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Geology and Hydrocarbon Resources of the Outer Western Carpathians and Their Foreland, Czech Republic

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ABSTRACT

The Western Carpathians in the territory of Moravia (the eastern part of the Czech Republic) and northeastern (Lower) Austria represent the westernmost segment of the entire Carpathian orogenic system linked to the Eastern Alps. Based on differences in their depositional and structural history, the Carpathians are divided into two primary domains: the Inner Carpathians deformed and thrusted in the Late Jurassic to Early Cretaceous, and the Outer Carpathians deformed and thrusted over the European foreland during the Paleogene and Neogene. These two domains are separated by the Pieniny Klippen Belt, which bears signatures of both these domains and stands out as a primary suture in the Western Carpathians. Only the Outer Carpathians, including the thin-skinned thrust belt partly overlain by the Vienna basin and the undeformed Neogene foredeep, are present in the territory of Moravia and, as such, are subjects of our deliberation.

The foreland of the Carpathians in Moravia is represented by the Bohemian Massif, which is a part of the West European plate. It consists of the Hercynian orogenic belt and the late Precambrian (Cadomian) foreland terrane of the Brunovistulicum. The unmetamorphosed sedimentary cover of the cratonic basement of the Bohemian Massif in Moravia extends through two plate-tectonic cycles, the Paleozoic Hercynian and the Mesozoic to Cenozoic Tethyan–Alpine. The Bohemian Massif continues far below the Carpathian foredeep and the thin-skinned Outer Carpathian thrust belt. Various deep antiformal structures have been identified in the subthrust plate by seismic methods and drilling. Some of these structures apparently formed during the Hercynian orogeny, whereas others are related either to the Jurassic rifting or to the compressional Alpine tectonics extending from the Late Cretaceous to Miocene. During the Laramide uplifting of the European foreland, in the Late Cretaceous

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to early Paleogene, two large paleovalleys and submarine canyons were cut into the foreland plate and filled with deep-water Paleogene strata.

The Carpathian orogenic system, as we know it today, evolved during the late Paleozoic, Mesozoic, and Cenozoic through the divergent and convergent processes of the plate-tectonic cycle. In the Outer Western Carpathians of Moravia, the divergent stage began in the Middle to Late Jurassic by rifting, opening of Tethyan basins, and development of the passive margins dominated by the carbonate platforms and basins. Further rifting and extension occurred in the Early Cretaceous. The convergent orogenic process in the Outer Carpathians began in the Late Cretaceous by the subduction of the Penninic–Pieninic oceanic basin and collision of the Inner Carpathians with the fragmented margins of the European plate. Since the Late Cretaceous, a major foreland basin dominated by the siliciclastic shelf and deep-water flysch sedimentation has formed in the Outer Carpathian domain. The Carpathian foreland basin, especially during the Late Cretaceous to the early Eocene, displayed a complex topography marked by an existence of intrabasinal ridges (cordilleras) such as the Silesian cordillera. We interpret them as preexisting rift-related crustal blocks activated during the Late Cretaceous–early Paleocene uplifting as foreland-type compressional structures.

During the Paleogene and early Miocene, the Upper Jurassic to lower Miocene sequences of the Outer Carpathian depositional system were gradually deformed and thrusted over the European foreland. The tectonic shortening occurred not only in the decoupled thin-skinned thrust belt but also at the deeper crustal level, where various blocks of the previously rifted margins were apparently at least partly accreted back to the foreland plate instead of being subducted.

Since the early Miocene, the synorogenic, predominantly deep-water flysch sedimentation was replaced by the shallow-marine and continental molasse-type sedimentation of the Neogene foredeep, which remained mostly undeformed. Also during the Miocene, the Vienna basin formed in the Carpathian belt of southern Moravia and northeastern Austria as a result of subsidence, back-arc extension, and the orogen-parallel pull-apart strike-slip faulting.

During its entire history, the evolution of Outer Western Carpathians in Moravia was significantly affected by the existence of two main structural elements, the Western Carpathian transfer zone and the Dyje–Thaya depression. The southwest–northeast-trending Western Carpathian transfer zone actually separated the Alps from the Carpathians. During the divergent stage, in the Early Cretaceous, the dextral motion in this zone accommodated a significant extension in the Outer Carpathian domain. Conversely, during the convergent stage in the Paleogene and Neogene, the sinistral transpressional motion in this zone facilitated the northeastern translation (escape) of the Carpathian belt and the opening of the pull-apart depocenter in the Vienna basin.

The northwest–southeast-trending Dyje–Thaya depression, in southern Moravia and northeastern Austria, formed, or at least was activated, during the Jurassic rifting. Within the fault-bounded limits of this depression, thick, organic-rich marls were deposited in the Late Jurassic, shallow-marine clastic strata were laid down and preserved in the Late Cretaceous, two paleovalleys were excavated in the Late Cretaceous–early Paleogene, and finally, the Vienna basin formed in the Miocene. The complex structural and depositional history of the depression and its surroundings created one of the most prolific petroleum systems in the entire Carpathian region, from which more than 850 million bbl of oil has been produced to date. Historically, the Vienna basin has been the dominant producer in Austria and Moravia. More recently, however, the subthrust European platform with multiple hydrocarbon plays has become the main producing province in Moravia. Some of the identified deep subthrust structures represent significant exploration prospects, which yet have to be tested.

INTRODUCTION

The main objective of this chapter is to provide a comprehensive and intelligible account on the geology and hydrocarbon resources of the Carpathian belt and its foreland on the territory of Moravia (eastern part of the Czech Republic) and northeastern Austria. To achieve this goal, we not only summarize the present knowledge on stratigraphy, structure, and petroleum systems in that particular area but also integrate it into a broader context of the Alpine–Carpathian geology and make it compatible with modern geological concepts. We also take this opportunity to present new original ideas, particularly on the structure and the depositional systems of the Outer Carpathians and their foreland.

The Western Carpathians in the territory of southern Moravia and northeastern Austria represent the westernmost segment of the entire Carpathian orogenic belt linked to the Eastern Alps. The location of this sector in the contact zone between the Alps and the Carpathians and in the proximity of the well-exposed Cadomian and Hercynian foreland terranes of the Bohemian Massif provides a unique opportunity to address some general problems of the Alpine–Carpathian geology. Among them, the correlation of various units of the Eastern Alps and Western Carpathians has been given a high priority. Southern Moravia is the only place in the entire Carpathian belt where the Paleogene strata have been found in their original unquestionable autochthonous position below the Neogene foredeep and the frontal zones of the thrust belt. Their existence not only enables better correlation with the Alpine Molasse but also provides a critical starting point for a more realistic palinspastic reconstruction of the Carpathian depositional system.

During both the divergent and convergent stages of the development of the Western Carpathians, the Austrian–Moravian sector functioned as a southwest–northeast-trending transfer corridor; we call it the Western Carpathian transfer zone, which separated the Alpine and Carpathian domains. The strike-slip movements in this zone were instrumental in the opening and closing parts of the Western Carpathian depositional system and in the late orogenic northeastward translation (escape) of the Carpathians and formation of the Vienna basin.

Special attention is paid to the existence of a southeast–northwest-trending Dyje–Thaya depression, which apparently formed during the Jurassic rifting at the territory of southern Moravia and northeastern Austria. Repeatedly subsiding during its geological history, this depression became one of the most important generative kitchens for hydrocarbons in the entire Carpathian system.

Thanks to a relatively shallow depth and the presence of important resources (coal and hydrocarbons), the subthrust European plate in Moravia has been explored intensively by drilling. It is, in fact, the most heavily drilled sector of the subthrust plate in the Alpine–Carpathian belt and possibly in the entire world and, as such, may serve as a model for other similar settings elsewhere.

In addition to the above-mentioned aspects specific to the Austrian–Moravian area, attention is paid to some general issues of the Outer Carpathian belt. Among them, the characterization of the Outer Carpathian foreland basin, which has evolved since the Late Cretaceous, seems to be the most important. Although bearing many primary characteristics of foreland basins elsewhere, the complex architecture of the Carpathian foreland basin, marked by the existence of internal ridges (cordilleras) and intervening troughs, as well as the prominent presence of both the deep-water flysch and the shallow-marine to continental molasse stages, sets it apart as a specific model among the foreland depositional systems.

In our discussions of the tectonic history of the Carpathians, we avoid using the traditional terminology of specific orogenic phases, such as Austrian, Pyrenean, Savian, etc. (e.g., Stille, 1936). Although pulses of higher and lower tectonic activity are well documented in various regions of Alps and Carpathians, it does not seem possible to correlate and generalize these events over the complex plate-tectonic setting of the entire Alpine–Carpathian system. The only phase we still use under its original name is the so-called Laramide orogeny marked by an extensive crustal uplifting and inversion of the Carpathian foreland plate at the Cretaceous–Paleogene transition. It is the predominantly vertical character of this far-reaching and deeply rooted compressional tectonism, which, in our opinion, singles out the Laramide orogeny from the progradational thin-skinned tectonism that shaped the Outer Carpathian thrust belt.

In addition, the term "sequence" is used for distinct depositional units without any reference to the specific terminology of the sequence stratigraphy.

The constructions of maps and cross sections are supported by seismic and well data. Stratigraphic records of 37 critical deep wells are listed in Appendix 1.

Finally, the local geographical and geological terms, as well as the names of authors in our article, are printed without diacritics. For better communication, lists of geographical and geological terms and authors' names with and without diacritics are provided in Appendix 2.

REGIONAL GEOLOGICAL SETTING

Part of the Tethyan–Alpine orogenic system of Europe, the Carpathian fold and thrust belt extends from the Eastern Alps in northeastern (Lower) Austria to southern Romania (Figure 1). Considering some regional differences in the evolution of the Carpathians as well as the diverse character of their foreland, the Carpathian belt is traditionally divided into three parts: the Western, Eastern, and Southern Carpathians. The Western Carpathians extend from the Danube

Figure 1. Generalized geologic map of the Alpine–Carpathian orogenic system of Europe. Study area is located in the box. The inner zones of the Alpine–Carpathian orogen are shown in blue. Modified from Picha (1996). Cross section of AA′ is shown in Figure 2.

Valley in Austria through Moravia and Slovakia to eastern Poland and the Ukrainian border (Uh River valley) and are bounded by the Hercynian West European platform. The Eastern Carpathians in Ukraine and eastern Romania are attached to the Late Proterozoic (Cadomian) East European platform. The boundary between the West and East European crustal domains is traditionally attributed to the Teisseyre–Tornquist line as marked by the course of the Polish–Danish trough. The Southern Carpathians constitute the southward-verging segment of the Carpathian belt bounded by the Moesian Platform. They extend westward to the Iron Gate on the Danube River in southwestern Romania.

The Carpathians, based on main differences in their depositional and structural history, have been divided into two main domains: the Inner Carpathians deformed and thrusted in the Late Jurassic to the Late Cretaceous and the Outer Carpathians deformed and thrusted during the Paleogene and Neogene (Figures 1, 2). In Romania, the Inner and Outer Carpathians are known as the Dacides and Moldavides, respectively (Sandulescu, 1988). Some authors, e.g., Plasienka (1995), further divided the Inner Carpathians into the Inner Carpathians sensu stricto, related to the subduction of the Meliata–Hallstatt ocean in the Late Jurassic, and the Central Carpathians consolidated in the Late Cretaceous. Because our article is not concerned with the geology of either of these systems, we prefer to use the simpler version, the Inner Carpathians, for both of them. Included into the Inner Carpathian domain are also the unfolded deposits of the Paleogene flysch basin and the Neogene molasse deposits of the Pannonian Basin that rest unconformably on the Mesozoic nappes of the Inner Carpathians.

The Inner and Outer Carpathians are separated by the Pieniny Klippen Belt (Figure 3), which has been traditionally considered to be a part of the Outer Carpathians (Ksiazkiewicz, 1977; Picha, 1996), although it bears signatures of both these domains. Structurally, the Pieniny Klippen Belt represents a complex suture, along which some elements of the oceanic and continental

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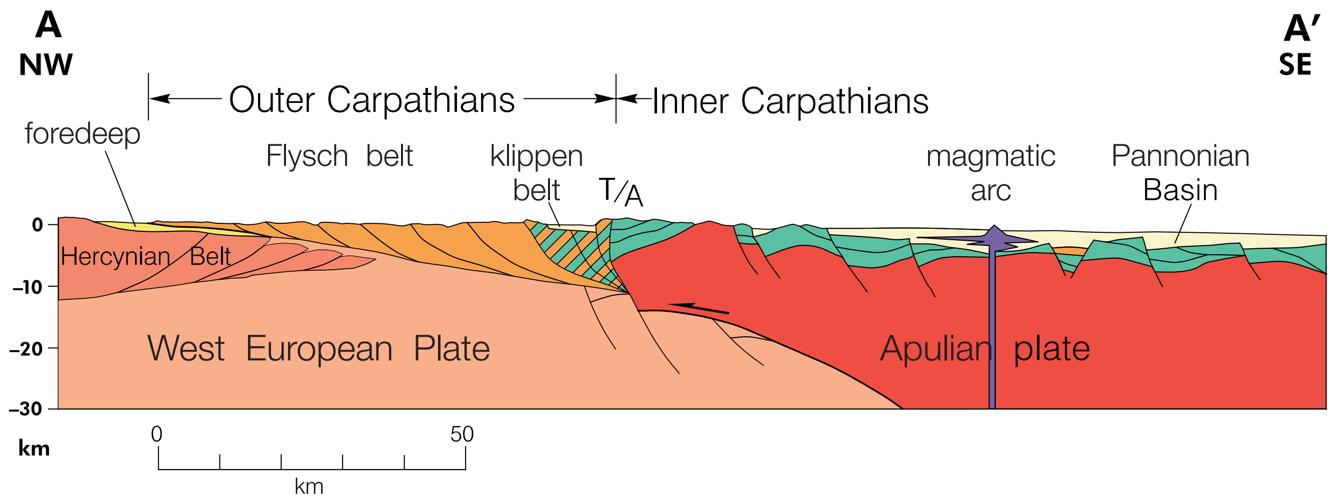

Figure 2. Generalized cross section through the Western Carpathians and their foreland. The term "Apulian plate" (Adria) is used here as a general name for microplates of the Carpathian region separated from North Africa. Location in Figure 1. Modified from Picha (1996).

lithosphere were subducted during the Late Cretaceous and early Miocene convergency (e.g., Grecula and Roth, 1978; Misik, 1979; Birkenmajer, 1986).

The Outer Western Carpathians thus consist of the allochthonous Pieniny Klippen Belt, the Flysch belt, the autochthonous Neogene foredeep, and the late orogenic and postorogenic Vienna basin, which rests transgressively on both the Outer and Inner Carpathians (Figure 3). Included into the Outer Carpathian domain are also the autochthonous Tethyan Jurassic, Cretaceous, and Paleogene strata preserved in their original depositional site on the Carpathian foreland. All these Outer Carpathian units, with the sole exception of the Pieniny Klippen Belt, are widely present on the territory of Moravia and, as such, remain the main subjects of our deliberations. The Pieniny Klippen Belt does not extend into this area; however, its complex depositional and structural history is critical for understanding the Outer Carpathian system in its entirety and, therefore, will be discussed as necessary.

The Flysch belt consists of a stack of several rootless tectonostratigraphic units, which, as a whole, is thrusted over the Neogene foredeep and the underlying European foreland. This wedge-shaped, southward-thickening thrust edifice comprises predominantly clastic sequences, whose age extends from the Jurassic to the early Miocene. It includes deposits of both the rifted continental margins and the synorogenic foreland basins, which formed after the Late Cretaceous collision of the Inner Carpathians with the fragmented margins of the West European plate.

The foreland of the Western Carpathians, including its part buried below the thrust belt, belongs to the West European plate (platform). In the territory of Moravia, the West European plate is represented by the Bohemian Massif, which consists of both Hercynian and Precambrian (Cadomian) terranes. Within the Outer Carpathian depositional system, other crustal blocks were present, such as the Silesian and Czorsztyn cordilleras, whose existence and character have been acknowledged only indirectly from the distribution of facies and composition of conglomerates in the Pieniny Klippen Belt and the Flysch belt. Tentatively, we interpret them as rifted away highstanding blocks of the European plate, further inverted during the Laramide uplifting of the foreland. Their adherence to either Hercynian or Cadomian systems may be debated.

THE HISTORY OF INVESTIGATION OF THE CARPATHIAN REGION IN MORAVIA

The geology of the Western Carpathians in Moravia has been studied for more than one-and-a-half centuries. The oldest contributions had a character of rather general observations by naturalists and travelers; more systematic geological investigation began in the second part of the 19th century. Most of the geological mapping and regional work was originally done by the Austrian Geological Survey (K.K. Geologische Reichsanstalt), which was founded in 1849. Foetterle (1866) published the geological map of Moravia and Silesia, and Hauer (1867–1874, 1869 [Western Carpathians]) published the first geological map of the Austro-Hungarian monarchy. Hohenegger (1852, 1855, 1861), based on his studies of macrofauna, laid down the foundation of stratigraphy of the Moravo-Silesian Beskydy Mountains, which, in principle, is valid to the present day.

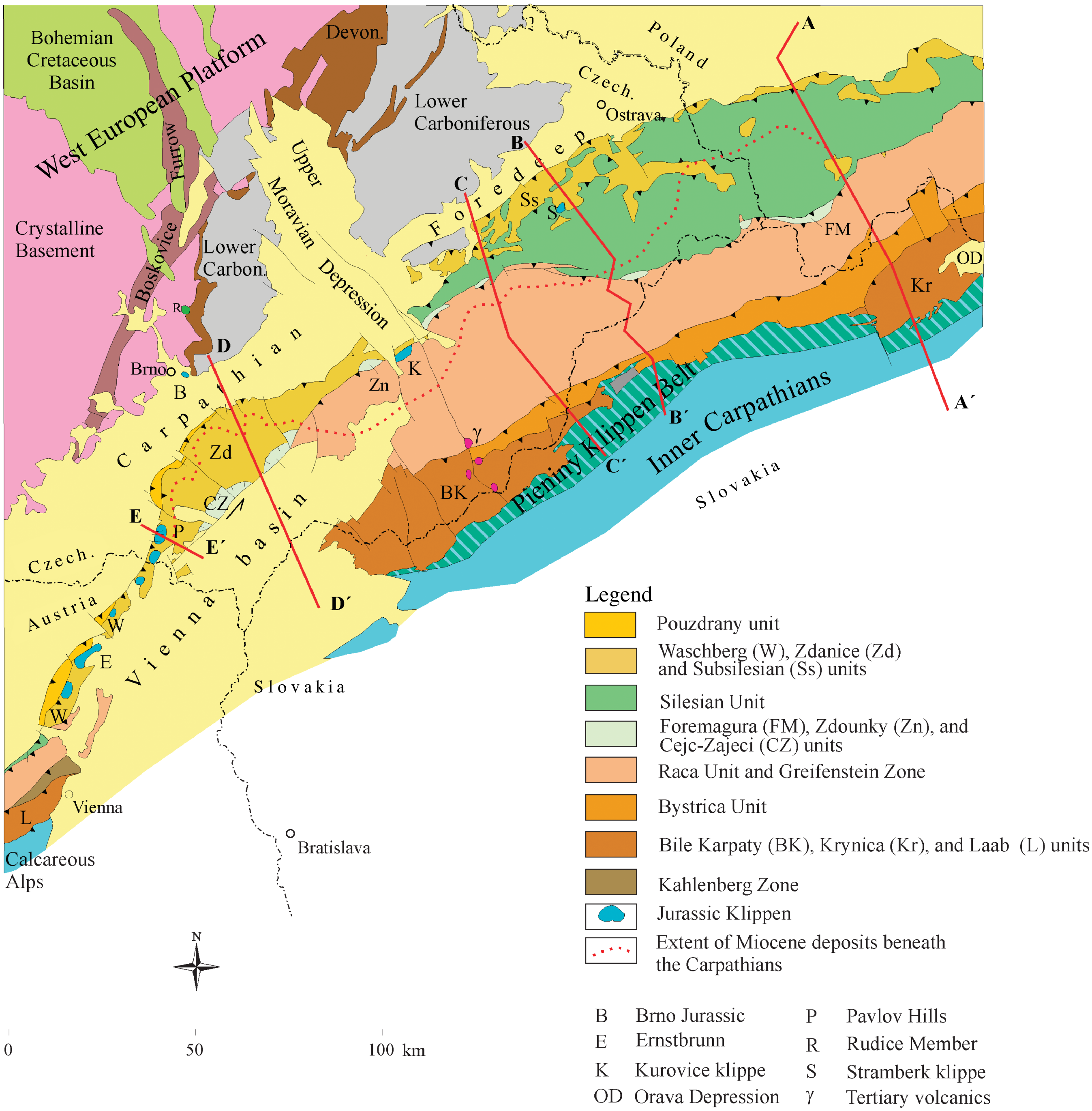

Figure 3. Surface geological map of the Western Carpathians and their foreland (West European platform) in northeastern Austria, Moravia (the eastern part of the Czech Republic), and western Poland. Cross sections of AA′, BB′, CC′, and DD′ are shown in Figure 20 (located on page 118).

In the second part of the 19th century, following the discovery of oil (e.g., Borislav oil field) in Galicia (presently southeastern Poland and southwestern Ukraine), the investigations of the Carpathian flysch belt intensified. Paul and Tietze (1877), Walter and Dunikowski (1883), and Paul (1890, 1893) established the stratigraphy of the flysch belt in Moravia, Slovakia, Poland, and Ukraine (Galicia). At the turn of the century, the Carpathian geology was significantly influenced by the nappe theory established in the Alps by Bertrand (1898), Argand (1911, 1916, 1924), and others. Based on Uhlig's (1897, 1903) observations, Lugeon (1903) published a nappe synthesis of the Carpathians, which, after much dispute, was accepted by Uhlig (1907). The existence of overthrusting was then proven by wells in the Beskydy foothills (Petraschek and Fuchs, 1912).

A new phase of investigation of the Outer Carpathians in Moravia and Silesia began after World War I. The Czechoslovak Geological Survey, as one of the successors of the Austrian Geological Survey, was established in 1919 in the newly founded Czechoslovak Republic. The geological mapping and studies in the Carpathians were motivated mainly by the search for oil and other mineral resources and the construction of railways. A significant factor in further development of the Carpathian geology was the Carpatho-Balkanian Association established in 1922. The Third Congress held in 1931 in Czechoslovakia was one of the most memorable. Beck and Gotzinger (1932), in their explanations to the geological map of the Ostrava–Karvina coal basin, interpreted the nappe structure of the Beskydy Mountains. Andrusov (1938) published the results of his investigation of the Pieniny Klippen Belt, which he considered to be the substratum of the Magura flysch.

A large expansion of geological activities both in the Carpathian belt and its foreland began after the Second World War. The geological activities were driven mainly by the search for raw materials and fuels, including coal, oil, and natural gas. The geological investigations were aided by the application of new techniques and scientific methods, including microbiostratigraphy, sedimentology, geochemistry, and geophysics. The application of seismic, gravimetric, and magnetic geophysical methods represented a significant step in interpreting the architecture of the deeper structural levels of the Carpathian thrust belt and its foreland. Of major importance was the publication of a comprehensive volume on geology of the Czechoslovak Carpathians by Andrusov (1959).

The completion of the geological maps and explanatory texts of all of Czechoslovakia on a scale of 1:200,000 during the 1950s and early 1960s was a major accomplishment. It was followed by a publication of several volumes on the regional geology of Czechoslovakia in Czech and English versions (Svoboda et al., 1966; Mahel and Buday, eds., et al., 1968). In recognition of these achievements, the 23rd International Geological Congress was held in Prague in 1968.

After the publication of general maps, the geological activity concentrated on the detailed investigation and mapping on scales of 1:50,000 and 1:25,000. An intensive exploration for coal and hydrocarbons under the Carpathian foredeep and the thrust belt, which began in the 1950s, invigorated the geological and geophysical activities in Moravia. Deep drilling (see the overview by Suk et al., 1991) and geophysical investigations (Ibrmajer and Suk, 1989) provided critical information about the stratigraphy and structure of both the Carpathian thrust belt and the subthrust plate. Large reserves of coal and several oil and gas fields have been found in various formations below the Carpathian thin-skinned belt. Numerous maps and papers, including the geological atlas of the Western Carpathians (Poprawa and Nemcok, 1988–1989), the palinspastic maps of the Western Carpathian Neogene (Kovac et al., 1998), and the accounts on the geodynamic evolution of the circum-Carpathian region (Golonka et al., 2000) and on the subthrust exploration for hydrocarbons (Picha, 1996), have been published. A great deal of work has also been done in the range of environmental geology, namely, engineering geology (construction of reservoirs, roads, and railroads) and hydrogeology. Traditionally, attention has been paid to geological hazards, especially landslides, which affect large areas of the Carpathian Flysch belt.

After the division of Czechoslovakia into two countries, the Czech Republic and Slovakia, in 1992, the geological mapping and regional investigations in the Moravian Carpathians and their foreland have been conducted by the Czech Geological Survey, which is also one of the primary sponsors of this publication.

GEODYNAMIC EVOLUTION OF THE OUTER WESTERN CARPATHIANS AND THEIR FORELAND IN MORAVIA (CZECH REPUBLIC)

The Tethyan–Alpine orogenic system, as we know it today, evolved during the late Paleozoic, Mesozoic, and Cenozoic through the processes of rifting of the former Pangea, opening of Tethyan oceanic basins, which separated Eurasia from Africa–Apulia, and the development of passive continental margins, followed by the subduction of the oceanic basins, continental collision, and formation of the Carpathian orogenic belt.

Plate kinematic reconstructions of the Alpine–Carpathian system have been published by Dewey et al. (1973), Biju-Duval et al. (1977), Tollmann (1978, 1980), Roth (1980b, 1986), Ziegler (1982, 1988), Dercourt et al. (1986), Le Pichon et al. (1988), Winkler and Slaczka (1994), Plasienka (1995), Yilmaz et al. (1996), Golonka (2000), and Golonka et al. (2000, 2003, 2006a), among others. The architecture of the Alpine orogen was newly interpreted, e.g., by Ziegler and Roure (1996). Still, some differences in the interpretation of the geodynamic history and structure of the region do exist. In this chapter, we concentrate on problems relevant to the Outer Western Carpathians and their foreland in Moravia and only briefly mention some general aspects of the geodynamic evolution of the Alpine–Carpathian region as a whole.

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The Hercynian Setting of the Western Carpathian Region

The Mesozoic and Cenozoic evolution of the Carpathian region resulted from the interplay of numerous lithospheric plates, such as the West and East European platforms, Adria (Apulia), Tisza, and Moesia. All of these plates were parts of the Pangea assembly, which formed during the Hercynian orogeny in the Late Carboniferous. Unlike the linear fold belts, such as the Uralides and the Appalachians, the Hercynian fold belt of Europe displays a complex arcuate architecture. This apparently resulted from the irregular geometry of the preexisting Ordovician to Early Carboniferous rift systems, from draping of the belt around several internal microcratons, as well as from the widespread wrench faulting and rotation of individual blocks (Ziegler, 1988). The curvature of the Hercynian belt is well evident in the Silesian–Moravian sector of the Carpathian foreland, where the generally northwest–southeast-trending front of the Hercynian belt suddenly turns into an anomalous north-northeast–south-southwest direction (orocline bend) apparently being deflected by the Cadomian block of the Brunovistulicum (Figure 4).

In the Moravian–Silesian area, the Hercynian convergent tectonism ended in the Late Carboniferous (Stephanian). The subsequent wrench faulting, rifting, and opening of grabens accompanied by magmatic activity began the process of the fragmentation of the previously assembled orogenic belt. This process is documented by the opening of the arcuate, generally north–south-trending Boskovice Furrow (graben) (Figure 3), filled with the Late Carboniferous (Stephanian C) to Early Permian (Autunian) clastic deposits (Svoboda, ed., et al., 1966; Maly, 1993; Jaros and Maly in Pesek et al., 2001). According to Malkovsky (1976) and Ziegler (1988), this late Hercynian fault system became reactivated time and again. It became a major factor during the Mesozoic rifting stage preceding the opening of the Tethys and the North Atlantic Ocean.

The Late Permian and Mesozoic Rifting and Formation of Passive Tethyan Margins

The evolution of the Eastern Alps–Western Carpathians began in the Late Permian to Early Triassic by rifting of the Tisza plate from Eurasia and the opening of the Meliata–Hallstatt ocean (Channell and Kozur, 1997; Golonka and Gahagan, 1997). The process of rifting and fragmentation of the Pangea assembly gradually spread through the entire Inner Carpathian zone and, in the Early to Middle Jurassic, reached into the Outer Carpathian zone. It led to the opening of the Penninic–Pieninic oceanic basin, the formation of the passive continental margins, and the establishment of a connection with the central Atlantic rift system. The newly formed continental margins were characterized by the development of carbonate platforms and intervening basins (Figures 5, 6).

