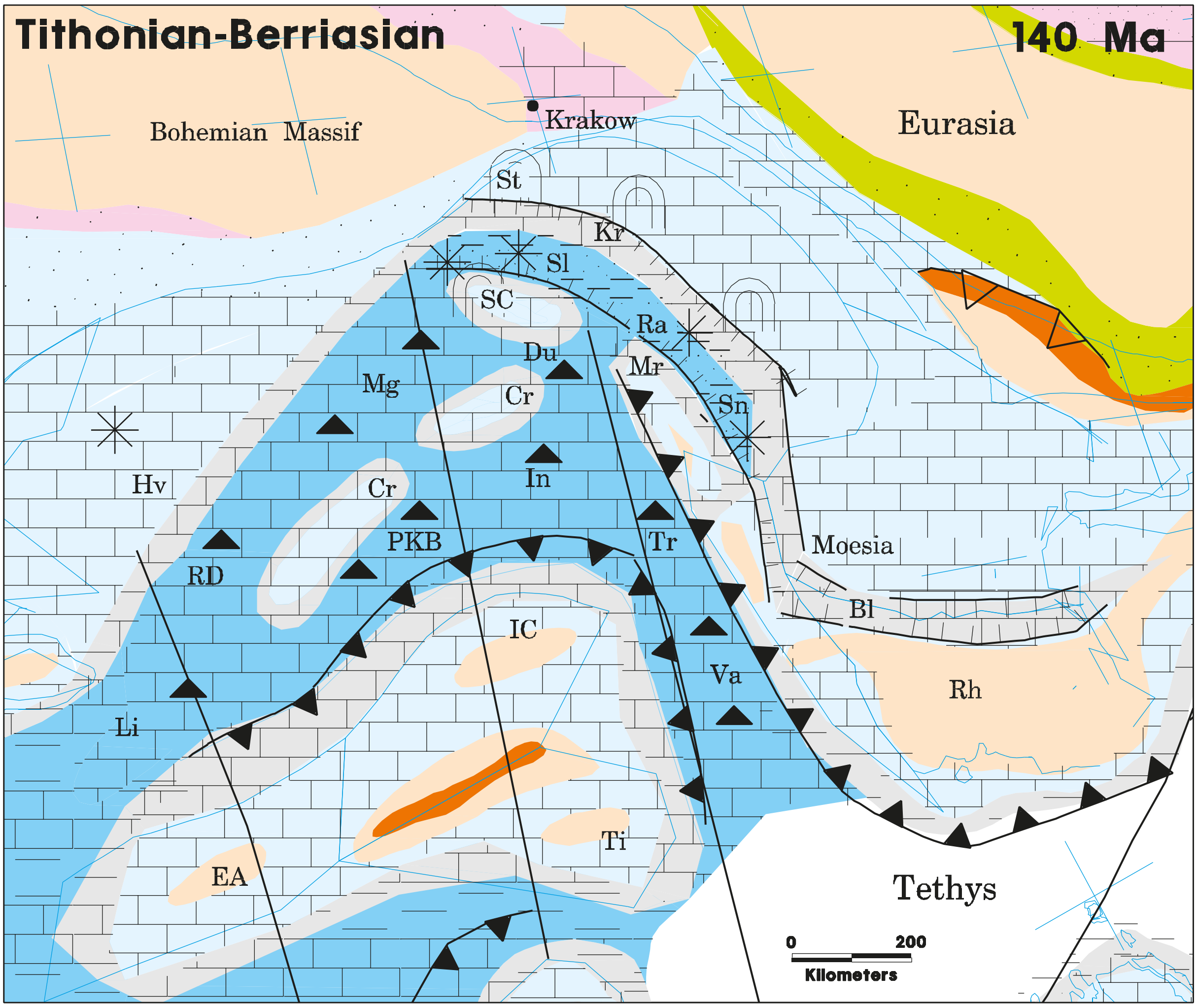

Fig. 7. Palaeoenvironment and lithofacies of the circum-Carpathian area during latest Jurassic–earliest Cretaceous; plates position at 140 Ma (modified from Golonka *et al*., 2006)
Abbreviations: Bl – Balkan rift; Cr – Czorsztyn Ridge; Du – Dukla Basin; EA – Eastern Alps; Hv – Helvetic shelf; IC – Inner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

of his antiecological usage of chemicals against extinct species. Following southwards the road crosses the Carpathian Foredeep dilled with Miocene molasse deposits. Springs of hydrosulphuric mineral waters are connected with the Miocene deposits. 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. Crossing the Sub-Silesian and Silesian Nappes route is reaching Magura Nappe within Myślenice town limits. From Myślenice the road is leading up the Raba river valley crossing the thrust-faults of the Magura Nappe. The morphology of the outer zone of the Magura Nappe, known as the Siary and Racza subunits, forms here isolated forested mountains of the Beskid Wyspowy (Island Beskid). During good weather the beautiful panorama of the Tatra mountains with highest Gerlach peak (2625 m) is visible and on the west an isolated cone of the Babia Góra Mt (1725 m), the highest mountain of the Polish Outer Carpathians. Between Gorce and Tatra Montains the Podhale region is situate. Podhale region is built of Central Carpathian Paleogene flysch deposits known in Poland as Podhale Flysch. General cross-sections trough this part of the Carpathians are presented on figures (Figs 21, 22) with lithostratigraphical scheme of the Carpathians (Fig. 23) with location of our outcrops (stop points).

In the vicinity of Andrychów, the route crosses the Outer Carpathian Main Thrust. The Outer Carpathian nappes are thrust over the Carpathian Foredeep Neogene rocks in this area. This part of the Outer Carpathians is known as the Beskid Mały Mountains. These mountains, exceeding 900 m in height, are built mainly of Cretaceous and Paleocene proximal flysch with thick sandstone bodies belonging to the Godula and Istebna formations. The first geological stop is located 5 kilometers south of Andrychów in the village of Rzyki.

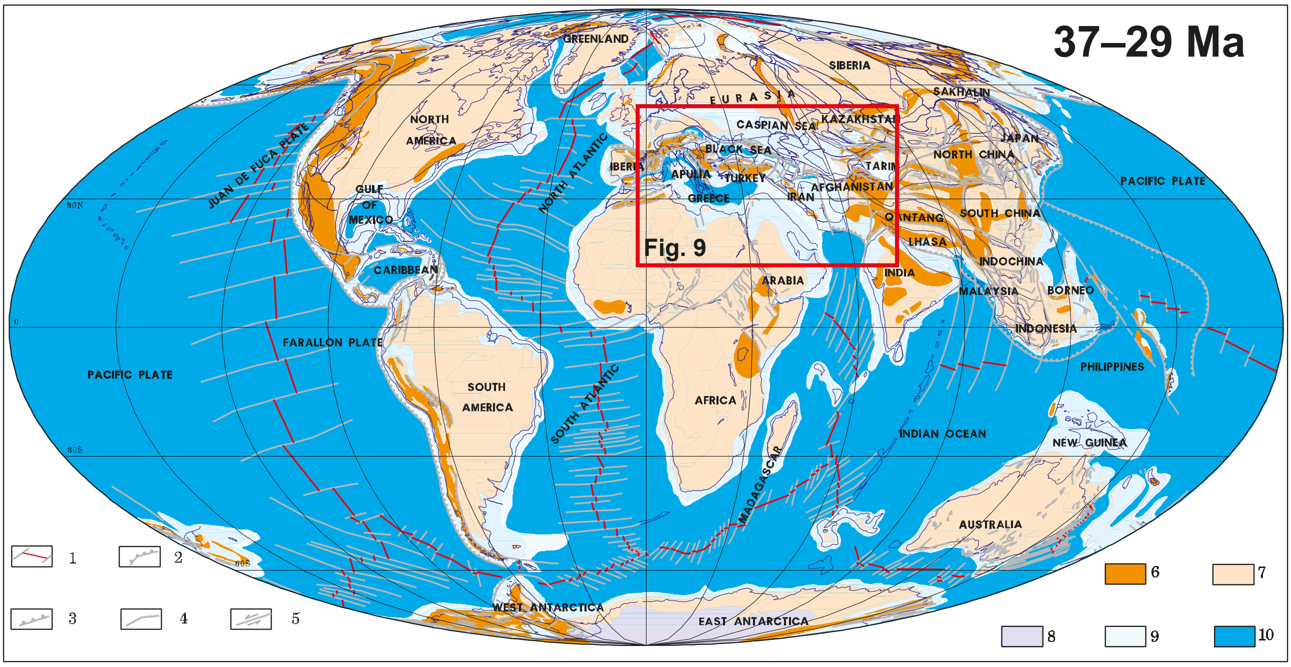

Fig. 8. Global plate tectonic map of the Oligocene (from Golonka, 2000; Golonka *et al*., 2006a; modified)
Explanations: 1 – oceanic spreading center and transform faults; 2 – subduction zone; 3 – thrust fault; 4 – normal fault; 5 – transform fault; 6 – mountains; 7 – landmass; 8 – ice cap; 9 – shallow sea and slope; 10 – deep ocean basin

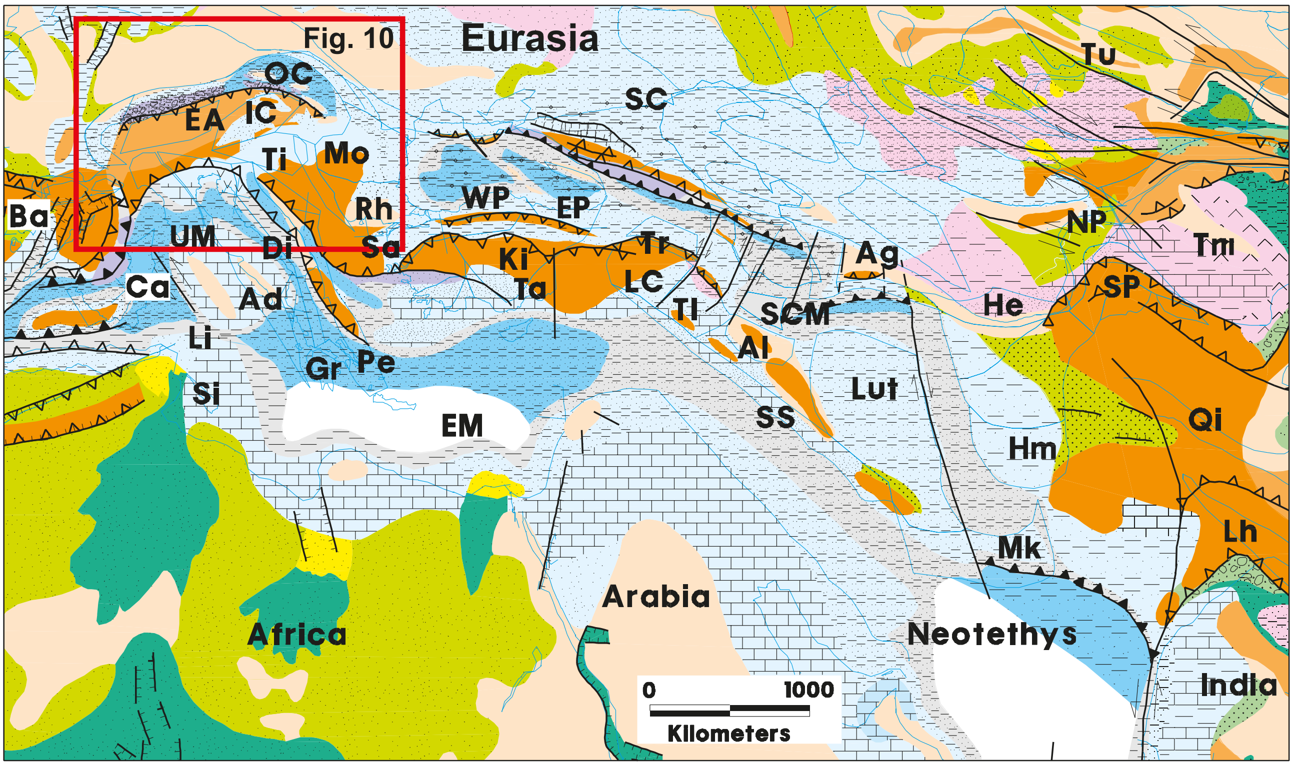

Fig. 9. Palaeogeography of the southern margin of Eurasia with lithofacies of the circum-Carpathian/Caspian areas during Oligocene time (plates' position about 30 Ma) (after Golonka, 2004; slightly modified)
Abbreviations of oceans/seas and plates names: Ad – Adria (Apulia); Ag – Aghdarband (southern 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; Mk – Makran; Mo – Moesia; NP – North Pamir; OC – Outer Carpathians; Pe – Pelagonian plate; 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. Explanations of colors and symbols – see Fig. 6

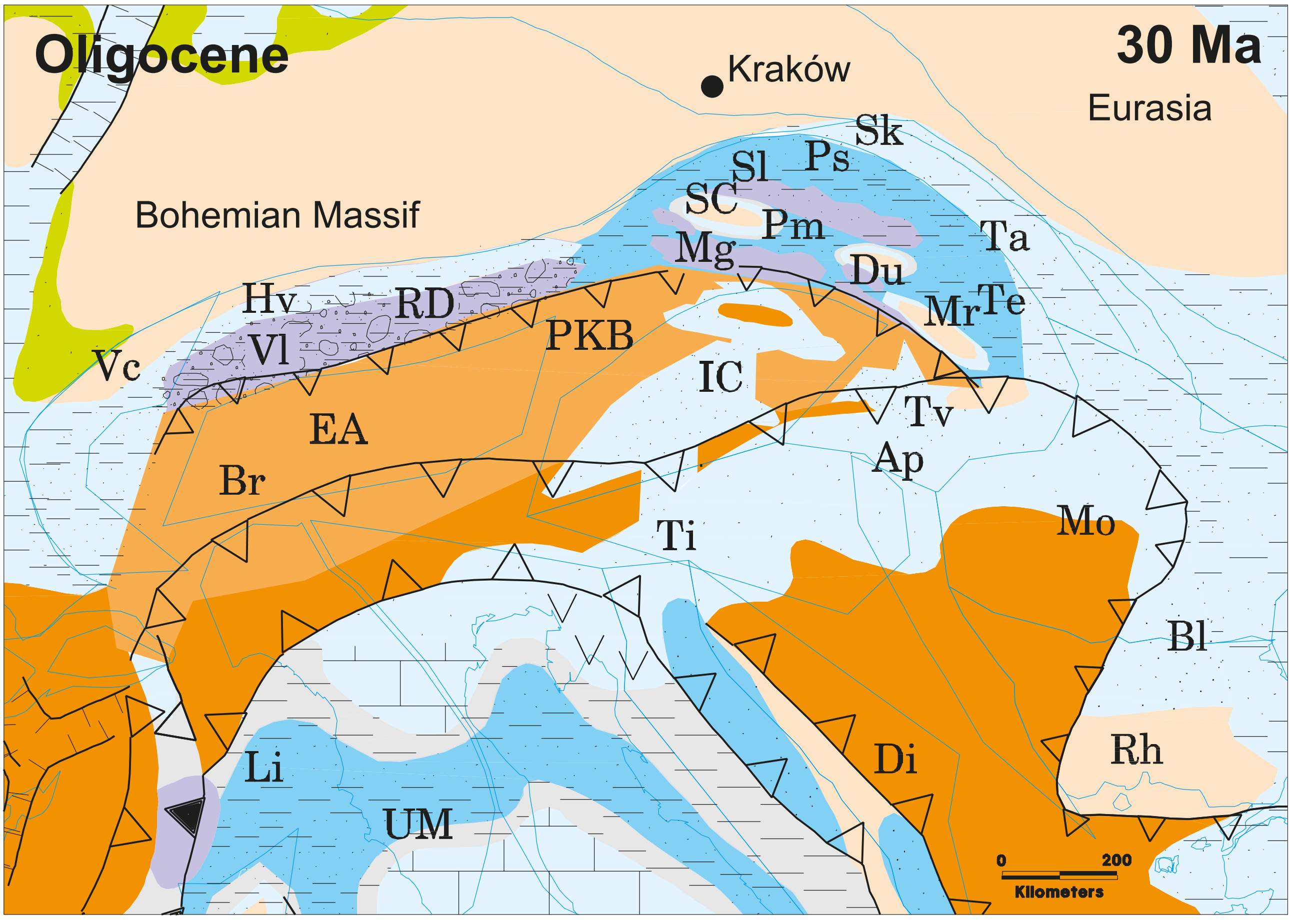

Fig. 10. Palaeogeography, palaeoenvironment and lithofacies of the circum-Carpathian area in Oligocene (plates' position about 30 Ma) (after Golonka *et al.*, 2000, 2006b; slightly modified)

Abbreviations: Ap – Apuseni Mts; Bl – Balkan Basin and fold belt; Br – Briançonnais terrane; Di – Dinarides; Du – Dukla Basin; EA – Eastern Alps; Hv – Helvetic shelf; IC – Inner Carpathians; Mg – Magura Basin; Mo – Moesia plate; Mr – Marmarosh Massif; Podh. Flysch – Podhale Flysch Basin; PKB – Pieniny Klippen Belt; Pm – Fore-Magura Basin; Ps – Sub-Silesian Ridge and slope zone; RD – Rhenodanubian Basin; Rh – Rhodopes; SC – Silesian Ridge (Cordillera); Sl – Silesian Basin; Sk – Skole Basin; Ta – Tarçau Basin; Ti – Tisza plate; Te – Teleajen Basin; Tv – Transylvanian Basin; UM – Umbria-Marche; Vc – Vercors zone; Vl – Valais trough. For other symbols and colors – see Fig. 6

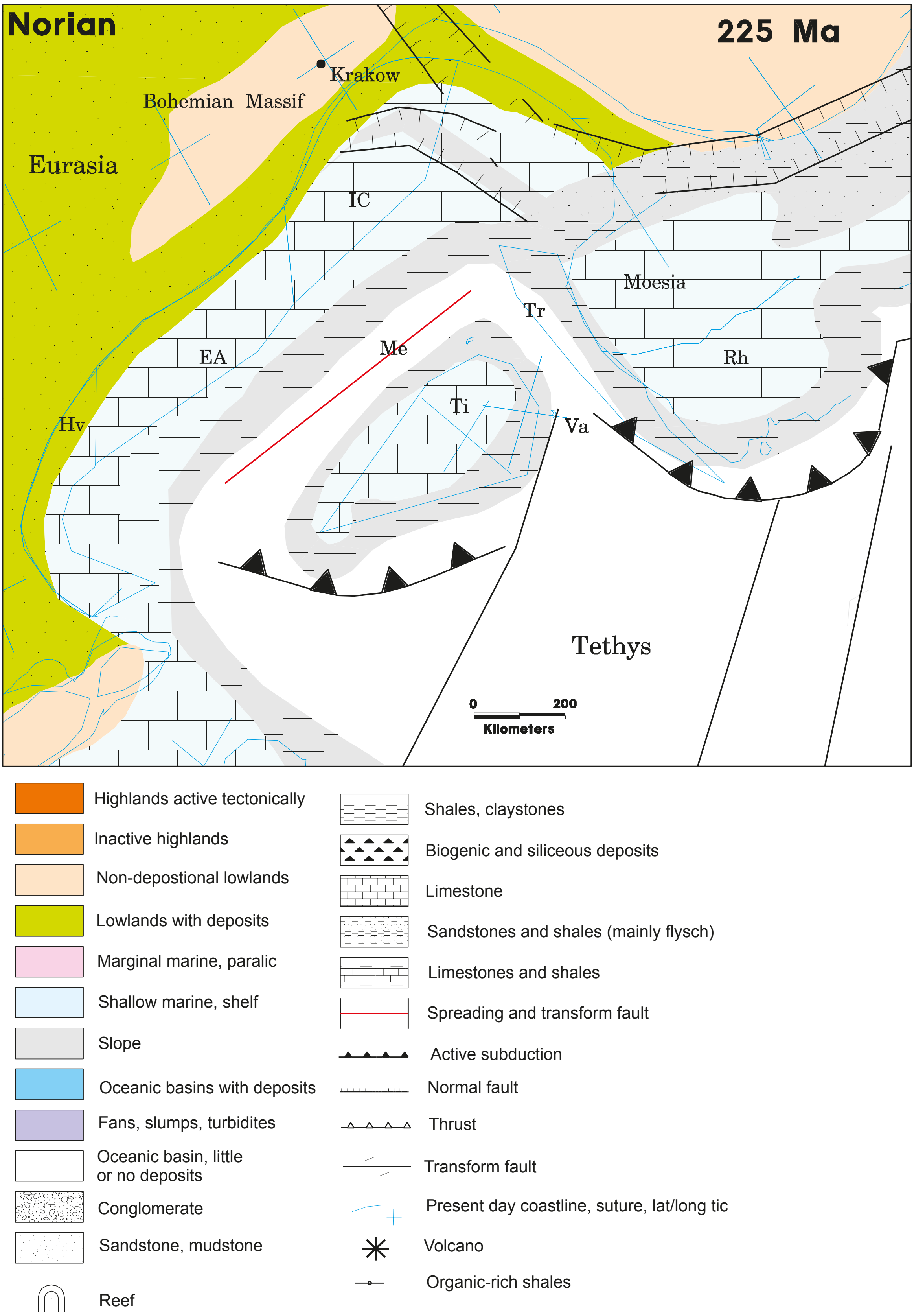

Fig. 11. Palaeogeography of the circum-Carpathian area during Late Triassic; plates position at 225 Ma (after Golonka *et al.*, 2005, modified)
Abbreviations of oceans and plates names: EA – Eastern Alps; Hv – Helveticum; IC – Inner Carpathians; Me – Meliata/Halstatt Ocean; Rh – Rhodopes; Ti – Tisa plate; Tr – Transilvanian Ocean; Va – Vardar Ocean

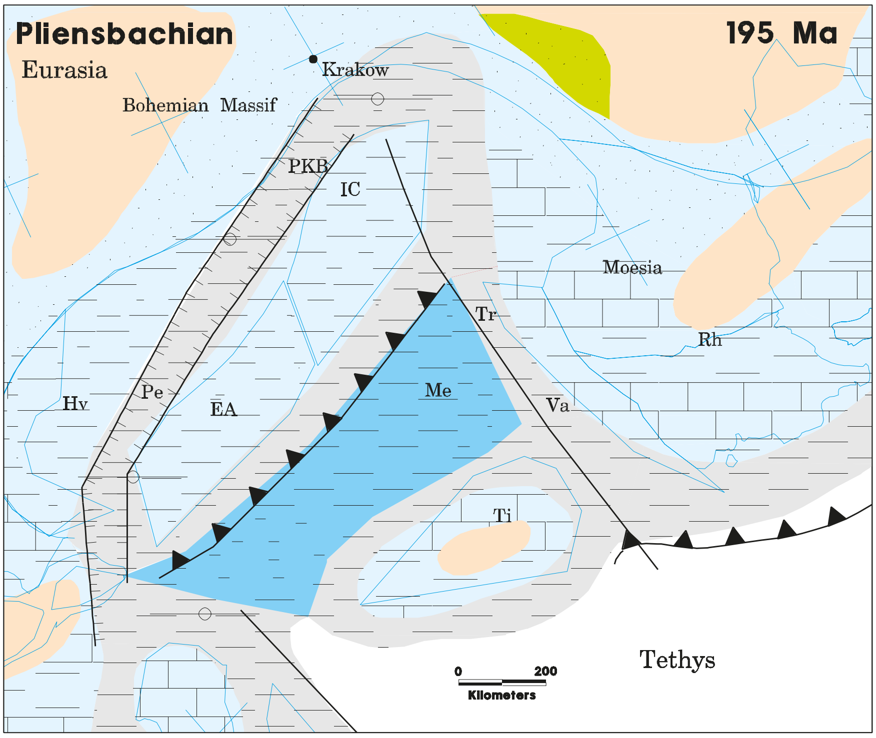

Fig. 12. Palaeogeography of the circum-Carpathian area during Early Jurassic; plates position at 195 Ma (after Golonka *et al*., 2005, modified)

Abbreviations of oceans and plates names: EA – Eastern Alps; Hv – Helveticum; IC – Inner Carpathians; Me – Meliata/Halstatt Ocean; PKB – Pieniny Klippen Belt rift (site of future Pieniny Klippen Belt/Magura Basin); Pe – Penninic rift (site of future Penninic Ocean); Rh – Rhodopes; Ti – Tisa plate; Tr – Transilvanian Ocean; Va – Vardar Ocean. For other symbols and colors – see Fig. 11

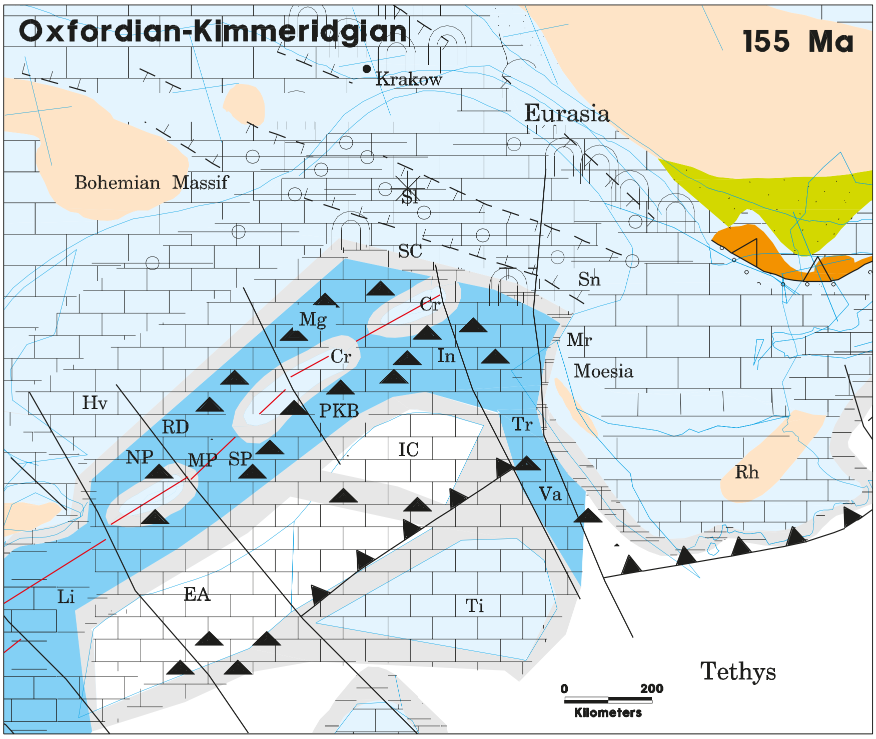

Fig. 13. Palaeogeography of the circum-Carpathian area during Late Jurassic; plates position at 155 Ma (after Golonka *et al*., 2005, modified)

Abbreviations of oceans and plates names: Cr – Czorsztyn Ridge; EA – Eastern Alps; Hv – Helveticum; IC – Inner Carpathians; In – Inacovce-Kricevo Ocean; Li – Ligurian Ocean; Mg – Magura Basin; MP – Midlle Penninic Ocean; Mr – Maramures Basin; NP – Northern Penninic Ocean; PKB – Pieniny Klippen Belt Basin; RD – Rhenodanubian Basin; Rh – Rhodopes; SC – Silesian Ridge/Cordillera; Sl – Silesian Basin; Sn – Sinaia Basin; SP – South Penninic Ocean; Ti – Tisa plate; Tr – Transilvanian Ocean; Va – Vardar Ocean. For other symbols and colors – see Fig. 11

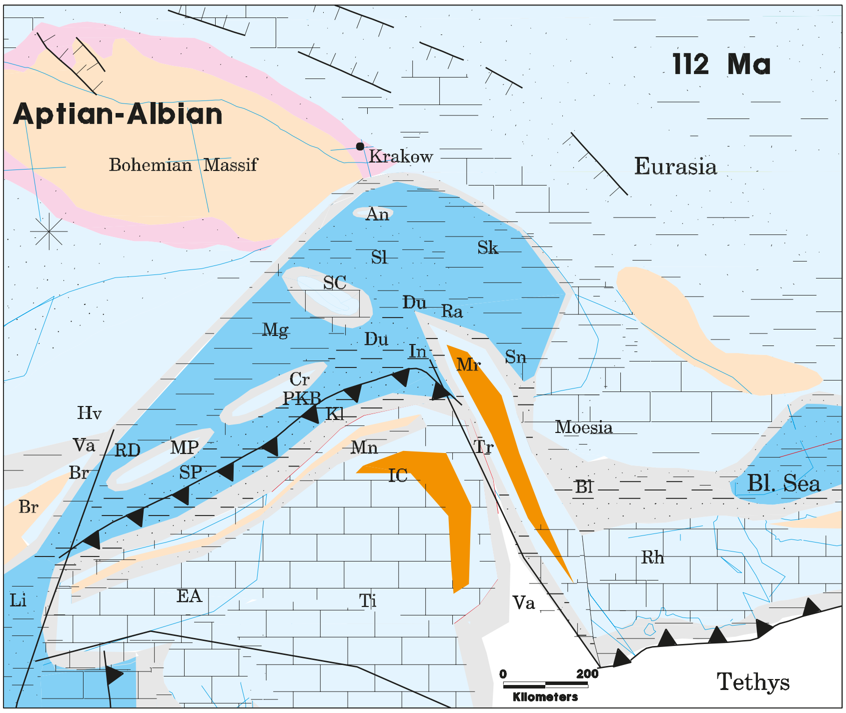

Fig. 14. Palaeogeography of the circum-Carpathian area during late Early Cretaceous; plates position at 112 Ma (after Golonka *et al.*, 2005, modified)

Abbreviations of oceans and plates names: An – Andrychów Ridge; Bl – Balcans; Bl. Sea – Black Sea; Br – Briancone Ridge/Basin; Cr – Czorsztyn Ridge; Du – Dukla Basin; EA – Eastern Alps; Hv – Helveticum; IC – Inner Carpathians; In – Inacovce-Kricevo Ocean; Kl – Klape Basin; Li – Ligurian Ocean; Mg – Magura Basin; Mn – Manin Basin; MP – Midlle Penninic Ocean; Mr – Maramures Ridge; PKB – Pieniny Klippen Belt Basin; Ra – Rahiv Basin; RD – Rhenodanubian Basin; Rh – Rhodopes; SC – Silesian Ridge/Cordillera; Sk – Skole Basin; Sl – Silesian Basin; Sn – Sinaia Basin; SP – South Penninic Ocean; Ti – Tisa plate; Tr – Transilvanian Ocean; Va – Vardar Ocean. For other symbols and colors – see Fig. 11

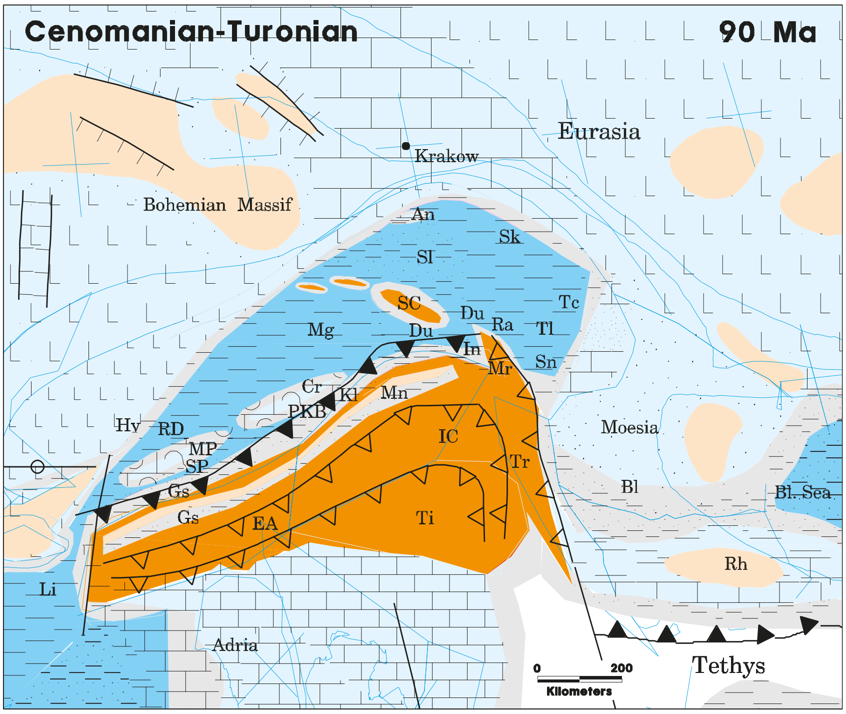

Fig. 15. Palaeogeography of the circum-Carpathian area during late Early Cretaceous; plates position at 112 Ma (after Golonka *et al.*, 2005, modified)

Abbreviations of oceans and plates names: An – Andrychów Ridge; Bl – Balcans; Bl. Sea – Black Sea; Cr – Czorsztyn Ridge; Du – Dukla Basin; EA – Eastern Alps; Hv – Helveticum; Gs – Gosau Basin; IC – Inner Carpathians; In – Inacovce-Kricevo Ocean; Kl – Klape Basin; Li – Ligurian Ocean; Mg – Magura Basin; Mn – Manin Basin; MP – Midlle Penninic Ocean; Mr – Maramures Ridge; PKB – Pieniny Klippen Belt Basin; Ra – Rahiv Basin; Rh – Rhodopes; SC – Silesian Ridge/Cordillera; Sk – Skole Basin; Sl – Silesian Basin; Sn – Sinaia Basin; SP – South Penninic Ocean; Tc- Tarcau Basin; Ti – Tisa plate; Tr – Transilvania. For other symbols and colors – see Fig. 11

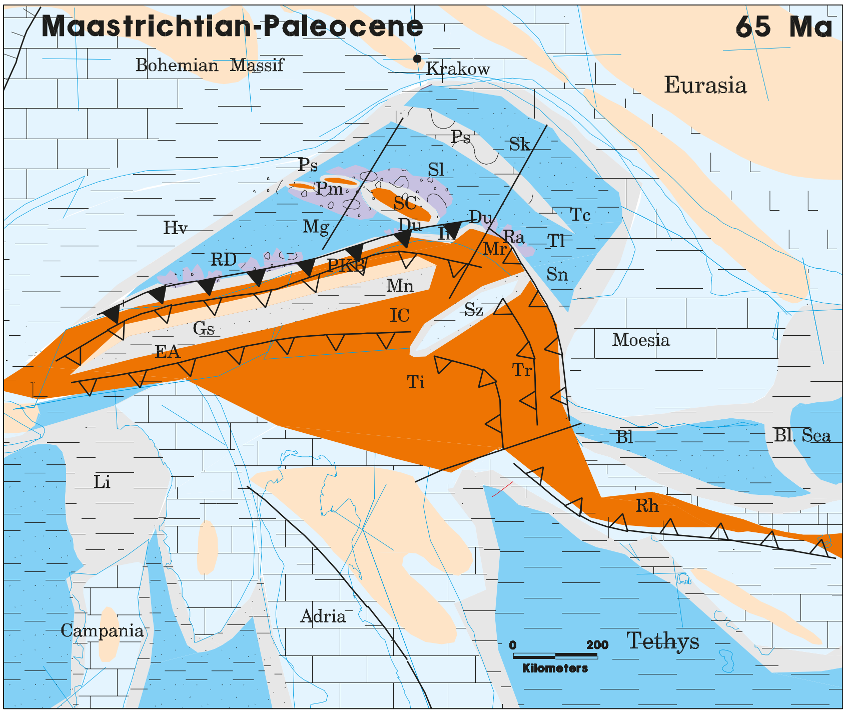

Fig. 16. Palaeogeography of the circum-Carpathian area during Late Cretaceous/Paleogene transition; plates position at 65 Ma (after Golonka *et al*., 2005, modified)

Abbreviations of oceans and plates names: Bl – Balcans; Bl. Sea – Black Sea; Du – Dukla Basin; EA – Eastern Alps; Hv – Helveticum; Gs – Gosau Basin; IC – Inner Carpathians; Li – Ligurian Ocean; Mg – Magura Basin; Mn – Manin Basin; Mr – Maramures Ridge; PKB – Pieniny Klippen Belt; Pm – Fore-Magura Basin; Ps – Sub-Silesian Ridge and slope zone; Ra – Rahiv Basin; RD – Rhenodanubian Basin; Rh – Rhodopes; SC – Silesian Ridge/Cordillera; Sk – Skole Basin; Sl – Silesian Basin; Sn – Sinaia Basin; Sz – Szolnok Basin; Tc – Tarcau Basin; Ti – Tisa plate; Tr – Transilvania. For other symbols and colors – see Fig. 11

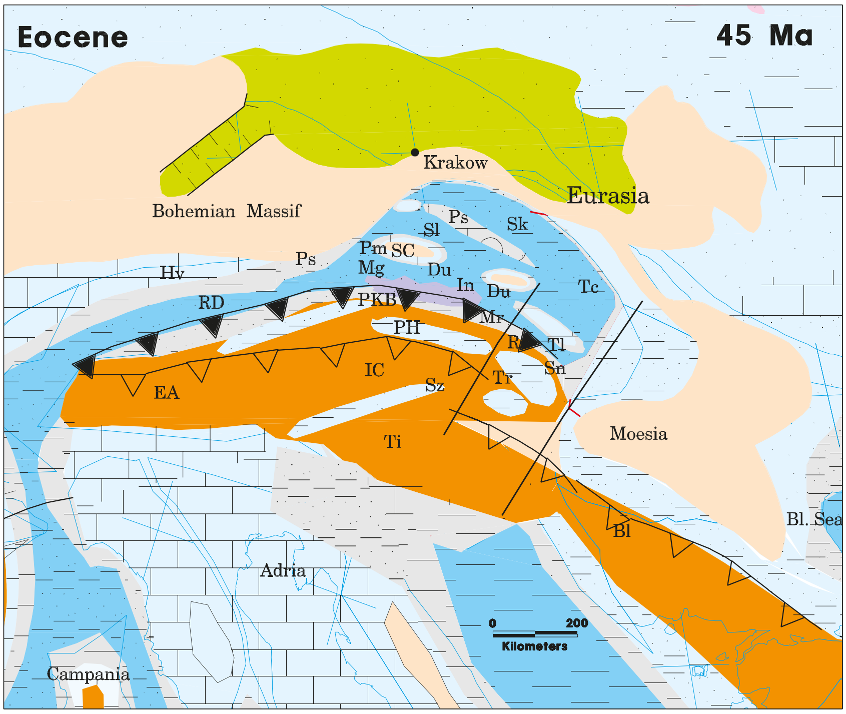

Fig. 17. Palaeogeography of the circum-Carpathian area during Early Eocene; plates position at 45 Ma (after Golonka *et al*., 2005, modified)

Abbreviations of oceans and plates names: Bl – Balcans; Bl. Sea – Black Sea; Du – Dukla Basin; EA – Eastern Alps; Hv – Helveticum; IC – Inner Carpathians; In – Inacovce-Kricevo; Mg – Magura Basin; Mr – Maramures Ridge; PKB – Pieniny Klippen Belt; Pm – Fore-Magura Basin; Ps – Sub-Silesian Ridge and slope zone; Ra – Rahiv Basin; RD – Rhenodanubian Basin; SC – Silesian Ridge/Cordillera; Sk – Skole Basin; Sl – Silesian Basin; Sn – Sinaia Basin; Sz – Szolnok Basin; Tc – Tarcau Basin; Ti – Tisa plate; Tr – Transilvania. For other symbols and colors – see Fig. 11

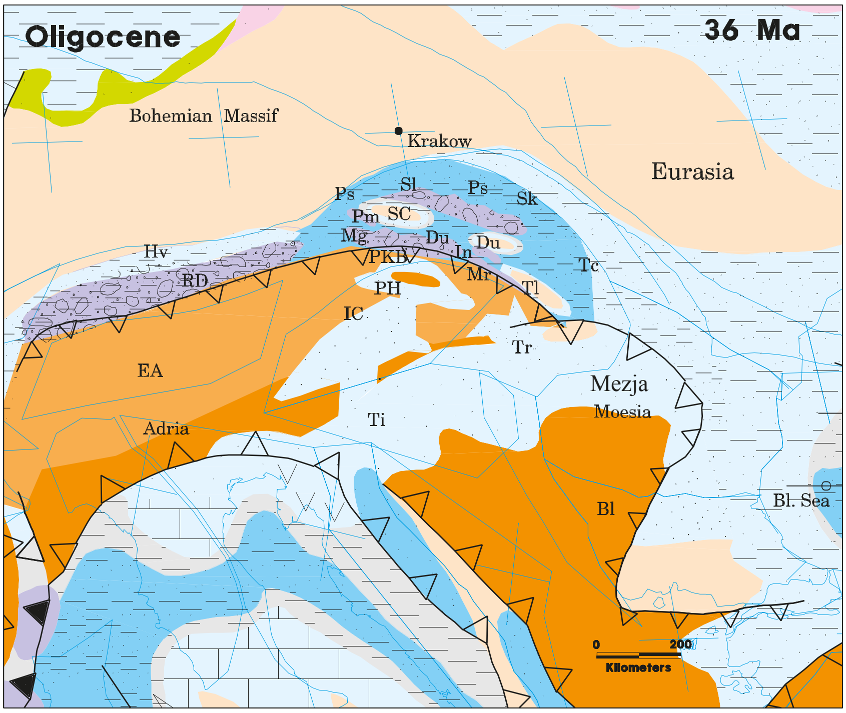

Fig. 18. Palaeogeography of the circum-Carpathian area during Oligocene; plates position at 36 Ma (after Golonka *et al.*, 2005, modified)

Abbreviations of oceans and plates names: Bl – Balcans; Bl. Sea – Black Sea; Du – Dukla Basin; EA – Eastern Alps; Hv – Helveticum; IC – Inner Carpathians; In – Inacovce-Kricevo; Mg – Magura Basin; Mr – Maramures Ridge; PH – Podhale Flysch Basin; PKB – Pieniny Klippen Belt; Pm – Fore-Magura Basin; Ps – Sub-Silesian Ridge and slope zone; RD – Rhenodanubian Basin; SC – Silesian Ridge/Cordillera; Sk – Skole Basin; Sl – Silesian Basin; Tc – Tarcau Basin; Ti – Tisa plate; Tr – Transilvania. For other symbols and colors – see Fig. 11

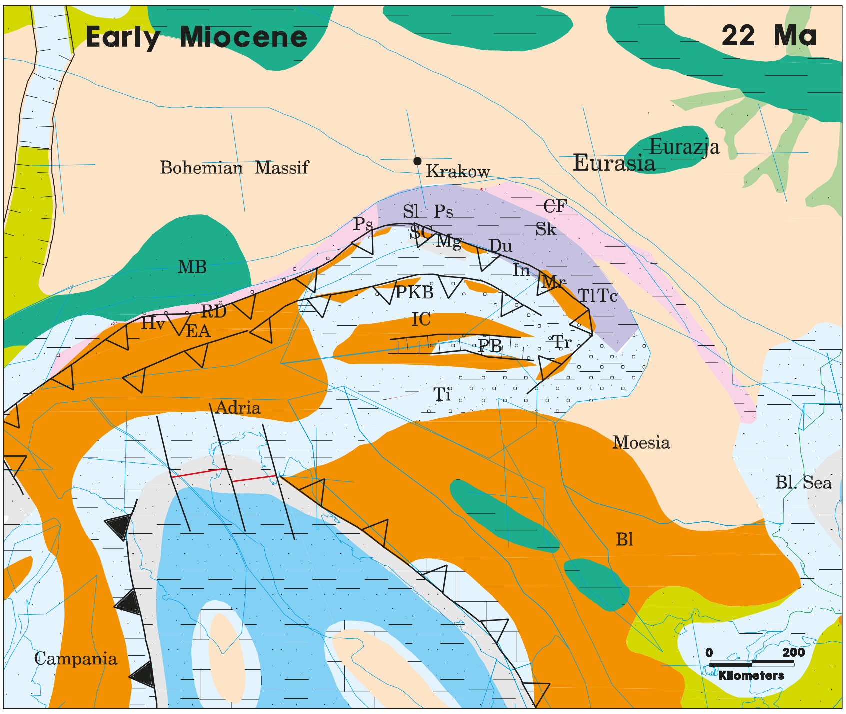

Fig. 19. Palaeogeography of the circum-Carpathian area during Early Miocene; plates position at 22 Ma (after Golonka *et al.*, 2005, modified).

