The General Geology of the Outer Carpathians, Poland, Slovakia, and Ukraine

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Geology and Hydrocarbon Resources of the Outer Carpathians, Poland, Slovakia, and Ukraine: General Geology

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ABSTRACT

The purpose of this chapter is to provide the general overview of the stratigraphy and tectonics of the Polish, Ukrainian, and adjacent parts of the Slovakian Outer Carpathians. The Polish and Ukrainian Outer Carpathians form the north and northeastern part of the Carpathians that expand from the Olza River on the Polish–Czech border to the Ukrainian–Romanian border. Traditionally, the Northern Carpathians are subdivided into an older range, known as the Inner Carpathians, and the younger ones, known as the Outer Carpathians. These ranges are separated by a narrow, strongly tectonized belt, the Pieniny Klippen Belt. The Outer Carpathians are made up of a stack of nappes and thrust sheets showing a different lithostratigraphy and tectonic structures. Generally, each Outer Carpathian nappe represented separate or partly separate sedimentary subbasin. In these subbasins, enormous continuous sequence of flysch-type sediments was deposited; their thickness locally exceeds 6 km (3.7 mi). The sedimentation spanned between the Late Jurassic and early Miocene. During the folding and overthrusting, sedimentary sequences were uprooted, and generally, only sediments from the central parts of basins are preserved.

The Outer Carpathian nappes are overthrust on each other and on the North European platform and its Miocene–Paleocene cover. In the western part, overthrust plane is relatively flat and becomes more and more steep eastward. Boreholes and seismic data indicate a minimal distance of the overthrust of 60–80 km (37–50 mi).

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The evolution of the Northern Outer Carpathian Flysch basins shows several tectonostratigraphic stages. The first period (Early Jurassic–Kimmeridgian) began from the incipient stage of rifting and formation of local basins. The next stage (Tithonian–Early Cretaceous) is characterized by rapid subsidence of local basins where calcareous flysch sedimentation started. The third period (Late Cretaceous–early Miocene) is characterized by compression movements, appearance of intensive turbiditic sedimentation, and increased rate of subsidence in the basins.

INTRODUCTION

The purpose of this chapter is to provide the general overview of the stratigraphy and tectonics of the Outer Carpathians in Poland, Ukraine, and adjacent Slovakia. The details concerning petroleum geology will be provided by other chapters in this Memoir by employees of the Polish Oil and Gas Company and the Ukrainian State Geological Research teams. These chapters include as follows:

- *Hydrocarbon resources of the Polish Outer Carpathians—reservoir parameters, trap types, and selected hydrocarbon fields: A stratigraphic review* by Dziadzio et al. (2006b, chapter 8)
- *Petroleum geology of the Boryslav–Pokuttya zone, the Ukrainian Carpathians* by Popadyuk et al. (2006, chapter 13)
- *The geology of the Weglowka oil field, Subsilesian unit, Polish Outer Carpathians* by Dziadzio (2006, chapter 14)
- *The Ciezkowice sandstone: Examples of basin-floor fan-stacking patterns from the main (upper Paleocene to Eocene) reservoir in the Polish Carpathians* by Dziadzio et al. (2006a, chapter 15)
- *Reconstruction of petroleum systems based on integrated geochemical and geological investigation: Selected examples from Middle, Outer Polish Carpathians* by Matyasik and Dziadzio, (2006, chapter 16)

A special chapter also deals with the geochemistry of the Carpathian source rocks: *The origin and habitat of hydrocarbons of the Polish and Ukrainian parts of the Carpathian province* by Kotarba and Koltun (2006, chapter 11).

THE POSITION OF THE OUTER CARPATHIANS IN THE ALPINE–CARPATHIAN FOLD AND THRUST BELT

The Polish and Ukrainian Outer Carpathians form the northeastern part of the great arc of mountains, which stretch more than 1300 km (807 mi) from the Vienna Forest to the Iron Gate on the Danube (Figure 1). 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.

Inner Carpathian terrane: A 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 (Figure 2) that are the prolongation of the lower, middle, and upper Austroalpine nappes, respectively. The nappes and the Hercynian basement are unconformably covered by middle Eocene–Oligocene flysch and early–middle Miocene marine and terrestrial (continental) molasses. Another terrane with the Hercynian basement, known as the Tisza–Dacia, is amalgamated with the Inner Carpathian terrane. According to Golonka et al. (2000), the Inner Carpathians, Eastern Alps, and Tisza–Dacia form Alcapa superterrane; according to, e.g., Kovac et al. (1998), Alcapa and Tisza constitute different plates.

North European platform: Large continental plate amalgamated during the Precambrian–Paleozoic. Proterozoic, Vendian (Cadomian), early Paleozoic (Caledonian), and late Paleozoic (Hercynian) fragments could be distinguished in the folded and metamorphosed basement of this plate. Beneath the Outer Carpathians, the sedimentary cover consists of the autochthonous upper Paleozoic, Mesozoic, and Cenozoic sequences covered by the allochthonous Jurassic–Neogene rocks (see Oszczypko et al., 2006).

These allochthonous rocks are uprooted and overthrust onto the southern part of the North European platform at a distance of at least 60–100 km (37–62 mi) (Ksiazkiewicz, 1977; Oszczypko and Slaczka, 1985). They form stacks of nappes and thrust sheets arranged in several tectonic units. In Poland, these allochthonous, mainly

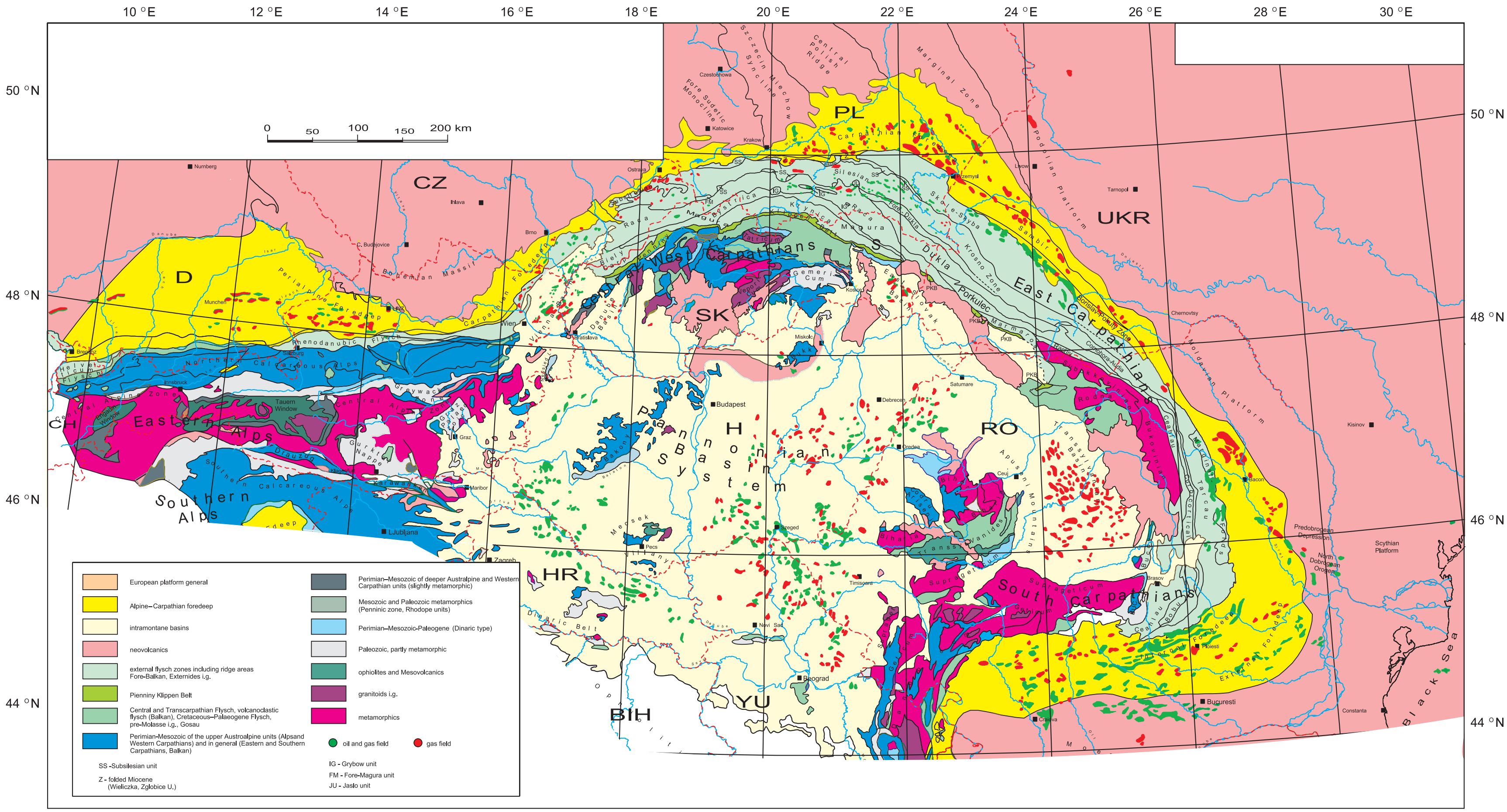

Figure 1. General overview and distribution of oil and gas in the Circum Carpathian Region of Central Europe.

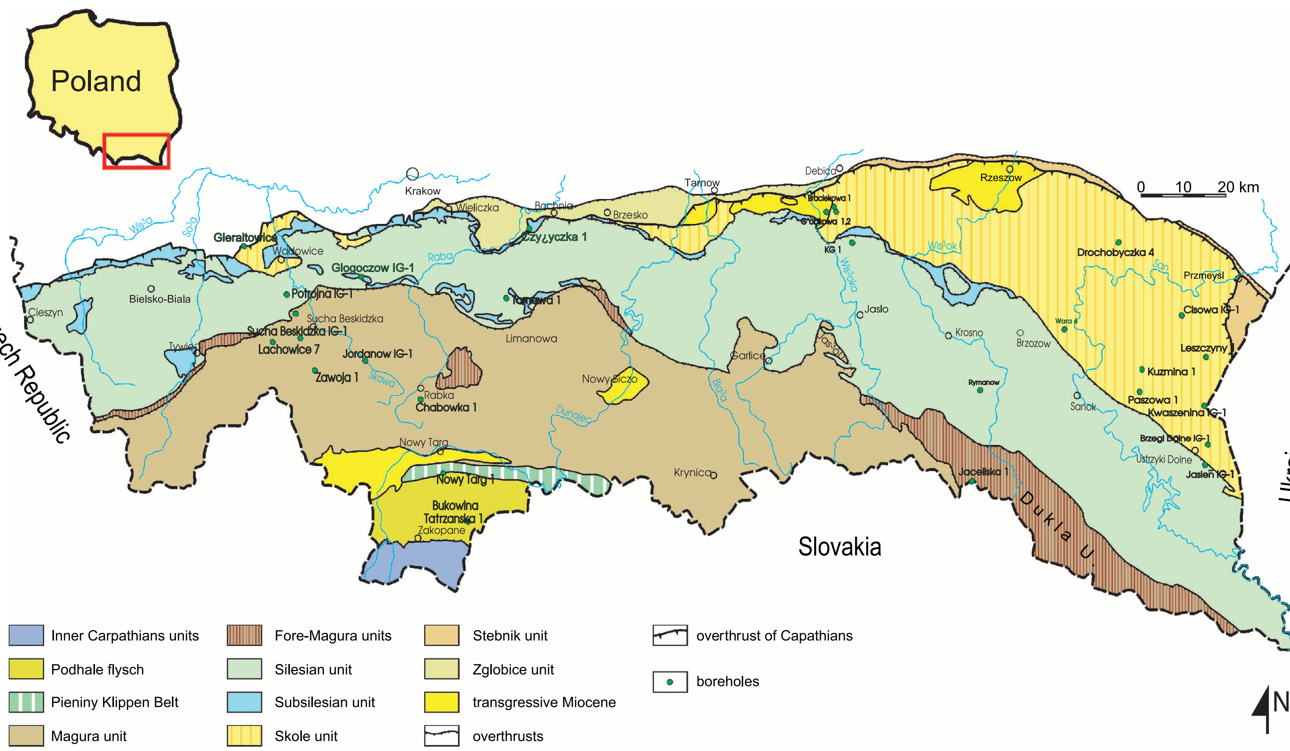

Figure 2. Tectonic scheme of the Polish Carpathians.

flysch units are being regarded as the Carpathian Flysch. Along the frontal Carpathian thrust, a narrow zone of folded Miocene deposits was developed (Figures 2, 3): the Zglobice unit in the Polish Carpathians, the Sambir nappe of the Ukrainian part of the Carpathians, and the Subcarpathian unit in Romania.

The Penninic realm is a part of the Alpine Tethys (e.g., Birkenmajer, 1986; Sandulescu, 1988; Oszczypko, 1992; Plasienka, 1999, 2000; Stampfli, 2001; Golonka, 2004; Golonka et al., 2006) that developed as a basin during the Jurassic between the Inner Carpathian–Eastern Alpine terrane and the 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 the 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 have been recently located in the suture zone between the Inner Carpathian terrane forming the Pieniny Klippen Belt; other sequences are involved in the allochthonous 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 also be regarded as belonging to the Outer Carpathians (e.g., Ksiazkiewicz, 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 involved in the Pieniny Klippen Belt, whose strongly tectonized structure is about 800 km (500 mi) long and 1–20 km (0.6–12 mi) wide, which stretches from Vienna in the west to the Poiana Botizei (Maramures, northeast Romania) in the east (Figure 1). The largest part of the northern subbasin forms the Magura unit, traditionally belonging to the Outer Carpathians. The Pieniny Klippen Belt is separated from the Magura nappe by the Miocene subvertical strike-slip fault (e.g., Birkenmajer, 1986, 1988).

The Transylvanian domain formed perhaps the extension of the eastern Tethys between the European platform and Tisza–Dacia terrane (Sandulescu, 1988; Sandulescu and Visarion, 2000). In Romania, it was truly an oceanic realm as indicated by the existence of ophiolites. It developed during the Triassic and was closed during the Cretaceous. In Poland, the presumed Transylvanian domain sedimentary sequences are represented only by pebbles in the flysch of the Pieniny Klippen Belt and the Magura unit.

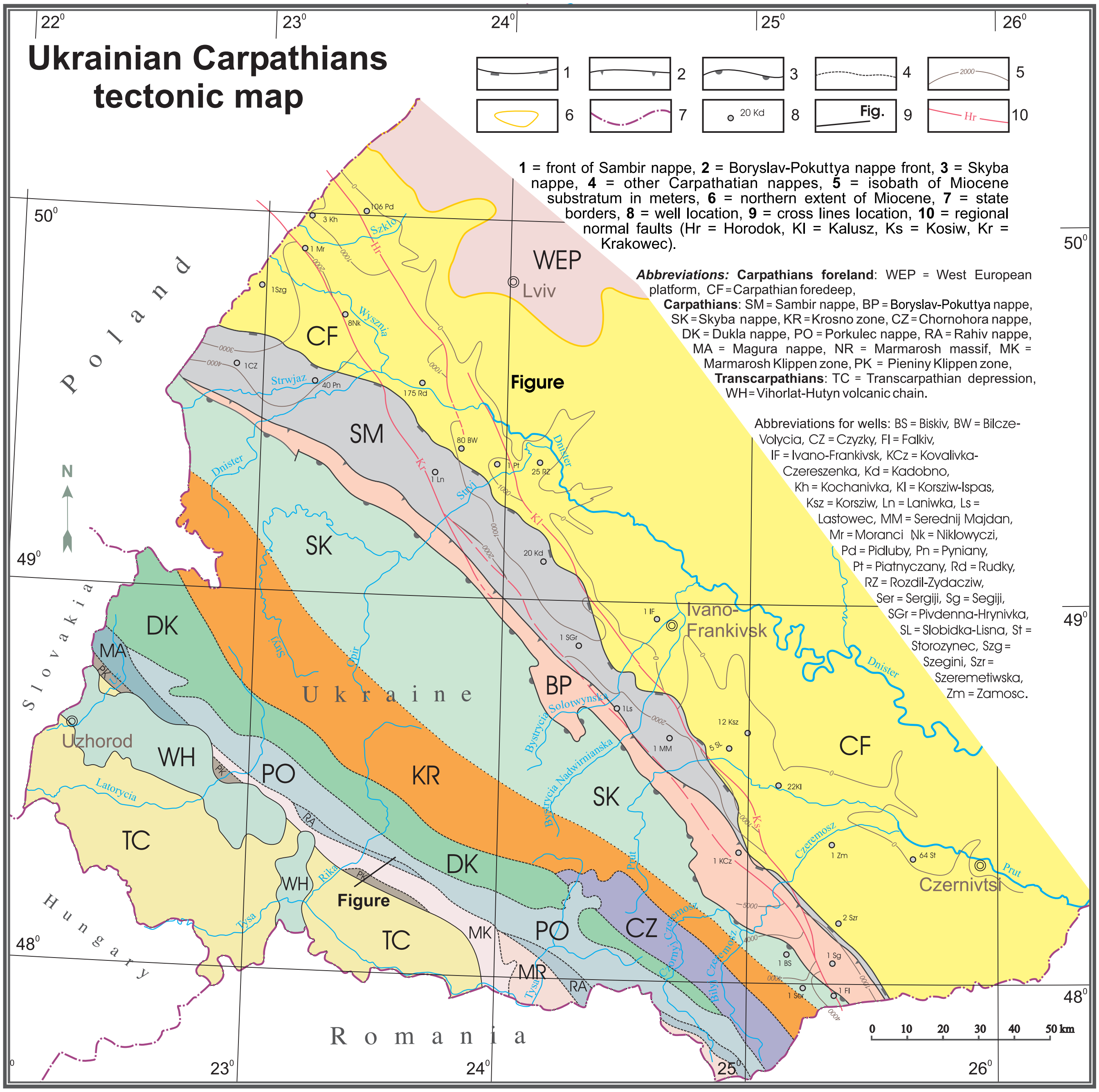

Figure 3. Ukrainian Carpathians tectonic map. Modified after Glushko and Kruglov (1986).

The Severin–Moldavidic realm (Balintoni, 1998), also known in Romania as the Outer Dacides and the Moldavides (Sandulescu, 1988; Stefanescu et al., 2006), developed in the North European platform as rift and/or back-arc basin. The Severin–Moldavidic 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, Subsilesian, Zdanice, and Skole tectonic units in Poland (Figure 2) and Czech Republic (Pescatore and Slaczka, 1984; Zytko et al., 1989; Stranik et al., 1993; Slaczka and Kaminski, 1998) and to Marmarosh, Rachiv, Porkulets, Chornohora, Skyba, and Boryslav–Pokuttya (Figure 3) nappes in Ukraine (Kruglov, 1989; Slaczka, 1996b). The subbasins in the Severin–Moldavidic are divided by ridges and uplifted zones. The most prominent one is the Subsilesian bulge between Silesian and Skole subbasins. Another ridge was located between Dukla and Silesian subbasins. The Subsilesian bulge is connected in the west with the slope sequences between the epicontinental part of the North European platform and the Outer Carpathian basins (Silesian and Magura basins). The Severin–Moldavidic basinal realm ends in Moravia, whereas slope sequences extend further westward. The Severin part (Severinides, see, e.g., Balintoni, 1998, 2001; or Outer Dacides, Sandulescu, 1988; Stefanescu et al., 2006) of the basin is represented in Ukraine by the Kaminnyi Potik and Rachiv units. This basin was closed during the Cretaceous. The Moldavic part (Moldavides, see, e.g., Sandulescu, 1988; Balintoni, 1998; Stefanescu et al., 2006) was closed during the Neogene.

The Getic–Marmarosh ridge (Golonka et al. 2003), also known as the Median Dacides (Sandulescu, 1988; Stefanescu et al., 2006), constitutes a fragment of the North European platform that rifted away during the opening of the Severin–Moldavidic basins. It includes Precambrian and early Paleozoic (Caledonian) granites and metamorphic rocks, late Paleozoic (Variscan) metamorphic rocks, as well as the late Paleozoic and Mesozoic sedimentary cover. The Getic–Marmarosh ridge separated the Severin–Moldavidic basin from the Transylvanian basin. Westward, the similar position has the Silesian ridge (Sandulescu, 1988; Oszczypko, 1992), which separated the Penninic–Magura realm and the Severin–Moldavidic realm. The eastern part of the Getic–Marmarosh ridge collided in the latest Early–Late Cretaceous with the Tisza–Dacia, forming several nappes. These nappes also included part of the Civcin–Severin–Moldavidic basins, the Rachiv and Porkulets units. The western part of the ridge was reorganized during the Late Cretaceous, forming the basement of the uplifted Silesian Cordillera, the Fore-Magura basin; a ridge separated Magura basin and Fore-Magura basin, as well as a marginal part of the Magura basin.

OUTER CARPATHIAN STRATIGRAPHY

The Outer Carpathians are made up of a stack of nappes and thrust sheets showing a different lithostratigraphy and tectonic structures (Figures 2, 3). Part of them can be traced along the whole of the Outer Carpathians, and a part can be traced only in the Eastern Carpathians. These tectonic units are overthrust one on the other from the south and built of sediments representing the time span between the Late Jurassic and the early Miocene (Ksiazkiewicz, 1962, 1977; Bieda et al., 1963; Mahel et al., 1968; Koszarski and Slaczka, 1976). They correspond to more or less separated sedimentary basins, and every basin generally displays a different lithostratigraphic development. During the overthrusting movements, the tectonic units became uprooted, and generally, only central parts of the basins are preserved.

The western and northern parts of the Outer Carpathians (Figure 4) consist of four main longitudinal units (basins): the Magura, Dukla–Fore-Magura, Silesian, and Skole belonging to two realms, the Penninic and Severin–Moldavidic, originally divided by the Silesian ridge (Cordillera). The other ridges and swells existed in the realms; the most distinctive were the Subsilesian swell and Andrychow ridge (Ksiazkiewicz, 1962). The Magura, Dukla, Silesian, and Skole sedimentary units (basins) extend eastward to the Ukrainian part of the Outer Carpathians. Several units, e.g., Marmarosh, Rachiv, Porkulets, Chornohora, Skyba (Skole), and Boryslav–Pokuttya units (Subbasin) in the Severin–Moldavidic realm, extend from Ukraine southeastward and southward to Romania; their correlation with the Western Carpathian units still remains somewhat uncertain and disputable. The relationship between the Magura unit and the Pieniny Klippen Belt in Ukraine and the adjacent part of Romania also remains speculative. Generally, the Magura unit is located between the Pieniny Klippen Belt (zone) and the Marmarosh zone (e.g., Sandulescu et al., 1981; Sandulescu, 1988; Zytko, 1999; Golonka et al., 2003); however, according to Kruglov (2001), fragments of the Magura nappe could also exist north from the Marmarosh zone.

In the Carpathian basins, enormous, continuous sequences of flysch were deposited mainly by different turbidity currents. The thickness of deposits locally exceeds 10 km (6.2 mi). These sediments were divided on a lithological basis into many formations with separate names. Commonly, the same formation has a different name in adjacent countries (Bieda et al., 1963; Geroch et al., 1967; Mahel et al., 1968; Slaczka, 1980; Kruglov, 1989, 2001; Pasternak, 1989; Golonka and Lewandowski, 2003). The source areas for the clastic material were situated on the outer and inner margin as well as on the intrabasinal ridges. The sedimentary sequences display lateral and vertical variations in the thickness and lithofacies, reflecting changes in subsidence and sedimentation rates and shifting of source areas.

Within the sedimentary sequences, generally, three main megastages of development exist, connected with the global events: an early stage from Jurassic to Albian, characterized by development of black shales; the second stage (Cenomanian–Eocene) characterized by the occurrence of red and variegated shales; and the final stage (Oligocene–early Miocene), when brown bituminous shales appeared and variegated shales disappeared.