According to Funk et al. (1987), the crustal stretching in the Tethyan realm was discontinuous, resulting in a system of structural highs with almost normal crustal thickness and structural lows with a thin continental or oceanic crust. This complexity might have

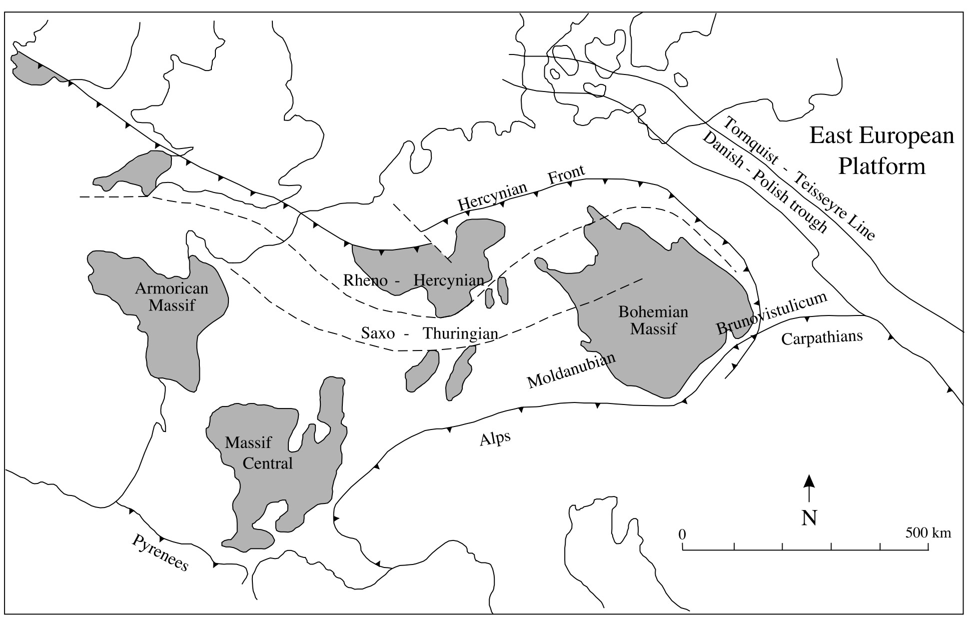


Figure 4. Hercynian system of Western and Central Europe. The Hercynian massifs are in gray pattern. The Bohemian Massif consists of the Rhenohercynian, Saxo-Thuringian, and Moldanubian zones of the Hercynian belt and of the late Precambrian terrane of the Brunovistulicum.

been, at least partly, related to the sinistral strike-slip movements associated with the opening of Tethys between the African and European continents (Boillot et al., 1984). In such a complex and relatively narrow system of Tethyan margins commonly dominated by transcurrent faulting, little space was present for the opening of broad oceanic domains (Plasienka, 1995). The structural and depositional pattern of the complex Tethyan continental margins thus differed markedly from the classical models elaborated for the relatively simple Atlantic-type continental margins.

The Pieninic oceanic basin fully opened by the Middle–Late Jurassic (Birkenmajer, 1986). From the Magura basin and the rest of the Outer Carpathian realm, the Pieniny ocean was separated by the Czorsztyn ridge (e.g., Birkenmajer, 1988). This positive structural feature was marked by a relatively shallow-water carbonate sedimentation, which contrasted with the deep-water basinal environment both north and south of the ridge (Golonka and Sikora, 1981). Golonka et al. (2006a) interpreted the Czorsztyn ridge as a midoceanic ridge. We tend to believe that the Czorsztyn ridge was rather a rifted-apart fragment of the European crust. Its position is similar to that of the Brianconnais zone of the Western Alps (Stampfli, 1993; Tomek, 1993), which is interpreted as a distal element of the European margin (Trumpy, 1988). The Magura flysch basin would then be an equivalent of the Valais trough, which separated the Brianconnais zone from the main part of the European margin. Our further postulations are based on an assumption that the substratum of the Magura basin was made mostly by an attenuated continental crust, and only the deepest zones might have been underlain by a basaltic basement (Figures 5, 6). However, such an interpretation is not accepted by everybody, and different views on this matter do exist even among the authors of this article.

In the Outer Carpathians, the process of divergency and subsidence peaked in the Hauterivian–Aptian. This period was marked by the deposition of black shales and silicites elsewhere in the Outer Carpathian realm. Teschenite intrusive and extrusive magmatism occurred in the highly attenuated Silesian basin.

The Dyje–Thaya Rift-related Depression of Southern Moravia and Northeastern Austria

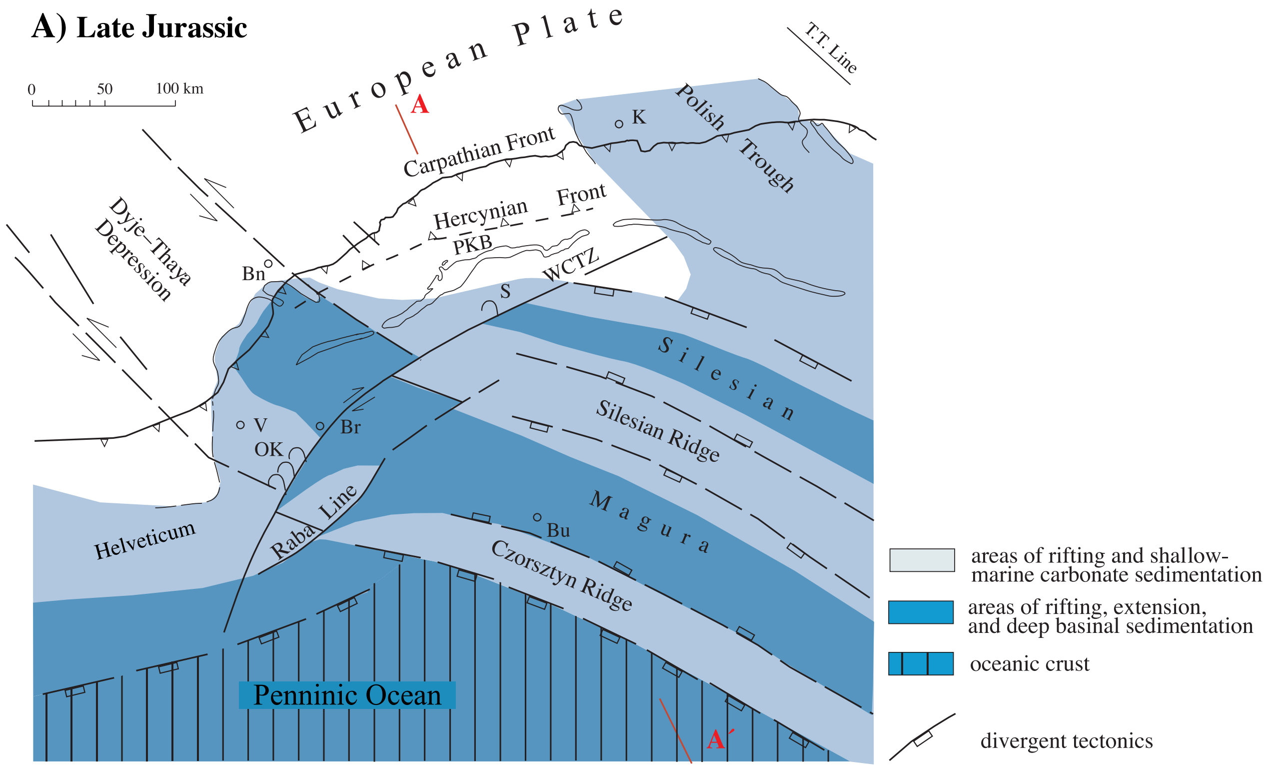
In southern Moravia and northeastern Austria, the Jurassic (Liassic–Doggerian) rifting and extension led to the opening of a deep depression filled with as much as 2000 m (6600 ft) of Jurassic and Upper Cretaceous strata, including as much as 1500 m (4950 ft) of organic-rich Malmian Marls (Picha et al. 1978; Picha, 1979a). This depression seems to trend in the northwest–southeast direction (Picha, 1979a) (Figure 5), parallel with the Danish–Polish trough (e.g., Pozaryski and Zytko, 1980) (Figure 4), the wrench fault system on the southwestern side of the Bohemian Massif (Nachtmann and Wagner, 1987; Schroder, 1987), and some other northwest–southeast-trending Jurassic and Cretaceous rift-related grabens of the Central and Western Europe (e.g., Ziegler, 1988). We call this major depression of the Carpathian foreland the Dyje–Thaya depression according to the river, which flows through it and is called Dyje in Czech and Thaya in German.

The Dyje–Thaya depression of southern Moravia and northeastern Austria is one of the most persistent structural features in the entire Carpathian system. Formed, or at least reactivated, during the Jurassic rifting, this depression and its northwest–southeast-trending bounding faults became a significant factor in the evolution of the Western Carpathian region. During most of the Early Cretaceous, the western part of the depression was uplifted and exposed, whereas sedimentation continued in the adjacent parts of the Silesian, Magura, and Pieniny realms. The subsidence and sedimentation in the depression resumed again toward the end of the Early Cretaceous (Albian) to be interrupted in the early Cenomanian and again at the Cretaceous–Paleogene transition during the Laramide uplifting and inversion of the foreland. Renewed subsidence occurred during the Paleogene to early Miocene. Finally, the Vienna basin formed in the realm of this depression during the Miocene. This repeated history of subsidence indicates that the continental crust of the depression had remained weakened and fragmented since the time of Jurassic rifting.

The complex structural and depositional evolution of the depression also created one of the most important petroleum systems in the entire Carpathian region, from which more than 850 million bbl of oil, generated mainly from the Jurassic source rocks, has been produced to date in Austria, Moravia, and Slovakia.

The Jurassic to Late Cretaceous Convergency in the Inner Carpathians and the Formation of the Outer Carpathian Foreland Basins in the Late Cretaceous

In the Inner Carpathians, the convergence began in the latest Triassic or the earliest Jurassic (Hettangian) by gradual subduction of the Meliata–Hallstatt ocean below the Tisza block (Bukk) (Kozur, 1991; Plasienka, 1995). During the Late Jurassic to the Early Cretaceous, the Meliata–Hallstatt ocean-derived units were deformed, thrusted, and accreted, and the subduction and deformation gradually progressed into the Penninic–Pieninic domain of the Inner Carpathians (Plasienka, 1995).



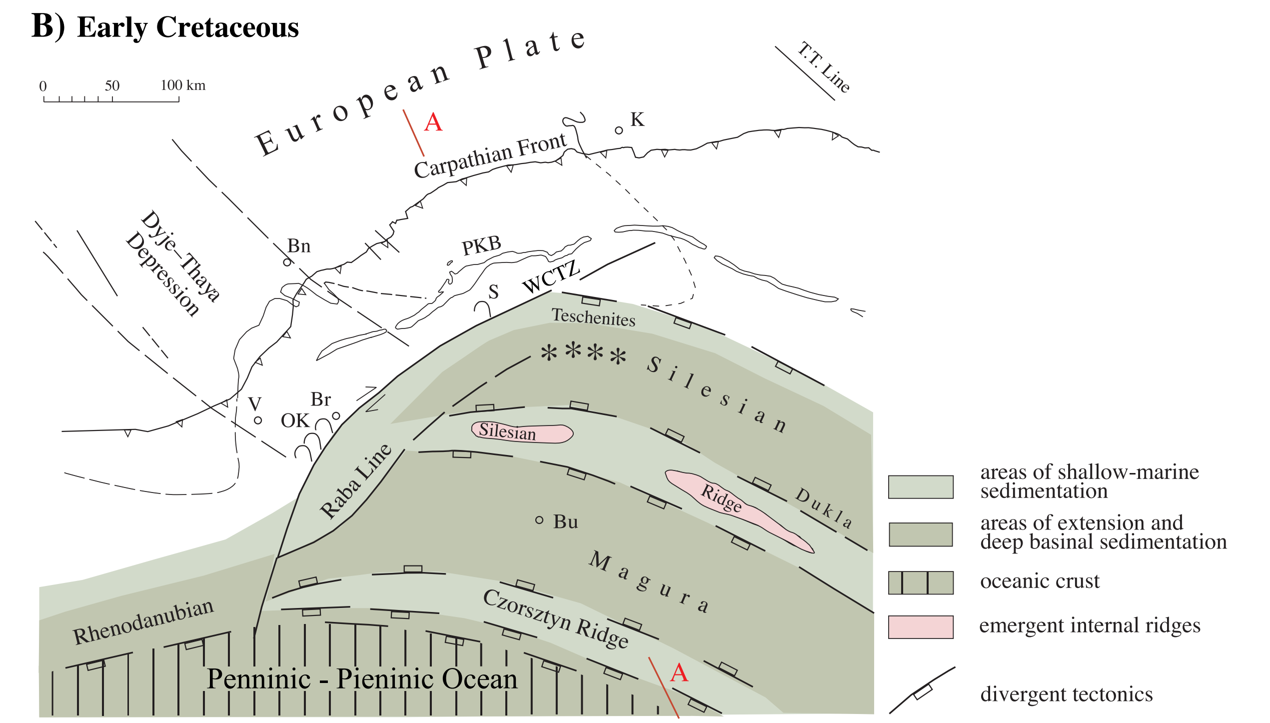
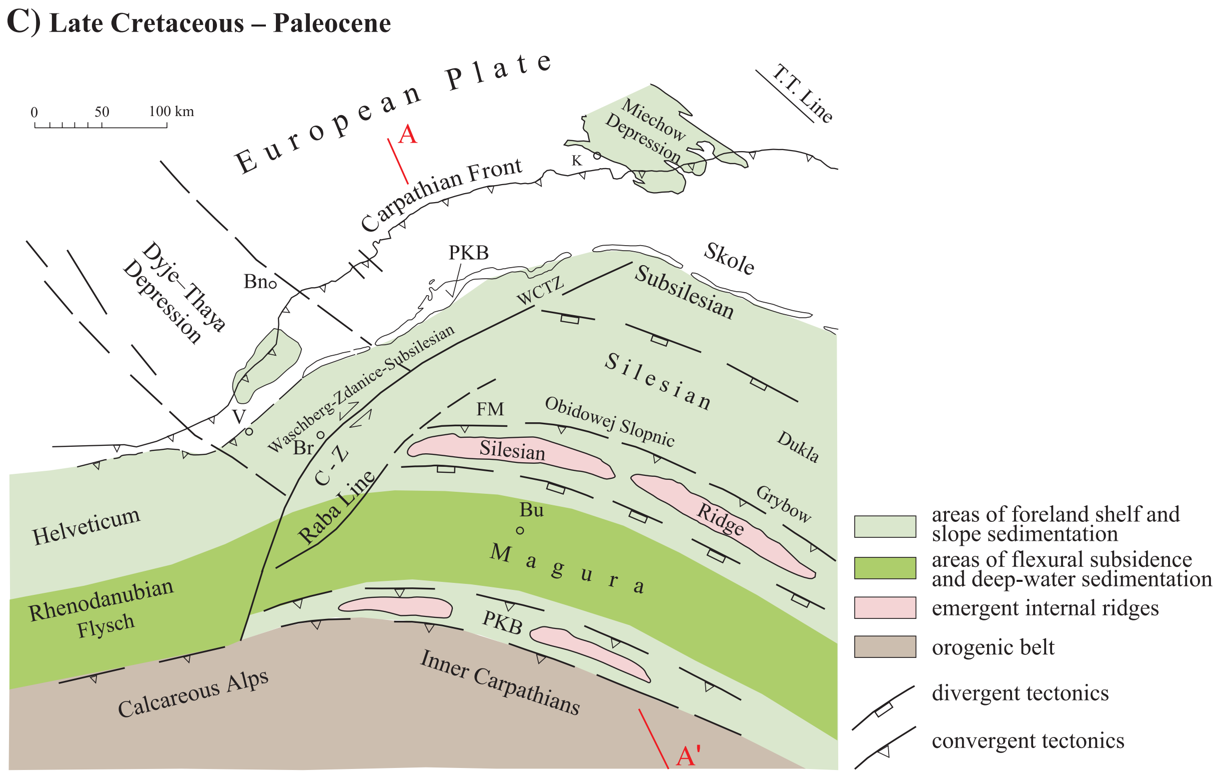


Figure 5. Palinspastic reconstruction of depositional systems of the Western Outer Carpathians during the Late Jurassic (A), the Early Cretaceous (B), the Late Cretaceous (C), and the late Oligocene (D). The distribution of the Middle to the Late Jurassic strata (A) indicates the existence of a northwest–southeast-trending structural pattern of rift-related grabens, such as the Dyje–Thaya depression in southern Moravia and northeastern Austria or the Polish trough and intervening ridges, e.g., the incipient Silesian ridge. In the Early Cretaceous (B), the depositional system was affected by the continuing rifting and extension as well as by the dextral motion and counterclockwise rotation along the southwest–northeast-trending Western Carpathian transfer zone (WCTZ), which separated the Carpathians from the Alps. During this period, the divergent depositional system of the Outer Carpathians reached its maximum extent.



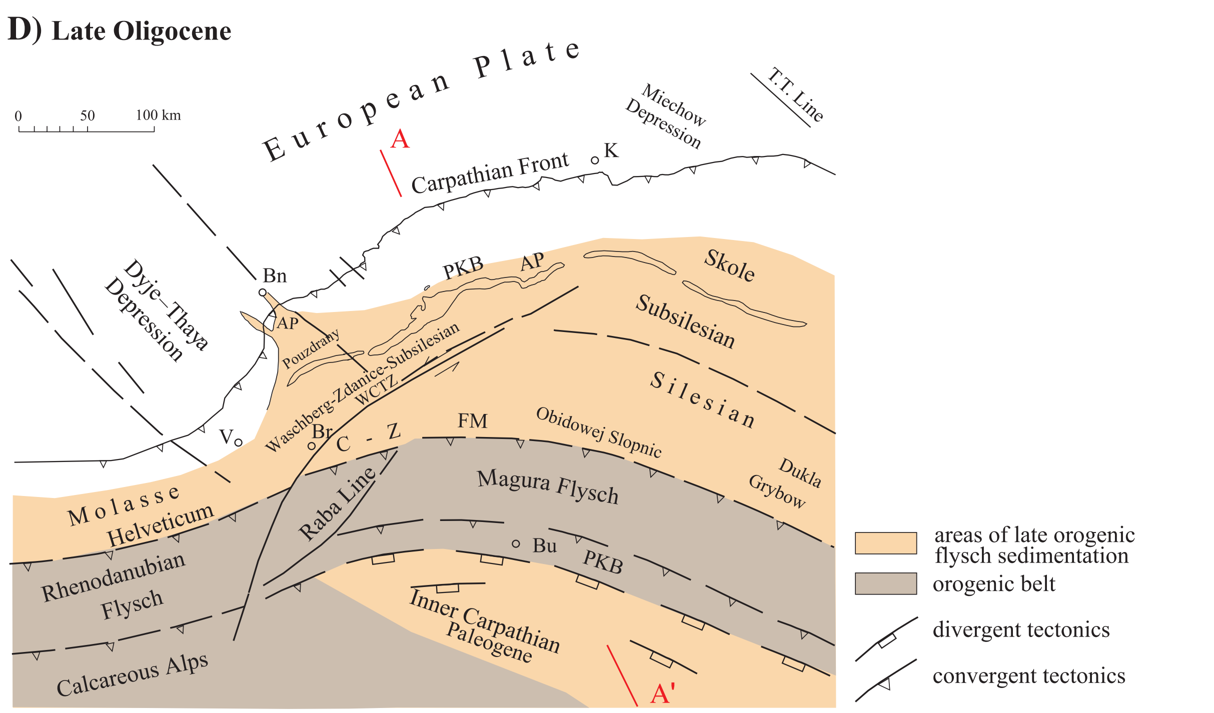
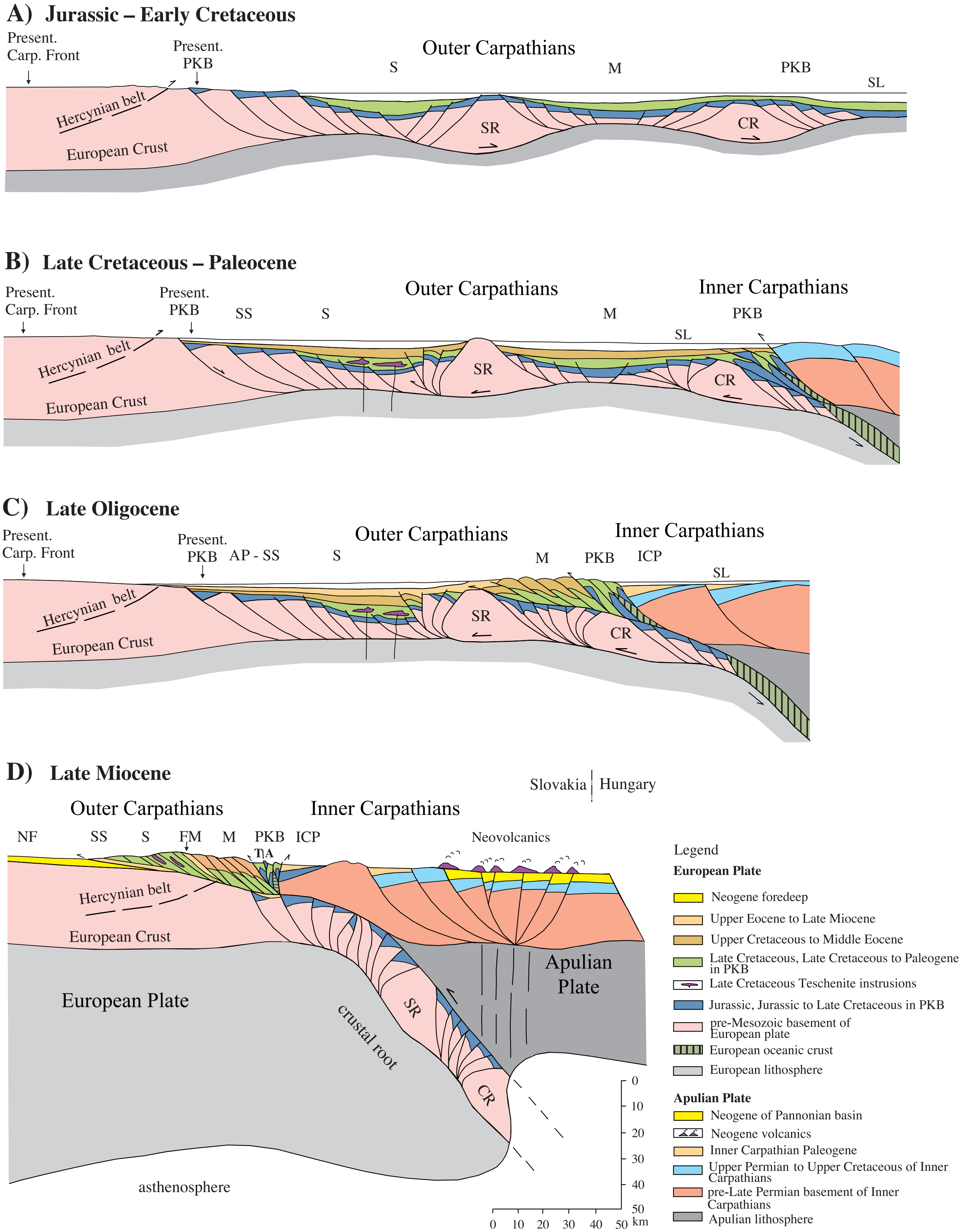


Figure 5. (cont.). In the Late Cretaceous (C), the divergent regime of the Tethyan margins changed into a convergent regime; and the motion along the Western Carpathian transfer zone reversed from dextral to sinistral. The sedimentary system spread farther northwest over the foreland. In the late Oligocene (D), the inner Magura unit was deformed and uplifted, whereas the Krosno-type flysch synorogenic sedimentation continued in the external zones of the Outer Carpathian system. The Inner Carpathian Paleogene basin formed on the top of the Inner Carpathian nappes. Line AA′ marks the section used for the geotectonic reconstructions in Figure 6. AP = autochthonous Paleogene; Bn = Brno; Br = Bratislava; Bu = Budapest; C-Z = Cejc–Zajeci unit; FM = Fore-Magura unit; K = Krakow; OK = Outer Klippen Belt carbonate buildups; PKB = Pieniny Klippen Belt; S = Stramberk carbonate buildup; V = Vienna; WCTZ = Western Carpathian transfer zone.


A) Jurassic – Early Cretaceous
Present. Carp. Front
Present. PKB
Outer Carpathians
S
M
PKB
SL
Hercynian belt
European Crust
SR
CR
B) Late Cretaceous – Paleocene
Present. Carp. Front
Present. PKB
SS
S
Outer Carpathians
M
Inner Carpathians
PKB
SL
Hercynian belt
European Crust
SR
CR
C) Late Oligocene
Present. Carp. Front
Present. PKB
AP - SS
S
Outer Carpathians
M
PKB
ICP
Inner Carpathians
SL
Hercynian belt
European Crust
SR
CR
D) Late Miocene
Slovakia
Hungary
Outer Carpathians
Inner Carpathians
NF
SS
S
FM
M
PKB
ICP
TA
Neovolcanics
Hercynian belt
European Crust
European Plate
crustal root
SR
CR
Apulian Plate
asthenosphere
0
10
20
30
40
50
0 10 20 30 40 50 km
Legend
European Plate
Neogene foredeep
Upper Eocene to Late Miocene
Upper Cretaceous to Middle Eocene
Late Cretaceous, Late Cretaceous to Paleogene in PKB
Late Cretaceous Teschenite instrusions
Jurassic, Jurassic to Late Cretaceous in PKB
pre-Mesozoic basement of European plate
European oceanic crust
European lithosphere
Apulian Plate
Neogene of Pannonian basin
Neogene volcanics
Inner Carpathian Paleogene
Upper Permian to Upper Cretaceous of Inner Carpathians
pre-Late Permian basement of Inner Carpathians
Apulian lithosphere

By the Senonian, all the Inner Carpathian basement-cored units, Gemeric, Veporic, and Tatric were thrusted and overridden by cover nappes of Silicic, Hronic, and Fatric units. Likewise, in the Alps, the basement-cored Austroalpine and Penninic nappes were emplaced on the Helvetic shelf of northern Europe (e.g., Tollman, 1978; Trumpy, 1980).

In the Pieniny Klippen Belt zone, the subduction of the oceanic and attenuated continental crust began in the Senonian (Figure 6). The detached Jurassic and Early Cretaceous strata were accreted toward the front of the Inner Carpathian nappe stack, whereas the sedimentation continued in a piggyback fashion, and synorogenic flysch was deposited at the front of the progressing nappes.

The deformation of the Inner Carpathians and their collision with fragmented margins of the European foreland in the Late Cretaceous marked the beginning of a major rearrangement of the Outer Carpathian depositional realm, which gradually passed from a divergent stage of rifts and passive margins to a convergent stage of prograding foreland basins and foredeeps. These two stages differ by the direction of the depositional progradation and supply of detrital material: in the passive margins, from the continent into the ocean, and in the foreland basin in the opposite direction, from the migrating thrust belt onto the continental foreland.

The depositional system of the foreland basin, which gradually spread over the Outer Carpathian domain and the adjacent part of the European foreland, was dominated by siliciclastic sedimentation of flysch and molasse. Although the synorogenic deep-water flysch sedimentation prevailed in the Late Cretaceous to early Miocene, the late orogenic and postorogenic shallow-marine and continental molasse-type sedimentation continued into the late Miocene and Pliocene.

The Late Cretaceous to Early Paleogene Laramide Uplifting of the Carpathian Foreland

The compressional stresses associated with the Late Cretaceous to early Paleocene deformation and thrusting of the Inner Carpathians and the Pieniny Klippen Belt were transmitted far into the European foreland, which was uplifted and deeply eroded. This Late Cretaceous–early Paleogene deformation of the Alpine–Carpathian foreland is commonly referred to as the Laramide orogenic phase or Laramide orogeny. During this event, some preexisting Jurassic to Lower Cretaceous extensional structures in the foreland as well as the Late Permian to Cretaceous Polish trough were inverted (Roure, et al., 1993; Kusmierek, 1994; Krzywiec, 2002). The Bohemian Massif, including the Carpathian foreland in Moravia, was dissected and, in part, uplifted along a set of wrench and reverse faults (e.g., Roth, 1978; Malkovsky, 1987). The magnitude of uplifting and erosion is well documented by the incision of two paleovalleys and submarine canyons, more than 1500 m (5000 ft) deep, in the Dyje–Thaya Jurassic rift-related depression of southern Moravia and northeastern Austria. The paleovalleys were filled with Paleogene marine deposits, which remained in the autochthonous position below the Carpathian thrust belt and the Neogene foredeep (see the section on the Autochthonous Paleogene and Paleovalleys). The Laramide uplifting and cutting of the deep paleovalleys further increased the structural complexity of the Jurassic Dyje–Thaya depression.

Within the Outer Carpathian depositional system, the Laramide phase led to a further uplifting of the inner ridges (cordilleras), as documented by a sudden influx of the coarse clastic material from the ridges into the Carpathian basins, most prominently into the Silesian and Magura basins.

The Laramide uplifting and inversion of the Carpathian foreland and the rise of internal ridges in the Outer Carpathian depositional domain during the Late Cretaceous to early Paleogene tectonism may be compared with the Late Cretaceous–early Paleogene (Laramide) foreland-type deformation in the North American Rocky Mountains in Colorado and Wyoming. The easternmost of these thick-skinned crystalline basement-involving structures is located more than 500 km (300 mi) east of the front of the thin-skinned Sevier orogenic belt in Utah and Wyoming. By analogy, the foreland-type deformation in the Carpathian

Figure 6. Geotectonic reconstruction of the depositional and structural history of the Western Outer Carpathians from the time of maximum divergence in the Aptian–Albian (A) to the time of the late orogenic convergence in the late Miocene (D). The present interpretation is based on assumptions that the Magura flysch was laid down on a continental and transitional crust, and that the Silesian and Czorsztyn ridges were rifted apart from the European plate and during the convergence accreted back to the European plate instead of being subducted. The term "Apulian plate" (Adria) is used here as a general name for microplates (terranes) of the Carpathian region separated from North Africa. The thickness of sedimentary formations is vertically exaggerated, the European crust is not internally differentiated, and the Inner Carpathians are shown schematically in three colors only. AP = autochthonous Paleogene, CR = Czorsztyn ridge, M = Magura flysch, NF = Neogene foredeep, PKB = Pieniny Klippen Belt, S = Silesian unit, SR = Silesian ridge, SS = Subsilesian unit, SL = sea level. The reconstructions are made along the section AA′ located in Figure 5.

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region apparently also extended for hundreds of kilometers from the thin-skinned front of the Inner Carpathians into the Outer Carpathian depositional system and its foreland.

The Internal (Intrabasinal) Ridges (Cordilleras)

When formed in the Late Cretaceous, the depositional realm of the Outer Carpathian foreland basin displayed a complex rugged relief of deep basins and intervening ridges (cordilleras), such as the Silesian and Andrychow ridges of the Western Carpathians and the Marmarosh Massif of the Eastern Carpathians. The Silesian ridge (cordillera) (Ksiazkiewicz, 1960) separated the Magura flysch basin from the more external Silesian and Subsilesian depositional realm and shed clastics into both of these depositional systems (Figures 5, 6). The less prominent Andrychow ridge, recognized in the Polish sector of the Western Carpathians (Ksiazkiewicz, 1960, 1977), emerged between the Silesian and Subsilesian-Skole basins.