Abbreviations of oceans and plates names: Bl – Balcans; Bl. Sea – Black Sea; CF – Carpathian Foredeep; Du – Dukla Basin; EA – Eastern Alps; Hv – Helveticum; IC – Inner Carpathians; In – Inacovce-Kricevo; MB – Molasse Basin; Mg – Magura Basin; Mr – Maramures Ridge; PB – Panonnian Basin; PKB – Pieniny Klippen Belt; Ps – Sub-Silesian Ridge; RD – Rhenodanubian; SC – Silesian Ridge/Cordillera; Sk – Skole Basin; Sl – Silesian Basin; Tc – Tarcau Basin; Ti – Tisa plate; Tr – Transilvania. For other symbols and colors – see Fig. 11

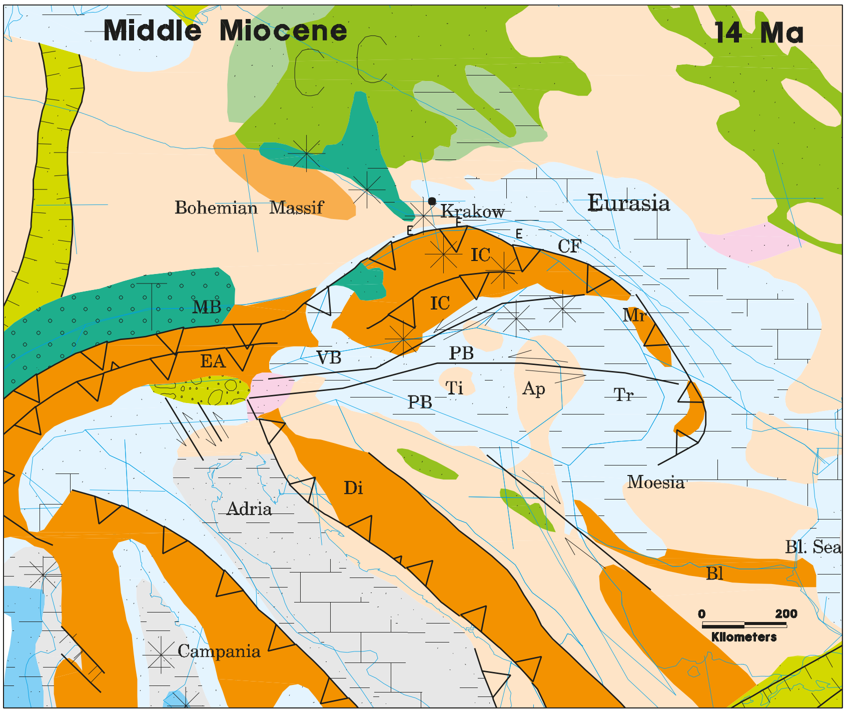

Fig. 20. Palaeogeography of the circum-Carpathian area during Middle Miocene; plates position at 14 Ma (after Golonka *et al*., 2005, modified).

Abbreviations of oceans and plates names: Ap –Apuseni Mts; Bl. Sea – Black Sea; CF – Carpathian Foredeep; Di – Dinarides; EA – Eastern Alps; IC – Inner Carpathians; MB – Molasse Basin; Mr – Maramures Ridge; PB – Panonnian Basin; Ti – Tisa plate; Tr – Transilvania; VB – Vienna Basin. For other symbols and colors – see Fig. 11

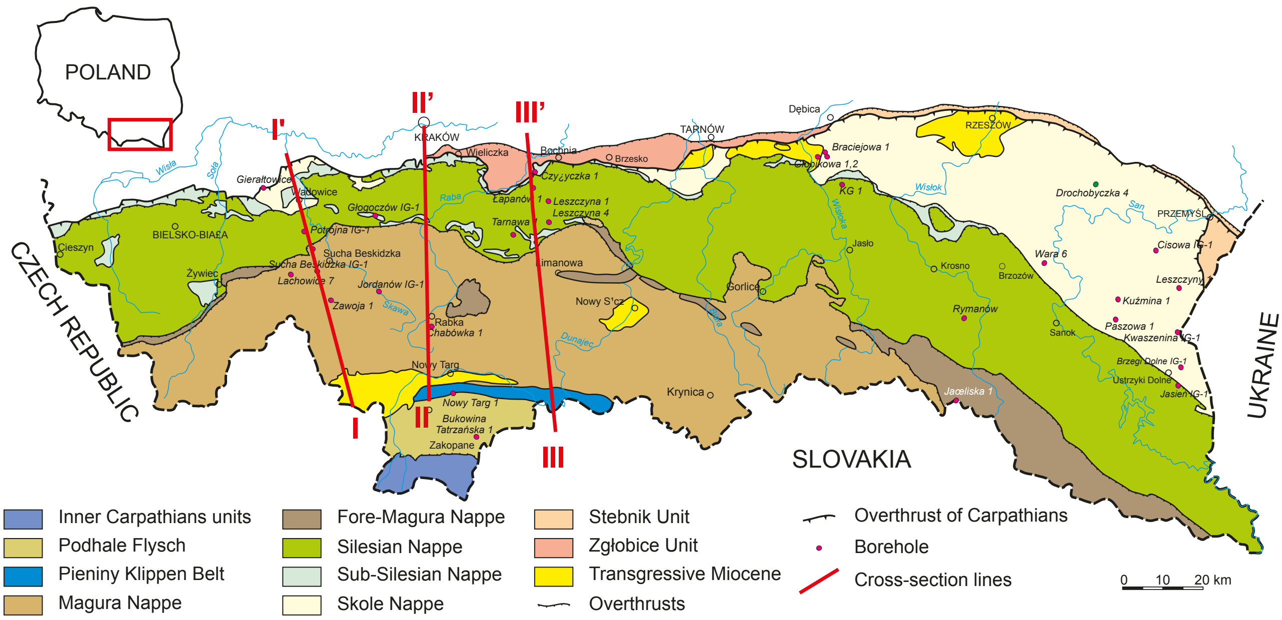

Fig. 21. Map of the Polish Outer Carpathians with the locality of cross-sections (see – Fig. 22) (after Żytko *et al*., 1989; Golonka *et al*., 2011, modified)

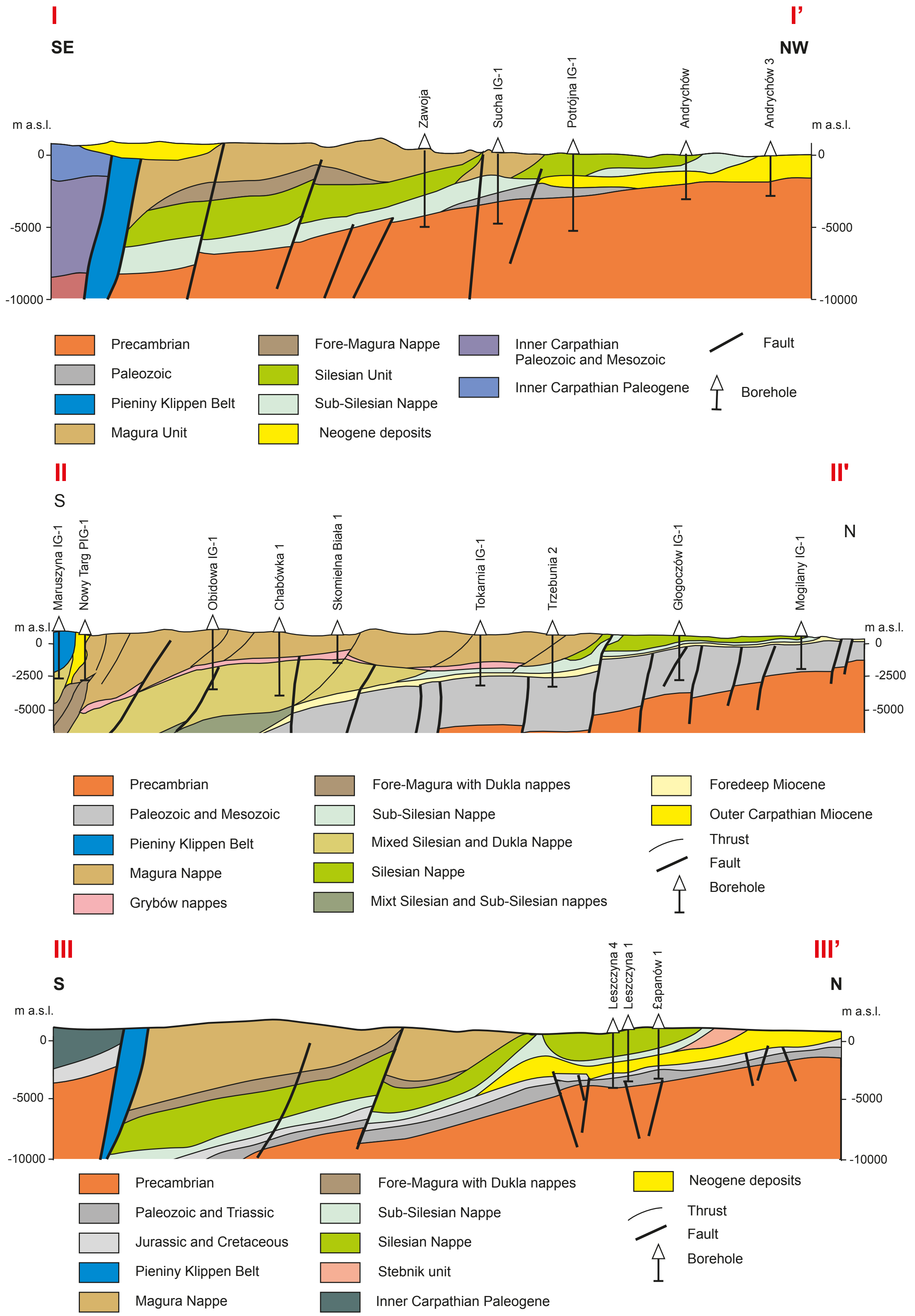

Fig. 22. Cross-sections through the western part of the Outer Carpathians and their foreland (after Golonka *et al.*, 2011, modified). Cross-section locations on Fig. 21

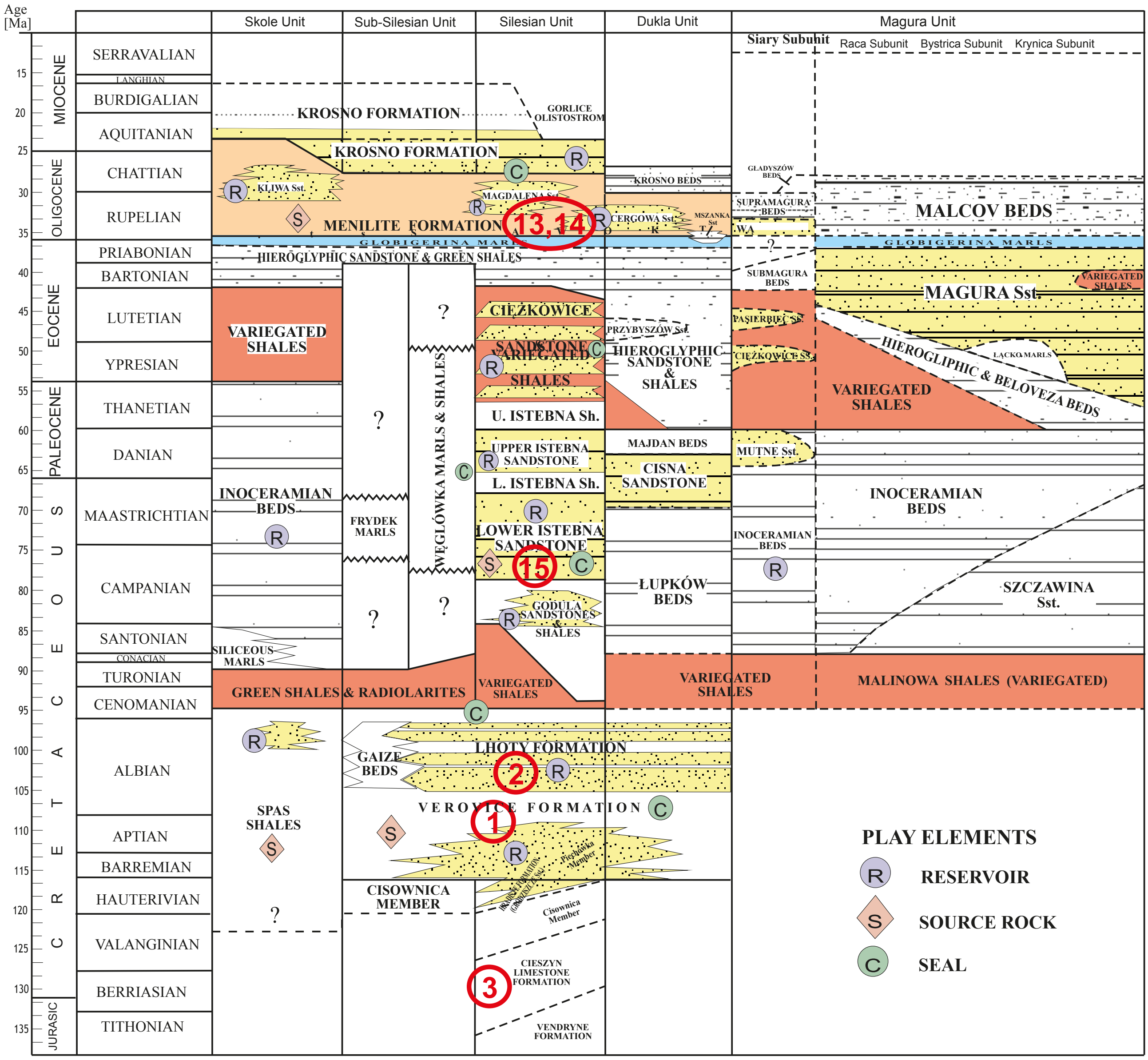

Fig. 23. Simplified lithostratigraphy of the Outer Polish Carpathians (after Koszarski *et al*., 1985; Dziadzio *et al*., 2001; modified) with the position of field trip stops

RZYKI (VEŘOVICE FORMATION) (FIGS 24–26)

Jan GOLONKA & Michał KROBICKI

Polish Carpathians, being a cradle of oil industry offer objects illustrating gathering of organic substance into organic-rich rocks, features of reservoirs – rocks containing oil and gas, oil seeps on surface, as well as historic sites of early industrial oil exploration. The classic site in Polish Outer Carpathians containing organic-rich rocks – source rocks for generation of oil is exposed in the Wieprzówka cascade site in Rzyki village (Fig. 24). This site is located near Andrychów town, about 80 km from the Krakow Balice Airport and close to Wadowice, the Pope John Paul II town. It can serve as an object illustrating geological processes such as the Oceanic Anoxic Event (OAE) for the general public. These events occur when the Earth's oceans become severely depleted of oxygen. During an anoxic event conditions in the vast areas of the oceans are similar to those existing today in the Black Sea. Much organic matter gathers on the bottoms of seas and oceans. It is believed that oceanic anoxic events are linked to specific palaeogeographic and palaeoclimatic conditions, changing oceanic current circulations, volcanic activity, and an increase of greenhouse gases and the subsequent global warming. Most of the Earth's oil and gas originated from organic-rich source rocks deposited during OAEs. The end of the Early Cretaceous represents one of these OAEs (Bralower *et al.*, 2002; Kratochvílová *et al.*, 2003; Golonka *et al.*, 2008b; Ślączka *et al.*, 2014).

The global sea-level was more than 200 hundred m above today's level, close to the maximum 1st-order highstand for the entire Earth history. This was also a period of increasing continental submergence. Volcanism activity release enormous amount of greenhouse gases, especially CO_2. Global greenhouse conditions prevailed. Hot, equable climates occurred, with generally humid continental interior settings. Organic-rich sediments were deposited in many marine basins worldwide. Typical OAE deposits are dark-grey or black; this color is caused by the high amount of carbon, measured as so-called Total Organic Carbon (TOC). Shales of the Veřovice Formation represent typical OAE-deposited rocks (Bralower *et al.*, 2002; Kratochvílová *et al.*, 2003, Golonka *et al.*, 2008b; Ślączka *et al.*, 2014).

The Veřovice Formation (Golonka *et al.*, 2008a, b and references therein) comprises typical black Cretaceous organic-rich deposits of the Silesian Series (Figs 25, 26). Its stratotype is located in the village of Veřovice in the Czech Republic; the Rzyki profile serves as a reference section. The formation age, evaluated from assemblages of dinoflagellates (Cieszkowski *et al.,* 2001, 2003; Skupien, 2003; Golonka *et al*., 2008a, b) and foraminifera (Bieda *et al*., 1963; Nowak, 1976; Szydło, 1997), is Barremian-earliest Albian. In the lithostratigraphic profile of the Silesian Series the Veřovice Formation is underlined by the Hradište Formation (Late Valanginian–Hauterivian) and stratigraphically covered by the Lhoty Formation (Albian–Early Cenomanian) (see Golonka *et al*., 2008a and references therein). This formation is tectonically deformed as a part of the Silesian Nappe. As mentioned above, rocks of the formation were deposited during the Early Cretaceous, under OAE conditions. Transgressions related to the highest Phanerozoic sea-level and upwelling contributed to the excessive nutrient supply. The Carpathian basins produced a large amount of organic matter, preserved due to the sedimentary conditions and to the limited supply of terrigeneous material. A Rock-Eval analysis of the Veřovice Formation from Rzyki revealed Total Organic Carbon – TOC content reaching 2.31 wt%. Organic-rich deposits are widespread in the Silesian Basin, with TOC values reaching 3–3.5 wt% in the Veřovice Formation in Moravia and in the Zasań area south of Kraków (Golonka *et al*., 2008b). These rocks were buried under a thick pile of younger, Upper Cretaceous and Paleogene flysch deposits, deformed into the imbricated series of nappes. This development led to the maturation of organic matter and expulsion of oil, as indicated by the Tmax values.

The section at the **Wieprzówka** cascade, forming Lower Cretaceous deposits of the Silesian Nappe, is located in the Beskid Mały "block" (Książkiewicz, 1977) (Fig. 25). Its geological surroundings were first depicted on the map by Książkiewicz (1951). Later the section was described by Ślączka & Kaminski (1998), Cieszkowski *et al*. (2001, 2003), Uchman (2004), Uchman & Cieszkowski (2008). Zieliński (2003) described the natural diversity and various, rich natural attractions of the whole Wieprzówka stream valley. He also noted the beauty of the waterfalls formed in the northern part of Rzyki. The stream valley of Wieprzówka in Rzyki was deeply dissected by fluvial erosion, probably as a result of extraction of gravel from the stream

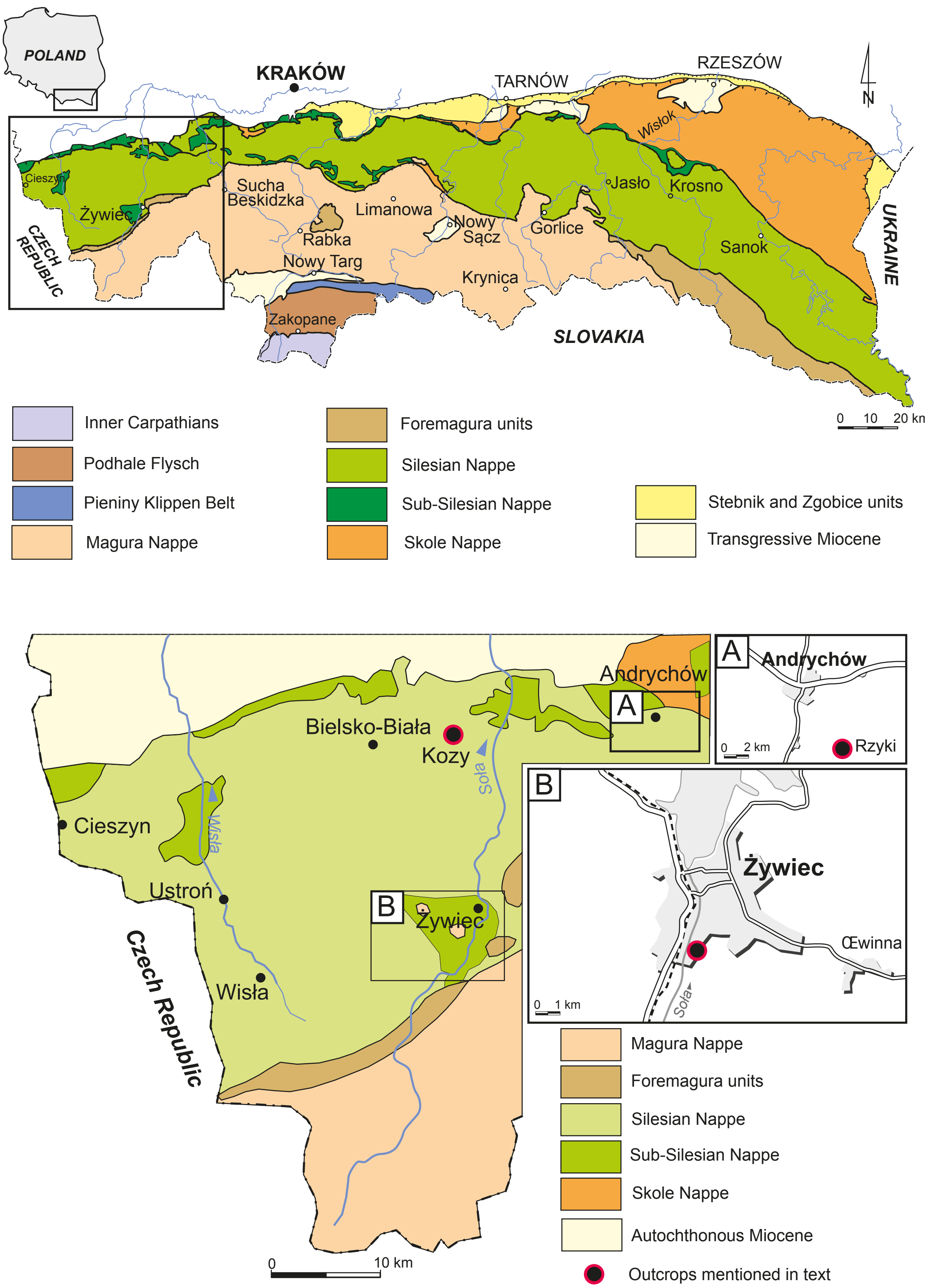

Fig. 24. Geological sketch between Andrychów and Żywiec with location of described outcrops (after Żytko *et al.*, 1989; simplified)

beds (Ślączka & Kaminski, 1998) and/or remodeling of the stream bed by the great flood in 1997 and later events. The stream erosion formed a kind of canyon a few m deep, with a series of scenic waterfalls and erosional potholes in the riverbed, and revealed the detailed development of the Lower Cretaceous Veřovice and Lhoty Formations (Fig. 25). These formations are gently folded and dip gently toward the south-southeast. They are cut by a system

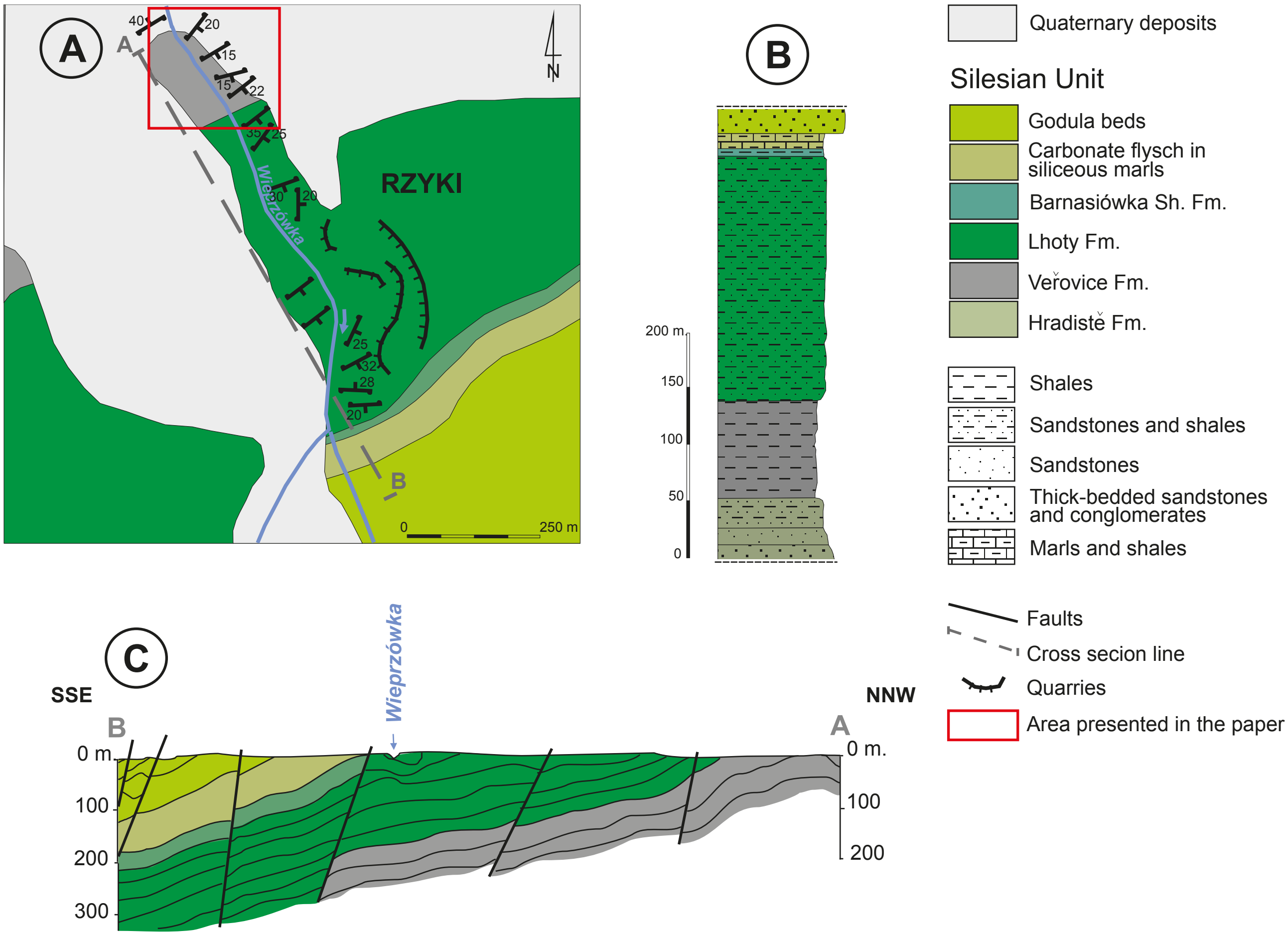

Fig. 25. Geological location of the Veřovice Formation's reference section in Rzyki
A – simplified geological map of the northern part of Rzyki (after Książkiewicz, 1951; Uchman & Cieszkowski, 2008, modified) – as in part B; **B** – geological log of the Cretaceous deposits showing the lithostratigraphic position of the Veřovice Formation in Rzyki; **C** – geological cross-section observed in northern part of Rzyki

of meso-scale faults. Most faults included within the system are normal, with down-thrown northern flanks, but some are reverse with northern thrust vergence. Transform faults, usually NNW–SSE oriented, also exist in this section. Fluvial erosion exploited the fault system, when forming the canyon. The uppermost part of the Veřovice Formation and the lower part of the Lhoty Formation are exposed in the cascade.

In the northern part of Rzyki, bordering on the NE with the village of Zagórnik, the section begins with the gently folded Veřovice Formation (uppermost Barremian–Aptian –earliest Albian). The upper part of the formation is exposed in the section, while the lower is covered by Quaternary fluvial deposits. The outcropping part of the Veřovice Formation is represented by carbonate-free black and dark grey, partly siliceous shaly mudstones with intercalations of thin- and very thin-bedded laminated coarser siltstones, fine-grained sandstones (Fig. 26). The sandstones are quartzitic, gray, with parallel or cross-ripple lamination. They have presumably been deposited by bottom currents in an abyssal plain sedimentary environment. Up the section, a few m thick, shaly-sandstone packages with thin- or medium-bedded sandstones are present within the mudstone and siltstone sequence. Some of the sandstones display features of turbidites. At the steep transition between the Veřovice and Lhoty Formations sandstones are more common. Occasional siderites and/or ankerites form thin layers, lenses, or oval concretions within black shales. In some places shales with siderites are disturbed by sub-marine slumps. Small aggregates of pyrite are also visible in the shales.

While the black shales of the Veřovice Formation at Rzyki are Lower Cretaceous anoxic deposits, trace fossils occur there occasionally (Ślączka & Kaminski, 1998; Cieszkowski *et al*., 2001, 2003; Uchman & Cieszkowski, 2008), indicating that living conditions periodically improved and that the sea-floor was colonized by animals. The trace-fossils were figured by Uchman (2004), Uchman in: Cieszkowski *et al*. (2001, 2003) and Uchman & Cieszkowski (2008). The trace fossils *Phycosiphon incertum*, *Chondrites inticatus*, *Planolites*, *Paleophycus* and locally *Thalassinoides* forms occur within some thin layers of usually lighter pelitic deposits. *Portovirgularia* traces have also been found in the section. It was supposedly produced by chemosymbi-

Fig. 26. The Veřovice Formation close to Rzyki with stream erosion has formed a kind of canyon a few m deep within the Veřovice Formation, series of small waterfalls built of black shales and laminated sandstones in the upper part of the Veřovice Formation. The shales display lower amounts of Total Organic Carbon, between 1–2 wt% in this part of the profile. Shales at the beginning of the Veřovice Formation section displaying the highest amount of Total Organic Carbon (TOC), reaching 2.31 wt%

otic bivalves that could burrow in anoxic sediment. A Barremian–Early Albian age of the Veřovice Formation in Rzyki was previously estimated from micropalaeontological data on foraminifera from other sections in the Outer Carpathian and on the superposition of this formation in the lithostratigraphic profile of the Silesian Series (Cieszkowski *et al*., 2001, 2003; Uchman & Cieszkowski, 2008). Recently dinocysts *Cerbia tabulata, Kiokansium polypes, Odontochitina operculata, Pseudoceratium gochtii, Pseudoceratium polymorphum* were determined by P. Skupien from the lower part of the Rzyki section. This assemblage indicated a Late Barremian–Aptian age.

Up the section the Veřovice Formation passes into the Lhoty Formation, Albian–Early Cenomanian in age. The Lhoty Formation here (Książkiewicz, 1951; Cieszkowski *et al*., 2001, 2003, Uchman & Cieszkowski, 2008) is developed as thin- and medium, and occasionally thick-bedded, quartzitic, dark sandstones with distinct parallel and cross lamination, inter-bedded with black, dark grey and greenish, often spotty shales.

At the beginning of the Veřovice Formation section the shales display the highest content of TOC, reaching 2.31 wt%. Upstream, at places where we encounter a series of small waterfalls built of black shales and laminated sandstones, the shales display lower contents of TOC, between 1–2 wt% in this part of the profile. TOC max values from the Veřovice Formation in the Wieprzówka waterfall profile indicate that a significant amount of oil was expelled, probably at the peak of orogenesis during Miocene times. The rocks in the profile were buried at that time by Carpathian imbricated nappes. This burial enhanced maturation and expulsion. Today the Wieprzówka rocks are exposed at the surface, therefore the maturation process has ceased entirely.

As mentioned above, the Veřovice Formation hosts a variety of tectonic deformations. We can easily identify rather small-scale faults and folds of different geometry. There is also visible a net of complementary joints clearly marked by white veins of calcite within black mudstones and gray sandstone. In some place coulisse or horse systems are presented. All the tectonic deformation resulted from the Miocene Alpine tectonism, which formed the Silesian Nappe as part of the Carpathian Mountains domain.

FIELD TRIP ▸ STOP 2

KOZY (LHOTY FORMATION)

▸FIGS 24, 27

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Typical Lhoty Formation are subdivided into three parts (Książkiewicz, 1962; Koszarski & Ślączka, 1973): (i) lower part (up to several dozen m in thickness) – coarse- and medium-grained sandstones with minor intercalations of exotic conglomerates, (ii) middle part (some hundred meters) – complex of medium- to thin-bedded sandstones and shales known from the whole Carpathian units, (iii) upper part (a dozen m maximum) – so-called Mikuszowice cherts developed as sandstones with spongiolites ("cherts") which locally occur in Carpathians.

In abandoned quarry in Kozy (Figs 24, 27) the middle part of the Lhoty Formation occur and is represented by 200-250 m thick sequence of distal turbiditic system of SM and MS facies. Medium- to fine-grained sandstones dominated usually as thin-bedded layers with fractionation gradations and different kind of laminations (parallel, convolute, cross-bedding etc) in several combinations of Bouma sequence (Bouma, 1962; see also – Bilan, 2013). SM facies is represented by sandstones with mudstones with flat and irregular bases with erosional character (Słomka, 1995). Numerous flute casts and trace fossils (e.g., *Arthrophycus tenuis*) (Cieszkowski *et al.*, 2001; Uchman, 2001) occur. Unrug (1959) described from this quarry sedimentological structures in this unit and Alexandrowicz & Lalik (1999) mentioned both sedimentary (e.g., coal, limestones) and crystalline rocks. Some authors distinguished several sub-facies of turbiditic system: mlSM – massive sandstones with parallel and cross-bedding lamination going into mudstones, gSM – fractional grained sandstones going into mudstones, glSM – fractional grained sandstones and parallel laminated and/or cross-bedded going into mudstones or lSM – parallel laminated sandstones and/or cross-bedded going into mudstones. Based on such sedimentological features we can interpret these facies as effect of sedimentation of muddy-sandy turbiditic currents with different kind of density and ratio of sedimentation of high-energy sea-bottom environments (Ghibaudo, 1992; Słomka, 1995) of submarine lobes. Some deformational deposits occur sporadically (facies F) which originated as submarine debris flows of typical deep-marine fans of turbiditic system (Mutti & Ricci Lucchi, 1972; Walter, 1978; Shanmugam & Moiola, 1988; Reading & Richards, 1994; Słomka, 1995; Bilan, 2013). In conclusion, in this quarry typical distal part of siliciclastic deep-marine fans and/or lobes occur with minor intercalations of apron-type/ramp system with linear source area (Gallowey, 1998). Additionally, analysis of direction of currents (by flute casts) indicate W and SW location of this area. The most probably Silesian Ridge/Cordillera has been linear source area with unstable architecture which supported origin of starved fans (un-channelized turbiditic desposits) (Bilan, 2013).

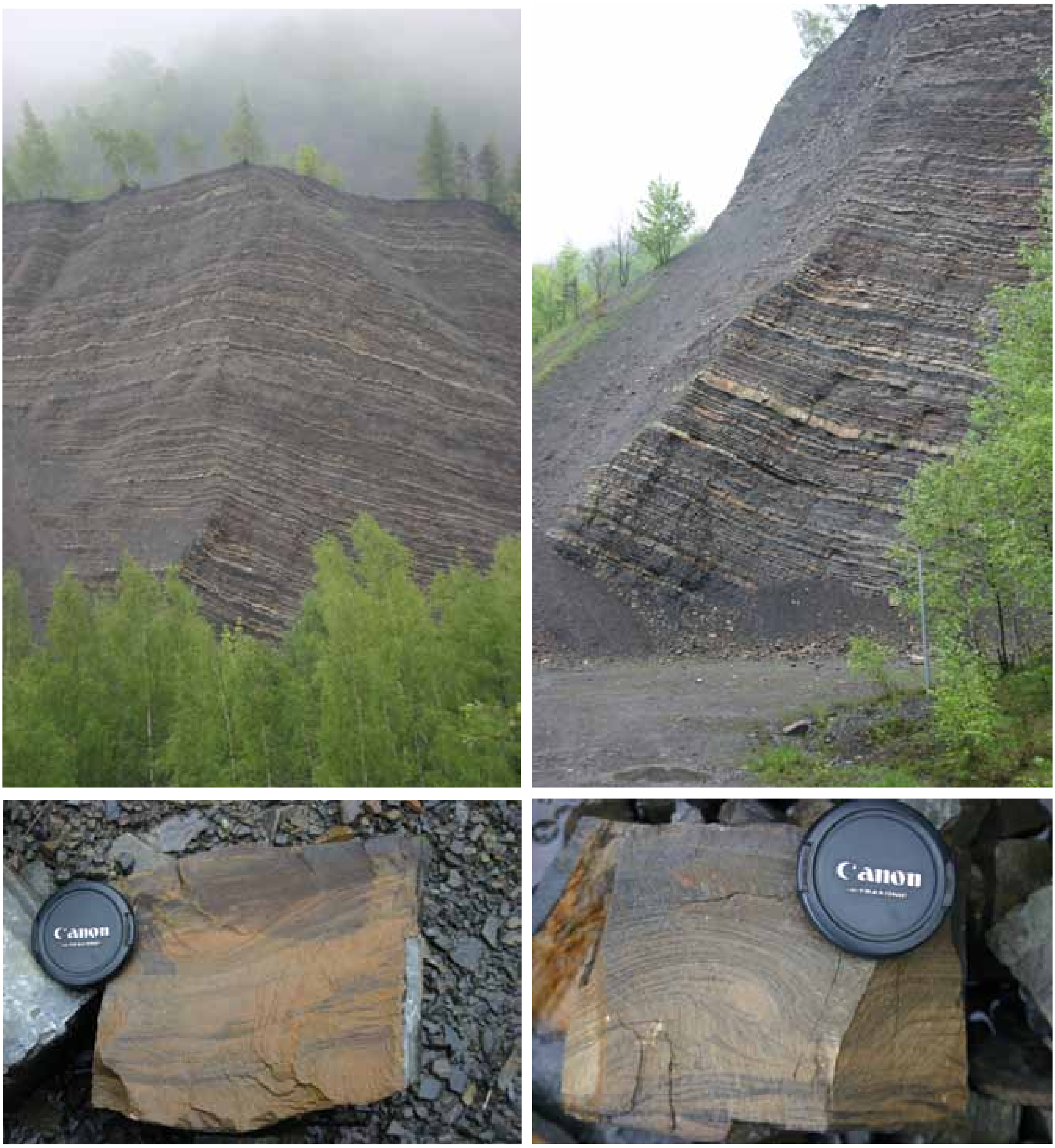

Fig. 27. The Lhoty Formation in the Kozy quarry with typical thin-bedded features of this formation and depositional structures

ŻYWIEC – SOŁA RIVER VALLEY (CIESZYN LIMESTONE FORMATION)

⊳FIGS 24, 28–32

Michał KROBICKI & Jan GOLONKA

Żywiec is a small town, location of the brewerey, founded by Habsburgs, which produces the best and most famous Polish beer, well known not only in Poland. The main secret of the Żywiec beer high quality is spring water used to its production. The water is explored from Żywiec Beskid mountain rangę, built of the flysch deposits of the Silesian Nappe. From the parking lot in the southern part of Żywiec, we are crossing the Koszarawa river and going to the outcrops along the eastern bank of the Soła river across the Solali paper plant nearby the brewery. These outcrops it is possible to observe the strongly folded and faulted, light-coloured Upper Jurassic–Lower Cretaceous Cieszyn Limestone Formation and dark teschenite sills.

In the western part of the Polish Carpathians region, between Cieszyn and Żywiec, is the best to study of the uppermost Jurassic and lowermost Cretaceous debris flow deposits of the Silesian Unit (Fig. 24). According to the newest lithostratigraphical scheme of the western part of the Outer Carpathians (Golonka *et al*., 2008) they belong to Vendryně Formation, Cieszyn Limestone Formation and Hradište Formation (Cisownica Shale Member) (Figs 28, 29) and cropping out in the Żywiec profile along the bank of the Soła River (Figs 29 upper part, 30).

Geological position of investigated outcrop(s) along the Soła River in Żywiec is very complicated by tectonic folding structures and was studied by several authors (Malik 1994; Słomka 2001; Golonka *et al*., 2006 with literature cited therein). To a large extent the pelitic limestones represent a *Coccolithus*-*Nannoconus* microfacies (Nowak, 1978), with numerous or scattered *Calpionella*, and also with large addition of calcitized radiolarians. Zoospores of *Globochaete alpina* Lombard, *Stomiosphaera, etc.* These limestones correspond to the rocks which arę known as “biancone” or “*Maiolica*”. The material of the detrital limestones consists of fragments or pebbles of zoogenic limestones, pelitic limestones, pelitic limestones with *Calpionella,* and also of fauna debris (plates and spines of echinoids, frag-mented Aptychi, valves of lamellibranchs, and brachiopods). Rather a big percentage of the materiał is also accounted for by foraminifers (including *Miliolidae* recognized by Szajnocha, 1922; *Trocholina),* cal-

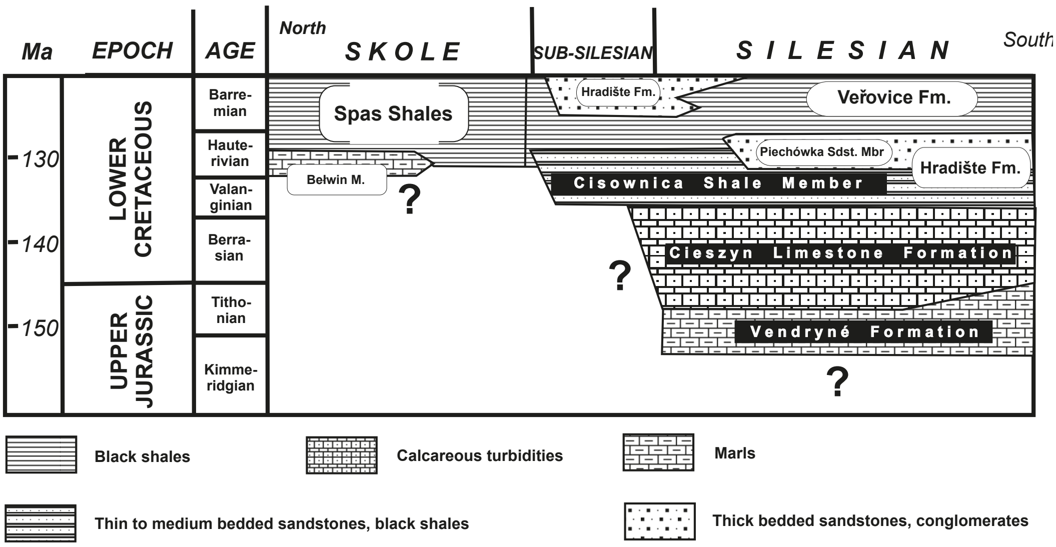

Fig. 28. Stratigraphical position of the Cieszyn Limestone Formation within Outer Carpathians Flysch units (after Słomka, 1986; modified)

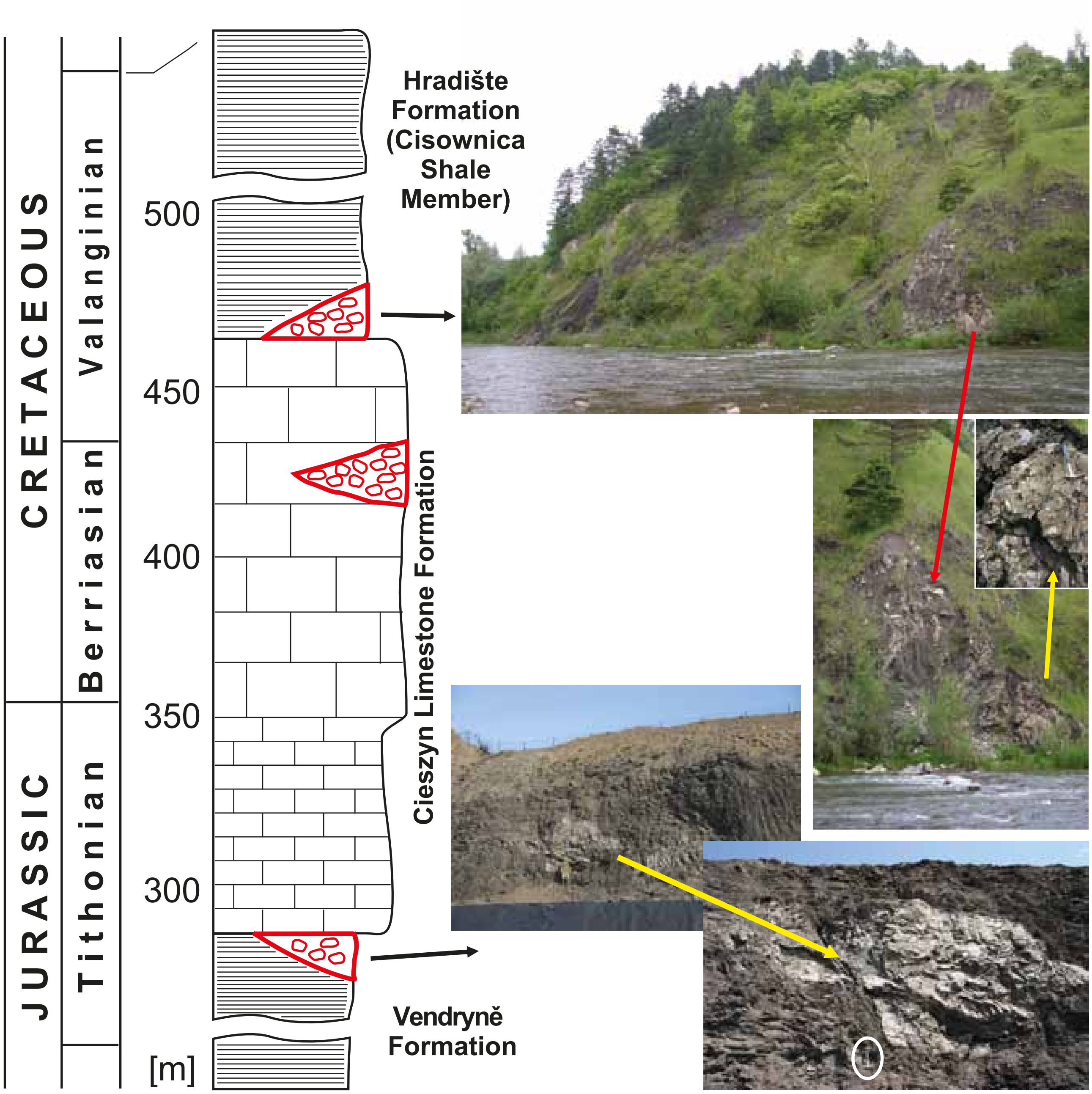

Fig. 29. General section of the Jurassic/Cretaceous boundary units of the Outer Carpathians in vicinity of Żywiec (after Słomka 1986a; changed and modified; Krobicki *et al.*, 2010) with position of debris flows and location of outcrops described in the text – Goleszów (lower photo), Soła River (upper photo); lithostratigraphy after Golonka *et al*. (2008)

careous algae, e.g. *Clypeina jurassica* Favre (Nowak, 1972, 1978), plates of planktonic crinoids *(Saccocoma* Agassiz). In the matrix of the bottom part of these limestones a large proportion of the materiał consists of *Cadosina* and *Stomiosphaera* (zone with *Colomisphaera cieszynica* Nowak) (Nowak, 1968); in the rest of the detrital limestones they are sparse and very scattered. But *Calpionella* are to be found everywhere here (Nowak, 1968, 1970, 1971, 1978), and there are zoospores of *Globochaete* throughout the whole sequence. Detrital quartz, autigenic quartz, and autigenic feldspars are an addition in the limestones (Peszat, 1959, 1966), while fragments of Upper Carboniferous coal from rather big concentrations in the higher part (Lower Cretaceous). The Cieszyn Limestone Formation display moderate diverse trace fossil assemblages dominated by *Chondrites targioni, Thalassinoides* and *Helminthopsis.* Other trace fossils (e.g., *Glocerichnus, Lorenzinia, Paleodictyon)* are rare. Upper part of turbidites are bioturbated up to 6.5 cm from the top. All these features, together with overall relatively light colour of marly shales, indicate well-oxygenated environment. The Soła River sequence of the Cieszyn Limestone Formation represents

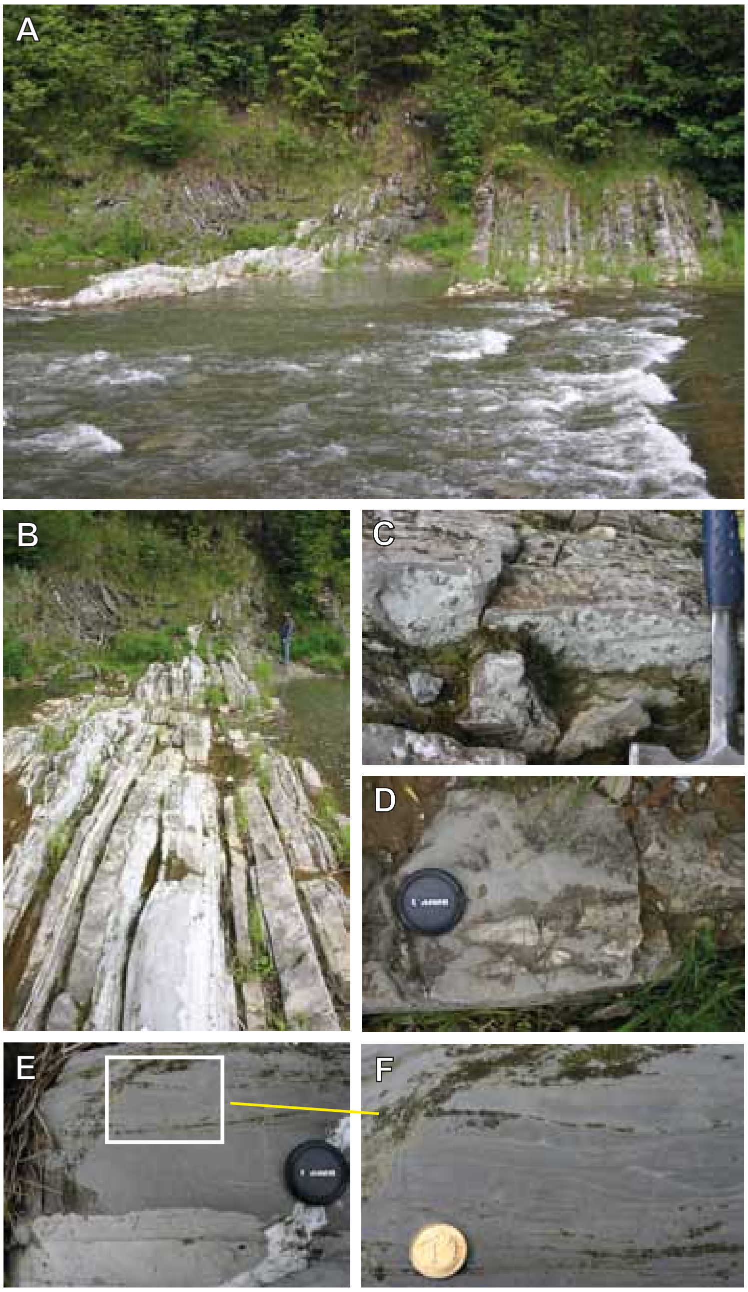

Fig. 30. Outcrop of thin-bedded micritic and detritic limestones of the Cieszyn Limestone Formation (**A**, **B** – Soła River – northern part) (with teschenitic sill), with numerous trace fossils on the soles of beds (**C**), with rare cherts (**D**) and delicate ripplemarks in fine-grained limestones (**E**, **F**) (after Waśkowska-Oliwa *et al.*, 2008)

a more proximal part of a vast submarine fan whose material was derived from the northern margin of the Carpathian basin.