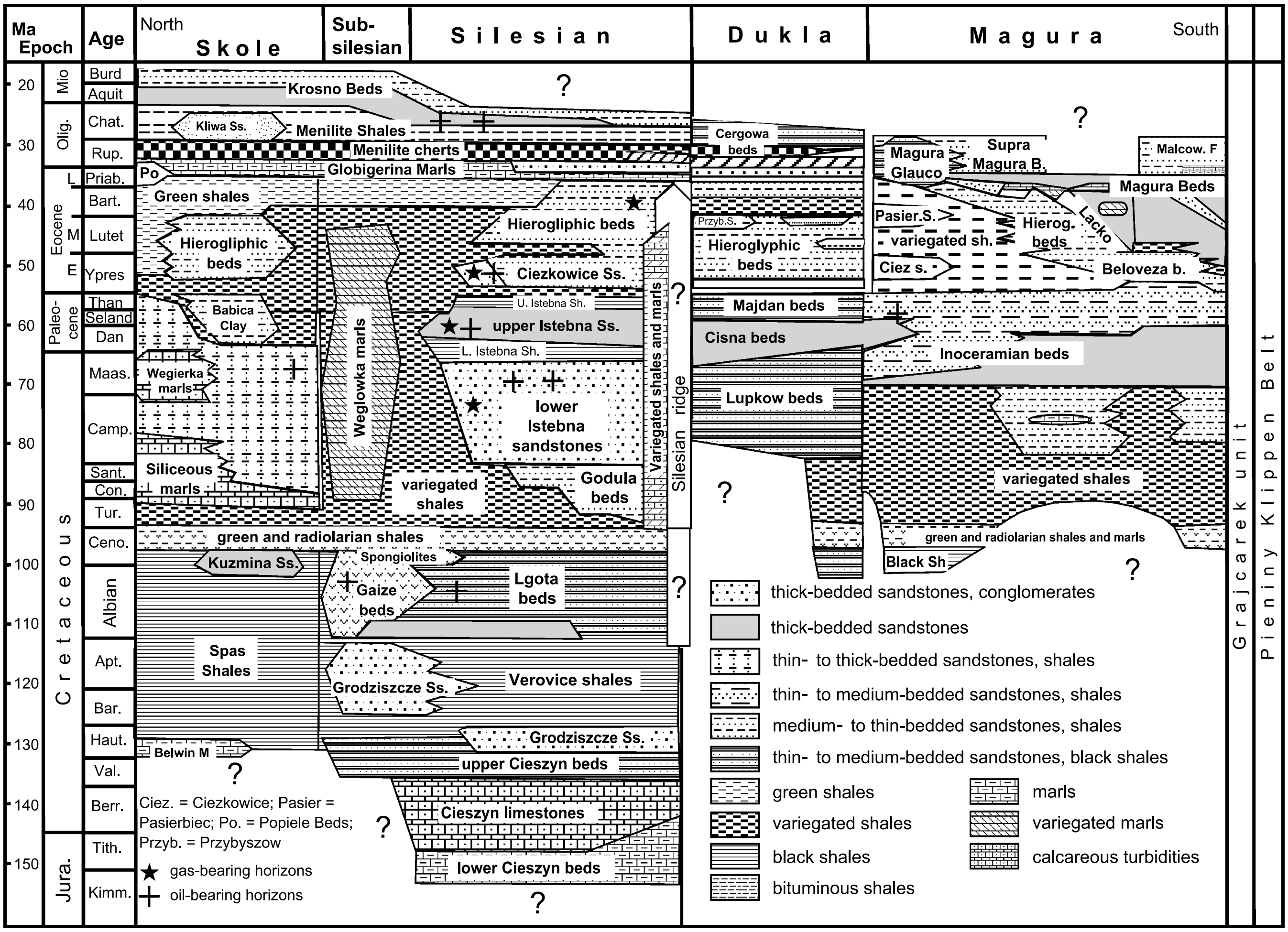

Figure 4. Lithology and chronostratigraphy of the Polish Outer Carpathians.

PENNINIC REALM

The Magura Nappe

The Magura nappe is the largest tectonic unit of the Western Carpathians and is linked with the Rhenodanubian flysch of the Eastern Alps. During the overthrust movements, the Magura nappe has been completely uprooted from its substratum along the ductile Upper Cretaceous rocks. The Tithonian–Barremian limestones and marls from the Kurovice Klippe are regarded as the basal portion of the marginal part of the Magura nappe in southern Moravia. More or less complete sections of this unit are known only from that part of the basin, which was incorporated into the Pieniny Klippen Belt (i.e., the Grajcarek, Birkenmajer, 1977; or Hulina unit, Burtan et al., 1981; Golonka and Sikora, 1981; Golonka and Raczkowski, 1984; see also Birkenmajer and Oszczypko, 1989). This succession is represented by Toarcian–Aalenian black shales, Bathonian–Oxfordian radiolarites, Kimmeridgian limestones (Ammonitico rosso), Tithonian–Barremian cherty limestones, Aptian spotty shales, Albian–Cenomanian black flysch, and Cenomanian green radiolarian shales (Birkenmajer, 1977). A single, Jurassic–Early Cretaceous Klippe of Hulina–Grajcarek type is also known from the Krynica unit (Golonka and Sikora, 1981).

The Albian–Cenomanian spotty shales remain in the southern margin of the Mszana Dolna tectonic window (Birkenmajer and Oszczypko, 1989). More recently, Hauterivian–Albian deposits have been recognized in a few localities in southern Moravia (Svabenicka et al., 1997). On the basis of facial differentiation in the Paleogene deposits, the Magura nappe in Poland, Slovakia, and Czech Republic (Moravia) has been subdivided into four facies-tectonic subunits: the Krynica, Bystrica (Nowy Sacz), Raca, and Siary (Figure 1, see also Koszarski et al., 1974). Additionally, in the west Slovakia segments of the Magura nappe, the Biele Karpaty unit has been distinguished.

The Cretaceous–Paleogene deposits of the Magura nappe may be subdivided into three turbidites complexes (see Oszczypko, 1992): Upper Cretaceous–Paleocene, lower–upper Eocene, and upper Eocene–Oligocene. Each of them begins with pelitic basinal deposits (variegated shales) that pass into thin- and medium-bedded turbidites with intercalations of allodapic limestones and marls, thick-bedded deposits, and finally, into thin-bedded turbidites. The Upper Cretaceous deposits begin with Cenomanian–Turonian, variegated, hemipelagic mudstones, with intercalations of thin-bedded turbidites (Malinowa Shale Formation). The thickness of the formation ranges from 70 m (230 ft) in the Raca subunit (Sikora, 1970) to as much as 200 m (660 ft) in the Krynica subunit (Birkenmajer and Oszczypko, 1989; Oszczypko et al., 1990). In the Raca and Bystrica subunits, sedimentation of variegated clays was finished at the Santonian–Campanian boundary, whereas in the Krynica subunit, it still existed in the Maastrichtian. The Malinowa Formation passes upward into thin-bedded turbidites with sporadic intercalations of thick-bedded sandstones. In the Grajcarek–Hulina unit, variegated shales pass upward into coarse-grained sandstones and exotic conglomerates (Burtan et al., 1984) of the Jarmuta Formation (Maastrichtian–Paleocene; see Birkenmajer and Oszczypko, 1989). This formation, as much as 400 m (1300 ft) thick, passes upward into medium- to thick-bedded calcareous turbidites of the Szczawnica Formation (Paleocene–lower Eocene), whose thickness also reaches 400 m (1300 ft). In the Krynica subunit, the thickness of the Jarmuta Formation reaches at least a few dozen meters (Oszczypko et al., 1990), whereas the thickness of the Szczawnica Formation oscillates about 300–400 m (1000–1300 ft) (Oszczypko et al., 1990, 1999a). In the more northern subunits of the Magura nappe, the Malinowa Formation is followed up in the section by thin- to medium-bedded turbidites of as much as 50 m (160 ft) in thickness, containing sometimes numerous 5–7- to 30-cm (2–3- to 12-in.)-thick intercalations of turbiditic limestones (Kanina beds; see Cieszkowski et al., 1989). These deposits pass upward into thick-bedded sandstones and conglomerates, which are as much as 100–400 m (330–1300 ft) thick (Szczawina sandstones) of the Maastrichtian–Paleocene age. The youngest unit of the complex, 100–200 m (330–660 ft) thick, is composed of medium- to thick-bedded turbidites of Paleocene age (Ropianka beds; see Malata et al., 1996), which are an equivalent of the Szczawnica Formation of the Krynica subunit.

The Ropianka beds are overlain by the lower to middle Eocene variegated, noncalcareous shales of Labowa Formation (Oszczypko, 1991). The thickness of this formation ranges from a few meters in the northern part of the Krynica subunit to as much as 130 m (430 ft) in the Siary subunit (see Sikora, 1970; Oszczypko, 1973, 1991). This formation represents the basinal plane facies, deposited below calcite compensation depth (CCD) (Oszczypko, 1991; Malata, 2001). The variegated shales pass upward into thin-bedded turbidites of the Zarzecze, Beloveza, and Hieroglyphic formations (Oszczypko, 1991; Oszczypko et al., 1999a). The thickness of these deposits varies from 500 to 600 m (1640 to 2000 ft) in the Krynica subunit (Zarzecze Formation), 350 m (1100 ft) in the Bystrica subunit (Beloveza Formation), and 200 m (660 ft) in the Raca subunit (Hieroglyphic beds).

In the Bystrica Subunit, the Beloveza Formation is overlain by thin- to medium-bedded turbidites with intercalations of the Lacko-type marls. In the Krynica, Bystrica, and Raca subunits, the youngest deposits of the Eocene complex belong to the Magura Formation, which is of lower to upper Eocene–Oligocene age. Thickness of this formation reached 1200–1400 m (4000–4600 ft) in the Krynica subunit, 500–2000 m (1600–6600 ft) in the Bystrica, and 1000 m (3300 ft) in the Raca subunit (Birkenmajer and Oszczypko, 1989; Oszczypko, 1999). The Magura Formation is represented by the thick-bedded turbidites and fluxoturbidites. The Magura Formation is locally overlain by the *Globigerina* marls (upper Eocene–lower Oligocene), Menilite shales, and the Malcov Formation (late Oligocene). In the northernmost part of the Magura nappe (Siary subunit), middle–upper Eocene variegated shales pass upward into the marls and thin-bedded flysch of the Zembrzyce beds, which contain a foraminiferal horizon of *Globigerina* marls (upper Eocene–lower Oligocene), thick-bedded glauconitic sandstones (Watkowa Sandstones) of the upper Eocene–lower Oligocene age, and finally, marls with intercalations of glauconitic sandstones (Budzow beds, Oligocene; see Oszczypko-Clowes, 2001). Traditionally, the Oligocene Malcov Formation was regarded as the youngest strata of the Magura nappe (Oszczypko-Clowes, 1998). However, in the vicinity of Nowy Sacz, the early Miocene Zawada Formation, belonging to the Magura nappe, has recently been discovered (Oszczypko et al., 1999b; Oszczypko-Clowes, 2001; Oszczypko and Oszczypko-Clowes, 2002). This formation reaches at least 450 m (1470 ft).

The Biele Karpaty unit was traditionally regarded as an equivalent of the Oravska Magura–Krynica subunit, which is in a similar tectonic position. Recently, Potfaj (1993) arranged the Biele Karpaty unit into Hluk and Vlara successions (groups). The Hluk succession, extended mostly in the southwest part of the Biele Karpaty Mountains, consists of several lithostratigraphic units. The Hluk Formation [Barremian(?)–Aptian], as much as 120 m (400 ft) thick, is represented by dark-gray, black-greenish shales and turbiditic limestones at the top. This formation passes upward into the Gbely (Kaumberg) Formation [Albian(?)–Cenomanian–Turonian], composed of variegated claystones and marls about 200 m (660 ft) thick. Higher up in sequence occurs the Puchov Formation [Coniacian(?)–Campanian–Maastrichtian], represented by 100 m (330 ft) of red and green shaly marls. These marls are interfingered with a 10-m (33-ft)-thick packet of gray-brown limy claystones, marls, and limestones (Antoninek Member, Campanian). The siciliclastic flysch begins with the Svodnice Formation (Paleocene–early Eocene), a 700-m (2300-ft)-thick complex of medium-bedded turbidites with thick intercalations of gray and brown-gray calcareous claystones. The upper part of the formation (Bzova Member, Paleocene–Eocene) contains a 600-m (2000-ft)-thick complex of thick-bedded, fine- to medium-grained sandstones, which resemble the Magura sandstones of the Krynica (Orava) subunit. Toward the north, the Svodnice Formation is replaced by thin-bedded, calcareous turbidites, about 600 m (2000 ft) thick (Nivnice Formation, late Paleocene), with two horizons of variegated shales at the base. This formation is overlain by the Kuzetov Formation (latest Paleocene to early Eocene), which is 250 m (800 ft) thick, as represented by thick horizons of the variegated shales and calcareous claystones and sandstone intercalations.

The Kaumberg (Lopenik) Formation (Campanian–Maastrichtian) belongs to the Vlara Group. It is built of thick sandy flysch with variegated claystones and marls at the base. In the upper part of the formation, two members were distinguished by Potfaj (1993). The Javorina Member (late Campanian–Maastrichtian) consists of thin-bedded turbidites with lenses of microconglomerates. The muscovite sandstones are composed of quartz, metamorphic rocks, and a large amount of carbonate clasts. The thickness of this member is about 600 m (2000 ft). The Drietomica Member (Maastrichtian) is characterized by the domination of thick-bedded, fine- to coarse-grained sandstones and microconglomerates of mixed carbonate-siliciclastic composition. The estimated thickness of this member is 200 m (660 ft). The Rajkovec beds [Paleocene–early Eocene(?)] are predominantly a sandstone-flysch sequence with thin intercalations of claystones and limy mudstones. The sandstones are fine to medium grained and of mixed carbonate-siliciclastic composition. The maximum thickness of the Rajkovec Member is as much as 500 m (1600 ft). The youngest deposits of the Vlara succession belong to the Chabova beds (late Paleocene–early Eocene), developed as thick beds of medium- to coarse-grained siliciclastic sandstones with carbonate clasts.

THE GETIC–MARMAROSH UNITS

The Marmarosh Unit

The Marmarosh unit is located in the Transcarpathian Ukraine next to the Romanian border (Figures 3, 5). The development of this unit differs from the other Outer Carpathian flysch units, and as was mentioned above, the exact position of this unit is disputable; it is placed adjacent to Pieniny Klippen Belt or north to the Magura unit. It consists of two parts: the Marmarosh Klippen zone (internal) and the Marmarosh crystalline massif (external).

The Marmarosh Klippen zone (Figures 3, 5) is situated southwest and west from the Marmarosh Massif. It is divided into two subzones: the external Vezhany subzone, containing olistolithes (Soimul olistostrome), and the internal Monastyrets subzone, built mainly by flysch. The giant blocks (as much as 1 km [0.62 mi]), olistolithes, and olistoplakes (Kruglov, 1965) occur in the marginal part of the Vezhany subzone (Soimul olistostrome). They are built of Paleozoic and Proterozoic schists and gneisses, Paleozoic, Triassic, Jurassic, and Barremian–Aptian dolomites and limestones of age, and Permian–Triassic quarzitic sandstones and conglomerates. The blocks of serpentinites, diabases, gabbro-diabases, pebbles and boulders of granites, quartz porphyries, and granitoids are also numerous there. The olistostromes pass upward into polimiktic conglomerates with a thickness of about 1000 m (3300 ft) covered by dark gray mudstones intercalated by thin sandstones (Soimul beds; Aptian–Albian). Complexes of thick sandstones are developed only locally. The Soimul beds pass upward into red marls (Puchov beds; Turonian–Campanian) and thin-bedded sandstones with green-gray and sparse red shales, correlated by Kruglov (1965, 1969) with Jarmuta beds (Maastrichtian) known from the Pieniny Klippen Belt in Poland and Slovakia. Paleogene sediments start with conglomerates and thick-bedded sandstones (Metovo beds) covered by thick-bedded marls of the Eocene. They pass into Oligocene black marls and shale (Dusina Beds) and thick-bedded gray sandstones. In the inner Monastyrets subzone, the lithostratigraphic profile starts with Paleogene variegated shales intercalated with thin-bedded and sporadically bedded sandstones (Shopurka beds) as much as 1000 m (3300 ft) thick. They are covered by thick-bedded sandstones (Dragov beds). According to Sandulescu et al. (1981), this sequence represents the southeast prolongation of the Magura unit.

The Marmarosh Massif is built by complex, dislocated, mesozonally metamorphosed Riphean–Vendian rocks and by sedimentary, volcanic, and epizonally

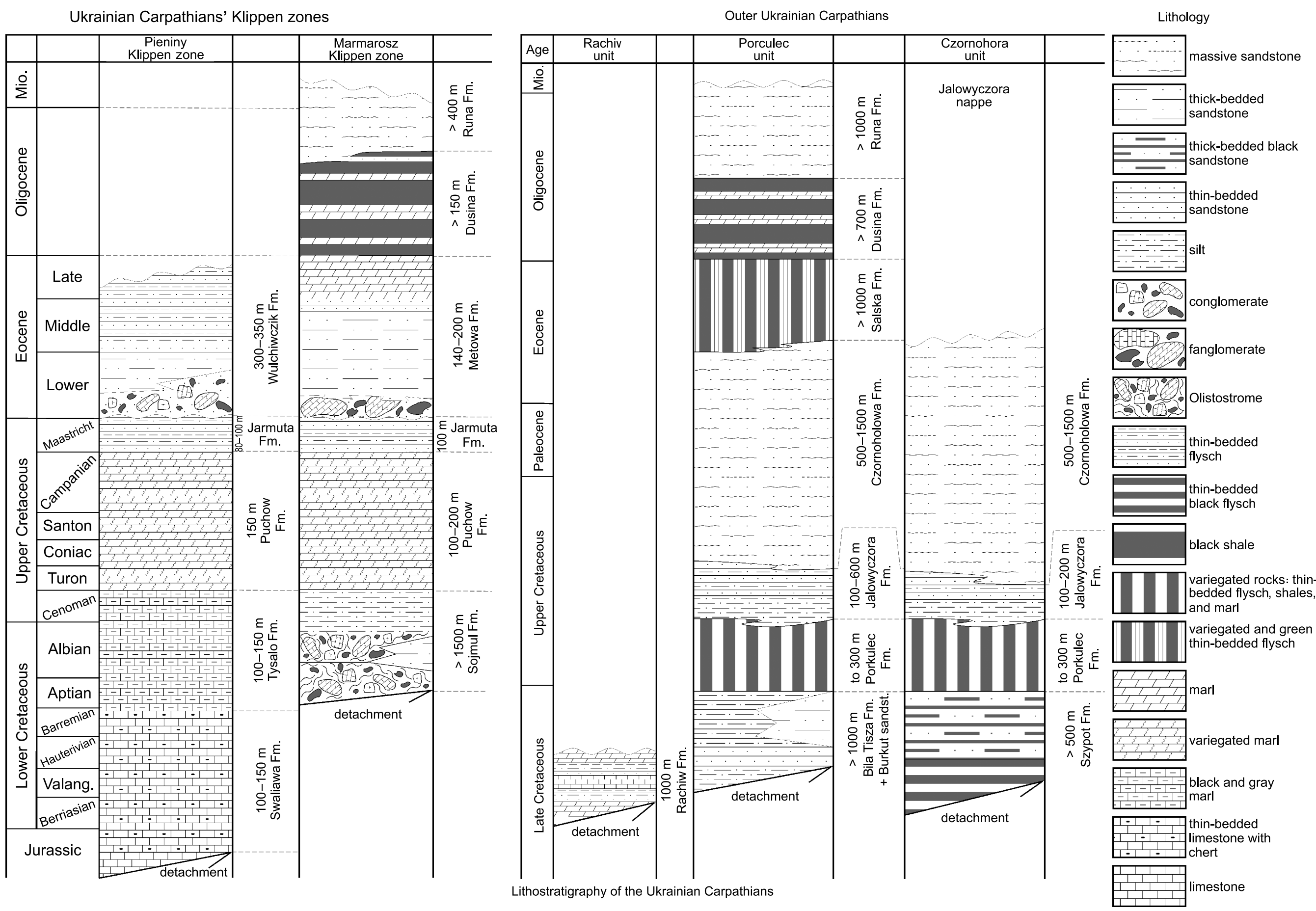

Figure 5. Lithology and chronostratigraphy of the internal part of the Ukrainian Carpathians.

metamorphosed Carboniferous, Triassic, and Jurassic formations. The Cretaceous conglomerates, organogenic limestones, and marls of age discordantly overlie older rocks. According to Romanian geologists, the Marmarosh Massif belongs to the large Getic realm (e.g., Sandulescu, 1988; Balintoni, 1998, 2001; Golonka et al., 2003) described above as the Getic–Marmarosh ridge. The metamorphic basement sequences of this realm are thrust over the flysch sequences of the Outer Dacides (Severinides). The nappe character of the Ukrainian part of the Getic–Marmarosh ridge was described in detail by Kropotkin (1991).

OUTER DACIDES–SEVERINIDES

The Kaminnyi Potik Unit

In the frontal part of the Marmarosh Massif, a tectonic thrust fold (Glushko and Kruglov, 1986) is present that is composed of dark, thin-bedded limestones, black shales, sandstones, and conglomerates (Kaminnyi Potik beds; Tithonian–Valanginian) containing the Upper Jurassic effusives of the basic type. It belongs probably to the basins of the Outer Dacides–Severinides and is the prolongation of the Black Flysch unit of the Romanian Carpathians.

Rachiv Unit

The Rachiv unit is represented mainly by the Valanginian–Hauterivian deposits (Rachiv beds) 1000 m (3300 ft) thick. The lower part is represented by black shales, calcareous thin- and medium-bedded turbiditic sandstones, and limestones passing upward into the complex that contains an intercalation of thick-bedded sandstones and conglomerates. Exotic blocks of Mesozoic limestones and diabases are also present. The whole sequence is terminated by black shales and thin-bedded calcareous sandstones. The uppermost part can also represent the Barremian. The Rachiv unit continues southward into the Cuk digitation of the Ceahlau nappe in the Romanian Carpathians containing basic volcanites.

Porkulets Unit

The oldest, Early Cretaceous deposits of the Porkulets unit (Figures 3, 5) are represented by gray shales, marly shales, thin- to thick-bedded laminated sandstones (Bila Tisza beds; Barremian–Albian, more than 1000 m [3300 ft] thick), and lenses of conglomerates that contain blocks of metamorphic and volcanic rocks (Bogdan and Bronka conglomerates). The Bila Tisza beds are locally replaced by massive, thick-bedded sandstones, more than 500 m (1600 ft) thick (Burkut conglomerates). The Porkulets beds, about 50–300 m (160–1000 ft) thick, are represented by uppermost Albian, Cenomanian, and Turonian red, green, and brown shales and marls intercalated by thin- and medium-bedded sandstones. They pass upward into a series of gray, medium-bedded sandstones, shales, and marls [Jalowychora beds; Coniacian, Santonian, Campanian(?)]. Their thickness increases to as much as 600 m (2000 ft) toward the northwest. In this area, they are known as lower Berezna beds. Higher up, a thick (as much as 1500 m [5000 ft]) complex of thick-bedded sandstones is present, representing the time span from late Campanian up to middle Eocene (Chornogolova beds). The higher part of the middle Eocene and upper Eocene is developed as green-gray and red shales (Skalska beds) with intercalations of Sol beds; as much as 1000 m (3300 ft), whereas the Oligocene is developed as black and yellowish marls and black shales (Dusina beds; as much as 700 m [2300 ft]) locally with intercalations of cherts and sandstones (Turica beds). The higher part of the Oligocene rocks is represented by a complex of thick-bedded sandstones as much as 1000 m (3300 ft) thick (Malavyzna beds). Toward the southeast, the Porkulets unit continues into the Bodoc digitation of the Ceahlau unit in Romania.