The sudden rise of the Silesian ridge in the Cenomanian–Turonian (Poprawa et al., 2002) is documented by the deposition of thick coarse clastics of the Godula and Istebna formations (Cenomanian–Turonian to Paleocene) in the Silesian basin. Only later, since the late Campanian, were the coarse clastics from the Silesian cordillera supplied into the Solan Formation (Maastrichtian to Paleocene) of the most proximal Raca subunit of the Magura flysch basin. The existence of a significant time gap between the beginning of the deposition of coarse cordillera-derived clastics into the Silesian and Magura units remains unexplained. In southern Moravia, where the Silesian unit is not present, the coarse clastics from the Silesian ridge were shed directly into the adjacent Cejc–Zajeci unit.

Attempts were made to reconstruct the geological history of these internal ridges from the composition of sandstones and conglomerates supposedly sourced from these ridges. Based on studies of Paleocene to Eocene conglomerates in the Cejc Zajeci unit in southern Moravia, Picha et al. (1966) suggested that the part of the Silesian ridge that sourced these conglomerates consisted of epizonal metamorphic rocks with stressed granites and mylonitized zones. The sedimentary cover was represented by Jurassic clastic and carbonate rocks, similar to those found in the Jurassic platforms and basins elsewhere in the Carpathian system. Fragments of crystalline rocks commonly found in some Jurassic limestones indicate that the Jurassic deposits, at least in some areas, were laid down directly on the crystalline basement. Sporadically found pebbles of basic effusive rocks might be assigned to the Permian or Triassic volcanism. In northern Moravia, the conglomerates of the Subsilesian unit contain abundant pebbles and cobbles of Paleozoic rocks (Boucek, 1952; Boucek and Pribyl, 1954, and others). The overall architecture of the inner ridges, their drainage system, and the way the quantities of coarse clastic were transported into the surrounding basins are still poorly understood. The formation of thick but spatially limited subsea fans, such as the Godula and Istebna formations in the Silesian unit, adjacent to the hypothetical ridges, would require an existence of major submarine canyon and channel systems connected with an active drainage network. Further studies of the architecture of these subsea fans combined with the application of modern analogs may provide a better understanding of the character of these ridges and the ways of transportation of the ridge-derived clastics into the flysch basins.

The origin and geological history of the internal ridges have been discussed in various publications. Sandulescu (1988) interpreted the Silesian ridge (cordillera) and corresponding Middle Dacides in the Eastern Carpathians as a thrust belt, which formed during the middle Cretaceous. Picha and Stranik (1999) explained the Silesian cordillera as a foreland-type compressional structure whose rise in the Late Cretaceous was caused by compressional stresses associated with the early collision of the Inner Carpathians with fragmented margins of Europe in the Late Cretaceous. These crustal deformations penetrated into the foreland along the deep decollements, possibly at the base of the continental crust, and led to compression and shortening of the previously attenuated European crust (Figure 6). At least some rise of the Silesian cordillera and other internal ridges might be related to the Laramide uplifting of the European foreland in the Late Cretaceous and early Paleogene (see previous section).

Based on our new paleogeographical reconstructions, we tend to believe that the internal ridges of the Outer Carpathian depositional domain may have initially formed during the Jurassic to Early Cretaceous rifting and extension of the European margins. Thus, already during the divergent stage, these highstanding crustal blocks (horsts) of the rift system separated various depocenters, e.g., the Silesian and Magura basins, in the passive continental margins (Figure 6). At least sporadically, these ridges sourced some flysch-type depositional sequences, e.g., the Hradiste Formation in the Silesian subbasin. The elevated ridges also became sites of the development of carbonate buildups and platforms, such as the Stramberk and Ernsbrunn–Pavlov Outer Klippen (Figure 5). Following the early Late Cretaceous to early Paleogene collision of the Inner Carpathians with the fragmented European margins, these ridges were uplifted and apparently deeply eroded, shedding quantities of coarse clastics into the adjacent flysch basins. Toward the end of the Eocene, the ridges submerged and were buried below the progressing Carpathian thrust belt (Figure 6). Whether they were subducted or accreted to the European crust, the explanation we prefer, remains a matter of discussion.

The Paleogene and Early Neogene Deformation and Thrusting of the Outer Carpathians

The structural and depositional evolution of the Outer Carpathian foreland basin and its equivalents in the Alpine region was primarily controlled by the subduction of the Penninic–Pieninic oceanic and attenuated continental lithosphere underneath the Eastern Alps–Inner Carpathian system (jointly referred to as ALCAPA). This process resulted in a northward shortening and stacking of thrust sheets and a gradual transfer of the depocenters of the foreland basins toward the European foreland. By the end of the Late Cretaceous, the Penninic–Pieninic oceanic lithosphere was consumed, and in the Paleocene, the subduction progressed into the Pieniny–Magura zone of the Outer Carpathians (Birkenmajer, 1986). The deformation and uplifting of the Magura flysch basin, as documented by deposition of Krosno-type strata in the external units, continued into the early Miocene. In the more external Silesian and Waschberg–Zdanice–Subsilesian zones, the deformation and thrusting began in the late Oligocene and lasted into the early Miocene (Karpatian) to the early middle Miocene (Badenian). By then, the Outer Carpathian Flysch belt in Moravia was thrusted over the undeformed Neogene foredeep and the European foreland. However, in the Vrancea Mountains of the Eastern Carpathians, the compressional Alpine tectonics continued into the Pliocene–Quaternary. According to Nemcok et al. (1998a), the Carpathian subduction was ended by a weak collision during the Pannonian (11.5–6.2 Ma).

The advancement of the compressional deformation and thrusting into the Outer Carpathian realm was accompanied by the extension and subsidence in the Inner Carpathian hinterland. Following the marine transgression in the middle Eocene (Lutetian), a deep-water flysch basin formed in the northern zones of the Inner Carpathian nappe stack (Figures 5D, 6). The sedimentation in this Inner Carpathian Paleogene basin continued into the early Miocene. With respect to its previously consolidated Inner Carpathian substratum, this was a postorogenic basin only marginally deformed by the transpressional tectonics along the contact zone with the Pieniny Klippen Belt. No comparable equivalents of the Inner Carpathian Paleogene basin are present in the Alpine region, where the more advanced continental collision apparently did not allow for any significant subsidence and extension during the middle to late Paleogene and early Miocene.

The Late Orogenic Extension and Subsidence in the Carpathian Orogenic System

The youngest phase of the evolution of the Western Carpathians is marked by the melting of the subducting lithosphere and the formation of magmatic arc accompanied by the asthenospheric upwelling (e.g., Lexa and Konecny, 1974; Stegena et al., 1975) and the opening of the back-arc Pannonian Basin (Figure 6D). The development of the Pannonian Basin began with the transtensional rifting and opening of deep pull-apart grabens filled with the middle Miocene (Badenian) strata. After the late Miocene (Pannonian), this fault-dominated extensional stage was followed by the thermally induced regional subsidence, which gave the Pannonian Basin its present form. According to Burchfiel and Royden (1982), the back-arc extension in the Pannonian Basin area was compensated for by a long-range thrusting of the Outer Carpathians over the European foreland and further bending of the Carpathian arc.

The northward thrusting in the Carpathian belt and the back-arc extension in the Pannonian Basin were further modified by a lateral, southwest–northeast-directed material movement (escape) from the Eastern Alps into the Carpathian–Pannonian realm (e.g., Le Pichon et al., 1988; Nemcok et al., 1998b; Ratschbacher et al., 1991a, b). The Pannonian Basin system thus developed in a combined back-arc and escape tectonic setting (Burchfiel and Royden, 1988).

In the early to late Miocene, the Vienna basin formed in the Outer Carpathian and partly Inner Carpathian nappe stack. Its formation and development is related to the combining effect of the continuous subsidence in the Dyje–Thaya depression and the transtensional wrench faulting associated with the northeastern translation (escape) of the Western Carpathians along the Western Carpathian transfer zone (see the next section). In addition, a small Pliocene Orava–Nowy Targ basin formed in a downwarped zone of the Pieniny Klippen Belt and adjacent zones of the Inner Carpathian Paleogene and the Magura flysch in the Orava–Podhale region.

In the late Miocene to Pliocene, the Upper Moravian Depression formed in central Moravia. Bounded by northwest–southeast-trending faults, it extends from the Hercynian foreland across the Neogene foredeep into the Carpathian thrust belt (Figure 3).

Although undeformed, these various late orogenic and postorogenic basins formed primarily in response to the tectonic movements in the orogenic belt and its foreland and, as such, have to be considered as components of the overall Carpathian orogenic system. Of those, only the Vienna basin and the Upper Moravian Depression are present in the territory of Moravia and, hence, the subject of a more substantial treatment in the following sections. The Pannonian Basin is more thoroughly discussed by Tari and Horvath (2006).

The Western Carpathian Transfer Zone

Most structural reconstructions and geophysical interpretations of the Western Carpathians in Moravia and Western Slovakia indicate an existence of a southwest–northeast-trending fault zone (wrench corridor) that roughly parallels the Pieniny Klippen Belt. Various lineaments, and fault zones referred in the literature as the Peripieniny lineament of Maska (Buday et al., 1961; Roth, 1980b), the Vah line, the Zahori fault (O. Fusan, J. Ibrmajer, and J. Plancar, 1979, personal communication), the Verona–Semmering–Vah fault system (Schenk et al., 1994), the Mur–Murtz–Zilina line of Bada et al. (2001), or even the Raba line and some other southwest–northeast-trending faults of western Hungary (Szafian et al., 1999) may, in fact, be elements of this broad structural corridor, which we propose to call the Western Carpathian transfer zone (Figure 5). We understand it as a wide structural zone in which southwest–northeast motion occurred along widely spaced (en echelon) individual wrench faults alternatively activated and newly created during the changing geodynamic settings of the main lithospheric units.

The lateral motion along the strike-slip faults in the transfer zone, predominantly dextral during the divergent stage and sinistral during the convergent stage, was an important factor in the structural and depositional evolution of Western Carpathians and their differentiation from the Eastern Alps. Among the many differences between the external zones of the Eastern Alps and those of the Western Carpathians, the absence of the Pieniny Klippen Belt in the Alpine sector seems to be the most obvious. The typical Pieniny Klippen Belt, as one of the most characteristic features of the Western Carpathians, does not continue into the Alpine domain beyond the projected trend of the transfer zone near the city of Vienna. In addition, the Early Cretaceous rifting and extension, which led to a wide opening of Outer Carpathian basins, the Silesian basin in particular, and possibly to an incipient formation of the internal ridges, was more prominent in the Carpathian domain. The continental collision began in the Western Alps in the late Eocene and led to a significant inversion of the European foreland plate and development of large foreland molasse basins. In the Carpathian realm, east of the transfer zone, the deep-water flysch sedimentation continued into the early Miocene, and the weak continental collision did not lead to a significant activation of the foreland plate and to the uplifting of the external massifs. The sinistral motion along the transfer zone also accommodated the late orogenic extrusion of the Western Carpathians toward the northeast in a process referred to as escape tectonics (e.g., Ratschbacher et al., 1991a, b; Nemcok, 1993; Sperner et al., 2002). The northeastern escape of the Western Carpathians, which began in the early Badenian (middle Miocene), reflects the final adjustment of various crustal blocks to orogenic stresses, which occurred during the late stages of the Alpine orogeny. Associated with this sinistral motion in the transfer zone is also the opening of the pull-apart Vienna basin in the Western Carpathian thrust belt.

Accommodated by the lateral motion in the transfer zone, thus, were all significant differences in the rate of divergency and convergency between the west–east-trending external Alps and the southwest–northeast-running Carpathians. Because of many uncertainties in the palinspastic reconstruction of the Western Carpathians and their foreland, the location of various elements of the Western Carpathian transfer zone can only be guessed at. Tentatively, based on our reconstruction, we would locate this broad southwest–northeast-trending transfer zone in a broad corridor extending from the deep strike-slip faults of the Vienna basin on the west to the faults on the western and eastern side of the present Male Karpaty range and the Raba line on the east.

STRATIGRAPHY AND STRUCTURE OF THE EUROPEAN FORELAND PLATE

The foreland of the Outer Western Carpathians is represented by the West European platform (plate), which is, in numerous publications, referred to as the Epivariscan platform (e.g., Stranik et al., 1993). It is a collage of Precambrian, Caledonian, and Hercynian cratonic terranes assembled in the late Paleozoic and partly covered by the Paleozoic, Mesozoic, and Cenozoic strata (Figures 7, 8). In the Moravian part of the Carpathian foreland, the unmetamorphosed sedimentary cover of the cratonic basement extends over two plate-tectonic cycles, the Paleozoic Hercynian and the Mesozoic to Cenozoic Tethyan–Alpine cycle (Figure 9).

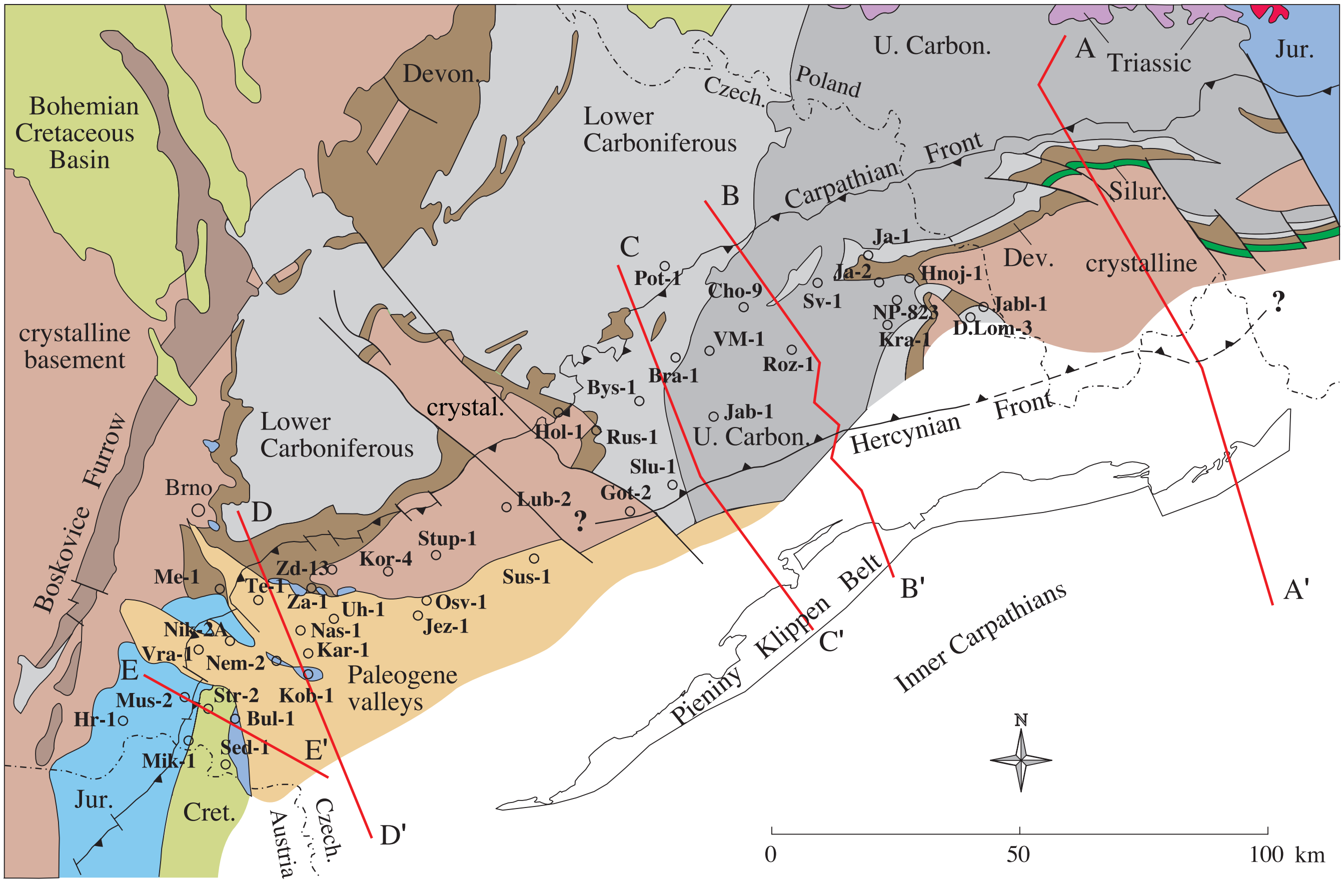

Figure 7. Pre-Neogene subcrop map showing the continuation of Paleozoic, Mesozoic, and Paleogene strata of the West European foreland plate (platform) below the Carpathian thrust belt. Modified from Poprawa and Nemcok (1988) and Picha (1996). Also shown is the anticipated front of the Hercynian orogenic belt below the Carpathian thrust belt. Stratigraphic records of all wells shown on the map are reported in Appendix 1. Cross sections of AA′, BB′, CC′, and DD′ are shown in Figure 20 (located on page 118).

The Cratonic Basement

Bohemian Massif, Brunovistulicum

At the territory of northeastern Austria, Moravia, and Silesia, the cratonic basement of the Carpathian foreland, both in front and under the Carpathian thrust belt, is represented by a complex terrane traditionally called the Bohemian Massif (Figure 4). It actually consists of two primary units: the Hercynian (Variscan) orogenic belt and the late Precambrian (Cadomian) terrane. In the Czech literature, this Precambrian terrane is most commonly known as the Brunovistulicum (Dudek, 1980), a term coined with an emphasis on the present similarity of various segments juxtaposed from southern Moravia to Vistula River in Poland. It is also called Brunnia, a broad paleogeographic term for a continental wedge between the typical Baltica and the Gondwana-related segments (Zapletal, 1931), Brno–Malopolska Block (Suk et al., 1984), having similar meaning as the voluminous assemblage of terrane segments of Brunnia by Zapletal or the East Silesian Massif of Ziegler (1989). This Cadomian assemblage of segments consists of the Brunovistulian, Malopolska, and Lysogory suspect terranes (or groups) (e.g., Valverde-Vaquero et al., 2000). The Hercynian orogenic system is further divided into the internal Moldanubian and Saxo-Thuringian and the external Rheno-Hercynian (Moravo-Silesian) zones (Figure 4). In most of the publications, however, the loosely defined term Bohemian Massif is used for both the Hercynian and late Precambrian terranes. Because, following the Hercynian orogeny and prior to the development of the Tethyan–Alpine system, the Cadomian and Hercynian terranes were accreted into a single West European plate, the usage of the common term Bohemian Massif for both of these terranes in the Alpine–Carpathian literature is acceptable.

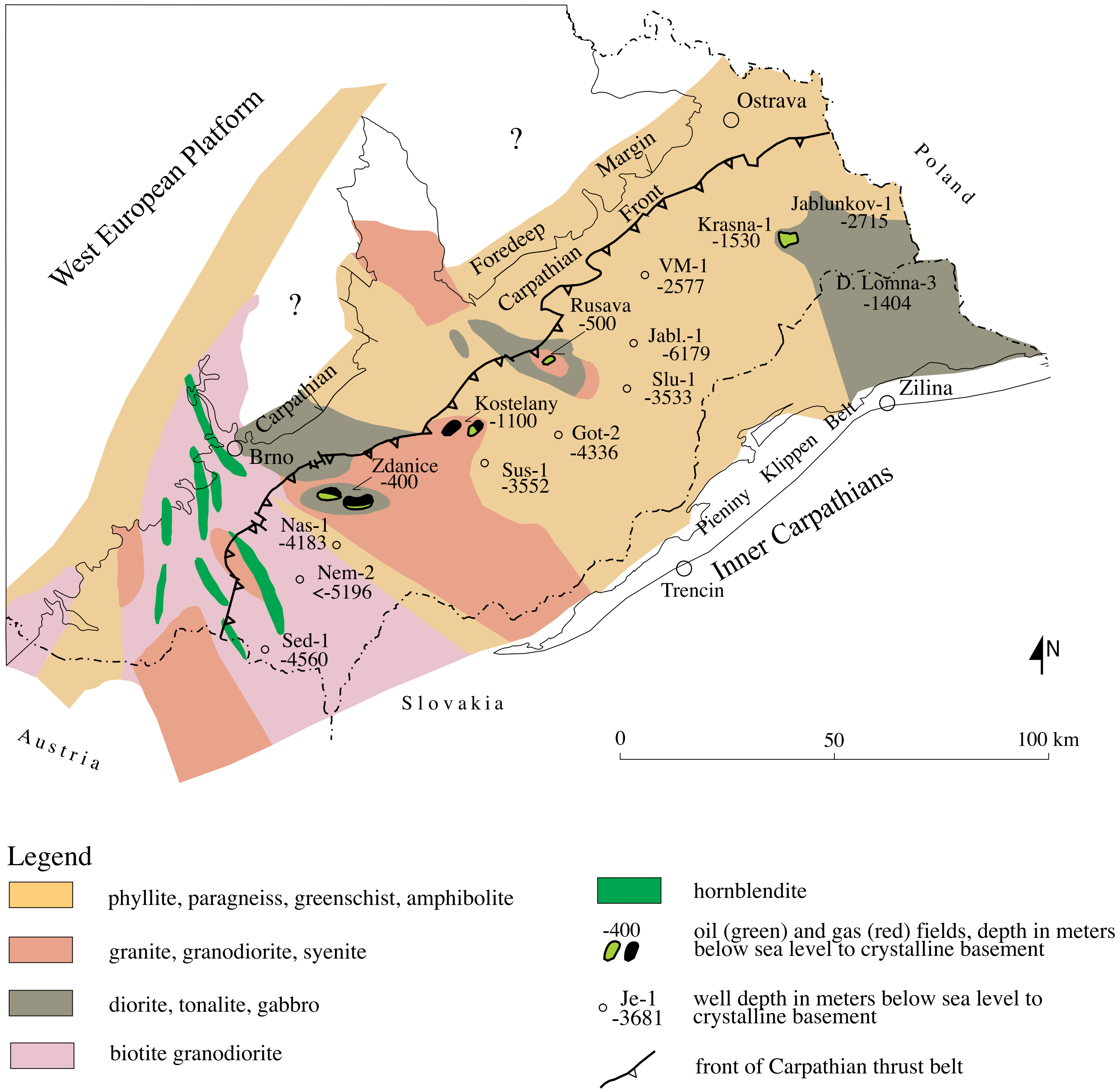

Figure 8. Geological map of the crystalline basement below the Neogene foredeep and the Carpathian thrust belt in Moravia. Modified from Dudek (1980). Stratigraphic records of wells shown on the map are reported in Appendix 1.

Because of the limited amount of seismic data and the overprint by younger tectonic events, the deep contact between the Hercynian system of the Bohemian Massif and the Brunovistulicum in its foreland is little known. Tentatively, the Sternberk–Benesov suture zone of northern Moravia and Silesia and the southern extension of the Boskovice graben in southern Moravia are considered as superficial expressions of such a main geological boundary (e.g., Dudek, 1980). Eastward, a series of Cadomian basement blocks continues into the Upper Silesian block and the Malopolska Massif and underneath the Carpathian Foredeep and thrust belt at least to the suture of the Pieniny Klippen Belt. The relation of the Brunovistulicum to the Carpathian–Pannonian block remains a subject of speculations. Zapletal (1954) suggested that the Brunovistulicum is a promontory of the Fenosarmatian platform inserted between the Bohemian Massif and the Western Carpathians. Others

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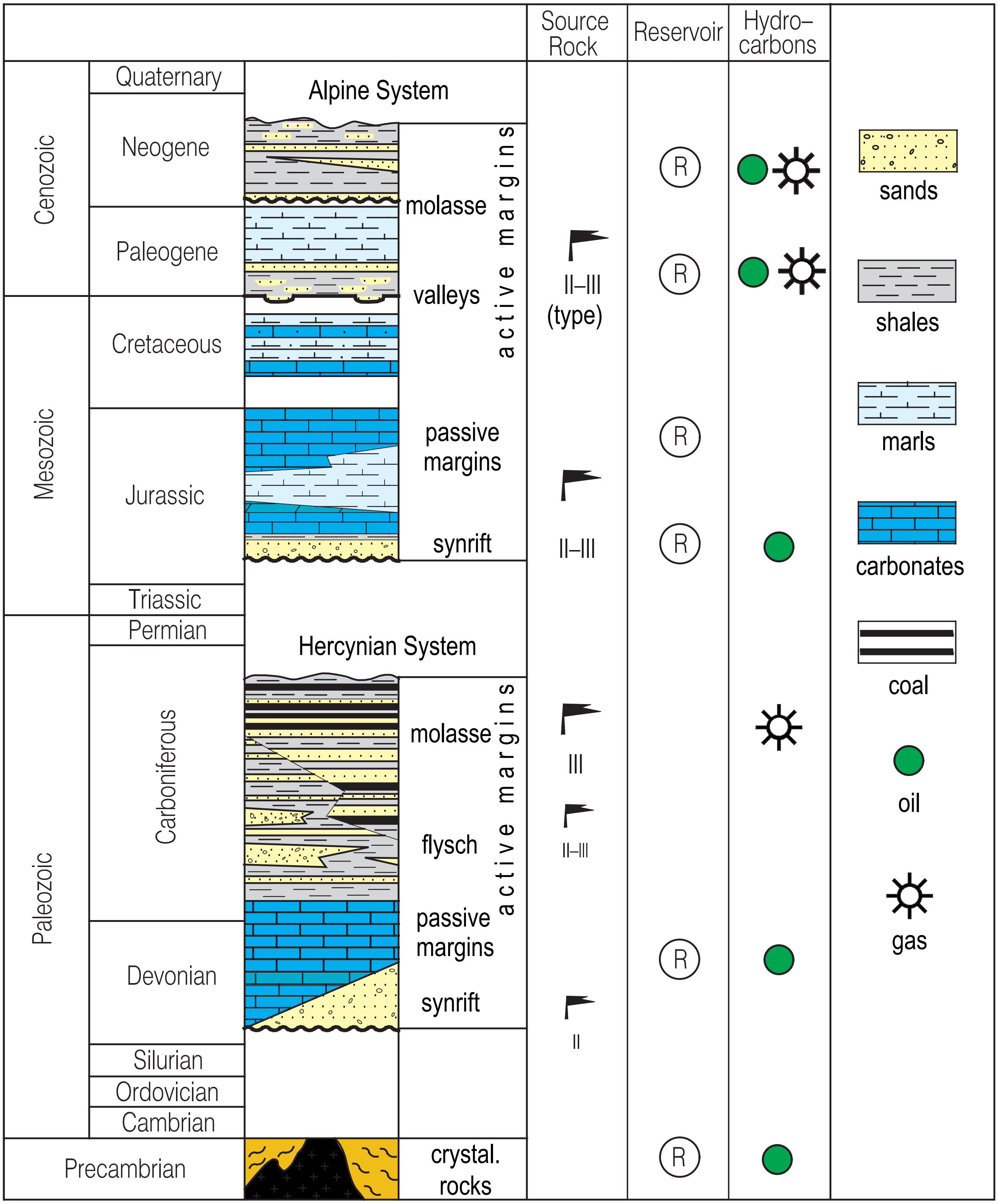


Figure 9. Stratigraphy and hydrocarbon habitat of the foreland zone of the European platform concealed underneath the Neogene foredeep and the Western Carpathian thrust belt in Moravia. Modified after Picha (1996).

believe that the Brunovistulicum is a microcontinent separated from the Gondwana at the beginning of the Paleozoic and positioned at the eastern side of the hypothetical PeriGondwana terrane of Avalonia (e.g., Kalvoda, 1995). The regional extent of the Brunovistulicum has been discussed by Buday and Suk (1989) and Suk (1993). The latest account on the structure of the Brunovistulicum was published by Gnojek and Hubatka (2001).

The Brunovistulicum and other Cadomian terranes thus actually make the essential part of the Western Carpathian foreland basement, both in front and below the thrust belt, from the Danube River in the southwest to the Teisseyre–Tornquist line separating the Western and Eastern European platforms in the northeast. In Moravia, the Brunovistulicum attains a very special position as a cratonic foreland for both the eastward-verging Hercynian and westward-verging Carpathian orogenic belts and as such was tectonically affected by both these orogenies. This is obvious especially on the western side of the Brunovistulian block, where the Early Devonian to Late Carboniferous deposits of the former passive margins of the Brunovistulicum, together with their crystalline basement, were detached and incorporated into the Hercynian thrust belt.

The crystalline basement of the Brunovistulicum consists of both magmatic and metamorphic rocks, the former prevailing in southern Moravia and the latter in central and northern Moravia. Several Cadomian plutons, such as the Brno, Zdanice, and Lubna (Kostelany) massifs, have been recognized and partly mapped below the Carpathian foredeep and thrust belt (Figure 8). The largest of them, the Brno massif, partly exposed in the foreland, is composed predominantly of granites and granodiorites, with subordinate quartz diorites, tonalites, diorites, and gabbros. Sandwiched in the middle of the Brno massif is a north–south-trending unit of tholeitic metabasalts of an oceanic character. Radiometric measurements (Dudek and Melkova, 1975; Finger et al., 2000) indicate that the granitic massifs originated during the final stages of the Cadomian orogeny (580–590 Ma). The metabasalts, including the quartz keratophyre suite and the secondary quartzites, are more than 725 Ma (Finger et al., 2000). Their contact with granites is tectonic, possibly accretionary (Hanzl and Melichar, 1997). The metamorphic rocks, prevalent in northern Moravia, are represented by various types of slightly migmatized paragneisses, greenschists, phylites, and amphibolites.

The relief map of the crystalline basement of the Brunovistulicum below the Carpathian foredeep and thrust belt (Gnojek and Hubatka, 2001) displays a rugged relief marked by southeastward-running erosional-like features, most commonly present in the southern and central Moravia and less visible in the northern Beskydy region.

Commercial accumulations of oil and gas have been found in the weathered and fractured surface of the granitic rocks of the Zdanice, Lubna, and Krasna elevations (Krejci, 1993; Blizkovsky et al., 1994).