More southern part of the long outcrop is built by mass movement deposits (up to 30 m thick) represented by dark grey/black marly shale matrix with abundant exotic rocks, both magmatic/metamorphic and sedimentary one (Carboniferous coals including); typical exotic-bearing gravelstone of the Cisownica Shale Member of the Hradište Formation (Valanginian in age). Some characteristic exotics are of uppermost Jurassic/lowermost Cretaceous deposits originated a little bit earlier in the Proto-Silesian Basin or surrounding regions. Sedimentological analysis suggests rapid sedimentation of these mass movement deposits with full transitional spectra from olistostromes to debris flows (Tokarski, 1947; Słomka, 1986b; Golonka *et al.*, 2006) (megaturbidites – *sensu* Malik, 1994). The thickness of debris flow deposits ranges from 2.5 to 30 meters. The share of the clastic framework does not exceed 30%. These sediments can be correlated with the facies A1.3 of Pickering *et al.* (1986) and facies GyM of Ghibaudo (1992). They include numerous fragments and pebbles of detrital and pelitic limestones of the Cieszyn Limestone Formation, organodetrital limestones, marly shales, Carboniferous and metamorphic rocks – granitic gneisses, gneisses and crystalline schists. Pebbles are randomly arranged in a mass of structureless, hard, marly silt. Generally, both the clays and embedded lumps of limestone have bends and folds closing towards the north suggesting that the sliding mass moved from the south.

Such type of redeposited material in olistostromes/debris flows indicates the building of the Silesian Ridge during the initial development of the active cordillera, at least

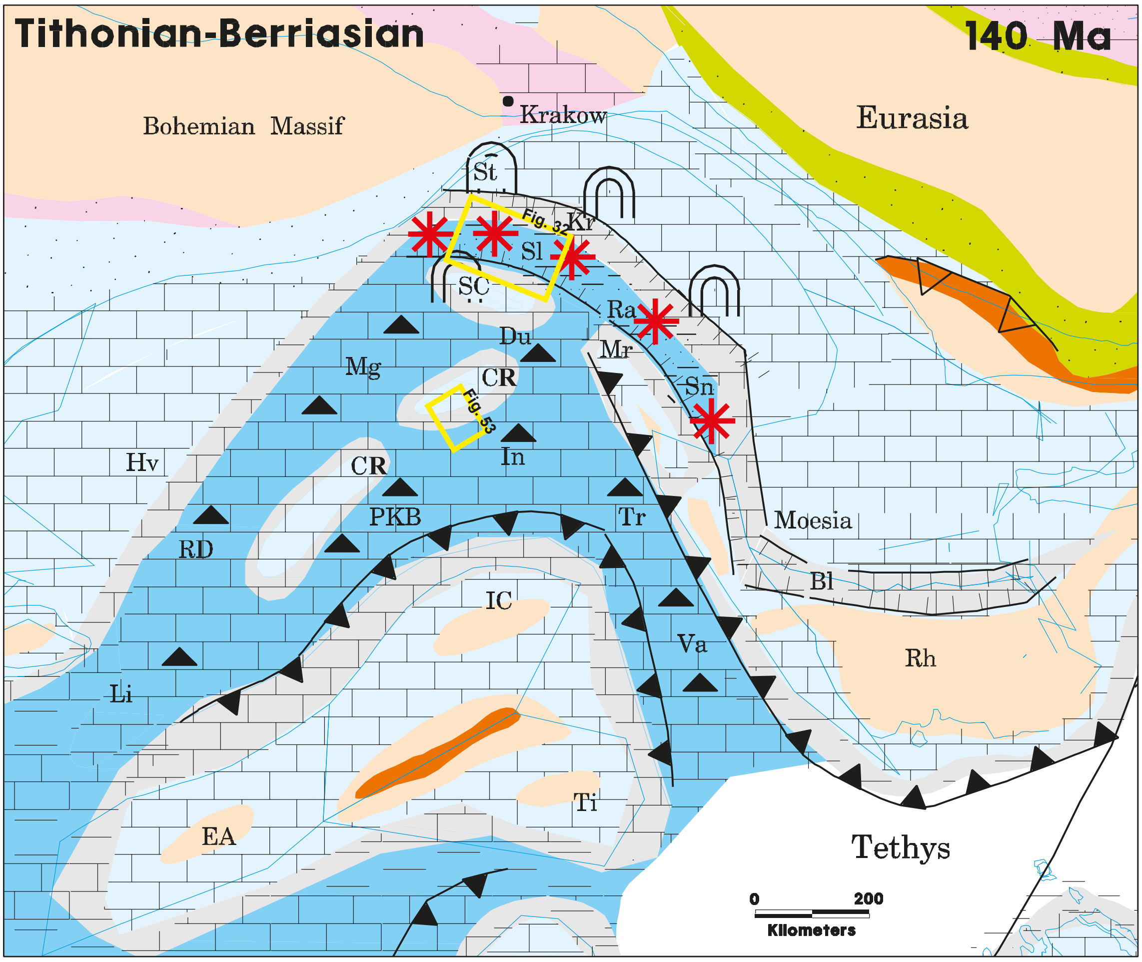

Fig. 31. Palaeoenvironment and lithofacies of the circum-Carpathian area during the latest Jurassic–earliest Cretaceous; plates position at 140 Ma (modified from Golonka *et al.*, 2006) with occurrence of rift-related magmatism (red stars). Abbreviations, Bl – Balkan rift; CR – Czorsztyn Ridge; Du – 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 – Ńtramberk Klippe; Ti – Tisa plate; Tr – Transylvanian Ocean; Va – Vardar Ocean. Explanations of colours and symbols – see Fig. 11

since Tithonian–Berriasian times. Tectonic activity caused the uplift of the Silesian Ridge and slope and continental rise of the Proto-Silesian Basin The deposits of the Cieszyn Limestone Formation were eroded again and redeposited as debris flows. Much greater participation of the coarse-grained facies of the upper part of the Cieszyn Limestone Formation and the appearance of mass-movement debris-flow deposits containing the fragments of the older rocks and exotics (both metamorphic and Paleozoic sedimentary rocks) suggest a higher rate of uplift during the latest Jurassic-earliest Cretaceous (Neo-Cimmerian) activity and "cannibalism" of the Proto-Silesian Basin (comp. Matyszkiewicz & Słomka 1994; Waśkowska-Oliwa *et al.*, 2008). Tectonic movements of the Silesian Ridge (and probable also opposite – Baška-Inwałd Ridge) were presumable connected with the development of initial rifting in the Proto-Silesian Basin, as documented by the presence of teschenitic magmatism (Grabowski *et al.*, 2003; Waśkowska-Oliwa *et al.*, 2008 with references therein) (Figs 31, 32). Such submarine magmatic processes took place mainly during Early Cretaceous times during first step of opening of the Proto-Silesian Basin (Ślączka & Słomka, 2001; Golonka *et al.*, 2006). During the Early Cretaceous tectonic activity a part of the basin was uplifted together with the Silesian Ridge (Cordillera) and Cieszyn Limestone Formation in this part was redeposited by debris flows (Słomka, 2001). Appearance of mass-movement debris-flow deposits containing the fragments of the older Cieszyn Limestone Formation and exotics of the basement rocks testify to higher rate of uplifting movements connected with Neo-Cimmerian activity (Osterwald Phase). The Early Cretaceous development of the Silesian Basin, perhaps from rifting into spreading phase, as suggested by this teschenitic magmatism (Narębski, 1990; Lucińska-Anczkiewicz *et al.*, 2000) was probably another effect of this Osterwald Phase.

In more northern part of this outcrop occur typical deposits of the Vendryně Formation with numerous, synsedimentationary deformed hard clasts of marls, which were interpreted as resedimentation effect, quite similar to the same-age deposits from Goleszów quarry (Malik, 1994). Chaotic type of sedimentation dominated during Late Jurassic times indicating early stages of the Proto-Silesian Basin opening with increased tectonic activity. The detritical material was supplied from two sources: from the Baška-Inwałd uplift separating the Proto-Silesian Basin and the Bachowice Basin located within the North European Platform, and from the island arcs within the Silesian Ridge separating the Proto-Silesian Basin and the Alpine Tethys (AT). The biogenic material originated within shallow-water reefal and carbonate platform zones was transported by turbiditic currents from the uplifted structures on the Proto-Silesian Basin margins into the deeper zones of this basin. Both the calciturbidites and calcifluxoturbidites formed, constituting the main lithosome within the younger lithostratigraphic unit – Cieszyn Limestone Formation. These deposits represent the oldest turbiditic currents sedimentation known from the Polish Outer Carpathian Basin.

Neo-Cimmerian tectonic events took place both in the AT and Proto-Silesian Basin. A big geotectonic reorganization, known as the Walentowa Phase, took place in AT during the latest Jurassic-earliest Cretaceous (Neo-Cimmerian) movements resulting in extensive gravitational faulting. Several tectonic horsts and grabens, documented by facies diversification, were formed. These rejuvenated some older structures and Middle/Late Jurassic (Meso-Cimmerian) faults which caused uplift of the shallow intrabasinal Czorsztyn pelagic swell (see below). The overregional significance of this geodynamic episode in the northernmost margin of the Tethyan Ocean is documented also by foundation of the Proto-Silesian Basin.

Such synsedimentary mass movement deposits are a key to understanding tectonic activity of the basins during their geotectonic history. Distribution of sedimentary breccias, mass flows; redeposited clasts are the main objects, which indicate time and mechanisms of origin of tectonic movements within sedimentary basins. Pulses of such activity are connected with wide-oceanic remobilization and are well known in several parts of the whole Alpine Europe. A lot of places of this region are full of very well recorded evidence of synsedimentary movements which originated during Jurassic–Early Cretaceous times. Other sedimentary features like: neptunian dykes, omission surfaces, condensation beds, redeposited shelly fauna, clastic sediments input to pelagic deposits as submarine wedges, olistostromes/olistoliths etc. also support such events. Such effects are strictly connected with activation and mobility of basin bottoms, especially during strong Alpine phases of tectonic revolutions, mainly of Middle–Late Jurassic/earliest Cretaceous (Meso- and Neo-Cimmerian) movements. Our knowledge on these types of tectonic activity in Polish part of the Carpathians is well documented both in the Outer (Flysch) Carpathians and in the Pieniny Klippen Belt.

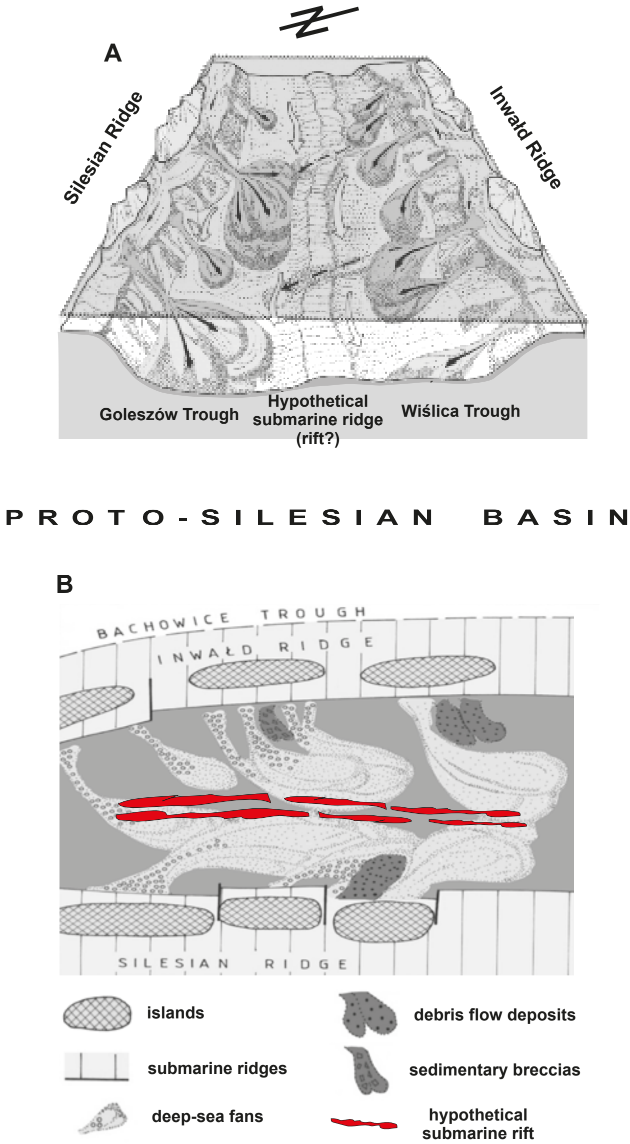

Fig. 32. Palaeogeographical blockdiagram of sedimentation of the oldest flysch deposits in the Proto-Silesian Basin (Jurassic/Cretaceous transition – Tithonian/Berriasian) (**A**) and its hypothetical palaeogeographical sketch (**B**) (after Słomka, 1986a; slightly modified)

In conclusion, Latest Jurassic–Earliest Cretaceous tectonic events took place both in the Pieniny Klippen Basin (AT) and Proto-Silesian Basin documenting over-regional significance of this geodynamic episode in the northernmost margin of the Tethyan Ocean. A big geotectonic reorganization, known as Walentowa Phase, took place in these two regions during Neo-Cimmerian resulting in extensive gravitational faulting. Several tectonic horsts and grabens were formed, rejuvenating some older faults which raised shallow intrabasinal Czorsztyn pelagic swell again and are documented by facies diversification. Additionally, these movements separated the basin into different zones with their own water circulation patterns, probably of an upwelling type. Volcanic activity (both intra-plate alkaline volcanism in the Ukrainian part of the Pieniny Klippen Belt – Krobicki *et al*., 2005, 2008 and Proto-Silesian rift-related magmatism) (Figs 31, 32) and change of oceanographical regimes (upwelling currents) most probable reflect also this geotectonic phenomena.

POLISH CARPATHIANS *VERSUS* JURASSIC ALPINE TETHYS (AT)

▹FIGS 5–7, 23, 28, 31–33

Jan GOLONKA & Michał KROBICKI

The Silesian Unit (Basin) of the Outer Flysch Carpathians are composed of Jurassic to Early Miocene flysch sequences. During the Alpine orogenic processes in Miocene times, the north-verging nappes were detached from their original basement (Ślączka *et al*., 2006). The Proto-Silesian (Severin-Moldavidic) Basin originated during Late Jurassic times together with syn- and post-rift deposits (Figs 6, 7). Part of this basin was included into the Silesian Unit, one of the Outer Carpathian nappes. The Silesian Ridge (Figs 7, 31, 32) has been an uplifted area, originally as a part of the North European platform separating during Jurassic-Early Cretaceous times AT and the Proto-Silesian Basin. Now it is known only from exotics and olistoliths occurring within the various allochthonous units of the Outer Carpathians. The shallow-water marine sedimentation prevailed on the Silesian Ridge during Late Jurassic and earliest Cretaceous times. The carbonate material was transported from the ridge toward the Proto-Silesian Basin. This basin developed within the North European Platform as rift and/or back-arc basin. Its basement is represented by the attenuated crust of the North European plate with perhaps incipient oceanic fragments. The sedimentary cover is represented by several sequences of Late Jurassic–Early Miocene age belonging recently to various tectonic units in Poland and Czech Republic. The Baška-Inwałd Ridge has been located on the opposite side of the Proto-Silesian Basin and its slope contains mainly carbonate deposits and originated as shoulder uplift separating the Bachowice Basin from the Proto-Silesian Basin. The Vendryně Formation (Kimmeridgian–Tithonian/Early Berriasian) represents the oldest deposits of the Silesian Unit (Nappe) (Figs 23, 28) (lithostratigraphy after Golonka *et al*., 2008). This formation is built of dark-grey marly shales with rare intercalations of redeposited detritical limestones containing fossils of shallow-marine fauna, mainly echinoderms and mollusks. Deposits forming huge sliding slices with numerous deformation structures indicating chaotic type of sedimentation occur within the profile. The rocks of the Vendryně Formation are exposed both in the classic type locality in Vendryně on the Czech side of the Silesian Unit and on the Polish territory (for example – in the abandoned quarry in Goleszów). They are covered by the Cieszyn Limestone Formation (Late Tithonian–Middle Valanginian) which are represented by (organo)detrital and pelitic limestones intercalated by grey/black shales. Limestones are usually thin-bedded with typical features of turbiditic deposits: sharp erosive base of beds, gradational fractionation, ripplemark-convolute cross-lamination in the top parts of beds, sometimes with numerous trace fossils on the soles of beds, flutcasts, delicate ripplemarks in fine-grained type of limestones, resedimented shales and carbonate clasts (often with fractionation) etc., and additionally with rare cherts (comp. Waśkowska-Oliwa *et al*., 2008). The younger, Middle Valanginian–Barremian Hradište Formation is represented mainly by grey/black shales with intercalations of very thin- to medium-bedded calcareous sandstones (lower part – Cisownica Shale Member; formerly Upper Cieszyn Beds) and thick-bedded sandstones and conglomerates (upper part of formation – Piechówka Sandstone Member; formerly Grodziszcze Sandstones). Also some rocks, formerly known as Wierzowskie/Veřovice Beds are now included into the Hradište Formation. The Hradište Formation is covered by the Veřovice Formation (Aptian) represented by dark and black shales and mudstones rich in organic matter. The younger is Lhoty Formation (Albian) representing synorogenic flysch-type deposits.

The AT, which constitutes important palaeogeographic elements of the future Pieniny Klippen Belt and Outer Carpathians (Figs 7, 31), developed as an oceanic basin, a continuation of the Central Atlantic, during Jurassic as a result of the Pangea break-up (Fig. 5). The Mesozoic rifting events caused the origin of oceanic type basins along the northern margin of the Tethys. The Inner Carpathian plate was detached from the Eurasian margin by this AT as part of the separation of Eurasia from Gondwana. It was also dissected by the rift system. The deeper water sediments, like radiolarites, were deposit in these rifts, while shallower water carbonate sedimentation prevailed in the uplifted areas. The central Atlantic and AT went into a drifting stage during the Middle Jurassic times. The oldest oceanic crust in the Ligurian–Piedmont Ocean was dated as late as the Middle Jurassic in the southern Apennines and in the Western Alps. Bill *et al*. (2001) date the onset of oceanic spreading of the AT by isotopic methods as Bajocian. The spreading phase follows the rifting during Early Jurassic times. The Jurassic rifting and spreading placed Triassic platform carbonate facies on the basin passive margins. Two major Late Jurassic basin, AT and Proto-Silesian Basin, were later included into the Carpathian thrust- and fold-belt. These basins were separated by Silesian Ridge. The NE part of AT was divided by the Czorsztyn Ridge into Pieniny Basin and Magura Basin, part of the Outer Carpathian Basins (Figs 7, 31). Major plate reorganization happened during the Tithonian time. The Central Atlantic began to expand into the area between Iberia and the New Foundland shelf. The Ligurian-Penninic Ocean reached its maximum width and the oceanic spreading stopped (Fig. 6). The Tethyan plate reorganization followed the global pattern. This reorganization was expressed by latest Jurassic/earliest Cretaceous tectonic movements, which affected both AT and Proto-Silesian Basin (Golonka *et al*., 2003).

The last stop of the first day of our excursion is in picturesque location with perfect view (if weather is good!) to almost whole Carpathians, including Babia Góra Mt, and general geology (Fig. 33).

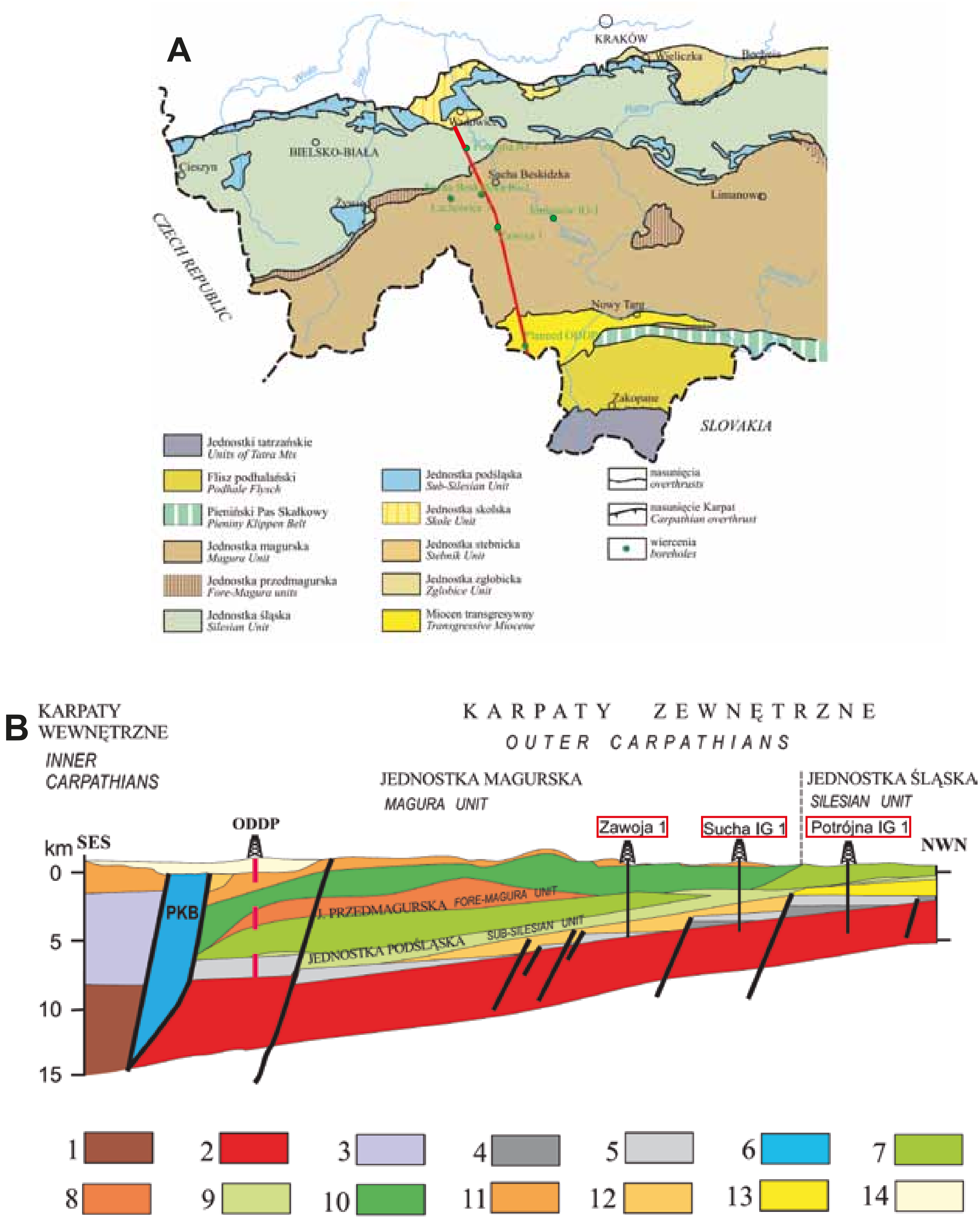

Fig. 33. Geological map of the western part of the Outer Flysch Carpathians (**A**) (after Oszczypko, 1998; Golonka *et al*., 2005; simplified) with line of cross-section (**B**)

Explanations of symbols: 1 – basement of the Carpathians; 2 – Proterozoic-Lower Paleozoic of the Bruno-Vistulicum; 3 – Mesozoic of the Inner Carpathians; 4 – Lower Paleozoic; 5 – Upper Paleozoic; 6 – Jurassic; 7 – Lower Cretaceous and Paleogene of the Silesian Unit; 8 – Cenomanian and Senonian; 9 – Upper Cretaceous and Paleocene of the Sub_Silesian Unit; 10 – Senonian and Paleocene; 11 – Eocene; 12 – Lower Miocene; 13 – Upper Miocene; 14 – Neogene

TATRA MOUNTAINS

▹FIGS 34–39

Michał KROBICKI & Jan GOLONKA

The Inner Carpathian Palaeozoic and Mesozoic rocks crop out in the Tatra Mountains (Figs 34–38). North of the Tatras, they are covered by the Central Carpathian Paleogene and known only from boreholes and geophysical data (Golonka *et al*., 2005b).

The Tatra Mts. 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 sq. km, with 175 sq. km lying in Poland and the rest in Slovakia. There are thirty-four summits with a prominence of at least 140m in the range that reach over 2000 m. Of these six reach 2500 m. The highest peak in the range Gerlach (Gerlachovsky stit) (2655 m), is in Slovakia.

The Tatra Mts 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 Paleogene (Figs 34, 40). The crystalline core form central and southern part of the Tatra Mts. Western, northern and north eastern parts are covered by autochtonous Mesozoic rocks (Fig. 34) and several allochtonous thrust sheets and small nappes. All these units are discordantly covered with a post-nappe transgressive succession of the Central Carpathian Paleogene Basin. The crystalline basement consists of the granitoids and the metamorphic envelope. Granitoids, represented by tonalities and granodiorites with subordinate amount of granites, originated during a continental collision between 360 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 Mts were affected by Variscan (Palaeozoic) and Alpine (Mesozoic–Cenozoic) tectono-metamorphic events. The best way to examine the Tatra's crystalline core is to take the trip by cable gondola car to the Kasprowy Wierch (1.987 m a.s.l.). The cable car constructed in 1936 and remodeled recently covers the distance of about 4.300 m and altitude difference about 900 m within 17 minutes. Kasprowy Wierch is located in the centre of the mountain range so everybody can enjoy magnificent panorama of both Western with Giewont, Czerwone Wierchy and Krywań as well as High (Eastern) Tatras is visible. The peak is built of the Paleozoic metamorphic rocks. The hiking trail will take more experienced tourists to Świnica (Figs 36, 37) and Orla Perć (*Eagle ridge*) with granitic mountain core. Less experienced tourists/geotourists can enjoy the trail downhill toward Gąsienicowa Hala and Zakopane. On this trail both metamorphic and sedimentary rocks can be observed. The site trip to Czarny Staw Gąsienicowy wills allows examining granitic rocks.

The Tatra Mts Mesozoic sedimentary rocks were deposited within the 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 palaeogeographic domains, partially reflected by present-day tectonic units: Tatric, Fatric, Veporic, Gemeric, Hronic and Silicic (Andrusov *et al*., 1973). The Tatra Mts rocks belongs to the Tatric, Fatric and Chronic domains. The Tatric domain is represented by the High-Tatric unit (Figs 34, 35) (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. 38). 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 Mt.), 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) (Fig. 35).

Fatric and Hronic domains are represented by the Krížna (Lower Sub-Tatric), Choč (Middle Sub-Tatric), and the (Strážov) (Upper Sub-Tatric) units occurring exclusively in thrust sheets, which overlie the High-Tatric units (Lefeld, 1999) (Fig. 34). 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 *Ammonitico Rosso*-type and *Maiolica* limestones. Basinal marlstones and limestones dominated in the Lower Cretaceous sediments in the western part of the Tatra Mts, while in the eastern part (Belanské Tatry) massive carbonates occurred (Lefeld, 1985; Wieczorek, 2000; Golonka *et al*., 2005b; Uchman, 2009; Jach *et al*., 2017) (Fig. 35).

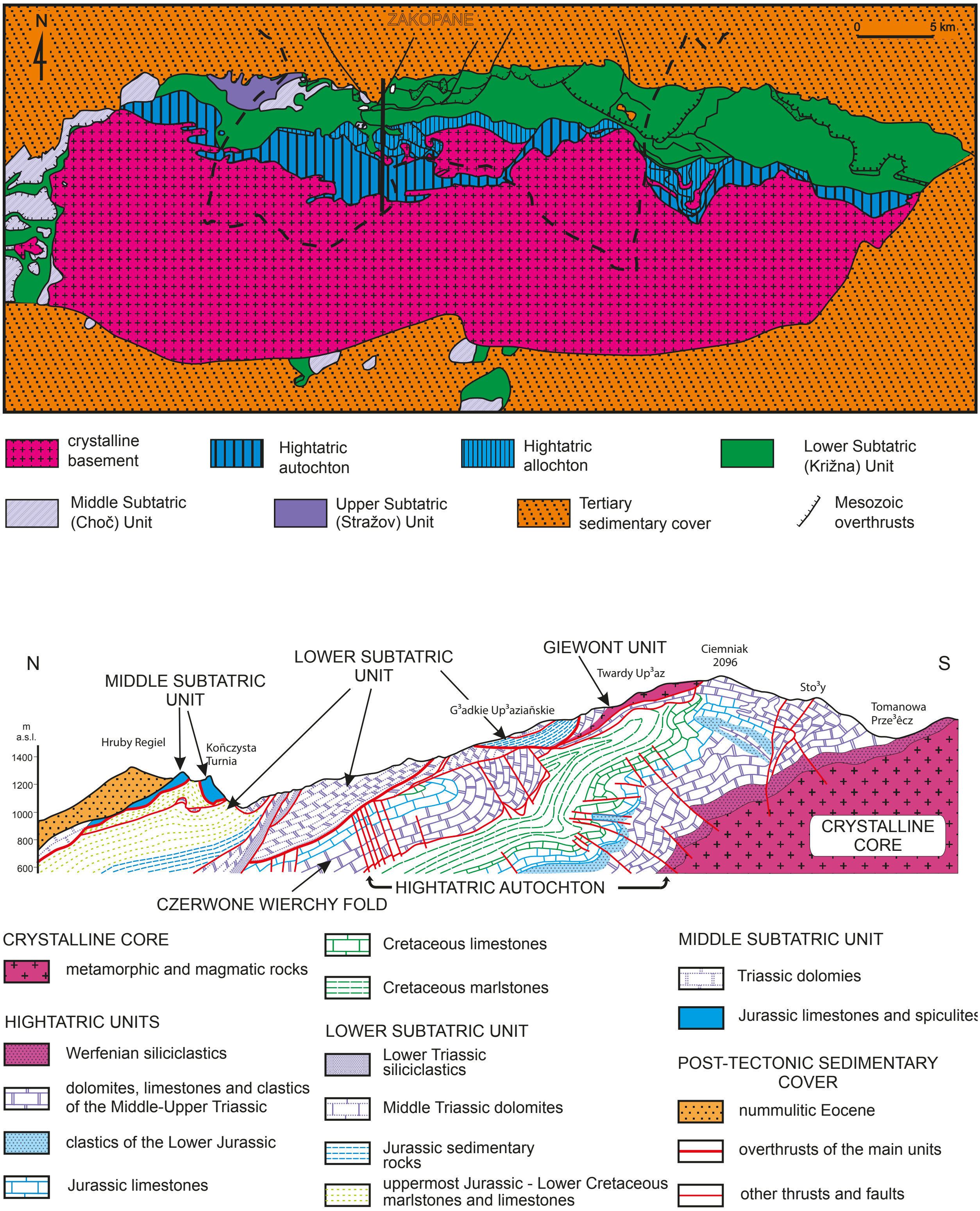

Fig. 34. Simplified geological map of the Tatra Mts (based on Bac-Moszaszwili *et al.*, 1979) (upper part) and geological cross-section along the eastern slopes of the Kościeliska Valley (after Kotański, 1961; Uchman, 2006)

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 (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, b; Iwanow & Wieczorek, 1987; Jurewicz, 2005).

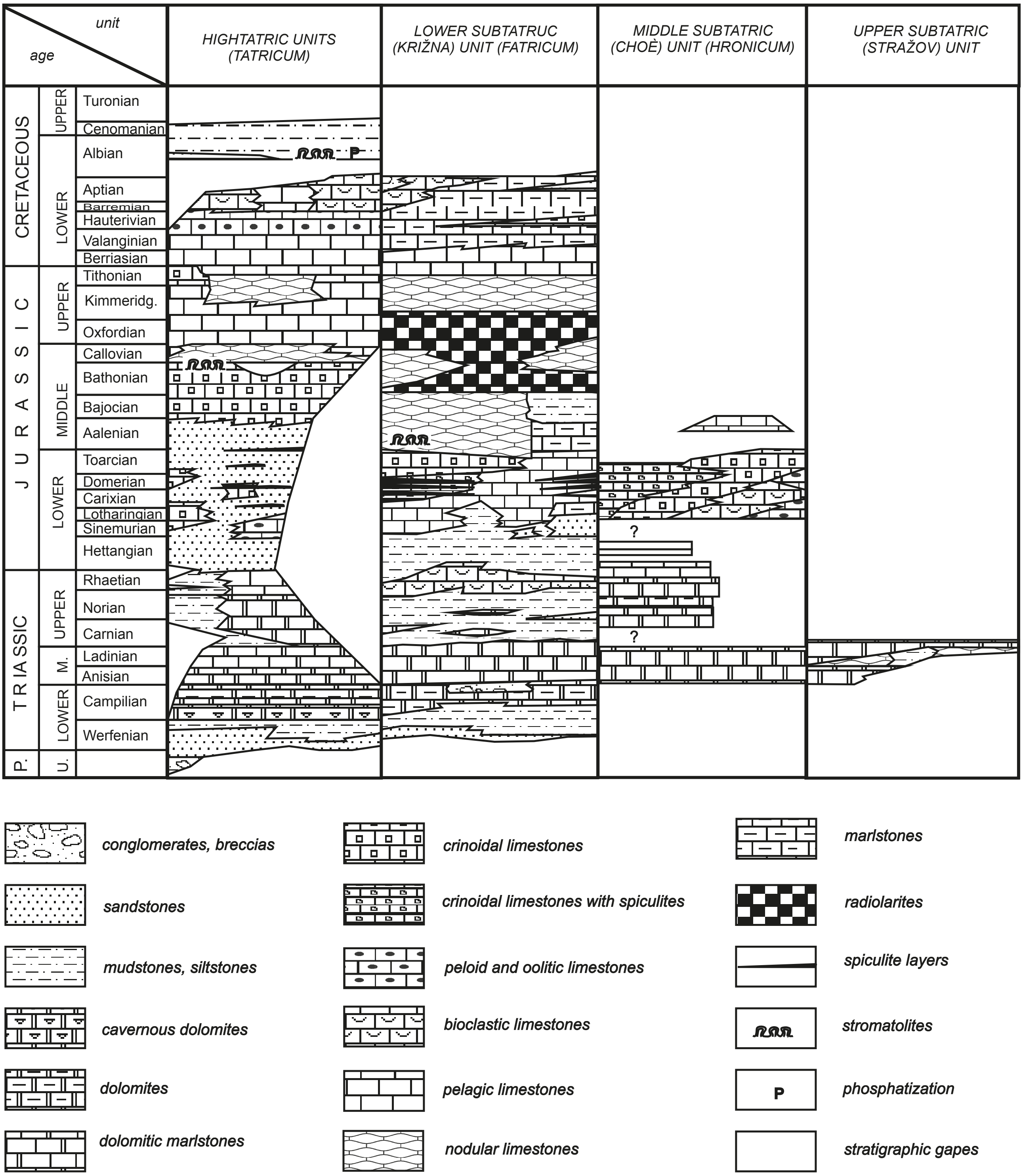

Fig. 35. Major Mesozoic facies of the Tatra Mts (after Uchman 2004, changed)

The best way to examine the Tatra's sedimentary rocks is to visit Kościeliska nd Chochołowska valleys. From the parking lot and bus stop in Kiry (Kościelisko) about 7 km from Zakopane town centre, the 6 km long trail will take us through Brama Kantaka, Brama Kraszewskiego, Polana Pisana to mountain lodge at Hala Ornak in the higher part of Kościeliska Valley. The entrance to the Valley is built ef Eocene Nummulitic limestones, further to the valley the Mesozoic rock of Krížna (Lower Sub-Tatric) nappe are exposed. They were beautifully sculptured by karst phenomena.The Western Tatra peaks south of Hala Ornak are built of Paleozoic metamorphic rocks. Tatra Mts belong today to Tatra National Pak and are well protected, by in the past, until XIX 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 Kościelika Valley with Kuźnice, the site of XIX century steel, today the site of lower station of Kasprowy Cable car.