Dukla Unit

The oldest, Early Cretaceous sediments are known only from the eastern part of the Dukla unit (Figures 4–6). They are represented by black shales, siliceous mudstones, and sandstones with intercalations of cherts and siderites (lower Shypot beds; Barremian–Aptian) passing upward into a series of siliceous sandstones and black shales (upper Shypot beds; Albian). In the Cenomanian, green and red shales with subordinate thin turbidites started to develop (lower Yalovetz beds). The Senonian sequence commences (Figure 6) with a 900-m (2900-ft) complex of dark shales and medium-bedded, fine-grained, calcareous sandstones (lower Berezna beds in Ukraine). This lithofacies, known in the southern part of the Dukla unit as Lupkow beds (Campanian–Maastrichtian and Campanian–Paleocene),

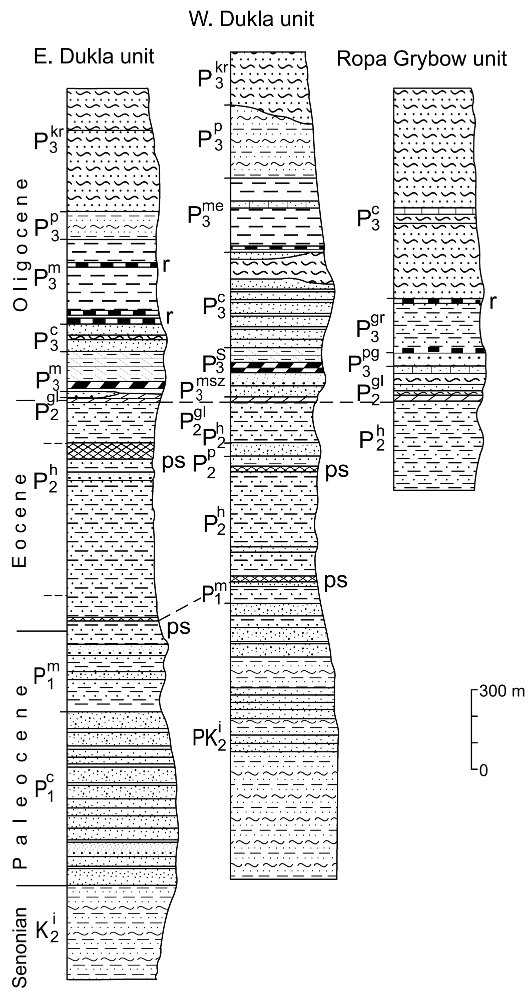

Figure 6. Comparative lithostratigraphic profiles of the Dukla and Grybow units. P_3^{kr} = Krosno beds; P_3^{p} = transition zone between Krosno and Menilite beds; P_3^{me} = Menilite shales; r = cherts; P_3^{c} = Cergowa sandstones; P_3^{s} = Jawornik marls; P_3^{msz} = Mszanka sandstones; P_2^{gl} = *Globigerina* marls; P_2^{h} = Hieroglyphic beds; ps = red and green shales; P_1^{m} = Majdan beds; PK_2^{c} = Inoceramian beds; P_1^{c} = Cisna beds; K_2^{l} = Cisna beds.

represents the oldest sediment of the Dukla unit in Poland and Slovakia (Lesko and Samuel, 1968; Slaczka, 1971; Korab and Durkovic, 1978). Uppermost Cretaceous and Paleocene deposits in almost the entire unit are represented by thick-bedded calcareous, coarse-grained, feldspathic sandstones and/or fine-grained, micaceous sandstones intercalated with dark clay shales (Cisna beds or upper Berezna beds in Ukraine, mainly Paleocene). This complex disappears toward the west and toward the north (data from Jasliska 2 borehole, southeast from Krosno). In the western part of the unit, between villages of Cisna and Jasliska, black shales and dark siliceous sandstones (Majdan beds; Paleocene) cover the Cisna beds. Locally, they contain the intercalation of thick-bedded sandstone complexes (Welka Berezna in Slovakia, Ljuta beds in Ukraine). Sometimes, these thick-bedded sandstones dominate. Starting from the youngest Paleocene, a striking change of sedimentation appears. The Eocene deposits are developed as a complex of green shales, medium to thin bedded, fine and medium grained, as much as 1000 m (3300 ft) thick (Hieroglyphic beds in Polish, sub-Menilite beds in Slovakian, and Strichava beds in the Ukraine parts of the Dukla unit) with intercalations of red shales (Sol and Vyshka beds in Ukraine). Locally, intercalations of thick-bedded sandstone lenses exist (as much as 200 m [660 ft]) (Przybyszow sandstones in Poland and Stavne beds in Ukraine). The upper Eocene Hieroglyphic type of sediments passes upward into the green shales and higher up into the *Globigerina* marls. At the beginning of the Oligocene, dark bituminous sediments started to develop in the Dukla unit as well as in the northern units. These sediments are represented by dark brown shales (Menilite shales) with horizons of cherts. Within these shales, several lithological members have developed. In the lowermost part, a lenticular complex (as much as 250 m [820 ft]) of thick-bedded, coarse-grained, or conglomeratic sandstones (Mszanka sandstones) is present; higher up, the complex-brown siliceous marls (Jawornik marls) and a thick complex (as much as 300 m [1000 ft]) of thick-bedded, medium-grained, calcareous sandstones intercalated by gray calcareous shales are present (Cergowa sandstones; Slaczka and Unrug, 1976). Toward the south, the sandstone complex is replaced by gray marly shales and thin-bedded, calcareous, laminated sandstones, calcareous shales, and subsequently replaced by a complex composed predominantly of shales. The dark Menilite shales pass gradually upward into a 1000-m (3300-ft) series of calcareous, thin-bedded, fine-grained sandstones and gray calcareous shales (Krosno beds; Oligocene), which terminate the sequence of the Dukla unit.

Fore-Magura Group of Units

In the Western Carpathians, north from the Magura unit, several units are characterized by the occurrence of the Upper Cretaceous–Paleocene sediments similar to those of the Magura unit and the Oligocene deposits similar to deposits from the Silesian unit. From the west, these are the Fore-Magura sensu stricto, Obidowa–Slopnice, Jaslo, and Grybow units. The relation between these units is not clear, but it is supposed that the Grybow unit was located in the more internal position than the Dukla unit or represents a prolongation of the southern part of the Dukla unit.

Fore-Magura Unit Sensu Stricto

This unit is known only from the western part of the Polish Carpathians in front of the Magura unit (Burtan, 1968; Golonka, 1981). The sequence commences with the Upper Cretaceous thick-bedded sandstones, which contain intercalations of green and gray shales. The lower and middle Eocene is developed as argillaceous variegated shales, locally with variegated marls, and the upper Eocene is developed as green shales. The Oligocene deposits are characterized by the occurrence of thick-bedded, gray, micaceous, and calcareous sandstones passing upward into gray, marly shales with subordinated thin, laminated micaceous sandstones (Krosno beds).

Obidowa–Slopnice Unit

This unit (Figure 7) is situated below the Magura nappe and is recognized only from the boreholes (Slopnice and Obidowa areas) (Cieszkowski et al., 1985). It shows a resemblance to the Dukla and Zboj units and probably represents its western equivalent. The oldest sediments of the Obidowa–Slopnice unit represent the Upper Cretaceous and are developed as thick-bedded, coarse-grained sandstones (1000 m [3300 ft] thick) showing similarity to a part of the Inoceramian beds in the Dukla unit. They pass upward into a complex with pebbly mudstones (Obidowa beds), covered again by thick-bedded sandstones (70 m [230 ft] thick) that belong to the Paleocene. The Eocene is developed mainly as green-gray shales and thin sandstones similar to those in the other Fore-Magura units. Intercalations of black shales, thick-bedded conglomerates, and sandstones reaching 700 m (2300 ft) (Rdzawka beds) occur in the upper Eocene deposits. The Eocene sequence is terminated by dark, black, calcareous mudstones and micaceous sandstones that show similarity to some

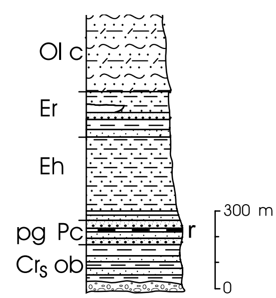

Figure 7. Lithostratigraphic profile of the Obidowa–Slopnice unit. Olc = Cergowa beds (Oligocene); Er = Rdzawka beds (late Eocene); Eh = Hieroglyphic beds (Eocene); pg Pc = thick-bedded sandstones (Paleocene); Cr$_s$ob = Obidowa beds (Late Cretaceous); r = cherts.

Oligocene deposits from the Dukla unit. This unit is correlated with the sediments found in Zboj 1 borehole near the Slovakian–Ukrainian border. That suggests a rather great extension of the Obidowa–Slopnice lithofacies and its position in front of the Dukla–Grybow units.

Grybow Unit

The Grybow unit (Figure 6) represents a western prolongation of the Dukla unit and is only known from the tectonic windows in the Magura nappe (Kozikowski, 1956; Sikora, 1970) and also from the boreholes. The sequence commences with the Eocene sediments that are developed as green, gray, and black shales, intercalated by thin- and medium-bedded sandstones (Hieroglyphic beds). The Eocene sequence terminates the *Globigerina* marls in a similar way as in other external units of the Outer Carpathians. The Oligocene is developed as gray marls with sporadic intercalations of sandstones (sub-Grybow marls, about 150 m [500 ft] thick). They are covered by black, hard shales with horizons of black hornstones at the top (Grybow shales, about 200 m [660 ft] thick). The youngest sediments are represented by gray, calcareous, thin- and medium-bedded sandstones with gray calcareous shales that prevail toward the top (Cergowa beds; as much as 400 m [1300 ft] thick). Toward the east, the Grybow unit passes probably into the inner part of the Dukla unit, and its prolongation can be the Porkulets unit. Both of them show similarity in the development of the Oligocene sediments.

Jaslo Unit

This unit, lately delimited in the Harklowa peninsula of the Magura nappe (Koszarski and Koszarski, 1985), is represented by Eocene variegated and green-brown shales covered by greenish gray marly shales with intercalations of sandstones (Dulabka beds; Eocene–Oligocene). The higher part of the Oligocene is developed as brownish, commonly siliceous shales followed by grayish marly shales and sandstones (Krosno beds). The occurrence of olistostromes is very typical, and some geologists (L. Jankowski, 2003, personal communication) state that all deposits of the Jaslo unit represent olistostromes in front of the Magura nappe.

Zboj Unit

The Zboj unit was found in borehole Zboj 1 at the depth of 3800–5002 m (12,500–16,410 ft) in Eastern Slovakia near the Ukrainian border below the Dukla unit (Korab and Durkovic, 1978). Mainly massive, homogenous, arkosic sandstones that are strongly siliceous represent this unit. Subordinately, complexes of greenish shales with thin-bedded sandstones and intercalations of mudstones with slump structures are present. These deposits, called Zboj beds, represent the upper Eocene and probably also the Oligocene. The possibility cannot be excluded that they represent the innermost part of the Silesian unit.

Chornohora Unit

The Chornohora unit (Figures 3, 5) is situated in the southeast part of the Ukrainian Carpathians and disappears toward the northwest. It can be correlated with the Audia zone in the Romanian Carpathians. Its situation outside the Dukla unit and the development of the Lower Cretaceous imply that the Chornohora unit can represent the prolongation of the inner part of the Silesian unit (Slaczka, 1959). However, according to one of the authors of this chapter (Kruglov), it can also form a prolongation of the Dukla unit. The Chornohora unit is divided in two subunits: the Skupiv subzone in the north and the Yalovychor, or the Hoverla, subzone in the south. This division is based on some differences in the development of the Late Cretaceous sediments. The oldest sediments (Barremian–Aptian) are represented by black, calcareous shales with siderites, fucoid marls, and thin-bedded sandstones (lower Shipot beds; as much as 300 m [1000 ft]) passing upward into series to as much as 200 m (660 ft) of dark, quartzitic sandstones with black shales (upper Shipot beds of Albian age). The Cenomanian–Turonian is developed, similarly to other Carpathian areas, as green and red shales, with thin sandstones and red marls (Porkulets beds; as much as 200 m [660 ft]). In the lower part of this sequence, intercalations of black shales, radiolarites, and siderites are present. The variegated sediments pass upward into gray shales, marls, and sandstones of Coniacian–Santonian age (Yalovychor beds; as much as 400 m [1300 ft]). The late Senonian is characterized in the Skupiv subzone by medium-bedded, gray sandstones and shales (Skupiv beds; as much as 500 m [1600 ft]) and in the Yalovychor–Hoverla subzone by a complex as much as 1000 m (3300 ft) of very thick- to thin-bedded sandstones and gray shales (Chornohora beds) that terminate the whole sequence of this subzone. The younger sediments are known only from the Skupiv subzone. Paleocene–early Eocene thin- to thick-bedded sandstones, limestones, and gray-green and red shales (Hniletz beds, about 400 m [1300 ft] thick) are present, covered by middle Eocene massive micaceous sandstones (Topilchany beds; as much as 250 m [800 ft] thick) and late Eocene thin- to medium-bedded sandstones and variegated shales (Probijeny beds; as much as 350 m [1150 ft]). The sedimentary sequence terminates by the Oligocene dark brown shales (Menilite beds). Recently, olistostrome formations of insignificant thickness were discovered in front of the Chornohora nappe. They are being compared with the Miocene facies of Slon in the Vinecisu strata in the Romanian Carpathians. This formation is characterized by the wide development of variegated, red rocks.

Silesian Unit

The oldest sediments of the Silesian (Figures 4, 8) are known only in the Moravia and Silesia areas in the Western Carpathians (see Picha et al., 2006). They are represented by the lower Tithonian dark gray, calcareous mudstones (lower Cieszyn shales) that began the euxinic cycle that lasted without major interruption until the Albian. Some of these mudstones represent slump deposits. The mudstones pass upward into turbiditic limestones and marls (Cieszyn limestones; Tithonian–Berriasian; as much as 200 m [660 ft] thick). The material of the detritic limestones was derived from the adjacent shallow-water, calcareous platforms, mainly from the north. The younger, Valanginian–Hauterivian dark gray, black calcareous shales with intercalations of dark, thin- and medium-bedded calcareous sandstones (upper Cieszyn beds; as much as 300 m [1000 ft] thick) are known from the whole Silesian unit. They generally pass into Barremian–Aptian black shales (Verovice shales; as much as 300 m [1000 ft]). During the Hauterivian and Barremian, several complexes (as much as 200 m [660 ft] thick) of sandstones and conglomerates have developed (Grodziszcze sandstones). During the late Tithonian and Early Cretaceous, the opening of the western part of the Silesian basin alkaline magma (teschenites association rocks) intruded the flysch deposits (Lucinska-Anczkiewicz et al., 2000). The black shale turbiditic sedimentation (Lgota beds; as much as 450 m [1500 ft]) developed in the Silesian basin in the beginning of the Albian and lasted during the whole Albian. It started with thick-bedded, coarse-grained sandstones and conglomerates that passed upward into thin- and medium-bedded, quarzitic sandstones intercalated by black, greenish shales. Locally, especially in the inner part of the Silesian basin, the Lgota beds are entirely developed as thin-bedded facies. North from Sanok, Lgota beds are replaced by fine-grained, light gray sandstones with sponge needles (Gaize beds, 150 m [500 ft] thick). At the beginning of the Cenomanian, slow and uniform sedimentation occurred, and green shales, with an intercalation of black shales and radiolarites, developed. Similar facies also developed in other parts of the Outer Carpathian flysch basin. During the Cenomanian–Turonian, the green shales and radiolarites passed upward to red and variegated shales. During the early part of the Late Cretaceous, a supply of clastic material commenced. The uppermost Turonian and the lower Senonian are represented by a very thick flysch series of Godula beds (approximately 2000 m [6600 ft] thick) developed as thick- and thin-bedded sandstones intercalated by green shales. The Godula beds occur mainly in the inner zone of the western and central parts of the Silesian unit. Toward the north and east, they are replaced laterally by variegated shales (Godula shales). The Godula beds are covered by another sandy complex, the Istebna beds (upper Senonian–Paleocene), which reaches 1700 m (5600 ft) locally. It is represented in the lower part by noncalcareous, coarse-grained, thick-bedded fluxoturbidites and conglomerates with subordinate intercalations of gray shales. The upper part is developed as a complex of slightly calcareous, thick-bedded, and coarse-grained conglomerates and sandstones covered by dark gray shales reaching 200 m (660 ft). These lithofacies are known from Moravia up to the Bieszczady Mountains in the east, where they were found in deep boreholes. The Paleocene–Eocene deposits of the Silesian unit are characterized by the occurrence, in the lower part, of a thick complex that attains 500 m (1600 ft) of thick-bedded fluxoturbidites and conglomerates in the form of elongated lenses more than 200 m (660 ft) thick (Ciezkowice sandstones) with a thin intercalation of red shales. The middle and upper Eocene consist principally of alternating green gray shales and thin-bedded sandstones (Hieroglyphic beds) that pass upward into green shales and *Globigerina* marls. Only in the southeast part of the Silesian unit is the upper Eocene developed as thick-bedded sandstones (Mszanka beds). Near the Roznow Lake, the Eocene deposits are more sandy and could be regarded as a separate unit, the Michalczowa unit (Cieszkowski, 1992).

During the early Oligocene, the sedimentation of brown, generally bituminous shales started that display horizons of cherts (Menilite shales, 200–400 m [660–1300 ft]). Locally in the southern area, the shales are replaced by thick-bedded sandstones (Magdalena sandstones) or partly replaced by thick-bedded calcareous sandstones (Cergowa sandstones). The Menilitic shales pass upward into the Krosno beds of the Oligocene–early Miocene age. They are developed in the lower part as thick- and medium-bedded, calcareous sandstones that pass upward into a complex of medium- to thin-bedded sandstones alternating with gray marly shales and covered by a complex of gray marly shales with subordinate thin-bedded sandstones. These lithofacies are diachronous across the Silesian unit (Jucha and Kotlarczyk, 1961; Koszarski and Zytko, 1961); in the northeastern part of this unit,

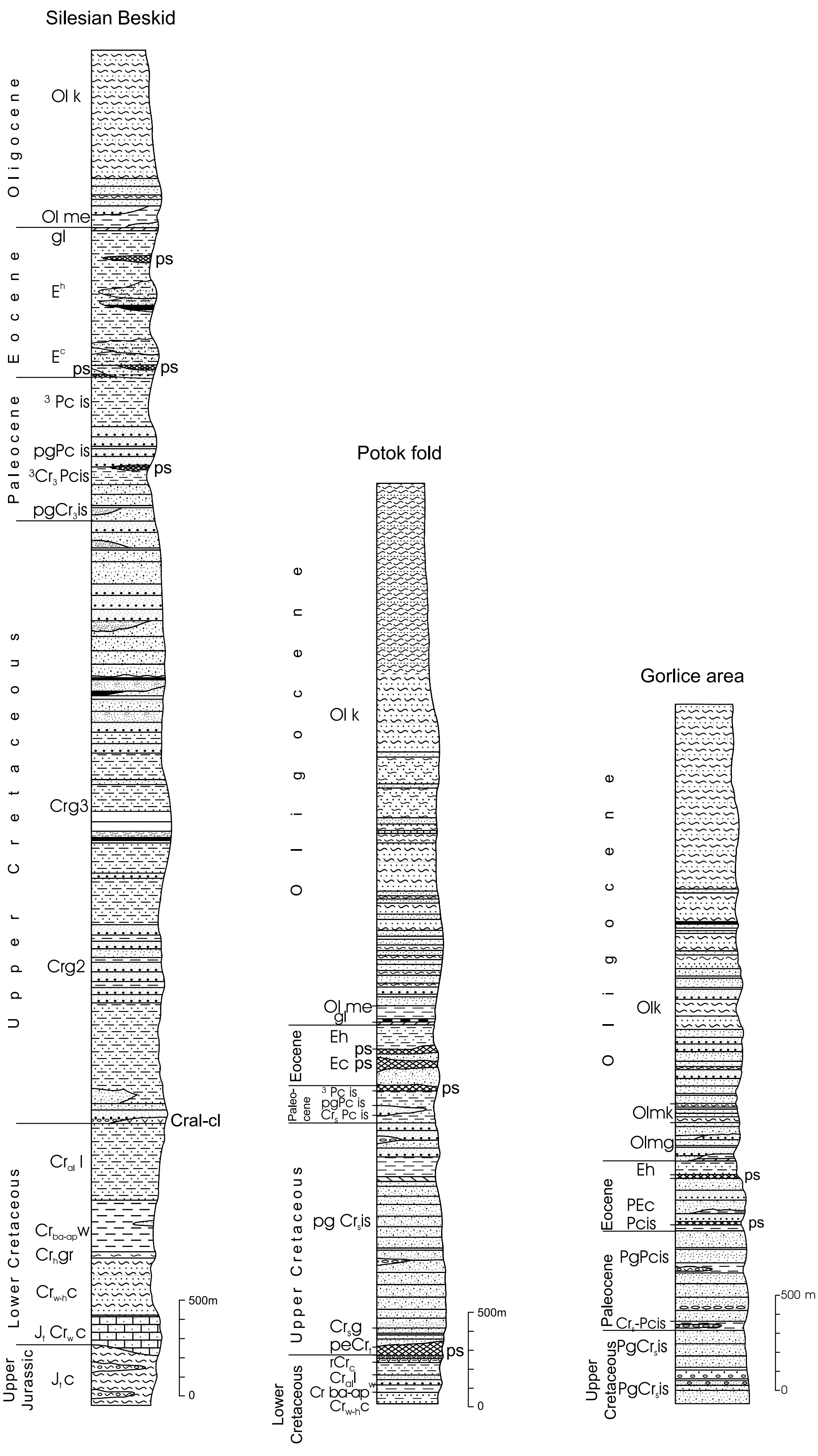

Figure 8. Comparative lithostratigraphic profiles of the Silesian unit from west to east. Olk = Krosno beds; Olme = Menilite beds; Olmg = Magdalena sandstones; gl = *Globigerina* marls; Eh = Hieroglyphic beds; Ec = Ciezkowice sandstones; ps = variegated shales; lPc is = upper Istebna shales; pgPcis = upper Istebna sandstones; lCr_3Pcis = lower Istebna shales; $pgCr_2is$ = lower Istebna sandstones; Cr_2g = Godula beds; Crg3 = upper Godula beds; Crg2 = middle Godula beds; Crg1-cl = lower Godula beds and white marls; $peCr_2$ = red shales (Godula shales); rCr = cherts; $Cr_{al.}l$ = Lgota beds; Crba-ap^w = Verovice shales; Cr_hgr = Grodziszcze sandstones; $Cr_{wh}c$ = upper Cieszyn shales; J_3Cr_wc = Cieszyn limestones; J_3c = lower Cieszyn shales.

they also represent the upper Miocene. However, in the inner part of the Silesian unit, in front of the Magura nappe, the Krosno beds also pass into the upper Miocene deposits that contain olistolithes derived from the frontal part of the Magura nappe (Slaczka and Oszczypko, 1987). In addition, in the southeastern part of Poland, in the Bieszczady Mountains in the lower part of the Krosno beds, horizons of olistostromes occur with blocks of metamorphic rocks and limestones, and near Sanok town, horizons of olistostromes occur with blocks of granites.