The Hercynian Cycle

The predominantly unmetamorphosed Paleozoic rocks of the external zones of the Hercynian belt thrusted eastward over the bulged Brunovistulicum are exposed in the Drahany and Nizky Jesenik highlands of Moravia and Silesia. The extension of Paleozoic strata, both allochthonous and autochthonous, further east is buried below the Neogene foredeep and the Carpathian thrust belt. However, numerous deep wells drilled in search of coal and hydrocarbons in this area provide information about the distribution of Paleozoic rocks below the Carpathian system (Figure 7). The stratigraphy and structure of the Paleozoic strata adjacent and below the Carpathian thrust belt have been discussed in numerous papers, e.g., Kettner (1950, 1970), Dvorak and Ptak (1963), Dvorak (1973, 1978, 1993, 1995), Zukalova (1976), Adamek et al. (1980), Zukalova and Chlupac (1982), Hladil (1986, 1988), Chlupac (1989, 1994), and Hladil et al. (1994, 1999).

Considering their geodynamic evolution, the Paleozoic strata of the Carpathian foreland could be divided into three major sequences: (1) the Lower Devonian to Lower Carboniferous, synrift and passive margins sequence; (2) the Lower Carboniferous synorogenic flysch (Culm) sequence; and (3) the Upper Carboniferous late orogenic and postorogenic sequence of the Upper Silesian coal basin, including the Namurian relicts in southern Moravia.

The Lower Devonian to Lower Carboniferous Synrift and Passive Margins Sequence

The Paleozoic Hercynian plate-tectonic cycle in eastern Moravia begins in late Early Devonian (Emsian) with continental rifting and deposition of terrestrial and shallow-marine synrift clastics, commonly especially in the older literature, called the ''Old Red Sandstone'' (Figure 10). The thickness of these deposits ranges from a few meters to as much as more than 1000 m (3300 ft), e.g., in the deep well Menin-1. Based on finds of acritarch fauna, some of the synrift deposits have been assigned to the Lower Cambrian (Jachowicz and Prichystal, 1997). However, the existence of presumably Cambrian clastics below the lithologically indistinguishable Devonian clastics in the same wells (e.g.,

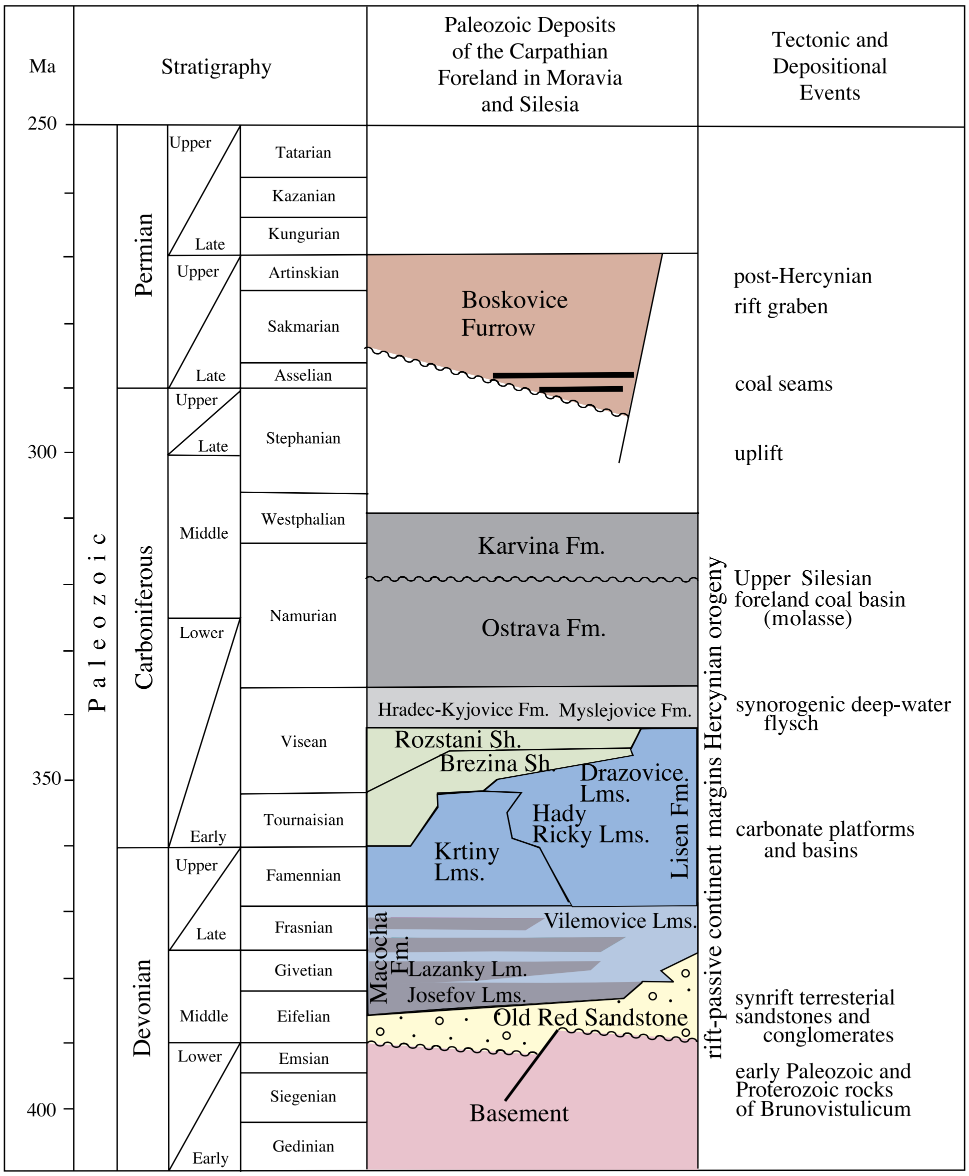


Figure 10. Stratigraphy of the Paleozoic strata of the European foreland plate in Moravia and Silesia, Czech Republic.

Menin-1) and the possibility of redeposition of the acritarch fauna make the presence of Cambrian strata disputable.

The synrift clastic sedimentation of terrestrial conglomerates, arcosic, and quartzitic sandstones was followed by a widespread marine transgression and development of carbonate platforms and ramps on the passive continental margins of the Brunovistulicum. The predominantly carbonate sedimentation continued through the Late Devonian into the Early Carboniferous (Tournaisian–Visean). The lower part of the carbonate section is represented by the Macocha Formation (Eifelian to Frasnian) deposited in an environment of large carbonate platforms with lagoons and reefs. Hladil (1983, 1986) recognized in the Macocha Formation four cycles (depositional sequences) separated by hiatuses marked by karstified surfaces, fossil soils, or layers of sandstones. These sequences commonly begin with dark-gray dolomitic limestones with brachiopods (Josefov Limestones), followed by dark-layered Amphipora limestones (Lazanky Limestones), and terminate with light-gray massive limestones with abundant coral and stromatoporoid fauna (Vilemovice Limestones). The overall thickness of the Macocha Formation is about 400 m (1300 ft); locally, it may exceed 1000 m (3300 ft). The highly diagenetically altered and karstified Vilemovice Limestones, especially their thinner equivalents that developed on paleogeographic highs, represent one of the most promising Paleozoic reservoirs for hydrocarbons, especially in the area underneath the Carpathian foredeep and thrust belt (Hladil et al., 1994).

The overlying Lisen Formation of the Famennian to Visean age is marked by the disappearance of coral and stromatoporoids apparently caused by the global Kellwasser cooling event at the Frasnian–Famennian transition. In the absence of carbonate buildups, the sedimentation tended to smooth the relief of the carbonate platforms. Unlike the underlying Macocha Formation, the Lisen Formation is characterized by a large diversity of facies. Several types of limestones, such as the nodular, micritic, and argillaceous Krtiny limestones, the turbiditic and argillaceous bituminous cherty Hady–Ricky Limestones, and the shallow-marine partly oolitic Drazovice Limestones, have been distinguished. The thickness of the Lisen Formation ranges from less than 50 to about 300 m (160 to about 1000 ft). Because of a fine-grained structure and a higher content of argillaceous material, the reservoir properties of the Lisen Formation are poor; the formation may be considered as a sealing horizon rather than a reservoir.

The overlying, mostly argillaceous deposits marked by occasional tuffaceous horizons and numerous hiatuses are known under the local names of Ostrov shale, Brezina shale, and Rozstani Shale. They represent a condensed transitional series between the predominantly carbonate sedimentation of the divergent continental margins and the clastic sedimentation of the convergent stage of the Hercynian cycle (Figure 10). This transition, however, did not occur simultaneously in the entire area, and consequently, the stratigraphic relationship of these transitional strata to both the underlying carbonates and the overlying younger clastic deposits might be diachronous. The most elevated blocks of the depositional environment have been emergent since the late Frasnian to the late Visean or even to the earliest Namurian.

The Lower Carboniferous Synorogenic Flysch (Culm)

At the onset of the Hercynian orogeny, in the Visean, the synorogenic clastic deposits supplied from the inner zones of the orogenic belt gradually spread over the carbonate platforms of the Hercynian foreland. They are represented by the deep-water turbiditic flysch facies, of alternating conglomerates, sandstones, and shales, called Culm.

The flysch deposits were laid down in a predominantly deep-water system of prograding synorogenic foredeeps. Their overall thickness decreases from several thousand meters at the western orogenic side of the depositional system, e.g., in the Nizky Jesenik Mountains, to only a few hundred meters on the distal eastern and northeastern side of the Hercynian foreland. In addition, the onset of the Culm sedimentation tended to be progressively younger toward the more stable zone of the Brunovistulican platform. In the zone adjacent to the present front of the Carpathians, the carbonate sedimentation may have lasted into the very late Visean (Dvorak, 1978), and the overlying Culm deposits are represented by a distal predominantly argillaceous synorogenic deep-water flysch facies of the Hradec–Kyjovice Formation and the Myslejovice Formation (Figure 10). The distribution of facies in the Myslejovice Formation indicates an existence of a large subsea fan system in this part of the Hercynian foredeep. The proximal part of this fan, located in the Drahany Highland, is made by the Racice and Lulec conglomerates of the channelized inner fan, which, toward the north-northwest, pass into a predominantly sandy turbiditic facies and even further into a distal, predominantly shaly facies of the foreland.

In the Carpathian region, the predominantly distal facies of the Lower Carboniferous strata of the Culm do not possess any source rock or reservoir properties; however, they may function as a regional seal for the potential underlying carbonate reservoirs.

The Upper Carboniferous, Upper Silesian Coal Basin

In the early Namurian, the deep-water flysch-type Culm sedimentation was replaced by the late and postorogenic shallow-marine and nonmarine molasse of the Upper Silesian coal basin, which formed in the distal zone of the Brunovistulican foreland in northern Moravia and Silesia. The transition from the marine Culm facies into the paralic molasse sedimentation of the Upper Silesian basin was gradual, reflecting on the gradual marine regression during the Namurian. As documented by remnants of the Upper Carboniferous strata in the Nemcicky-1 well in southern Moravia and occurrences of pebbles of Carboniferous rocks in the Cretaceous and Paleogene conglomerates of the Outer Carpathians, the original extent of the Upper Carboniferous molasse basin was apparently much larger. Situated on the outer side of the Hercynian orogen, the Upper Silesian basin has a similar position as some other coal basins of Western Europe, e.g., in Wales, Belgium, and Ruhrland.

A systematic investigation of the Upper Silesian coal basin began in the 1860s. Stur (1875, 1885) described marine fauna and flora of the Carboniferous strata. Jicinsky (1885) and Folprecht and Pateisky (1928) published monographic studies on the basin. After the Second World War, systematic work was done by Havlena (1964, 1982), Rehor and Rehorova (1972), Dopita and Havlena (1977), and Dopita and Kumpera (1993), among others. The present knowledge of the basin has been compiled in a comprehensive monography by Dopita et al. (1997).

Stratigraphically, the Upper Carboniferous strata in Moravia extend from Namurian A to Westphalian A. They are divided into the Ostrava and Karvina formations, each of them further subdivided into several members (Figure 10).

The Ostrava Formation, as much as 3200 m (10,500 ft) thick, belongs to the lower Namurian. It is a coal-bearing paralic molasse consisting of numerous cyclothems, formed through the transition from the fluviatile to lacustrine, swampy, and lagoonal environment. A typical cyclothem begins with the deposition of basal fluviatile sandstones and/or conglomerates, followed by lacustrine siltstones, a coal seam formed in a swampy environment and completed with deposition of lagoonal clays with fresh water, brackish, and marine fauna. Four major marine horizons in the Ostrava Formation are used as primary correlation horizons. The correlation of strata is further helped by the presence of several horizons of acidic tuffs and tuffites (Tonsteins). Swarms of dikes and sills of acidic and intermediate magmatic rocks are found locally.

As many as 500 coal seams are present in the Ostrava Formation, of which about one quarter is mineable. The overall thickness of the Ostrava Formation and, with that, also the number and thickness of coal beds decrease eastward toward the foreland.

In southern Moravia, the Ostrava Formation is known from deep wells. In the Damborice-1 well, the Ostrava Formation, 523 m (1715 ft) thick, evolves gradually from the underlying Myslejovice Formation of the Culm without any apparent interruption. However, in the Nemcicky-1 and Nemcicky-2 wells, the 150-m (500-ft)-thick basal conglomerates of the 1300-m (4300-ft)-thick Ostrava Formation rest on an erosional surface of the thin shaly Myslejovice Formation. The upper part of the formation consists of depositional cycles with coal seams similar to those found in the northern part of the Upper Silesian basin. The existing evidence, however, would exclude the presence of marine horizons in this part of the basin.

The overlying Karvina Formation, several hundred meters thick, also consists of numerous cycles, but unlike the Ostrava Formation, it does not have any marine horizons and therefore represents a typical continental coal-bearing molasse. Twenty-three freshwater horizons important for correlation have been distinguished in the formation. The coal seams, of which about 90 are mineable, are typically thicker but less numerous than in the Ostrava Formation. The Karvina Formation is more widely present in the Polish part of the Upper Silesian basin.

In the Czech part of the Upper Silesian basin, the sedimentation ended in the lower Westphalian, whereas in the Polish territory, sedimentation continued into the Stephanian. The top of the Karvina Formation in the Czech sector of the basin is deeply eroded and covered by Miocene deposits. Two large paleovalleys, Detmarovice and Bludovice, were cut into the Upper Carboniferous strata, apparently during the Late Cretaceous–Paleocene Laramide uplifting of the Carpathian foreland, and filled with Karpatian to Badenian detrital deposits.

The Upper Silesian basin was partly deformed during the last stages of the Hercynian orogeny, which, in the Moravo-Silesian zone, came to an end during the Westphalian. The Westphalian strata are only lightly folded, and the Stephanian strata remained undeformed. Locally, they rest transgressively on the older, moderately folded Westphalian strata. The frontal compressional structures, such as the Orlova fold, are bounded by reverse faults rather than the low-angle thrust faults.

The Upper Silesian basin is a significant black coal-producing province. Evidence exists that coal was used there during the last ice age by people of the Gravettian culture, about 23,000 yr ago. The present length of the subsurface tunnels is in a range of several thousands of kilometers. The first deep well was drilled in 1867; the present length of all wells drilled in the basin exceeds 1800 km (1100 mi).

The Upper Carboniferous strata of the Upper Silesian coal basin, with the exception of some weathered surfaces, are not considered to be reservoir rocks for hydrocarbons. However, they represent a significant source of gas, some of which accumulated in the overlying Tertiary deposits. So far, any attempts to produce commercial quantities of the coalbed methane, both in the Czech and Polish parts of the basin, have not been successful.

The Extent of Hercynian Deformations underneath the Outer Western Carpathians

Compressional Hercynian structures underneath the Carpathian foredeep and the Flysch belt in eastern Moravia have been reported by Jurkova (1979), Mencik (in Poprawa and Nemcok, 1988–1989), Cizek and Tomek (1991), Stranik et al. (1993), Picha (1996), Grygar and Jelinek (1999), and Hladil et al. (2000). Using well data and regional seismic lines, Picha (1996) tentatively mapped the front of the Hercynian belt below the Carpathian belt in northeastern Moravia. The frontal edge of the Hercynian belt, as mapped, roughly parallels the front of the overlying Carpathian belt but verges in an opposite southwest direction (Figure 7). The continuation of the Hercynian structures to southern Moravia and Austria, however, remains uncertain. Outside the Carpathian realm, the Hercynian thrusting has been recognized in the Drahany Highland west of Brno, where the high-grade metamorphic rocks of Moldanubicum are thrusted over the low-grade metamorphic rocks of Moravicum. Similarly, in the Moravian karst north of Brno, the Cadomian granitic rocks of the Brno massif are thrusted over Devonian carbonates (Kettner, 1950). Below the Carpathian thrust belt, the existence of a steep Hercynian thrust was proven by the Nemcicky wells (Krejci et al., 2002). South of the tectonically induced Nesvacilka paleovalley, the recognition of the Hercynian structures in the Carpathian foreland is further complicated by the absence of Paleozoic strata (Figure 7), which, in that area, were either eroded or initially not deposited. Tentatively, we interpret some of the structural features seen on the seismic lines in southern Moravia as Hercynian, thus extending the frontal zone of the Hercynian belt to the western margins of the Vienna basin (Figure 20 DD′, shown on page 118).

An existence of moderate Hercynian deformation in a zone adjacent to the Carpathian belt in southern Poland has been reported, among others, by Dadlez et al. (1994). Kotanski (1997) proposed to call this zone, where the Hercynian folds overprint the older Caledonian and Cadomian deformations, the Peri-Variscicum (Peri-Hercynicum).

The frontal anticlinal structures of the Hercynian belt are bounded by reverse faults and high-angle thrust faults rather than by the low-angle thrust faults. Typically, the thrusting involves both the Paleozoic (Devonian to Late Carboniferous) strata and the Cadomian crystalline basement. We interpret these thick-skinned folds as foreland-type structures detached along a deeper decollement in the crystalline basement. They differ markedly from the thin-skinned structures known from the external zone of the Hercynian belt in the Odra Hills and the Nizky Jesenik Mountains, as reported by Cizek and Tomek (1991). This implies that in the external Hercynides in northern Moravia and adjacent zones of Poland, two sets of compressional structures are present: the thin-skinned structures, comprising only the sedimentary rocks, and the thick-skinned foreland-type structures involving the crystalline basement. In that sense, the frontal Hercynides resemble some other orogenic belts, e.g., the Rocky Mountains of North America or the Ural Mountains of Russia, where both types of structures have been recognized.

We caution, however, that not all deep antiformal structures bounded by northwestward-dipping reverse faults visible on seismic lines are necessarily of Hercynian age. The origin of some of the antithetic reverse faults (with respect to the Carpathians) and associated anticlinal structures in the subthrust basement may also be related to the Laramide (Late Cretaceous–Paleogene) uplifting of the foreland or to the late orogenic transpressional strike-slip faulting, commonly identified in the Pieniny Klippen Belt. Depending on the distribution of compressional stresses, the sinistral motion of crustal blocks along the Western Carpathian transfer fault zone might have generated both synthetic and antithetic compressional structures. One such questionable structure (well documented on the seismic sections 8HR, 2T) is the Orava structure. The antithetic faults bounding this structure on its eastern side may be either Hercynian, as we interpret it on the AA′ cross section (Figure 20, shown on page 118), or much younger, related to the Alpine convergent orogeny. Krzywiec (2001) explained similar south-verging structures, seen on seismic lines in the eastern part of the Polish Carpathian foreland, by rotation of basement blocks along inherited synthetic southward-dipping normal faults. Such a rotation will induce extension on the northern side of the rotating blocks and compressional reverse faulting at their southern edges.

The potential Hercynian structures below the thin-skinned Carpathian belt may represent an interesting target for exploration. These structures and potential reservoirs in them were in place prior to the generation and migration of hydrocarbons in the Carpathian system. Unless further disrupted by younger tectonism, they may still hold hydrocarbons (Picha, 1996).

Similarities of the Hercynian and the Carpathian Orogenic Belts in Moravia

The external zone of the Hercynian belt in Moravia is a mirror image of the Outer Carpathian belt. Despite the different age, the late Paleozoic for the Hercynian belt and the Mesozoic–Cenozoic for the Carpathian belt, the opposite verging frontal zones of these two orogenic belts display many similarities in their depositional and structural setting (Figure 7). Both the Hercynian and the Carpathian external zones are represented by the synorogenic and late to postorogenic flysch and molasse sequences. These are separated from the older, more internal zones by narrow structurally complex sutures, the Pieniny Klippen Belt in the Carpathians and the Sternberk–Benesov suture in the Hercynian belt. The territory of central and northern Moravia thus provides a unique opportunity to study the frontal zones of these two major orogenic belts of Europe in the same relatively small area.

The Tethyan–Alpine Cycle

In the Carpathian foreland of Moravia and northeastern Austria, the Mesozoic to Cenozoic Tethyan–Alpine cycle began in the Early to Middle Jurassic with continental rifting and extension, followed by the marine transgression and development of passive continental margins dominated by carbonate sedimentation in the uppermost Middle and Late Jurassic (Figure 9).

No direct evidence of the existence of a Triassic strata in the Outer Carpathians and their foreland in Moravia is available, although clasts of Triassic detrital, carbonate, and possibly volcanic rocks have been reported from conglomerates in the Carpathian Flysch belt (Sotak, 1985). Considering the long-distance thrusting and orogen-parallel lateral translation of various units of the Flysch belt, the original source of these Triassic rocks should be sought in areas quite removed from the present location of the conglomeratic bodies in the Flysch belt. The most likely source would be the Silesian cordillera, a tectonically active ridge that separated the external units of the Outer Carpathians from the Magura flysch or the collisional zone between the Inner and Outer Carpathians, which comprised elements of the Czorsztyn ridge, Pieniny Klippen Belt, and the frontal units of the Inner Carpathians.

During most of the Early Cretaceous, the area of southern Moravia and northeastern Austria was uplifted. The marine carbonate sedimentation marginally resumed in the Aptian–Albian, followed by a major Cenomanian transgression and deposition of siliciclastic rocks. During the Laramide uplifting and erosion, at the Cretaceous–Paleogene transition, large paleovalleys were excavated in the Carpathian foreland of Moravia and later filled with transgressive Paleogene strata of the Carpathian foreland basin, which has gradually evolved since the latest Cretaceous. The geological history of the foreland region ended with the formation of the Carpathian foredeep in the early to middle Miocene.

During the various phases of the Alpine–Carpathian orogeny, most of the Mesozoic and Cenozoic strata were deformed and thrusted to form the Carpathian thrust belt. Only on the foreland periphery, some of the Jurassic, Upper Cretaceous, and Paleogene strata, as well as most of the Neogene foredeep, remained in their autochthonous position. Although integral parts of the Carpathian depositional assembly, these autochthonous strata are dealt with in this section on the European foreland, whereas all the deformed and thrusted Jurassic to Neogene sequences, including the undeformed Vienna basin, are discussed in the section on the Outer Carpathian thrust belt.

The Jurassic Rifting and Passive Margins

With the exception of a few smaller exposures near Brno, the autochthonous Jurassic strata on the territory of southern Moravia and northeastern Austria are buried below the Neogene foredeep and the Carpathian thrust belt (Figures 3, 7). Numerous wells, both in Moravia and Austria, indicate that the autochthonous Jurassic strata are confined to a relatively narrow zone of the Carpathian foreland between Brno and the Danube Valley. Picha (1979a) interpreted this zone, roughly identical with the extent of the Vienna basin, as a remnant of a larger northwest–southeast-trending rift-related depression, which originally extended farther into the Bohemian Massif. We call it the Dyje–Thaya depression (Figure 5A). On its northern side, this depression is bounded by a fault that parallels the axis of the Nesvacilka graben and the superimposed Nesvacilka Paleogene valley and submarine canyon and, on its southern side, by faults defined by, e.g., Zimmer and Wessely (1996) on the northeastern side of the Bohemian Massif spur south of the Danube Valley (Figure 5). These northwest–southeast-trending faults, which apparently formed or were activated during the Jurassic extensional phase, seem to continue further into the Bohemian Massif or at least are on a trend with some other significant northwest–southeast faults in the massif. As indicated by the erosional remnants of Jurassic rocks in the vicinity of Brno (Figure 7) and along the Luzice (Lusatian) fault in northern Bohemia, the Jurassic strata may have originally extended across the Bohemian Massif into northern Europe (Picha, 1979a). The continuation of the Jurassic depression toward the southeast underneath the deeper zones of the Carpathian thrust belt is not known, but the palinspastic reconstruction of the Outer Carpathian depositional system (Figure 5A) indicates that the Jurassic strata of the Dyje–Thaya depression most likely represent an extension of the Silesian–Magura–Pieniny depositional system into the foreland.

The distribution of carbonate platforms and basins along the northwestern margin of the depression (Figure 5A, B) and orientation of rift-related fault blocks (e.g., Wessely, 1988) indicate an existence of southwest–northeast-oriented structural trends in the presumably northwest–southeast-running Dyje–Thaya depression. Such a discrepancy may be explained by an existence of a transtensional rather than an orthogonal tectonic regime during the rifting phase and opening of the depression. The southwest–northeast lateral motion in the Western Carpathian transfer zone and the Late Cretaceous–early Paleocene uplifting and inversion of the Carpathian foreland might have further modified the original structural grain. The resulting distribution and orientation of Jurassic faults and rotated fault blocks in the Dyje–Thaya depression might then be complex. The detailed mapping of the structural pattern of the Jurassic strata based on three-dimensional (3-D) seismic data may be of great importance for any future exploration of deeper subthrust structures within the depression.

The Dyje–Thaya depression and the Polish Trough (Dadlez, et al., 1995) are the two areas in the foreland of the Western Carpathians where significant Jurassic rift-related faults have been positively identified and mapped. In the Alpine Molasse foredeep southwest of the Dyje–Thaya depression and west of the Bohemian Massif spur, most of the faults affecting Jurassic strata in the subthrust plate (e.g., Wagner, 1996; Zimmer and Wessely, 1996) apparently formed by the flexural downbending of the European foreland during the formation of the Neogene Molasse Basin rather than by the Mesozoic rifting and extension. North of the Nesvacilka paleovalley, in central and northern Moravia, and the adjacent part of Poland, the Jurassic strata are not known. They were either not deposited in this area or later removed by erosion. Only in the deep zone adjacent to the Pieniny Klippen Belt, some normal faults visible on seismic lines might be considered as potential rift-related structures. In addition, the Miocene faults of the Carpathian foreland in western Poland, according to Krzywiec (2001), only seldom show any evidence of being reactivated from preexisting Jurassic rift-related structures. However, in the area adjacent to the Polish trough, large normal faults, dissecting both Paleozoic basement and the Miocene foredeep infill, have been interpreted as Mesozoic rift-related structures reactivated during the Miocene (Krzywiec, 2001). In the Polish trough, the Oxfordian–Kimmeridgian event represents the second phase of rifting, which followed the previous Late Permian–Triassic phase of crustal and lithospheric thinning associated possibly with extensive regional wrenching (Dadlez et al., 1995). The existence of Jurassic rifting in the Dyje–Thaya depression and the Polish trough and its apparent absence in the area between these two features and in the eastern Alpine foreland strengthen the assumption that during the Jurassic, the northwest–southeast-trending rift-related faulting and sedimentation might have prevailed in the Western Carpathian region (Figure 5A).

The autochthonous Jurassic strata of the Carpathian foreland in southern Moravia were studied by Elias (1981), Adamek (1986, 2002), and Elias and Wessely (1990), among others. The Jurassic sequence begins with synrift terrestrial and deltaic deposits more traditionally known as the Gresten Formation but also named the Divaky Formation by M. Elias (1971, personal communication) and assigned to the Doggerian age (Figure 11). Their thickness ranges from tens of meters on the uplifted edges of rotated fault blocks to more than 1000 m (3300 ft) on the subsided sides of the fault blocks and in the local grabens. So far, the very thick synrift clastic deposits are known only from the territory of Austria, where they are subdivided into four members: the lower and upper quartzarenite series and lower and upper shaly horizons, representing the alternating continental and deltaic sequences, respectively (Elias and Wessely, 1990). The basal transgressive member contains intercalations of coaly shales and coals. The sandy facies of the Gresten Formation proved to be a good reservoir in the Damborice oil field and the Uhrice gas field.

In the Callovian, the terrestrial and deltaic synrift sedimentation of the Gresten Formation was followed by a further marine incursion and gradual development of the predominantly carbonate depositional environment of the passive continental margins. The transition was marked by deposition of 70–250-m (230–820-ft)-thick dolomitic sandstones known as the Nikolcice Member in Moravia and the Hoflein beds in Austria. In the early Oxfordian, two main facies, (1) the marginal, high-energy carbonate platform facies, and (2) the deeper basinal anoxic facies, evolved at the territory of southern Moravia and northeastern Austria (Stranik, et al., 1968; Brix, et al., 1977; Ladwein, 1988; Elias and Wessely, 1990) (Figure 11).

The marginal carbonate facies, including the erosional remnants of Jurassic strata on the exposed margins of the Bohemian basin, formed in a shallow sea along the western side of the Jurassic basin. At the surface localities near Brno, e.g., Stranska skala, Hady, Nova hora, and Svedske sance, the Jurassic limestones rest directly on the granitic rocks of the Brno massif or the Devonian limestones. These basal sandy and brecciated limestones pass into the bedded micritic, cherty, and bioclastic (crinoidal) limestones with glauconite and phosphates. The rich, shallow-marine fauna, studied by Oppenheimer (1907, 1926, 1932), among others, would assign these strata to the Callovian and Oxfordian. The known thickness of these Jurassic deposits near Brno is about 50 m (160 ft). In the area below the Carpathian foredeep and the frontal zone of the thrust belt, the carbonate facies is represented by the Altenmarkt Group, in Moravia formerly known as the Hrusovany Limestones, Novosedly Limestones and Dolomites, and Pasohlavky Limestones. This platform sequence, about 450-m (1476-ft)-thick, begins with bedded limestones and dolomites, an equivalent of the Vranovice Limestones of the basinal facies, which grade upward into bioclastic limestones and finally into algal and coral patch reefs alternating laterally with oolitic and bioclastic limestones. Basinward, the carbonate facies is fringed by a transitional facies of marly limestones distinguished in Austria as the Falkenstein Formation.