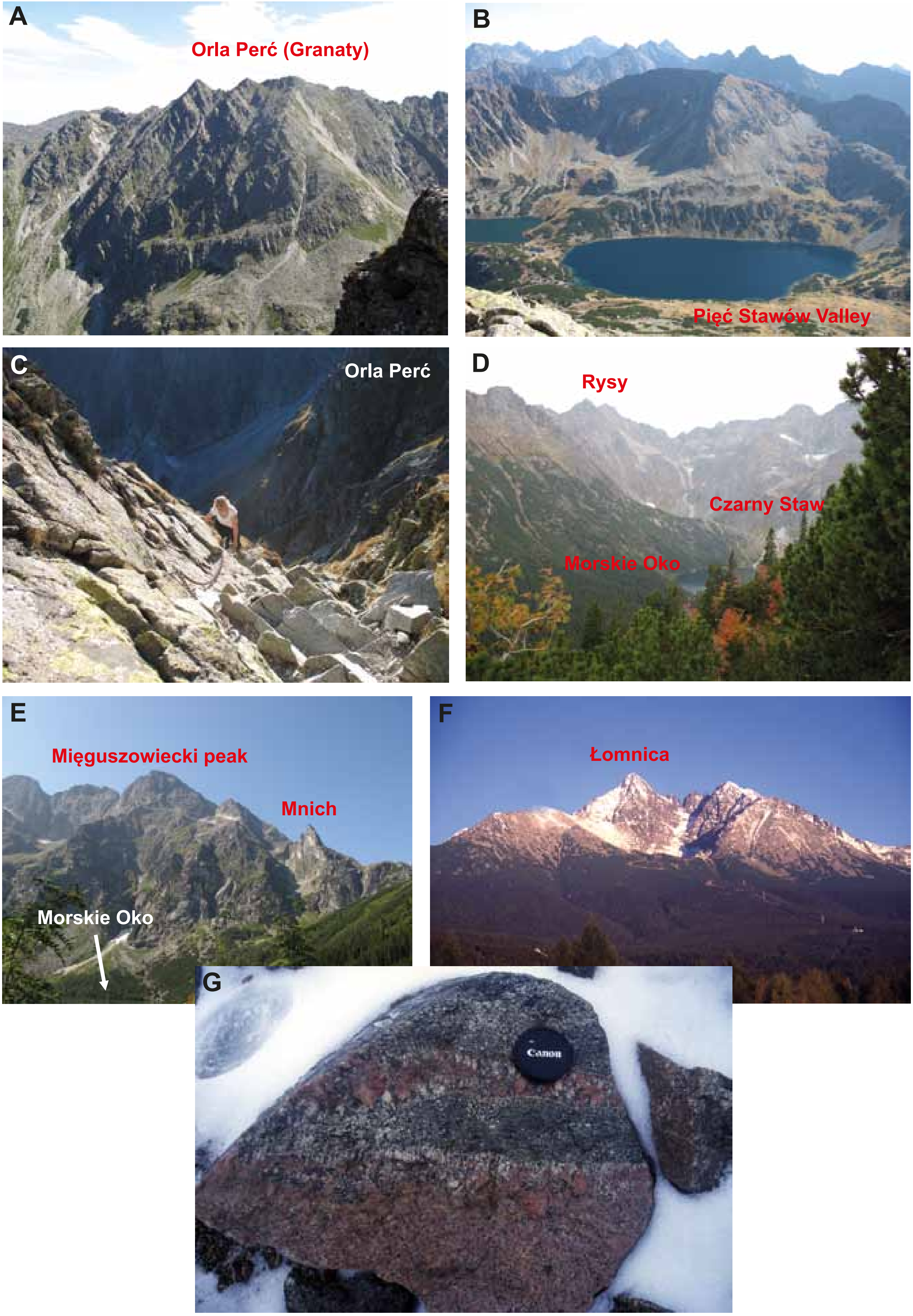

Fig. 36. Landscapes of the High Tatra Mts with Hercynian granitoidic (and pegmatitic – **G**) rocks

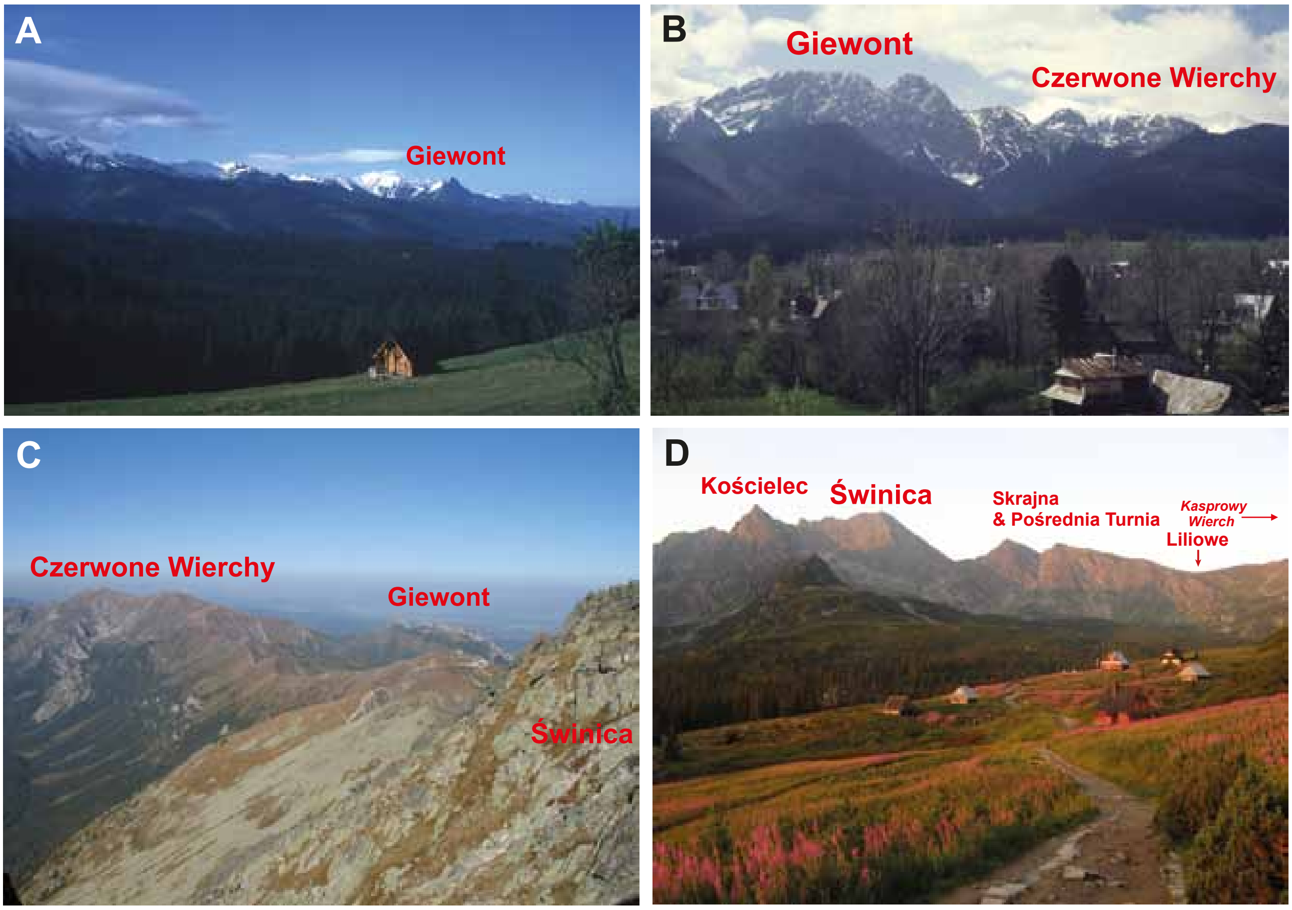

Fig. 37. Panoramic view of the Tatra Mts with most characteristic shape of the Giewont Mt (from north – **A**, **B** and from south – **C**) and the Gąsienicowa Hala area during sunrise (**D**), including Liliowe Pass (**D**)

The last episode of geological history of the Tatra Mts is connected with Quaternary glaciations. 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. The most beautuful glacial lake are located in the broad High Tatra valley, like Sucha Woda Valley with mentioned above Czarny Staw Gąsienicowy, Five Lakes Valley or Rybi Potok Valley with Morskie Oko Lake (Fig. 36). 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 tu public in Poland, including Jaskinia Mroźna (the Frosty Cave) with electric ligh, however the most interesting cave accessible to the public is Belianska jaskyňa in Tatranska Kotlina in Slovakia.

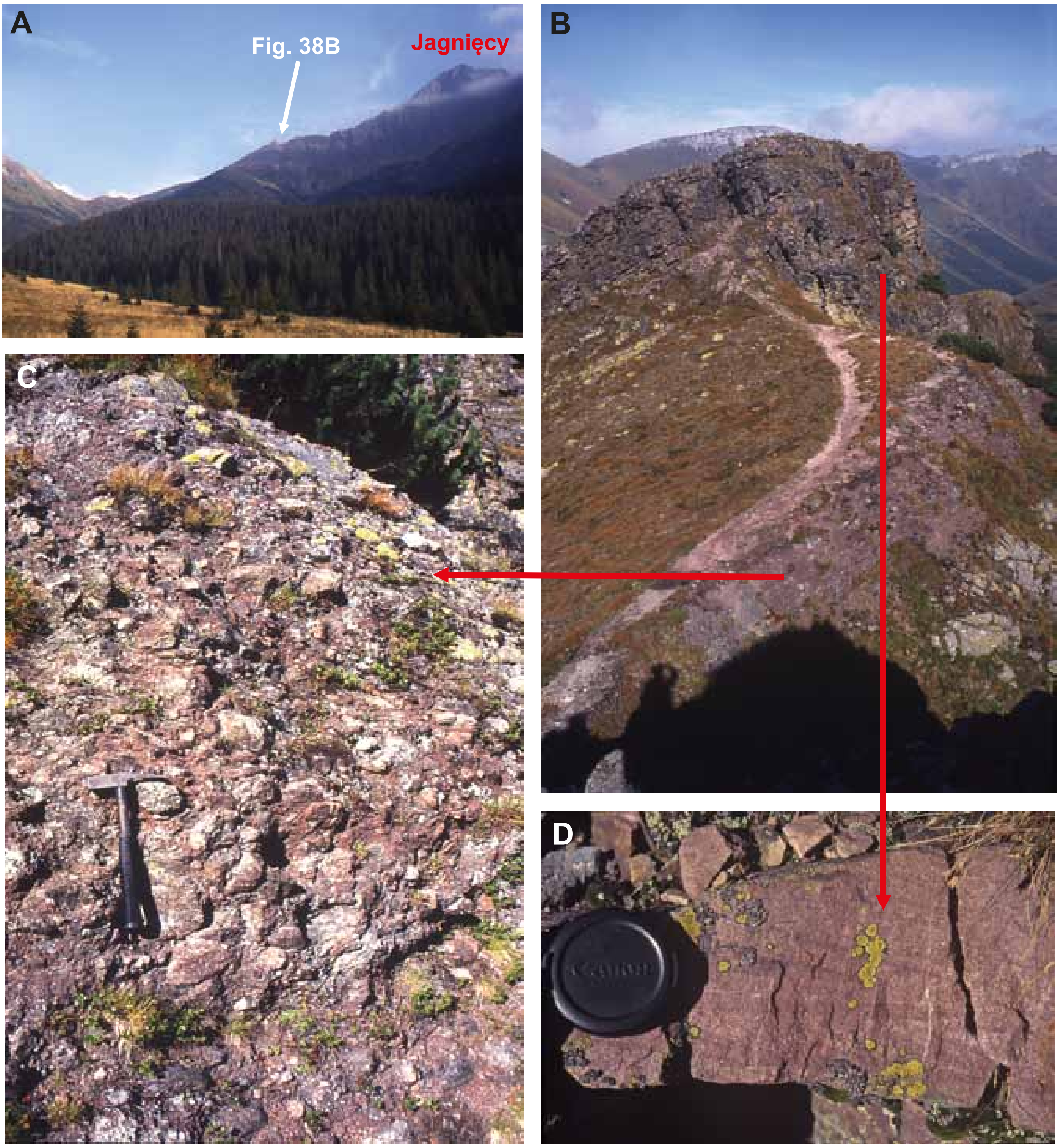

Fig. 38. Panoramic view of the Jagnięcy Mt (**A**) (Slovakian part of the Tatra Mts) with granitoidic basement covered by the oldest sedimentary rocks in the Tatra Mts – *Verrucano* facies (**B**, **C**) and the lowermost Triassic Verfen-type quartzic sandstones with cross-bedding structures (**D**)

FIELD TRIP ▸ STOP 4

LILIOWE PASS (LOWER TRIASSIC, NAPPE STRUCTURES)

▸FIG. 37D

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 Mts. The peak of the Kasprowy Wierch is built by Hercynian (Carboniferou s) crystalline rocks that form an isolated tectonic island (so-called "Goryczkowa hat" = Goryczkowa Crystalline Island) 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).

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.

FIELD TRIP ▸ STOP 5

POD CAPKAMI QUARRY ("NUMMULITIC EOCENE")

▸FIGS 34, 39–40

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After our visit on the top of Kasprowy Wierch and vicinity, we came back down by cable car and close to Zakopane sport center (near ski jumps – Mała, Średnia and Wielka Krokiew) we can visit one of the best location of the Paleogene, post-orogenic deposits, represented by *Nummulites*-bearing carbonates (so-called "Nummulitic Eocene").

The Paleogene is subdivided to two parts in the Podhale region. Older is represented by the so-called "Tatra Eocene" (Middle–Upper Eocene) and consists of conglomerates (Passendorfer, 1958, 1983) and carbonate rocks (nummulitic limestones) (Roniewicz, 1969, 1979; Bac-Moszaszwili *et al*., 1979; Kulka, 1985; Bartholdy *et al*., 1995, 1999) (Fig. 40). Younger complex (Oligocene) is represented by flysch deposits (the Podhale Flysch) (see below). A stratigraphical gap is suggested between the "Tatra Eocene" and the overlying flysch deposits (Gedl, 1999a). According to Olszewska & Wieczorek (1998) the whole Paleogene deposits in Podhale region represent four distinct facies: the Middle Eocene basal conglomerates originated as deltaic and cliff-type facies (0–100 m in thickness), the Upper Eo-

Fig. 39. Paleogene (Eocene) sedimentary cover of the Mesozoic nappes in the Pod Capkami quarry with typical nummulitic rudstones of so-called "Nummulitic Eocene"

cene sublittoral nummulitic limestones of carbonate platform (max. 100–150 m), the uppermost Eocene hemipelagic marls with planktic foraminifera (ca. 20 m), and the Oligocene flysch as typical turbiditic sedimentation with coarsening upward sequence (2700–3400 m).

In abandoned Pod Capkami quarry thick- and medium-thick well-bedded dolomites and limestones occur with several intercalations of nummulitic-rich rudstones represented one of the typical facies of Eocene carbonates in the Tatra Mts. These rocks were object of very intensive both stratigraphical, palaeontological and palaeoecological investigations since at the beginning of XX century (e.g., Kuźniar, 1910; Małecki, 1956; Bieda, 1959, 1960, 1963; Alexandrowicz & Geroch, 1963; Roniewicz, 1969; Kulka, 1985; Głazek *et al*., 1998; Olszewska & Wieczorek, 1998; see also – Machaniec *et al*., 2009 with reference cited therein). Generally, relatively shallow-water palaeoenvironments took place during sedimentation of these deposits with several parts of ancient carbonate platform, based on both of facies/sequence studies (e.g., Bartholdy *et al*., 1995, 1999) and palaeocological evaluations (e.g., so-called Arni model – Kulka, 1985).

FIELD TRIP › STOP 6

BUKOWINA TATRZAŃSKA (PODHALE PALEOGENE FLYSCH, GEOTERMY)

› FIGS 40–47

Michał KROBICKI [1,2], Jan GOLONKA [2] & Józef CHOWANIEC [1]

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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 Paleogene. 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 (Olszewska & Wieczorek, 1998; Golonka *et al*., 2005b). Directly on the transgressive deposits of the calcareous Eocene there occur stratigraphically younger strata of the Paleogene, i.e. the Podhale flysch. The oldest part of this flysch – the Zakopane Formation (Lexa *et al*., 2000) is well exposed in the streams in the marginal part of the Tatra Mts. The uplift of the Tatras, dated using apatite fission tracks, took part probably during the Miocene (15–10 Ma) (Golonka *et al*., 2005b).

The Podhale region is very attractive area both by its close connection to Tatra Mts on the one, southern side, and to the Pieniny Mts on the opposite, northern side. Such geographical position of this region is useful to visit both mentioned regions, which are prime attractions in southern Poland (Krobicki & Golonka, 2008). But Podhale area with gentle hills, wild forests, clean rivers and streams and perfect landscapes to both, southern and northern directions, offers also high spectacular geological objects of the Carpathians, by study of Paleogene flysch deposits. Maybe first impression indicates very monotonous sequences of flysch-type rocks, but several features of these sedimentary units and their tectonic regimes surprisingly give us introduction to understanding complicated history of the origin of the Podhale Paleogene Flysch.

The Podhale Paleogene Flysch belongs to larger unit, so-called Central Carpathian Paleogene Basin (Marschalko, 1987), which lies within the West Carpathian Mountain chain (Figs 3, 4, 40) (Soták & Janočko, 2001; Soták *et al*., 2001; Golonka *et al*., 2005a with literature cited therein). Polish part of this basin was filled mainly by very thick sequence of flysch-type rocks, comprising up to *ca.* 3400 m of conglomerates, sandstones, mudstones and siltstones of turbiditic submarine fans origin (Westfalewicz-Mogilska, 1986; Marschalko, 1987; Wieczorek, 1989; Soták & Janočko, 2001). In polish nomenclature this region was classically termed as the Podhale Flysch (Radomski, 1958). Recently it occurs between Pieniny Klippen Belt (on the north) and Tatra Mts (on the south). Flyschoidal rocks occupied depression and built wide synclinal structure (Fig. 40).

The Podhale Basin is considered as a fore-arc basin, located at the NE border of the North Pannonian unit, probably developed above a B-subduction zone of the European plate (Tari *et al*., 1993). Eocene volcanic rocks around the Balaton lineament in Hungry indicate some possibilities of connection with this subduction. Some palaeogeographical reconstruction suggested that Central Carpathian region has been located some hundreds kilometers to the southwest of its present position (Csontos *et al*., 1992). During the Oligocene time the convergence between Africa and Eurasia took place (Figs 8–10). The tectonic collision of the Apulia and the North European platform (Eurasia in general sense) has been the main process resulted in transport of several plates in Alpine-Carpathian system. From the global and European point of view it was connected with plate's reorganization during the Alpine orogeny and geodynamic reaction to the origin and movements of several accretionary prisms in Alpine-Carpathian arc (Golonka, 2004) (Figs 8–10). In this time metamorphism of the Penninic nappes in the Alps reached peak thermal conditions at about 30 Ma (Kurz *et al*., 1996). The Paratethys Sea was formed in Europe and Central Asia, ahead of the northward moving orogenic belts (Dercourt *et al*., 1993). The Paratethys Sea included orogenic foredeep as well as remnants of older oceanic basins and epicontinental platforms of the Peri-Tethys area. There was a transition from flysch to mo-

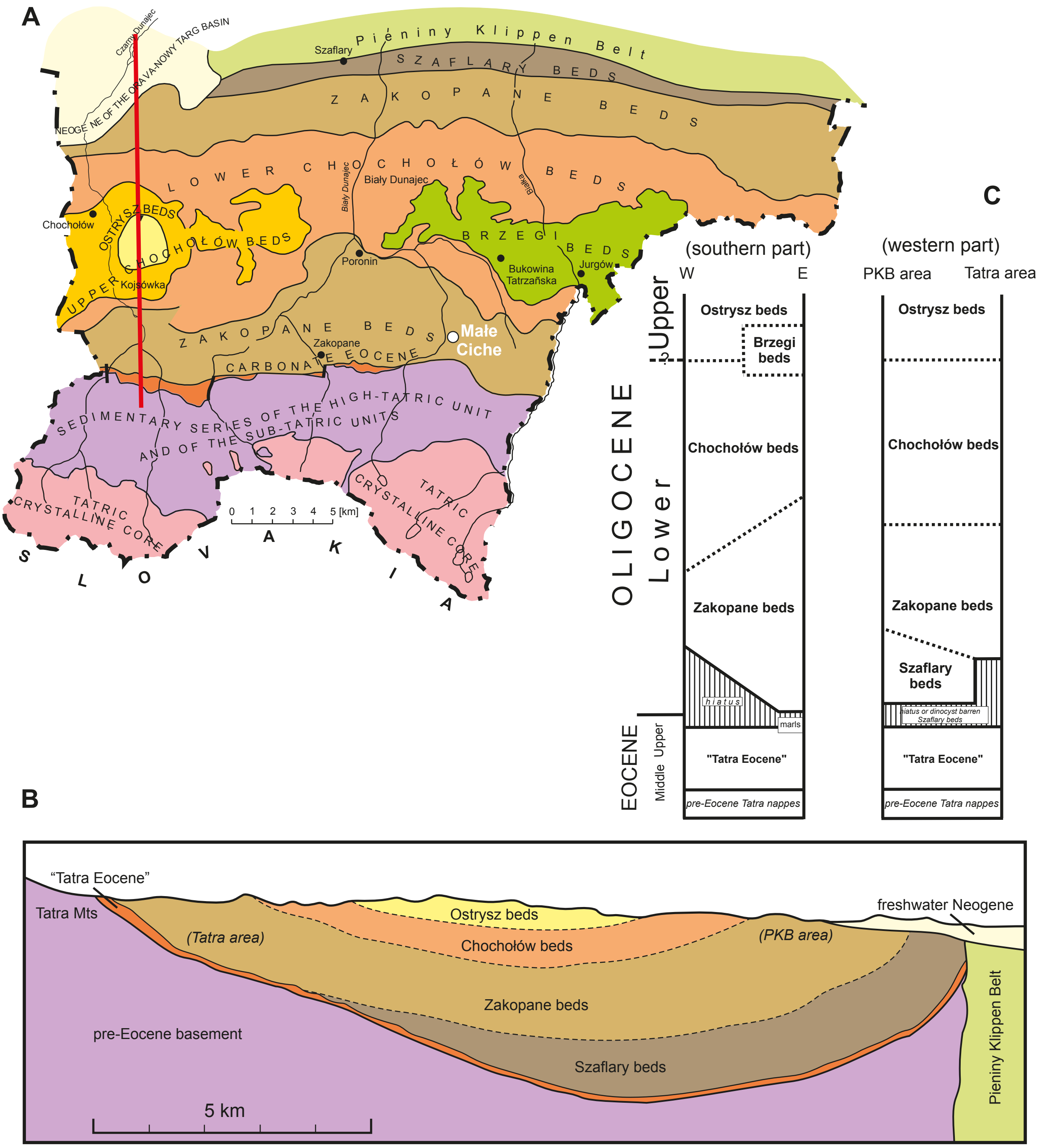


Fig. 40. Geological sketch of the Tatra Mts and the Podhale Paleogene Flysch (**A**) with marked of cross section (red line) (after, Chowaniec, 1989; Chowaniec & Kępińska, 2003; slightly modified), cross section by western part of area (**B**) (after Gedl, 2000a), and lithostratigraphy of the Podhale Paleogene (**C**) (after Gedl, 2000a)

lasse type of sedimentation in the basins in the Alpine-Carpathian realm (Golonka, 2004). A little bit later, during the Miocene times, collisions continued in the area between Africa and Eurasia. Apulia and the Alpine-Carpathian terranes were moving northwards, colliding with the European plate, until 17 Ma (Decker & Peresson, 1996). This collision caused the foreland to propagate north. The north to NNW-vergent thrust system of the eastern Alps was formed. Oblique collision between the North European plate and the overriding western Carpathian terranes led to the development of an outer accretionary wedge, the build-up of many flysch nappes (Outer Flysch Carpathians) and the formation of a foredeep (Ślączka, 1996; Kováč *et al.*, 1998). These nappes were detached from their original basement and thrust over the Palaeozoic–Mesozoic deposits of the North European platform (Fig. 3). This process was completed during the Late Miocenes time in the area of the Vienna Basin and then progressed northeastwards (Oszczypko, 1999; Golonka *et al.*, 2004, 2006b).

The Podhale region is the northern part of the Central Carpathian Paleogene Basin, which includes Liptov, Orava and Spiš depressions. It is built up of Paleogene strata underlain by mostly calcareous Mesozoic rocks. The lithostratigraphic section of these deposits has been recognized by several deep boreholes (Figs 3, 40). In subsequent years, the boreholes provided very advantageous information. The results of investigations showed that the sub-Paleogene substratum is an extension of geological-structural elements of the Tatra massif, to which the Sub-Tatric (Krížna, Choč) and High-Tatric nappes belong. It is generally accepted, based on different evidences both from the field and boreholes, that the Tatra Mts have been covered by the Podhale Paleogene flysch deposits, later eroded during the uplift of the Tatra massif. Moreover, in logs of some deep drillings (Sokołowski, 1973; Chowaniec, 1989) the facies elements similar to certain rock types of the Pieniny successions and deposits of uncertain affinity were found. After the retreat of the Late Cretaceous sea, a subsequent transgression took place in the Middle Eocene that resulted in the formation of conglomerates and limestones in the initial phase. These deposits form the basal member of the Podhale Paleogene. Then, typical flysch deposits were formed. Sediments of the calcareous Eocene are known from numerous natural exposures situated at the outlets of valleys draining the Tatra Mts and from drillings made in the Podhale Basin. Stratigraphically younger of the Paleogene, i.e. the Podhale flysch occur directly on the transgressive deposits of the calcareous Eocene. The largest thickness of the Podhale flysch, *ca.* 3000 m, was recorded in borehole Chochołów PIG-1 (Golonka *et al*., 2005a).

In the Slovak Orava, the Eocene–Oligocene sequence is known as the Podtatranská Group (Gross *et al.*, 1984, 1993), which is an equivalent of the Podhale Flysch in Poland. The basal Borové Formation lies transgressively on the Mesozoic cover of the Malá Fatra, Tatra, and Choč mountains. The lithology of this formation is variable, being strongly dependent on the character of the substratum upon which it was deposited. It is composed of breccias, sandstones, and carbonates, sporadically with large foraminifers. The thickness of the entire formation varies from few centimetres to several tens of metres. The Szaflary beds, occurring in the northern part of the basin (Figs 40, 41), are generally assigned to the oldest flysch members (Kępińska, 1997; Chowaniec & Kępińska, 2003). Shaly flysch strata of the Zakopane beds (Fig. 42), in turn, belong to the younger members. The Slovak equivalent of the Zakopane beds is the Huty Formation, which comprises mainly pelitic, sandstone (also some breccias) strata, only few centimeters thick, in contrast to several tens of centimetres to m thick clayey beds. Localities exposing dark-brownish Menilite-like silty claystones occur at several places within the Huty Formation. The Chochołów beds (Zuberec Formation in Slovakia), overlying the Zakopane beds, is of typical flysch facies, with variable sandstone/shale ratio (Fig. 43). Fine-grained breccias and even slumped conglomerates are common. Submarine slumps are of sandy matrix with dispersed clasts of sandstones, siltstones, claystones, and limestones. The total thickness of this unit is about 500–900 m, and its age was determined as the Early Oligocene (Gedl, 1999a, 2000a) (by Slovakian geologists as Late Eocene to Early Oligocene – Gross *et al.*, 1993). The uppermost formation in the Podtatranská Group in Slovakia is the Biely Potok Formation. Its equivalent in Poland is known as the Ostrysz beds, forming the culmination of Ostrysz Mt in the western Podhale (Fig. 40). This formation consists of coarse-grained sandstones and subordinate claystone strata. The sandstones are mainly siliciclastic with clayey matrix, bearing only small percentage of carbonates. The thickness of the Biely Potok Formation is up to 700 m, and its age is Late Oligocene. At few places (in Slovakia only), the Pucov Conglomerates occur (e.g. south of the Oravský Podzámok). This member consists of blocky conglomerates bearing various Mesozoic carbonate clasts cemented with reddish sandy-pelitic matrix. Longitudinal, narrow bodies of conglomerates are incised into the Zuberec and Huty formations, and also to the Mesozoic substratum. The Pucov Conglomerate is interpreted as a channel fill supplied from the nearby southern source. The thickness of 170 m was documented by a borehole log (Gross *et al.*, 1993).

Palaeogeographical position of the Podhale Paleogene Basin is showed in global, European and Carpathian perspectives (Figs 5, 6, 18) with full geodynamic context and facies distribution during the Oligocene times. Janočko & Kováč (2003; see also references therein) suggested that the initial evolutionary stage of the basin was due to oblique convergence during the retreat of subduction boundary, which resulted in compressional regime in front of the advancing upper plate and extension in the plate's inner part. The opening of the Central Paleogene Basin was related to this extension. The front of the Inner Carpathian plate served as a source area for sedimentation of the Szaflary beds. In the presently narrow zone along the Pieniny Klippen Belt, the Paleogene strata were deformed into slices and folds (Fig. 41A–C). The wide area (15–20 km?) of the northern rim of the Inner Carpathian Paleogene is missing. It is that part, where the lateral input of the material from the source had produced proximal sediments. One has to bear in mind, however, that palaeomagnetic data point to significant (70–110°) counterclockwise rotations within the Carpathian Paleogene Basin (see, e.g. Grabowski & Nemčok, 1999; Márton, 2003; Csontos & Vörös, 2004; Golonka *et al*., 2005a). According to Golonka *et al.* (2005a), an analysis of exotic clasts supports this rotation, indicating that neither the present-day Tatricum nor the Sub-Tatric (Križna and Choč) nappes were source areas for the Pieniny and Magura flysch during Paleogene time. This question requires new independent

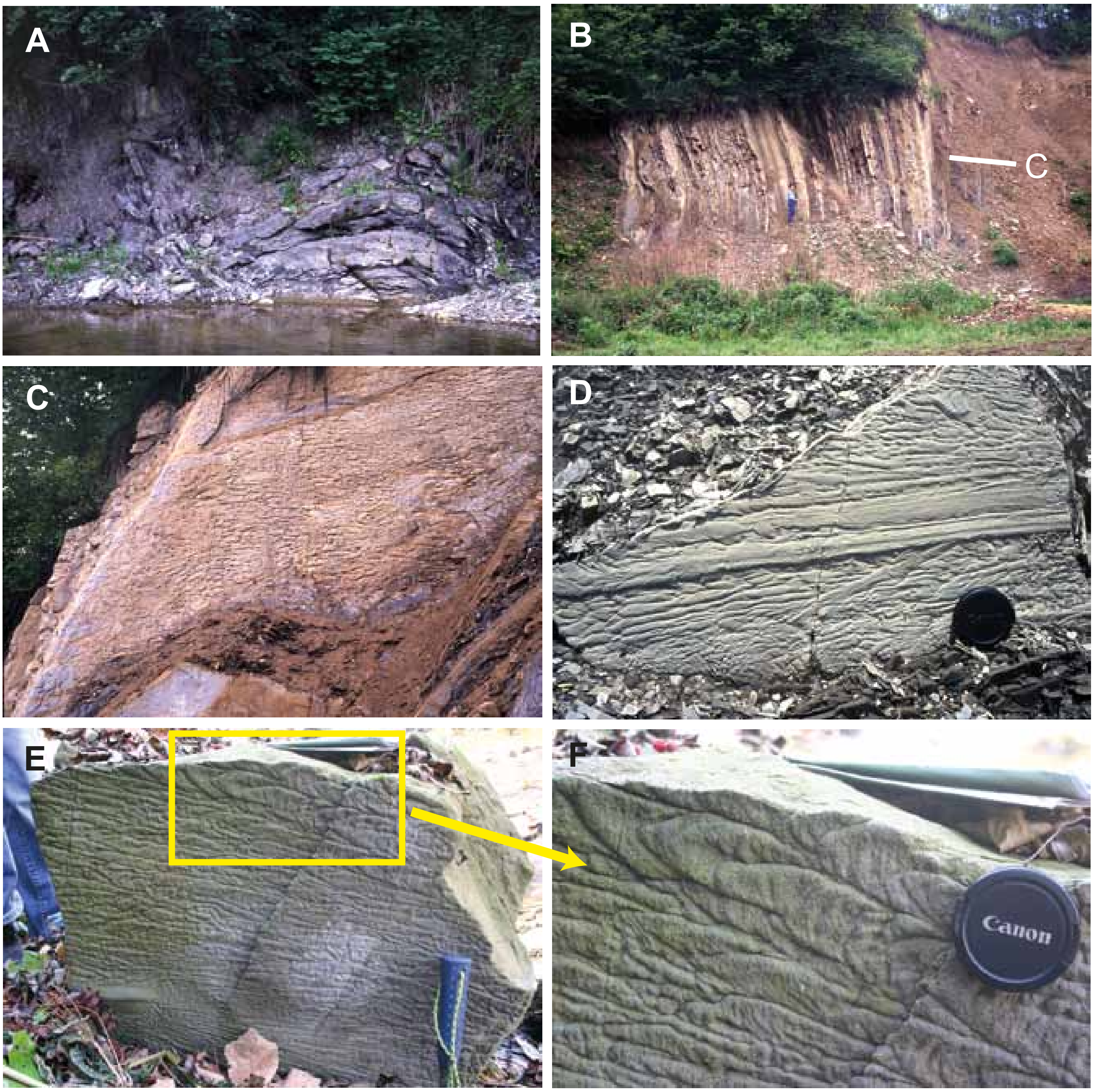


Fig. 41. Strongly folded Szaflary Beds (**A**) or in vertical position (**B**) in contact zone with Pieniny Klippen Belt [Niedzica and Slovakia respectively] with conglomerates/coarse grained sandstones on the sole surface of beds (C) and perfect preserve mechanical hieroglyphs: groove casts (**D**) and moulds of scales arranged in flow-parallel rows and longitudinal ridges (**E**, **F**) (comp. Dżułyński, 1996, 2001) (**D–F**: **D** – Słowacja; **E**, **F** – Leśnica stream (after Krobicki & Golonka, 2008b)

research, but nevertheless it calls for a substantial correction in estimations of the genuine palaeogeographic location of alimentary areas. Perhaps the missing part of the Central Paleogene Basin is located somewhere within the Tisza plate. The Transylvanian Paleogene (e.g., Săndulescu *et al.*, 1981; Ciulavu & Bertotti; 1994; Meszaros, 1996) and certain parts of the Szolnok flysch Paleogene sequences, situated in the marginal part of the Tisza unit (e.g. Nagymarosi & Báldi-Beke, 1993), display similarities with the Central Carpathian Paleogene. This problem requires further investigations. In the Neogene, the Inner Carpathian plate rotation wiper effect led to significant deformation along the plate boundary, which resulted in a complex tectonic pattern along the present-day boundary between the Central Carpathian Paleogene and the Pieniny Klippen Belt. The present-day Podhale Basin is an asymmetric basin, delimited by the Tatras in the south, and by a steep fault along the Pieniny Klippen Belt in the north. According to Soták & Janočko (2001), the structural pattern of the Central Carpathian Paleogene Basin includes basement-involving fault zones, like the Margecany and Muran faults. Extensional features, like half-grabens and listric and antithetic faults are to be found in the Hornad, Periklippen and Poprad Depressions, while structures re-

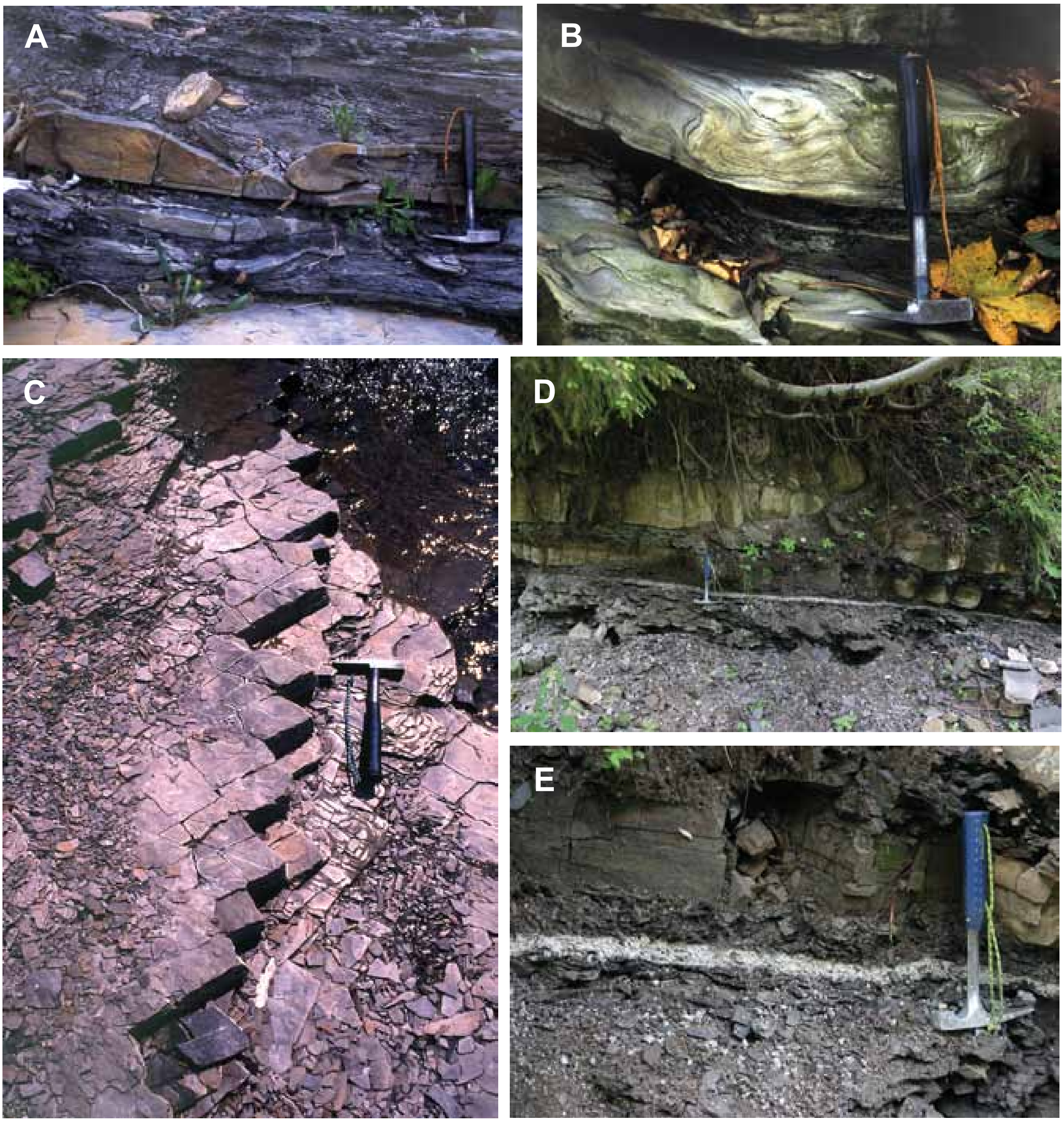


Fig. 42. Thin-bedded Zakopane Beds with slump structures (**A**) and convolutions (**B**) and good visible joints (**C**) (Murzasichle village), as well as thin tuffitic intercalations in grey shales (**D**, **E**) (Małe Ciche) (after Krobicki & Golonka, 2008b)

lated to retro-wedge thrusting, transform faulting, and strike-slip tectonics occur in the Šambron Zone (Mastella *et al.*, 1988; Kovač & Hók, 1993; Ratschbacher *et al.*, 1993; Nemčok & Nemčok, 1994; Marko, 1996; Sperner, 1996, 2002; Plašienka *et al.*, 1997; Soták & Janočko, 2001; Janočko & Kováč, 2003).

The above-mentioned counterclockwise rotation of the Alcapa plate was compensated by dextral shearing in a transpressional zone between the Alcapa and North European plates (Ratschbacher *et al.*, 1993; Soták & Janočko, 2001). The present-day northern boundary was caused by amputation by a transform fault related to this rotation.

In the Polish part of the Carpathian Paleogene has been divided the following informal lithostratigraphical units, which correspond to mentioned above Slovakian units (comp. Gedl, 1999a, 2000a) (Fig. 40):

– "**Tatra Eocene**" (Borove Formation in Slovakia);
– **Szaflary beds** (hasn't good equivalent in Slovakian Paleogene) but Soták & Janočko (2001) and Soták *et al*. (2001) suggested correlating them to Šambron Beds (eastern part of the Central Carpathian Paleogene Basin – to E from Vyšné Ružbachy). Szaflary beds are developed only in northernmost part of the Podhale Flysch, close to the Pieniny Klippen Belt, and therefore

usually are strongly folded up to vertical position (Fig. 41). Thin-bedded sandstones of the Szaflary beds consists normal flysch-type sedimentary structures such as: groove casts (Fig. 41D) moulds of scales arranged in flow-parallel rows and longitudinal ridges (Fig. 41E, F);

- **Zakopane beds** (Huty Formation), dominated by dark grey shales and thin layers of sandstones (sometimes with synsedimentary slump structures of submarine mass movements – Fig. 42A) and convolution (Fig. 42B), relatively rich in thin intercalations of tuffites (Fig. 42D, E). In this unit are well visible joints (Fig. 42C) which indicate tectonic stress plane;
- **Chochołów beds** (Zuberec Formation) with full spectrum of sedimentological (i.a., flut casts, parallel(tractional)-lamination in sandstone) (Fig. 43A–D) and palaeontological (carbonized piece of wood and/or trace fossils) (Fig. 43E–G) features, typical for turbiditic systems;
- **Ostrysz beds** (Biely Potok Formation) is the youngest unit of the Polish Paleogene Flysch deposits, and occur locally in westernmost part of the Podhale region (Fig. 40).

According to biostratigraphical data all flysch-type units are Oligocene in age (Gedl, 1999a, 2000a).

Several branches of geological sciences could be study in Podhale Flysch region:

Sedimentological features of the Podhale Paleogene are typical for submarine turbiditic processes which produced fan-shape lobes and/or aprons (Grzybek & Halicki, 1958; Radomski, 1958; Marschalko & Radomski, 1960; Picha, 1964; Koszarski & Sikora, 1971; Krysiak, 1976; Krawczyk, 1980; Westfalewicz-Mogilska, 1986; Marschalko, 1987; Wieczorek, 1989; Soták, 1998; Soták & Janočko, 2001; Soták *et al*., 2001 with literature cited therein) and very well correspond to classical “flysch” literature (e.g., Mutti & Ricci Lucchi, 1975; Słomka, 1986; Stow, 1986; Mutti & Normark, 1987; Ghibaudo, 1992; Reading & Richards, 1994; Lowe, 1997; Shanmugam, 2000). Some sedimentological problems are the objects of hot disscusions and controversion (see – Gross & Köhler, 1993; Wieczorek, 1989, 1995).

Biostratigraphy has long history, since thirties-fifties years of last century up to recent, and were based on different type of microfossils [foraminifers (both nummulitic and planktonic forms), calcareous nannoplankton, and dinocysts] (Bieda, 1959; Blaicher, 1973; Dudziak, 1983, 1986, 1993; Gedl, 1995, 1999a, 2000a, b, 2004a; Olszewska & Wieczorek, 1998). Correlation between different biostratigraphical schemes was recently done by Gedl (2000a) who suggested Oligocene age for all flysch units. However, Garecka (2005) on the basis of calcareous nannoplankton, proposed for youngest Ostrysz beds earliest Miocene age.

Petrography and ***mineralogy*** (including analysis of ***tuffitic*** layers). These primary researches were objects of interest several generations of geologists. The first focus was directed into the petrographical character of flysch rocks, their mineral composition and fraction, and distribution main and accessory minerals (e.g. Kosiorek-Jaczynowska, 1959; Bromowicz & Rowiński, 1965; Wieser, 1973; Roniewicz & Westfalewicz-Mogilska, 1974 and references cited therein). More intensive have been studied exotic-bearing deposits and exotics from them, both crystalline rocks and sedimentary one, mainly carbonates (Chowaniec & Golonka, 1980; Chowaniec & Wieser, 1981; Burtan *et al*., 1984; Chowaniec, 1985). More recently investigations of tuffitic rocks have been concentrated on their mineral character and geochemical meaning (Głazek *et al*., 1998). About 30 tuffitic horizons have been recognized in the Oligocene Podhale flysch deposits, which indicate intensive volcanic activity within Carpathian region (Michalik & Wieser, 1959). But the source area of such piroclastic material is still enigmatic (Sobień, 2005) and maybe connected with Inner Carpathian volcanism which originated by activity of subduction processes during geodynamic events in this part of Alpine-Carpathian orogenic system.

Structural geology indicates that the reorientation of orogenic system (arc) and some rotation of isolated parts of mountains are most important to understand the movement events in post-orogenic tectonic activities. On the end of such investigations are neotectonic researches, which based on very young tectonic processes (Halicki, 1963; Pokorski, 1965; Morawski, 1972, 1973; Pepol, 1972; Mastella, 1975; Mastella & Ozimkowski, 1979; Mastella *et al*., 1988).

Organic geochemistry – the Central Carpathian Paleogene Flysch (including Podhale Flysch) not belongs to high potential source of oil area, in spite of the region is well recognized for accumulation of such type of energy. The basin-fill sequence has poor (TOC – Total Organic Carbon – ≤0.5%) to fair (TOC <1.0%) quality of source rocks (e.g., Soták *et al*., 2001 with literature cited therein; see also – Watycha, 1968; Mastella & Koisar, 1975).

In the Podhale Flysch we can see useful ***palaeontological*** and ***palaeoecological*** data record, enabling step by step a detail reconstruction of origin and evolution of Podhale Basin as only small part of the whole Central Carpathians Paleogene basins system. We can mainly use, as mentioned earlier, microfossils to determine palaeoenvironmental regimes in sedimentary basins as follow: bathymetry, palaeosalinity, bottom conditions, oxygen content both on the sea-floor and in water column, palaeocurrents, palaeoclimate, sedimentation rate, quantity and quality of nutrients etc. (e.g., Roniewicz & Pieńkowski, 1977; Pieńkowski & Westfalewicz-Mogilska, 1986; Gedl, 1999b, 2000a).