Toward the east, near the Polish–Ukrainian border, as an effect of plunging of the Silesian unit, the deposits that are older than the Oligocene almost completely disappear from the surface, and the unit passed in the Ukrainian Carpathians into the internal part of the Krosno zone (the Turka subzone that wedges out near the Black Tisza). In addition, a change in lithofacies exists. The Late Cretaceous (Coniacian–Maastrichtian) sediments are represented by a 200-m (660-ft) sequence of dark, medium-bedded quartzitic sandstones and dark shales that pass into the complex of dark medium- to thick-bedded sandstones and dark greenish shales (Soimy beds; Paleocene–middle Eocene; 400–500 m [1300–1600 ft]). The latter lithofacies of the Krosno zone are similar to those known from the inner part of the Silesian unit from the wells Wetlina Geological Institute (IG 1) and Wetlina IG 2 near the village of Wetlina in the southeastern Polish Carpathians. The late Eocene and Oligocene–early Miocene deposits (as much as 2000 m [6600 ft]) are similar to those of the Silesian and Skole–Skyba units.

Andrychow Ridge

This unit is represented by several huge blocks on the boundary between the Silesian and Subsilesian units, near the town of Andrychow. Probably, they are remnants of a calcareous platform that was situated between Silesian and Subsilesian sedimentary areas or represented a part of Subsilesian substratum. The composition of the klippes differs from the adjacent units, although the Upper Cretaceous sediments show a similarity to the sequences of the Subsilesian unit. The nonflysch, calcareous facies are very characteristic for the Andrychow ridge sequences (Ksiazkiewicz, 1951; Gasinski, 1998). The basement of the ridge was made up of granite-gneiss or mylonitized rocks. The sedimentary sequences are represented by crinoidal and shallow-water limestones of Late Jurassic age covered by transgressive early Campanian conglomerates and marls, limestones, and shaly marls of Campanian and Maastrichtian age. Paleocene and early Eocene are represented by organogenic limestones and shales. The more basinal or slope facies are represented by Maiolica-type Late Jurassic–Early Cretaceous cherty limestones (Olszewska and Wieczorek, 2001).

Subsilesian Unit

During the Early Cretaceous and the lower part of the Late Cretaceous, the sedimentation in the Subsilesian realm (Figures 3, 9) had a basinal character, very similar to the Silesian basin, and both basins were connected. The Subsilesian realm became generally uplifted later on, and on the slope of this uplift, marly lithofacies developed. The oldest known deposits (Barremian–Aptian) of this unit are euxinic black shales (Verovice shales), with intercalations of complexes of thick-bedded, coarse-grained, and conglomeratic sandstones (Grodziszcze sandstones; as much as 200 m [660 ft]). These sediments show similarity to the deposits of the same age in Silesian subbasin. During the Albian, turbiditic sedimentation developed which displayed a thinning-upward sedimentary sequence. The lower part of the sequence is represented by a complex of thick-bedded, fine- to coarse-grained sandstones more than 150 m (500 ft) thick, with intercalations of black shales (lower Lgota beds). These sandstones are overlain by a 400-m (1300-ft)-thick complex of fine-grained sandstones varying in thickness. A part of these beds represent typical gaizes, with numerous spicules (Gaize beds), in places with lenses of spongiolites. At the beginning of the Cenomanian, green radiolarian shales and radiolarites developed, similar to the adjacent basins. They are covered by variegated shales of the Cenomanian–Turonian age that pass upward into a thick complex (about 700 m [2300 ft]) of red and green marls (Weglowka marls) that are Senonian to middle Eocene (Mitura and Birecki, 1966). In the western part of the Subsilesian unit, intercalations of sandy and conglomeratic complexes of the upper Senonian and/or Paleocene are present. During the late Senonian, gray marls (Frydek marls), commonly with exotic rocks (Burtan et al., 1984), developed in this area. The marly complex passed upward into variegated shales and/or a series of shales and thin-bedded sandstones terminated by green shales and *Globigerina* marls representing the uppermost part of the Eocene. The Oligocene begins, as in adjacent areas, with brown, bituminous shales (Menilite beds) that grade upward into a complex of thick- and medium-bedded, calcareous sandstones and marly shales (Krosno beds). Locally, near the Zegocina village, gray mudstones similar to the Krosno beds of the early Miocene have been found (W. Gorczyk, 2003, personal communication).

Toward the east, the Subsilesian, Late Cretaceous variegated marls extend probably to the Ukrainian Carpathians. A narrow strip of marly deposits at the

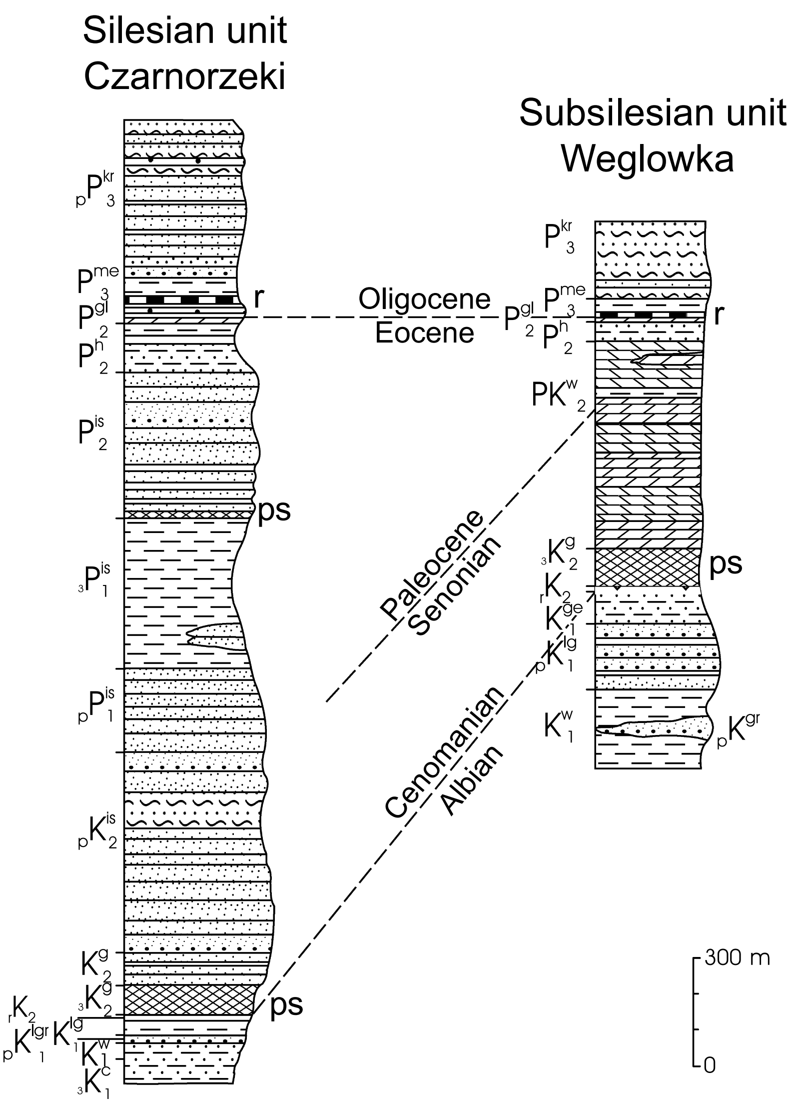

Figure 9. Comparative lithostratigraphic profiles of the Silesian and Subsilesian units, north from the Krosno town. $_pP^{kr}{}_3$ = Krosno beds, sandstones; $P^{kr}{}_3$ = Krosno beds; P_3^{me} = Menilite shales; r = cherts; P_2^{gl} = *Globigerina* marls; P_2^{h} = Hieroglyphic beds; PK_2^{w} = Weglowka marls; ps = red and green shales; P_2^{c} = Ciezkowice sandstones; $_lP_1^{is}$ = upper Istebna shales; $_pP_1^{is}$ = upper Istebna sandstones; $_pK_2^{is}$ = lower Istebna beds; K_2^{g} = Godula beds; $_rK_2$ = radiolarian cherts; K_1^{lg} = Lgota beds; K_1^{ge} = Gaize beds; $_pK_1^{gr}$ = Grodziszcze sandstones; K_1^{w} = Verovice shales; K_1^{c} = Cieszyn beds.

Rozluch scale can represent a prolongation of the Weglowka marls.

To the west, the Subsilesian sediments and especially red marls continue to the Helvetic zone of the Alpine foreland. In the Moravian Carpathians, the Subsilesian lithofacies are partly replaced by sediments of the Zdanice unit (Picha et al., 2006).

Zdanice Unit

For a description of this unit, see Picha et al. (2006).

SKOLE–SKYBA UNIT AND BORYSLAV–POKUTTYA UNITS

The Skole–Skyba units and more external Boryslav–Pokuttya unit (Figures 4, 5, 10) are described together because they developed in the same basin and show generally similar sequences of sediments. The differentiation of both units is based mainly on their tectonic position.

The oldest sediments of the Skole unit (Figures 4, 5) are represented by anoxic black shales, thin, laminated siltstones with layers and lenses of siderites (Spas shales; Barremian–Albian) (Kotlarczyk, 1985; Kruglov, 2001). The subordinate intercalations of fine-grained, thin-bedded, quartzitic, and sparse calcareous sandstones also occur here. The Kuzmina borehole, located southwest from Przemysl, encountered a complex of thick-bedded sandstones several tens of meters thick in the upper part of the Spas shales (Kuzmina sandstones, Borys and Cisek, 1989). At the boundary between the Albian and Cenomanian, a thin layer of green radiolaritic shales developed, which passed upward to the red shales and whitish, siliceous turbiditic marls (Holownia marls). Above them, a thick complex of siliciclastic

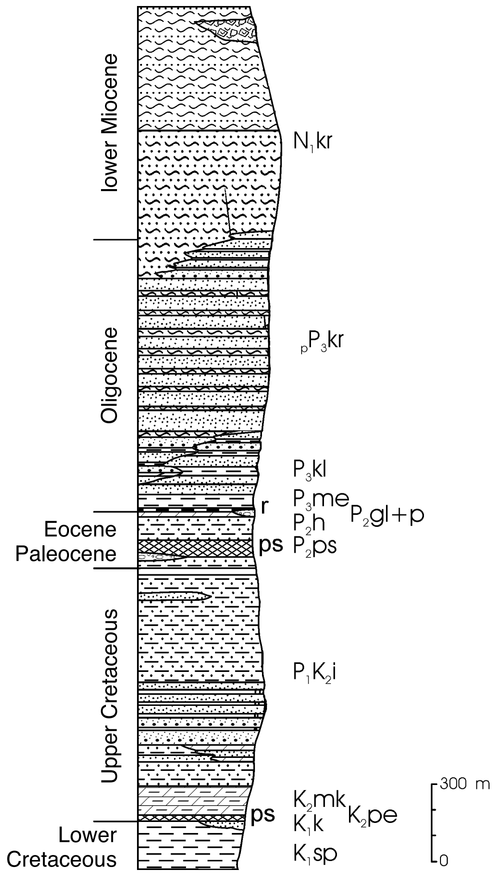

Figure 10. Lithostratigraphic profile of the Skole unit. N_1kr = Krosno beds; $_pP_3kr$ = sandstones, Krosno beds; P_3kl = Kliwa sandstones; P_3me = Menilite shales; r = cherts; P_2gl + p = *Globigerina* marls and Popiele beds; P_2h = Hieroglyphic beds; P_2ps = variegated shales; P_1K_2 i = Inoceramian beds; K_2mk = Holownia marls; ps = red and green shales, radiolarian cherts; K_1k = Kuzmina sandstones; K_1sp = Spas shales.

turbidites is developed (Inoceramian beds = Ropianka Formation [Kotlarczyk, 1978] in Poland, Stryj beds in Ukraine; Turonian–Paleocene; more than 1500 m [5000 ft] thick). It is represented by thin-bedded, calcareous sandstones alternating with gray shales in the lower part. The middle part consists of thick-bedded sandstones, and the upper part consists of thin- and medium-bedded, calcareous sandstones and gray shales. Locally, in the western part of the Skole–Skyba unit, in the uppermost part (Paleocene) of the sequence, lenses of black pebbly mudstones (Babica Clays) are present. Above the turbiditic sequence, a horizon of red and green shales appears (Jaremche horizon; Paleocene). The higher part of the Paleogene displays distinct lateral changes of lithofacies. In the Polish and inner part of the Ukrainian area above the red horizon, developed green and gray clayey shales with intercalations of thin layers of sandstones are present (Hieroglyphic beds in Poland, Witwica beds in Ukraine; late Paleocene–middle Eocene about 150 m [500 ft]). In the outer part of the Ukrainian part of Skyba and in the Boryslav–Pokuttya units, the red horizon is covered by thick-bedded fluxoturbidites and conglomerates (Jamna beds; late Paleocene; as much as 350 m [1150 ft]). These sediments pass upward into thin-bedded turbidites and shales (Maniava beds; early Eocene; as much as 350 m [1150 ft]) covered by a complex of thick-bedded turbidites (Wyhoda beds; middle Eocene; as much as 300 m [1000 ft]) and locally by a series of light limestones and calcareous turbidites (Pasieczna beds; middle Eocene; as much as 200 m [660 ft]). The sedimentation of green shales with subordinate thin sandstones (Bystryci Formation) with local intercalations of gray marls and slump deposits with exotic material (Popiele beds) prevailed during the late Eocene. Marls with *Globigerina* sp. (*Globigerina* marls; late Eocene) terminate these sequences.

The Oligocene sequence begins with dark brown, bituminous shales (Menilite beds) with cherts in its lower part and lenses of thick, coarse-grained, fluxoturbiditic sandstones (Kliwa sandstones) as much as 250 m (800 ft) thick. The Menilite beds pass gradually upward through a transitional zone into a complex of thick-bedded, medium-grained, calcareous sandstones (lower Krosno beds, as much as 800 m [2600 ft] thick). They are overlain by medium- and thin-bedded, calcareous sandstones and gray marly shales. Toward the top of the sequence, sandstones gradually disappear, and the upper part of the Krosno beds is represented mainly by marly shales (upper Krosno Beds; early Miocene). The lithological boundary is diachronous. It migrates upward from the inner to the outer part of the Skole–Skyba unit. The thickness of the Krosno beds exceeds 2400 m (7900 ft). In the outer area of the Ukrainian part of the Skyba unit and in the Boryslav-Pokuttya unit, the dark brown shales are divided by gray marls and thin sandstones (Lopianetz beds; Rupelian–Eggerian; as much as 200 m [660 ft]) into lower and upper Menilite beds. They are covered by gray marly shales and sandstones (Polianycia Formation; Eggenburgian, more than 600 m [2000 ft]). In the Boryslav-Pokuttya unit above the Polanytsia beds, marls and shales (Vorotyshcha beds) are present with evaporates (gypsum and salt) and with lenses of conglomerates (Sloboda conglomerate).

OUTER CARPATHIAN TECTONICS

During the late Oligocene and Miocene orogenesis, several nappes corresponding to the lithostratigraphic units were formed with prevalent northern direction of thrusting in the Western Carpathians, northeastern and eastern in the Eastern Carpathians (e.g., Koszarski et al., 1974; Ksiazkiewicz, 1977) (Figures 1–3, 11–13).

In the Western Carpathians, from the south, they are

1) Magura nappe
2) Fore-Magura Group of nappes that consist of Fore-Magura unit, namely, the Obidowa–Slopnice unit, Grybow unit, and Jaslo unit
3) Dukla nappe
4) Silesian nappe
5) Subsilesian nappe
6) Skole–Skyba nappe

In the Ukraine Carpathians, they are

1) Magura nappe
2) Marmarosh unit
3) Rachiv nappe
4) Porkulets nappe
5) Dukla nappe
6) Chornohora nappe
7) Silesian–Krosno zone
8) Skyba unit
9) Boryslav–Pokuttya nappe

The Magura Nappe

The Magura nappe is the largest tectonic unit of the Outer Western Carpathians (Figure 1) linked with Rhenodanubian flysch of the Eastern Alps (Schnabel, 1992). This unit runs as an arc from the Wienerwald in Austria through Moravia, Slovakian and Polish Beskids, and Eastern Slovakia. It narrows and disappears east of Uzhorod (Transcarpathian Ukraine and adjacent Romania) beneath the Miocene volcanic rocks (Ksiazkiewicz, 1977).

At the east of Latorica River, the prolongation of the Magura nappe is uncertain and speculative. The Monastyrets–Petrova unit of the Marmarosh Klippen zone, situated between the Shopurka and Latorica Rivers, could be regarded as the prolongation of the Magura nappe in the Eastern Carpathians (Sandulescu et al., 1981; Sandulescu, 1988; Bombita and Pop, 1991; Bombita et al., 1992; Oszczypko, 1992; Zytko, 1999; Airoldi, 2001). The eastern termination of this unit is known from the Poiana Botizei area in the Romanian Carpathians.

During the overthrust movements, the Magura nappe has been completely uprooted from its substratum mainly along ductile Upper Cretaceous rocks (Birkenmajer 1986; Oszczypko, 1992). On the base of the facial differentiation of the Paleogene deposits, the Magura nappe has been subdivided into four facies-tectonic subunits: the Krynica, Bystrica, Raca, and Siary (see Koszarski et al., 1974). The Krynica and Bystrica subunits form regional thrust sheets. These separate thrust sheets are especially visible in the western part of the Magura nappe in Slovakia. Slovak geologists use the term ''Magura Group of Nappes'' (e.g., Kovac and Plasienka, 2002). These thrust sheets are also well visible in the adjacent part of Poland in Beskidy west of Babia Gora (Golonka, 1981; see also Zytko et al., 1989). Eastward in the very well-recognized Polish sector of the Magura nappe, the independent subsidiary nappes are hard to distinguish (Ksiazkiewicz, 1977). In the western prolongation of the Magura nappe (Slovak and Czech Republics), the position of the Krynica zone is occupied by the Bile Karpaty zone (Potfaj, 1993). The Magura nappe is separated from the Pieniny Klippen Belt in Poland and the adjacent part of Slovakia by a steeply dipping southward strike-slip boundary (Birkenmajer, 1986; Plasienka, 1999b). Westward, this boundary has a mixture of strike-slip and thrust character (Picha, 1996; Kovac and Plasienka, 2002). The Magura nappe is flatly thrust over its foreland, and made up of the Fore-Magura group of units and partly by the Silesian unit. The amplitude of the overthrust is at least 50 km (30 mi), and the post-middle Badenian thrust displacement is more than 12 km (7 mi) (Oszczypko and Zuchiewicz, 2000; Oszczypko, 2001). The northern limit of the nappe has an erosional character. The subthrust morphology of the Magura foreland is very distinctive. The shape of the northern limit of the Magura nappe and the distribution of the tectonic windows inside the nappe are connected with the denivellation of the Magura basement. As a rule, the embayments of the marginal thrust are related to transverse bulges in the Magura basement, whereas the peninsulas are located in the depressions of the basement (Oszczypko, 2001). The zone of the tectonic windows, which are connected with the uplifted Fore-Magura basement, is located at a distance of 10–15 km (6–9 mi) south from the northern limit of the unit (e.g., Mszana Dolna, Szczawa, Kleczany, Ropa, Ujscie Gorlickie, and Swiatkowa tectonic windows). The Obidowa unit occurs in the tectonic windows, which is regarded as the western prolongation of the Dukla unit (Cieszkowski et al., 1985) and the Grybow unit, known also as the Kleczany–Pisarzowa unit (Kozikowski, 1958). These units are composed predominantly of the late Eocene–Oligocene deposits and sometimes of the Upper Cretaceous–Paleocene deposits. The Mszana Dolna tectonic window is the biggest. It is situated in the central part of the Polish Carpathians. In the Mszana Dolna tectonic window, the Grybow unit is made up of several, south-dipping thrust sheets, composed of hinterland-dipping duplexes (see also Mastella, 1988). This duplex structure had developed during the middle Miocene thrusting, between the Magura and Dukla units, which

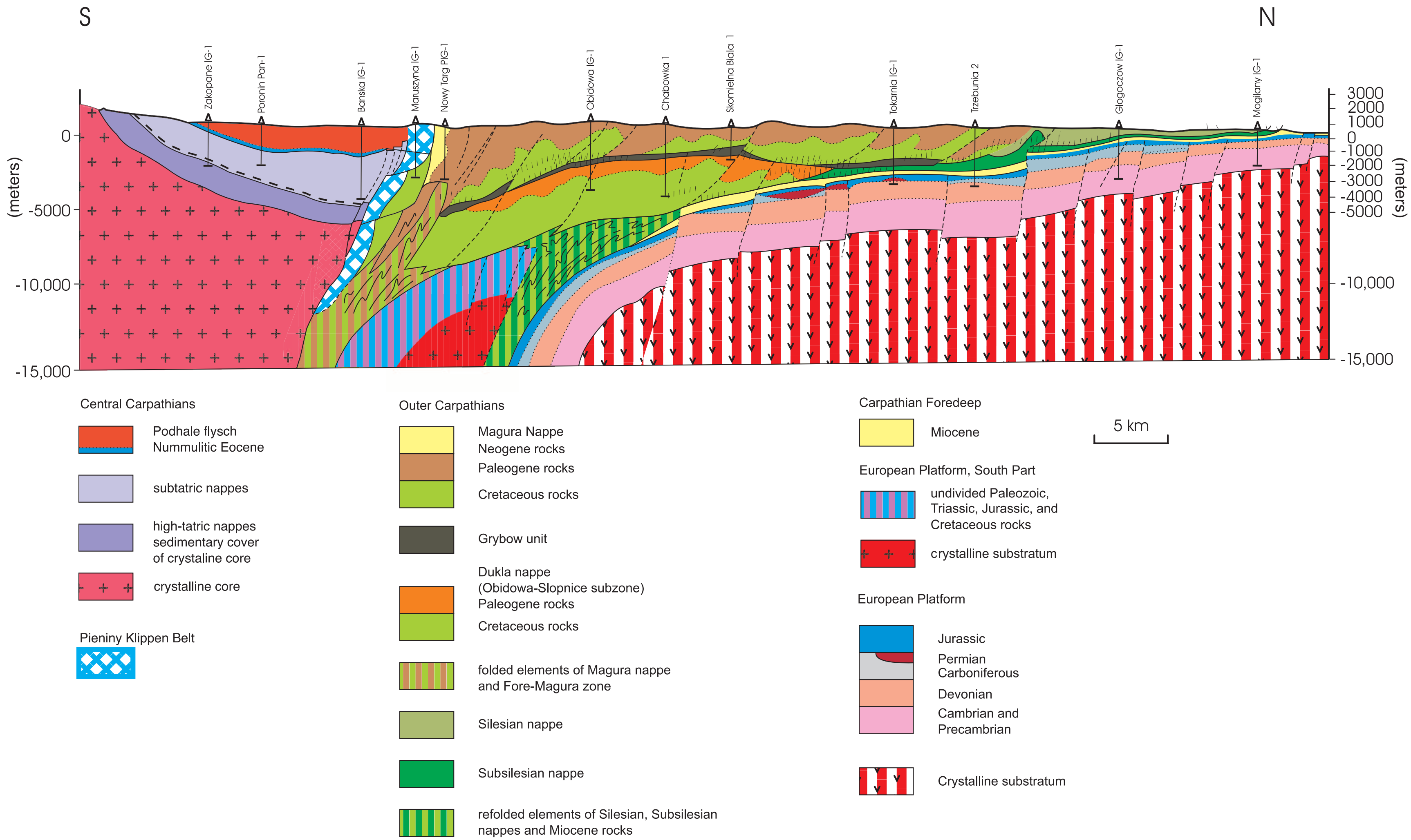

Figure 11. Cross section through the western part of the Polish Carpathians between Krakow–Zakopane (M. Cieszkowski, 2002, personal communication).