In the basinal facies, the full carbonate sedimentation began with the deposition of the fine-grained, cavernous, and partly silicified calcareous dolomites and limestones called the Vranovice Limestones (formerly the lower carbonate series in Austria). The thickness of

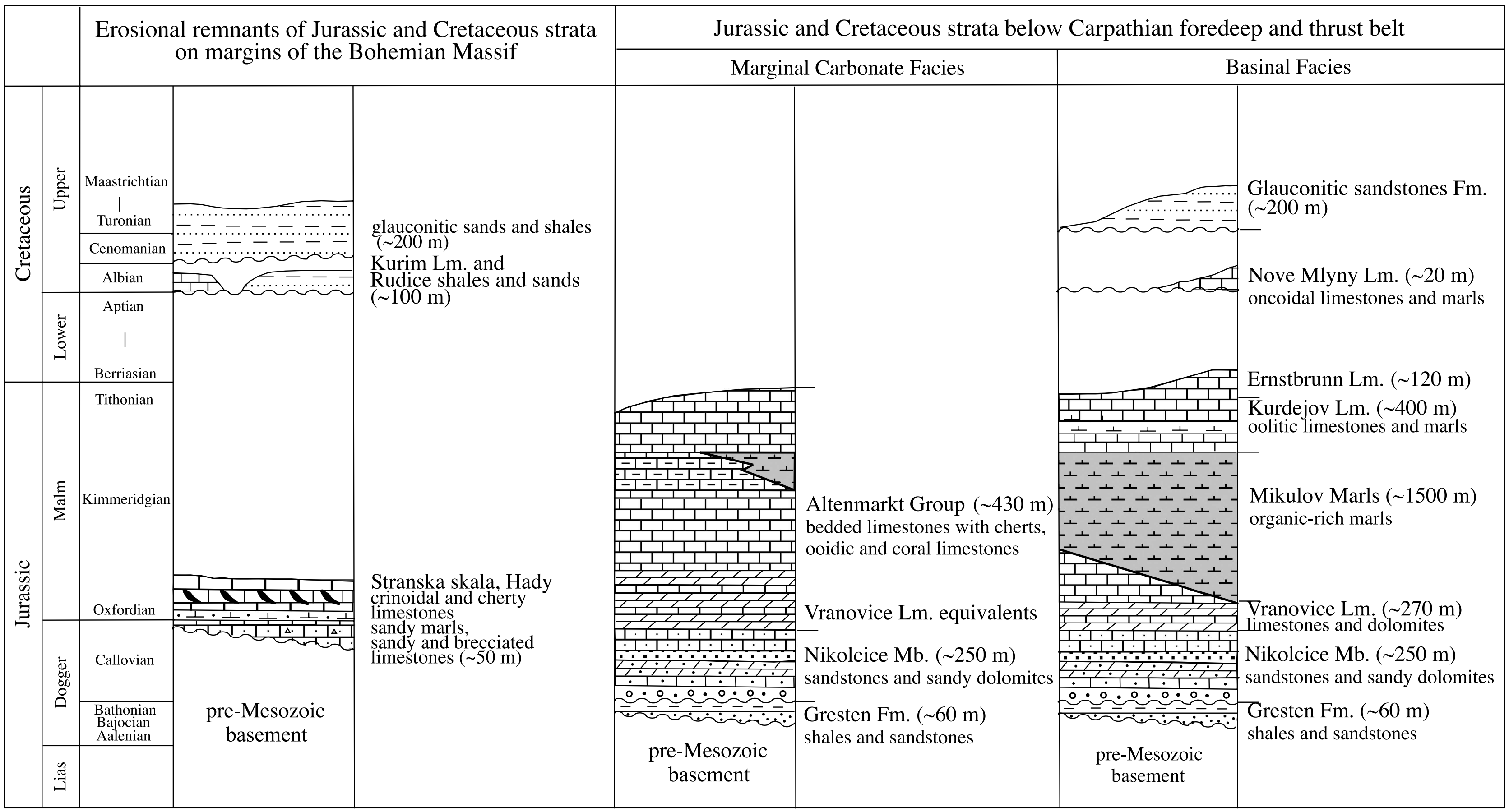


Figure 11. Stratigraphy of the autochthonous Mesozoic strata of the European foreland plate predominantly concealed below the Neogene foredeep and the Western Carpathian thrust belt in Moravia.

these strata, assigned to Oxfordian, decreases eastward toward the deeper parts of the basin. As documented by the newly discovered Zarosice oil field in central Moravia, the dolomitic rocks of the Vranovice Limestones have good reservoir properties.

The Vranovice Limestones pass gradually into the full basinal facies represented by the monotonous sequence of dark, organic-rich marls with subordinate lenses and intercalations of organodetritic limestones. They are called Mikulov Marls and correlated with the Klentnice Formation of the Outer Klippen. As established in numerous wells, the thickness of these strata increases basinward to as much as 1500 m (4950 ft). The Malmian Mikulov Marls represent a world-class source rock, which sourced most of the oils found elsewhere in northeastern Austria and southern Moravia (Ladwein, 1988; Francu et al., 1996, Picha and Peters, 1998).

Upward, the Mikulov Marls gradually pass into an approximately 400-m (1300-ft)-thick sequence of dark organodetritic limestones, dolomites, and minor marls called the Kurdejov Limestones. They are followed by the upper Tithonian, organodetritic, partly dolomitized limestones with rich mollusk, algal, and coral fauna. In wells, Wessely (in Brix et al., 1977) and Adamek (1986) named these strata the Ernstbrunn Limestones, as an equivalent of the well-known allochthonous Ernstbrunn Limestones of the Outer Klippen in northeastern Austria and southern Moravia (Figure 3).

Similarly, the autochthonous Mikulov Marls show many similarities with the coeval allochthonous Klentnice Formation of the Outer Klippen. Whether these two sets of strata, autochthonous and allochthonous, originated in a single depositional system, from which the allochthonous Klentnice Formation and the Ernstbrunn Limestones were tectonically detached, or whether the allochthonous strata were tectonically transported from a more distal carbonate environment remains uncertain. Apparently, other more internal, carbonate platforms and basins formed on the rifted and fragmented continental margins of Europe (Picha, 1996), from which, during the Alpine orogeny, the tectonic klippen might have been detached, integrated into the Outer Carpathian belt, and tectonically transported for a considerable distance.

The Cretaceous Period

The deposition of the Ernstbrunn Limestones in the Tithonian was followed by a major regression and hiatus, which, in the area of southern Moravia and northeastern Austria, lasted through most of the Lower Cretaceous. A new limited marine incursion, documented by rare finds of Aptian to Albian limestones at Kurim and sands and shales at Rudice both north of Brno, occurred toward the end of the Lower Cretaceous (Figure 11). The relicts of the Aptian to Albian limestone breccias near Kurim, described by Krystek and Samuel (1979), are found in open fractures of granitic rocks of the Brno massif. The locality is situated in a northwest–southeast-trending fault zone, which delineates the northeastern margin of the Dyje–Thaya depression. The upper Albian bioclastic and micritic limestones with algae and oysters have been also reported by Rehanek (1984) from the deep well Nove Mlyny-1 (Nove Mlyny Limestones by Adamek, 1986). No equivalents of these Lower Cretaceous rocks have been found in the tectonic klippen of the Pavlov and Ernsbrunn Hills in southern Moravia and northeastern Austria. Most of the European foreland in Moravia apparently remained exposed during the Early Cretaceous.

The uplifting of the foreland during the Early Cretaceous might have been associated with the fundamental reorganization of the Outer Carpathian realm from one dominated by the northwest–southeast-trending rift pattern of the Middle to Late Jurassic to another associated with the activation of the northeast–southwest-trending Western Carpathian transformation zone. On the Carpathian side of this zone, the dextral motion along the sheer zone led to further rifting, extension, and formation of a system of grabens and ridges (Figure 5), whereas the area west of the transfer zone in southern Moravia and northeastern Austria was uplifted and remained relatively stable throughout the rest of the Mesozoic history.

In the Cenomanian, a major global transgression engulfed most of the European platform, including its marginal parts adjacent to the Tethyan realm. The Upper Cretaceous strata are known from the Dyje–Thaya depression in southern Moravia and northeastern Austria and from the Miechow depression in Poland (Figures 5C, 7). In Moravia and Austria, they have been encountered in numerous wells below the Waschberg–Zdanice nappe. There, the Upper Cretaceous strata rest transgressively on the karstified Jurassic carbonates. The basal glauconitic sandstones pass upward into the calcareous claystones and siltstones with beds of sandy limestones and sandstones. Typically, these strata are more than 200 m (660 ft) thick; the highest known thickness of 517 m (1696 ft) was recorded in the well Ameis-1 in Austria. According to Stranik et al. (1996), the age of these deposits in Moravia extends from the Turonian to the early Campanian. In Austria, the Upper Cretaceous strata are known as the Klement Supergroup, which is subdivided into the Ameis (Cenomanian–Santonian) and the Poysdorf (Campanian–Maastrichtian) formations (Fuchs and Wessely, 1977, 1996). Both lithologically and stratigraphically, these autochthonous strata may be compared with the Klement and Palava formations (Stranik et al., 1996) found on the top of the Tithonian Ernstbrunn Limestones in the tectonic klippen of the Pavlov Hills. As indicated by faunas, these Upper Cretaceous strata of southern Moravia and northeastern Austria represent a transitional facies between the epicontinental deposits of the boreal sea of northern Europe, such as the Brezno Formation of the Bohemian Massif, and the coeval deposits of the Tethyan continental margins. Their deposition seems to be related more to the global Cenomanian transgression rather than to a progression of the Tethyan Sea onto the European foreland. Nevertheless, the distribution of these strata in the Dyje–Thaya depression in southern Moravia and northeastern Austria and in the Miechow trough in Poland indicates that the deposition and/or preservation of these strata was at least partly controlled by the renewed subsidence in these depressions.

The Autochthonous Paleogene and Paleovalleys

The autochthonous Paleogene in southern Moravia is represented predominantly by the depositional fill of two large paleovalleys buried below the Neogene foredeep and the Carpathian thrust belt (Figures 7, 12). These features, also called Nesvacilka and Vranovice depressions or canyons, were originally interpreted as tectonic grabens (Homola et al., 1961; Adam et al., 1968; Nemec, 1973), with some erosional component in their formation (Dlabac and Mencik, 1964). Picha (1974, 1979a) and Picha et al. (1978) fully recognized their erosional origin and interpreted them as paleovalleys and submarine canyons. In addition to these two main valleys, several smaller side tributaries, mainly located on the northeastern side of the Nesvacilka canyon, have been mapped (Jiricek, 1987, 1994). Indications exist that a similar paleovalley, named the Tulln canyon by Jiricek (1994), may exist at the Austrian territory west of Vienna (Figure 22, shown on page 128). The recognition of the Paleogene depressions as erosional features has been obscured by their location in a broader northwest–southeast-trending Dyje–Thaya Jurassic tectonic depression (Picha et al., 1978) (see previous section). In particular, the Nesvacilka paleovalley is located in the confines of an asymmetrical northwest–southeast-trending Jurassic rift-related Nesvacilka half graben (Picha 1996), bounded on its southwestern side by a steep fault, which apparently marks the northeastern margin of the Dyje–Thaya depression (Figure 13). In the literature, the relationships of these two genetically and morphologically different features are commonly confused, and the term Nesvacilka graben has been alternatively used for both the rift-related Jurassic graben and the Paleogene valley. So far, no indication is found that a similar rift-related graben does exist below the Vranovice paleovalley.

The tectonically controlled valleys are interpreted (Picha, 1979a, 1996) as being excavated by rivers during the Late Cretaceous to early Paleogene Laramide uplifting of the Carpathian foreland and later inundated by the sea and converted into estuaries and submarine canyons through which clastic material was transported into the foreland basin. Like some of the modern river-submarine channel systems, e.g., the Amazon and Niger, the Paleogene valleys in Moravia evolved preferentially in the preexisting depression, and their course was affected by an active fault system.

The Paleogene valleys are as much as 12 km (7 mi) wide and more than 1500 m (4900 ft) deep. By drilling and seismic data, these paleovalleys have been traced for more than 40 km (24 mi). In their size, the paleovalleys in southern Moravia are comparable with the Grand Canyon of Arizona or the Hudson submarine canyon of the East Coast of North America. They are cut into Mesozoic and Paleozoic carbonate and clastic strata and the Precambrian crystalline basement rocks and are filled with Paleogene clastic deposits (Figures 12, 13). The excavation of these gigantic valleys testifies to a major uplifting of the Carpathian foreland and sea level changes during the Late Cretaceous and early Paleogene.

The sedimentary fill of the canyons, a few hundred meters to more than 1500 m (4900 ft) thick, is made predominantly of organic-rich (1–9% total organic carbon [TOC]), laminated mudstones and siltstones with subordinate channelized sands, proximal and distal turbidites, and debris flows. The coarser clastics, sandstones, and conglomerates are distributed mainly in the lower axial parts of the Nesvacilka valley, where they form a discontinuous turbiditic sequence as much as 300 m (1000 ft) thick. They were deposited during the early active phase, dominated by erosion and transport of clastics into the foreland basin (Figure 14). The overlying mudstones then represent the later phase of abandonment and hemipelagic drape sedimentation. So far, no significant coarser deposits have been found in the less-drilled Vranovice valley. The common presence of slump folds, pebbly mudstones, and chaotic slump bodies indicates that the mass movement was a significant factor in sediment transport inside the canyons. The uppermost parts of the canyon fill are commonly deformed and possibly detached and tectonically transported along the base of the allochthonous thrust sheets (Picha, 1979a).

Because of a complex and discontinuous depositional pattern, Picha (1979a, b) assigned the whole fill, including its tectonized upper part, to one lithostratigraphic unit and called it the Nesvacilka Formation after the Nesvacilka-1, 2, and 3 wells, in which the

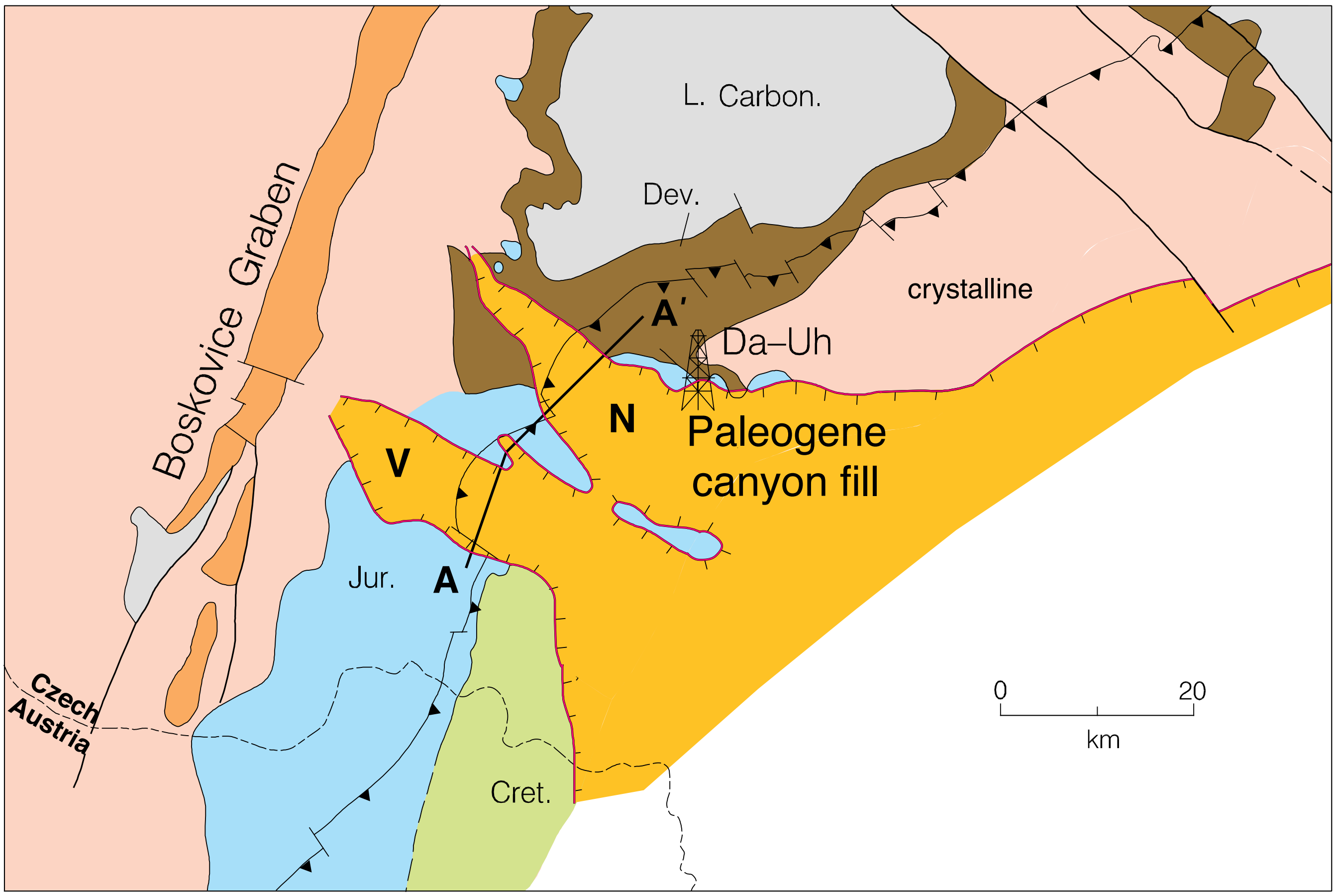


Figure 12. Pre-Neogene subcrop map showing the Nesvacilka (N) and Vranovice (V) paleovalleys cut into the European foreland plate, filled with Paleogene deposits, and later buried below the edges of the Western Carpathian thrust belt and the Neogene foredeep. Da-Uh marks the location of the Damborice and Uhrice oil and gas fields, respectively (Picha, 1996). Cross section of AA′ shown in Figure 13.

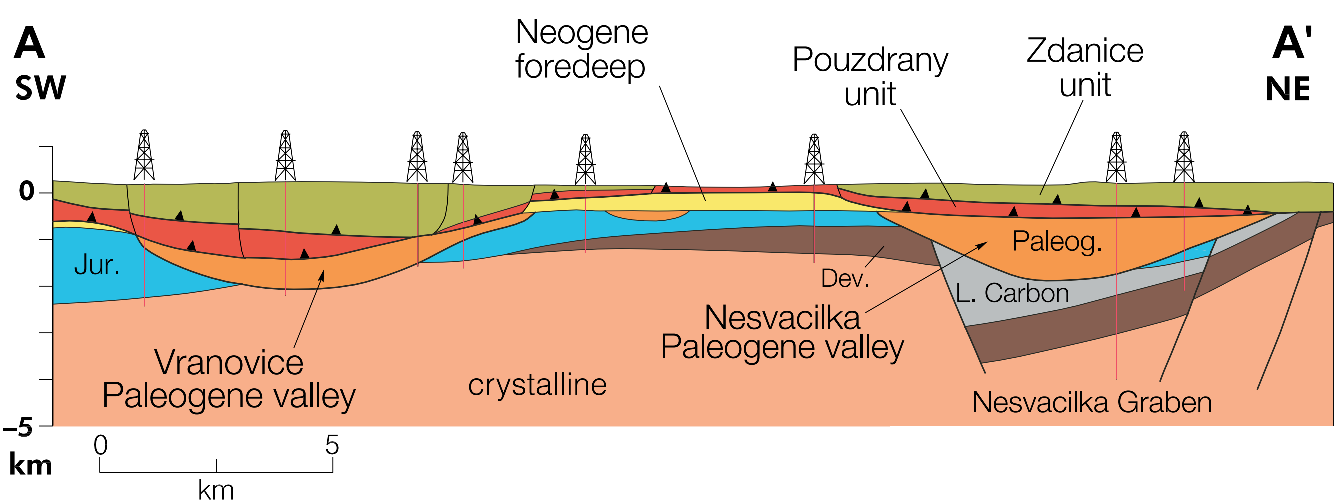


Figure 13. Cross section through the Vranovice and Nesvacilka Paleogene valleys-submarine canyons along the line AA′ (location in Figure 12). The erosional features are cut into the Mesozoic and Paleozoic strata and the Precambrian crystalline basement rocks of the Brunovistulicum. Note the position of the Nesvacilka paleovalley in the Jurassic rift-related Nesvacilka graben.

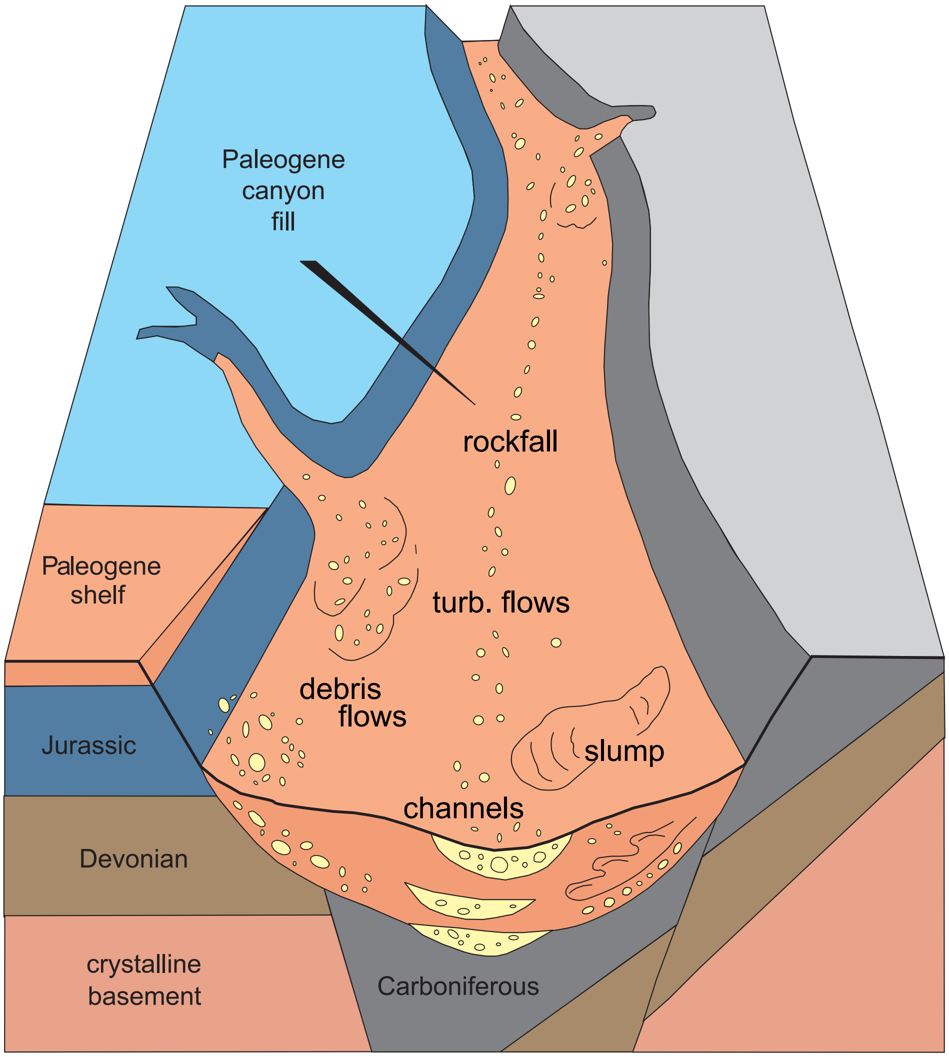


Figure 14. Schematic reconstruction of the deep-water depositional system of the Nesvacilka paleovalley and submarine canyon located in the confines of the Nesvacilka graben. The dynamic depositional regime of the early stages of the development of the submarine canyon was characterized by the activity of gravity-driven currents (turbidites and debris flows) and widespread slumping. The best reservoirs are found among the channelized sands. Modified after Picha (1979a).

autochthonous Paleogene was originally defined (Holzknecht in Homola et al., 1961). Later, Jiricek (1994) and Rehanek (1994) introduced a new name, the Damborice Group, for the whole autochthonous valley fill and subdivided it into two formations, keeping the term Nesvacilka Formation for the hemipelagic upper part of the sequence and using the name Tesany Formation for the turbidity-dominated lower part of the sequence.

The age determination of the paleovalleys fill has become a subject of discussions as well. Based on paleontological evidence, Pokorny (in Dlabac, 1946), Grill (1947), Holzknecht (in Homola et al., 1961), and Cicha and Hozknecht (1964) dated the autochthonous Paleogene deposits as late Eocene to early Oligocene. Hanzlikova (in Picha et al., 1978) confirmed the late Eocene to early Oligocene age but also pointed to a complex character of microfaunal assemblages that, in addition to highly diversified indigenous species, include faunas redeposited from the older, Jurassic, Cretaceous, and early Paleogene strata. Hamrsmid and Krhovsky (1987), Holzknecht and Krhovsky (1987), Hamrsmid et al. (1990), and Krhovsky et al. (1990) disregarded the possibility of redeposition of all the Late Cretaceous and early Paleogene microfaunas and extended the stratigraphic range of the Nesvacilka Valley fill from the Late Cretaceous (Campanian to Maastrichtian) to early Paleogene (Paleocene to early Eocene). Later, Jiricek (1994) reassigned the age of the paleovalley fill to the early Paleocene to early Oligocene.

From the regional viewpoint, the late Paleogene age of the paleovalley fill seems to be more logical. The marine incursion and depositional filling of paleovalleys could then be associated with the late Eocene transgression, which initiated the development of the Molasse basin in the Alpine foreland. Such an interpretation also enables better correlation of the autochthonous Paleogene with the allochthonous Pouzdrany unit, as well as with the lower marine Molasse of the Alps in Austria and Bavaria (Picha and Stranik, 1999). Picha (1979b) loosely correlated the autochthonous

Paleogene with the allochthonous late Eocene to early Oligocene Pouzdrany Formation of the Pouzdrany unit and with the coeval part of the Nemcice (Submenilitic) Formation and the early Oligocene Menilitic Formation of the Zdanice unit (Figure 17A, shown on page 89). All these formations belong to the late Eocene to early Oligocene depositional sequence of the Carpathian foreland basin, the autochthonous Nesvacilka Formation with the highest proportion of coarser clastics being associated with the estuarine to submarine canyons environment and the coeval deposits in the Pouzdrany and Zdanice units being deposited in the open shelf and slope to basin environments, respectively (Figure 5D).

Questionable with respect to the known stratigraphy of the valley fill is the position of the Malesovice Formation, recognized in one well at the northwestern end of the Vranovice paleovalley (Ctyroky, 1987). This formation, made by only a few meters of brownish gray marine shales with Egerian fauna, most likely represents the youngest members of the paleovalley fill, which may be correlated with the upper part of the Uhercice Formation of the allochthonous Pouzdrany unit (Figure 17A, shown on page 89). Another less likely solution would be to interpret these strata as the oldest members of the Neogene foredeep. However, because the evolution of the Neogene foredeep in that particular region began later during the Eggenburgian (Burdigalian) transgression, such an explanation would not be consistent with the overall regional stratigraphic pattern.

The rivers, which cut these paleovalleys as well as the successor submarine canyons, must have funneled large amounts of clastic material into the foreland basin. According to Picha (1969), the Late Cretaceous to the late Eocene kaolinite-rich pelitic deposits of the external Zdanice–Subsilesian and Silesian units were, to a great extent, supplied from the deeply weathered foreland of the Bohemian Massif.

Large paleovalleys buried below the Neogene foredeeps and the Carpathian thrust belt are not unique to southern Moravia. They have been identified in northern Moravia (Detmarovice and Bludovice valleys) and elsewhere in eastern Poland. However, unlike the Nesvacilka and Vranovice canyons, these valleys are filled with the middle Miocene (Badenian) deposits. Furthermore, the large Hystria and Kamchia depressions in the Black Sea shelf of Romania and Bulgaria, respectively, were apparently at least partly shaped by erosion. They are filled with 2.5–3.0 km (1.5–1.8 mi) of Eocene and Oligocene deposits.

The sedimentary fill of the paleovalleys and adjacent shelf deposits represent the northernmost marginal facies of the Carpathian foreland basin in the Paleogene. As indicated by the presence of coeval organic-rich marginal facies in the Alpine Molasse and in the Black Sea region, these deposits may have extended along the entire length of the Alpine–Carpathian belt (Picha, 1996). In the Carpathian region, with the exception of southern Moravia, these marginal Paleogene strata would be buried below the thin-skinned thrust belt. Toward the south, this autochthonous marginal Paleogene facies may extend to the present Pieniny Klippen Belt (e.g.,Tomek, 1976, 1979). In the Polish Carpathians, limited occurrences of assumed autochthonous Paleogene deposits have been described by Moryc (1995) from wells near Rzeszow and by Oszczypko (1997) and Oszczypko et al. (2000) from the Zawoja-1 well in the Orava region. Oligocene to early Miocene (NN1 zone) conglomerates and mudstones have been reported from the eastern slope of the Pribor–Tesin Ridge (Andrychow area) by Oszczypko and Oszczypko-Clowes (2003). At least some of these marginal Paleogene facies, including the paleovalley fills, would represent a shallow-marine lateral equivalent to the deep-water Menilitic (Dysodile) shales, which are believed to be the main source rocks in the Carpathians (ten Haven et al., 1993; Lafargue et al., 1994; Dicea, 1995). However, unlike the more distal Menilitic shales, which are detached and integrated into the Carpathian thrust belt, the marginal Paleogene facies may still be found in the autochthonous or parautochthonous positions below the allochthonous belt.

After being buried below the thrust belt in the early Miocene, these organic-rich Paleogene strata may have generated significant quantities of oil and natural gas, which would have migrated into any structural or stratigraphic trap, which, at the time of the migration, existed both in the Carpathian thrust belt and the underlying European foreland. The assumed presence of paleovalleys and organic-rich Paleogene deposits underneath the Carpathian thrust belt may provide a critical ingredient to a variety of potential hydrocarbon plays and, thus, greatly widen the prospectivity of the entire Carpathian system.