Palaeogeographical reconstruction and facies context of the Paleogene flysch deposits originated in the Central Carpathian Paleogene Basin is strictly connected with geodynamic regimes of formation of the Alpine-Carpathians

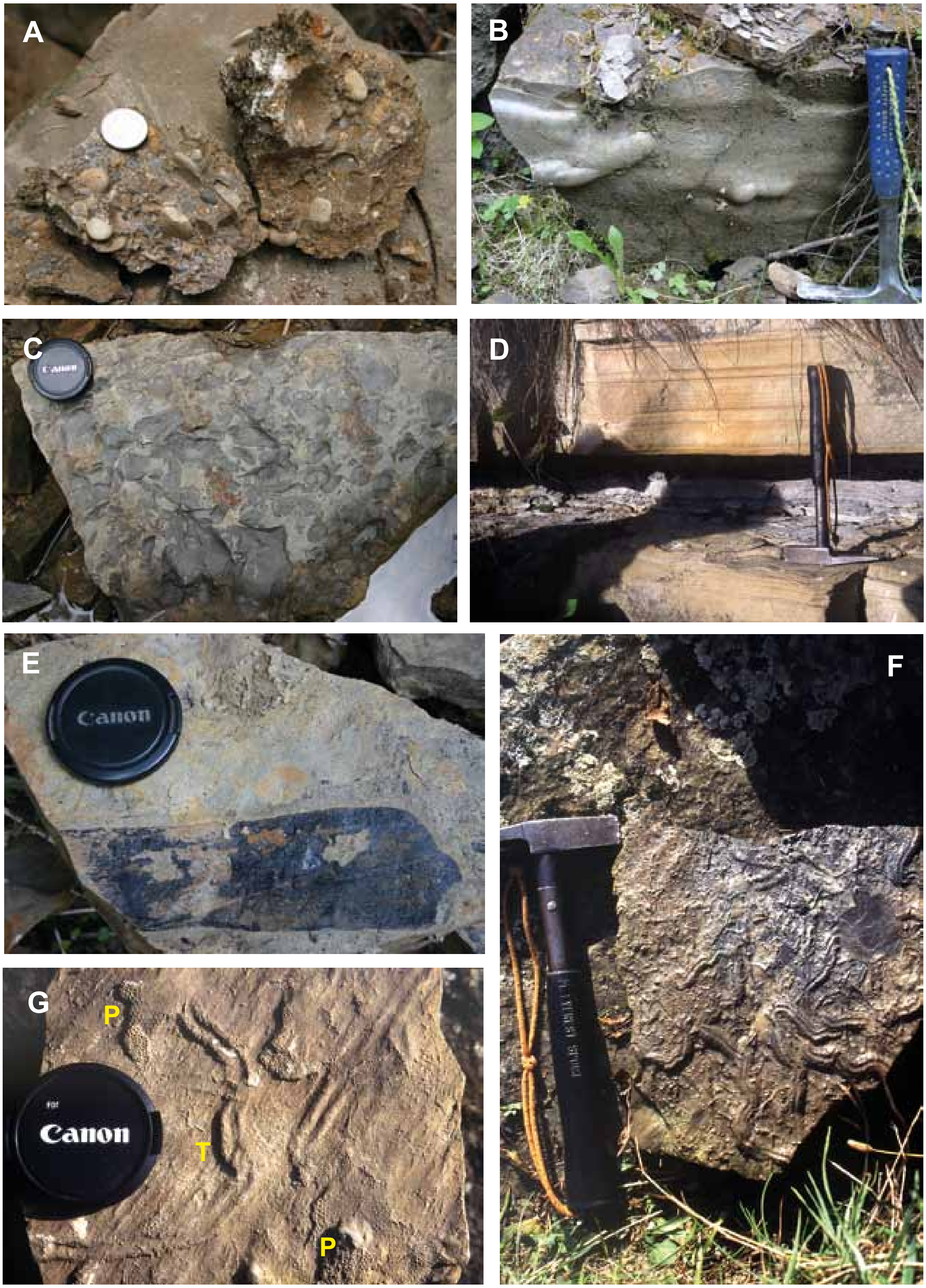


Fig. 43. Sedimentological and palaeontological features of the Chochołów Beds: **A** – conglomerates with large pebbles of the base of thick bed with gradational fractionation (Słodyczków stream); **B** – large flut casts on the base surface of fine-grained sandstone bed (Słodyczków stream); **C** – big shale clasts in lowermost part of fine-grained sandstone bed (Słodyczków stream); **D** – parallel (tractional)-lamination in sandstone (Habovka; Slovakia); **E** – large piece of wood (carbonized) (Słodyczków stream); **F**, **G** – trace fossils (*Taphrhelminthopsis* – T and *Paleodictyon* – **P**) on the base of thin-bedded sandstone (Slovakia) (after Krobicki & Golonka, 2008b)

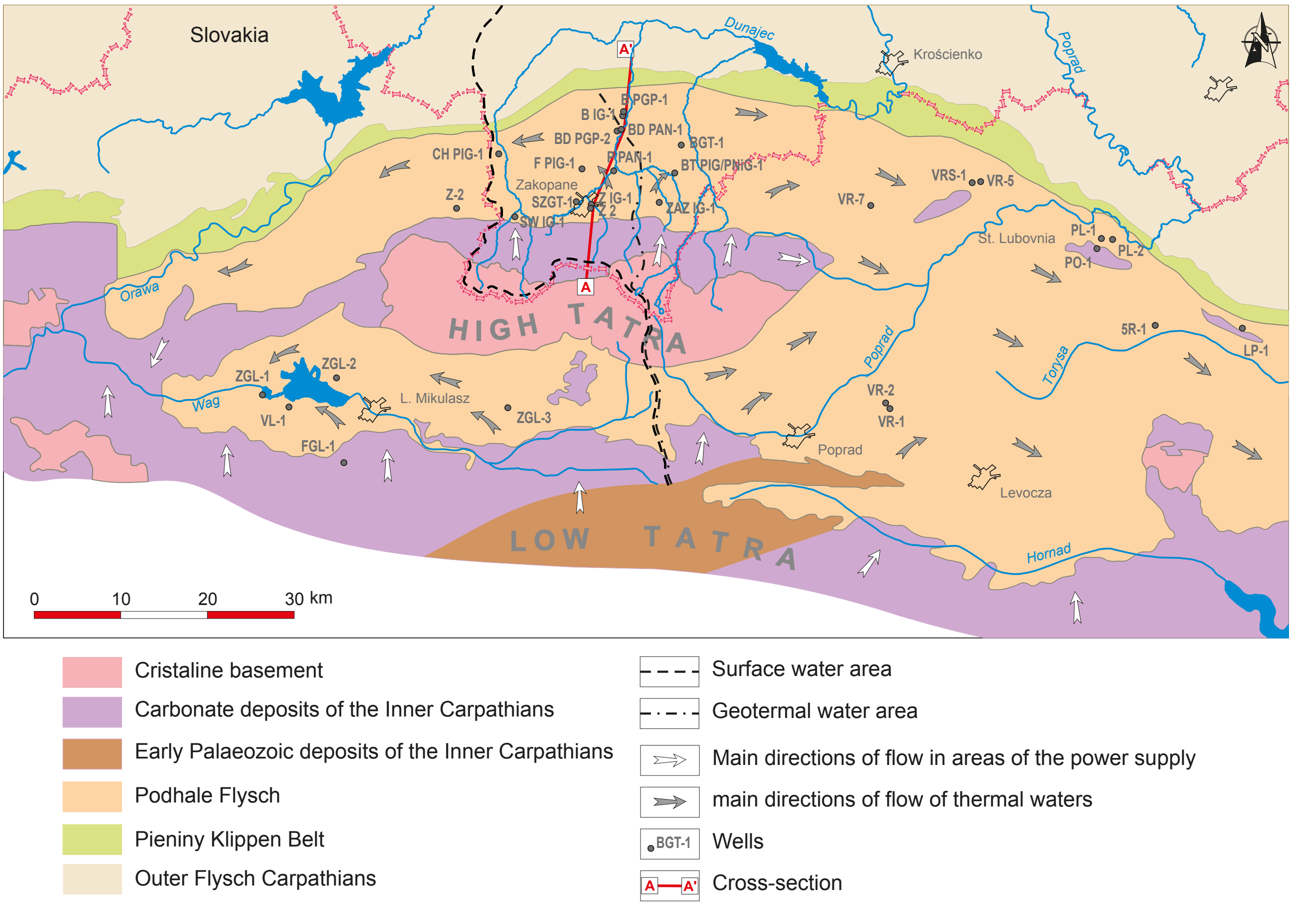


Fig. 44. Geologic map of the Inner Carpathians with the lines of the hydrogeologic cross-section (see – Fig. 45) (after Szklarczyk in: Chowaniec, 2009, modified)

arc in the light of Eurasian and Africa movements and collisions of smaller terranes (i.a., Marschalko, 1981; Tari *et al.*, 1993; Golonka, 2004; Golonka *et al.*, 2000, 2006; Soták & Janočko, 2001; Soták *et al.*, 2001 with references cited therein) (Figs 8–10). In local, regional perspective development of the Podhale Flysch Basin and its filling by thick flysch sequence have been connected with strong subsidence tendency (especially during sedimentation of fine- and/or coarse-grained sandstones of Szaflary, Chochołów and Ostrysz beds). This subsidence was close correlate with strong erosion of uplifted source areas, which produced very large bulk of fresh sediments, which have been moved by turbiditic currents, debris flows and other gravitational mass movements down to deep basin floor, and produced lobes and turbiditic fans interrupted by catastrophic sedimentation of chaotic conglomerates (e.g., Pucov Formation) (Radomski, 1956, 1958; Marschalko, 1981, 1987; Westfalewicz-Mogilska, 1986; Wieczorek, 1989; Soták & Janočko, 2001 with literature therein). Reconstruction of location of source areas is one of the most interesting problems in such flysch filling region and is mainly based on orientation of palaeocurrents interpreted by measurements of mechanical hieroglyphs (mainly flutcasts, groove casts, flow-parallel rows and longitudinal ridges etc.) (Figs 41D–F, 43B). According to such investigation, during the Late Oligocene the Central Carpathian Paleogene Basin was filled up by sand-rich fans, for example. In Podhale region the flysch sequences form large elongate fan with generalized direction of sediment dispersal based on palaeocurrent orientation and proceeding from W and SW to E and NE (based on data from Dżułyński & Radomski, 1955; Radomski, 1958, 1959; Marschalko & Radomski, 1960; Krysiak, 1976; partly Westfalewicz-Mogilska, 1986) (but earlier, during the Early Oligocene time, northern source area has been active – see Krysiak, 1976; Gedl, 2000a). Such palaeocurrent system exhibit a downfan distribution of submarine fan facies indicated by directions of hieroglyphs, distribution of facies zones and evolution of facies tracts. Youngest deposits known recently from Podhale region are Upper Oligocene Ostrysz beds (Gedl, 2000a) (after Garecka, 2005 – even up to lowermost Miocene), but most probably sedimentation in this basin continued later (during Early–Middle? Miocene times), but younger deposits are limited today (Cieszkowski, 1992, 1995). In fact, we don't know when precisely the sedimentation has finished {?Middle Miocene [Middle Sarmatian (=Serravalian)]?} (Cieszkowski, 1995). The burial, thermal history of this region is recorded in change of clay minerals (identification of maximum palaeotemperatures by measuring the ratio of smectite to illite in mixed-layer clay minerals separated from shales) and indicates deep burial episode just after sedimentation of Ostrysz beds (Late Oligocene) (Środoń *et al*., 2006; Środoń, 2007). On the other hand, the orogenic uplift and exhumation effects and moving up of the Podhale Flysch region could be measured by so-called apatite fission track analyses. This study indicates that uplift rates were rather slow and variable during Miocene time and not to exceed 0.4 mm/year (Anczkiewicz *et al*., 2005; Środoń *et al*., 2005). Based on such results, the calculations about the erosion and removal of huge thickness of enigmatic deposits [even more than 7 km! (between 5.5 and 6.9 km in eastern part of the Podhale Basin) – twice bigger than general thickness known today] younger than Late Oligocene (earliest Miocene) (Środoń *et al*., 2005, 2006; Środoń, 2007) are surprisingly and high controversial (comp. Poprawa & Marynowski, 2005). These authors suggested that calculated apparent thickness of removed deposits is significantly overestimated with compare to any realistic range of its value. On the other hand, Hurai and his co-authors (2006), using the fluid inclusions for recalculation of thermal and pressure history of burial process, suggested also similar value for fore-arc Podhale basin (5.4–6.6 km eroded rock column). For Outer Flysch Carpathians (Dukla Unit), as typical accretionary wedge, they proposed even about 11 km of rocks removed from recent erosion surface.

Geophysical studies in Carpathians were focuses mainly on the Outer Flysch Carpathians and only few surveys concentrated on the Podhale Flysch region. Deep soundings of magnetotelluric survey describe the deep-rooted lithospheric blocks and give general view of deep basement of Carpathians (Czerwiński & Stefaniuk, 2001; Czerwiński *et al*., 2003; Golonka *et al*., 2005a). Another geophysical works are concentrated on palaeomagnetic researches, which for example can explain post-orogenic rotation of regions and/or their parts. In Podhale case, results of such investigations indicate counterclockwise movements of the Podhale block along the peri-Pieniny strike-slip faults (Grabowski & Nemčok, 1999; Márton *et al*., 1999).

Geothermal energy (see below as well) is type of "clean energy" and is very useful and that's why the Podhale region is more and more attractive. All waters below the zones of active flow of meteoric waters have elevated temperatures and by definition are regarded as thermal. However only Mesozoic and Paleogene formations of the Podhale Basin contain renewable thermal waters, which are recharged in the outcrop area in the Tatras at the elevations of above *ca.* 1000 m a.s.l. (Figs 44, 45). They flow to the north under thick flysch cover up to the obstacle formed by the Pieniny Klippen Belt where they are divided into two back flows to the south, with the natural discharge in the Danube watershed in Slovakia (Fig. 45). The hydrogeologic data of wells with thermal waters are given in Table 1. The highest temperature measured at well heads is 86°C, the highest TDS content is *ca.* 3 g/dm^3, and the highest yield is *ca.* 550 m^3/h (Fig. 46). Thermal waters in the Outer Polish Carpathians are known from a number of deep wells. They represent buried waters of high TDS contents and moderate temperatures of 27 to 42°C. Their exploitation is generally impossible due to the pollution endanger of surface waters. Only mineral and thermal water in Ustroń is exploited for therapeutical purposes as the used water is removed back to Devonian formation by special injection well (Fig. 46). Some thermal wells situated near the recharge area contain measurable tritium contents (Chowaniec *et al*., 2009). In some boreholes the terrestrial heat flow values have been determined at 55.6 and 60.2 mW/m^2 (Zakopane IG-1 and Bańska IG-1 wells respectively – see Fig. 3B) (Chowaniec & Kępińska, 2003; see also – Jaromin *et al*., 1992a, b). The temperatures at depths of 2–3 km amount to over 80–90°C, which is markedly higher than those resulting from the normal geothermal gradient (between 1.9 and 2.3°C per 100 m). Kępińska (1997) suggests that increased upflow of heat and/or hotter fluids from the deeper part of the system occur along discontinuity planes. Very sharp contact of the Podhale Flysch zone with the Pieniny Klippen Belt at northern side is presumable a key to understanding such phenomena. The thermal anomalies recorded on the tectonic contact between the Podhale Flysch Basin and the Pieniny Klippen Belt is about 2–3°C higher than average background values (Pomianowski, 1988). The main collectors of geothermal water is both nummulitic Paleogene and Mesozoic basement mainly built by Triassic sedimentary rocks and belong to different tectonic Tatric units (Wieczorek & Barbacki, 1997; Chowaniec & Kępińska, 2003). These are the main reasons of development geothermal spring/spa in this region, including Bukowina Tatrzańska village with Termy BUKOVINA (Fig. 47).

Quaternary geology *(including Late Neogene)* and ***geomorphology*** is the youngest history of the Podhale Basin (e.g., Klimaszewski, 1952; Zuchiewicz, 1998) which concern both the Neogene Orava Basin (Baumgart-Kotarba, 1992, 1996, 2001; Baumgart-Kotarba *et al*., 2001) (Figs 4, 40) filled mainly by freshwater clastic deposits, and central part of Podhale (Birkenmajer, 1978). Palaeoclimatic reconstruction of this region are based on palynological studies (in Pliocene time – Oszast, 1973) and malacological one (Holocene – Alexandrowicz W.P., 1997). The malacological assemblages are mainly connected with calcareous tufa (Alexandrowicz, W.P., 1997, 2001) and therefore, this Late Glacial and Holocene deposits and travertines are most important to documentation of the youngest geological history of the Podhale region, changes of climate and palaeoenvironments. Some molluscan assemblages (mainly land gastropods) which abundantly occur in such type of deposits are very useful for documentation of

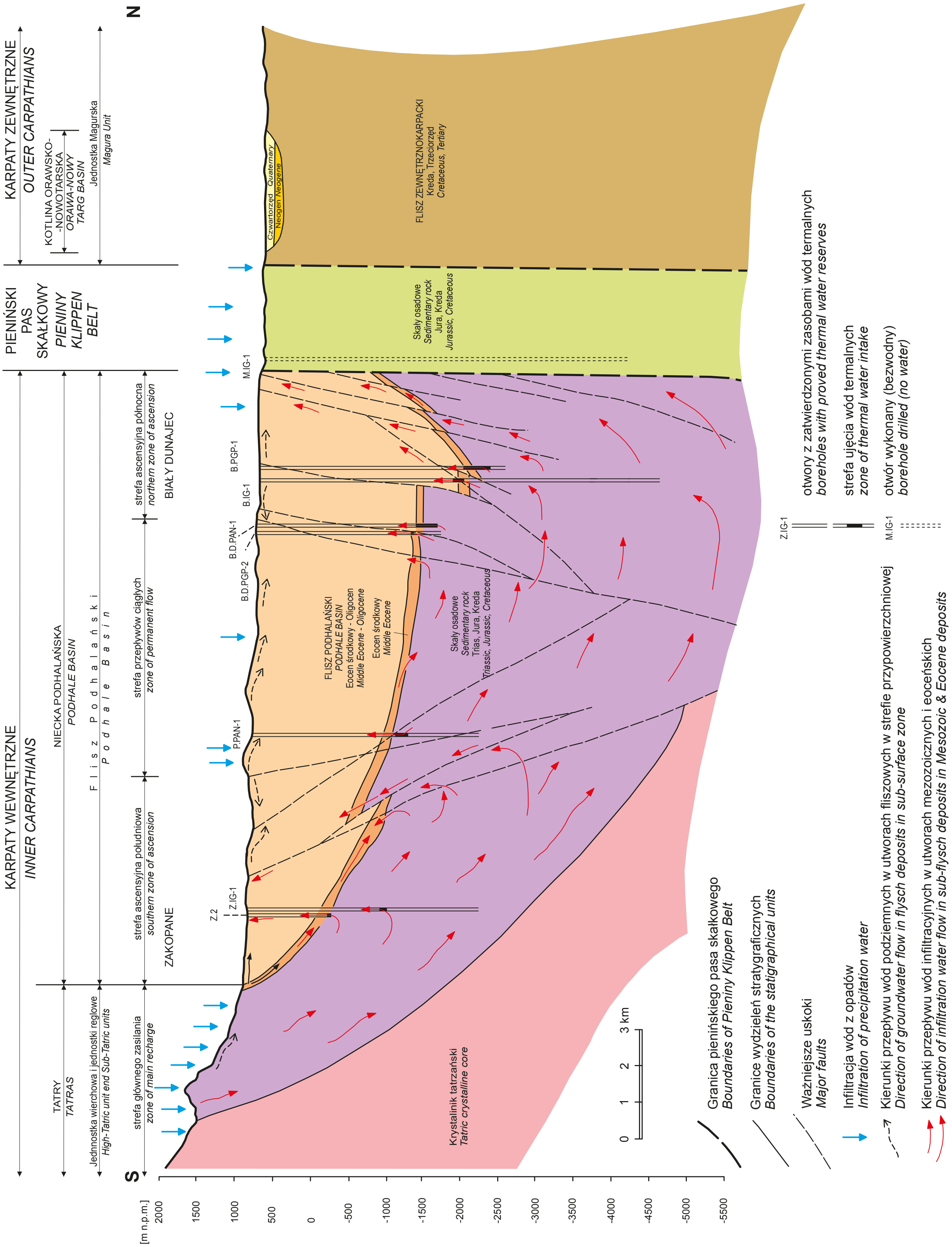


Fig. 45. Geological cross-section of the Podhale Basin with schematic flow lines (after Chowaniec, 2009) (cross-section line is shown in Fig. 44)

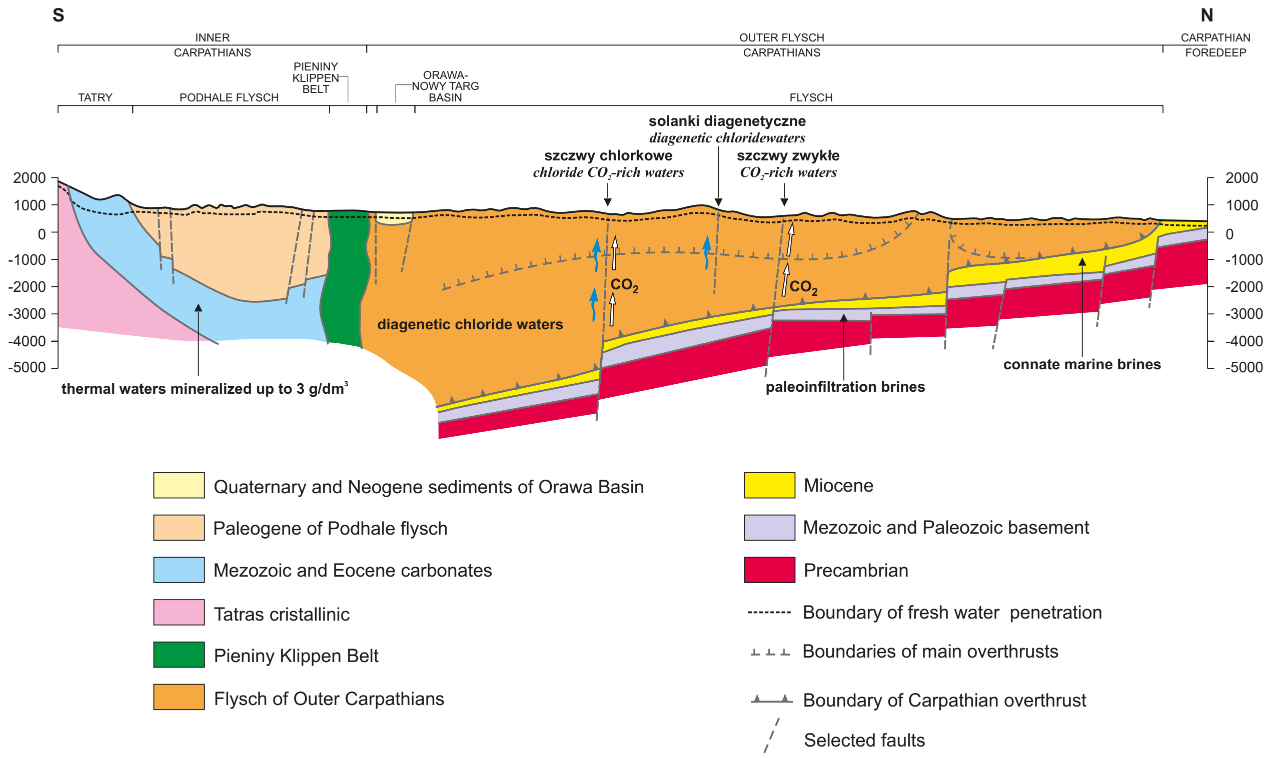


Fig. 46. Simplified geological cross-section through the western part of the study area with schematically shown waters of different origin

continuous changes during the last 11 000 years (Alexandrowicz, W.P., 1997, 2003). On the other hand, a lot of localities of peatbogs in the Orava-Nowy Targ area can help to understand some Quaternary vegetation/climate history of this region (Wagner, 2005; Ptaszek, 2008). Usually, the thickness of these peatbogs ranges 1.1–7.1 metres, and in this area occur 26 peatbogs (Wagner, 2005). Another investigations of such young (in geological sense of course) period, up to recent, are based on neotectonic movements in Podhale (Zuchiewicz, 1998), aerial photographs interpretation and landform analysis (Ozimkowski, 1987, 1991) and also interpretations of landslides (Bober, 1971; Mastella, 1976; Kukulak, 1988). Podhale is also a region of earthquakes reaching the magnitude 5.1 in the Richter scale and causing damages in the nearby villages (Chrustek & Golonka, 2005). These earthquakes are related to the modern fault's activity.

Nature protection (geological heritage). In the Podhale region only 2 localities are formally protected by Polish law as nature monuments: spherosiderites from Zakopane beds in Chłabowski stream near Zakopane and waterfall (7 m high) formed on horizontal layers of Zakopane beds in Kacwin village (eastern Podhale) (Chowaniec & Golonka, 1981; Alexandrowicz, Z., 1997, 2000). Six additional outcrops are planned to protect as documentary sites of inanimate nature: Dzianisz (synsedimentary slump structures), Bustryk (the same as above), Stasikówka (perfect outcrop of flysch sequence), Kotelnica near Zakopane (slump structures), Karpenciny (same as above) and Gliczarów (calcareous tufa) (Alexandrowicz, Z., 2000; Alexandrowicz, W.P., 2003, 2004). Several others geological and geomorphological outcrops/landforms in the Podhale Flysch need to protection in law as good both educational and scientific objects.

Fig. 47. General view of the Termy BUKOVINA resort

DUNAJEC RIVER RAFTING (MESOZOIC HISTORY OF THE PIENINY KLIPPEN BELT)

› FIGS 48–63

The Pieniny Klippen Belt (PKB) is situated at the boundary between Outer Flysch Carpathians and Inner Carpathians forming strongly tectonized terrain about 600 km long and 1–20 km wide, which stretches from Vienna to the West up to Romania to the East (Fig. 2). Present day confines of the PKB are strictly tectonic. They may be characterized as (sub)vertical faults and shear zones (Fig. 3), along which a strong reduction of space of the original sedimentary basins took place (Birkenmajer, 1986; Golonka & Krobicki, 2006; Krobicki & Golonka, 2006). The PKB tectonic components of different age, strike-slip, thrust as well as toe-thrusts and olistostromes were mixed together, giving the present-day melange character of the PKB, where individual tectonic units are hard to distinguish.

GENERAL HISTORY OF THE PIENINY KLIPPEN BELT

Michał KROBICKI & Jan GOLONKA

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, b 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) (Figs 31, 48, 53).

The **earliest stage** of the basin history is enigmatic and documented only by exotic pebbles in the Cretaceous-Tertiary flysch. The pelagic Triassic limestones in the exotic pebbles in the Pieniny Klippen Belt (Birkenmajer *et al.*, 1990) and the Outer Carpathian Flysch (Magura Unit, see Soták, 1986) could have originated in this oldest stage of basin. These pebbles indicate the possibility of an existence of enigmatic embayment of the Vardar–Transilvanian Ocean which separated the Tisa (Bihor–Apuseni) block from the Moesian Eastern European Platform (Săndulescu 1988; Golonka & Krobicki, 2004). The other interpretation of these pebbles origin involves the rotation of the Inner Carpathian plate (Golonka, 2005).

The **oldest Jurassic** rocks known only from the Ukrainian and Slovakian part of the Pieniny Klippen Belt (Krobicki *et al.*, 2003; Schlögl *et al.*, 2004) 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), which we will see in Homole Gorge, as well as dark marls and spotty limestones of widespread Tethyan *Fleckenkalk*/*Fleckenmergel* facies, indicate the oxygen-depleted conditions (Birkenmajer, 1986; Tyszka, 1994, 2001). 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).

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

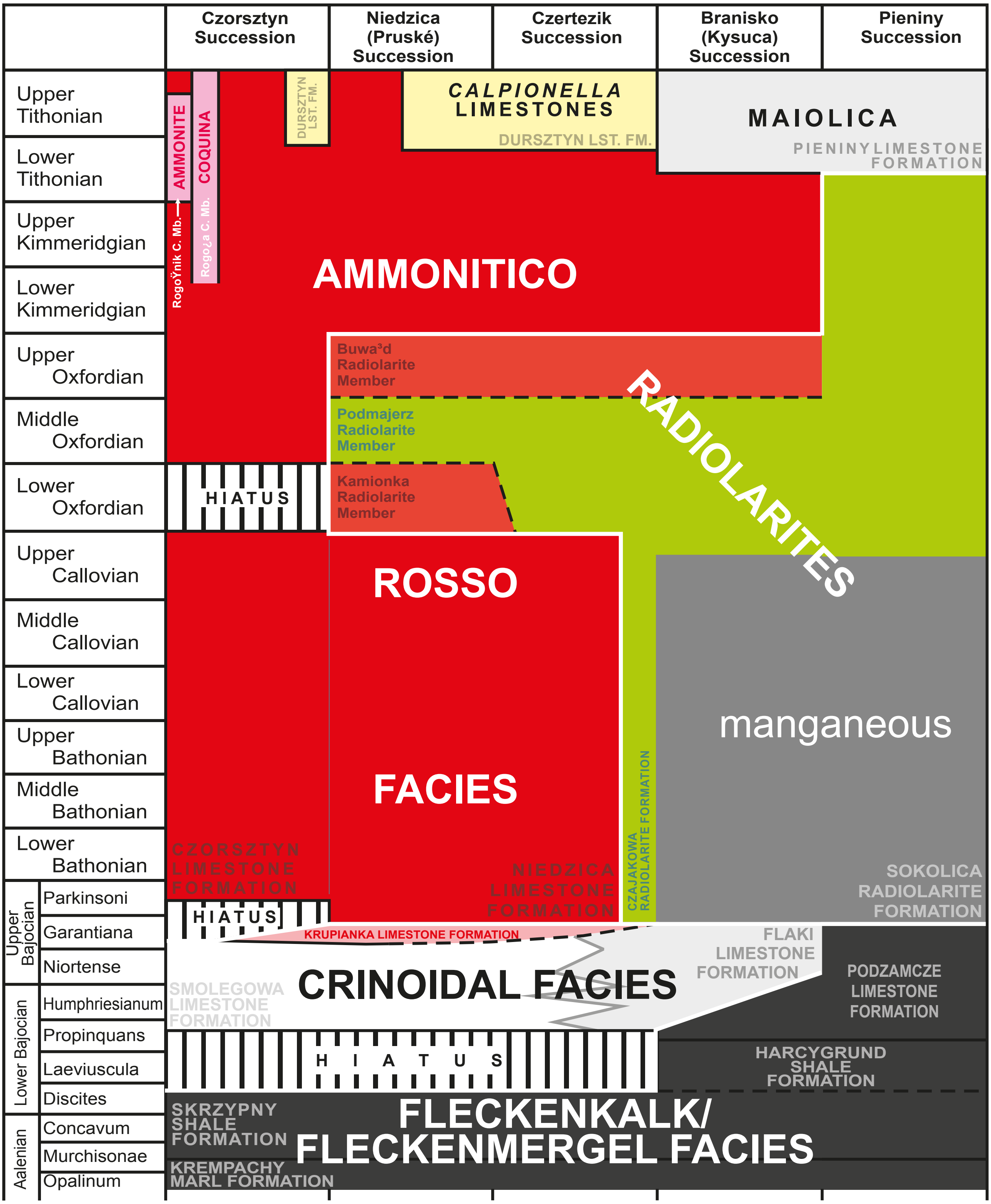


Fig. 48. Stratigraphical correlation between Jurassic lithofacies (lithostratigraphic units) of the Pieniny Klippen Belt successions (after Wierzbowski *et al*., 2004; supplemented by Krobicki & Wierzbowski, 2004)

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 involving, among the others, mixed siliceous-carbonate sedimentation took place later, mainly during Oxfordian times. 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).

The Czorsztyn Succession during **latest Jurassic–Early Cretaceous** time (Tithonian–Berriasian) includes 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, 1996). 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, 1996; Aubrecht *et al*., 1997; Krobicki & Słomka, 1999; Golonka & Krobicki, 2002; Plašienka, 2002; Golonka *et al*., 2003). This tectonic activity was caused by formation and destruction of submarine tectonic horsts in the Carpathian basins (Birkenmajer, 1958, 1975, 1986; Michalík & Reháková, 1995; Krobicki, 1996; Krobicki & Słomka, 1999; Golonka *et al*., 2003; Krobicki *et al*., 2006). Additionally, the palaeogeographical and palaeoclimatological evidences (*e.g.*, phosphate deposits) suggest upwelling zone in northern Tethys (Birkenmajer, 1986, 1988; Krobicki, 1996; Golonka & Krobicki, 2001). 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). These white-grey, micrite well bedded calpionellid-bearing limestones built now highest part of the Pieniny Mts (*e.g.*, Trzy Korony Mt, Sokolica Mt etc).

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 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, 1983, 1984, 1987a,b; Gasiński, 1991; Birkenmajer & Gasiński, 1992; Bąk, K., 1998; Bąk, M., 1999). During this syn-orogenic stage of the development of the Pieniny Klippen Belt Basin these flyschoidal deposits developed as submarine turbiditic wedges, fans and canyon fills (Radwański, 1978; Birkenmajer, 1986) with several episodes of debris flows with numerous exotic pebbles took place (Late Albian–Early Campanian).

The Pieniny Klippen Basin was closed at the **Cretaceous/Tertiary** 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 Pieniny Klippen Belt is geomorphologic view of tectonically isolated klippes of Jurassic and Cretaceous hard rocks surrounding by softer shales, marls and flysch deposits. Process of megabrecciation and megaboudinage is clear from regional point of view where Pieniny Klippen Belt is very narrow tectonic structure between

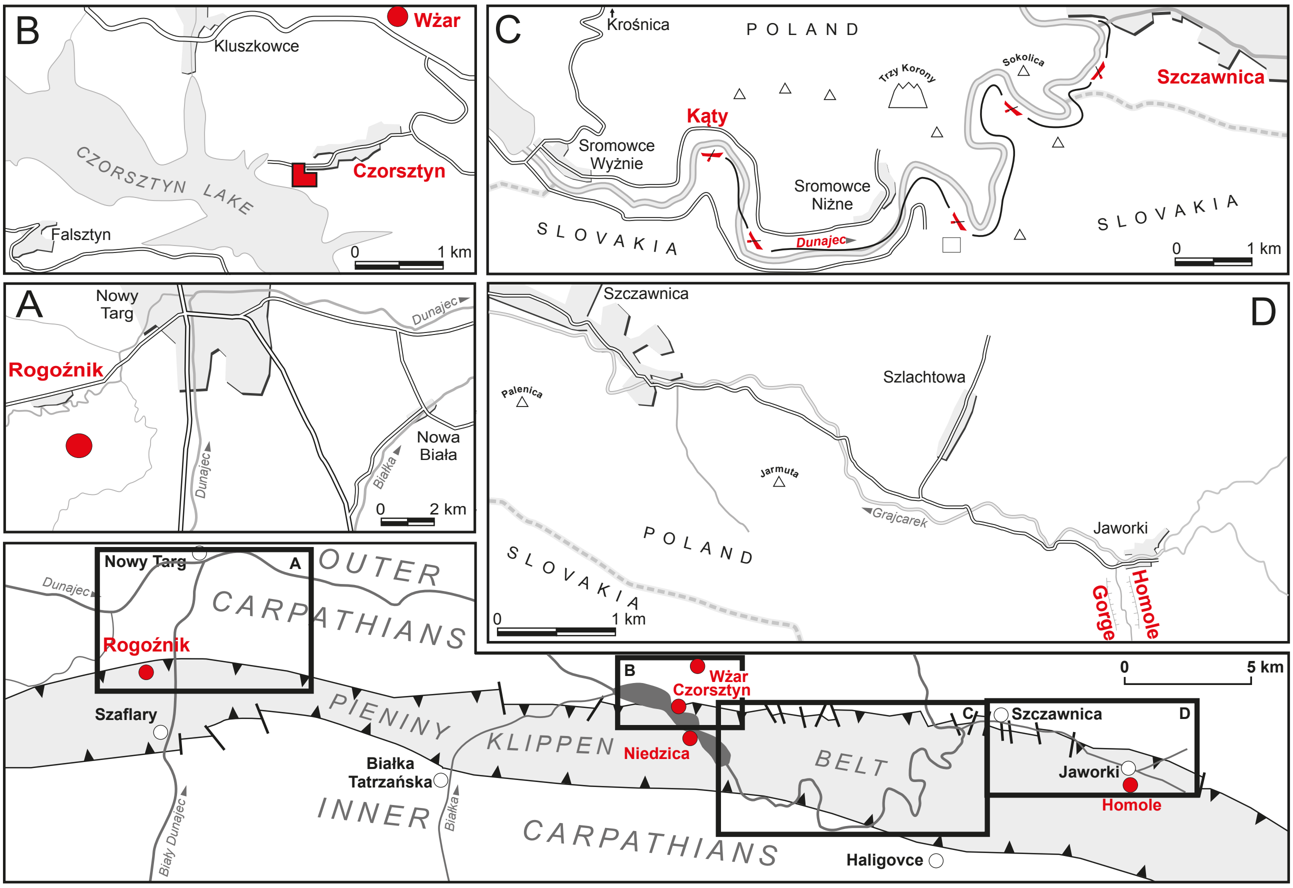


Fig. 49. Polish part of the Pieniny Klippen Belt and locations of visited outcrops – stop points

two huge part of Carpathians: Inner on the south and Outer (Flysch) on the north. The Pieniny Klippen Belt was formed as the melange in the suture zone between Inner Carpathian-Alpine (Alcapa) terrane and the North European plate. Part of allochthonous Outer Carpathian units and perhaps fragments of basement were also located in this suture zone. Finally, with the eastward movement of the Alcapa plate, system of strike-slip faults originated (Birkenmajer, 1983).

The last important event in the Pieniny Klippen Belt 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) (see Wżar Mt stop). They formed so-called Pieniny Andesitic Line (PAL).

RAFTING TROUGH DUNAJEC RIVER GORGE

Jan GOLONKA & Michał KROBICKI

One of the major attraction of the Pieniny Klippen Belt region is the rafting through the Dunajec River Gorge (Golonka & Krobicki, 2007, see also Alexandrowicz & Alexandrowicz, 2004) (Figs 49C, 50). 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.

The Kąty's harbor has good connection with main touristic points in the Pieniny Mts and is easily accessible both from Zakopane (from south) and from Kraków (from north). Kąty are located very close to famous medieval castles (Czorsztyn Castle – Fig. 59 and Niedzica Castle), which occupied good accessible and visible places just above artificial Czorsztyn Lake.

First viewpoint is the **Macelowa Mt** (Fig. 50A) there strongly folded Jurassic–Cretaceous deposits are visible on the southern slope of Macelowa Mt, where the Pieniny Succession is completely overturned. The oldest Oxfordian radiolarites occupy topmost part of the Macelowa Mt, grey cherty limestones of the *Maiolica* facies (Pieniny Limestone Formation) occupy topmost and middle part of this mountain in completely overturned position). Therefore, the youngest part of this succession occurs in lowest (topographically) position and is represented by the Late Cretaceous *Globotruncana*-bearing marls of the *Scaglia Rossa*-type deposits (Fig. 11A) (Birkenmajer, 1979; Bąk, 2000) (Jaworki Formation). Red marls and marly limestones with greyish intercalations of calcareous sandstones and siltstones dominate in this outcrop. This

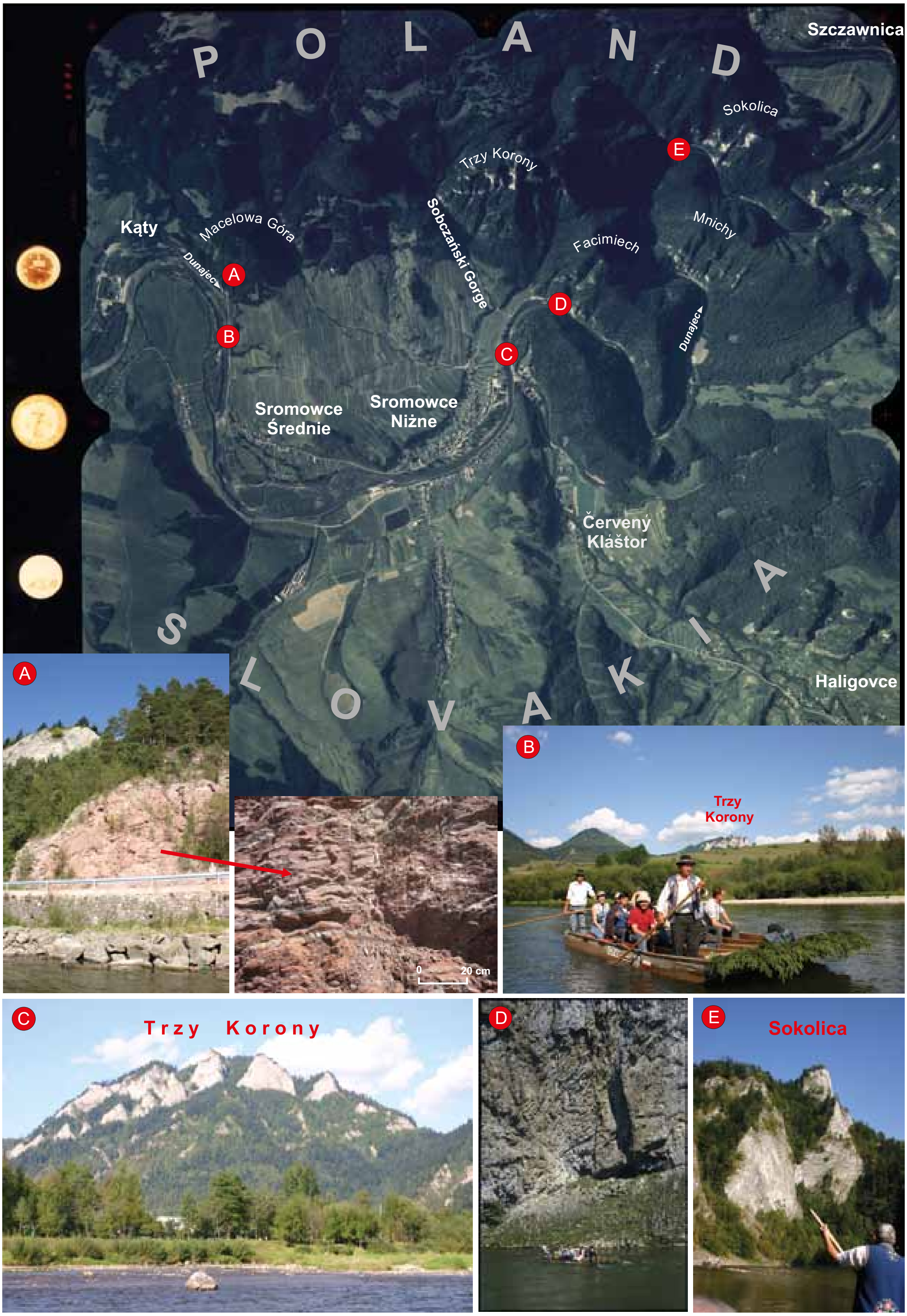


Fig. 50. Aerial view of the central Pieniny Mts and Dunajec River Gorge with points of photos: **A** – Upper Cretaceous red marls of the Scaglia Rossa-type facies (Macelowa Mt); **B**–**E** – views to Trzy Korony Mt, Mnichy Mt and Sokolica Mt built by uppermost Jurassic–lowermost Cretaceous *Maiolica*-type well-bedded limestones, usually strongly tectonic folded, with cherts (see – detail) over Dunajec River Gorge (after Krobicki & Golonka, 2008a)

is youngest part of multicoloured (green-variegated-red) globotruncanid marls of so-called Macelowa Marl Member of the Jaworki Formation with good foraminiferal Late Cretaceous biozonation (*Dicarinella concavata – D. asymmetrica* foraminiferal zones of the Upper Coniacian–Santonian) (Bąk, 2000). These deposits originated during last episode of evolution of the Pieniny Klippen Basin, when unification of sedimentary facies took place within all successions. The Tethyan Ocean *Scaglia Rossa* type facies (= *Couches Rouge* = *Capas Rojas*) wide-spread in Late Cretaceous indicate a wide connections between several branches of this ocean. In close view we can see red marls and marly limestones with greyish intercalations of calcareous sandstones and siltstones of turbiditic origin connected with distal parts of submarine fans, typical flysch and flyschoidal sedimentation (comp. Mutti & Ricci Lucchi, 1975; Słomka, 1986; Stow, 1986; Mutti & Normark, 1987; Ghibaudo, 1992; Reading & Richards, 1994; Lowe, 1997; Shanmugam, 2000). The Jaworki Formation is composed (stratigraphically) by green-variegated-red globotruncanid marls with perfect studied biostratigraphy (Bąk, 2000). Such type of facies is wide known both from Alpine and Apennine geology. The primary seaways between several parts of the Tethyan Ocean, especially the above mentioned Alpine Tethys (Golonka, 2004), which existed during the Late Cretaceous times are very well documented by these facies occurrences.

Passing by the Macelowa Mt the river crosses through the major vertical strike-slip fault separating the Pieniny Klippen Belt from the rolling hills of the Central Carpathians Paleogene Flysch region. These hills are built of Central Carpathian Paleogene flysch rocks deposited within the Spiš depression of the Central Carpathian Paleogene Basin.

At Sromowce Niżne village the Dunajec River enters again the Pieniny Klippen Belt. The most famous and beautiful peak of the Pieniny Mts – **Trzy Korony Mt** (Three Crowns) (982 m a.s.l.) (Fig. 50B, C) is good visible at the beginning of the Dunajec River Gorge (Birkenmajer *et al.*, 2001). In local, folk nomenclature, the Trzy Korony peak is known as (from left to right): Kaśka (Kate), Zośka (Sophie) and Kudłata Maryśka (Hairy Mary). The very steep walls of this peak are formed by strongly folded thin-bedded, grey, cherty limestones of the *Maiolica* (=*Biancone*) facies of the Pieniny Limestone Formation (Kimmeridgian-Albian) of the Pieniny Succession. The Trzy Korony Mt belongs to the most frequently visited and honored mountain peaks in Poland. The magnificent panorama from the summit allows to see several mountain ranges: the Małe Pieniny Mts, the Tatra Mts, the Spiska Magura, the Sądecki Beskid and the Gorce. The Pieniny Limestone Formation build steep another cliffs (**Grabczycha, Ostra Skała, Grabczycha Niżnia** and **Wyżnia** which are located at the entrance of the proper Dunajec River Gorge etc). The so-called **Zbójnicki Skok** or **Janosikowy Skok** (**Janosik's** or **highland robber's jump**) is the narrowest fragment of the Gorge. The river depth is about 8 m in this place. Janosik was a legendary chief of XVIIIth century highland robbers. He is famous and known as Polish and Slovak counterpart of Robin Hood. Many books, plays and movies were dedicated to Janosik. The plays include musical "Na Szkle Malowane (Painted on the Glass)", which has been continuously performed for thirty years in Bratislava, perhaps establishing the world record. According to legends, Janosik, when chased by policemen, escaped by jumping across the Dunajec River.

The origin of the Dunajec River Gorge is connected with young, mainly Neogene history, related to the neotectonic movements. 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 Mts followed the meandering stage of the Dunajec River. Then, during episode of the uplifting of the Pieniny Mts, such shape of the river have been conserved, when more competent Late Cretaceous marls and marly limestones, softer than Jurassic/Early Cetaceous cherty limestones, which occurred between several tectonic slices, have been easily eroded by the Dunajec River water, during the fault-related uplift. Origin of strongly faulted, usually thin-bedded *Maiolica*-type cherty limestones is connected with this tectonic activity of Alpine orogeny (Fig. 50E). 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). Some of the faults were still active during the Quaternary and were connected with rare earthquakes (Baumgart-Kotarba, 1996, 2001; Zuchiewicz *et al.*, 2002 and references therein). 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). Detail description of origin and age of the Dunajec River Gorge, with the review of structural and geomorphological features of the Pieniny Mts and formation of magnificent cliffs of the Gorge, was published by Birkenmajer (2006) (see also – Zuchiewicz, 1982).