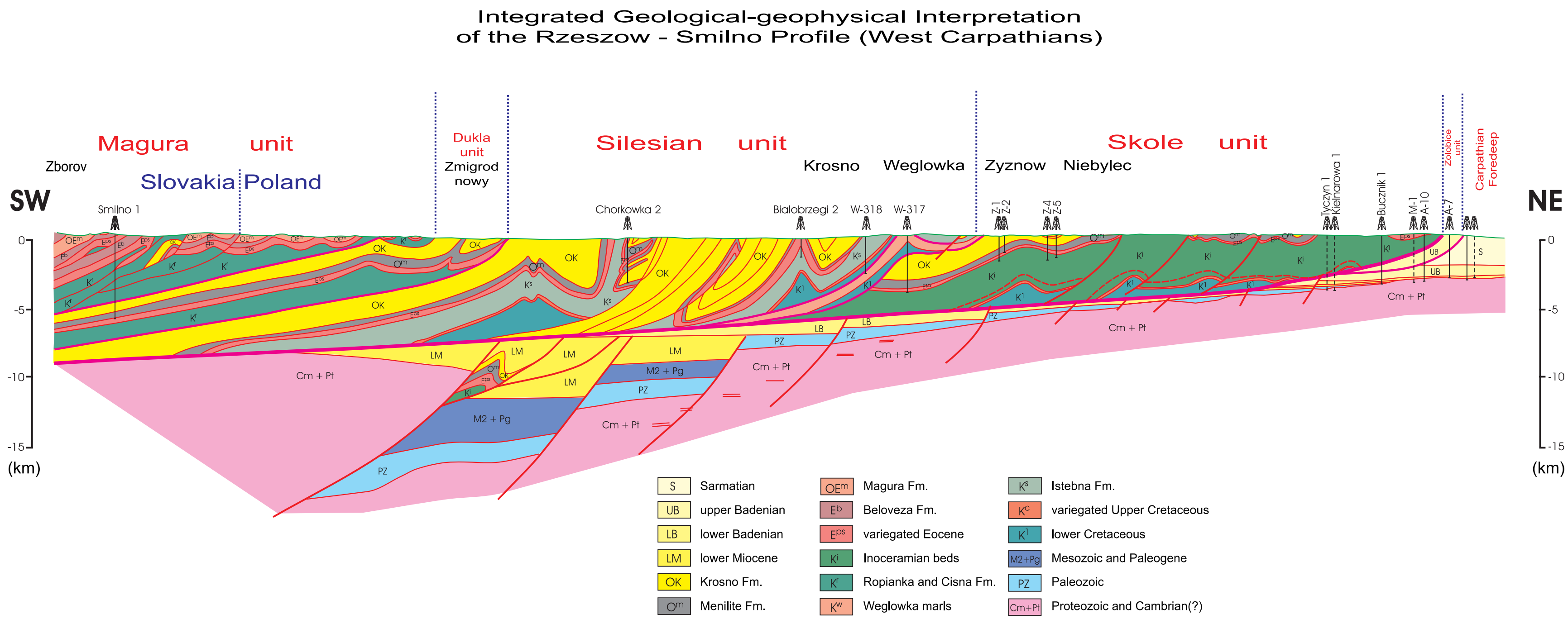

Figure 12. Cross section through the eastern part of the Polish and Slovak Carpathians between Rzeszow and Smilno (modified after Oszczypko et al., 1998).

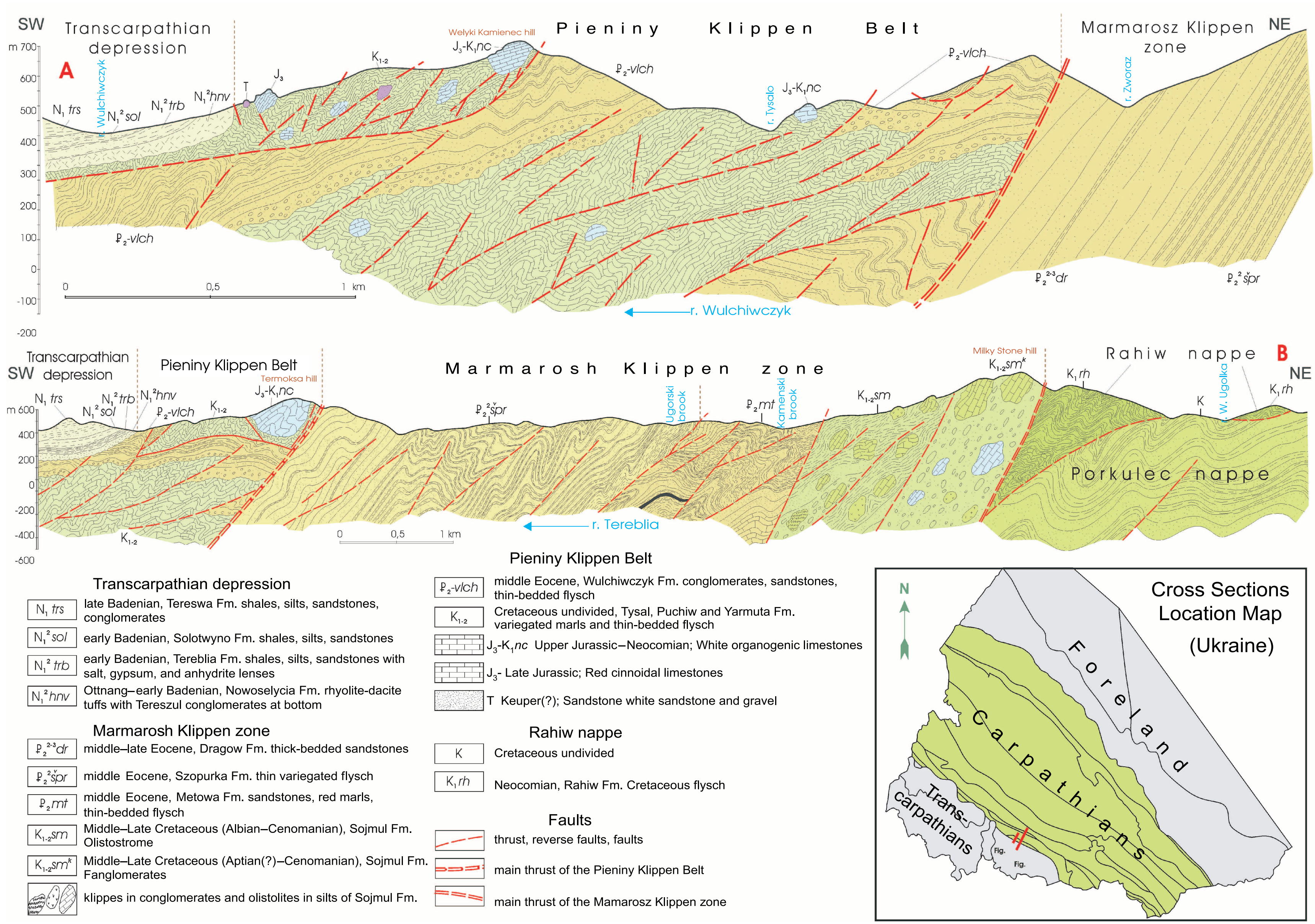

Figure 13. Cross sections through the inner part of the Ukrainian Carpathians.

formed the roof and floor thrust, respectively (Oszczypko, 2001). South of the zone of tectonic windows, the inclination of the Magura thrust surface increases, and at the northern boundary of the Pieniny Klippen Belt, the thickness of the Magura nappe is more than 5 km (3 mi). The Krynica, Bystrica (Nowy Sacz), Raca, and Siary subunits coincide, to a large extent, with the corresponding facies zone. In the Magura unit, three structural complexes can be distinguished. The first was formed during the Late Cretaceous–Paleocene, the second was formed during the early to late Eocene, and the last was formed during the Oligocene to early Miocene (Oszczypko and Thomas, 1985; Oszczypko, 1992, 1999; Malata et al., 1996). These complexes revealed a decreasing degree of tectonic deformation from the base to the top of the nappe. In the area surrounding the Mszana Dolna and Szczawa tectonic windows, the basal part of the nappe, made up of Upper Cretaceous–Paleocene flysch rocks, is strongly deformed. In the lower to upper Eocene flysch of the Raca and Krynica subunits, the broad west–east-trending synclines and narrow anticlines dominate. The southern limbs of synclines are commonly reduced and overturned. In the Bystrica (Nowy Sacz) subunit, subvertical thrust sheets are common. Both the northern limbs of the anticlines and the southern limbs of the synclines are tectonically reduced and commonly overturned. In the Krynica and Raca subunits, the youngest (Malcov Formation, late Eocene–early Oligocene), weakly deformed deposits of the Magura nappe unconformably overlie the older Eocene flysch deposits.

The Biele Karpaty (White Carpathians) were traditionally regarded as an integral part of the Magura nappe and correlated with the Oravska Magura (Krynica subunit), which is in a similar tectonic position with respect to the Pieniny Klippen Belt. According to Potfaj (1993), the Biele Karpaty forms a separate subunit. This subunit is overthrust onto the Raca and Bystrica subunits of the Magura (at least to 25 km [15 mi]). The inner structure of the Biele Karpaty unit has a fold and thrust character in its western part. The biggest thrust, known as the Javorina nappe, forms a flat-lying body approximately 50 km (30 mi) long, about 900 m (2900 ft) thick, and 20–25 km (12–16 mi) wide.

The Marmarosh Unit

The Marmarosh unit stretching along the Pieniny Klippen Belt is divided into two parts:

the Marmarosh Klippen zone and Marmarosh crystalline massif
the Marmarosh Klippen zone

The Marmarosh Klippen zone is situated northeast from the Pieniny Klippen zone; its northwest part is overthrust onto the Rachiv and Porkulets nappes, and its southeast part is overthrust onto the Marmarosh crystalline massif. The surface of the overthrust, traced by several boreholes, is generally steep, and the whole zone dips steeply toward the southwest; only in the area of Chornyj Cheremosh does the overthrust become locally flat, and its visible distance is more than 3.5–4 km (2.1–2.5 mi). The Marmarosh Klippen zone is divided into two subzones on the base of tectonic structures and age of folding: the external subzone of Vezhany, which contains huge olistolithes (Soimul olistostrome), and the internal subzone of Monastyrets, which is built mainly by flysch. The first subzone is made up of a few thrust folds, whereas the second subzone is a large monocline with small secondary folds. The internal subzone is the prolongation of the Petrova nappe and is regarded by some geologists as the southeast prolongation of the Magura unit (Sandulescu, 1988; Oszczypko, 2001).

The Marmarosh crystalline massif is the northwest prolongation of units known from the Romanian Carpathians (Kropotkin, 1991; Golonka et al., 2003) as Bucovinian–Getic nappes (Sandulescu et al., 1981) or Median Dacides (e.g., Stefanescu et al., 2006). It is overthrust on the Black Flysch unit or Rachiv unit and dipping under the Marmarosh Klippen zone. The massif displays a nappe structure, and two subnappes were distinguished (Kruglov, 2001): the Bilyi Potik (the lower subnappe) and the Dilove (the upper subnappe). Minimal displacement magnitude of the Dilove nappe is more than 10 km (6 mi), whereas in the Romanian Carpathians, its displacement of more than 25 km (16 mi) is observed. In front of the massif, the narrow Kaminnyi Potik unit is present, composed of calcareous-terrigenous deposits containing the Upper Jurassic effusives of the basic type. Glushko and Kruglov (1986) consider this thrust sheet as a part of the massif.

The Rachiv Nappe

The Rachiv nappe is represented by a narrow zone made up of secondary folded Lower Cretaceous flysch dipping generally toward the southwest under the Kaminnyi Potok–Marmarosh unit and is overthrust on the Porkulets nappe. This overthrust is very gentle in the Chyvchyny area and becomes very steep to the northwest near the village of Dovhe. Magnitude of the overthrust is apparently great, and its minimal value, according to the existing data, is of several kilometers. The Rachiv nappe forms a direct continuation of the upper internal thrust sheet or the Ciuc digitation of the Ceahlau nappe in the Romanian Carpathians. Toward the northwest, it disappears under Vezany subzone north from the village of Dovhe, and its west continuation is not known.

Porkulets Nappe

The Porkulets nappe is the largest among the internal thrust sheets of the Ukrainian Carpathian Flysch. It overthrusts the Dukla nappe and also the Chornohora nappe farther to the southeast. The magnitude of horizontal displacement of the Porkulets nappe reaches at least 12 km (7 mi) locally. This displacement is evidenced by the existence of a tonguelike tectonic semioutlier in the Mount Petros area and by borehole data. According to the data from the Chornoholova parametric borehole, the thrust surface is steep and can be estimated as 80° in the inner part of the Porkulets nappe. The Porkulets nappe plunges to the southwest, below the fragmented strip of the Rachiv nappe and at the westernmost part below the Magura nappe. The thrust surface is steep, and its dip can be estimated as 80°. The inner part of the Porkulets nappe is characterized by a general submergence of the axis of folds from the southeast, where only Early Cretaceous deposits are exposed to the northwest where younger deposits are also visible. It abruptly terminates near the Slovak border, and its northwest prolongation is not clear.

The four subzones separated by cross-faults are distinguished based on tectonic structures and stratigraphic profiles. They are, from the northwest to the southeast, Chornoholova, Turya Polyana, Lysychiv, and Bila Tisza. A distinctive feature of the Chornoholova subzone is its general elevation and planar structures. Toward the southeast, this subzone terminates on a transverse wrench fault that is manifested as the Turya sigmoid (flexure) in the more external Dukla nappe. The thrust surface of the Turya Polyana subzone is steep and can be estimated as 80°. This zone consists of two thrust folds: the internal Turycia and the external Zvir thrust folds. A wrench fault separates this subzone from the next, more uplifted Lysychiv subzone, the main subzone of the Porkulets nappe. This subzone is made up of several thrust folds secondarily folded and cut by several cross-faults and is terminated by a cross-fault along the Luzanka River.

The Bila Tisza subzone, built mainly of the Early Cretaceous deposits, is located further to the southeast. The distinctive feature of this subzone is the presence of the greenstone basic igneous and tufogenic rocks in its frontal part. The Bila Tisza subzone is strongly tectonized, and its internal structures are still not satisfactorily recognized. It prolongs southeastward in the Romanian territory into the Bodok digitation of the Ceahlau nappe.

Dukla Nappe

The Dukla nappe stretches from the Polish to Ukrainian Carpathians. In its southeastern part, this nappe consists of several imbricated, thrust-faulted folds (Figure 12) with a northwest–southeast strike. The Dukla nappe is maximally elevated in the east. The axis of folds plunges gradually toward the northwest, and eventually, the whole nappe disappears below the Magura nappe. Data from boreholes (Zboj 1, Smilno 1) show that the Dukla nappe extends under the Magura nappe far to the south. Southeast from the Slovak–Ukraine border, where the Magura nappe disappears, the inner part of the Dukla nappe is hidden below the Porkulets nappe. From the more external Silesian nappe and/or Zboj unit, the Dukla nappe is separated by a thrust plane more distinct in the eastern than in the western part. Data from the deep boreholes Jasliska 2 and Wetlina 3 show that the thrust plane in the Polish part of the nappe is very steep. However, data from the deep borehole Zboj 1 indicate that thrust exceeds 15 km (9 mi), and the thrust plane beneath the more internal part of the Dukla nappe becomes more flat.

In the Polish and Slovakian part of the Dukla nappe, two subunits can be distinguished: the internal and the external subunits. The folds in the internal subunit are generally gently dipping toward the southwest and the unit's overthrust is low dipping, whereas in the external subunit, folds are steep and commonly with a reversed (southwestern) vergence. The internal subunit disappears on the border between Slovakia and Ukraine; however, the fact that this subunit prolongs into the Porkulets nappe cannot be excluded.

In the southeastern Ukrainian part, the Dukla nappe has a general direction of submergence from the southeast to the northwest. It is made up of several thrust folds commonly with a southwest vergence. The presence of the gentle brachysynclinal folds is the most distinctive structural feature in the Ukrainian part of the Dukla unit. The best known brachysyncline of the Polonyna Runa was interpreted as a late Oligocene superimposed trough or as an isometric fold formed by an uplift of a crystalline basement at its basis or as a rootless tectonic outlier. Kruglov and Tarasenko (1984) interpret it as a normal brachysyncline with disharmonic folds at its base because of diverse physical properties of the substratum. Three structural-facial en echelon subzones are delimited in the Dukla nappe: the Stuzhycia in the northwest, the median Luzhanka, and the Blyznyca in the southeast. The Dukla nappe disappears from the surface in the valley of Bily Cheremosh, and its continuation toward the southeast is unknown. It could be connected with the Romanian Teleajen nappe, which is assigned to the Moldavides (Sandulescu et al., 1981).

Beneath the Dukla nappe, a separate tectonic unit, the Zboj unit, was described from the borehole Zboj 1, situated in east Slovakia. Only a fragment of a limb of anticline represents this unit. Its internal structures and relation to the more outer tectonic units, especially the Silesian nappe, are unknown.

Fore-Magura Group of Units

The small tectonic units encountered in the Western Carpathians in front or beneath the Magura nappe belong to this group. The Fore-Magura nappe, located near the town of Zywiec, occupies the most western position in the Fore-Magura group of units. It consists of two narrow, asymmetrical anticlines with thrust faults, the inner anticline being strongly deformed and disharmonic. Toward the east, this unit disappears completely. Further toward the east, another unit, the Grybow unit, is exposed in several tectonic windows in the Magura nappe, from the Mszana Dolna tectonic window to the Smilno tectonic window in Slovakia. It was also encountered in several boreholes below the Magura nappe. This unit is strongly folded, with several disharmonic thrust-faulted folds. The Grybow unit is itself thrust over the Obidowa–Slopnice unit or on the Silesian unit. The Obidowa–Slopnice unit is present in several boreholes between the Obidowa and Slopnice (Obidowa IG 1 and Chabowka 1) and in several boreholes in the Slopnice area. The strata of this unit are fairly gently dipping toward the south, commonly without any intense tectonic deformation, except in the higher part. The Jaslo nappe (Koszarski, 1985) forms a rather poorly defined unit existing in front of the Magura nappe to the west of the town of Jaslo. It was distinguished in the Harklova and Luzna peninsulas of the Magura nappe. The Jaslo nappe is flatly thrust over the Silesian nappe with complicated internal structures. However, L. Jankowski's (2003, personal communication) opinion states that rocks regarded as the Jaslo nappe represent olistostromes in the youngest deposits of the Silesian unit. Some of these units (Grybow and Slopnice-Obidowa) display a general affinity and probably belonged to a bigger nappe, which was divided into separate units during the Neogene folding.

Chornohora Nappe

The Chornohora nappe stretches from the Romanian border to the Teresva–Chorna Tisza area, where it disappears under the Dukla nappe. It is connected to the Audia nappe in Romania. Two smaller elements are distinguished here, according to the character of tectonic structures and differences in their lithological development: these are the Skupiv subzone in the north and the Yalovychor, or the Hoverla, subzone in the south. The first one is represented by a complicated secondary folded and faulted monoclinal structure dipping toward the northwest. The more internal subzone is made up of numerous narrow, long, commonly steep thrust folds. It is also called the subzone of small thrust folds. Existing data imply that the Chornohora nappe was displaced more than 10 km (6 mi) to the northeast along a gentle overthrust on the Krosno zone. The Chornohora nappe was completely penetrated at a depth of 3800 m (12,500 ft) by the Hryniava parametric borehole, which was started in the Skupiv zone, 6 km (4 mi) from the nappe front. After penetrating through the Chornohora nappe, the borehole reached into the Oligocene–Lower Cretaceous deposits of the Verhovyna depression belonging to the Krosno zone. The Chornohora nappe thrust surface is dipping at an angle of 30°.

Silesian Nappe

The Silesian nappe stretches from the Moravia area (Czech Republic) to Ukraine, where it loses its individuality. In the western segment of the Polish Carpathians, the Silesian nappe is flatly overthrust onto the substratum (Figure 11). Within the Silesian nappe, several tectonic windows are present where the Subsilesian nappe is exposed. Toward the east, the thrust plane gradually plunges, and at the same time, the character of the tectonic structures changes in the Silesian nappe. In the western part, the structures are generally shallow and gently folded, whereas toward the east, they pass into long, narrow, steeply dipping, imbricated folds. The southern part of the Silesian nappe is hidden beneath the Magura nappe and the Dukla–Fore-Magura nappes (Figures 11, 12).

To the west of the Sola River, near the western border of Poland, the Silesian nappe is composed of two subunits: the Cieszyn subunit is built of a strongly folded Lower Cretaceous strata, whereas the Godula subunit is composed of Upper Cretaceous and Paleogene deposits that dip monoclinally southward. This part of the Silesian nappe is cut by several transverse faults. Farther to the east, the Cieszyn subunit and the Godula subunit join, and the Silesian nappe is built of several gently folded structures. East of the Dunajec River, these structures pass into imbricated folds. The more important structures are the Stroze, Jankowa, and Ciezkowice folds. Small oil fields are associated with these folds. Toward the east, the Stroze fold passes into the broad Gorlice fold (Figure 12), where one of the oldest oil fields in the Carpathians exists. Within the culmination of the next fold to the north, the Ciezkowice–Biecz fold, small oil fields were found. The eastern part of the Silesian nappe, east of the Wislok river, plunges toward the southeast and is represented by a synclinorium (Central Carpathian Synclinorium), which is made up mainly of Oligocene deposits (Wdowiarz, 1985). The Central Carpathian Synclinorium is made of several long, narrow, imbricated, thrust-faulted folds that are commonly disharmonic (Figure 12). These folds are cut by several transverse faults that divide them into separate blocks. The folds display, along the strike, several axial culminations, where, along the northern and southern margins of the synclinorium, the Cretaceous and Eocene strata are exposed. Several folds and thrust folds were distinguished in the Central Carpathian Synclinorium: the Folusz–Bukowica–Fore-Dukla zone, Zboiska, the Lubatowka–Iwonicz Spa–Tokarnia fold, the Osobnica–Bobrka Rogi–Suche Rzeki fold, the Lubienka–Mokre–Zatwarnica fold, the Roztoki–Potok–Turaszowka–Kroscienko–Tarnawa–Wielopole–Czarna fold, the Sanok–Zmiennica–Strachocina–Czarnorzeki fold, and the Ustianowa–Miedzybrodzie–Grabownica fold. Several oil fields are connected with these folds (Slaczka, 1996a).

The Central Carpathian Synclinorium passes into the Krosno zone on the territory of Ukraine. Southeast from the town of Sanok, the amplitudes of marginal thrusts of the Silesian nappe and of the Subsilesian nappe diminish and eventually terminate, and this zone passes into a normal fold. Therefore, the prolongation of the boundary between the Silesian and Skole units toward the southeast is not so clear as it is in the west. The Silesian nappe prolongs on the Ukrainian territory into the Krosno zone.

The boundary between the Krosno zone and the Skyba nappe in Ukraine is generally placed along the large rupture dislocation that stretches along the internal boundary of the Skyba nappe and ends somewhere in the median part of the Ukrainian Carpathians. However, the dislocation does not limit but merely cuts the large synclinorium buildup of thick Oligocene–lower Miocene flysch strata and the intramountain molasse of the Krosno beds.

Within the Krosno zone, three subzones are distinguished according to structural and lithological-facial peculiarities. The most inner subzone (the Bytliana subzone) is characterized by the development of the Oligocene olistostrome formations. It forms a narrow strip along the front of the Dukla nappe and corresponds generally to the Fore-Dukla subunit of the Polish Carpathians. Pre-Oligocene formations are exposed only at the Stryi headwaters in the core of the Smole structure.