The Carpathian Neogene Foredeep

The Carpathian Neogene foredeep in Moravia is a part of Paratethys as originally defined by Laskerev (1924) and later further elaborated by Cicha et al. (1975a, b), Rogl and Steininger (1983), and Rogl (1998), among others, as an autonomous, depositional system separated from the Mediterranean realm by the rising Alpine–Carpathian mountain chain. However, episodic communication between these two areas did exist, especially in the middle Miocene. At this time, through the Rhine Graben, the Paratethys was also connected with the North Sea. Geodynamics and paleoclimatic changes of the Central Paratethys were discussed by Cicha and Kovac (1990), among others.

With respect to its somewhat isolated position, an autonomous regional stratigraphic nomenclature, distinct from the general Mediterranean classification, has been established for the Central Paratethyan realm (e.g., Cicha et al., 1975a, b; Steininger et al., 1988).

The evolution of foreland basins and foredeeps has been, to a great extent, controlled by the preexisting structural pattern of the foreland, which varied significantly along the length of the Alpine–Carpathian thrust belt. Typically, prominent depocenters marked by a high rate of subsidence alternate with areas of limited subsidence and deposition. In the Alpine–Carpathian region, the Bavarian–Austrian Molasse with more than 4 km (2.4 mi) of Neogene deposits, the Boryslav area in Western Ukraine with up to 7 km (4.2 mi) of Neogene deposits, and, the deepest of all, the Focsany depression in Romania with 8–10 km (5–6 mi) of Neogene deposits (Sandulescu, 1984) represent prominent depocenters. However, the segments of the foredeep, which evolved on more stable crustal blocks of the foreland, such as the Bohemian Massif on the territory of Moravia, are characterized by a low rate of subsidence and sedimentation.

As, in time, the collision along the Carpathian arc moved progressively eastward, so did the rate of sedimentation in the foredeep (Meulenkamp et al., 1996; Oszczypko, 1998). In Austria and southern Moravia, the highest rate of subsidence occurred in the Eggenburgian to Karpatian, in northern Moravia and western Poland in the Badenian, and in Romania in the Dacian and Romanian.

In the Carpathian terminology, the term foredeep is used for all principally undeformed late orogenic and postorogenic, shallow-marine and nonmarine Neogene deposits found in front, underneath, and marginally on the top of the frontal zone of the Alpine–Carpathian thrust belt and also in tectonic slices incorporated into the thrust belt. The Neogene foredeep in Moravia displays a broad range of depositional environments, including the typical depocenters adjacent to the front of the thrust belt as well as the distal facies deposited on the European foreland. For such an assembly of depositional settings, the term foreland basin would be more proper, but for the sake of nomenclatorial unity, we continue to use the traditional term Neogene foredeep for all the undeformed late orogenic and postorogenic sequences laid down on the Carpathian foreland. The northwestern landward margin of the Neogene foredeep in Moravia is delineated by outcrops of the Bohemian Massif (Figure 3). The southeastern margin is outlined on the surface by the front of the Carpathian thrust belt, but the undeformed strata of the Neogene foredeep continue farther below the thrust belt. At least 30 km (18 mi) of overthrusting over the undeformed strata of the Neogene foredeep has been verified by deep drilling (Stranik et al., 1993; Elias and Palensky, 1997; Oszczypko et al., 2006).

As established by numerous wells, the overall thickness of the Neogene foredeep on the territory of Moravia ranges in several hundred meters. Only locally, e.g., at the edge of the thrust belt near the Austrian border, in the Bludovice and Detmarovice depressions and paleovalleys in the Ostrava region, and possibly in the Roznov area underneath the Carpathian nappes (Palensky et al., 1995), the thickness of the foredeep strata exceeds 1000 m (3300 ft). The deepest zone of the foredeep is typically situated along the present erosional edge of the thrust belt, and from there, the thickness of Neogene strata generally decreases toward both the foreland and the hinterland underneath the Carpathian nappes.

The stratigraphic range of the undeformed Carpathian Foredeep in Moravia extends from the Eggenburgian (Burdigalian) to Badenian (Langhian and early Serravalian) (Jiricek and Seifert, 1990). In addition, the latest Miocene and Pliocene strata deposited in the Upper Moravian Depression are considered to be a part of the foredeep. In a broad sense, all these deposits can be correlated with the upper marine and upper nonmarine Alpine Molasse of Switzerland, Bavaria, and Austria (Wagner, 1996, 1998; Burkhard and Sommaruga, 1998). They also can be correlated with similar deposits of the Neogene foredeep on the territory of Poland (Oszczypko, 1998; Oszczypko et al., 2006).

The stratigraphic subdivision and correlation of the Neogene strata is based primarily on biostratigraphy. In terms of the Paratethyan stratigraphy, three major depositional units(?), (1) the Eggenburgian–Ottnangian, (2) the Karpatian, and (3) the Badenian, might be distinguished (Figure 15). Each of these units is characterized by a specific depositional regime, apparently dependent on the progression of the Carpathian thrusting and global sea level changes, as well as on the preexisting structural pattern.

The lithostratigraphy of the Carpathian foredeep in Moravia has not been fully developed. The fragmentary nomenclature we present in our chapter, including the chart in Figure 15, may undergo significant changes in the future.

The Eggenburgian–Ottnangian (Burdigalian)

The Eggenburgian to Ottnangian strata in Moravia are distributed mainly in the southern part of the Carpathian Foredeep between the Austrian border and Brno. They are also found on top of the Zdanice nappe

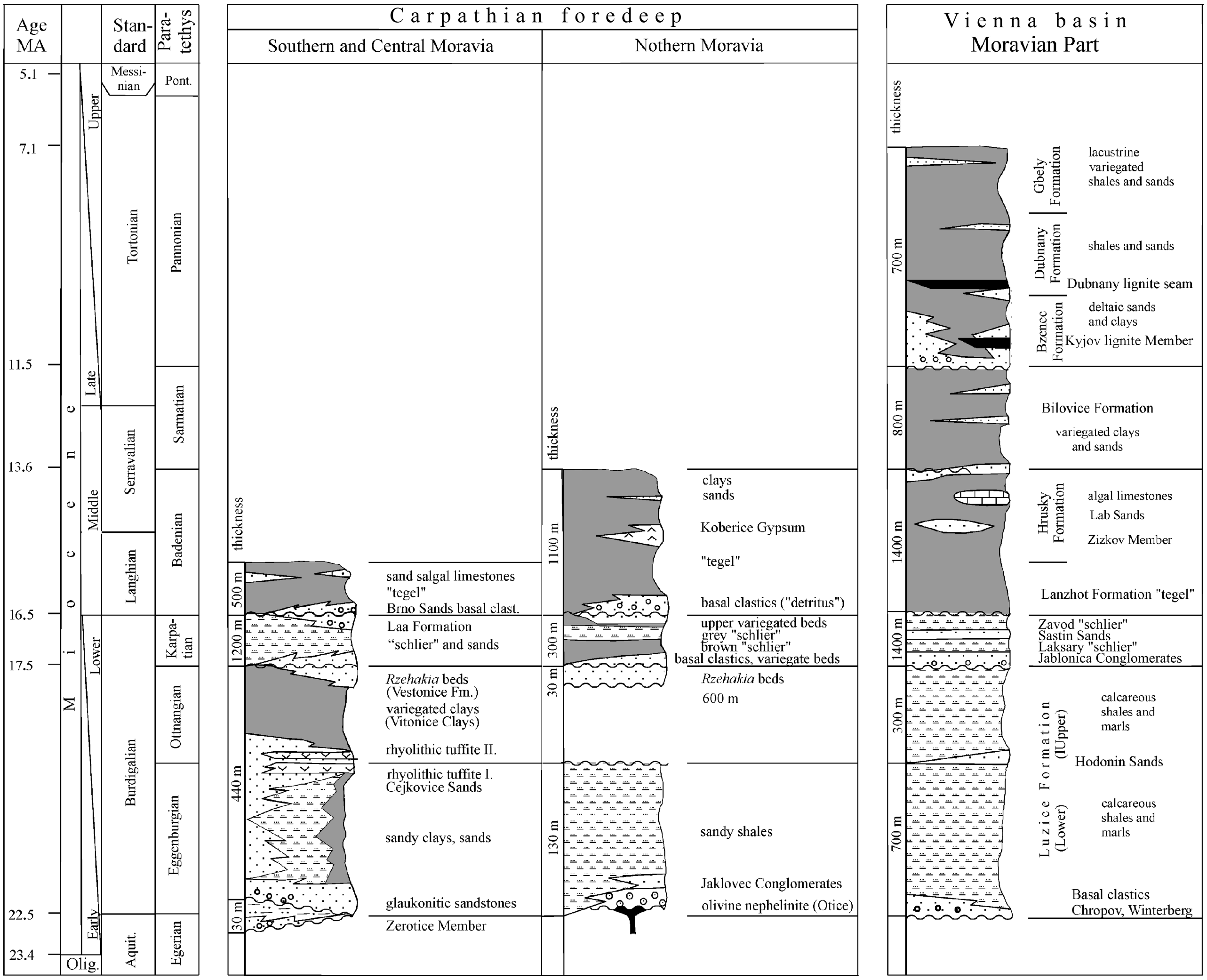


Figure 15. Stratigraphy of the Neogene foredeep and the Vienna basin in Moravia.

(Sakvice Marls and Pavlovice Formation) and in the Vienna basin (Luzice Formation). The distribution of the Eggenburgian and Ottnangian strata in southern Moravia was, to a great degree, controlled by the preexisting northwest–southeast-trending fault system of the Jurassic Dyje–Thaya depression, which underlies the Carpathian foredeep and the thrust belt. In northern Moravia, the Eggenburgian fluvial sands and overlying marine shales are known only from the Detmarovice depression and paleovalley and from a few small erosional remnants (Jurkova et al., 1983).

The Eggenburgian–Ottnangian strata are exposed along the western margin of the foredeep. They rest transgressively on kaolinized crystalline rocks of the Bohemian Massif marked by a rugged erosional relief with ridges and canyons. In the eastern zone adjacent to the thrust belt, the Eggenburgian–Ottnangian strata are covered by younger Miocene deposits. Their extent and stratigraphy were established from well data, e.g., during the exploration of the Dolni Dunajovice gas field (Adamek, 1979).

Two main facies, a marginal, shallow-marine and a deeper basinal facies, have been distinguished in the Eggenburgian deposits of southern Moravia (Ctyroky 1982, 1991, 1993). A great lithological variability characterizes the shallow marginal facies; the basal coarse conglomerates and kaolinitic sands are followed by silts, sands, shales, and occasional coaly shales. At the western margin of the basin near Znojmo, the sporadically variegated proluvial clastic deposits were described and named the Zerotice Member by L. Prachar (1970, personal communication). P. Ctyroky and J. Adamek (1988, personal communication) assigned these strata to the Egerian(?) to Eggenburgian (Figure 15). In the basinal facies, the basal transgressive conglomerates and glauconitic sands are overlain by alternating sands and shales. The Ottnangian deposits, otherwise similar to those of the underlying Eggenburgian strata, were laid down in a shallower, more restricted depositional environment (Jiricek, 1983) marked by the presence of dark shales with remnants of fish. Characteristic for the Ottnangian strata is the presence of the *Rzehakia* beds represented by sandstones and gravels with pebbles of Jurassic limestones and endemic mollusks (*Rzehakia socialis*). The *Rzehakia* beds, recently renamed by Adamek (2003) the Vestonice Formation, mark a transgressional-regressional event (Brzobohaty and Cicha, 1993), whose stratigraphic adherence to either Ottnangian or Karpatian, at least in the outer zone of the foredeep in Moravia, is not clear. In northern Moravia, the Ottnangian strata were recognized recently below the Carpathian nappes near Frydek–Mistek (P. Ctyroky, 1996, personal communication). The maximum known thickness of the Eggenburgian–Ottnangian strata of about 440 m (1400 ft) has been recorded in a zone adjacent to the front of the Carpathian thrust belt near the Austrian border.

Occasional layers of rhyolitic tuffs associated with the arc volcanism of the Carpatho-Pannonian province are found in the uppermost Eggenburgian and the lowermost Ottnangian strata in Moravia (Nehyba, 1997; Nehyba and Roetzel, 1999).

The Karpatian (Upper Burdigalian)

The chronostratigraphic stage Karpatian is used in terms of the original definition (Cicha et al., 1967), although the new lithostratigraphic division of strata deposited in the time span of the Karpatian stage in the Moravian sector of the foredeep is also considered (Brzobohaty et al., 2003).

In the Karpatian, the paleogeography of the foredeep changed significantly. The front of the deformation progressed far onto the foreland, and the most external units of the Outer Carpathian belt were thrusted over the inner zones of the Neogene foredeep. Related with the flexural downbending of the crust in front of the thrust belt, the foredeep took the shape of a continuous depositional wedge, which rimmed the front of the Western Carpathian thrust belt along its entire length (Cicha and Kovac, 1990). The highest rate of subsidence occurred in southern and central Moravia, where the thickness of the Karpatian deposits may exceed 1200 m (3900 ft). Whether all this thickness is depositional or partly enhanced by tectonic stacking remains uncertain.

The predominantly marine Karpatian deposits of southern Moravia are attributed to the Laa Formation defined by Kapounek et al. (1960) in the Molasse Basin of northeastern Austria. The lower part of the formation, represented by laminated shales and silts, known as ''schlier'' and recently named the Musov Member (Adamek et al., 2003), was deposited in a deeper neritic environment of the transgressive stage. The upper part of the Laa Formation, named the Novy Prerov Member (Adamek, 2003), is characterized by a higher proportion of sands and silts. It was laid down in a deeper neritic to bathyal environment during the maximum extent of the Karpatian transgression. With a distinct discordance, the Novy Prerov Member is overlain by a transgressive sequence, the Ivan Member (Adamek, 2003), believed to be the highest part of the Karpatian.

In northern Moravia, the Karpatian strata transgressed over a more diverse relief, marked by the existence of the large Bludovice and Detmarovice paleovalleys and intervening ridges. The basal sands of different age and composition pass upward into the variegated and brown silty shales deposited in lagoons and coastal lakes with variable salinity. They are followed by marine gray laminated silty shales (schlier) (e.g., Jurkova and Novotna, 1974; Palensky, 1988; Palensky et al., 1995), which correspond to the Stryszawa Formation (Slaczka, 1977) in Poland. For the latest information on the stratigraphy of the Karpatian stage of the Central Paratethys, we refer to a book by Brzobohaty et al., 2003.

Toward the end of the Karpatian, the foredeep shrank to a narrow and shallow basin stretched along the front of the thrust belt. Variegated shales with anhydrite and unsorted sands were deposited in northern Moravia, whereas shales, sands, and gravels with clasts of rocks brought in from the Carpathian belt were laid down in central Moravia.

In the uplifted southeastern margin of the Bohemian Massif, an extensive erosion of Miocene deposits led to the development of a rugged relief with deep northwest–southeast-trending valleys. An uninterrupted sedimentation between the Karpatian and Badenian continued only at the territory of Austria (Grund Member) and locally in southern Moravia (Cicha and Ctyroka, 1995).

The Badenian (Langhian–Serravalian)

The Badenian began with a local transgression originally limited to the deepest zone of the foredeep along the front of the Carpathian front. Only gradually, the lower Badenian transgression spread over the more distal zones of the foredeep and adjacent parts of the Bohemian Massif. After a brief retreat of the sea, the whole foreland submerged, and the Badenian sea spread over vast areas of the Bohemian Massif.

The Badenian sedimentation began with continental breccias followed by marine sands and gravels known under local names, e.g., Brno Sands in the surroundings of Brno or ''detritus'' in the Ostrava region

(Jurkova, 1971). Calcareous clays called ''tegel'' were deposited in deeper and farther offshore-located parts of the basin. Following the second transgression, mainly calcareous clays (tegels) were laid down in the deeper parts of the basin, whereas thin biostromes and bioherms of algal (Lithothamnion) and bryozoan limestones formed on elevated parts of the relief. Volcanic ashes from the Carpathian–Pannonian volcanic province fell repeatedly into the basin (Nehyba, 1997).

The rate of subsidence varied greatly in various parts of the foredeep and the foreland. In contrast with the Karpatian, the subsidence of the foredeep in the Badenian was more intense in northern Moravia, where as much as 1100 m (3600 ft) of the lower Badenian strata accumulated in the Bludovice and Detmarovice deep depressions and paleovalleys. At the same time, a chain of southwest–northeast-trending elevations, previously known as the Slavkov–Tesin Ridge (Dlabac and Mencik, 1964), kept rising. According to Krejci et al. (2002), these elevations have diverse origin and do not represent a forebulge.

In the Opava subbasin of northern Moravia, the lower Badenian is represented by basal clastics, followed by variegated clays with intercalations of lignite. Subsequent deposition of gray calcareous shales with rich foraminiferal fauna marks the maximum extent of the lower Badenian sea. At this time, basaltic volcanism was active in the area.

In southern Moravia, the sedimentation ended in the lower Badenian, whereas in the northern part of Moravia, the sedimentation continued until the end of the Badenian (Figure 15). There, the middle Badenian gray clays are overlain by evaporites (Koberice Gypsum) as much as 65 m (213 ft) thick (Cicha et al., 1985). The precipitation of evaporites marks the middle Badenian shallowing of the sea and the formation of restricted evaporitic conditions (saline crisis) that occurred elsewhere in the outer basins of the Central Paratethys (Oszczypko et al., 2006). The upper Badenian clays with abundant remnants of plants and with thin beds of limestones at the base represent the terminal marine sediments in the northern part of the Carpathian Foredeep in Moravia. Since then, only continental fluviatile and lacustrine deposits of different ages accumulated in local depressions, both on the Bohemian Massif (e.g., in the Boskovice Furrow) and in the Carpathian Foredeep (e.g., in the Upper Moravian Depression). Among them are also the gravels with tectites, e.g., at Dukovany.

The Late Miocene to Pliocene Upper Moravian Depression

In the late Miocene and in the Pliocene, a nonmarine basin, the Upper Moravian Depression, formed in the Carpathian foreland and the marginal zone of the thrust belt (Figure 3). The bounding faults of this depression trend in the northwest–southeast direction, perpendicular to the direction of the Carpathian belt. These faults, extending from the foreland of the Bohemian Massif across the Neogene foredeep into the marginal parts of the thrust belt, apparently follow the preexisting fault system reactivated during the postorogenic rebound of the foreland plate. The sedimentary fill of the Upper Moravian Depression consists of fluviatile and lacustrine deposits, which may be divided into the lower and upper parts (Ruzicka, 1989). The lower part of the fill is represented by as much as 100-m (330-ft)-thick reddish deposits, supplied from the deeply weathered surface. The equally thick upper part is composed of monotonous gray and green clays and sands with numerous layers of coaly clays and lignite. The subsidence in the Upper Moravian Depression continued into the Pleistocene.

The Paleovalleys with Neogene Fill

Numerous paleovalleys filled with Miocene strata have been reported from the Carpathian foreland. On the territory of the Czech Republic, large, tectonically controlled Bludovice and Detmarovice paleovalleys are cut into the upper Paleozoic strata of the Upper Silesian coal basin and filled with Karpatian to Badenian strata of the Carpathian foredeep (Jurkova, 1961, 1976). These features, as well as the intervening Ostrava–Karvina and Pribor–Tesin ridges, trend generally in the west–east direction.

Paleovalleys with Neogene fill have been identified in the Carpathian foreland in eastern Poland (Czernicki and Karnkowski, 1987), in Western Ukraine (Sovchik and Vul, 1996), and in the Moesian Platform of Romania (Paraschiv, 1979). Although no obvious Paleogene sediments have been found in these paleovalleys in the territory of Poland, Krzywiec (2001) believes that these features in eastern Poland and the adjacent part of Ukraine were incised in the Paleogene in conjunction with the inversion and erosion of the Polish trough. This, however, would not determine the time of excavation of paleovalleys in other parts of the Carpathian foreland. Whether both the paleovalleys with the Paleogene fill, like the Nesvacilka and Vranovice canyons of southern Moravia, and those filled with Neogene deposits were cut at the same time during the Late Cretaceous to early Paleogene uplifting of the Carpathian foreland and filled with either Paleogene or Neogene strata or whether there were successive stages of paleovalleys cutting and filling remains to be discovered. Paleogeographic reconstruction of the Paleogene strata in the Alpine Molasse and in southern Moravia would indicate that the Paleogene transgression might have been limited to the southern zone of the foreland, including the Nesvacilka and Vranovice paleovalleys and the Hystria and Kamchia depression in the Black Sea shelf. The younger, Miocene transgressions then progressed farther north, into the paleovalleys of northern Moravia, eastern Poland, and Western Ukraine. The different regional extent of Paleogene and Neogene transgressions may thus explain why various paleovalleys, possibly incised at the same time, during the Late Cretaceous–Paleocene uplifting of the Carpathian foreland, were filled with either Paleogene or Neogene strata.

The Structure of the Neogene Foredeep

The Neogene strata of the foredeep remain generally undeformed. Just along the very front of the Waschberg–Zdanice–Subsilesian nappe, e.g., in central Moravia and in the Pavlov Hills of southern Moravia, the Neogene strata of the foredeep are detached and integrated into the imbricated frontal zone of the thrust belt. The structural pattern of the Neogene foredeep is dominated by a system of synthetic and some antithetic faults formed during the flexural downbending of the foreland in front of the progressing thrust belt. The presence of antithetic faults seems to be critical for the formation of structural traps (Adamek, 1979). However, the character of some of these faults is not clear and remains a subject of disputes. One of them is the antiformal feature, which underlies the Dunajovice gas field located in the Neogene foredeep west of Mikulov (Figure 26, shown on page 137). On its western side, the structure is bounded by an eastward steeply dipping Vestonice fault, which penetrates through the Jurassic strata into the crystalline basement but does not continue directly into the overlying Neogene foredeep deposits, which drape over this structure. The thinning of the lower Miocene (Eggenburgian) strata over the structure would suggest that the fault was active during the early Miocene. Adamek (1979) and Kostelnicek et al. (2006) interpret this structure as an antithetic westward-dipping normal fault. However, the overall geometry of the structure, as seen on the seismic data, suggests that this might be rather a steep reverse and/or strike-slip fault, whose origin may be related to the compressive or transpressive stresses, which propagated into the foreland from the progressing thrust belt at a deeper structural level.

The structural pattern of the Neogene foredeep in Moravia has been only partly mapped and interpreted (e.g., Jurkova, 1961, 1971; Jurkova and Novotna, 1974; Jurkova, 1976; Elias and Palensky, 1997). The full evaluation of the structural architecture of the foredeep may yet provide new opportunities for the exploration for hydrocarbons.

THE STRATIGRAPHY AND DEPOSITIONAL SYSTEM OF THE OUTER WESTERN CARPATHIAN THRUST BELT

The Outer Western Carpathian thrust belt may be defined as a stack of rootless, thin-skinned nappes thrusted over the Neogene foredeep and the underlying West European platform (Figures 2, 3). In a broad sense, it consists of the Pieniny Klippen Belt, the Flysch belt, and the successor Vienna basin, which, mostly undeformed, rests on the top of the Carpathian belt. The extent of tectonic shortening and the character of the original depositional sites of various thrust sheets are little known and remain subjects of alternate geodynamic reconstructions. Different views also exist on the character of the crust underlying the depositional sites, which apparently varied from continental to transitional and, according to some views, possibly to oceanic, and on the amount of its shortening, accretion, and subduction.

In terms of the Carpathian terminology, the Flysch belt includes not only the typical deep-water turbiditic flysch deposits of alternating shales and sandstones found mainly in the Magura flysch and the Silesian units but also the hemipelagic marginal deposits characteristic for the external Pouzdrany, Waschberg–Zdanice–Subsilesian units, and parts of the Silesian unit. The definition of Carpathian Flysch belt is thus based more on structural rather than lithological criteria. In that respect, the Carpathian terminology differs from that used in the Alpine region. There, the term flysch is used exclusively for typical synorogenic deep-water turbiditic deposits of the inner zones of the depositional system, whereas the hemipelagic and shallow-marine strata of the more external zones of the Outer Alpine depositional system are referred to as Helveticum or even included into the Alpine Molasse (Figure 16).

In the Alpine nomenclature, the term Molasse is traditionally used for undeformed and marginally deformed (imbricated Molasse) late Eocene to Oligocene marginal marine deposits (lower marine molasse) of the external zones of the depositional system, as well as for the shallow-marine and terrestrial Miocene deposits of the Neogene foredeep (lower nonmarine, upper marine, and upper nonmarine molasse). In the Carpathians, the deformed Paleogene to early Miocene (Egerian) strata are included into the Flysch belt, whereas the undeformed Neogene deposits are attributed to the Carpathian foredeep.

The depositional assemblies of the Outer Carpathians comprise sequences formed during both the divergent and convergent stages of the evolution of the Tethyan–Alpine system. The preorogenic depositional assembly of divergent margins encompasses the Jurassic to the Lower Cretaceous carbonate and siliciclastic

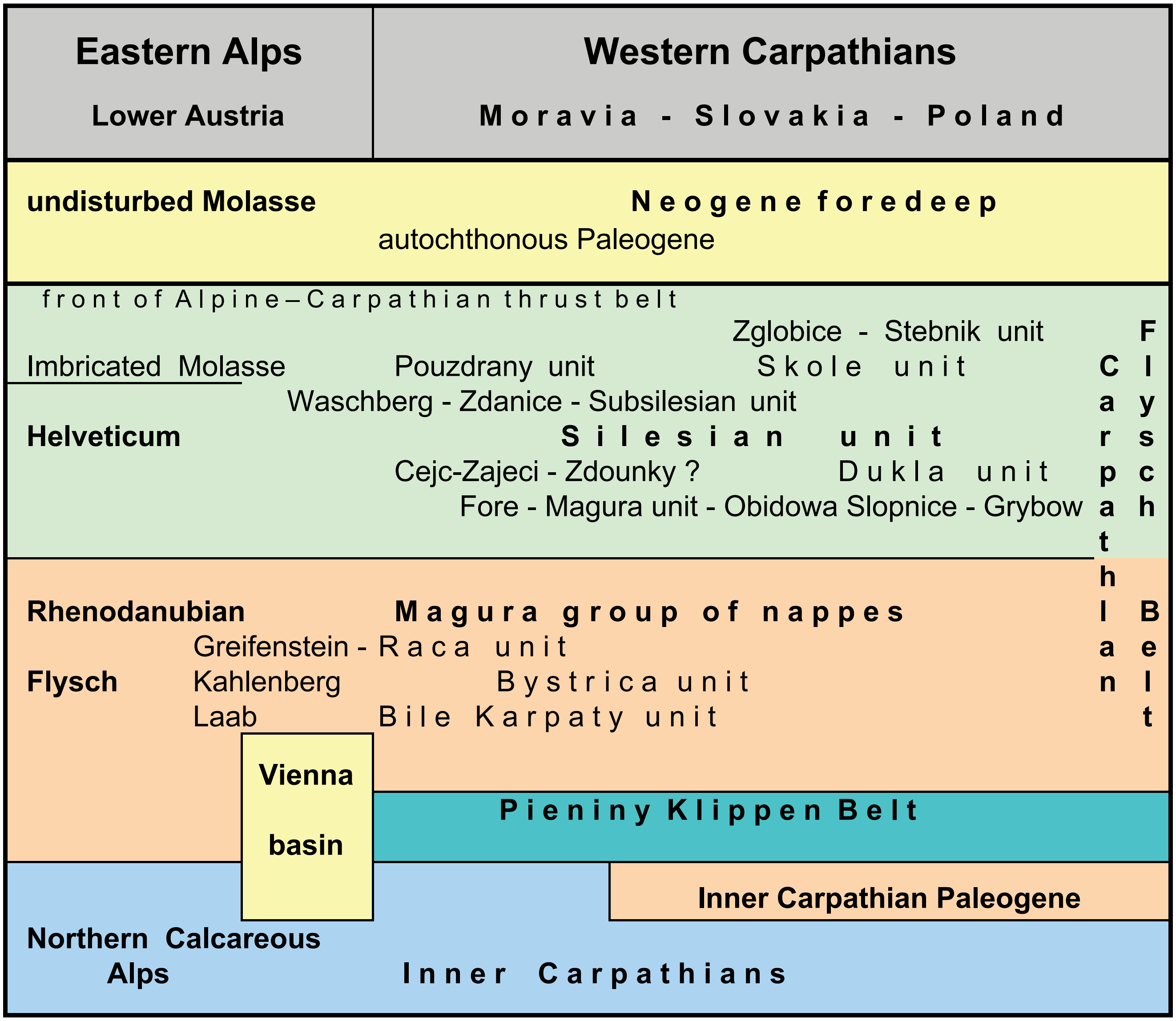


Figure 16. Regional distribution and correlation of principal tectonostratigraphic units of the external Eastern Alps in northeastern Austria and the Outer Western Carpathians in Moravia (Czech Republic), Slovakia, and Poland. Arranged from the foreland (top) to the hinterland (bottom).

strata, including some deep-water turbidites. However, with the exception of the Silesian unit, these strata are seldom exposed on the surface. Only scattered information about their character and distribution are available from wells and tectonic klippen and olistoliths incorporated into the thrust belt. The Late Cretaceous to early Miocene synorogenic flysch and molasse deposits of the convergent tectonic setting represent the bulk of the Flysch belt depositional assembly. They were deposited and redeposited in a series of troughs and foredeeps and sourced from internal ridges (cordilleras), the prograding thrust belt, as well as from the European foreland.

The conversion from the divergent to the convergent tectonic setting in the Outer Carpathian realm, which resulted from the collision of the Inner Carpathians with the rifted and attenuated margins of Northern Europe in the Late Cretaceous, was gradual and not always clearly defined. In the outermost zones, e.g., in the autochthonous Paleogene, the Pouzdrany, and the Waschberg–Zdanice–Subsilesian domains, the Late Cretaceous and Paleogene synorogenic sequences are separated from the Jurassic to the Lower Cretaceous strata of the passive margins by a major hiatus. In the more internal Silesian and Magura units, however, the sedimentation continued without major interruption from one regime to another. The transition from the divergent to the active convergent setting, especially in the external zones of the Outer Carpathian depositional system, was also partly obscured by the worldwide transgression and high-water stand in the Cenomanian–Turonian. Possibly, the best indicator of the inception of the synorogenic regime is the sudden appearance of thick packages of coarse turbiditic sands, such as the Godula sandstones in the Silesian unit or Solan sandstones in the Magura unit (Figure 17A, B).