On the right bank of the Dunajec River the medieval **Czerwony Klasztor** (**Red Monastery**) is visible, where one of the Camelot's pharmacist (Brother Cyprian – Franciszek Ignacy Jeszke) collected a huge quantity of plants (282 species), both from the Pieniny and Tatra mountains. The herbarium guide has the plant names in Greek, Latin, Polish, Slovakian, and German. The Pieniny highlanders believed that brother Cyprian was a sorcerer, which could fly on the home-made wings from the Trzy Korony Mt to the yard of the monastery (Tłuczek, 2004). The original name of the Red Monastery was "Lachnicki", its present name is derived from the red roofs of the buildings. It is located in the St. Andrew valley (elevation 453 m a.s.l.), at the Lipnik and Dunajec rivers and is surrounded by the houses of Czerwony Klasztor village (Tłuczek, 2004). The Red Monastery construction has started in 1330. In 1705 it was remodeled the Camelot order.

The Dunajec River is winding within the Gorge. The guides usually ask tourists to figure the direction of the river bends. The very sharp bend (145°) is at the foot of the **Barsztyg Mt**, on the right bank of the river. The **Facimiech Mt** built of the Pieniny Limestone Formation on the Polish side of the river form a large peninsula surrounded by Dunajec. The cherty limestones of the **Facimiech Mt** cliffs (783 m a.s.l.) are vertically arranged like most of the rocks in the Pieniny Klippen Belt. According to the boat guides, the rocks shape resembles the eagle (the Polish National emblem) and the nun. The karst processes formed the 20 m long cave and a shallow, but spacious hole under the Eagle, which serves as a shelter for boats during thunderstorms and heavy rains (Tłuczek, 2004).

One of the most spectacular views along rafting trip is **Mnichy** (**Seven Monks**) and **Sokolica Mt** (**Falcon's Mt**) (747 m a.s.l) belongs to the most beautiful Pieniny peaks. The name of this mountain is derived from numerous falcons, which used to nest here. The summit built of rigid cherty limestones lies 312 m above the river level (435 m a.s.l.). Southern cliffs of the mountains almost vertically tower over the river. The Sokolica is frequently visited by hikers because of magnificent view of the Tatra Mts and the ancient, 500 years-old pine forest (Tłuczek, 2004). The last *Maiolica* cliff on the left bank of the Dunajec River – **Hukowa Skała** (**Bang Rock**) ends the Dunajec River Gorge. In the XIXth century tourists announced they arrival by pistol and mortar shots. This tradition ended with the introduction of law restricting gun possession. The boat trip passes the Polish-Slovak border and ends at the harbor on the right bank of the river in Szczawnica town. **Szczawnica** is a spa town famous for its mineral waters. These mineral waters have been known since the medieval time, but first written remarks came from XVI century. In the XIXth century the Szczawnica owner Jozef Salay turned the small highland village into a health resort. Today Szczawnica is visited by hundreds thousands of spa visitors and tourists. 85% of town is located in the low and medium mountains.

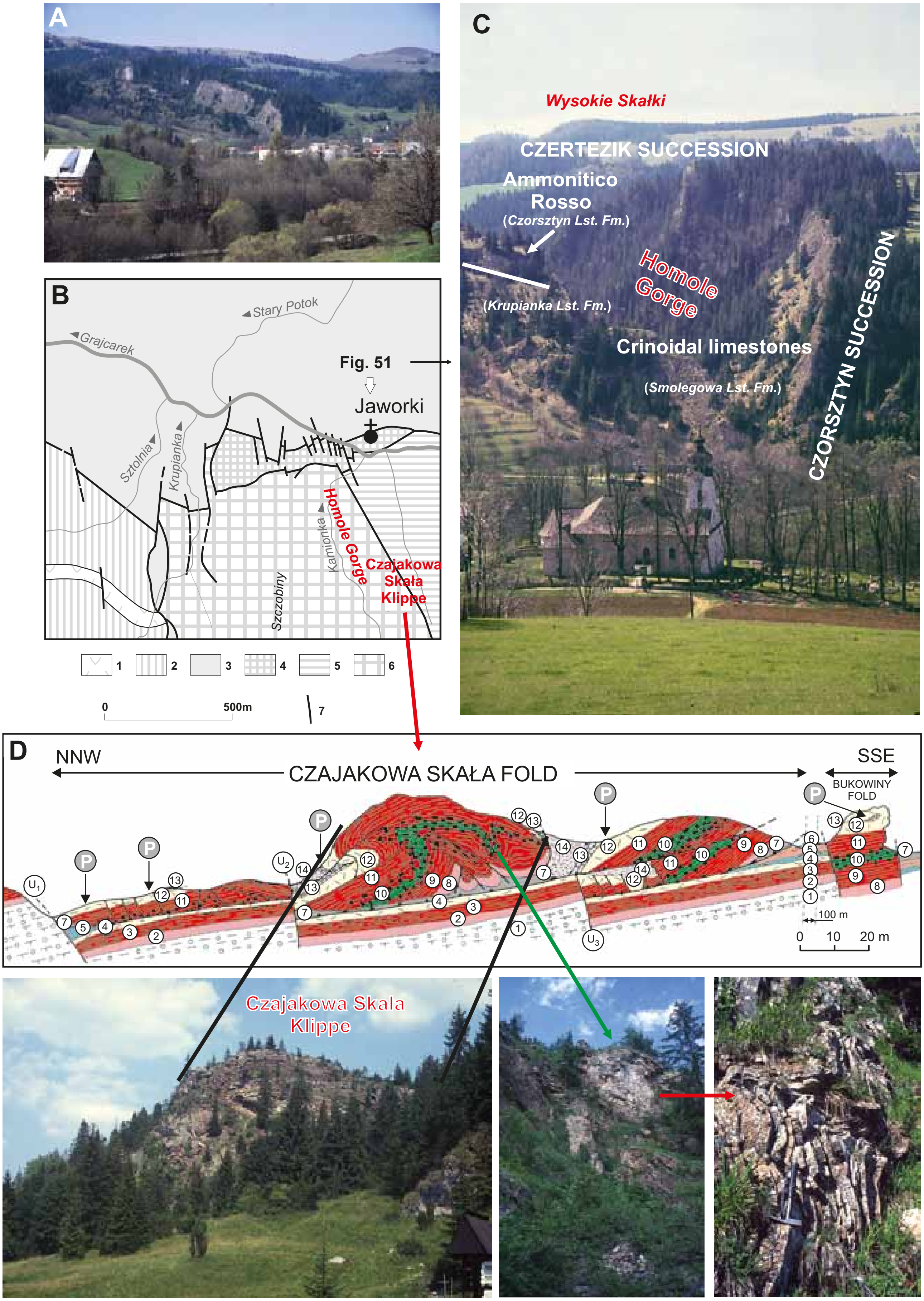
FIELD TRIP › STOP 8

HOMOLE GORGE NEAR JAWORKI VILLAGE (NAPPE STRUCTURE)

› FIGS 49D, 51–54

Michał KROBICKI & Jan GOLONKA

The famous deep Homole Gorge is cutting through the rocks of the so-called Homole tectonic block south of the Jaworki village (Fig. 49D). The origin this block is speculative and a subject of scientific debate. There is a variety of opinions, from the autochthonous position within the Czorsztyn ridge (e.g. Birkenmajer, 1986) through the nappe thrust over the other tectonic units (e.g. Książkiewicz, 1977; Golonka & Rączkowski, 1984; Jurewicz, 1997, 2005) to the olistoliths (Cieszkowski & Golonka, 2005, 2006; Cieszkowski *et al*., 2008). In the Homole Gorge up to near 100 m thick section of white crinoidal limestones of the Smolegowa Limestone Formation of the Czorsztyn Succession is exposed (Birkenmajer, 1963, 1977) (Figs 51, 52). These limestones are overlain by red crinoidal limestones (of the Krupianka Limestone Formation) and *Ammonitico Rosso*-type nodular limestones of the Czorsztyn Limestone Formation, which reach maximum 20 m in thickness. Both Smolegowa and Krupianka formations are Bajocian (Middle Jurassic) in age whereas condensed nodular limestones of the *Ammonitico Rosso* facies (Czorsztyn Limestone Formation) represent Middle Jurassic to Early Cretaceous. The decrease of sedimentation rate in pelagic sedimentation regime happened during late Middle–Late Jurassic times. This phenomenon is recorded by the deepening-upward sequence of deposits. The famous Pieniny Klippen Belt tectonic fold and thrust structures can be observed in the Czajakowa Skała Klippe in the upper part of the Homole Gorge (Birkenmajer, 1970, 1979) (Fig. 51). The Niedzica Nappe is there thrust here over the thick Czorsztyn Unit. Several beds representing the red nodular limestones of the *Ammonitico Rosso*-type of the Niedzica Limestone Formation and Czorsztyn Limestone Formation as well as intercalated radiolarites of the Czajakowa Radiolarite Formation are strongly tectonically disturbed, forming the overturned fold (Birkenmajer, 1970; Jurewicz, 1994). The type locality of the Czajakowa Radiolarite Formation occurs within this klippe. The Czajakowa Skała Klippe also shows a complete sequence of the Jurassic deposits of the Niedzica Succession (Birkenmajer, 1977; Wierzbowski *et al*., 1999). The oldest black and grey marly shales with spherosiderite concretions of the Skrzypny Shale Formation (Birkenmajer, 1977; Tyszka, 1994) are exposed here. These deposits are overlain by yellowish-greyish-red and dark cherry-greenish crinoidal limestones. The contact between black shales and crinoidal limestones is sharp and irregular. The ammonite fauna was found in the lowermost part of the crinoidal limestones of the Smolegowa Limestone Formation. These very precise biostratigraphical data suggest a hiatus between black shales and crinoidal limestones, spanning the Laeviuscula and a part of Propinquans ammonite zones of the Lower Bajocian (Middle Jurassic) (Krobicki & Wierzbowski, 2004; Krobicki, 2006). The small outcrops of these black shales occur below Czajakowa Skała Klippe within local landslide. The best outcrops of these rocks within the Pieniny Klippen Belt in Poland occur in the Krupianka stream, which run the Homole Gorge, westward. Their thickness reach 4–5 metres. The shales contain a lot of spherosiderites and sometimes perfectly preserved ammonites. The ammonite fauna indicates latest of Aalenian and/or earliest of Bajocian (Middle Jurassic) age of black shales, representing oxygen-depleted facies of the *Fleckenmergel*-type. The Cretaceous strata are the youngest part of this sequence, and especially the Early Cretaceous (Berriasian) deposits are extremely interesting by phosphatic event (large phosphatic macrooncoid pavements) (Fig. 51) which indicates upwelling currents in the Pieniny Klippen Basin in this time (Figs 53, 54).


A
B
Stary Potok
Grajcarek
Fig. 51
Jaworki
Sztolnia
Krupianka
Homole Gorge
Kamionka
Czajakowa
Skała
Klippe
Szczobiny
1
2
3
4
5
6
0
500m
7
C
Wysokie Skałki
CZERTEZIK SUCCESSION
Ammonitico
Rosso
(Czorsztyn Lst. Fm.)
(Krupianka Lst. Fm.)
Homole
Gorge
Crinoidal limestones
(Smolegowa Lst. Fm.)
CZORSZTYN SUCCESSION
D
NNW
SSE
CZAJAKOWA SKAŁA FOLD
BUKOWINY
FOLD
100 m
0
10
20 m
Czajakowa Skala
Klippe

←

Fig. 51. 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 Paleogene; 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, 2008a)

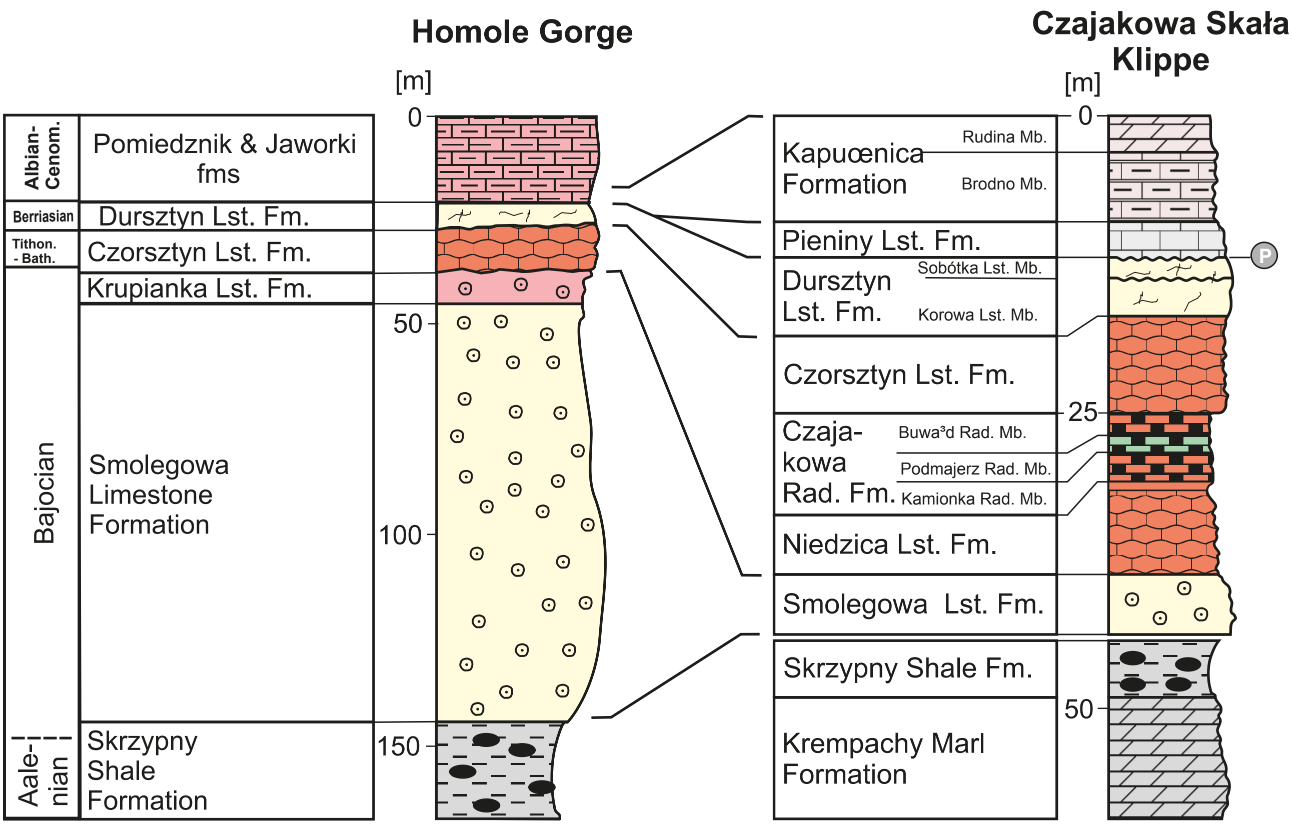


Fig. 52. Lithostratigraphical columns of the Czorsztyn Succession of the Homole Gorge (left) and Niedzica Succession of the Czajakowa Skała klippe (right) (after Birkenmajer, 1977; modified). Explanations of symbols – see the text (after Krobicki & Golonka, 2008a)

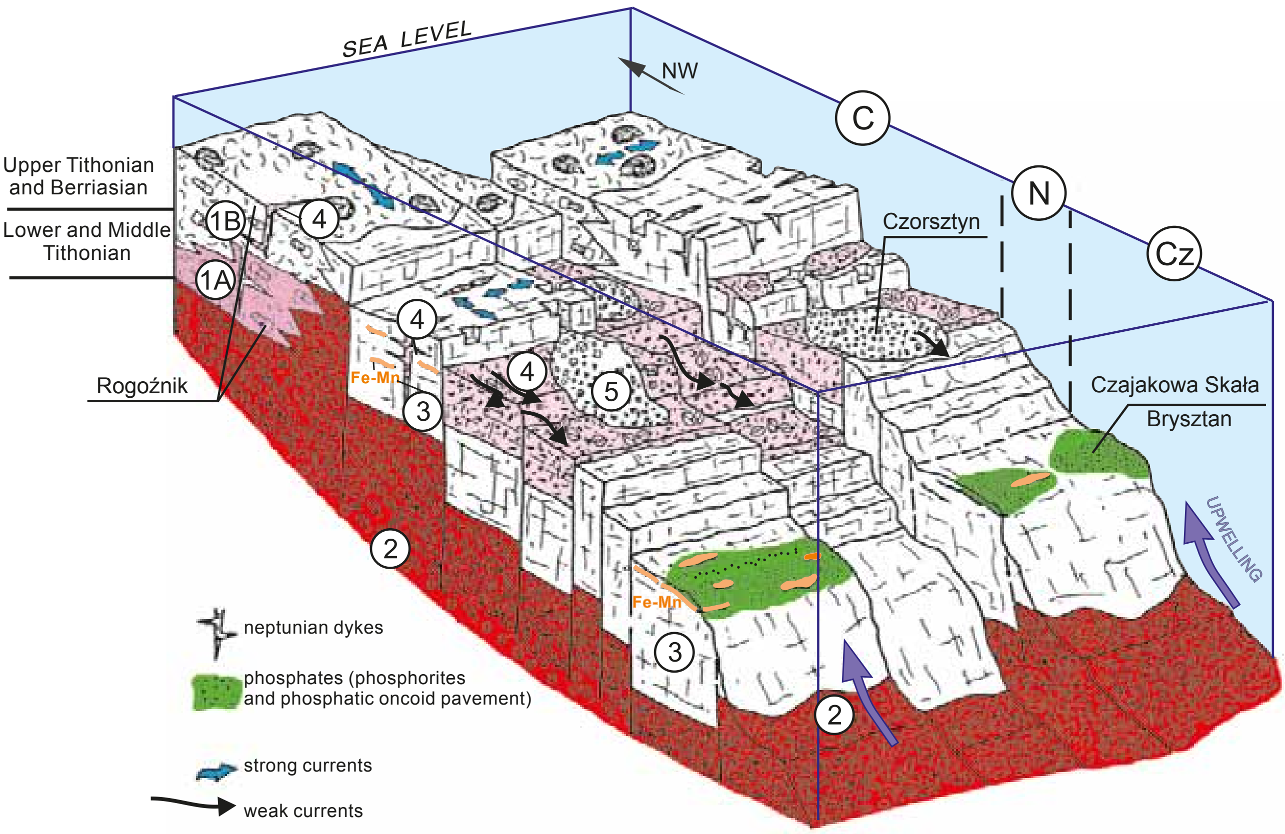


Fig. 53. Model of sedimentation on the intraoceanic Czorsztyn pelagic swell in Berriasian times with effects of pronounced Neo-Cimmerian tectonic movements (after Krobicki 1996; modified).
Abbreviations: 1 – Rogoźnik Coquina Member (A – sparitic coquina; B – micritic coquina); 2 – Czorsztyn Limestone Formation (*Ammonitico Rosso* facies); 3 – Sobótka Limestone Member (Dursztyn Limestone Formation); 4 – Harbatowa Limestone Member (Łysa Limestone Formation); 5 – Walentowa Breccia Member (Łysa Limestone Formation); successions: C – Czorsztyn; N – Niedzica; Cz – Czertezik (after Krobicki *et al*., 2010; slightly changed)

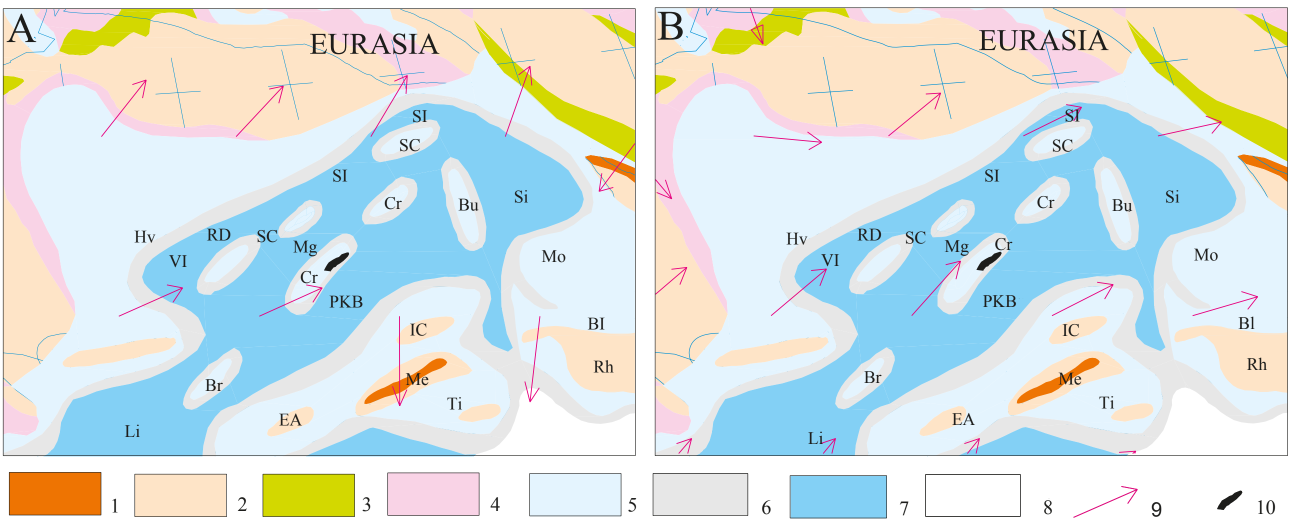


Fig. 54. Palaeoenvironments, wind direction and upwelling zones of the Carpathian area during Tithonian–Berriasian time (palaeogeography after Golonka *et al*., 2000, modified). **A** – summer Northern Hemisphere; **B** – winter Northern Hemisphere
1 – mountains/highlands (active tectonically); 2 – topographic 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 – Sinaia Basin; Sl – Silesian Basin; Ti – Tisa Plate; Vl – Valais Trough (after Golonka & Krobicki, 2001)

ROGOŹNIK (JURASSIC/CRETACEOUS AMMONITE COQUINA)

▸FIGS 55–58

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The Czorsztyn Succession of the PKB is best exposed at the Rogoźnik Klippe (Figs 49A, 55-57), Homole Gorge (Figs 49D, 51, 52) and Czorsztyn Castle (Figs 49B, 59, 60). The Rogoźnik (or Rogoża) klippen (Alexandrowicz *et al*., 1997) lies within the nature reserve, located south of the village of Rogoźnik, and of the Wielki Rogoźnik stream (**this is protected area – please, do not hammer the rocks**). The locality comprises a small klippe as well as the neighboring abandoned quarry (Fig. 55A, B). The former is the type locality of the Rogoźnik Coquina Member, the latter of the Rogoża Coquina Member (Birkenmajer, 1977). Both these lithostratigraphic units constitute a special ammonite coquina-type development of the Upper Jurassic deposits markedly different from coeval nodular limestones especially well represented in the Czorsztyn Succession of the Pieniny Klippen Belt (cf. Kutek, Wierzbowski, 1986). The klippen at the Rogoźnik village are well known since the 19th century (e.g. Zittel, 1870; Neumayer, 1871; Uhlig 1890; for full list see Arkell, 1956; see also – Birkenmajer, 1963), as the Pieniny Klippen Belt the richest Upper Jurassic ammonite locality with very detail biostratigraphic documentation (Kutek & Wierzbowski, 1986; Wierzbowski *et al*., 2006), and with rich brachiopod fauna as well (Krobicki, 1994; Wierzbowski *et al*., 2006). The succession of ammonites (Fig. 57) extends from Lower Tithonian up to the lowermost Berriasian. The discussed deposits of the Rogoźnik Coquina Member consist of pink to reddish coquinas with voids filled with white sparry calcite and numerous fossils (mostly ammonites, and their debris, aptychi, brachiopods, crinoid fragments) and less common gastropods, echinoids, bivalves, solitary corals, sponges, fish teeth.

In the studied section of the Rogoźnik Coquina Member, the Lower–Middle Tithonian brachiopod fauna differs distinctly from the Upper Tithonian–Berriasian one (see Fig. 58) (Krobicki, 1994). High percentage of both rhynchonellids (*Lacunosella*) and dallinid *Dictyothyropsis* in the Upper Tithonian–Berriasian strata suggests apparently shallower depositional environment of these rocks in comparison with the older part of the section.

Difference between Lower–Middle Tithonian and Upper Tithonian–Berriasian brachiopod assemblages known from this section reflect change of palaeoenvironmental conditions. One of the consequent quantitative and qualitative changes of brachiopod fauna composition is the replacement of primarily dominating pygopids s.l. (*Pygope* and *Nucleata*, Lower–Middle Tithonian) by "shallow-water" assemblage with an important role of rhynchonellids (*Lacunosella*) and the dallinid *Dictyothyropsis* (Upper Tithonian–Berriasian) (Fig. 58). Quantitative and qualitative changes in brachiopod fauna assemblages dependent on palaeogeographic position of specific facies in the palinspastic reconstruction of the Pieniny Klippen Belt may be applied to the determination of palaeoecological conditions basing on domination/presence/absence analysis of given taxons in succeeding lithofacies. In this sense, the main diagnostic features are: the presence of rhynchonellids of the genus *Lacunosella*, dallinid *Dictyothyropsis tatrica*, as well as trend of quantitative changes in the occurrence of pygopids s.l. (*Pygope* and *Nucleata*). Rhynchonellids (*Lacunosella hoheneggeri* and *L. zeuschneri*) commonly occur in shallow-water, reef-like, Tithonian–Berriasian and Lower Valanginian sediments (so-called tramberk limestones) of the Outer Carpathian basins (Nekvasilová, 1977). The *Lacunosella hoheneggeri* was regarded by Ager (1965) as typical of sublittoral zones. A few *Dictyothyropsis tatrica* specimens also have their equivalents in the tramberk limestones (Barczyk, 1979, 1991). Pygopids s.s. were usually interpreted as deep-water organisms (Ager, 1976). Dieni and Middlemiss (1981) described an abundant collection of these brachiopods from the Venetian Alps, from pelagic limestones of *Maiolica* facies which are the facial equivalent of the cherty limestones of Branisko and Pieniny successions (Pieniny Limestone Formation). Recently, it was suggested that pygopids s.s. could occupy also shallow waters over the seamounts (? = Czorsztyn pelagic swell) (Ager, 1993).

In the Czorsztyn Succession the significant amounts of *Lacunosella* suggest generally shallower deposition environment of the enclosing sediments in comparison with those strata in which rhynchonellids are rare or absent. From the other side, the increasing percentage of pygopids s.l. in the assemblage suggests deeper environments. Usually, if *Lacunosella* representatives constitute about 25% of the assemblage, both *Pygope* and *Nucleata* are subordinate and *vice versa*.

Fig. 55. General view of Rogoźnik quarry and klippen (**A**), quarry with Jurassic–Cretaceous boundary (**B**) with typical micritic ammonite-rich coquina (**C**) and Jurassic–Cretaceous boundary in famous Rogoźnik ammonite coquina: **D** – boundary between Jurassic and Cretaceous beds in the Rogoźnik Klippe; **E** – typical ammonite sparry coquina full of ammonites usually filled by sparry calcite (**F** — arrow) with typical ornamentation of uppermost Jurassic ammonite *Simocosmoceras* (**G**)

Fig. 56. Detail biostratigraphical zonation of the Rogoża coquina (after Rehakova & Wierzbowski, 2011)

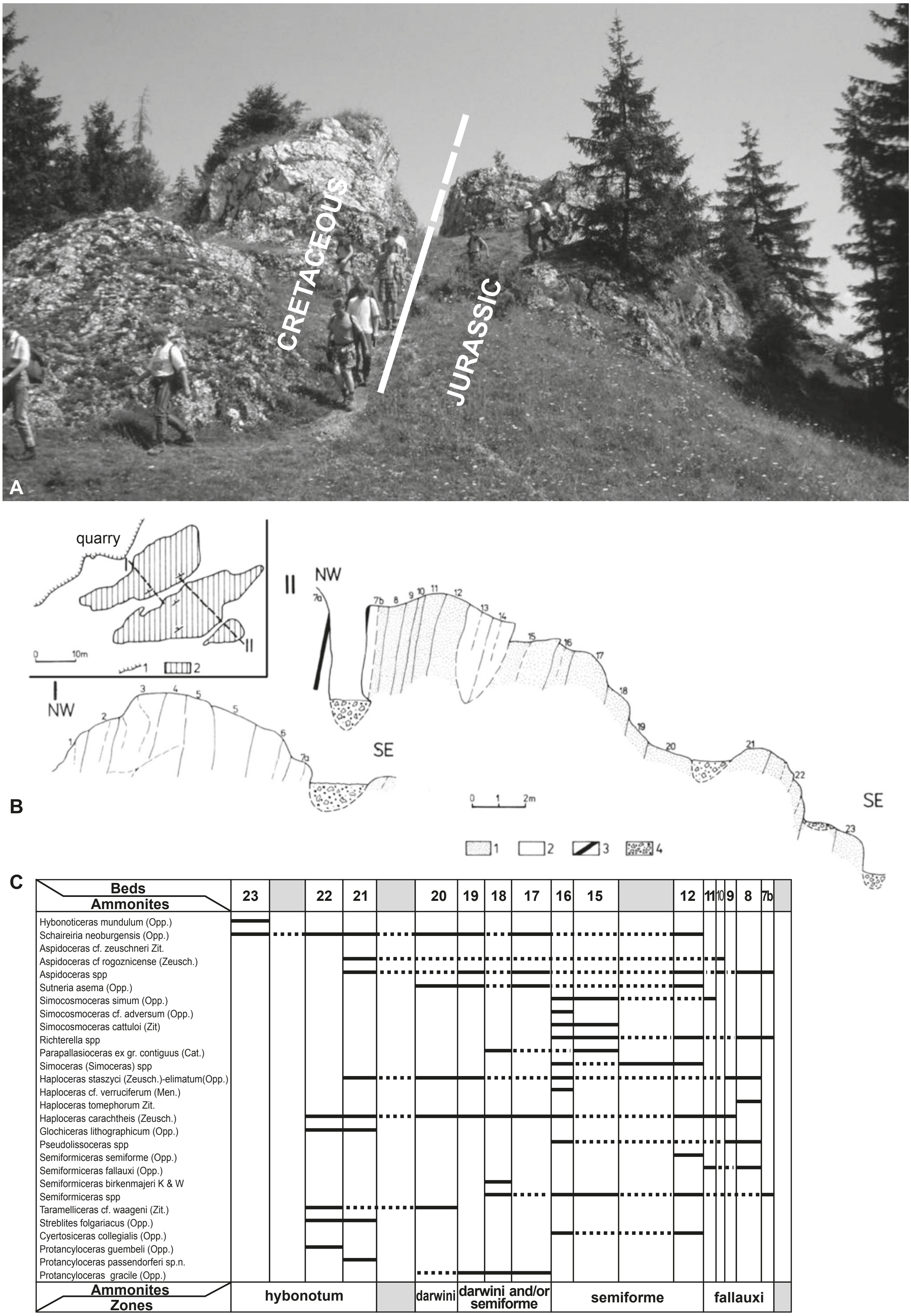

Fig. 57. Cross-section of the Rogoźnik Klippen (**A**), their detail section (**B**) and precise ammonite biostratigraphy (**C**) (after Kutek & Wierzbowski, 1986; modified)

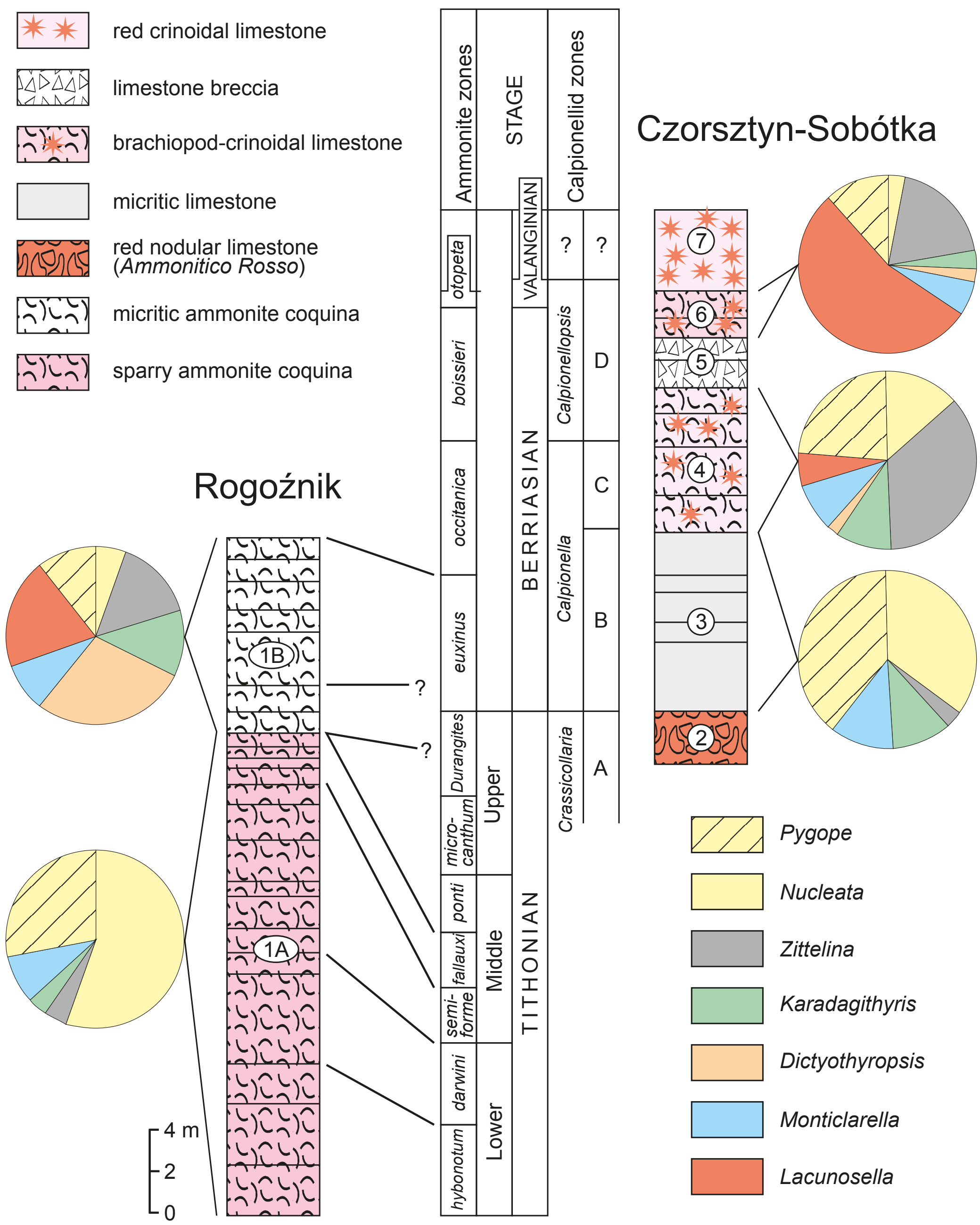

Fig. 58. Correlation of Tithonian–Berriasian deposits from Rogoźnik Klippen and Czorsztyn-Sobótka outcrops with pie cherts of brachiopod assemblages (stratigraphy after Kutek & Wierzbowski, 1986; Wierzbowski & Remane, 1992; lithostratigraphy after Birkenmajer, 1977)

Key: 1 – Rogoźnik Coquina Mbr (Czorsztyn Limestone Fm.); 2 – Czorsztyn Limestone Fm.; 3 – Korowa and Sobótka Limestone mbrs (Dursztyn Limestone Fm.); 4-6 – Łysa Limestone Fm. (4 – Harbatowa Limestone Mbr; 5 – Walentowa Breccia Mbr; 6 – Kosarzyska Limestone Mbr); 7 – Spisz Limestone Fm. (after Krobicki, 2006; modified). For detailed description, see text

CZORSZTYN CASTLE (JURASSIC–CRETACEOUS DEPOSITS OF THE CZORSZTYN SUCCESSION) (FIGS 49B, 59, 60)

Michał KROBICKI & Jan GOLONKA

Caution: This is a protected area - please, do not hammer the rock!!

The Czorsztyn Castle klippen are one of the most famous geological site of the Pieniny Klippen Belt with full sequence of Czorsztyn Succession from the Middle Jurassic up to Upper Cretaceous deposits, rich in invertebrate fossils such as: ammonites, brachiopods, crinoids, calpionellids, foraminifers, described and illustrated by numerous authors since beginning of the XIX century (e.g., S. Staszic, L. Zejszner, E. Suess, M. Neumayr, K. A. Zittel, V. Uhlig and others) (Uhlig, 1890; Birkenmajer, 1963, 1977, 1979, 1983; Barczyk, 1972a,b; Głuchowski, 1987; Krobicki, 1994, 1996; 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) of the Czorsztyn Succession (Birkenmajer, 1977; see also Birkenmajer, 1963) (Fig. 59). New data on Middle–Upper Jurassic macrofauna and microfacies development (Wierzbowski *et al.*, 1999) indicate that the oldest are grey crinoid limestones of the Smolegowa Limestone Formation. These are well-bedded grainstones and the youngest beds are cross-bedded (Fig. 59D) well recorded shallow marine origin of these limestones. Gastropod trace fossils found in the base of these limestones supported such idea (Krobicki & Uchman, 2004). 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 and Wierzbowski (2004), whereas the nodular limestones directly overlying crinoid limestones of the Krupianka Limestone Formation in the Czorsztyn Castle Klippe (Fig. 59E, F) 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 ferromanganese crust, very typical feature for this boundary surface, known from several other outcrops in the Pieniny Klippen Belt, both in Polish, Slovakia and Ukrainian part of the region. The overlying nodular limestones correspond already to the Czorsztyn Limestone Formation (red nodular limestone). These limestones are very well visible just under the wall of the castle and within it (so-called "lower castle"). The lowermost part of the nodular limestones of the Czorsztyn Limestone Formation exposed in the Czorsztyn Castle Klippe yielded the rich ammonite faunas. They were collected bed-by-bed and have been use to detailed biostratigraphicaly dated levels. 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). In several other places of Alpine-Appenines and Carpathian-Dinarides arcs such Middle-Upper Jurassic *Ammonitico Rosso* facies are known very well (for example, famous Julia's loggia in Verona is built by so-called *Ammonitico Rosso Veronese* and half of this beautiful city as well). Jurassic–Cretaceous sequence of the Czorsztyn Succession of the Czorsztyn Castle Klippen gave also possibility to correlate two biostratigraphical schemes for Jurassic strata – ammonite on the one side and calpionellid on the other – first time in the whole Carpathians (Wierzbowski & Remane, 1992). Spectacular distribution of Middle and Upper Jurassic microfacies (within

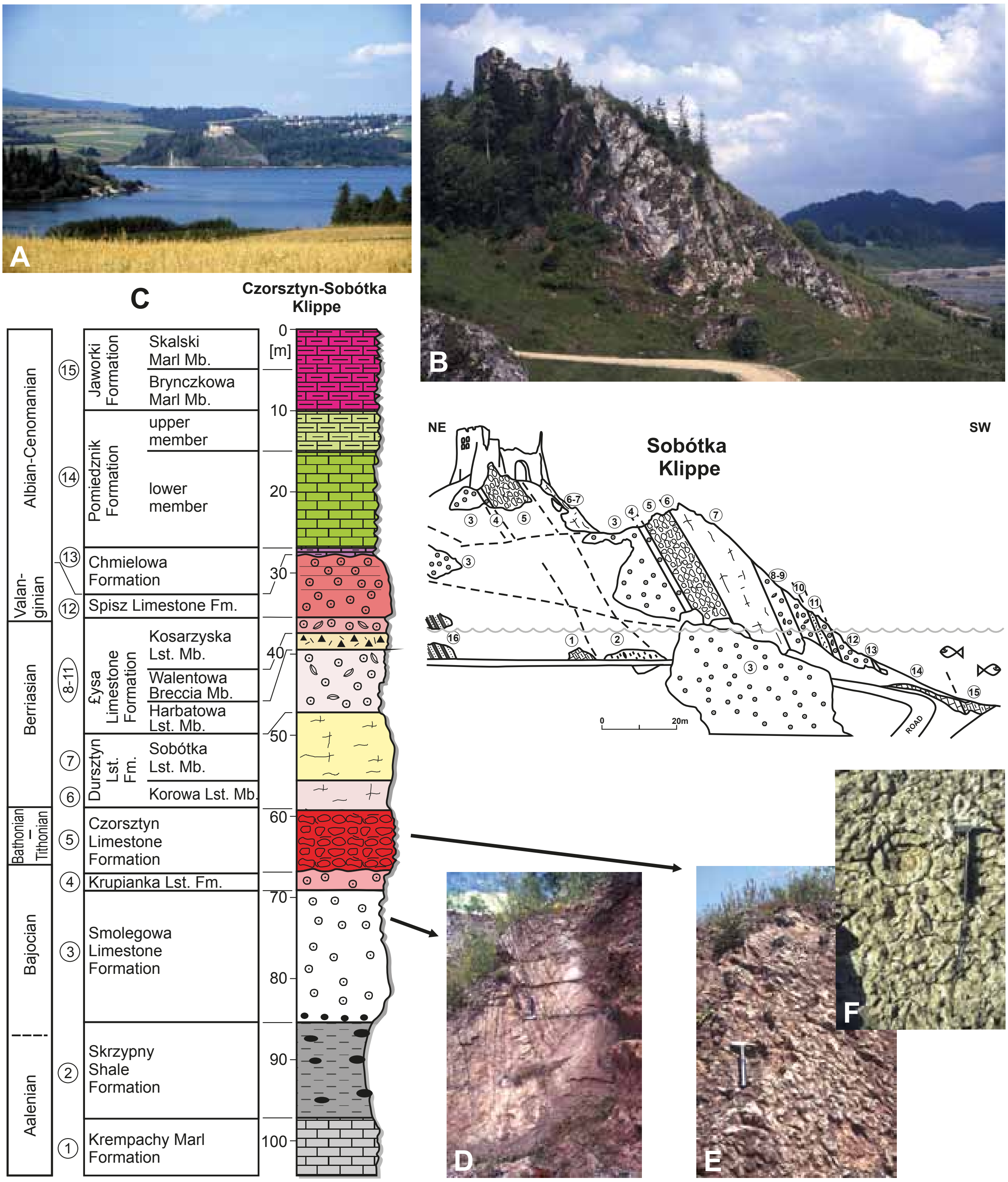

Fig. 59. A – panoramic view on the Czorsztyn Castle over artificial Czorsztyn Lake; **B** – a past view of the Sobótka Klippe (state – 1992) with geological sketch of this section and the Czorsztyn Castle Klippe (Czorsztyn) (after Birkenmajer, 1963, 1979, modified). **C** – lithostratigraphic column of the Czorsztyn Succession in the Czorsztyn-Sobótka Klippe and the Czorsztyn Castle Klippe (after Birkenmajer, 1977); **D** – cross-bedded structure within white crinoidal limestones of the Smolegowa Limestone Formation (Czorsztyn Castle Klippe); **E** – typical nodular feature of red *Ammonitico Rosso*-type limestones with large ammonites; **F** – Czorsztyn Succession in the Czorsztyn-Sobótka Klippe and the Czorsztyn Castle Klippe (after Birkenmajer, 1977)
Explanations: 1 – Krempachy Marl Fm. (spotty limestones/marls of -type facies); 2 – Skrzypny Shale Fm. (black shales with spherosiderites); 3 – Smolegowa Limestone Fm. (white crinoidal limestones); 4 – Krupianka Limestone Fm. (red crinoidal limestones); 5 – Czorsztyn Limestone Fm. (red nodular limestones of the *Ammonitico-rosso* type facies); 6-7 – Dursztyn Limestone Fm.: 6 – Korowa Limestone Mbr. (pink micritic limestones); 7 – Sobótka Limestone Mbr. (white micritic limestones); 8-11 – Łysa Limestone Fm.: 8-9 – Harbatowa Limestone Mbr. (brachiopod-crinoidal limestones); 10 – Walentowa Breccia Mbr. (sedimentary limestone breccia); 11 – Kosarzyska Limestone Mbr. (crinoidal-brachiopod limestones); 12 – Spisz Limestone Fm. (red crinoidal limestones); 13 – Chmielowa Fm. (violet limestones); 14 – Pomiedznik Fm. (marls/limestones); 15 – Jaworki Fm. (variegated marls); Magura Succession (Grajcarek Unit): 16 – Szlachtowa Fm. (black flysch) (after Krobicki & Golonka, 2008c)

red nodular limestones of the Czorsztyn Limestone Formation) has been made first time here as well (i.a.: the *Bositra* shell filament microfacies – with subordinate juvenile-gastropod microfacies: uppermost Bajocian-lowermost Bathonian; the *Globuligerina* microfacies: Oxfordian, and the *Saccocoma* microfacies: Kimmeridgian-Lower Tithonian) (Wierzbowski *et al*., 1999). The whole Cretaceous strata were visible previously (from Berriasian limestones up to Santonian marls), including very characteristic syn-sedimentary limestone breccias of the so-called Walentowa Breccia Member of the **Łysa** Limestone Formation (Berrisian in age) (Fig. 60) which indicates the earliest Cretaceous (Neo-Cimmerian) tectonic movements in this part of the Tethys (Fig. 53).