The Turka subzone that wedges out near the Black Tisza occupies the intermediate location. It is a nappe thrust over the most external Slavsko–Verhovyna subzone. This subzone is elevated at its southeastern part (at the interflow of the Black Tisza and the Rika) and is submerged at its northwestern part. It was also proposed to relate the area located to the northeast from this nappe (Slavsko–Verhovina subzone) already to the Skyba nappe.

A distinctive feature of the Slavsko–Verhovyna subzone is its general submergence. The subzone is divided into the Slavsko depression in the northwest and the Verhovyna depression in the southeast. These elements are separated by a gentle transversal elevation located between the headwaters of the Tereblia and the Bystrytsia Nadvirnianska.

Andrychow Zone

In the western part of the Outer Carpathians near the town of Andrychow, along the Silesian nappe, several huge blocks exist built mainly by Jurassic limestones. They were regarded as tectonic klippen that were sheared off during the movements of the Silesian nappe (Ksiazkiewicz, 1977); however, new data suggest that they are olistolithes in the uppermost part of the Krosno beds of the Subsilesian nappe (Slaczka, 1998).

The Subsilesian Nappe

The Subsilesian nappe is exposed in a narrow, strongly tectonized zone in front of the Silesian nappe and also in the Western Polish Carpathians in several tectonic windows in that nappe. The presence of the Subsilesian nappe was also established in numerous boreholes beneath the Silesian and the Magura nappes (Figures 11, 12).

In the frontal part of the Silesian nappe, north of the town Krosno, the Subsilesian nappe is exposed in the Weglowka tectonic half window. Deep wells connected with the Weglowka oil field show that the tectonic window is built of a refolded thrust-faulted anticline. The Subsilesian nappe is steeply overthrust onto the Skole nappe. Further to the east, the Subsilesian nappe forms once more a narrow zone in front of the Silesian nappe (Figure 12). Near the town of Ustrzyki Dolne, the Subsilesian nappe disappears from the surface, and the frontal part of the Silesian nappe becomes a thrust-faulted fold and eventually joins with the Skole nappe. A possibility also exists that tectonic prolongation of the Subsilesian nappe is the Rosluch thrust fold in Ukraine.

THE SKOLE–SKYBA NAPPE

The Skole–Skyba nappe forms a large portion of the eastern part of the Northern Carpathians and is overthrust onto the Miocene sediments that cover the North European platform (Figure 12). This nappe is the largest unit in the Outer Carpathians (as much as 40 km [25 mi] wide) and is the only structural unit that stretches from Poland, where it is called the Skole nappe, through the whole of the Ukrainian Carpathians, where it is named the Skyba nappe, and passes into Romania as the Tarcau nappe. The most distinctive structural feature of the zone is the occurrence of large thrust-folds skybas thrust over each other in the northeast direction and traced for several hundreds of kilometers along the stretch of the Carpathian arc. The width of such skybas ranges from a few kilometers to as much as 12 km (7 mi).

Toward the west, the Skole–Skyba nappe becomes more and more narrow, and eventually, the Skole nappe plunges under the Subsilesian and Silesian nappes near the town of Brzesko, and in this area, these nappes reach the Carpathian margin. Its prolongation further toward the west is not clear. However, sometimes, the marginal flysch unit, near the town of Wadowice, is considered as a western prolongation of the Skole nappe.

The outer part of the Skole nappe is elevated and consists of numerous, narrow, steep, thrust-faulted anticlines where the Lower Cretaceous Spas shales appear at the surface (Figure 12). The synclines are narrow and made up of the Menilite or Krosno beds. Southwest from the town of Przemysl, the folds create a sigmoidal arc that reflects a similar sigmoidal bend of the Carpathian margin. Near the town of Rzeszow, the marginal part of the Skole nappe is covered by Miocene molasses, which form a piggyback basin. Farther toward the west, near the town of Pilzno, the northern part of the Skole nappe is probably folded together with the Miocene cover.

The inner part of the Skole nappe is represented by a synclinal area made of several folds (Tyrawa–Lodyna, Paszowa–Wankowa, Witrylow, and Wara) with broad synclines composed of the Oligocene–lower Miocene Krosno beds (Figure 12). In this part of the Skole nappe, the deepest borehole in the Polish Carpathians, the Kuzmina 1 (7541 m; 24,740 ft), was situated. Near the town of Strzyzow, the axis of folds plunges, and synclines join together, creating a vast Strzyzow synclinorium.

The structures observed in the western part of the nappe continue toward the southeast. Six thrust folds (skybas), secondary folded, were distinguished in the Skole–Skyba nappe in Ukraine (external thrust folds: Marginal, Oriv, and Skole; and internal thrust folds: Parashka, Zelemjanka, and Rozhanka). The frontal parts of thrust folds in Skole–Skyba nappe are commonly made up of the Upper Cretaceous deposits (Holovnia and Stryi beds) and sometimes of Paleogene formations. The inner parts of three external thrust folds are made of the Oligocene and the lower Miocene deposits (Menilite shales), whereas the inner parts of the internal thrust folds are represented by the lower Miocene Krosno beds. Two inner thrust folds are distinctly traced only to the Lomnytsia River. Farther to the east, they are hidden below the Oligocene lower Krosno beds. The Marginal (Berehova) thrust fold disappears in the area of the Pokuttya Carpathians, where it was completely destroyed by erosion. Hence, only two thrust folds are distinctly traced up to the Romanian border: the Oriv and the Skole thrust folds. The continuation of the more inner thrust folds is not very clear.

The Skole–Skyba nappe is thrust over the Boryslav–Pokuttya nappe in the east and over the folded and/or autochthonous Miocene covering the North European platform in the west. The thrust plane of the Skyba nappe is different in diverse regions and changes from very gentle (where the thickness of the nappe is small) to very steep. Especially, a steep thrust plane exists at the most internal part of the nappe, where it has been discovered in recent years by superdeep boreholes. In the Bytkov region, the magnitude of a vertical displacement of the Skyba nappe exceeds 17 km (11 mi). However, in the west, the thrust plane is more flat. The Kuzmina 1 borehole (Figure 2), situated 30 km (18 mi) from the northern boundary of the Skole–Skyba nappe, reached the thrust plane at a depth of 6399 m (20,994 ft). Further to the west, in borehole Szufnarowa 1 (Figure 2) situated 15 km (9 mi) from the northern boundary, the thrust plane was encountered at the depth 4000 m (13,000 ft).

The Skole nappe is locally covered by Miocene molasses, which form piggyback basins. Near the town of Pilzno, the northern part of the Skole nappe is probably folded together with the Miocene cover.

Boryslav–Pokuttya Nappe

This nappe developed only in the Eastern Carpathians. Within the Ukrainian segment, it is commonly designated as the Boryslav–Pokuttya nappe, and on Romanian territory, it is designated as the Marginal fold unit. This nappe is composed of Cretaceous–lower Miocene flysch, as well as early Miocene molasse. Therefore, that unit is commonly regarded as the inner foredeep zone.

The Boryslav–Pokuttya nappe consists of four main thrust folds: the Boryslav, the Maydan, the Pokuttya, and the Runhur (Glushko and Kruglov, 1986). They have a very complicated, refolded, internal structure typical for duplexes. The majority of the folds display a northeast vergence as well as an absence of tucked-in limbs. Very commonly, two or three folds are overthrust one on top of the other and display stacked structures. The presence of thick clay and salt-containing deposits substantially complicates the local structures of those folds. Sometimes, two or three digitations of such folds (partial thrust sheets) could be delimited in the profile. Sometimes, a diagonal strike of the main folds in relation to the main structure of the nappe could be observed, and five to seven separate groups of such folds are distinguished. The largest groups of folds are the Blazhiv, the Boryslav, the Dolyna, the Bytkiv, and the Pokuttya–Dzviniany.

The Boryslav–Pokuttya nappe is overthrust on the folded Miocene [Sambir (Stebnik) nappe] laying on the autochthonous Miocene of the Ukrainian platform (Kruglov, 1999). The thrust plane is very steep or even overturned near the surface, but at depths of 3–4 km (2–2.5 mi), it became almost flat. The overthrust, documented by boreholes, is of a magnitude of at least 20 km (12 mi). The continuation of the Boryslav–Pokuttya nappe below the Polish part of the Skole–Skyba nappe is still controversial. No real traces of this structure have been present. If it exists, it is hidden far to the south below the Carpathians.

THE DEEP STRUCTURE OF THE OUTER CARPATHIANS

The deep structure of the Polish Outer Carpathians and its basement has been recognized by deep borehole results as well as by magnetotelluric, gravimetric, magnetic, geomagnetic, and deep seismic sounding profiles (Woznicki and Sucha, 1989; Guterch et al., 2001; Stefaniuk, 2002; Stefaniuk and Slaczka, 2002).

Deep boreholes (as much as 9000 m [29,500 ft]), which reached the Carpathian substratum, were drilled along the Outer Carpathians (Figures 2, 11, 12) (see also Oszczypko et al., 2006; Picha et al., 2006). They allowed the recognition of the deep structures of the Carpathians, the depth of Carpathian thrust plane, its minimal range, as well as the character of the substratum. First of all, they proofed the thin-skinned character of the Outer Carpathians orogen, which is thrust over the autochthonous Miocene deposits covering the eastern and western parts of the North European platform. They also documented the occurrence of several uprooted nappes thrust upon each other and the existence of new tectonic and lithostratigraphic units that were not known from the surface data.

Generally, the thrust plane of the Carpathians dipped slowly to the south in their western part (see Oszczypko and Tomas, 1985). The borehole Bystre IG 1 located 30 km (18 mi) southward from the northern margin of the Carpathians crossed this plane at the depth of −3131 m (−10,272 ft) below sea level. Toward the east, this plane plunges gradually. West of Krakow, borehole Zawoja 1, also situated 30 km (18 mi) from the Carpathian margin, crossed the plane on the depth of −3225 m (−10,580 ft) below sea level. In the eastern part of the Polish Carpathians (Oszczypko and Tomas, 1985), borehole Brzozowa 1, situated 10 km (6 mi) from the margin, reached substratum at the depth of −2575 m (−8448 ft) below sea level. Borehole Szufnarowa 1, 15 km (9 mi) from the margin, passed the plane at depth of −3455 m (−11,335 ft) below sea level, and borehole Kuzmina 1, 25 km (16 mi) from the margin, reached the plane at the depth of −6885 m (−22,588 ft) below sea level. The southeast plunging of the thrust plane of the Carpathians continues also in the Ukrainian Carpathians, where borehole Schevthenkovo 1, situated 16 km (10 mi) from the Carpathian margin, did not reach the Carpathian flysch up to the depth of 7520 m (24,672 ft).

Seismic data provide similar values of the depth of the thrust plane. These seismic data were obtained from hundreds of reflection and refraction profiles crossing the Outer Carpathians, especially along their outer part.

Because of data from deep boreholes, new tectonostratigraphic units were found, which were hidden completely below the nappes and not known previously from the surface data. Part of them are an important factor in the hydrocarbon exploration. In the middle part of the Outer Carpathians (Figure 2, 11), several boreholes, e.g., Obidowa IG 1, Chabowka 1, and Slopnice 20, encountered a new unit (Obidowa–Slopnice) completely hidden below the Magura nappe. In addition, in the eastern part of Slovakia, the borehole Zboj 1 came across an unknown unit (Zboj) below the Dukla nappe.

The magnetotelluric soundings in the Polish Carpathians revealed a high resistivity horizon connected to a consolidated crystalline basement (Kucharski et al., 1990; Rylko and Tomas, 1995; Stefaniuk, 2002). According to these authors, the depth of the crystalline basement ranges from 4 to 8 km (2.5 to 5 mi) in the northern part of the Carpathians, dips to approximately 15–20 km (9–12 mi) at its deepest point, then rises to about 8–10 km (5–5 mi) in the southernmost part of the orogen. The axis of the basement depression is located along the line Orava–Nowy Targ–Krynica–Dukla–Jablonka and more or less coincides with the axis of the gravimetric minimum. The obtained results are generally in agreement with the seismic and borehole data.

The results of gravimetric studies show distinct gravity minimum (as much as 60 mGal) along the Carpathian Flysch. The axis of this anomaly runs approximately parallel to the marginal Carpathian thrust in the west, along the northern boundary of the Pieniny Klippen Belt. East from the town of Nowy Targ, it shifts toward the northwest along a perpendicular dislocation and diagonally crosses the structures of the Outer Carpathians. In the Krosno area, the en echelon relocation of the axis toward the north is observed, probably along a strike-slip fault. Further eastward in the Ukrainian Carpathians, the axis runs toward Staryj Sambir and almost approaches the outer margin of the Carpathians. That gravity minimum is related either to the presence of low-density molasse deposits below the Carpathians or to the structures in Moho (Woznicki and Sucha, 1989). The integrated geophysical modeling along the Rzeszow–Bardejov geotraverse (Figure 12) (Oszczypko et al., 1998) documented that regional gravimetric low south of Krosno is a result of a combined effect of the thick Carpathian nappes, which cover thick early Miocene molasses, and the occurrence of the Mesozoic and Paleogene deposits related to the passive margin of the European platform.

The geomagnetic soundings provided the interesting data that helped to recognize the Carpathian substratum. They revealed the presence of a zone of zero values of the Wieses vectors (see Woznicki and Sucha, 1989) that runs south from the Pieniny Klippen Belt near Kroscienko upon Dunajec; westward, it crosses the klippen zone and runs beneath the inner part of the Outer Carpathians. This zone of zero values is connected to a high conductivity body 2.5–6 km (1.5–3.7 mi) thick, at a depth of 15–30 km (9–18 mi). It probably indicates the position of the southern extent of the North European platform and its contact with the Inner Carpathian basement (see Stranik et al., 1993) and probably the southern extent of the Outer Carpathian overthrust. The observed data in the Eastern Carpathians imply a change of the depth of the asthenosphere that sinks from 80 km (50 mi) in the west to more than 150 km (93 mi) in the Eastern Carpathians.

The deep seismic profile located in the Slovakian Carpathians (see Tomek, 1993) gives an approximate thickness of the lithosphere and depth of the Moho below the Carpathians. It shows that two groups of south-dipping reflectors north of Pieniny Klippen Belt exist that probably are related to the middle Miocene subduction of the Moldavidic part of the Severin–Moldavidic realm (see also Tomek and Hall, 1993). The upper reflection between 1 and 3 s (about 4.5–8 km [2.8–5 mi]) probably belongs to a plate boundary between the upper plate (the thrust Penninic realm, mainly the Magura wedge) and the accretionary wedge complex (Moldavides or Dukla–Silesian–Subsilesian group of units). The lower reflectors represent the crystalline basement of the lower plate (North European plate) and its sedimentary cover.

The position of the crust–mantle boundary (Moho) has been recognized along several seismic profiles (Woznicki and Sucha, 1989; Guterch et al., 2001). The depth to the Moho discontinuity generally ranged from 30 to 40 km (19 to 25 mi) at the front of the central part of the Carpathians and increases to 50 km (31 mi) south of the Nowy Sacz. South of the Pieniny Klippen Belt, this value decreases to 36–37 km (22–23 mi) and to only 24–30 km (15–18 mi) beneath the Pannonian Basin. The consolidated basement with velocity $V_p > 6.0$ km/s (3.7 mi/s) is situated at a depth of 10–18 km (6–11 mi) beneath Carpathians and 5–8 km (2.8–5 mi) beneath the Pannonian Basin. The Moho discontinuity in Ukrainian part of the Outer Carpathians is determined at a depth of about 60 km (37 mi), with the P-wave velocity at 8.25 km/s (5 mi/s). It implies the existence of a cross-fault between the Western and Eastern Carpathians. The seismic data show the existence of a low-velocity zone beneath the Tatra that can be connected with the prolongation of the Outer Carpathians flysch below the Tatras. According to Lillie et al. (1994), the thickness of the lithosphere in the Polish Carpathians ranges from 160 km (99 mi) near Krakow to 100 km (62 mi) in the Pieniny Klippen Belt, with a temperature at the depth of the Moho discontinuity ranging from 400°C beneath the Pieniny Klippen Belt to 660–700°C beneath the Outer Carpathians.

Jurassic to Early Miocene Tectonostratigraphic Evolution of the Outer Carpathian Flysch Basins

The Outer Carpathian basin (Figure 14) was created during the course of disintegration of the North European plate. It was connected with rifting and a formation of thinned or oceanic crust in the Penninic realm–Magura basin (Early–Middle Jurassic) and also in the Severin Moldavides realm–Outer Dacides and Silesian basin (Jurassic–Cretaceous boundary) (see Birkenmajer, 1986; Rakus, et al., 1988; Sandulescu, 1988; Tolman, 1990; Dercourt et al., 1993; Golonka et al., 2000, 2003, 2004, 2006). During the sedimentary-tectonic evolution of the Outer Carpathian basins, several periods can be distinguished.

The first period began from the incipient stage of rifting of the southern part of the European plate and the formation of local basins (Early Jurassic–Kimmeridgian). The new configuration of plates replaced the older one, represented by the Transylvanian domain. The Transylvanian oceanic basin developed during the Triassic

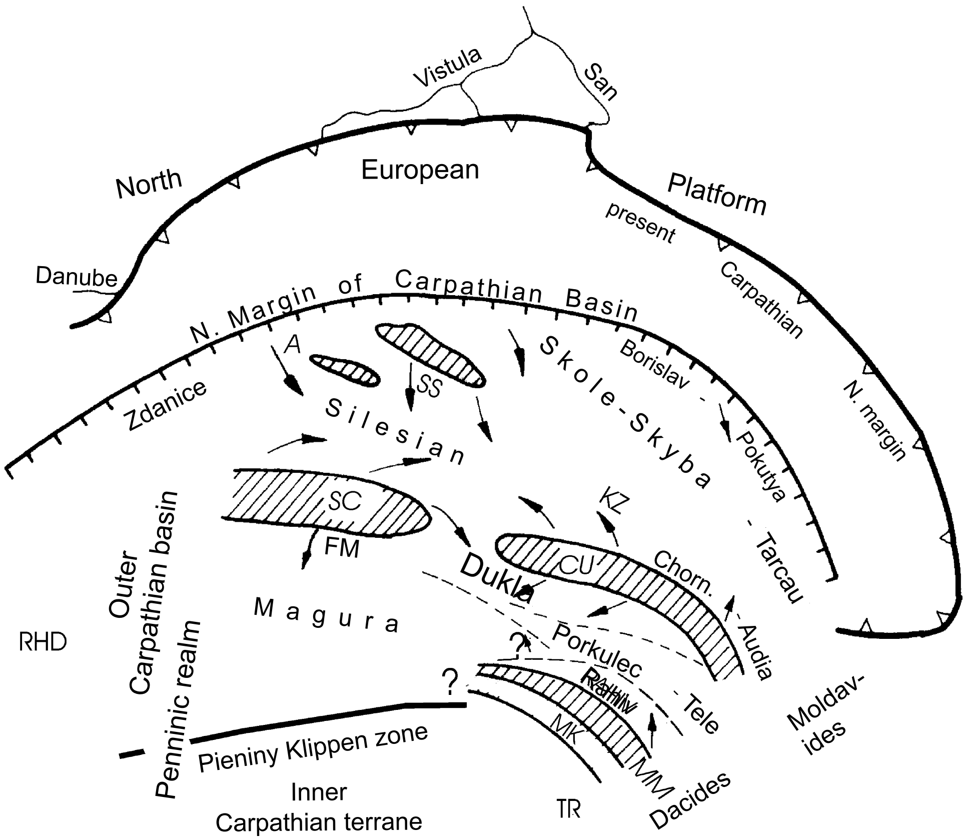

Figure 14. Sketch of probable distribution of ridges and subbasins in Outer Carpathian basin, partly palinspastic (based on Slaczka, 2000). A = Andrychow ridge; CU = Kuman ridge; FM = Fore-Magura basin; MK = Marmarosh Klippen; KZ = Krosno zone; MM = Marmarosh massive; RHD = Rhenodanubian flysch; SC = Silesian Cordillera; SS = Subsilesian realm; Chorn. = Chornohora subbasin; Tele. = Teleajen subbasin.

and was closed during the Cretaceous. The exotic material of the Late Triassic pelagic spotty limestones, which occur as pebbles in Cretaceous–Paleogene gravelstones in the Pieniny Klippen Belt (Birkenmajer et al., 1990) and Magura unit (Sotak, 1986), could have originated in this basin. The Transylvanian basin position and its relation to the other parts of the Tethys and Vardar Ocean, Meliata–Halstatt Ocean, Dobrogea rift, and Polish–Danish Aulacogen remain quite speculative. The Penninic domain, which belongs to the Alpine Tethys, was opening during the Early–Middle Jurassic. The Alpine Tethys, that is, the Ligurian, the Penninic, and the Pieniny Klippen Belt and Magura oceans, constitutes the extension of the Central Atlantic system. Bill et al. (2001) date the onset of oceanic spreading of the Alpine Tethys by isotopic methods as Bajocian. According to Winkler and Slaczka (1994), the Pieniny data fit well with the supposed opening of the Ligurian–Penninic Ocean. During the Early Jurassic rift stage, the single basin existed in the future Pieniny–Magura realm. During the Bajocian–Bathonian, the Czorsztyn ridge originated. Plasienka (2002) postulated the thermal uplift above the distal, subcrustal part of detachment fault. The origin of the Czorsztyn ridge is coeval with the spreading phase of the Pieniny–Magura Ocean. However, it should be stressed that the existence of the oceanic crust beneath the Magura basin could be disputable (Winkler and Slaczka, 1994). The occurrence of the mafic (basalt) intrusions in the eastern termination of the Czorsztyn ridge in Novoselica Klippen (Lashkevitsch et al., 1995) seems to support the thermal origin of the ridge related to the oceanic spreading. The orientation of the Pieniny Ocean was southwest–northeast (see discussion in Aubrecht and Tunyi, 2001; Golonka and Krobicki, 2001). The Pieniny–Magura Ocean was divided into the northwestern and southeastern branches. The deepest parts of both basins are documented by deep-water, extremely condensed, Jurassic–Early Cretaceous pelagic limestones and radiolarites. The shallowest ridge sequences are known as the Czorsztyn succession. In this succession, the Early Jurassic *Posidonia* marls are followed by Middle Jurassic–earliest Cretaceous crinoidal and nodular limestones and Late Cretaceous Ammonitico rosso and marly facies. The transitional slope sequences between the deepest basinal units and ridge units consist of mixed cherty, limestone, and marly facies. The northeastern part of the basin developed later into the Magura basin. The detailed study of the basinal facies (Golonka and Sikora, 1981) revealed an enormous condensation of *Nannoconus* limestones and radiolarites. The Upper Jurassic–Lower Cretaceous profiles do not exceed more than a dozen or so meters. In the external case, 2 m (6.6 ft) was deposited during a time span of 50 m.y. This Jurassic Ocean was connected with the older, Triassic Transylvanian Ocean. A junction of the eastern and Atlantic Tethys existed perhaps in the Eastern Slovakian–Ukrainian Carpathians and is represented by the Inacovce–Krichevo unit (Sotak et al., 2000). The opening of the Severin–Moldavidic basins was related to the gradual closing of the Transylvanian Ocean, which disappeared entirely during the Cretaceous. Shallow-water, calcareous sediments prevailed in the Severin–Moldavidic realm. Riftogenic magmatic rocks were connected with the Jurassic period: calc-alkaline rocks (andesites of Zegocina; Slaczka, 1998) and basic, ultrabasic rocks, e.g., Biala Woda, Kaminnyi, Potok, and Poiana Botizei, along the southern margin of the platform.