Reasoning on the fact that the convergent depositional system of the Outer Carpathians gradually progressed toward the foreland and was supplied with detrital material mainly from the tectonically activated inner ridges and the progressing orogenic belt, Picha and Stranik (1999) defined the entire convergent depositional system since the Upper Cretaceous to Neogene as a foreland basin. Despite the possibility that during the Late Cretaceous to early Paleogene, the innermost part of this foreland basin, particularly the Magura deep-water depositional site, might have been still at least partly underlain by an attenuated transitional crust or even, as some believe, an oceanic crust, the term foreland basin seems to be justified. Under such a broad definition, the Carpathian foreland basin would include both the Late Cretaceous to early Miocene synorogenic, predominantly deep-water flysch sequences constituting the Flysch belt and the early to late Miocene late orogenic and postorogenic shallow-marine and continental molasse deposits of the Neogene foredeep and of the Vienna basin. In that respect, the Alpine–Carpathian foreland basin, by having two stages, namely, the deep-water flysch and the shallow-marine and continental molasse, compares well with other foreland basins of the Tethyan system, such as the Apennines and the Dinarides–Hellenides, or with the Ouachitas of North America (Golonka et al., 2006b). However, it differs from some other classical foreland basins, such as the well-studied Cretaceous to Paleogene Rocky Mountain foreland basin of North America, which does not have the typical deep-water flysch phase and is filled only with molasse-type deltaic and other shallow-marine and continental deposits sourced from the Sevier orogenic belt as well as the foreland structures of the Rocky Mountains.

The terms foreland basin and foredeep are commonly used interchangeably for various foreland depositional settings. In this chapter, following the previous usage by Picha and Stranik (1999), we keep the term foredeep for a fast subsiding wedge-shaped depocenter bordering the structural front of the thrust belt and supplied primarily from the emerging thrust belt. The foredeeps are formed typically by flexural downbending of the foreland plate caused by the tectonic load of the progressing accretionary prism (Allen and Allen, 1990). The foreland basin, as defined, for example, by DeCelles and Giles (1996), is a more complex depositional setting, which, in addition to the foredeep proper, may also include the flexural forebulge and the broad landward part of the basin supplied predominantly from the foreland. That means that the foredeep is merely the innermost, deepest, and most active part of a foreland depositional setting of the foreland basin. As the front of deformation moved toward the foreland, so did the foredeep, the forebulge, and the rest of the foreland basin.

The typical foreland basin succession records a marine transgression, followed by a deepening phase, and later by regression to alluvial and lacustrine conditions (Lickorish and Ford, 1998). The foreland basin thus passes from an underfilled to an overfilled stage. The Carpathian foreland basin, in general terms, fits this classical model with the hemipelagic and deep-water turbiditic flysch deposits representing the underfilled synorogenic stage and the molasse shallow-marine and continental deposits exemplifying the overfilled stage. However, only the outer zones of the Carpathian foreland basin began with transgression over the foreland, whereas in the inner zones, the sedimentation passed from the divergent to the convergent stage continuously without any major interruption.

As indicated by the distribution of various lithostratigraphic facies, the structural and depositional pattern in the Carpathian basin was complex. The basin architecture was characterized by a rugged topography, partly inherited from the previous divergent stage and further enhanced by the foreland-type crustal uplifting, caused by compressional and transpressional orogenic stresses. The depositional systems in the foreland basin were thus determined not only by the progression of the orogenic belt, global fluctuations of sea level, and climatic changes, but also by the orogenic activation of preexisting basement structures.

THE WESTERN CARPATHIAN FLYSCH BELT

The thin-skinned Outer Western Carpathian Flysch belt consists of numerous tectonostratigraphic units and subunits. In the Polish Carpathians, Nowak (1927) separated them into three groups: the marginal, central (Menilite–Krosno), and inner (Magura) groups. However, the delimitation of the marginal group, which, in Moravia, will be represented only by the Pouzdrany unit, proved to be questionable. The Pouzdrany unit, as well as the autochthonous Paleogene, may be better classified as a continuation of the Alpine Molasse into the territory of southern Moravia. To simplify the matter, Swidzinski (1934) distinguished only two groups of units: the external and internal. The external or Menilite–Krosno group of units is characterized by the Late Cretaceous to late Eocene succession of variegated shales, early Oligocene menilitic silicites, and the late Oligocene to early Miocene Krosno-type flysch. In a broad sense, based mainly on their structural position, these units are comparable with the Helvetic units of the Alps. In Moravia, this external group includes the Pouzdrany, Waschberg–Zdanice–Subsilesian, Silesian, Cejc–Zajeci, Zdounky, and Fore-Magura units (Figures 3; 16; 17A, B). Thick Late Cretaceous to Eocene deep-water flysch deposits characterize the internal Magura unit, which, during the sedimentation, was separated from the external units by the Silesian ridge. The Magura unit, also referred to as the Magura group of nappes, is correlated with the Rhenodanubian Flysch of the Alps (Elias et al., 1990).

The stratigraphic extent of these tectonic units of the Carpathian Flysch belt varies but, with the exception of the Silesian nappe, only seldom encompasses the complete stratigraphic range of the original depositional assembly from the Jurassic to the early Miocene. The various units of the Carpathian flysch belt are composed predominantly of the Late Cretaceous to Miocene clastic deposits, both shelf and deep-water flysch facies, with only a token presence of Jurassic and Lower Cretaceous carbonates and clastics. These older synrift and continental margin strata laid down in a structurally diversified setting of rift-related basins and platforms might have been cut off by the sole detachment and left behind in their autochthonous position. Occasionally, tectonic slivers of these Jurassic and Cretaceous carbonate and clastic strata were detached and incorporated into the Carpathian Flysch belt as the so-called ''tectonic klippen.'' Typically well exposed on the surface, these klippen provide valuable information about the character and distribution of Jurassic and Cretaceous strata along the margins of the European plate now deeply buried below the thrust belt.

Some incompetent depositional sequences were reduced tectonically, smeared, or piled in duplexes at the base of major thrust sheets. The stratigraphic extent of sequences was also reduced by erosion, which occurred either prior or during and after the deformation. The reconstruction of the full original depositional record thus requires integration of fragments of evidence gathered from numerous sites of tectonically separated units. Without strong support from biostratigraphy, such a reconstruction would be very difficult if not impossible.

The Pouzdrany Unit

The marginal Pouzdrany unit, as defined by Cicha et al. (1964, 1965), is the outermost tectonostratigraphic unit of the Carpathian thrust belt in Southern Moravia. It appears as a narrow imbricated thrust sheet trending along the frontal edge of the Waschberg–Zdanice nappe in southern Moravia and northeastern Austria (Stranik, 1996) (Figure 3). However, it has not been officially recognized as a separate unit in Austria, where its equivalents are traditionally included into the Waschberg zone.

The Pouzdrany unit comprises strata ranging from the late Eocene to the early Miocene. They have been divided into five lithostratigraphic units: (1) the Pouzdrany Marls, (2) the Uhercice Formation, including the diatomites, (3) the Boudky Marls, (4) the Krepice Formation, and (5) the Sakvice Marls (Figure 17A, B).

The oldest known strata in the Pouzdrany unit are the brown fossiliferous shales and *Globigerina* Marls described as the Pouzdrany Marls by Rzehak (1880, 1895a) and dated as the late Eocene to early Oligocene by Pokorny (1954), Cicha (1975), Krhovsky (1981), Krhovsky et al. (1992), and others. Locally, these predominantly pelitic strata contain intercalations and concretional bodies of limestones with rich late Eocene faunas (Oppenheim, 1922; Ctyroky, 1966). These limestones have been named the Moutnice Limestones and correlated with the Hollingstein Limestones of the Waschberg zone in Austria by Ctyroky (1966). The Pouzdrany Marls are overlain by diatomites with cherty concretion and pelagic micritic limestones and dolomites, followed by brown and gray shales with limonite and jarosite coatings and gypsum crystals (weathering products of pyrite) and scattered lenses of glauconitic sands with shark teeth and brachiopods. The whole sequence, about 200 m (660 ft) thick, was studied by Pokorny (1960), Cicha et al. (1965), and Stranik (1981) and named the Uhercice Formation by Krhovsky (1981). The diatomites have been correlated with the Menilitic Formation of the Zdanice unit and assigned to the lower Oligocene by Picha and Stranik (1999). In the autochthonous Paleogene series, the early Egerian Malesovice Member might be considered as an equivalent of the Uhercice Formation (Figure 17A, B).

The Uhercice Formation is followed by an approximately 30-m (100-ft)-thick sequence of gray pelagic marls with dolomitic concretions. Cicha et al. (1964, 1965) named these strata the Boudky Marls and correlated them with the Egerian (Aquitanian) Michelstetten beds (Grill, 1962) of the Waschberg zone in Austria. Cicha (1975) and Cicha et al. (1975b) assigned them to the late Egerian. Krhovsky (1981) and Stranik et al. (1981a) dated the Boudky Marls as the late Egerian to early Eggenburgian. Picha and Stranik (1999) tentatively suggested the late Oligocene to early Miocene (Egerian) age for these open-marine marls as the most likely. The Boudky Marls may represent a distal landward facies of the Outer Carpathian foreland basin. It might have been deposited on or close to a foreland

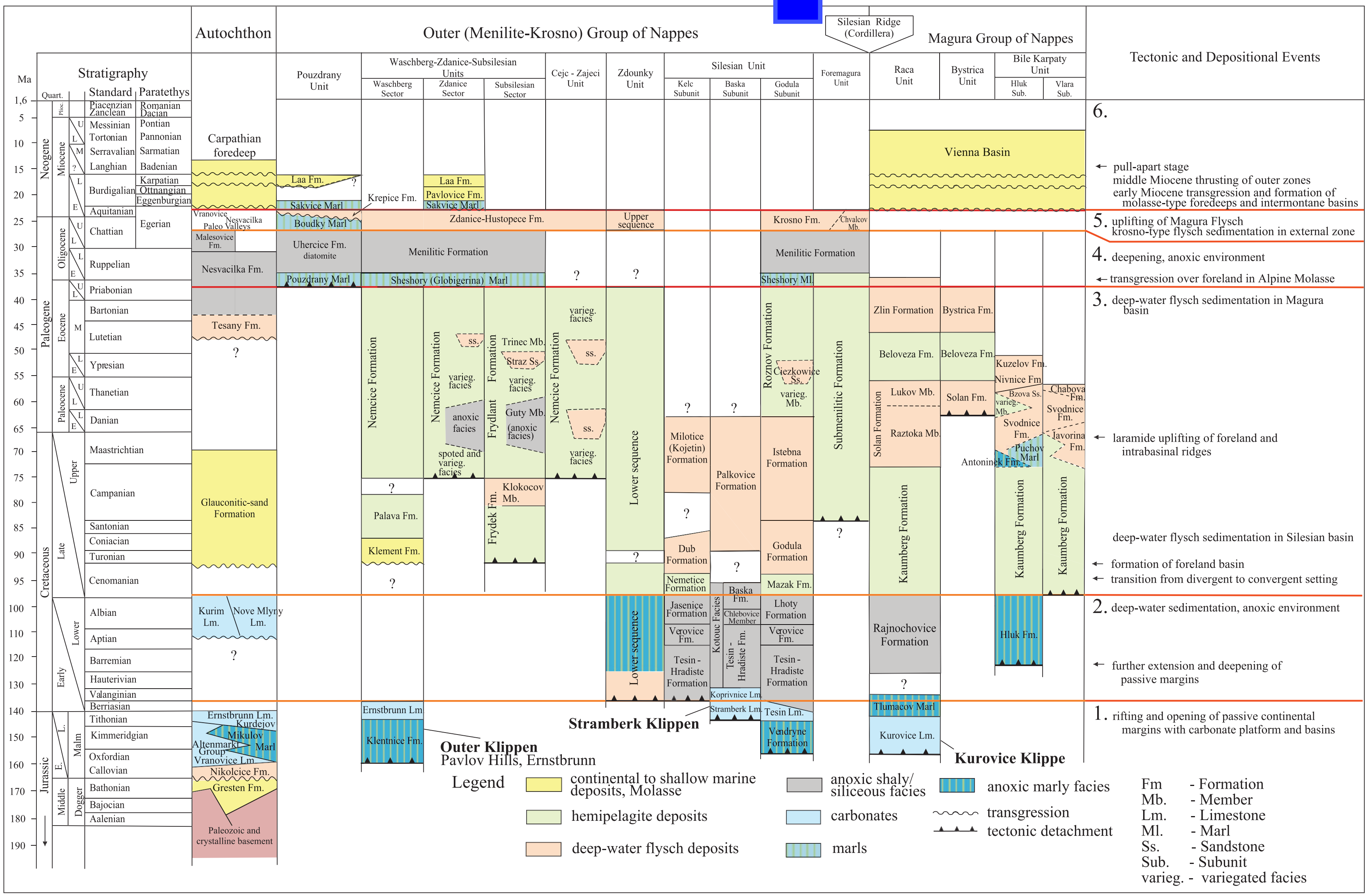


Figure 17. (A) Stratigraphic correlation of principal tectonostratigraphic units of the Outer Western Carpathians in Moravia. Major depositional sequences distinguished and numbered 1–6.

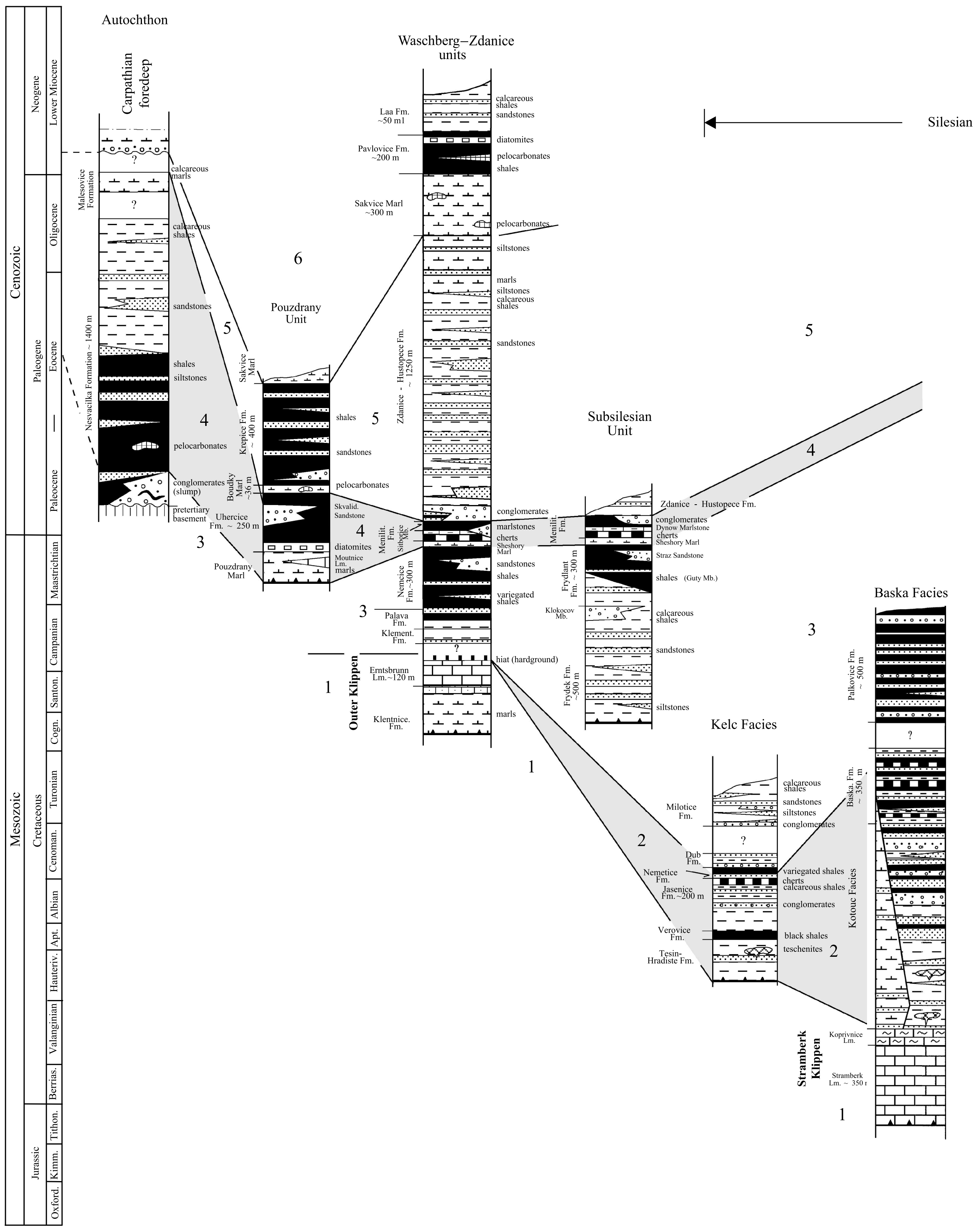
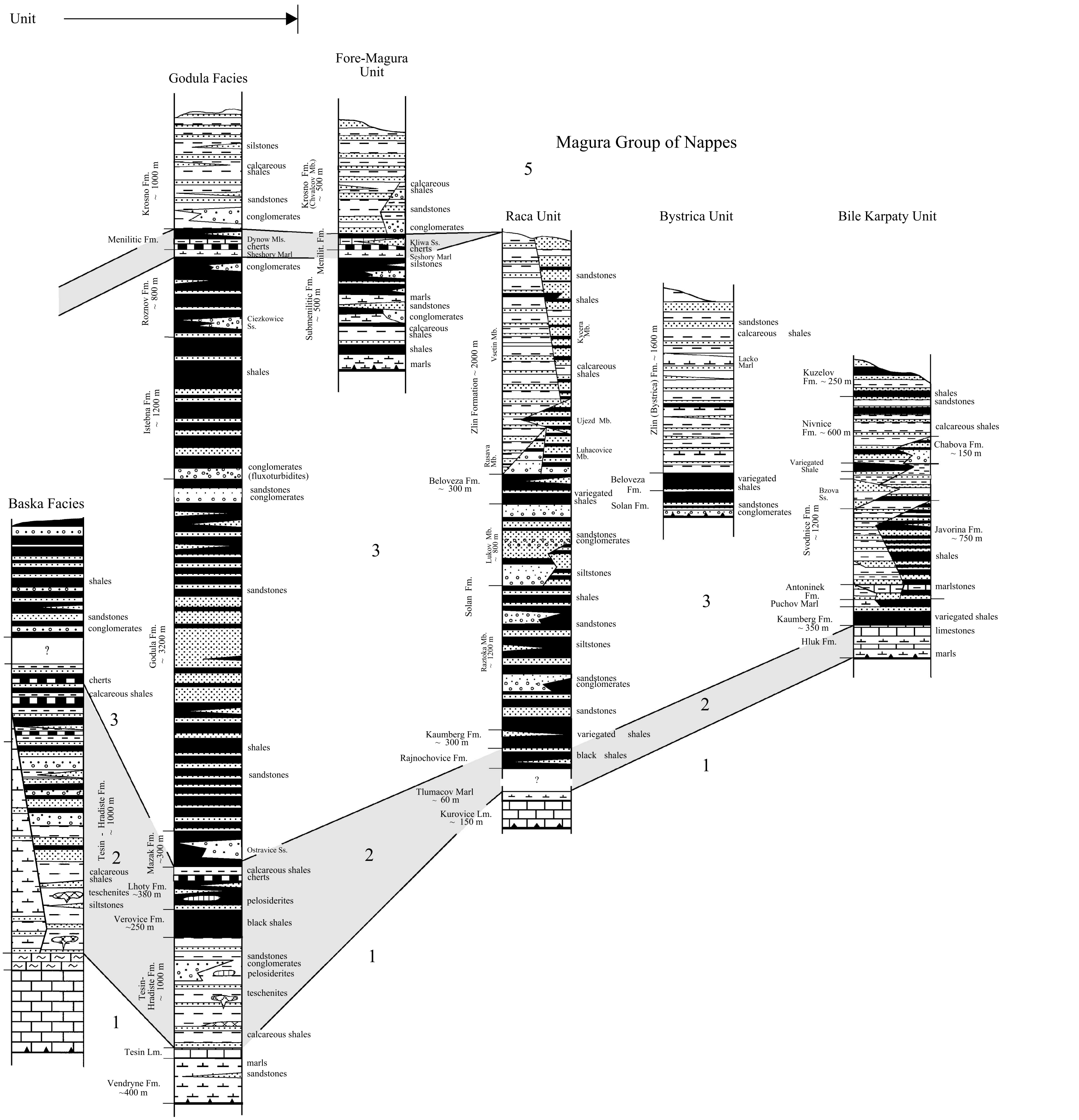


Figure 17. (cont.) (B) Lithologic columns of various tectonostratigraphic units of the Outer Western Carpathian Flysch belt in Moravia. For time (chronostratigraphic) correlation of various stratigraphic units and major depositional sequences 1–6, see Figure 17A. The lithologic columns for the Godula facies of the Silesian unit and all units of the Magura group of nappes are reduced by 50% with respect to other columns.


Unit
Godula Facies
Fore-Magura Unit
Magura Group of Nappes
5
Raca Unit
Bystrica Unit
Bile Karpaty Unit
Baska Facies
Krosno Fm. ~ 1000 m
silstones
calcareous shales
sandstones
conglomerates
Menilitic Fm.
Dynow Mls.
cherts
Sheshory Marl
conglomerates
Roznov Fm. ~ 800 m
Ciezkowice Ss.
shales
Istebna Fm. ~ 1200 m
conglomerates (fluxoturbidites)
sandstones conglomerates
Godula Fm. ~ 3200 m
sandstones
shales
sandstones
Mazak Fm. ~300 m
Ostravice Ss.
calcareous shales
cherts
Lhoty Fm. ~380 m
pelosiderites
Verovice Fm. ~250 m
black shales
Tesin-Hradiste Fm. ~ 1000 m
sandstones
conglomerates
pelosiderites
teschenites
calcareous shales
Tesin Lm.
marls
sandstones
Vendryne Fm. ~400 m
Krosno Fm. (Chvalcov Mb.) ~ 500 m
calcareous shales
sandstones
conglomerates
Menilit. Fm.
Kliwa Ss.
cherts
Seshory Marl
siltstones
Submenilitic Fm. ~ 500 m
marls
sandstones
conglomerates
calcareous shales
shales
marls
3
shales
sandstones conglomerates
?
cherts
calcareous shales
3
Tesin - Hradiste Fm. ~ 1000 m
2
calcareous shales
teschenites
siltstones
1
2
1
Zlin Formation ~ 2000 m
Vsetin Mb.
Kycera Mb.
sandstones
shales
calcareous shales
Ujezd Mb.
Rusava Mb.
Luhacovice Mb.
Beloveza Fm. ~ 300 m
variegated shales
Solan Fm.
Lukov Mb. ~ 800 m
sandstones conglomerates
siltstones
shales
Raztoka Mb. ~ 1200 m
sandstones
siltstones
sandstones conglomerates
sandstones
Kaumberg Fm. ~ 300 m
variegated shales
Rajnochovice Fm.
black shales
?
Tlumacov Marl ~ 60 m
Kurovice Lm. ~ 150 m
Zlin (Bystrica) Fm. ~ 1600 m
sandstones
calcareous shales
Lacko Marl
Beloveza Fm.
variegated shales
Solan Fm.
sandstones conglomerates
3
2
1
Kuzelov Fm. ~ 250 m
shales sandstones
Nivnice Fm. ~ 600 m
calcareous shales
Chabova Fm. ~ 150 m
Variegated Shale
Bzova Ss.
Svodnice Fm. ~ 1200 m
Javorina Fm. ~ 750 m
shales
Antoninek Fm.
marlstones
Puchov Marl
Kaumberg Fm. ~ 350 m
variegated shales
limestones
Hluk Fm.
marls

bulge, which formed in response to the tectonic loading and downbending of the foreland crust in front of the progressing Carpathian thrust belt in the late Oligocene and early Miocene.

The Boudky Marls are overlain by the flyschlike Krepice Formation (Cicha et al., 1964, 1965), a sequence as much as 400 m (1300 ft) thick of alternating shales and fine micaceous sandstones and siltstones. Based on similarities in lithology, the Krepice Formation has been interpreted as a distal facies of the Oligocene to early Miocene (Egerian) Zdanice–Hustopece Formation of the Zdanice unit (Cicha et al., 1964, 1965; Picha, 1979b). This is a typical synorogenic turbiditic flysch facies, whose base may become progressively younger toward the foreland, thus reflecting on the gradual progression of deformation and foredeep deposition. An existence of a local erosional event at the base of the Krepice Formation (Stranik et al., 1981a) would support such an interpretation. The distal Krepice Formation of the Pouzdrany unit may, in fact, correlate only with the upper part of the Zdanice–Hustopece Formation of the Zdanice unit (Figure 17A, B). The underlying Boudky Marls, possibly deposited on the foreland bulge, may then be coeval with the lower part of the Zdanice–Hustopece Formation elsewhere (Picha, 1979b; Stranik, 1983). Because of the lack of diagnostic fossils and the structural complexity of the Pouzdrany unit, the stratigraphic position of the Krepice Formation has been disputed. The gray marls and calcareous shales overlying the Krepice Formation have been compared with the Eggenburgian Sakvice Marls of the Zdanice unit (Stranik and Molcikova, 1980) (Figure 17A, B). Such a superposition would support the Egerian age of the underlying Krepice Formation. However, based on the presence of rusty-colored limonitic siltstones and claystones with sparse laminae of diatoms in its upper part, Krhovsky et al. (1995) compared the Krepice Formation with the Ottnangian to Karpatian Pavlovice Formation of the Zdanice unit (Stranik, 1983). Considering the overall lithologic similarity with the Zdanice–Hustopece Formation and the stratigraphic position below the Eggenburgian strata, the interpretation of the Krepice Formation as a distal facies of the Zdanice–Hustopece Formation of the Egerian age seems to be more likely.

Cicha et al. (1965) defined the Pouzdrany unit as a tectonic sliver detached from the more internal zones of the autochthonous Paleogene series and tectonically transported at the front of the thrust belt during the youngest early Miocene phase of thrusting. The obvious spatial proximity of the depositional sites of these two units (Figure 5D) and similarities, both lithological and faunal, of their upper Eocene to lower Oligocene strata would advocate such an interpretation. Because no equivalents of the younger, lower Miocene members of the Pouzdrany unit, the Boudky Marls and the Krepice Formation, have been found in the autochthonous series, it is speculated that these younger members of the sequence were detached and tectonically transported at the front of the thrust belt with the rest of the Pouzdrany unit, whereas the older members, mainly the Eocene fill of the paleovalleys, were mostly preserved in their autochthonous position.

Stratigraphically, the strata of the Pouzdrany unit would correlate with the Lower Marine Molasse of the Alps. Structurally, the Pouzdrany unit may be compared with the imbricated (Gefaltette) Molasse of Austria, as described, e.g., by Wagner (1996).

The Waschberg–Zdanice–Subsilesian Units

This major frontal thrust system of the Outer Western Carpathian belt consists of three sectors: the Waschberg sector (unit) of northeastern Austria and southernmost Moravia (south of the Dyje River), the Zdanice sector (unit) of southern and central Moravia, and the Subsilesian sector (unit) of northern Moravia (north of the Upper Moravian Depression) and western Poland (Figure 3). On the Polish territory, the Subsilesian unit may be traced to the Bielsko–Biala area; its continuation further to the east is not clear. Included in the Subsilesian unit in Poland are also some Lower Cretaceous strata (Slaczka et al., 2006), which Elias (1998) compares with the coeval strata of the Kelc subunit of the Silesian unit of central Moravia. Together with some elements of the Silesian unit, the Waschberg–Zdanice–Subsilesian thrust system is comparable with the Helvetic nappes of the Alps.

The inner structure of the rootless, tectonically transported Waschberg–Zdanice–Subsilesian units is complex. This lowermost unit of the Outer Western Carpathian nappe stack consists of numerous imbricates, duplexes, and partial thrust sheets of the Upper Cretaceous to lower Miocene strata. In the Washberg sector of northeastern Austria and southernmost Moravia, tectonic slivers (klippen) of Jurassic marls and carbonates and Upper Cretaceous (Turonian to Campanian) clastic rocks are tectonically incorporated into the thrust sheets of younger, Upper Cretaceous to lower Miocene sequences (e.g., Juttner, 1922, 1928, 1933, 1942; Stranik, 1963; Stranik et al., 1999, and others). These so-called ''Outer Klippen'' of the Palava Hills in Moravia and of the Ernstbrunn area in Austria were detached from the underlying European foreland and tectonically integrated into the frontal zones of the Carpathian thrust belt during the last stages of the thrusting.

For better clarity and in compliance with some traditional views, the Outer Klippen of the Waschberg sector in Moravia, the Waschberg–Zdanice sector, and the Subsilesian sector of the Waschberg–Zdanice–Subsilesian thrust system in Moravia are discussed in three separate sections. The Waschberg sector of northeastern Austria is not covered in our chapter.

The Outer Klippen of the Waschberg Sector

The Outer Klippen of the Waschberg unit varies in size from small tectonic slivers to large bodies several kilometers long and several hundred meters thick. They consist of several formations of the Late Jurassic and Late Cretaceous age (Figure 17A, B). The oldest Klentnice Formation of Oxfordian to early Tithonian age (Hanzlikova, 1965b) is composed predominantly of dark-gray marls and calcareous shales with an upward-increasing proportion of fine-grained limestones. The poorly preserved fauna of ammonites, bellemnites, brachiopods, pelecypods, crinoids, sponges, bryozoans, and ostracods was described by K. Matzka (1934, personal communication), Bachmayer (1957), Pokorny (1959, 1973), and Vasicek (1971a), among others. Both biostratigraphically and lithologically (Elias, 1991), the Klentnice Formation is comparable with the Mikulov Marls of the autochthonous foreland plate. The known thickness of the Klentnice Formation does not exceed 200 m (660 ft). The marls of the Klentnice Formation are organic rich; however, because of their limited extent and the insufficient burial history, they have practically no potential as source rocks for hydrocarbons.