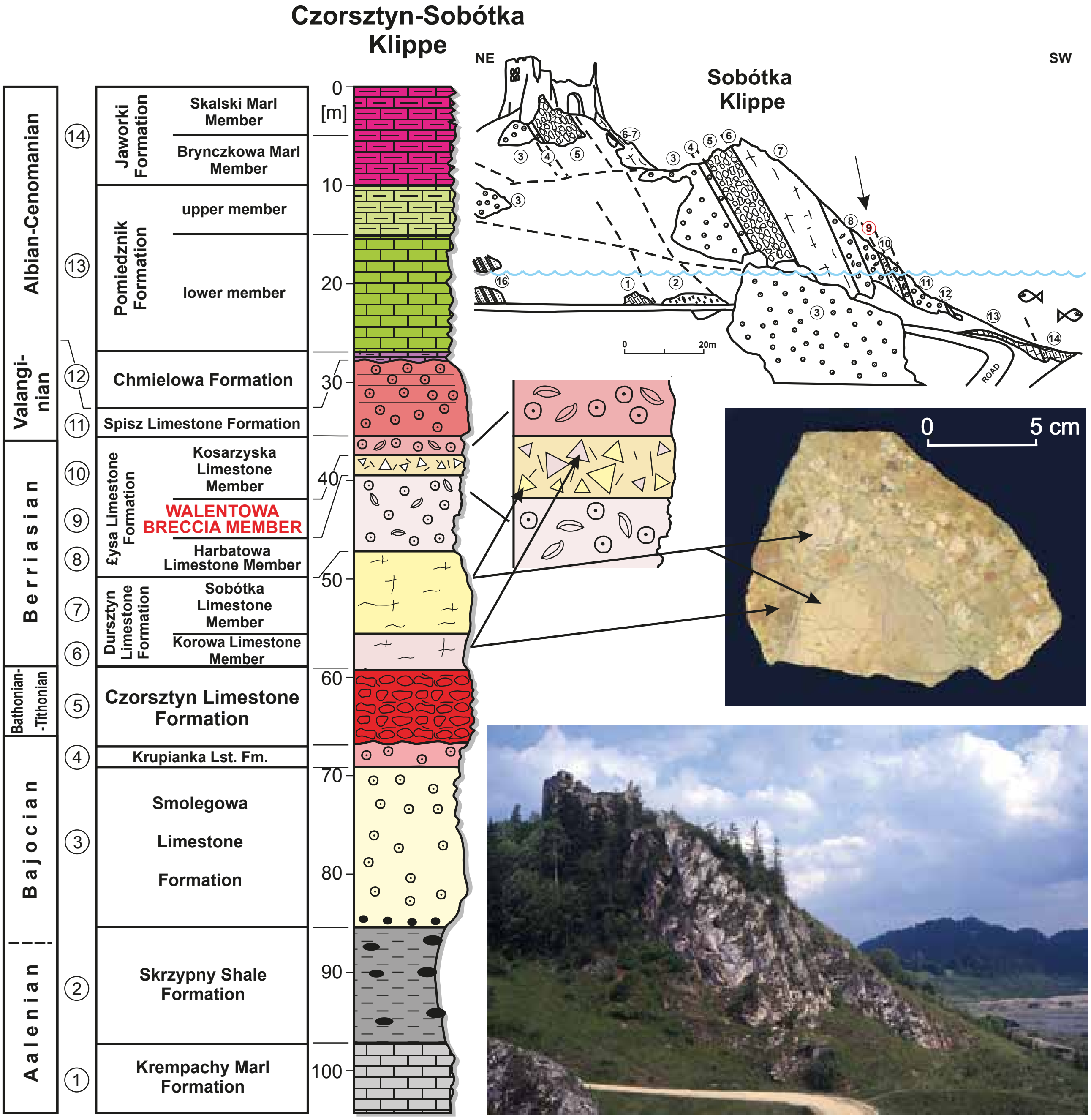

Fig. 60. Stratigraphical section of the Czorsztyn-Sobótka Klippe (PKB) with indication of position of the Walentowa Breccia Member of the Łysa Limestone Formation of the Czorsztyn Succession (lithostratigraphy after Birkenmajer, 1977, slightly modiefied) (photo – state in 1992)

Explanations of lithology: 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. 59) (after Krobicki *et al*., 2010)

FIELD TRIP ▹STOP 11

FLAKI RANGE (JURASSIC DEPOSITS OF THE BRANISKO SUCCESSION)

▹FIGS 61, 62

Michał KROBICKI & Jan GOLONKA

Note: (**this is Pieniny National Park – please, do not hammer the rocks**).

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) (Fig. 61). These rocks are surrounded by less resistant Upper Cretaceous marls and flysch siliciclastics belonging to different tectonic units of the Pieniny Klippen Belt. 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; Birkenmajer *et al.*, 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) (Fig. 62). 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 (2002, 2004) presented their ichnofabric from the Triassic and Jurassic of Japan.

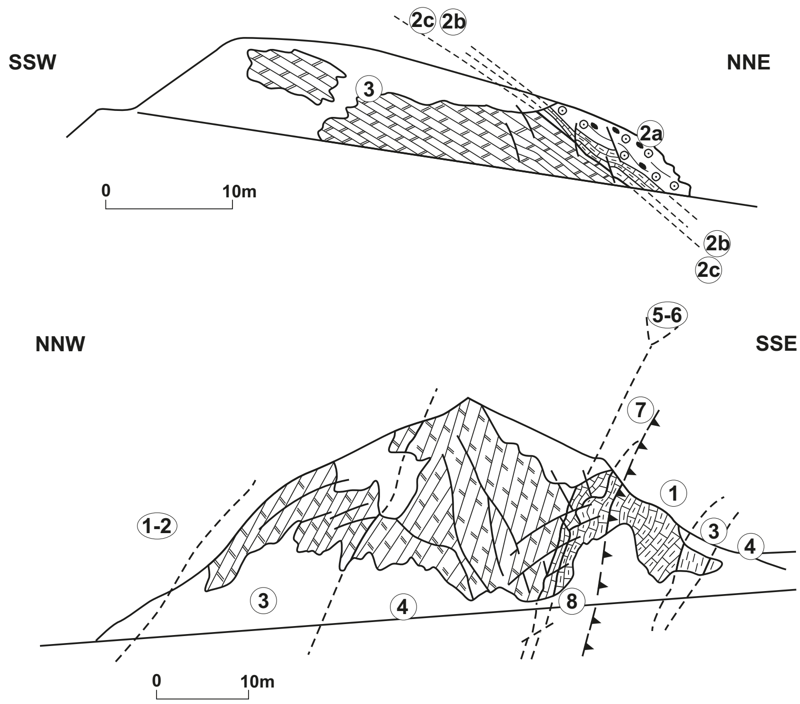

Fig. 61. View of the Flaki Range sections; Branisko Succession (next page: **A** – western side; **B** – eastern side) and general sketch of studied sections (after Birkenmajer *et al.*, 1985)

Fig. 61. Cont. View of the Flaki Range sections; Branisko Succession (**A** – western side; **B** – eastern side) and general sketch of studied sections (previous page) (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)

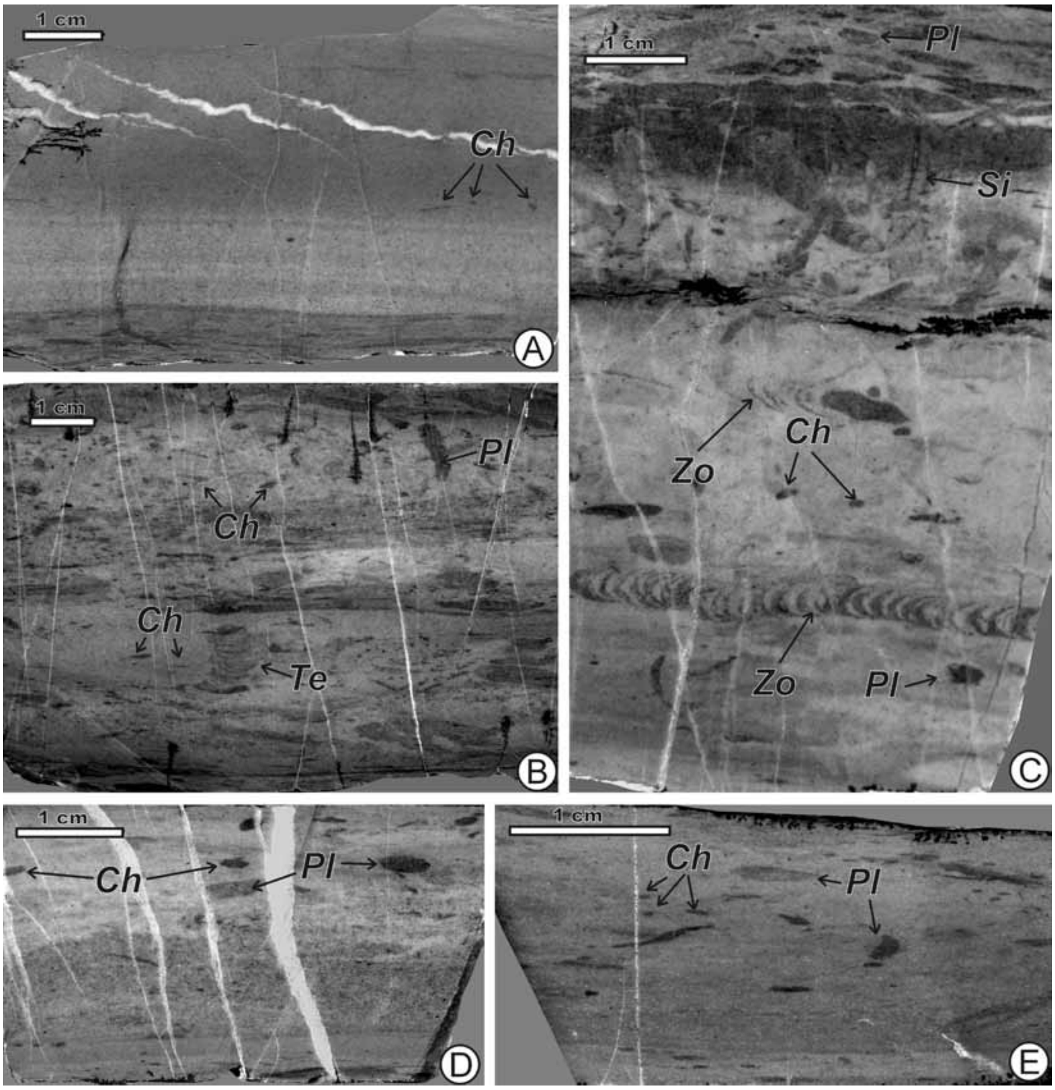

Fig. 62. Trace fossils within Sokolica Radiolarite Fm. in the Flaki Range, Branisko Succession: Ch – *Chondrites*; Pl – *Planolites*; Si – *Siphonichnus*; Te – *Teichichnus*; Zo – *Zoophycos* (after Krobicki *et al.*, 2006)

WŻAR MT (MIOCENE ANDESITES AND PANORAMIC VIEW)

▹FIGS 49B, 63

Jan GOLONKA & Michał KROBICKI

The most famous outcrop (artificial one – abandoned quarry) of the Middle Miocene volcanism of the Pieniny Mts occur on the Wżar Mt, near Snozka pass, and is represented by two generation of intrusive dykes and sills. In half of the XX century several pioneer researches were done both geologically, mineralogically/petrographically and geophysically (e.g. Wojciechowski, 1950, 1955; Birkenmajer, 1956a,b, 1958; Małoszewski, 1957, 1958; Kardymowicz, 1957; Małkowski, 1958; Gajda, 1958; Kozłowski, 1958). The Neogene volcanic activity in Carpathian–Pannonian region was widespread. The Pieniny Andesite Line is an about 20 km long and 5 km wide zone, which cut both Mesozoic–Paleogene rocks of the Pieniny Klippen Belt and Paleogene flysch of the Magura Nappe of the Outer Flysch Carpathians. Andesites occur in the form of dykes and sills. At the Wżar Mt two generations of andesitic dykes occur (Youssef, 1978). Numerous older dykes are sub-parallel to the longitudinal distribution of the Pieniny Klippen Belt 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, b).

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

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

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

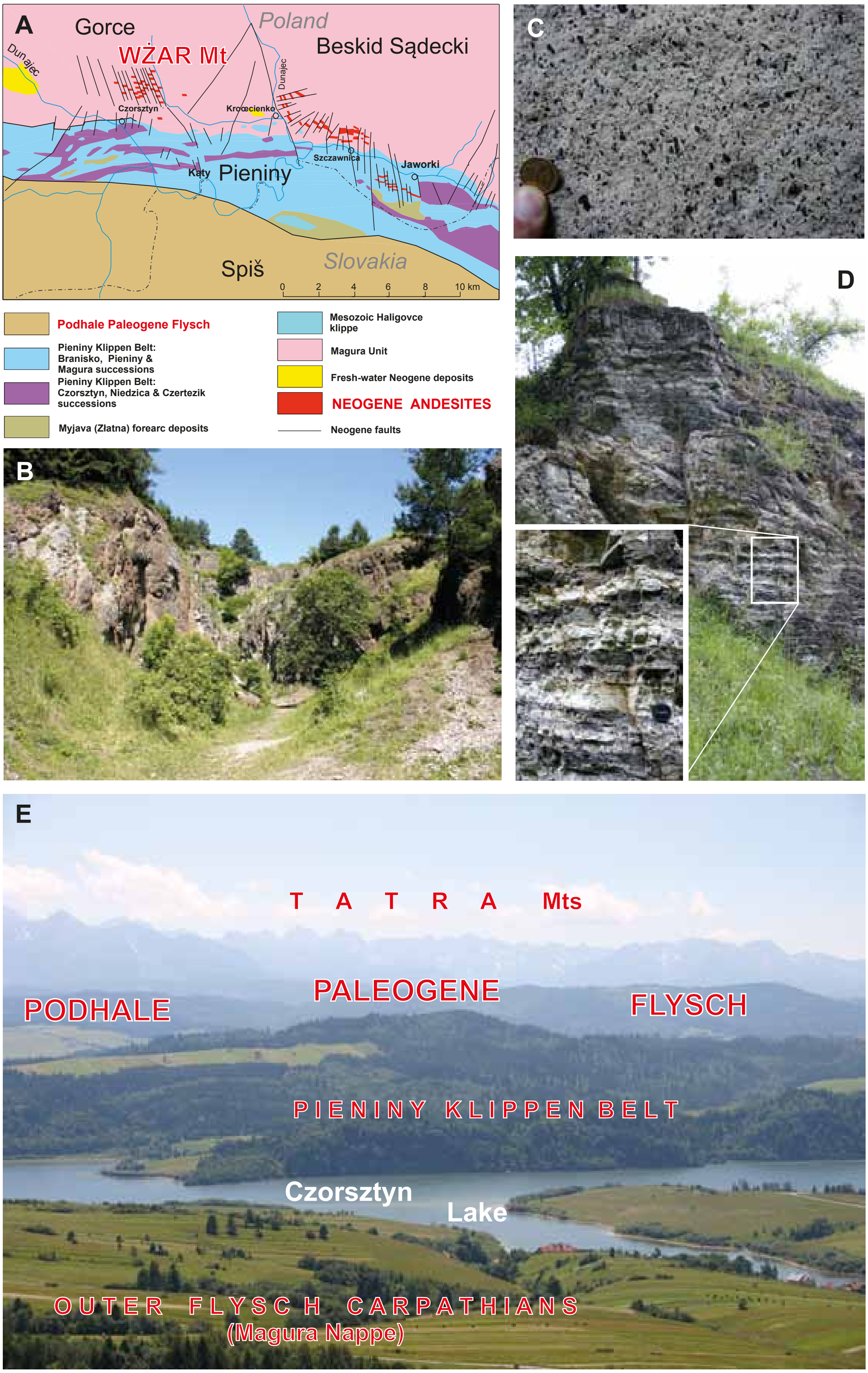

Fig. 63. Geological position of Miocene andesites of the so-called Pieniny Andesite Line: **A** – geological sketch of the Pieniny Klippen Belt (Polish sector) and surrounding regions (after Birkenmajer, 1979; simplified) with location of Wżar Mt; **B** – main entrance to abandoned quarry; **C** – andesites with piroxenes and amphibolites; **D** – thermally change of flysch deposits of the Magura Unit (Outer Flysch Carpathians) on the contact with andesites; **E** – general view of Inner Carpathians from topmost part of the Wżar Mt

FIELD TRIP ▹ STOP 13

SKRZYDLNA QUARRY (OLIGOCENE MENILITE FORMATION)

▹FIGS 64–69

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Skrzydlna is a village about 50 km to the south east of Kraków (Fig. 64). In the southern part of the village, near the local road between Skrzydlna and Kasina Mała, a quarry is located, in which almost 200 m thick succession of the Menilite Formation is exposed. This Oligocene succession forms an organic-rich lithofacial unit that is widely widespread in the Carpathians (Fig. 65). It is composed mainly of dark brown shales, chert and siliceaous marls with minor intercalation of sandstones. Thickness of these deposits varies from about 100 m in southern part of their occurrence up to 550 m in the northern part (Kuśmierek, 1990). However, in some areas they contain also thick sandstone bodies that may occasionally comprise half of their total thickness. A unique value of the Menilite Formation lies in relatively high organic matter content of up to 20% (e.g. Curtis *et al.*, 2004) thus they represent the most important source rock for hydrocarbons in the entire Carpathian realm. Thick sandstone complexes within these Beds are, in turn, significant reservoir rocks. Polish Carpathians constitute one of the oldest petroleum province in the world where oil exploitation dates back to the middle of 19th century.

The Skrzydlna quarry is situated in a structurally complex area. The Menilite Formation visible in the quarry

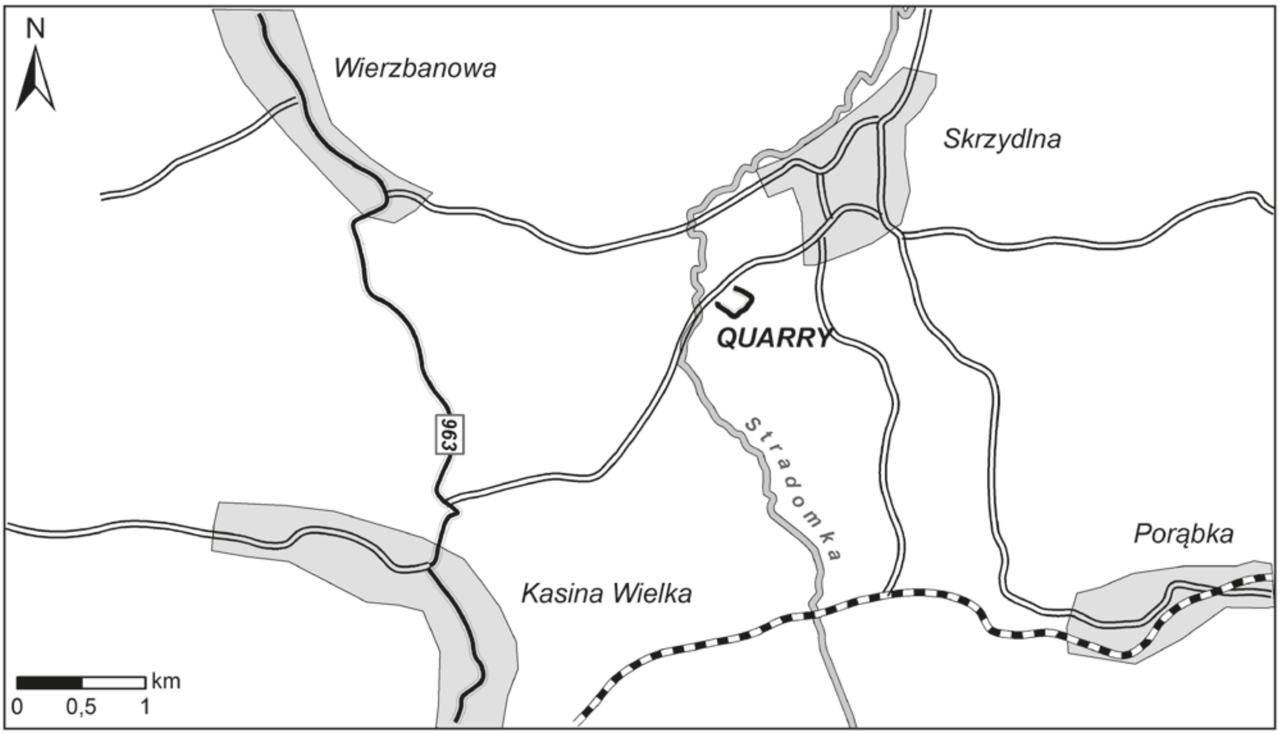

Fig. 64. Location of the Skrzydlna quarry

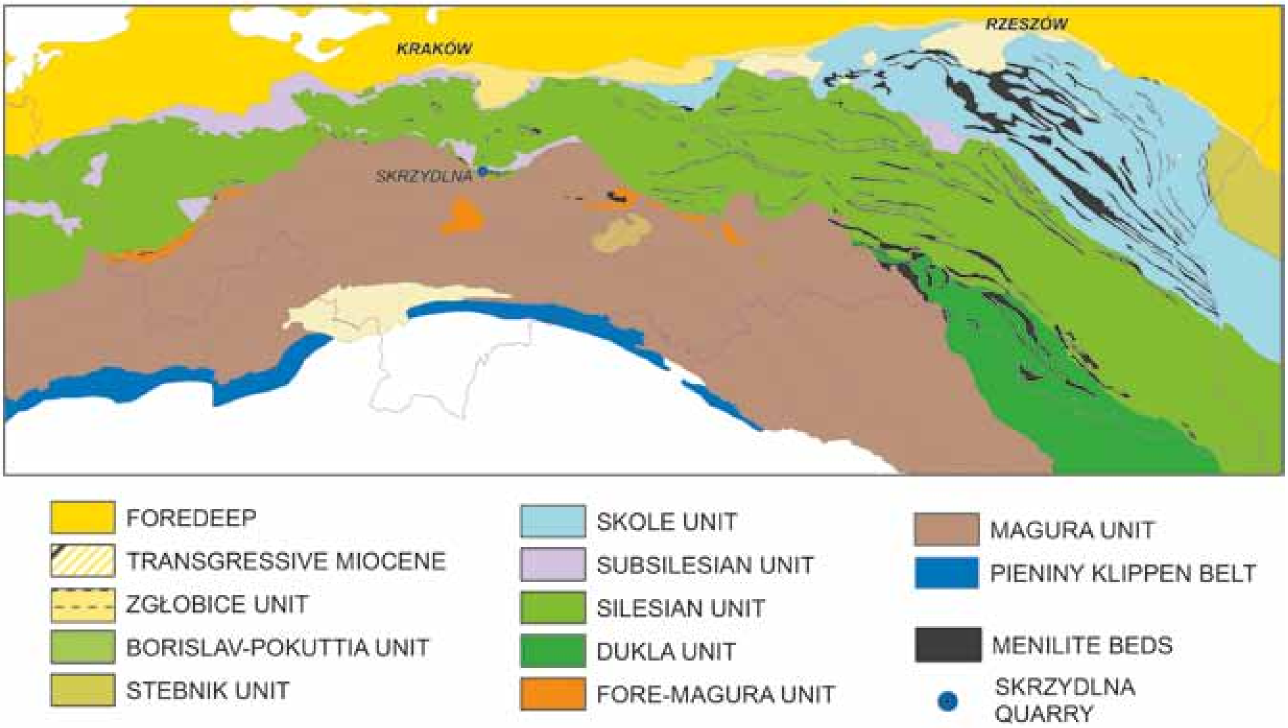

Fig. 65. Occurrence of the Menilite Formation within the Polish segment of the Carpathians

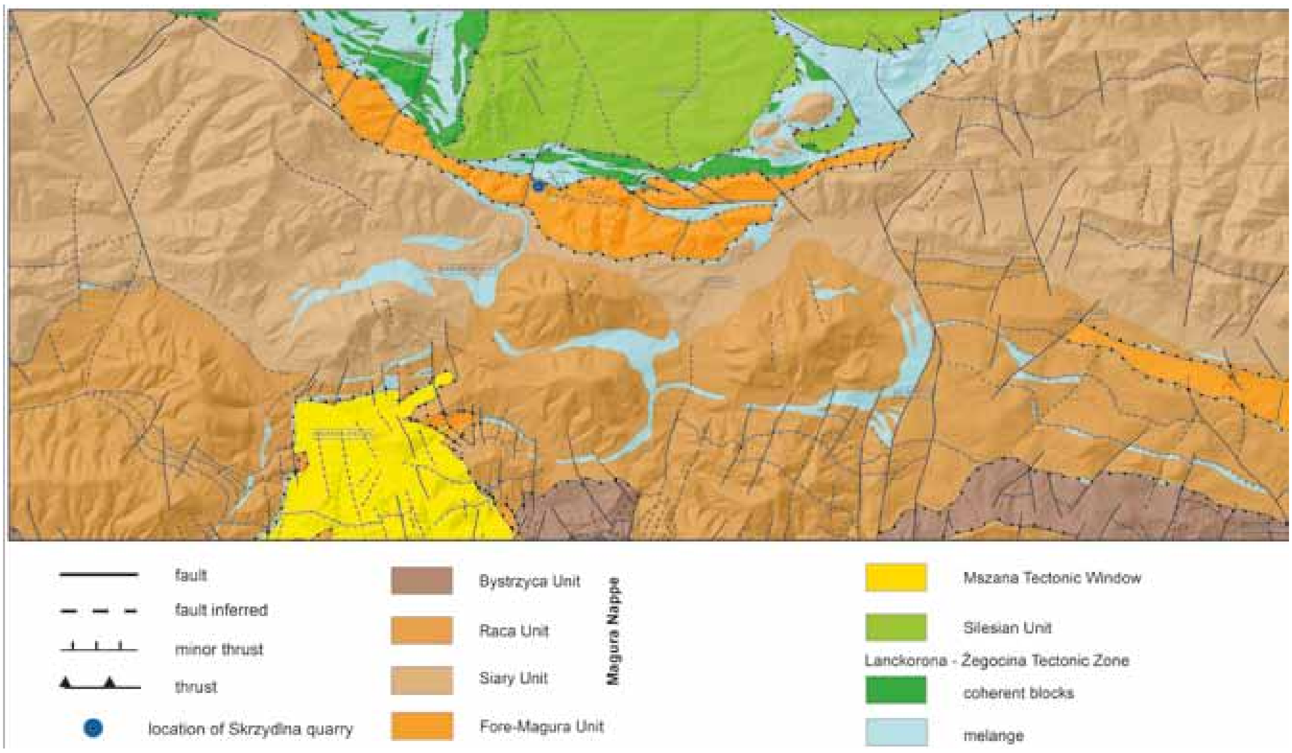

Fig. 66. Regional tectonic sketch with the location of the Skrzydlna quarry (based on Żytko *et al.*, 1989, modified)

together with covering them Krosno Beds form a narrow zone of a steeply dipping beds. This zone belongs to one of the Fore-Magura Tectonic Units (i.e. Dukla or Grybów Unit) (Fig. 66), although its geological setting has been differently interpreted (e.g. Burtan, 1974; Polak, 1999; Cieszkowski *et al.*, 2012; Jankowski & Margielewski, 2012). In the area of Skrzydlna the Fore-Magura Unit strikes along NW–SE direction right at the front of the Magura Nappe that forms here a tectonic bay. To the SE the Unit is hidden under the Magura Nappe but more to the SE its extension appears in a several elongated tectonic windows within the Nappe.

Structural development of the Fore-Magura Unit in Skrzydlna is connected with the Lanckorona–Żegocina Zone that borders the Unit from the north. It is a zone of a tectonic mélange that has been formed as a result of diapiric-type migration of the less competent formations along the strike-slip fault (Golonka *et al*., 2011) or in a multistage deformation process involving out-of-sequence thrusting, strike-slip and normal faulting (Jugowiec-Nazarkiewicz & Jankowski, 2001; Jankowski, 2012). Thus, after an emplacement of the Fore-Magura Unit in Skrzydlna as a thrust sheet in a compressional phase of forming of the Carpathian accretionary wedge, the Unit has been influence by faulting along Lanckorona-Żegocina Zone. That is the reason for belt arrangement of tectonic components of the Unit, as well as steeply dipping beds observable in a quarry.

The quarry is actively mined, therefore the amount and nature of some details that can be observed is bound to change with time; this guide describes the situation observed early in 2017. The succession exposed at Skrzydlna is a highly heterogeneous association of several facies complexes, which reflects radical changes in tectonic controls on sedimentation in the Early Oligocene (Cieszkowski, 2006; Polak, 2000). A general view of the quarry face reveals dark-coloured complex of anoxic menilite shales with a thick interbed of orange-weathering sandstone topped by whitish weathering limestone (Fig. 67). This is overlain by orange in colour conglomerate succession (olistostrome) quickly grading into a F–U turbidite sequence (Fig. 68). The beds are overturned steeply dipping to the left (N), with soles facing upwards. Their dip gradually decreases to the right, across the width of the quarry face.

Concerning the age of the succession, calcareous nannofossil assemblages from marly interbed in the lower, shaly part of the exposed succession represents nannoplankton zone NP 21 (Late Eocene). A limestone bed within turbidites in the uppermost part of the outcrop provided the youngest species, which indicate zone NP 22 (Early Oligocene) (Siemińska *et al*., 2017).

The lowermost complex of carbonaceous menilite shales and siliceous shales contains infrequent very thin intercalations of current-rippled turbidite sandstones (Bouma Tcde; Tce) and cherts, and includes several thin sandstone dykes. In the upper part, there is a prominent 9 m thick fine-to medium grained sandstone interlayer composed of massive, amalgamated intervals, devoid of mudstone interbeds, bounded by sharp, erosional lower boundary. Following another interval of the menilite facies with cherts, there is a 10 m thick pelitic limestone complex composed of thin beds, locally silicified and containing solitary cross-laminated lenses of medium to coarse-grained sandstone.

The menilite shales complex represents very low energy deposition in deep, calm, anoxic basin and interrupted by one stage of influx of sandy material deposited by high-density, rapidly decelerating turbidity currents in tectoni-

Fig. 67. General view of the NE part of the Skrzydlna quarry with lower part of the Menilite succession dominated by dark shaly deposits

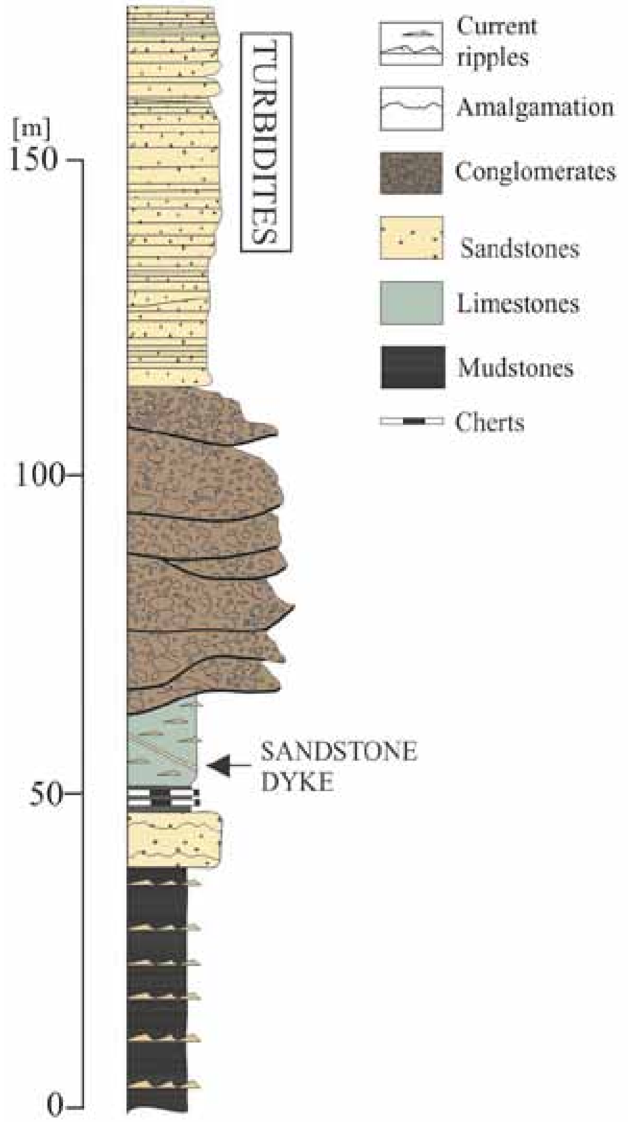

Fig. 68. Schematic section of lithological succession at Skrzydlna quarry

cally inactive basin. The sandstone dykes injected into the menilite shales are interpreted here as results of liquefaction of sandstone beds caused by seismic events. These two phenomena combined, we interpret as the earliest signals of uplift initiated in the source area, which was accompanied by seismic events. The massive sandstone interval is interpreted as a turbidite distributary channel, eroded and extended far from the source into the anoxic basin plain. The features of the limestone complex suggests hemipelagic deposition interrupted at intervals by traction currents reworking limited quantity of sandy material that formed starved current ripples. This unit is also traversed by sandstone dykes that represent three generations: two intersecting thin veins and one 30 cm thick injection and suggested to reflect (together with the massive sandstone interbed) the earliest signs of approaching uplift responsible for the origin of the overlying coarse clastic succession.

The overlying olistostrome sequence consists of amalgamated debris flow beds in which rich sandy matrix with extrabasinal granules and pebbles supports isolated blocks (olistoliths) exceeding 0.5 m in length (Fig. 69). The olistostrome base is uneven, grooved and erosionally incised into the underlying limestone complex; grooves are NW–SE oriented. Intrabasinal fragments include plastically deformed sheets of dismembered sandstone beds and clasts of black menilite shales. Among extrabasinal olistoliths a 10×2.5 m slide block of Jurassic limestone is emplaced in the middle of the complex. Two prominent layers of black shale mark breaks in intervals of high-energy sedimentation. Up the olistostrome sequence, the texture of the matrix is irregularly fining and the structure becomes more orderly.

Transition from the olistostrome to the overlying turbidite complex is marked by a few hybrid beds, each composed of turbidite sandstone overlain by linked debrite (wacke rich in mud clasts), and some other beds have features characteristic for sustained gravity flows: hyperpycnally supplied turbidites. The succession above, deposited by mainly surge-type turbidity currents, consists of three fining-upwards cycles. Two of these begin with very thick, amalgamated, massive sandstone beds deposited by high-density turbidity currents that fill channels incised by 2.5 m into the underlying strata. Bouma sequences, representing normal to diluted turbidites, appear above with Ta-e rhythms fining to Tce in the uppermost associations of thin-bedded sandstones and shales. The turbidite sandstones are quartzarenites with common admixture of coalified plant detritus. Overall, this 60 m thick turbidite succession is forms a fining-upwards sequence composed of smaller scale fining-upwards cycles.

Extreme facies contrasts between the underlying, fine-grained deposits, which settled in anoxic environment, and the succeeding olistostrome complex deposited by mass gravity flows delivering coarse debris and oxidised waters suggest rapid uplift of the source area (Wendorff *et al*., 2015). Abundance of dark shale intraclasts indicates involvement of the basin plain, the proximal part of which was transformed into slope adjacent to the elevated source. A broad lithological variety, age-affinity and shapes of extraclasts within the olistostrome imply complex structure of the source area, which included Mesozoic rocks. Topography of the source zone is suggested to have been pronounced and varying with time. Very-well rounded boulders, some prolate in shape, suggest reworking of large blocks on the abrasion platform. A huge angular block of the Jurassic-age limestone may reflect collapse of cliff face. Slide sheets of sandstone with composition different form the conglomerate matrix implies slumping and sliding of proximal slope sediments. All these features suggest that the succession originated as a result of inter-

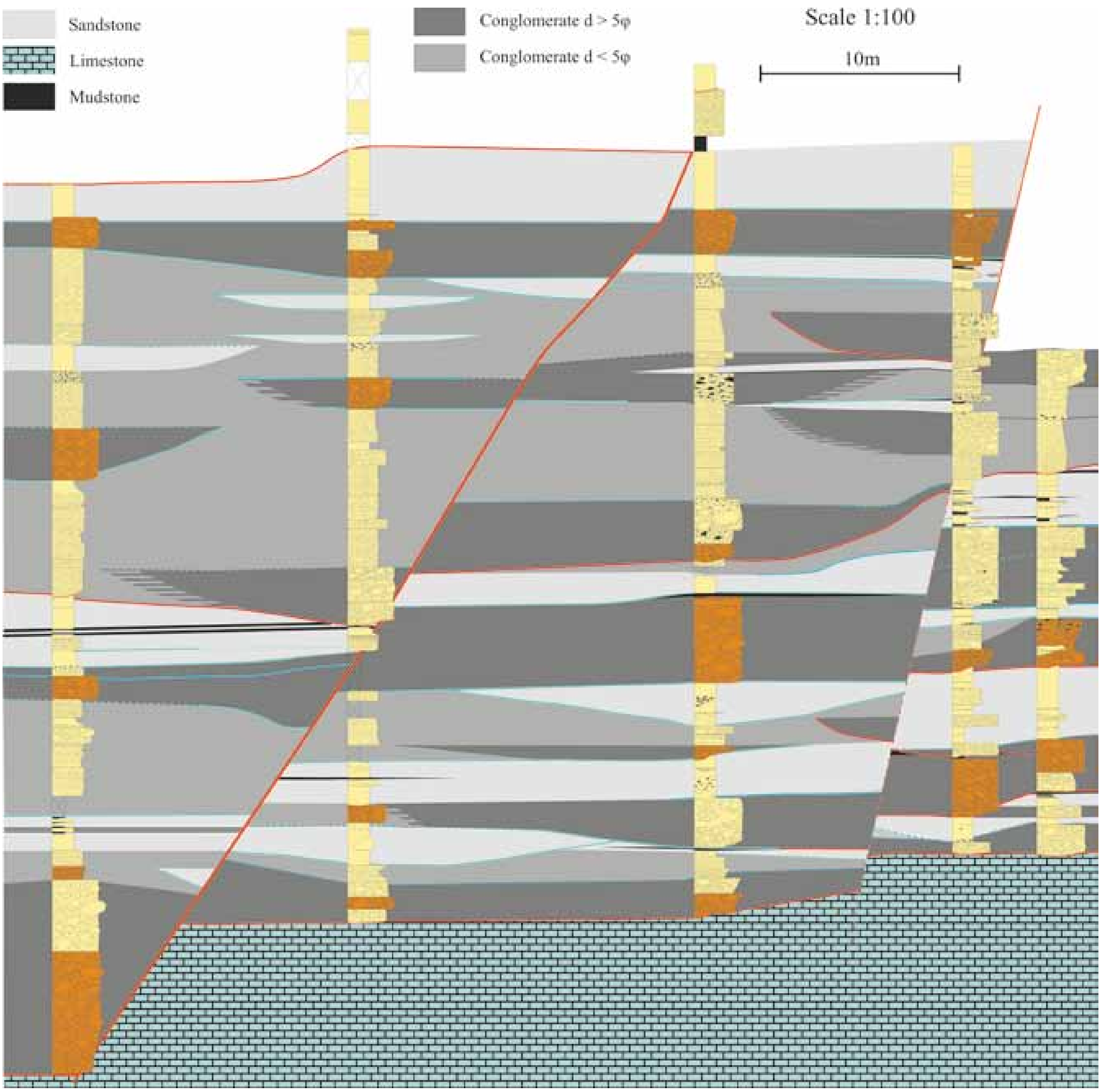

Fig. 69. Detailed sedimentological logs of Skrzydlna quarry outcrop within five exploitation levels

play of mass-transport deposition (MTD) and sedimentation out of a variety of turbidite flows.

From the basinal sedimentary infill point of view, the sudden appearance of the olistostrome within low-energy anoxic basin sediments suggests that it forms an MTD complex infilling a large erosional feature. Lateral and vertical facies distribution in the olistostrome complex reflect significant hydrodynamic changes within and between successive flows. The first-order F–U sequence composed of second-order F–U sequences within the overlying turbidite succession imply formation of a retrograding submarine fan. The age determination places the olistostrome succession in the period of climatically-controlled global Eocene-Oligocene sea-level fall. Therefore, the olistostrome genesis needs to be linked to tectonic uplift the effect of which was amplified by marine regression. Furthermore, the observed nannoplankton assemblages may reflect eustatic fluctuations of sea-level and temporally low salinity of the environment, which could be related to tectonic movements, climatic changes and gradual isolation of Paratethys in Late Eocene and during Oligocene (Švábenická *et al*., 2007).

KOBIELNIK QUARRY (OLIGOCENE MENILITE FORMATION)

›FIGS 70, 71

Jan GOLONKA & Michał KROBICKI

In the Kobielnik abandoned quarry thin-bedded flyschoidal rocks crop out. The oldest part is calcareous fine-grained sandstones and marls with numerous cherts. The main part of the quarry is occupied by the Menilite Formation, thin-bedded, brown, grey and black siliceous marls and shales intercalated with rare coarse-grained sandstones/fine-grained conglomerates (Barmuta *et al*., 2014). Large fragments of Upper Carboniferous coal are common in these beds (exotic clasts). In some places natural seeps of bituminous oil occur on joint surface of sandstones (Fig. 71). On several bedding surfaces we can observe abundant fish scales characteristic of the clupeids (Szymczyk, 1978).

SOME REMARKS ON THE MENILITE FORMATION

The Oligocene sequences commenced with dark brown bituminous shales and cherts with locally developed sandstone submarine fans or a system of fans up to several kilometers long. The upper boundary of the bituminous shales is progressively younger towards the north and shales pass gradually upwards into a sequence of micaceous, calcareous sandstones and grey marls, thinning upwards. The rocks representing this time interval belong to Lower Tejas III from the sequence stratigraphy point of view (Golonka & Kiessling, 2002; Cieszkowski *et al*., 2006). Two basins were present during these times: Magura and Krosno. The Menilite Formation dominated within the Krosno Basin (Oszczypko, 1991; Oszczypko *et al*., 2005; Cieszkowski *et al*., 2006; Ślączka *et al*., 2006).

The Menilite Formation was deposited during favourable conditions for organic-richness. The major processes responsible for such richness are: high biologic productivity, non-dilution of organic richness by clastic sedimentation, and preservation of organic matter within its depositional environment (Golonka *et al*., 2009; Kotlarczyk & Uchman, 2012, with literature cited therein).

PROCESSES FOR ENHANCED ORGANIC RICHNESS BIOLOGIC PRODUCTIVITY

(Nutrient Concentrating Processes and Settings)

A. Enhanced Nutrient Concentrations
- Terrestrial Input of Nutrients
- Coastal Upwelling
- Open Water Upwelling
- B. Evaporitic Settings
- Silled Basins
- Shelf/Platform Depressions
- Rifts on Flooded Continental Platforms
- Mid and High Latitude Deserts

C. Restricted Geographic Configuration
D. Terrestrial Kerogen Influx
E. High Latitude Effect (Oceanic Convergence).

DEPOSITIONAL PRESERVATION (of Organic Material in Depositional Environment)

A. Actively Subsiding Depocenter at Time of Deposition
B. Maintenance of Anoxia
- Positive Water Balance (Fresh-Water Influx)
- Salinity Stratification
- Thermal Stratification
- High Productivity
- Restricted Circulation (Deep, Narrow Trough or Silled Basin)
- C.Isolation Factor
- Distance of Basin from PalaeoShoreline
- (Coastlines, Shelves, Epeiric Seaways)
- Local Uplift Deflecting Drainage away from Basin (Rifts).