The next stage (Tithonian–Valanginian) is characterized by a rapid subsidence of local basins (Slaczka, et al., 1999), where calcareous flysch sedimentation started (Cieszyn area; Cieszyn beds and Eastern Carpathians Kamiennyi Potik–Black Flysch and Rachiv units; Slaczka, 1976). This calcareous detritus was probably derived from the erosion of the reef carbonate platform and deposited by turbidity currents. Pelagic, deep-water sedimentation occurred mainly in the Magura basin at the same time. The Penninic–Pieniny–Magura Ocean reached its maximum width in the latest Jurassic and then stopped spreading. Subduction jumped to the northern margin of the Inner Carpathian terranes and began to consume the Pieniny Klippen Belt Ocean (Birkenmajer, 1986; Golonka et al., 2000). The closing process of the Pieniny Klippen Belt basin is a matter of discussion, as summarized by Golonka and Krobicki (2001) according to earlier elaborations (e.g., Birkenmajer, 1986; Plasienka, 1999; Golonka et al., 2000). The latest Jurassic blueschist metamorphic rocks found as pebbles (exotics) in the Albian flysch in the Pieniny Basin indicate the existence of a subduction below the northern margin of the Inner Carpathian plate (Fayrad, 1997).

The following stage (Valanginian–Aptian) discerned the continuous spreading of the basin and regional subsidence. Sedimentation of black shales embraced almost the whole Outer Carpathian basin with local submarine clastic fans. Clastic material was derived or from the outer margin of the basin or from the uplifted remnants of the North European platform in the basin. Before the Albian, the Magura basin was separated from the Silesian basin by the Silesian submerged ridge (latest Jurassic–earliest Cretaceous). Basalt rocks started to intrude in the Eastern Carpathians (Kamiennyi Potik–Black Flysch; see Sandulescu, 1975), and diabase, teschenites, and syenites started to intrude in the Western Carpathians (Ksiazkiewicz, 1977). In the latter area, it lasted up to the Aptian (Lucinska-Anczkiewicz et al., 2002).

The next stage (Aptian–Albian) is characterized in the Eastern Carpathians by compressional movements that caused folding and overthrusting in the Marmarosh area during the Aptian and development of huge olistolithes in Marmarosh Klippen zone. The existence of Late Jurassic metamorphosed limestones also suggests periods of metamorphism during that time. It is still not certain whether the folding processes also embraced more outer areas (the Cordillera in front of the Dukla subbasin and Silesian Cordillera). On the western and more outer subbasins during the Albian, acceleration of the turbiditic sedimentation occurred. Only in the Skole–Skyba zone was continuation of black siliceous shales observed. At the beginning of the Aptian, a ridge between Silesian–Subsilesian subbasins and Skole subbasin developed, which gave a material to gaize sandstones in the Subsilesian area. South-dipping subduction was active on the southern margin of the Magura basin. Consumption of the Penninic–Pieniny Ocean led to the development of accretionary prism in front of the moving plates. In the Albian, synorogenic flysch developed in the Magura basin (Golonka and Sikora, 1981).

The following period lasted during the Cenomanian–Turonian. It was characterized by subsidence of all the subbasins, cessation of clastic sedimentation, slow and uniform sedimentation, and well-oxygenated conditions. It marked the beginning of a period of intensified, well-oxygenated, deep-water circulation in Tethyan and Proto-Atlantic Oceans (Slaczka and Kaminski, 1998). This period was characterized by very low rates of sedimentation (0.5–5 m/m.y.; 1.6–16 ft/m.y.; Poprawa et al., 2002). At the turn of the Cenomanian, radiolarian shales followed by red clays with intercalations of basinal turbidites were deposited below the CCD (Oszczypko, 1999). The oxic conditions prevailed, and the appearance of the red and green shales like the Malinowa Formation in the Magura basin (Birkenmajer and Oszczypko, 1989) is characteristic. In the northern and middle part of the Magura basin, this type of sedimentation persisted up to the Campanian, whereas in the Krynica zone, this type of sedimentation persisted up to the Maastrichtian (Oszczypko, 2001). The rate of sedimentation of variegated shales oscillated between 15 and 25 m/m.y. (49 and 82 ft/m.y.). In the northern part of the Outer Carpathian, in the Skole–Skyba subbasin, this type of sedimentation terminated during the Turonian (Kotlarczyk, 1985).

The Late Cretaceous–early Miocene period is characterized by compressional movements, appearance of intensive turbiditic sedimentation, and increased rate of subsidence in the basins. Late Cretaceous Laramian tectonic movements, very intensive in the Inner Carpathians, resulted in the enlivening of intrabasinal source areas and in the creation of several sedimentary subbasins, each with specific deep-water clastic sedimentation. The uplift of the southern margin of the North European platform resulted in a tectonic inversion, a general regression of the sea, and erosion. The eroded material supplied the Skole subbasin, and the first appearance of intensive clastic sedimentation during the Turonian occurred in this outermost subbasin.

In the more southern subbasin, the Silesian subbasin, clastic sedimentation appeared on the Turonian–Coniacian boundary in its western part. The clastic material was delivered from the Silesian ridge (Cordillera) and rejuvenated along a system of longitudinal faults, and several clastic submarine fans or faulted submarine slope aprons developed, each with a specific heavy-mineral composition (Wieser, 1970). Probably, the Silesian Cordillera was, in reality, a system of ridges divided by more local basins, some of which later became the Fore-Magura units. The most coarse-grained sediments were deposited along the southern deepest part of the Silesian subbasin. Toward the north and northeast, these deposits were gradually replaced, at first by variegated shales, and eventually by variegated marls that deposited on a slope of the Subsilesian ridge. This slope, also known as Subsilesian basin, was occupied by a pelagic marly sedimentation from the Coniacian–Santonian to Paleocene. Sedimentation of variegated marls also occurred on the southern slope of the Silesian subbasin (Slaczka and Gasinski, 1985) that belonged to the Silesian ridge system, which separated the Silesian subbasin from more inner Fore-Magura subbasins. A diachronous migration of the first appearance of clastic material in the Silesian subbasin from the west toward the east occurred, near the Polish–Ukrainian border; it started during the Santonian–Maastrichtian, and it was probably connected with the migration of the uplifting of the Silesian Cordillera from the west toward the east. In the Dukla and Porkulets subbasins, clastic sedimentation started during the Coniacian from the source areas that were created along the northeastern margin of the Dukla subbasin on the prolongation of the Silesian Cordillera and continued toward the southeast.

During the Maastrichtian–Paleocene, a considerable reorganization of the Magura basin occurred. This was connected both with the compression at the southern margin of the basin and an inversion of the northern margin (Silesian uplifted ridge system). It was accompanied by a deposition of the upper Senonian–Paleocene flysch (the so-called Inoceramian beds). The rate of sedimentation oscillated between 50 and 75 m/m.y. (160 and 250 ft/m.y.). This clastic sedimentation started in the Magura basin during the Campanian along its northern margin and during the Maastrichtian along the southern margin, where clastic material was delivered generally from the south (Oszczypko, 1975). At the convergent, southern margin of the Magura basin, subduction of the oceanic and thinned continental crust beneath the Czorsztyn ridge occurred during the Maastrichtian and Paleocene, and the Pieniny Klippen Belt was formed. As a result, the Magura basin became trenchlike in its character.

The intensive sedimentation from uplifting source areas lasted generally up to the end of the Paleocene, except in the Silesian subbasin, where it lasted until the late Eocene. Such distribution of the first period of clastic sedimentation implies that, in response to the tectonic movement in the Inner Carpathians, several bulges were created in their foreland, earlier in more distal parts and consequently later and later toward the inner margin of the Outer Carpathian basin. Since the Paleocene–early Eocene, the accretionary prism began to develop in the southern margin of the Magura basin (Oszczypko, 1992, 1999, 2001), close to the fold and thrust Pieniny Klippen Belt. The moving load in front of this accretionary prism has caused subsidence and a progressive northward shift of depocenters. The early Eocene axis of deposition was located in the Krynica zone and then, during the middle and late Eocene, it migrated northward, toward the Bystrica and Raca zones, respectively. In this part of the basin, a narrow and very long submarine fan (a few hundred kilometers) was formed. The clastic immature material of the fan was supplied from a southeast direction and was derived from the erosion of the exposed part of the accretionary prism. During the early to middle Eocene, the deepest part of the basin, commonly beneath the CCD, was located in the northern part of the basin. The rate of sedimentation varied from 10–15 m/m.y. (32–50 ft/m.y.) on the abyssal plain to 75–100 m/m.y. (246–330 ft/m.y.) in the outer fan and between 200 and 300 m/m.y. (660 and 1000 ft/m.y.) in the area affected by the middle fan-lobe system (Oszczypko, 2001). Simultaneously, along the northern margin of the basin (Siary zone), small fans developed. In the Moldavidic realm during the early and middle Eocene, the intense tectonic movement ceased, and the remaining Outer Carpathian subbasins were dominated by a hemipelagic deposition (green and variegated shales) with an intercalation of thin-bedded turbidites. Only locally within the southeast part of the Silesian and in Dukla subbasins did sporadically clastic fans develop. Sedimentation in the Magura basin was strongly influenced by the development of an accretionary prism along its inner margin (see also Misik et al., 1991).

In the late Eocene, the southern part of the Magura basin was affected by tectonic movements (see Ksiazkiewicz and Lesko, 1959; Oszczypko and Zytko, 1987; Royden and Baldi, 1988; Seifert, 1992) and connected with the continuous subduction of the southern margin of the basin beneath the Pieniny Klippen Belt (Oszczypko, 1992). At the turn of Eocene, in the whole Outer Carpathian basin, unification of the sedimentary condition occurred, and green shales and, later on, *Globigerina* marls were deposited.

As a result of the late Eocene tectonic movements, the Carpathian basins were partly isolated from the Mediterranean area. This was the beginning of the Early Parathethys formation (Kovac et al., 1993). This isolation caused the deposition of the black shales that are rich in organic matter (Menilite shales). This type of anoxic sedimentation persisted in the Skole basin until the Oligocene–Miocene boundary. The Magura accretionary wedge was completed before the late Oligocene. The late Oligocene folding and northward thrusting in the Magura nappe was almost contemporaneous with that of the Northern Calcareous Alps and Rhenodanubian flysch. Simultaneously, the southern part of the Magura nappe was transformed into the piggyback basin flooded during the Aquitanian–early Burdigalian (Oszczypko et al., 1999b; Oszczypko and Oszczypko-Clowes, 2002). At this point in time, the front of the overriding Magura nappe reached the Silesian ridge, which collapsed as a result of the lithosphere flexure, and in the southern part of the Silesian subbasin, in front of the advancing nappe, huge olistolithes were deposited (Slaczka and Oszczypko, 1987). The Outer Carpathian flysch basin attained features of a relatively deep-water foreland basin (Golonka et al., 2000). During the Oligocene, the main area of subsidence and deposition of the Menilite and Krosno beds were shifted from the Dukla to the Silesian subbasin (Slaczka, 1969). In the next stage (early Burdigalian), the axis of deposition was transferred to the Skole; the frontal part of the Silesian–Subsilesian nappes became uplifted, and the inner part subsided, creating probably locally piggyback basins. Blocks of Weglowka marls slipped into the Skole subbasin from the uplifted Subsilesian nappe. In the Boryslav–Pokuttya residual flysch basin, the turbiditic deposition was replaced by salt-bearing molasses deposits (Worotysche Formation). Sedimentation in this basin was probably completed during the Ottnangian, whereas the whole former flysch basin was uplifted. According to Veto (1987), the anoxia in the Carpathian flysch basin was caused by very high plankton productivity, whereas nutrients necessary for it were most likely supplied by upwelling.

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REFERENCES CITED

Airoldi, C., 2001, The Pienides in Maramures—Sedimentation, tectonics and paleogeography: Cluj-Napoca, Romania, Cluj University Press, 156 p.

Aubrecht, R., and I. Tunyi, 2001, Original orientation of neptunian dykes in the Pieniny Klippen Belt (Western Carpathians): The first results: Contributions to Geophysics and Geodesy, v. 31, p. 557–578.

Balintoni, I., 1998, An evolutionary model for the rift of external Carpathian flysch basin: Studia Universitatis Babes-Bolyai: Geologia, v. 42, p. 119–131.

Balintoni, I. C., 2001, The structure of the Romanian territory, *in* I. I. Bucur, S. Filipescu, and E. Sasaran, eds., Algae and carbonate platforms in the western part of Romania: 4th Regional Meeting of International Fossil Algae Association, August 29–September 5, 2001: Field Trip Guide Book: Cluj-Napoca, Romania, Cluj University Press, p. 1–8.

Bieda, F., S. Geroch, L. Koszarski, M. Ksiazkiewicz, and K. Zytko, 1963, Stratigraphie des Karpates externes polonaises: Biuletyn Instytutu Geologicznego, v. 182, p. 5–174.

Bill, M., L. O'Dogherty, J. Guex, P. O. Baumgartner, and H. Masson, 2001, Radiolarite ages in Alpine–Mediterranean ophiolites: Constraints on the oceanic spreading and the Tethys–Atlantic connection: Geological Society of America Bulletin, v. 113, p. 129–143.

Birkenmajer, K., 1977, Jurassic and Cretaceous lithostratigraphic units of the Pieniny Klippen Belt, Carpathians, Poland: Studia Geologica Polonica, v. 45, p. 1–158.

Birkenmajer, K., 1986, Stages of structural evolution of the Pieniny Klippen Belt, Carpathians: Studia Geologica Polonica, v. 88, p. 7–32.

Birkenmajer, K., 1988, Exotic Andrusov Ridge: Its role in plate-tectonic evolution of the West Carpathian fold belt: Studia Geologica Polonica, v. 91, p. 7–37.

Birkenmajer, K., and N. Oszczypko, 1989, Cretaceous and Palaeogene lithostratigraphic units of the Magura nappe, Krynica subunit, Carpathians: Annales Societatis Geologorum Poloniae, v. 59, p. 145–181.

Birkenmajer, K., H. Kozur, and R. Mock, 1990, Exotic Triassic pelagic limestone pebbles from the Pieniny Klippen Belt of Poland: A further evidence for early Mesozoic rifting in West Carpathians: Annales Societatis Geologorum Poloniae, v. 60, p. 3–44.

Bombita, G. E., and G. Pop, 1991, Mesozoic formations from Poiana Botizii, Pieniny Klippen Belt of Romania: Geologica Carpathica, v. 42, p. 139–146.

Bombita, G., E. Antonescu, E. Malata, C. Muller, and T. Neagu, 1992, Pieniny-type Mesozoic formation from Maramures, Romania (second part): Acta Geologica Hungarica, v. 35, p. 117–144.

Borys, Z., and B. Cisek, 1989, New perspectives on finding hydrocarbon deposits in the Lower Cretaceous sandstones in the Skole unit: Nafta (Katowice), v. 10–12, p. 142–147.

Burtan, J., 1968, The pre-Magura scale: XXIII International Geological Congress (Prague, 1968): Guide to Excursion C44—Geology of the Polish Flysch Carpathians: Warsaw, Poland, Wydawnictwa Geologiczne, p. 14–16.

Burtan, J., J. Golonka, N. Oszczypko, and Z. Paul, 1981, Mapa Geologiczna Polski, Arkusz Nowy Sacz: Instytut Geologiczny geological map of Poland (in Polish): Warszawa, Geological Institute Publishing House, scale 1:200,000, 1 sheet.

Burtan, J., J. Chowaniec, and J. Golonka, 1984, Wstepne wyniki badan nad egzotycznymi skalami weglanowymi z zachodniej czesci polskich Karpat fliszowych (Preliminary results of studies on exotic carbonate rocks in the western part of the Polish Flysch Carpathians): Biuletyn Instytutu Geologicznego, v. 346, p. 147–159.

Cieszkowski, M., 1992, Michalczowa zone: A new unit of the Fore-Magura zone, Outer West Carpathians, south Poland (in Polish with English summary): Zeszyty Naukowe Akademia Gorniczo-Hutnicza, v. 1463, p. 1–125.

Cieszkowski, M., A. Slaczka, and S. Wdowiarz, 1985, New data on structure of the Flysch Carpathians: Przeglad Geologiczny, v. 6, p. 313–333.

Cieszkowski, M., N. Oszczypko, and W. Zuchiewicz, 1989, Upper Cretaceous siliciclastic-carbonate turbidites at Szczawa, Magura nappe, West Carpathians, Poland: Bulletin of the Polish Academy of Sciences. Earth Sciences, v. 37, p. 231–245.

Dercourt, J., L. E. Ricou, and B. Vrielynck, eds., 1993, Atlas Tethys paleoenvironmental maps: Paris, Gauthier-Villars, 307 p.

Dziadzio, P. S., 2006, The Geology of the Weglowka oil field, Subsilesian unit, Polish Outer Carpathians, *in* J. Golonka and F. J. Picha, eds., The Carpathians and their foreland: Geology and hydrocarbon resources: AAPG Memoir 84, p. 467–476.

Dziadzio, P. S., M. A. Enfield, M. P. Watkinson, and S. J. Porebski, 2006a, The Ciezkowice Sandstones: Examples of basin-floor fan-stacking patterns from the main (upper Paleocene to Eocene) reservoir in the Polish Carpathians, *in* J. Golonka and F. J. Picha, eds., The Carpathians and their foreland: Geology and hydrocarbon resources: AAPG Memoir 84, p. 477–496.

Dziadzio, P. S., et al., 2006b, Hydrocarbon resources of the Polish outer Carpathians—reservoir parameters, trap types, and selected hydrocarbon fields: A stratigraphic review, *in* J. Golonka and F. J. Picha, eds., The Carpathians and their foreland: Geology and hydrocarbon resources: AAPG Memoir 84, p. 259–291.

Fayrad, S. W., 1997, Petrological model for blueschist facies metamorphism in the Pieniny Klippen Belt, *in* P. Grecula, D. Hovorka, and M. Putis, eds., Geological evolution of the Western Carpathians: Mineralia Slovaca Monograph, Bratislava, p. 155–162.

Gasinski, M. A., 1998, Campanian–Maastrichtian palaeoecology and palaeobiogeography of the Andrychow Klippes, Outer Carpathians, Poland: Wydaw: Krakow, Uniwersytetu Jagiellonskiego, 90 p.

Geroch, S., A. Jednorowska, M. Ksiazkiewicz, and J. Liszkowa, 1967, Stratigraphy based upon foraminifera of the Western Polish Carpathians: Biuletyn Instytutu Geologicznego, v. 211, p. 185–267.

Glushko, W. W., and S. S. Kruglov, 1986, Tectonic map of the Ukrainian Carpathians: Kiev, Ministerstvo Geologii, scale 1:200,000, 6 sheets.

Golonka, J., 1981, Arkusz Bielsko-Biala; Objasnienia do mapy geologicznej Polski (Geological map of Poland, explanations; Bielsko-Biala sheet): Warsaw, Geological Institute Publishing House, p. 1–63.

Golonka, J., 2004, Plate tectonic evolution of the southern margin of Eurasia in the Mesozoic and Cenozoic: Tectonophysics, v. 381, p. 235–273.

Golonka, J., and M. Krobicki, 2001, Upwelling regime in the Carpathian Tethys: A Jurassic–Cretaceous palaeogeographic and paleoclimatic perspective: Geological Quarterly, v. 45, p. 15–32.

Golonka, J., and M. Lewandowski, eds., 2003, Geology, geophysics, geothermics and deep structure of the West Carpathians and their basement: Institute of Geophysics Polish Academy of Sciences Monographic Volume, v. M-28, no. 363, p. 1–184.

Golonka, J., and W. Raczkowski, 1984, Arkusz Piwniczna; Objasnienia do szczegolowej mapy geologicznej Polski (Detailed geological map of Poland) (in Polish): Warsaw, Geological Institute Publishing House Wydawnictwa Geologiczne, p. 85, scale 1:50,000, 1 sheet.

Golonka, J., and W. Sikora, 1981, Microfacies of the Jurassic and Lower Cretaceous sedimentarily thinned deposits of the Pieniny Klippen Belt in Poland (in Polish with English abstract): Biuletyn Instytutu Geologicznego, v. 31, p. 7–37.

Golonka, J., N. Oszczypko, and A. Slaczka, 2000, Late Carboniferous–Neogene geodynamic evolution and paleogeography of the circum-Carpathian region and adjacent areas: Annales Societatis Geologorum Poloniae, v. 70, p. 107–136.

Golonka, J., M. Krobicki, N. Oszczypko, A. Slaczka, and T. Slomka, 2003, Geodynamic evolution and palaeogeography of the Polish Carpathians and adjacent areas during neo-Cimmerian and preceding events (latest Triassic–earliest Cretaceous), *in* T. McCann and A. Saintot, eds., Tracing tectonic deformation using the sedimentary record: Geological Society (London) Special Publication 208, p. 138–158.

Golonka, J., L. Gahagan, M. Krobicki, F. Marko, N. Oszczypko, and A. Slaczka, 2006, Plate-tectonic evolution and paleogeography of the circum-Carpathian region, *in* J. Golonka and F. J. Picha, eds., The Carpathians and their foreland:

Geology and hydrocarbon resources: AAPG Memoir 84, p. 11–46.

Guterch, A., M. Grad (reporters) and Polonaise '97 and Celebration 2000 Working Group, 2001, New deep seismic studies of the lithosphere in Central Europe: Polonaise '97 and Celebration 2000 seismic experiments: Biuletyn Panstwowego Instytutu Geologicznego, v. 396, p. 61.

Jucha, S., and J. Kotlarczyk, 1961, Le series des couches a Menilite et des couches de Krosno dans le flysch des Karpates: Pace Geologiczne Polska Academia Nauk, Oddzial a Krakowie, v. 4, p. 1–115.

Korab, T., and T. Durkovic, 1978, Geology of the Dukla unit (east Slovakian flysch) (in Slovakian with English summary): Bratislava, Geologicky Ustav D. Stura, 194 p.

Koszarski, L., ed., 1985, Geology of the middle Carpathians and the Carpathian foredeep: Guide to Excursion 3, Carpatho-Balkan Geological Association XIII Congress, Krakow, Poland, 254 p.

Koszarski L., and A. Koszarski, 1985, Dukla unit and its relation to Silesian nappe: Marginal zone of the Magura nappe: Tectonic windows with Grybow unit series: Jaslo nappe, *in* L. Koszarski, ed., Geology of the Middle Carpathian Foredeep, Guide to excursion 3, Carpatho-Balkan Association XIII Congress, Cracow, Poland, Geological Institute, Poland, p. 185–225.

Koszarski L., and A. Slaczka, 1976, The Outer (Flysch) Carpathians: The Cretaceous, *in* S. Cieslinski, ed., Geology of Poland, v. I, Stratigraphy part 2: Instytut Geologiczny, Warszawa, p. 495–498, 657–679, 740–748.

Koszarski, L., and K. Zytko, 1961, Jaslo shales within the Menilite-Krosno series in the Middle Carpathians: Biuletyn Instytutu Geologicznego, v. 166, p. 87–213.

Koszarski, L., W. Sikora, and S. Wdowiarz, 1974, The Flysch Carpathians: Polish Carpathians, *in* M. Mahel, ed., Tectonics of the Carpathian–Balkan regions: Bratislava, Geologicky Ustav Donuza Stura, p. 180–197.

Kotarba, M., and Y. V. Koltun, 2006, The origin and habitat of hydrocarbons of the Polish and Ukrainian parts of the Carpathian province, *in* J. Golonka and F. J. Picha, eds., The Carpathians and their foreland: Geology and hydrocarbon resources: AAPG Memoir 84, p. 395–442.