NON-DILUTION OF SEDIMENTED ORGANIC MATTER (Low Sedimentation Rate)

A. Proximity to Orogenic Belts during Interval of Source Rock Deposition

B. Drainage Conduits into Depocenter from Uplifted Areas

C. Rate Influence by Climatic Belts, e.g. Wet Zones.

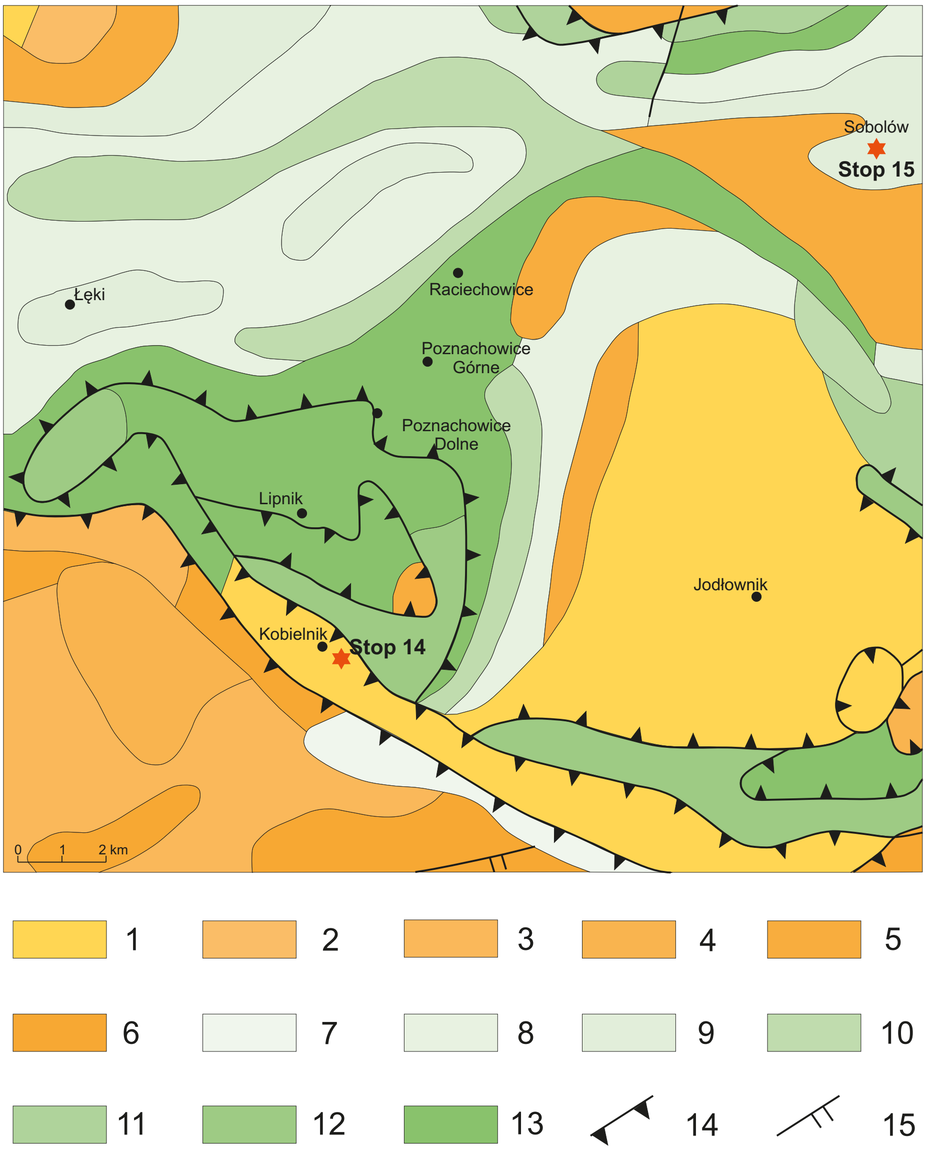

Fig. 70. Sketch of geological map of the Kobielnik and Sobolów (after Lexa *et al.*, 2000)
Explanations: 1 – Menilite and Krosno formations; 2 – Sub-Magura and Supra Magura members; 3 – Pasierbiec and Osielec Sandstones; 4 – Hieroglyphic Formation; 5 – Ciężkowice Formation; 6 – Beloveza Beds; 7 – Inoceramian Formation; 8-10 – Istebna Formation; 11 – Lhota Formation; 12-13 – Cieszyn-Hradište formations; 14 – first order overthrust lines; 15 – second order overthrust lines

Much of the organic matter produced in the oceans eventually settles into deeper water. A greater part of this material is oxidized during settling, consumed by benthic or planktonic organisms, or undergoes strong degradation in the sediments. A variable part of the primary production, however, is buried and preserved. The relative importance of these processes depends on the level of organic matter production, the depth of the water column, the rate of sedimentation, and the availability of oxidants (Golonka *et al.*, 2009).

High or increased levels of primary production of organic matter by photosynthesis within single-celled marine algae in the surface waters of the ocean supports an increased flux of organic carbon to the sea floor. This process is invoked as one of the primary controls on the origin of organic-rich sediments. The primary production of organic matter by planktonic organisms is governed by solar radiation and nutrient supply. Light attenuation restricts photosynthesis to the euphotic zone, which ranges in depth from 100-120 m in clear, open oceans to only a few m in turbid and near-shore areas. The euphotic zone is also limited to a depth of only about 20–35 m in plankton-rich, stagnant areas like the Azov and Black Seas. The euphotic zone usually has low concentrations of dissolved nutrients, because these are consumed by the phytoplankton. Deeper water, below the euphotic zone, is enriched in nutrients by bacterial degradation of organic debris (fecal pellets and dead organisms) as it sinks to the ocean floor. Sustained primary production can occur only if the nutrient supply into the euphotic zone is maintained. Nutrients can be supplied to the euphotic zone by wind-driven mixing of deeper water, by upwelling of intermediate water beneath areas of surface

Fig. 71. Abandoned quarry in the village of Kobielnik. **A** – general view of the Menilite Formation; **B**, **C** – cherty marls in the lower part of the Menilite Formation; **D** – thin bed of coarse-grained conglomerates as intercalations within the Menilite Formation; **E** – thin bed of fine-grained sandstones with Upper Carboniferous clasts of coal; **F** – natural efflux of bituminous oil on joint surface of sandstone (after Krobicki *et al*., 2012)

water divergence, and in coastal areas by lateral inflow of nutrient-rich river waters. The origin of many organicrich rocks has been attributed to upwelling. This is because upwelling zones are rich in dissolved nutrients necessary to sustain high organic productivity (Picha & Stranik, 1999; Golonka *et al*., 2009).

The Krosno Basin during Oligocene times fulfilled the conditions for organic productivity. Palaeoclimate modelling indicates wind directions favorable to open water upwelling, symmetric and circular. The computer modeling was confirmed by actual observations from the Czech Republic (Picha *et al*., 2006). According to the quoted authors the Menilite Formation was deposited in a zone of proliferation of marine life (diatoms). Nutrient supply was caused by upwelling of nutrient-rich deep waters under anoxic conditions, The siliceous phyto- and zooplankton production is accompanied by nektonic organisms, mainly fish (Jerzmańska & Kotlarczyk, 1976; Kotlarczyk *et al*., 2006; Picha *et al*., 2006; Bieńkowska-Wasiluk, 2010). The depositional environment of the formation may be compared to that which existed along the active margins of coastal California, where the Monterey Formation of Miocene age formed in the palaeoenvironmental conditions favorable to the deposition of organic-rich diatomites. Restricted geographic conditions also enhanced these processes.

The most important factor operating to preserve organic matter in sediment is the reduction or removal of oxygen from the bottom layers of water on the sea floor. Most major source rocks, with the exception of prodelta shales and some turbidites and upwelling-related source rocks, show evidence of having been deposited in anoxic or suboxic conditions known also from Carpathian basins, especially from the Krosno Basin (e.g., Picha & Stranik, 1999; Golonka & Picha, 2006; Kotlarczyk & Uchman, 2012). Even though oxidation occurs in both oxic and anoxic conditions at similar rates, anoxia at the bottom of the water column can help to preserve organic matter. This is done through restriction (due to a lack of oxygen) of deposit feeders, which represent a major catalyst to oxidation efficiency. Another way anoxic pore water aids preservation is that many organic molecules (liquid hydrocarbons, lipids, lignins) are more stable in anaerobic conditions and are resistant to anaerobic degradation (Golonka *et al*., 2009).

In general, as the rate of deposition of fine-grained sediment increases, the organic matter content of the sediment also increases. This general rule holds when the sedimentation rate is not excessive and follows from the above relationships and because rapid sedimentation decreases the time organic matter is exposed on the sea floor or in the top few m of the sediment column (Golonka *et al*., 2009). Most of the Menilite Formation represents fine-grained sedimentation. With the passage toward the sandy Krosno Formation, the organic richness disappears.

A critical control on the content of organic matter in marine sediments is the rate of organic matter accumulation on the ocean floor versus the rate of sedimentation of terrigenous and skeletal mineral matter. The organic carbon concentration in sediments is then ultimately determined by the amount that the organic matter is diluted by inorganic sediment. For rich source rocks the sedimentation rate of organic matter exceeds that of mineral matter, which is usually very low. Preservation of the record of organic productivity is also associated with very limited influx of detrital material both from the foreland and the orogenic belt to the Krosno Baisn during the sedimentation of the Manilite Formation shales (Picha *et al*., 2006). This non–dilutional factor contributed to the final organic richness of the Manilite Formation.

The actual geochemical characteristics of the Menilte Formation were outlined by Kotarba & Koltun (2006), Lewan *et al*. (2006) and Kotarba & Nagao (2008). According to these authors the Total Organic Carbon (TOC) content ranges from 0.18 to 17.25% (mean 4.48%) in the deposits occurring now within the Silesian Nappe. Type II, algal oil-prone kerogen dominates throughout the formation; types I and III are less frequent. Organic production caused by algae (mainly dinoflagelates and diatoms) is indicated by biomarkers and stable carbon isotope analyses. Some biological markers were derived from gymnosperms and angiosperms. It indicates the influx of terrigenous organic matter from ridges surrounding the Krosno Basin (Kotarba & Koltun, 2006).

FIELD TRIP ▸ STOP 15

SOBOLÓW QUARRY (ISTEBNA FORMATION)

▸FIGS 70, 72

Michał KROBICKI

In the Sobolów quarry thick-bedded sandstones crop out. They are poorly sorted, medium-grained and in some places with amalgamated beds but in intercalated grey or greenish sandy mudstones plant detritus occur. Flute casts are rare and trace fossils are represented mainly by Ophiomorpha rudis (decapod crustacean burrows – see Rajchel & Uchman, 2012).

This deposits belong to the Upper Cretaceous–Paleocene Istebna Formation (comp. Figs 23, 72) which is subdivided to four informal units: Lower Istebna Sandstone (Campanian–Maastrichtian in age), Lower Istebna Shale (the uppermost Maastrichtian–Paleocene), Upper Istebna Sandstone and Upper Istebna Shale (both Paleocene in age) (Rajchel & Uchman, 2012; Strzeboński, 2015 with literature cited therein). In this quarry the Lower Istebna Sandstone occur (the most probably its middle part – Fig. 72). According to several older investigations, summarised by Strzeboński (2015), their sedimentological features (cf. Strzeboński, 2015) and palaeoenvironmental interpretations (including trace fossils – see Rajchel & Uchman, 2012) indicate deep-sea turbiditic system. By their non-calcareous sandstones and mudstones their deposition area was presumably below calcite compensation depth (CCD), on the bathyal and/or abyssal zones , maybe even deeper than 3000 m as suggested some authors (see discussion – Rajchel & Uchman, 2012), which is a little bit controversial idea.

Contrary to previously see source rocks of the Menilite Formation, the Lower Istebna Sandstone are one of the best reservoir system in whole Outer Flysch Carpathians (Fig. 23).

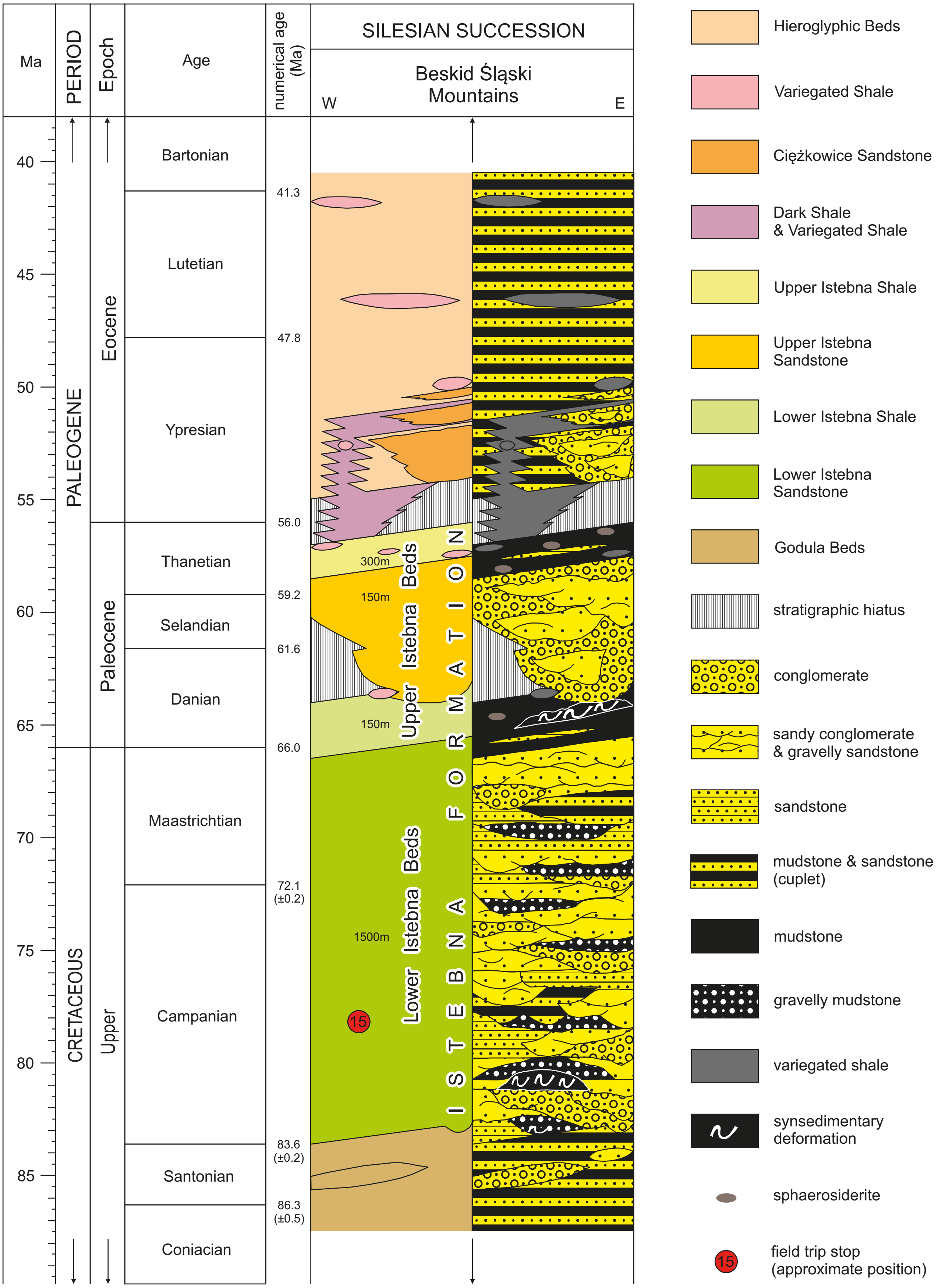

Fig. 72. Scheme of lithostratigraphic units for the Upper Cretaceous–Eocene part of the Silesian Succession (after Strzeboński, 2015; slightly modified)

BOCHNIA AND WIELICZKA SALT MINES (MIOCENE EVAPORITES)

› FIG. 72

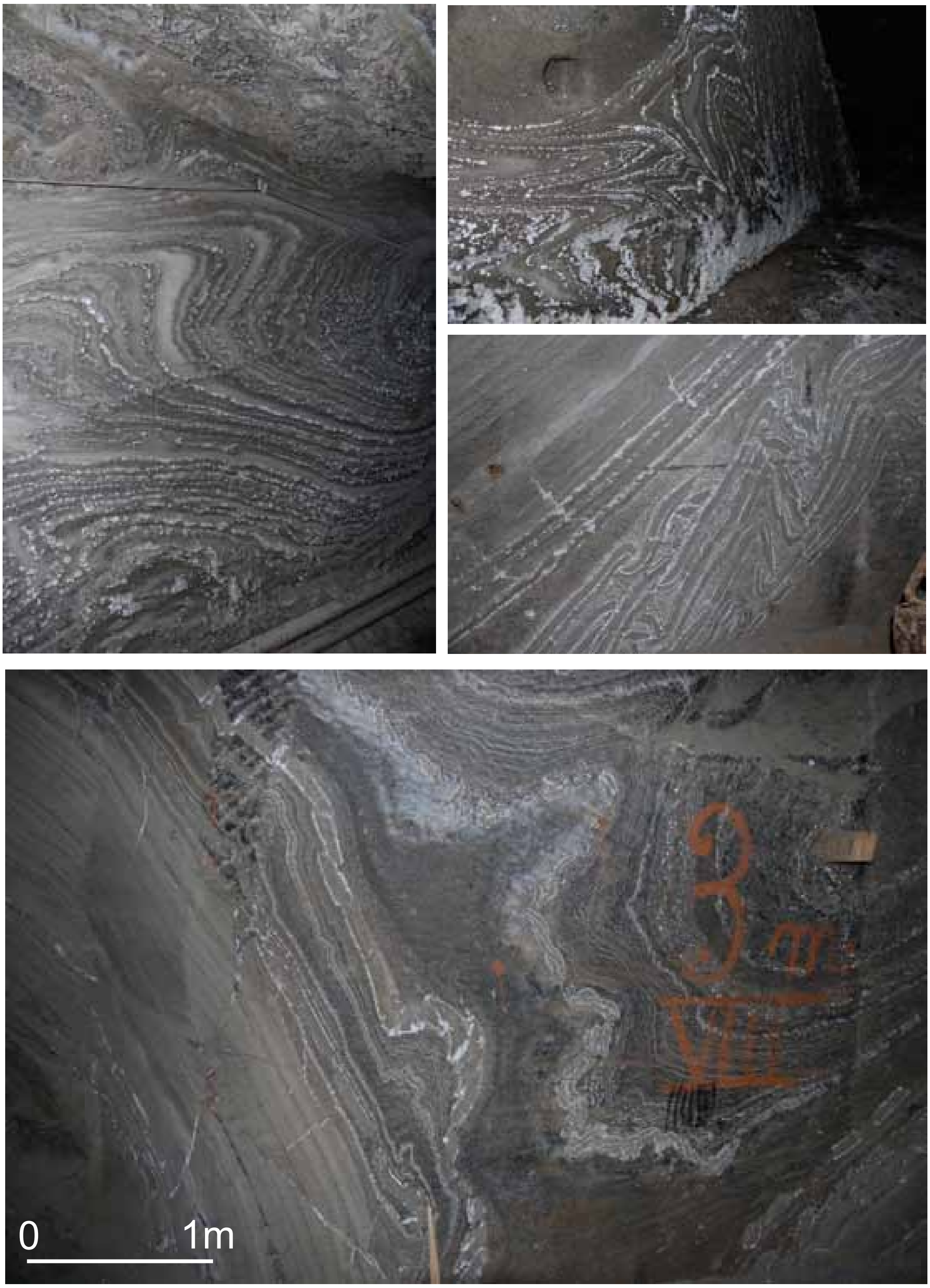

Fig. 72. Tectonic structures within folded Miocene deposits of the folded part of the Carpathian Foredeep strata (Wieliczka)

Salt Mines at Bochnia and Wieliczka

Aleksander Garlicki[1]

Marine Miocene deposits are present in southern Poland and cover the area of the entire foreland of the Polish Flysch Carpathians, extending without interruption from Silesia, through the Cracow region to the Paleozoic massif of the Kielce region to the north and to the eastern state boundary. The thickness of the discussed deposits varies greatly ranging from a few hundred metres in the western and northern parts to over 3000 m in the eastern part of the Carpathian foreland (Fig. 1).

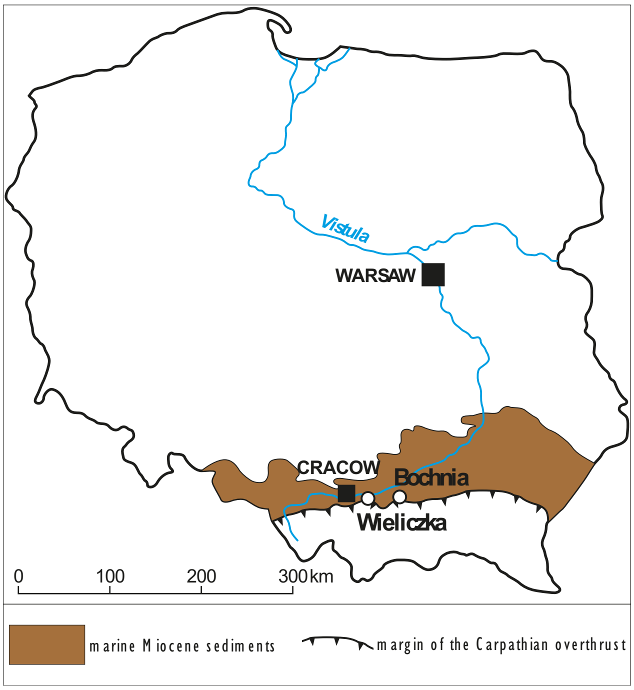

Fig. 1. Extent of marine Miocene deposits in Poland (after Garlicki, 1974; modified)

The marine Miocene consists of clays, sands and evaporites. The evaporites comprise only a small part of the vertical profile of marine sediments, but they are widespread in the sedimentary basin and thus form an important marker bed. The Miocene salt-bearing formation containing the evaporite horizon belongs to the Badenian (stage M4) and is subdivided into 3 members:
- ❑ Skawina Beds (underlying evaporites),
- ❑ Wieliczka Beds (evaporites),
- ❑ Chodenice Beds (overlying evaporites).

The Skawina Beds range in thickness from a few meters up to 150 m. They are usually represented by marly claystones and marly, clayey shales, and less frequently by siltstones with cement composed partly of evaporites. Within these beds numerous intercalations of chemically deposited dolostones occur, as well as abundant carbonized plant remains.

A transition to overlying evaporites is gradual, with slow increase of chloride and sulphate minerals. The thickness of the Wieliczka Beds range from 40 m to about 200 m, with an eastward increase in thickness. In normal profile of evaporites five cyclothems can be distinguished. The youngest cyclothem is present only in the central part of the sedimentary basin. In general, each cyclothem begins with claystones or clayey-anhydritic rocks, with abundant admixture of silt and carbonized plant fragments, followed by anhydritic claystones of nodular and banded structure, which in turn are followed by clayey-anhydritic shales that are fine-laminated and thinly banded. The uppermost part of each cyclothem consists of rock salt layers, except for some profiles of the youngest, fifth cyclothem, where the uppermost section is developed as anhydrite, instead of salt. In the Badenian basin between Wieliczka and Bochnia the intensity of chemical sedimentation was never high enough to cause the precipitation of potassium-magnesium salts.

The Chodenice Beds (overlying evaporites) are usually developed as sandy, marly, clayey shales, with numerous dolostones intercalations in the lower part and tuff intercalations in the upper part. The total thickness of the Chodenice Beds varies from 100 m to about 1000 m.

The youngest Badenian formation (Grabowiec Beds) is developed as sands and sandstones. These strata are fairly undisturbed and in the area adjacent from the north to the salt mines at Bochnia and Wieliczka overlie discordantly clayey rocks of the Chodenice Beds (Figs. 2, 9).

Within the Wieliczka Beds, three facies have been distinguished:
- ❑ carbonate — littoral facies comprising organic limestones, mixed carbonate and detrital rocks;
- ❑ sulphate facies, comprising anhydrite-gypsum and sulfur-bearing deposits;
- ❑ chloride facies, containing rock salt with anhydrite and clay-anhydrite rocks.

The areal extent of the chloride facies is smaller than the other one. In the Upper Silesia the chloride facies covers an oval shaped area and is surrounded by the sulphate facies. Farther east, along the Carpathian boundary, chloride facies extend from Wieliczka to Tarnów and occur in the vicinity of Pilzno and Przemyśl (Garlicki, 1974, 1979).

In the Badenian salt-bearing formation of the Carpathian foreland two main units can be distinguished,

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namely an autochthonous unit and overthrust one (the latter sometimes is called an allochthonous element). In the substratum of both Miocene units, the Mesozoic strata developed as marls and limestones have been stated in the western part of the Wieliczka deposit and north of the Bochnia mine. During the Late Miocene, strata in the allochthon were folded in front of the Carpathian nappes and thrust from the south over the autochthon. These tectonic features can be observed in numerous cross-sections through the marginal zone of the Miocene in front of the Carpathian thrust belt in Poland. As a result of these intense disturbances in the overthrust unit, one can observe recumbent folds passing into imbrications, strong deformations of salt layers resulting in increase and decrease of their thickness, and even coarse breccia composed of salt clay with blocks of rock salt. The final stage of these disturbances was an uplift of folded strata to the surface. This was the origin of the Miocene salt deposits, among others at Bochnia and Wieliczka (Figs. 2, 9).

The Miocene salt deposits were the cradle of the Polish mining industry. In the vicinity of Bochnia and Wieliczka the first traces of salt production go back to the Neolithic period (ca 3500 BC). In the earliest times salt was obtained there by roasting and boiling to dryness salt brine obtained from the natural surface salt springs and brine wells. Boiling kettles from the tenth and eleventh centuries were discovered during recent archaeological excavations. Written documents show that as early as the eleventh century there was a great brine producing centre at Wieliczka (called *Magnum Sal* in Latin). Rock salt, however, was discovered several years later. Surface salt springs and shallow wells must have become depleted quickly. In the course of deepening the brine wells rock salt was discovered in Bochnia about 1248 and at Wieliczka in the second half of the 13th century. By the 13th century rock salt mines operated at Bochnia and Wieliczka (Jodłowski et al., 1988). These salt mines were the King's property and were administered by managers designated by royal orders. During medieval history of Poland, the salt mines were the main source of income to the royal treasury. On the other hand both mining towns were granted special royal privileges, which established their leading position in the country for several hundreds years. In 1978 the Wieliczka Salt Mine was included by UNESCO in the first list of the World Cultural and Natural Heritage. In 1994 it was recognized by state's authorities as a Polish historical monument. Few years later in 2000, the Bochnia Salt Mine was also recognized as a historical monument of Poland.

It is worth mentioning that Bochnia and Wieliczka Salt Mines (from the Middle Ages known as *the Cracow Saltworks*) are the oldest still operating factories in Poland. Of course, their main tasks nowadays are quite different than those carried out during passed 800 years.

Bochnia

Geological setting of the Bochnia salt deposit has been presented in several publications (Poborski, 1952; Garlicki, 1968). The best description accompanied by numerous cross-sections and maps one can find in the monograph by Poborski (1952).

In the vicinity of Bochnia, in front of the Carpathians, two Miocene anticlines occur (Fig. 2). The main, northern fold is called the Bochnia anticline, whereas the parallel, southern is called the Uzbornia anticline. Cores of these anticlines are built up of flysch sediments which indicate that the evaporites of Bochnia were deposited upon flysch sediments. The salt deposit is situated in an almost vertical northern limb of the Bochnia anticline, about 40 km east of Cracow. The length of the deposit is about 7 km; the width varies from some dozen metres to 200 m. The deepest part of the mine is about 460 m. Westward continuation of the Bochnia salt deposit, about 6 km long, has been discovered in the years 1956–1968 in localities: Łapczyca, Moszczenica, Siedlec and Łężkowice. The internal structure of these deposits reveals some features similar to the Bochnia deposit. In vertical profile four cyclothems have been distinguished, comparable with those in other salt deposits.

Part of the deposit recognized at Łężkowice used to be exploited between years 1968–1990 using the method of leaching salt beds with water through boreholes from the surface. Generally, production of salt in the Bochnia mine was finished in 1990.

The inner tectonics of the Bochnia deposits reveals a unique accumulation of steep folds of high amplitude. Such intense folding was connected with a partial tectonic squeezing of certain limbs of folds resulting in plastic translocation, shearing of salt off

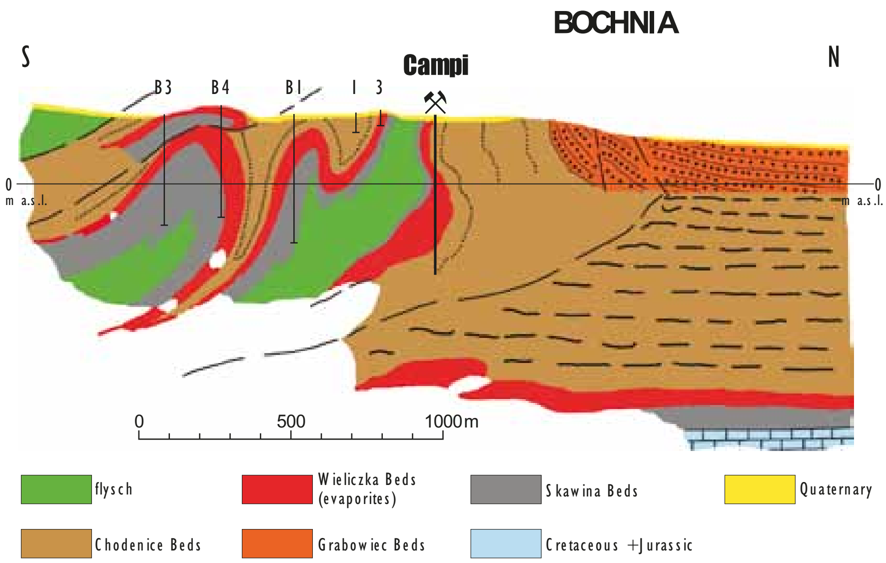

Fig. 2. Cross-section through the shaft Campi (after Garlicki, 1968; modified)

Fig. 3. Shaft Campi. Figs. 3, 5 photo by K. Stompór

Fig. 5. St. Kinga chapel

more rigid rock layers, accumulation of salt in fold bends, and finally piercing of fold bends by the salt mass. In some cases bigger accumulations of salt took place due to thinning out of the barren beds (separating the salt beds in an undisturbed profile). The richest part of the deposit mainly composed of middle salts, was situated at the depth about 200–450 m below surface, generally between the first and tenth mine levels.

Within lithostratigraphic profile of the deposit following rocks can be distinguished (from the bottom):

- ❏ basal anhydrite
- ❏ lower zuber
- ❏ southern salts
- ❏ shaley marly claystone
- ❏ upper zuber
- ❏ anhydritic claystone
- ❏ argillaceous anhydritic shale with beds of crystal salt
- ❏ middle salts
- ❏ anhydritic claystone
- ❏ northern salts
- ❏ shaley marly claystone
- ❏ top anhydrite

The primary thickness of these sediments was about 70 m and they may be well correlated with five cyclothems distinguished in the Wieliczka salt deposit (Garlicki, 1968, 1979).

An access to the mine took place through several shafts: Floris, Gazaris, Regis, Sutoris, Campi, Trinitatis. Currently operate two main output shafts: Sutoris and Campi (Fig. 3). Both shafts are connected by several mine working levels. In the western part of the mine there is also an intake shaft Trinitatis.

In the Bochnia Salt Mine there are numerous artifacts of old mining from the 17^{th} to 19^{th} centuries. Of particular interest are chapels, old chambers and galleries carved out in salt which are situated in the upper part of the mine. Some wooden tools and appliances from Bochnia have become a part of per-

Fig. 4. Stalactites in a cross-cut section. Level Lobkowicz. Figs. 4, 6 photo of the Bochnia Salt Mine

Fig. 6. Ważyn chamber

Fig. 7. Wooden slide and steps heading into Ważyn chamber. Figs. 7, 8 photo of the Bochnia Salt Mine

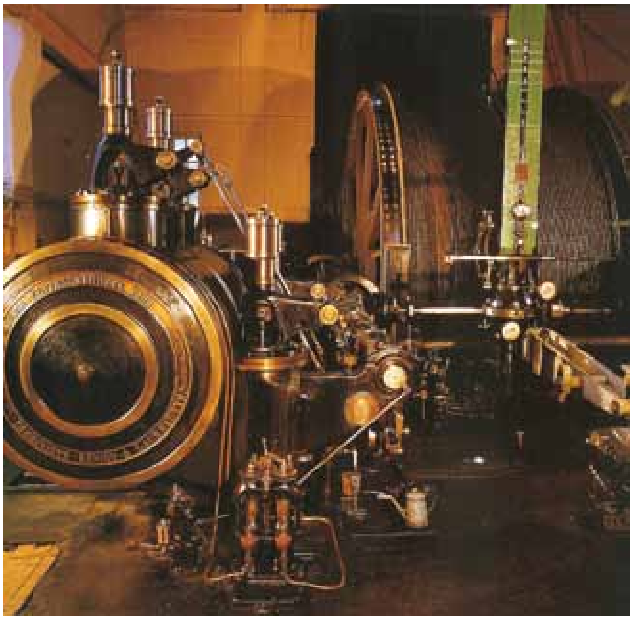

Fig. 8. Old steam winding engine of the shaft Campi

manent exhibition in the underground museum in Wieliczka. Since 1993 the touring route in the Bochnia salt mine was arranged between two main shafts: Sutoris and Campi. These two shafts are connected by longitudinal gallery of the first mine level called August. The level founded at the beginning of the 18th century is situated at the depth of 212 m below the surface, reaching the length of 3 km. Section open to tourists, over 2 km long, leads through old workings: galleries, crosscuts and internal shafts in the mine.

Some of these workings are protected by the wooden lining and filled up with barren rocks. To the touring route have been included old mine workings of the neighboring parallel levels (Wernier, Lobkowicz and Sienkiewicz). The main objects of the touring route are chambers: Wernier, Christian, Stanetti, Rabsztyn, St. Kinga and Ważyn) as well as galleries and cross-cuts, mine staircases, wooden hauling gears and other tools, sculptures and paintings. In old cross-cuts natural geological forms have been preserved (Fig. 4).

On the level August, after 1747 was built the largest chapel in the Bochnia mine and dedicated to St. Kinga (Fig. 5). It is about 31 m long, 21 m wide; an average height is 5 to 6 m. The chapel is furnished with two altars and several sculptures carved both in wood and rock salt. On the walls and ceiling very distinct features of salt tectonics are exposed.

On the level Sienkiewicz (about 250 m below the surface) after 1697 in several stages the Ważyn chamber was developed (Fig. 6). In recent years this chamber was remodeled with the use of modern mining machinery; it is about 300 m long and adapted to the functions of a sanatorium, recreation, sport and social events. The subterranean sanatorium with a unique microclimate is an ideal place for treatment of respiratory diseases, bronchial asthma, and allergies. The sanatorium provides to the visitors overnight stays and such medical facilities as equipment for inhalations. A stable temperature inside the Ważyn chamber is 14 to 16 °C. Direct access from the August level to the Ważyn chamber provides connecting shaft equipped with a lift for visitors. Another access to the Ważyn chamber is also possible by dip heading furnished with steps and wooden slide for children (Fig. 7).

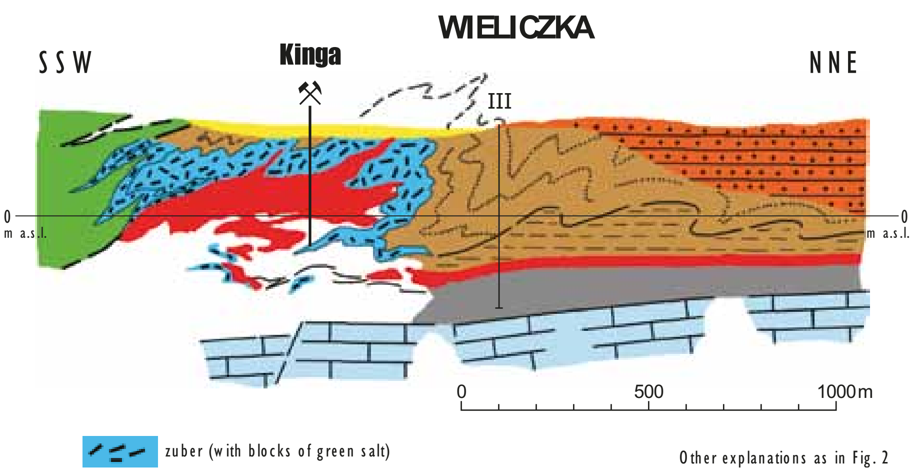

Fig. 9. Cross-section through the shaft Kinga (after Garlicki, 1968; modified)

Fig. 10. Shaft Daniłowicz. Figs. 10, 11 photo by A. Grzybowski

A special attraction furnished for the visitors is underground trip along the August level between two shafts (Sutoris and Campi). The train on rails is hauled by the battery-powered locomotive. It is said that it is the only case in the world with the train passing through the central part of the church (it means through middle of the St. Kinga chapel).

On the surface, close to the shaft Campi there is exposed steam engine for many years operating in the shaft. It was manufactured in 1909 in Silesia and has been preserved in excellent technical condition up to now (Fig. 8).

During last few years the number of tourists visiting the Bochnia mine has been increasing gradually. In 2007 the number of visitors already exceeded 130 000 persons.

Wieliczka

Geological structure of the Wieliczka salt mine has been already described in numerous papers (Gaweł, 1962; Poborski & Skoczylas-Ciszewska, 1963; Garlicki, 1974, 1968; Kolasa & Ślączka, 1985; Ślączka & Kolasa, 1997; Bukowski, 1997). The most comprehensive publication is that of Gaweł (1962), in which numerous detailed cross-sections have been presented. The salt deposit at Wieliczka situated 13 km south of Cracow, is 1 km wide, about 6 km long, over 425 m deep, and consists of two essential parts (Fig. 9). The upper one is developed in the form of coarse breccia (*boulder deposit*) composed mainly of salty clays (*zuber)*, with blocks of coarse-grained salt, called green salt. These blocks of irregular shape and various thicknesses in some places reach an extension of more than 150 m. Some authors (Kolasa & Ślączka, 1985; Ślączka & Kolasa, 1997) have stated that the boulder deposit was formed due to submarine gravitational slumps and flows (olistostrome), developed on a tectonically active basin slope. This part is supposed to be the facies equivalent of the lower one. The lower part of the deposit is developed as a complex of salt layers strongly folded, deformed and thrust over one another, being usually called *stratified* or *bedded* part of the deposit. Within the lower part of the Wieliczka salt deposit three main anticlines have been distinguished (southern, central and northern). They are elongated northwards, reduced in thickness and form a kind of scales (imbricated folds) piercing up into the boulder deposit. East of the Daniłowicz shaft the northern scale is transformed into a domal structure of the Crystal Caves. The salt layers of the stratified deposit are interbedded with anhydrite and anhydritic clays. In this part of deposit occur considerable complexes of the shaft salt and spiza salts, which used to be very important subjects of exploitation up to the 20th century. From the south, flysch deposits in the form of tongue-shaped wedges are squeezed into the inner part of the salt deposit (Garlicki, 1974). Western part of the Wieliczka salt deposit called Barycz, about 1 km wide and 2 km long is separated from the underground mine by 200 m wide safety pillar. During 1923–1998 there took place exploitation of salt by leaching salt through boreholes carried out from the surface, whereas the Wieliczka salt mine stopped an output of salt in 1996.

Reconstructed normal stratigraphic profile of the stratified part of the deposit is as follows (from the bottom):

- anhydritic claystones and siltstones,
- the oldest salts (varigrained with admixture and intercalations of clay, silt and sand);
- salty sandstones and siltstones, partly conglomeratic;
- set of green layered salts (numbered I–V), intercalated with anhydritic claystones;
- shaft salt (coarse-grained salt devoid of mineral admixtures but containing traces of gaseous hydrocarbons);
- lower spiza salts (or spizum salts);
- central intercalation (anhydritic claystone);
- upper spiza salts (both lower and upper spiza salt are coarse-grained, banded, with intercalations of sandy anhydritic clays);
- claystones, siltstones and sandstones, anhydritic in the upper part.

The primary thickness of this sequence was about 70 m. The lowermost part of the profile (including the oldest salts) corresponds to the first and second cyclothems recognized within autochthonous unit of evaporites. Green layered salts, shaft salt and lower spiza salts belong to the third cyclothem, whereas the fourth cyclothem contains the upper spiza salts. At the

Fig. 11. Horse-powered treadmill. Polish type

Fig. 12. St Kinga chapel. Figs. 12, 13, 14 photo by A. Grzybowski

bottom of evaporites and within the lower part of the green layered salts three thin tuff intercalations have been distinguished, being the important marker beds (Garlicki & Wiewiórka, 1981).

Historical salt mine at Wieliczka consists of two parts:

- ❑ touristic route, comprising galleries, chambers, and other mine workings occurring from the first to third level (from 64 m to 135 m below the surface);
- ❑ exhibition of underground museum located on the third level.

The first part is managed and subordinated to the governmental join-stock company Salt Mine Wieliczka, the second is administered by the Cracow Saltworks Museum, representing Ministry of Culture and National Heritage.

Fig. 13. Saurau chamber

The Wieliczka Mine has several shafts (Daniłowicz, Kinga, Kościuszko, Regis, Górsko) and mine working levels, from the first situated at 64 m below surface to the last one at the depth of 327 m. In the upper part of the mine the old exploited area comprises a network of chambers and galleries the total length of which exceeds 200 km. In this part of the mine an original underground touring route and museum have been arranged. The touring route (about 2 km long) starts at main touristic shaft Daniłowicz (Fig. 10) and leads to the depth of 135 m (third mine level) passing through galleries staircases, artificial lakes, chambers, and chapels carved out of salt blocks during the 17th to 20th centuries. The underground museum has several sections showing both the history of salt mining in Poland and geology of the Wieliczka salt deposit.

The large exhibition of Muzeum Żup Krakowskich (the Cracow Saltworks Museum) situated in chambers: Russeger, Maria Teresa, Miejska, Kraj, Karol and Modena on the third level of the mine, presents various methods of salt extraction, safety systems, outstanding works of sacred art, geology of salt and sulfur deposits in Poland, results of archaeological excavations in the vicinity of Wieliczka, history of salt industry in southern Poland, old mining mechanisms and equipment for transport and many others.

Fig. 14. Wall of the Crystal Cave

Among mining equipment of special interest are various types of horse-powered treadmill (Fig. 11) and a wide scope of tools used in salt mining.

In geological section a great attraction are Permian color salts, specimens of big crystals and numerous fossils well preserved inside the rock salt.

One of the largest and most beautiful is the chapel of St. Kinga founded at the depth of 101 m (Fig. 12). It is 54 m long, 18 m wide and 12 m high. The walls of the chapel are decorated with sculptures and bas-reliefs representing Biblical scenes (e.g. *The Flight to Egypt*, *Herod's Sentence*, *The Slaughter of the*

Fig. 15. Group of geologists inspecting underground exposures. Spiza salts, chamber Wałczyn, first level. Photo by S. Klimowski

Innocents, *The Miracle at Cana of Galilee*, *The Last Supper*).

On the main altar of the chapel there is exposed monumental sculpture of St. Kinga carved out in translucent salt and the background of the figure is made of pure large salt crystals (taken from the Crystal Caves). The floor of St. Kinga chapel is one uniform flat plane of salt, with secondary carved channels in form of separate tiles. Interior of the chapel is illuminated by chandeliers made of pure crystals of salt.

Besides St. Kinga chapel there are many famous chambers and namely: St. Anthony, Drozdowice, Michałowice, Weimar, Warszawa, Saurau, Pieskowa Skała, J. Piłsudski, E. Barącz, S. Staszic, N. Copernicus. Chamber Saurau is an example of conservation works using modern technologies in order to preserve this object of old mining for the next generations (Fig. 13). Some chambers have been filled up with saturated brine and equipped with special illumination (light and sound). Among the largest underground workings of the Wieliczka mine Warszawa chamber is worth mentioning, situated at the depth of 125 m. It is intended for cultural and sport events with accommodation for about 1000 persons.

At the depth of ca 80 m there occur unique Crystal Caves. During the 18th and 19th century intense mining works were carried out in order to explore new salt reserves in the eastern part of the mine. Special attention was paid to seepages and outflows of fully saturated brine. Protecting works against water influx hazard resulted in the discovery of complex system of fissures and cavities in the Upper and Lower Crystal Caves. At the end of the 19th century the chamber Baum-Schwind was discovered. There occurred very pure and large crystals of halite (Fig. 14). Their edges sometimes reach about 50 cm (Alexandrowicz, 2000).

For the youngest visitors there have been arranged sets of sculptures carved in rock salt presenting dwarfs, legends, fables, famous persons etc.

Over 50 years ago, some 200 m underground, a sanatorium for patients with bronchial asthma and allergic diseases has been established. The microclimate in the chambers left after extractions of salt produces excellent effects and sensational curing results.

Since the Middle Ages the salt mine in Wieliczka has been visited by many famous citizens of Poland as well as foreigners. Their signatures are exposed in many documents and Visitors' Books.

Each year over one million tourists visit the Wieliczka Mine, which is one of Poland's top touristic attractions.

An integral part of the Cracow Saltworks Museum is the Saltworks Castle (called *Saline Castle*), erected on the surface, which used to be the historic seat of the Saline Authorities from the 13th to 20th century. Currently it houses the Cracow Saltworks Museum with its offices, exhibition rooms, collections and library. There are permanent and temporary exhibitions in the Castle. There commonly take place such special events as conferences, workshops, shows, games and concerts. The Saltworks Museum is also a scientific institute taking care of underground exhibitions, carrying out research and editorial activities.

Final remarks

In the underground workings of both salt mines students of the University of Science and Technology (AGH) carry on their field exercises in surveying, mining, geology and underground mapping. Some big chambers in the Bochnia and Wieliczka salt mines are the places of sport tournaments (e.g. championships in basketball, tennis, handball, volleyball, football) or such scientific and social events as conferences, seminars, symposia, meetings, classes for pupils of schools, temporary exhibitions and fairs. For these purposes some chambers have been specially equipped with facilities (conference rooms, projectors, screens, panel discussion rooms, cafeterias and even banquet halls). Besides regular group of tourists, sometimes special trips for those connected with mining professions are guided to selected places of interest (Fig. 15).

In order to preserve these mines for the future generations, continuous long lasting and very expensive maintenance works are required. Unfortunately, due to these works many interesting geological exposures must be cover up by the wooden lining.

Both chapels of St. Kinga at Bochnia and Wieliczka offer unforgettable experiences to visitors, but they also serve religious purposes. Every year solemn masses are celebrated in these churches, at least on Christmas Eve, on the St. Kinga's Day (24th July) and on St. Barbara's Day (4th December).

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