Kotlarczyk, J., 1978, Stratigraphy of the Ropianka Formation in the Skole unit of the Flysch Carpathians (in Polish with the English summary): Polska Academia Nauk Oddzial w Krakowie. Komisja Nauk Geologicznych Prace Geologiczne, v. 108, p. 1–82.

Kotlarczyk, J., 1985, An outline of the stratigraphy of marginal tectonic units of the Carpathian orogen in the Krakow–Przemysl area: Guide to Excursion 4. Carpatho-Balkan Geological Association—XIII Congress, Krakow, Poland, 171 p.

Kovac, M., and D. Plasienka, eds., 2002, Geological structure of the Alpine–Carpathian–Pannonian junction and neighbouring slopes of the Bohemian Massif: Bratislava, Comenius University, 84 p.

Kovac, M., A. Nagymarosy, J. Sotak, and K. Sutovska, 1993, Late Tertiary paleogeographic evolution of the Western Carpathians: Tectonophysics, v. 226, p. 401–415.

Kovac, M., A. Nagymarosy, N. Oszczypko, A. Slaczka, L. Csontos, M. Marunteanu, L. Matenco, and M. Marton, 1998, Palinspastic reconstruction of the Carpathian–Pannonian region during the Miocene, *in* M. Rakus, ed., Geodynamic development of the Western Carpathians: Bratislava, Geological Survey of Slovac Republic, p. 189–217.

Kozikowski, H., 1956, Pisarzowa unit, a new tectonic unit of the Polish Flysch Carpathians: Biuletyn Instytutu Geologicznego, v. 110, p. 93–121.

Kozikowski, H., 1958, Geologia Centralnej Depresji Karpackiej między Żmigrodem a Sanokiem: Acta Geologica Polonica, v. 8, p. 471–500.

Kropotkin, P. N., 1991, Nappes of the Maramures zone of the Eastern Carpathians: Geotectonics, v. 25, p. 46–52.

Kruglov, S. S., 1965, O prirode marmaroshskikh utiosov Sovetskikh Karpat (About the nature of the Marmarosh Klippens of the Soviet Carpathians) (in Russian): Geologicheskiy Sbornik Lvovskogo Geologiceskogo Obsestva, v. 9, p. 41–54.

Kruglov, S. S., 1969, Pro stratyhrafichne polozhennya pudynhovych piskovykiv i hravelitiv v zoni peninskyh skel (Zakarpattia) (About stratigraphical position of pudding sandstones and gritstones within the Pieniny Klippen zone (Zakarpatia) (in Ukrainian): Proceedings of the Academy of Sciences of the Ukrainian SSR, v. 10, p. 874–877.

Kruglov, S. S., 1989, Geodynamics of the Ukrainian Carpathians: Geologica Carpathica, v. 40, p. 101–123.

Kruglov, S. S., 1999, V zashchitu Predkarpatskoho prohiba (In advocacy of the pre-Carpathian Foredeep) (in Russian): Geodynamics (L'viv), v. 1, p. 88–91.

Kruglov, S. S., 2001, The problems of tectonics and paleogeodynamics of western Ukraine (a critical survey of the new publications) (in Ukrainian): L'viv, Ukrainian Interdisciplinary Committee for Tectonics, 83 p.

Kruglov, S. S., and W. I. Tarasenko, 1984, Brachysynclinal folds in the Sovjet Carpathians (in Russian): Gielogiceskij Zurnal, v. 2, p. 28–37.

Ksiazkiewicz, M., 1951, Kreda Karpat Zewnętrznych, *in* M. Ksiazkiewicz, ed., Regionalna Geologia Polski, Karpaty, Stratygrafia: Rocznik Polskiego Towarzystwa Geologicznego, v. 1, p. 86–110.

Ksiazkiewicz, M., ed., 1962, Geological atlas of Poland: Stratigraphic and facial problems: Instytut Geologiczny: Warsaw, Poland, Wydawnictwa Geologiczne, 16 maps.

Ksiazkiewicz, M., 1977, Tectonics of the Carpathians, *in* W. Pozaryski, ed., Geology of Poland, v. IV, Tectonics: Warsaw, Poland, Wydawnictwa Geologiczne, p. 476–604.

Ksiazkiewicz, M., and B. Lesko, 1959, On relation between the Krosno and Magura Flysch: Bulletin of the Polish Academy of Sciences Series, Chemistry, Geology, and Geography, v. 7, p. 773–780.

Kucharski, R., S. Doktor, M. Graniczny, M. Molek, and B. Dabrowska, 1990, Wgłębna budowa geologiczna Karpat w świetle kompleksowej analizy teledetekcyjno-geofizycznej: Przegląd Geologczny, v. 11, p. 469–476.

Lashkevitsch, Z. M., A. P. Medvedev, and Y. Z. Krupskiy, 1995, Tectonomagmatic evolution of Carpathians (in Russian): Kiev, Naukova Dumka, p. 1–113.

Lesko, B., and O. Samuel, 1968, Geology of the Eastern Slovakian

Flysch (in Slovakian with English summary): Bratislava, Slovenska Akademia ved, 256 p.
Lillie, R. J., M. Bielik, V. Babuska, and J. Plomerova, 1994, Gravity modelling of the lithospere in the eastern Alpine–Western Carpathian–Pannonian Basin region: Tectonophysics, v. 231, p. 215–235.
Lucinska-Anczkiewicz, A., A. Slaczka, I. M. Villa, and R. Anczkiewicz, 2000, $^{39}Ar/^{40}Ar$ dating of the teschenite association rocks from the Polish Outer Carpathians: Mineralogical Society of Poland Special Papers 17, p. 241–242.
Lucinska-Anczkiewicz, A., I. M. Villa, R. Anczkiewicz, and A. Slaczka, 2002, $^{40}Ar/^{39}Ar$ dating of alkaline lamprophyres from the Polish Western Carpathians: Geologica Carpathica, v. 53, p. 45–52.
Mahel, M., et al., 1968, Regional geology of Czechoslovakie: Part II, The West Carpathians: Praha, Geological Survey of Czechoslovakia, 723 p.
Malata, E., 2001, Interpretacja biostratygraficzna zespołów malych otwornic w środkowej części płaszczowiny magurskiej (polskie Karpaty zewnętrzne), Rozprawa doktorska w Instytucie Geologicznym Uniwersytetu Jagiellońskiego: Ph.D. dissertation, Jagiellonian University, Krakow, 273 p.
Malata E., T. Malata, and N. Oszczypko, 1996, Litho- and biostratigraphy of the Magura nappe in the eastern part of the Beskid Wyspowy Range (Polish Western Carpathians): Annales Societatis Geologorum Poloniae, v. 66, p. 269–283.
Mastella, L., 1988, Structure and evolution of Mszana Dolna tectonic window, Outer Carpathians, Poland (in Polish with English summary): Annales Societatis Geologorum Poloniae, v. 58, p. 53–173.
Matyasik, I., and P. S. Dziadzio, 2006, Reconstruction of petroleum systems based on integrated geochemical and geological investigations: Selected examples from the Middle, Outer Carpathians in Poland, *in* J. Golonka and F. J. Picha, eds., The Carpathians and their foreland: Geology and hydrocarbon resources: AAPG Memoir 84, p. 497–518.
Misik M., M. Sykora, and J. Jablonsky, 1991, Strihovske zlepence a juhomagurska kordilera: Zapadne Karpaty: Seria Geologicka, v. 14, p. 7–72.
Mitura, F., and T. Birecki, 1966, Geological structure of the Carpathians between Korczyna and Domaradz (in Polish with English summary): Prace Instytutu Naftowego, 3, Instytut Naftowy, Krakow, Poland, 52 p.
Olszewska, B., and J. Wieczorek, 2001, Jurasic sediments and microfossils of the Andrychow Klippes (Outer Western Carpathians): Geologica Carpathica, v. 52, p. 217–228.
Oszczypko, N., 1973, Budowa geologiczna Kotliny Sądeckiej: Biuletyn Instytutu Geologicznego, v. 271, p. 101–197.
Oszczypko, N., 1975, Exotic rocks in the Palaeogene of the Magura nappe between Dunajec and Poprad rivers, Carpathians, Poland: Rocznik Polskiego Towarzystwa Geologicznego, v. 45, p. 403–431.
Oszczypko, N., 1991, Stratigraphy of the Palaeogene deposits of the Bystrica subunit (Magura nappe, Polish Outer Carpathians): Bulletin of the Polish Academy of Sciences. Earth Sciences, v. 39, p. 415–431.
Oszczypko, N., 1992, Late Cretaceous through Paleogene evolution of Magura basin: Geologica Carpathica, v. 43, p. 333–338.
Oszczypko, N., 1999, From remnant oceanic basin to collision-related foreland basin—A tentative history of the Outer Western Carpathians: Geologica Carpathica, v. 40, special issue, p. 161–163.
Oszczypko, N., 2001, Magura unit—General geology, *in* K. Birkenmajer and M. Krobicki, eds., Carpathian palaeogeography and geodynamics: A multidisciplinary approach: Field Trip C, 2nd Meeting of the Association of European Geological Societies, Krakow 2001, Panstwowy Instytut Geologiczny, p. 173–177.
Oszczypko, N., and M. Oszczypko-Clowes, 2002, The new findings of lower Miocene deposits in the Nowy Sacz area (Magura nappe, Polish Outer Carpathians): Geological Quarterly, v. 46, p. 15–30.
Oszczypko, N., and A. Slaczka, 1985, An attempt at palinspastic reconstruction of Neogene basin in the Carpathian foredeep: Annales Societatis Geologorum Poloniae, v. 55, p. 55–75.
Oszczypko, N., and A. Thomas, 1985, Tectonic evolution of marginal part of the Polish Flysch Carpathians in the middle Miocene: Kwartalnik Geologiczny, v. 29, p. 109–128.
Oszczypko, N., and W. Zuchiewicz, 2000, Jointing in Eocene flysch strata of the mid-eastern Magura nappe, Polish Outer Carpathians: Implications for the timing deformation: Slovak Geological Magazine, v. 6, p. 441–456.
Oszczypko, N., and K. Zytko, 1987, Main stages of evolution of the Polish Carpathians during the late Palaeogene and Neogene times, *in* Y. G. Leonov and V. E. Khain, eds., Global correlation of tectonic movements: Chichester, J. Wiley and Sons Ltd., p. 187–198.
Oszczypko, N., J. Dudziak, and E. Malata, 1990, Stratygrafia osadów płaszczowiny magurskiej (kreda-paleogen) w Beskidzie Sądeckim, Karpaty: Studia Geologica Polonica, v. 97, p. 109–181.
Oszczypko, N., P. Krzywiec, M. Lemberger, M. Stefaniuk, K. Pietsch, and H. Trygar, 1998, Integrated geological-geophysical interpretation of the Rzeszow-Smilno profile (Western Carpathians) (abs.): Carpathian–Balkan Geological Association XVI Congress, Vienna, 1998, Abstracts, p. 446.
Oszczypko, N., E. Malata, and M. Oszczypko-Clowes, 1999a, Revised position and age of the Eocene deposits on the northern slope of the Gorce Range (Bystrica subunit, Magura nappe): Slovak Geological Magazine, v. 5, p. 235–254.
Oszczypko N., A. S. Andreyeva-Grigorovich, E. Malata, and M. Oszczypko-Clowes, 1999b, The lower Miocene deposits of the Raca subunit near Nowy Sacz (Magura nappe, Polish Outer Carpathians): Geologica Carpathica, v. 50, p. 419–433.
Oszczypko, N., P. Krzywiec, I. Popadyuk, and T. Peryt, 2006, Carpathian foredeep basin (Poland and Ukraine): Its sedimentary, structural, and geodynamic evolution, *in* J. Golonka and F. J. Picha, eds., The Carpathians and their foreland: Geology and hydrocarbon resources: AAPG Memoir 84, p. 293–350.
Oszczypko-Clowes, M., 1998, Late Eocene–Oligocene calcareous nannoplankton and stable isotopes (^{13}C, ^{18}O) of the

Globigerina marls in the Magura nappe (West Carpathians): Slovak Geological Magazine, v. 4, p. 107–120.
Oszczypko-Clowes, M., 2001, The nannofossils biostratigraphy of the youngest deposits of the Magura nappe (east of the Skawa river, Polish Flysch Carpathians) and their palaeoenvironmental conditions: Annales Societatis Geologorum Poloniae, v. 71, p. 139–188.
Pasternak, S. I., 1989, Stratotypes of the Cretaceous and Palaeogene deposit of the Ukrainian Carpathians: Kijev, Naukova Dumka Publishers, 204 p.
Pescatore, T., and A. Slaczka, 1984, Evolution models of two flysch basins: The Northern Carpathians and the southern Apennines, Tectonophysics, v. 106, p. 49–70.
Picha, F. J., 1996, Exploring for hydrocarbons under thrust belts: A challenging new frontier in the Carpathians and elsewhere: AAPG Bulletin, v. 80, no. 10, p. 1547–1564.
Picha, F. J., Z. Stranik, and O. Krejci, 2006, Geology and hydrocarbon resources of the Outer Western Carpathians and their foreland, Czech Republic, *in* J. Golonka and F. J. Picha, eds., The Carpathians and their foreland: Geology and hydrocarbon resources: AAPG Memoir 84, p. 49–175.
Plasienka, D., 1999, Tektonochronológia a paleotektonický model jursko-kriedového vývoja centralnych Západných Karpat (Tectonochrology and paleotectonic model of the Jurassic–Cretaceous evolution of the Central Western Carpathians): Bratislava, Veda, 127 p.
Plasienka, D., 2000, Paleotectonic controls and tentative palinspastic restoration of the Carpathians realm during the Mesozoic: Slovak Geological Magazine, v. 6, p. 200–204.
Plasienka, A., 2002, Early stages of tectonic evolution of the Pieniny Klippen Belt: Geolines, v. 14, p. 75–77.
Popadyuk, I., M. Vul, Y. Ladyzhensky, and P. Shpak, 2006, Petroleum geology of the Boryslav–Pokuttya zone, the Ukrainian Carpathians, *in* J. Golonka and F. J. Picha, eds., The Carpathians and their foreland: Geology and hydrocarbon resources: AAPG Memoir 84, p. 455–466.
Poprawa, P., T. Malata, and N. Oszczypko, 2002, Ewolucja tektoniczna basenów sedymentacyjnych polskiej części Karpat zewnętrznych w świetle analizy subsydencji: Przegląd Geologiczny, v. 50, p. 1092–1105.
Potfaj, M., 1993, Position and role of the Biele Karpaty unit in the Flysch Zone of the Western Carpathians (in Slovakian with English summary): Geologicke Prace, Spravy, v. 98, p. 55–78.
Rakus, M., M. Misik, J. Michalik, R. Mock, T. Durkovic, T. Korab, R. Marschalko, J. Mello, M. Polak, and J. Jablonsky, 1988, Paleogeographic development of the West Carpathians: Anisian to Oligocene: Memoires de la Societe Geologique de France, Nouvelle Series, v. 154, p. 39–62.
Royden, L. H., and T. Baldi, 1988, Early Cenozoic tectonics in paleogeography of the Pannonian and surrounding regions, *in* L. H. Royden and F. Horvath, eds., The Pannonian Basin: A study in basin evolution: AAPG Memoir 45, p. 1–16.
Rylko, W., and A. Tomas, 1995, Morfologia powierzchni stropowej podłoża krystalicznego Karpat w świetle badañ magneto-telurycznych: Kwartalnik Geologiczny, v. 39, p. 1–16.
Sandulescu, M., 1975, Essai de synthese structurale des Carpathes: Bulletin de la Societe Geologique de France, v. 17, p. 299–358.
Sandulescu, M., 1988, Cenozoic tectonic history of the Carpathians, *in* L. Royden and F. Horvath, eds., The Pannonian Basin: A study in basin evolution: AAPG Memoir 84, v. 45, p. 17–25.
Sandulescu, M., and M. Visarion, 2000, Crustal structure and evolution of the Carpathian–western Black Sea area: Romanian Geophysics, v. 7, suppl. 1, p. 228–231.
Sandulescu, M., H. Krautner, I. Balintoni, D. Russo-Sandulescu, and M. Micu, 1981, The structure of the East Carpathian (Moldavia–Muramures area): Carpathian–Balkan Geological Association, XII Congress Bucharest–Romania 1981, Guide to Excursion B1, p. 1–92.
Schnabel, W. G., 1992, New data on the flysch zone of the Eastern Alps in the Austrian sector and new aspects concerning the transition to the flysch zone of the Carpathians: Cretaceous Research, v. 13, p. 405–419.
Seifert, P., 1992, Palinspastic reconstruction of the Easternmost Alps between upper Eocene and Miocene: Geologica Carpathica, v. 43, p. 327–331.
Sikora, W., 1970, Geology of the Magura nappe between Szymbark Ruski and Nawojowa: Biuletyn Instytutu Geologicznego, v. 235, p. 5–121.
Slaczka, A., 1959, Stratigraphy of the Bystra scale, Middle Carpathians (in Polish with English summary): Biuletyn Instytutu Geologicznego, v. 131, p. 203–286.
Slaczka, A., 1969 Final stages of geosynclinal development in the southeast-part of the Polish Carpathians: Acta Geologica Academiae Scientiarum Hungaricae, v. 13, p. 331–335.
Slaczka, A., 1971, Geology of the Dukla unit (in Polish with English summary): Prace Instytutu Geologicznego, v. 1, p. 1–63.
Slaczka, A., ed., 1976, Atlas of paleotransport of detrital sediments in the Carpathian–Balkan Mountains system: Warszawa, Geological Institute, scale 1:2,000,000, 6 sheets.
Slaczka, A., 1980, Les Carpathes flysheuses, *in* R. Osika, ed., La geologie et les gites mineraux en Pologne: Biuletyn Instytutu Geologicznego, v. 251, p. 491–497, 604–615.
Slaczka, A., 1996a, Oil and gas in the Northern Carpathians, *in* G. Wessely and W. Liebl, eds., Oil and gas in Alpidic thrust belts and basins of Central and Eastern Europe: European Association of Geoscientists and Engineers Special Publication 5, p. 187–195.
Slaczka, A., 1996b, Oil and gas in the Ukrainian part of the Carpathians and their foredeep, *in* G. Wessely and W. Liebl, eds., Oil and gas in Alpidic thrust belts and basins of Central and Eastern Europe: European Association of Geoscientists and Engineers Special Publication 5, p. 17–21.
Slaczka, A., 1998, Age of the andesitic rocks in the Subsilesian unit (Outer Carpathians) (abs.): XVI Congress, Carpatho-Balkan Geological Association, Vienna, August 30–September 2, 1998, Abstracts, p. 559.
Slaczka, A., 2000, Dukla unit: A key between Western and Eastern Carpathians: Terra Nostra, Schriften der Alfred-Wegener-Stiftung 2000/1, Geology 2000, Vienna, April 14–17, 2000, p. 109.

Slaczka, A., and M. A. Gasinski, 1985, Paleobathymetric model of the upper Senonian flysch sediments of a part of the Silesian basin (Polish Carpathians): Cretaceous Research, v. 6, p. 207–218.
Slaczka, A., and M. A. Kaminski, 1998, Guidebook to excursions in the Polish Flysch Carpathians: Grzybowski Foundation Special Publication 6, p. 1–171.
Slaczka, A., and N. Oszczypko, 1987, Olistostromes and overthrusting in the Polish Carpathians: Annales Instituti Geologici Hungarici, v. 70, p. 287–292.
Slaczka, A., and R. Unrug, 1976, Trends of textural and structural variations in turbidite sandstone (Oligocene, Outer Carpathians): Rocznik Polskiego Towarzystwa Geologicznego, v. 46, p. 55–75.
Slaczka, A., N. Oszczypko, E. Malata, and M. Cieszkowski, 1999, An early history of the Outer Carpathians basin: Geologica Carpathica, v. 50, p. 170–172.
Sotak, J., 1986, Stratigraphy and typology of the Upper Triassic development in outer units in the West Carpathians (reconstructions from redeposits localised in the Silesian Cordillera area): Knihovnicka Zemniho Plynu a Nafty, v. 31, p. 1–53.
Sotak, J., A. Biron, R. Prokesova, and J. Spisiak, 2000, Detachment control of core complex exhumation and back-arc extension in the east Slovakian basin: Slovak Geological Magazine, v. 6, p. 130–132.
Stampfli, G. M., ed., 2001, Geology of the western Swiss alps—A guide book: Memoires de Geologie, Lausanne, v. 361, p. 1–195.
Stefanescu, M., O. Dicea, A. Butac, and D. Ciulavu, 2006, Hydrocarbon geology of the Romanian Carpathians, their foreland, and the Transylvanian basin, *in* J. Golonka and F. J. Picha, eds., The Carpathians and their foreland: Geology and hydrocarbon resources: AAPG Memoir 84, p. 521–567.
Stefaniuk, M., 2002, Basic structures of the Carpathian basement as a result of magnetotelluric sounding interpretation: Geologia Carpathica, v. 53, special issue (CD-ROM).
Stefaniuk, M., and A. Slaczka, 2002, Interpretation of the flysch cover structures based on magnetotelluric data (Polish Outer Carpathians): Geologica Carpathica, v. 53, special issue (CD-ROM).
Stranik, Z., J. Dvorak, O. Krejci, P. Muller, A. Prichystal, M. Suk, and C. Tomek, 1993, The contact of the North European platform with the West Carpathians: Journal of the Czech Geological Society, v. 38, p. 21–29.
Svabenicka, L., M. Bubik, O. Krejci, and Z. Stranik, 1997, Stratigraphy of Cretaceous sediments of the Magura group of nappes in Moravia (Czech Republic): Geologica Carpathica, v. 48, p. 179–191.
Tolman, A., 1990, Paleogeographic maps and profiles in the Eastern Alps and relationship of the Eastern Alps and neighboring terrains, *in* A. E. M. Nairn, ed., Evolution of the northern margin of Tethys: The results of International Geological Correlation Project 198, v. 3: Memoires de la Societe Geologique de France, Nouvelle Series, v. 154, p. 23–38.
Tomek, C., 1993, Deep crustal structure beneath the Central and Inner West Carpathians: Tectonophysics, v. 226, p. 417–431.
Tomek, C., and J. Hall, 1993, Subducted continental margin in the Carpathians of Czechoslovakia: Geology, v. 21, p. 535–538.
Veto, I., 1987, An Oligocene sink for organic carbon: Upwelling in the Paratethys? Palaeogeography, Palaeoclimatology, Palaeoecology, v. 60, p. 143–153.
Wdowiarz, S., 1985, On same aspects of the geological structure and oil and gas accumulation of the Carpathian synclinorium: Biuletyn Instytutu Geologicznego, v. 350, p. 5–52.
Wieser, T., 1970, Exotic rocks from the deposits of the Magura nappe: Biuletyn Instytutu Geologicznego, v. 235, p. 123–161.
Winkler, W., and A. Slaczka, 1994, A Late Cretaceous to Paleogene geodynamic model for the Western Carpathians in Poland: Geologica Carpathica, v. 45, p. 71–82.
Woznicki, J., and P. Sucha, 1989, Geophysical map of the Western Outer Carpathians and their foreland, *in* D. Porawa and J. Nemcok, eds., Geological atlas of the Western Outer Carpathians and their foreland: Warszawa, Panstwowy Instytut Geologiczny, Table 6.
Zytko, K., 1999, Correlation of the main structural units of the Western and Eastern Carpathians (in Polish with English summary): Prace Panstwowego Instytutu Geologicznego, v. 168, p. 135–164.
Zytko, K., et al., 1989, Geological map of the Western Outer Carpathians and their foreland without Quaternary formations, *in* D. Poprawa and J. Nemcok, eds., Geological atlas of the Western Carpathians and their foreland: Warszawa, Poland, Panstwowy Instytut Geologiczny, Table 1.