The Klentnice Formation passes gradually into the overlying Ernsbrunn Limestones of the Tithonian to Berriasian and, according to Elias and Eliasova (1984), up to Hauterivian(?) age. The lower part of the formation is dominated by brecciated organodetritic limestones with matrix of calcareous shales and occasional large clasts of limestones up to several meters in diameter. This facies may represent a detrital talus of a carbonate platform dominated by gravitational transport, including slides, debris flows, and turbidites. Thick-bedded, partly dolomitized calcarenites (locally oolitic) and micritic limestones, which apparently originated in the shallow-water environment of the carbonate platform, make up the upper part of the formation. Occasional hardgrounds and karstifications testify to sporadic emergencies of parts of the platform. The rich fauna, in addition to forms found in the Klentnice Formation, also includes fragments of corals, stromatoliths, and calcareous algae and fish. The known thickness of the Ernsbrunn Limestones is about 120 m (400 ft).

Traditionally, the Ernstbrunn Limestones and the underlying Klentnice Formation of the Outer Klippen have been interpreted as a tectonically detached part of a carbonate succession, which evolved on the rifted passive margins in the Oxfordian and Tithonian. An alternative interpretation by Elias and Eliasova (1984) assumes that the Ernstbrunn Limestones represent a pile of carbonate debris derived from a preexisting hypothetical Tithonian Pavlov platform and redeposited into the Zdanice basin in time of a eustatic drop of the sea level.

Equivalents of Ernsbrunn Limestones, if preserved in an autochthonous position underneath the Carpathian thrust belt, would represent a potential reservoir conveniently charged by hydrocarbons from the underlying organic-rich basinal deposits of the Klentnice Formation.

The Tithonian to Berriasian Ernsbrunn Limestones in the Pavlov Hills and the Waschberg zone are transgressively overlain by the Turonian–Coniacian Klement Formation (Klement Supergroup in Austria) composed of shales, glauconitic sandstones, and sandy limestones. Both lithologically and biostratigraphically, the Klement Formation resembles the Upper Cretaceous epicontinental boreal deposits of northern Europe, specifically the Brezno Formation of the Bohemian Massif (Stranik et al., 1996). The maximum known thickness of the Klement Formation in southern Moravia is 32 m (105 ft). The juxtaposition of the Ernstbrunn and the Klement formations documents the existence of a stratigraphic gap, which most likely lasted from the Valanginian to the Cenomanian and is marked by distinct hardgrounds (Stranik et al., 1996). In that sense, the marginal depositional zone of the Outer Klippen differed from the Silesian basin (Godula subunit), where the sedimentation continued from the Late Jurassic to the early Neogene without any significant interruption (Figure 17A, B).

The overlying Palava Formation (Stranik et al., 1996), previously known as the Mucronata Marls (Abel, 1899), is composed of gray calcareous shales assigned to the late Coniacian to Campanian (Stranik et al., 1996). Fuchs and Wessely (1996) dated the Mucronata Marls of the Outer Klippen in northeastern Austria as the late Maastrichtian. Despite the differences in dating, the Palava Formation and the Mucronata Marls in both these areas are believed to have evolved without an interruption from the underlying Klement Formation (Klement Supergroup in Austria). An existence of a hiatus in the Santonian (Glaessner, 1931) has thus not been proved (Stranik et al., 1999).

The Waschberg–Zdanice Units (Sectors) in Moravia

Excluding the Jurassic and Upper Cretaceous strata of the tectonic klippen, the deposits of the Waschberg–Zdanice units in southern Moravia (the Austrian part of the Waschberg sector is not discussed) range in age from the Late Cretaceous (Campanian–Maastrichtian) to the early Miocene (Egerian to Karpatian). These strata have been divided into six lithostratigraphic formations: (1) the Nemcice Formation, including the Sheshory Marls; (2) the Menilitic Formation, including the Subchert Marls, the Menilitic Cherts, the Dynow Marls, and the Sitborice Member; (3) the Zdanice–Hustopece Formation; (4) the Sakvice Marls; (5) the Pavlovice Formation; and (6) the Laa Formation (Figure 17A, B). The first three formations represent the essentially continuous Late Cretaceous to early Miocene (Egerian) sequence of the Western Carpathian Flysch belt; the other three belong to the late orogenic transgressional Miocene molasse-type deposits comparable to the coeval deposits of the Vienna basin.

The oldest Nemcice Formation of southern Moravia, in recent publications more commonly known as the Submenilitic Formation (Swidzinski, 1948; Roth, 1962), was defined by Rzehak (1880), and its fauna was described by Rzehak (1895b) and Oppenheim (1922). It is composed predominantly of hemipelagic, gray, green, brown, black, and red calcareous shales with subordinate laminae and lenticular bodies of siltstones and sandstones. The proportion of sandstones increases from the northwest to the southeast and markedly steps up in the Cejc–Zajeci zone, interpreted by present authors as a separate unit of the Outer Carpathian Flysch belt. The age of the Nemcice Formation extends from the Late Cretaceous (Campanian–Maastrichtian) to the early Oligocene. However, the oldest Upper Cretaceous strata have been reported mostly from the innermost Cejc–Zajeci zone (unit) (Picha et al., 1968).

In the late Eocene–early Oligocene transition, the deepening and submergence of sources of clastic material, combined with global cooling, created an environment favorable for accumulation of biogenic deposits. In the external Waschberg–Zdanice–Subsilesian and Silesian basins, this new regime led to the deposition of the *Globigerina* Marls (Grzybowski, 1897), which was renamed to Sheshory Marls by Vialov (1951). In the Zdanice sector (unit), the thickness of these predominantly brown laminated marls and shales does not exceed 30 m (100 ft). They correlate well with the coeval Pouzdrany Marls of the Pouzdrany unit and also with the upper part of the autochthonous Nesvacilka Formation (Figure 17A, B). The overall depositional character of Nemcice Formation, including the Sheshory Marls, refers to an underfilled basin, in which the subsidence outpaced the sediment supply.

The Sheshory Marls pass gradually into a variety of biogenic deposits summarily assigned to the Menilitic Formation (Glocker, 1843). Stranik (1981), following the subdivision of the Menilitic Formation in Poland (Jucha and Kotlarczyk, 1961), distinguished in the Menilitic Formation of Moravia the Subchert Member, the Chert Member (Menilitic Cherts), and the Dynow Marls. He also named the uppermost predominantly shaly part of the Menilitic Formation the Sitborice Member (Figure 18).

The Subchert Marls, about 10 m (33 ft) thick, is represented by brown stratified marls and shales with fish scales. Vertically, they pass into laminated cherts and organic-rich siliceous shales of the Chert Member (Menilitic Cherts), which represents the most characteristic member of the Menilitic Formation. In the outermost zones of the Zdanice unit, the cherts are partly substituted by diatomites and in the marginal Pouzdrany unit fully replaced by diatomites (Figure 19). This would suggest that the typical Menilitic Cherts of the Zdanice–Subsilesian units were formed by the diagenetic alteration of laminated diatomites. The Menilitic Cherts reach their maximum thickness of approximately 4 m (13 ft) in the northwestern part of the Zdanice unit. Their thickness decreases both toward the south into the Waschberg zone and toward the east. The Menilitic Cherts have not been found in the Cejc–Zajeci unit.

According to Picha and Stranik (1999), the Menilitic Cherts were deposited in a zone of upwelling of nutrient-rich deep waters and proliferation of marine life (diatoms), combined with anoxic conditions on the bottom of the sea and a very limited influx of detrital material both from the foreland and the orogenic belt (Figure 19). The sedimentation was essentially a function of siliceous phyto- and zooplankton manufacture. The presence of remnants of fish with light-emitting organs (Kalabis, 1949) indicates that the depositional environment of the Menilitic Formation was relatively deep. Based on studies of the ichthyofauna, Brzobohaty (1981) and Gregorova (1988) suggested a mesopelagic depositional environment with a water depth between 200 and 1000 m (660 and 3300 ft). Picha and Stranik (1999) compared the depositional environment of the Menilitic Cherts to that which prevailed along the active margins of coastal California, where conditions for deposition of organic-rich diatomites have repeatedly occurred since the Late Cretaceous to the Miocene. The most prominent Monterey Formation of Miocene age was deposited in a neritic to bathyal environment (water depth 50–1200 m [160–4000 ft]) of silled basins formed along the active continental margins of North America (Smith, 1968).

The Menilitic Cherts are overlain by pelagic limestones and marls of the Dynow Marls (10 m; 33 ft), which are followed by a predominantly argillaceous noncalcareous sequence of the Sitborice Member

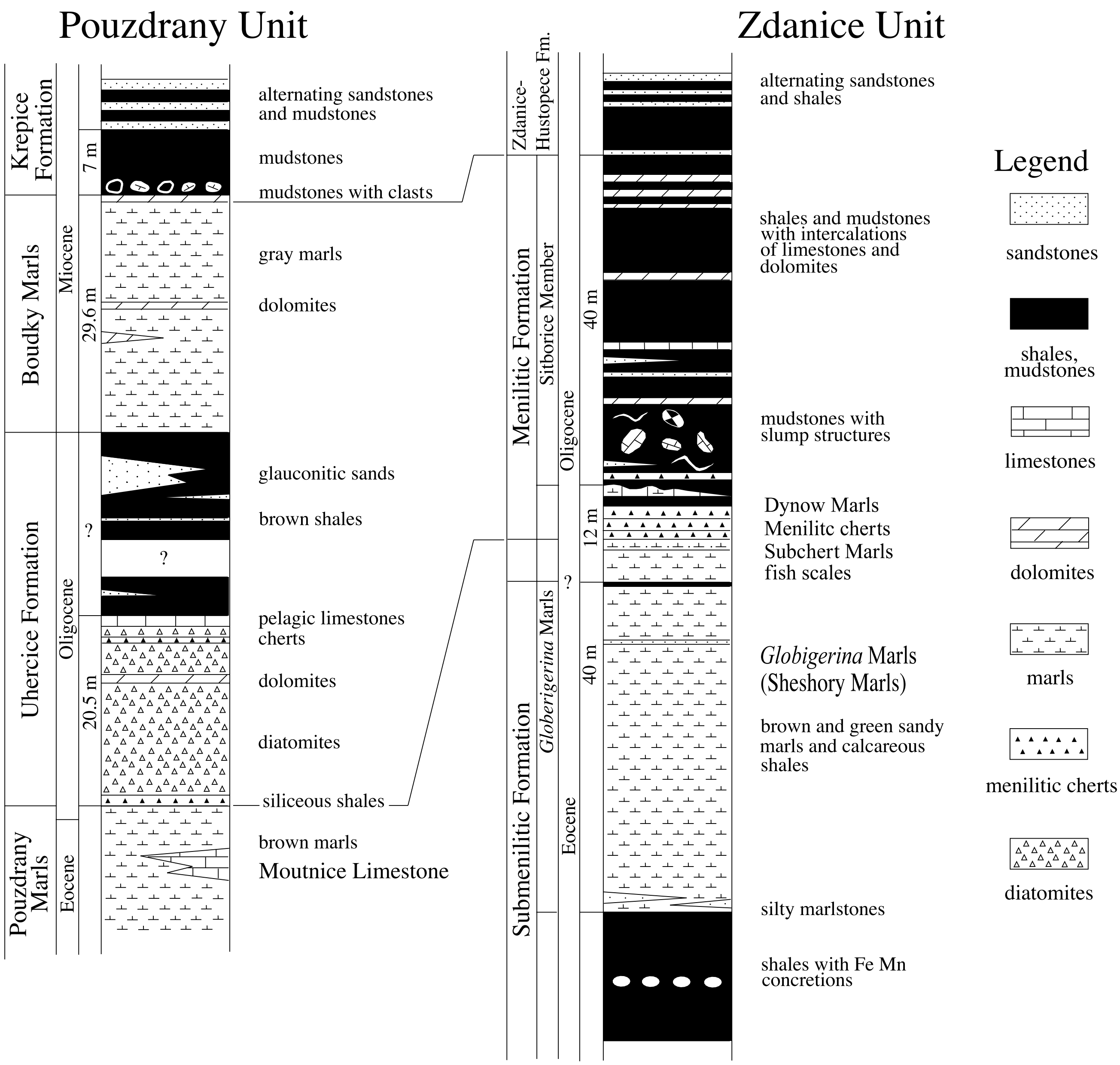


Figure 18. Correlation of the upper Eocene to the lower Miocene strata in two measured sections, one in the Pouzdrany unit at the Wine cellars in the village Pouzdrany and the second in the Zdanice unit in the road cut near the village Velke Nemcice. Modified after Picha and Stranik (1999).

(Stranik, 1981). The base of this sequence is commonly marked by an erosional surface, most likely associated with the activity of turbidity currents and debris flows. In its lower part, the Sitborice Member is made mostly of brown noncalcareous shales with occasional slump bodies of mudstones and debris flows with intraclasts of Menilitic cherts. Locally, large blocks of carbonates and crystalline rocks, apparently derived from the adjacent slopes or from the slump-prone deposits of the submarine canyons, are found in this sequence. The upper part of the Sitborice Member is represented by alternating brown and green shales with manganese and jarosite coatings, scattered crystals of anhydrite and gypsum, and intercalations and concretions of limestones and dolomites (Stranik et al., 1981b). Picha and Stranik (1999) correlated this upper part of the Menilitic Formation with the lithologically similar Uhercice Formation of the Pouzdrany unit

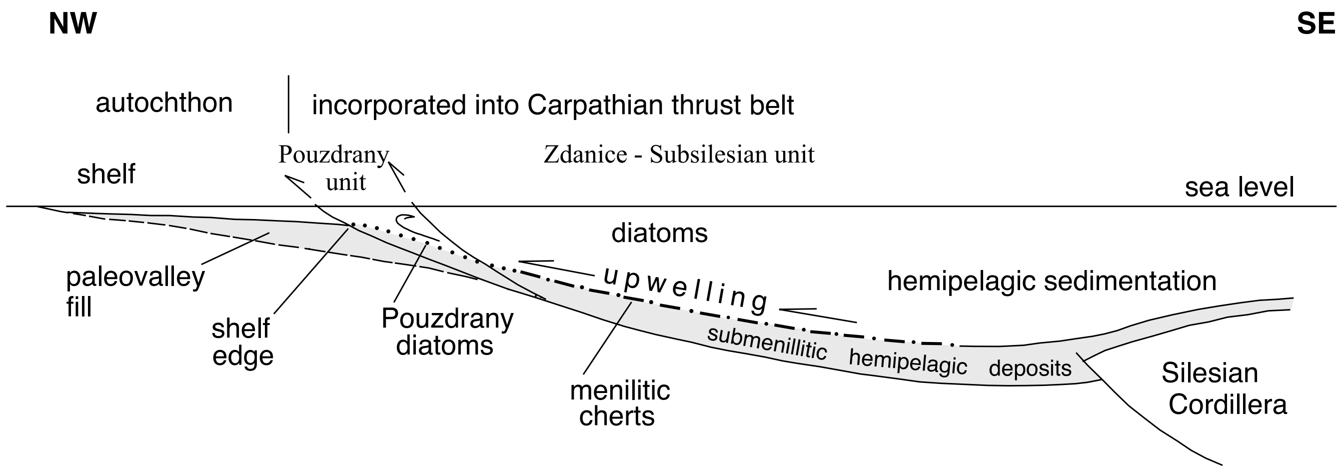


Figure 19. Depositional setting of the lower Oligocene organic-rich diatomites and Menilitic Cherts of the Pouzdrany and Zdanice units, respectively. The slope-to-basin environment was characterized by upwelling of nutrient-rich waters, proliferation of diatoms, and prevalence of anoxic conditions on the bottom (Picha and Stranik, 1999, reprinted with permission).

(Figure 17A, B). The overall thickness of the Menilitic Formation in the Zdanice unit ranges from several tens to 200 m (660 ft).

The presence of submarine slump conglomerates and debris flows in the Sitborice Member of the Zdanice unit and in the Menilitic Formation of the Subsilesian unit marks the end of the quiet period and the beginning of a new phase of tectonism. In the Outer Western Carpathians, this tectonic phase resulted in the deformation and uplifting of the Magura flysch and the formation of the Krosno-type synorogenic foredeeps.

Because of the lack of diagnostic fossils both in the Menilitic Formation and the overlying Krosno-type flysch, the exact timing of these events and even the age of the Menilitic Formation have become a subject of discussion. The Menilitic Formation of the Zdanice unit was assigned variably to the late Eocene (Pokorny, 1947), to the Eocene–Oligocene boundary (Hanzlikova in Mahel and Buday, eds., et al., 1968), and to the early Oligocene (Rzehak, 1922; Pokorny, 1960; Cicha et al., 1971; Hanzlikova, 1981; Krhovsky, 1981; Bubik, 1987; Jurasova, 1987; Krhovsky et al., 1992). The deposition of the Menilitic Formation in the Outer Carpathians falls into a critical period in Earth history, when the greenhouse climates of the Eocene were replaced by the icehouse conditions associated with a major extinction in the early Oligocene (Prothero et al., 2000). The onset of deposition of the Menilitic Cherts marks the beginning of the cooling period on the northern hemisphere (Krhovsky et al., 1992).

The organic-rich Menilitic Formation represents one of the most important source rocks for hydrocarbons in the entire Carpathian realm. In the Zdanice–Subsilesian unit, however, with the exception of the inner zones buried deeper below the Silesian and Magura nappes, the organic-rich strata of the Menilitic Formation remain immature.

The Menilitic Formation of the Zdanice unit is overlain by the as much as 1200-m (4000-ft)-thick Zdanice–Hustopece Formation (Chmelik and Matejka in Kalasek et al., 1963). It is a typical late orogenic Krosno-type turbiditic sequence of gray shales with laminae of siltstones and beds of sandstones and local conglomerates. The proportion of sands and shales varies both vertically and horizontally from predominantly shaly (Hustopece Marls; Rzehak, 1881) to predominantly sandy facies (Zdanice Sandstones; Paul, 1893) with a widespread transitional facies between these two end members.

The overall northwestward fining and thinning of turbiditic sequences, as well as the orientation of the paleocurrent markings, indicate that the Zdanice–Hustopece Formation was deposited in a system of northwestward-prograding and overlapping subsea fans. Occasional bodies of conglomerates, interpreted as debris flows and channel fills, are more commonly found in the proximal southeastern parts of these fans. In addition to abundant larger clasts of sandstones reworked from the Magura flysch, these conglomerates contain pebbles of crystalline rocks and Mesozoic shales and carbonates, most likely supplied from the tectonically activated (inverted) crustal blocks located at the front of the Magura thrust system. In composition, the conglomerates of the Zdanice–Hustopece Formation differ markedly from the Paleocene–Eocene conglomerates found in the pericordilleran Cejc–Zajeci unit. These older conglomerates were apparently sourced from the Silesian cordillera (Picha et al., 1966; Sotak, 1992).

Similarly, the monotonous garnet-dominated assemblages of heavy minerals of the upper Oligocene to lower Miocene Zdanice–Hustopece Formation, as well as of the Krepice Formation of the Pouzdrany unit, differ markedly from the complex garnet-zircon-tourmaline-staurolite assemblage of heavy minerals found in the older, Upper Cretaceous to lower Oligocene strata of the Zdanice and Pouzdrany units and in the autochthonous Paleogene (Krystek in Homola et al., 1961; Picha, 1963, 1965; Stranik et al., 1968) (Table 1).

As documented by differences in composition of heavy minerals and conglomerates (Picha et al., 1966), the primary restructuring of the Outer Carpathian depositional system in the late Oligocene also led to primary changes of the provenance of the clastic material. The older pre-late Oligocene strata were supplied from the foreland and the uplifted Silesian ridge, whereas the younger, late Oligocene flysch deposits were recycled mainly from the uplifted inner zones of the Flysch belt, particularly the Magura flysch.

In addition, most of the microfauna found in the Zdanice–Hustopece Formation was redeposited from the older Cretaceous and lower Paleogene flysch deposits, whereas the diagnostic indigenous species are scarce. The age determination of the formation, thus, to a great extent, depends on dating of the underlying and overlying strata and the overall tectonostratigraphic position of the formation. Considering all these aspects, the age of the Zdanice–Hustopece Formation was assigned to the late Oligocene to early Miocene (Egerian) (Cicha et al., 1964, 1975b).

The Zdanice–Hustopece Formation and its equivalent, the Krepice Formation of the Pouzdrany unit, are typical synorogenic flysch sequences deposited in a system of foredeeps, which formed by flexural downbending of the foreland plate in front of the northwesterly progressing nappe stack of the Magura flysch. In a broader sense, these formations belong to the late Oligocene to early Miocene Krosno-type synorogenic depositional system characteristic for the external units of the Outer Carpathian belt.

The deposition of the Zdanice–Hustopece formation in Chattian–Aquitanian (Egerian) marks the end of the continuous sedimentation in most of the Waschberg–Zdanice–Subsilesian zone, although in the Dyje–Thaya depression of southern Moravia and the Waschberg zone, the sedimentation continued into the Burdigalian (Eggenburgian, Ottnangian, and Karpatian). These strata, however, belong to a new molasse-type phase associated with a major marine transgression and formation of the Neogene foredeep in the foreland and the Vienna basin in the hinterland of the Outer Carpathian thrust belt in Moravia.

Table 1. Assemblages of heavy minerals in the autochthonous unit and in various formations of the Pouzdrany and Zdanice allochthonous units in southern Moravia (Krystek in Homola et al., 1961; Picha, 1963). Note the principal differences in composition of heavy-mineral assemblages between the pre-late Oligocene (Nesvacilka, Pouzdrany, and Submenilitic formations and Sitborice Member) and the late Oligocene and early Miocene (Krepice and Zdanice–Hustopece formations) (Picha and Stranik, 1999). All numbers are percentages.

	Autochthonous Unit (%)	**Pouzdrany Unit (%)**		**Zdanice Unit (%)**		
	Nesvacilka Formation Eocene–Lower Oligocene	**Pouzdrany Formation** Eocene–Lower Oligocene	**Krepice Formation** Upper Oligocene–Lower Miocene	**Submenilitic Formation** Upper Cretaceous–Upper Eocene	**Sitborice Member** Lower Oligocene	**Zdanice–Hustopece Formation** Upper Oligocene–Lower Miocene
Garnet	40.6	17.3	80.5	31.3	26.4	79.1
Zircon	30.7	23.1	3.1	33.3	25.1	4.0
Staurolite	4.9	25.2	3.8	4.1	13.3	6.7
Tourmaline	7.9	12.1	1.7	7.6	16.0	2.2
Rutile	6.5	10.7	4.9	8.9	8.9	2.7
Kyanite	2.3	7.3		1.7	4.7	1.2
Anatase	0.7			1.8		0.4
Apatite	5.5			10.1	3.9	1.4
Monazite	0.3			0.4		
Amphibole				0.2		
Other	0.9		1.5		1.7	

In the Zdanice unit, this new phase began with deposition of as much as 200 m (660 ft) of shallow-marine calcareous shales and marls, which were assigned to lower Burdigalian (Eggenburgian) and correlated with the Luzice Formation of the Vienna basin by Benesova et al. (1963) and Cicha and Picha (1964) and called Sakvice Marls by Cicha et al. (1975b) (Figure 17A, B). These strata are found mainly in an east–west-trending depression (Figure 3), which is generally identical with the extent of the Vranovice paleovalley buried underneath the Zdanice nappe. This coincidence may be explained by the preservation of the Sakvice Marls in a zone of a higher subsidence apparently related to a higher compaction rate of the underlying paleovalley fill. Another much smaller occurrence of the Sakvice Marls is located near the village of Kobyli, just in front of the Cejce–Zajeci thrust unit (Cicha et al., 1964). The Sakvice Marls, in some areas, e.g., at the type locality near Sakvice, seem to evolve from the underlying Zdanice–Hustopece Formation without interruption (Benesova et al., 1963); in other areas, however, the Sakvice Marls rest transgressively on the underlying older strata without any evident angular discontinuity. Elsewhere in Moravia, the Eggenburgian strata represent a transgressive sequence, which was laid down on the folded and thrusted Magura flysch in the Vienna basin area or on the newly submerged zone of the foreland as a basal sequence of the late orogenic and postorogenic molasse-type foredeep.

Locally, e.g., in the vicinity of the Zajeci railway station, the Eggenburgian Sakvice Marls pass vertically into a 300-m (1000-ft)-thick sequence of greenish and brownish gray claystones with lenses of silty Fe and Mn pelocarbonates followed by diatomites (Pokorny, 1961; Benesova et al., 1963; Cicha and Picha, 1964). Stranik (1983) named this sequence the Pavlovice Formation and assigned it to the Eggenburgian to Karpatian (Figure 17A, B). The Pavlovice Formation is overlain by gray calcareous silty shales with rich microfauna, which Pokorny (1961), Benesova et al. (1963), and Cicha and Picha (1964) attributed to the Karpatian. Stranik (1983) assigned these about 100-m (330-ft)-thick strata to the Laa Formation (including the Korneuburg Member) (Figure 17A, B), as known from the Neogene foredeep and the Vienna basin in Austria (Grill, 1953, 1962; Kapounek et al., 1960; Wessely, 1998).

During the last stages of the Carpathian thrusting in Moravia, at the end of the early Miocene, the lower Miocene (Eggenburgian to Karpatian) strata, together with the underlying strata of the Zdanice and Pouzdrany units, were gently folded and tectonically transported toward the foreland. The level of deformation of these lower Miocene deposits seems to be similar to that of the underlying relatively competent Zdanice–Hustopece Formation. The older, less competent strata of the Nemcice and Menilitic formations were deformed more intensely (Hrouda and Stranik, 1985; Stranik, 1999).

The Subsilesian Unit (Sector)

The Subsilesian unit of northern Moravia comprises strata ranging in age from the Late Cretaceous (Turonian–Maastrichtian) to the early Miocene (Egerian). They have been divided into four lithostratigraphic units: (1) the Frydek Formation, (2) the Frydlant Formation, (3) the Menilitic Formation, and (4) the Zdanice–Hustopece Formation (Figure 17A, B).

The oldest, Turonian to Paleocene Frydek Formation, about 500 m (1600 ft) thick, is made by a monotonous sequence of gray laminated silty shales and siltstones with a rich foraminiferal microfauna (Hanzlikova, 1969). Turbiditic sandstones and slump bodies of conglomerates with fragments of corals are found locally, especially in the upper part of the formation in the vicinity of Pribor, where they are known as the Klokocov Member. Elias (1998) interpreted these coarse deposits as a proximal facies of the lower slope environment, whereas he assigned the finer predominantly pelitic strata of the Frydek Formation to a deep-water basinal facies. According to the senior author of this article, the Frydek Formation may be interpreted as a distal facies of a major submarine fan system, whose more proximal facies are represented by the Istebna Formation of the Silesian unit. The Klokocov Member would then be a local facies related to a proximity of the locally uplifted ridge holding the Jurassic Stramberk reef complex (fragments of corals).

The overlying Frydlant Formation (Elias, 1998), also known as the Submenilitic Formation, is facially more diversified. Hanzlikova et al. (1963) distinguished three facies in this formation: (1) the anoxic facies (Guty Member of E. Mencik, M. Elias, I. Zurkova, F. Jurasova, and L. Rybarova, 1973, personal communication) of dark-gray shales, which prevailed in the late Paleocene; (2) the variegated facies (Roth, 1962) of gray, green, and red shales, which dominated in the late Eocene; and (3) the transitional facies of spotted gray, green, and red shales (spotted facies of Roth, 1962). Elias (1998) added a facies of thick sandstones and conglomerates (Straz Sandstones of Hanzlikova et al., 1955), which may be interpreted as channel and overbank deposits, similar to those found in the Cejc–Zajeci unit.

The Menilitic Formation of the Subsilesian unit, otherwise similar to that of the Zdanice unit, locally includes about 100-m (330-ft)-thick facies of slump conglomerates and debris flows. They contain clasts of crystalline basement rocks, Devonian and Carboniferous carbonates and clastics, as well as intraclasts of Upper Cretaceous to upper Eocene strata derived from the margins of the Subsilesian basin proper. The nannoplankton from the locality Dolni Tesice indicates the early Oligocene age of the Menilitic Formation (Jurasova, 1974). Thin intercalations of laminated shaly limestones (Jaslo Limestones, Uhlig, 1882, 1883b; Jucha, 1958) occur in the upper part of the Menilitic Formation near Bystrice nad Olsi (Nowak, 1965). Bubik (1987) compared them with the Oligocene Zagorza horizon (Haczewski, 1989) of the Polish Carpathians. The overall thickness of the Menilitic Formation in the Subsilesian unit may reach as much as 200 m (660 ft).

The Zdanice–Hustopece Formation (Elias, 1998), previously called the Krosno Formation, is coeval and lithologically similar to the Zdanice–Hustopece Formation of the Zdanice sector, but its extent and thickness are limited. The known thickness of this formation in the Subsilesian sector does not exceed several tens of meters.

Additional information about the stratigraphy and structure of the Subsilesian unit may be found in publications by Hanzlikova et al. (1953), Mencik and Pesl (1955), and Mencik and Hanzlikova (1983).

The Facies Variations along the Strike of the Waschberg–Zdanice–Subsilesian Units

The character of the Waschberg–Zdanice–Subsilesian nappe system significantly changes along the strike of the Carpathian thrust belt. The Upper Cretaceous strata are prominently present everywhere in the Subsilesian sector, whereas their extent in the Waschberg–Zdanice sector is limited. However, the late Oligocene to early Miocene Krosno-type flysch deposits are dominant in the Zdanice and partly in the Waschberg sectors and only marginally present in the Subsilesian sector of northern Moravia. The lower Miocene Sakvice, Pavlovice, and Laa formations, typical for the Waschberg–Zdanice sectors, have not been found in northern Moravia. Their presence seems to be limited to the confines of the Dyje–Thaya depression. The Pouzdrany marginal unit has been recognized only in the Zdanice sector, but it is apparently also present in the Waschberg zone of Austria (Stranik, 1996) but evidently absent in northern Moravia. Finally, the Jurassic and Cretaceous tectonic klippen is found only in the Waschberg zone.

These stratigraphic and facial differences obscured the correlation and led to the development of separate terminologies for the various sectors. The most apparent terminalogical differences exist between the Austrian and Czech parts of the Outer Carpathians (Figure 16). Whereas in Austria, the Waschberg zone is generally considered to be a part of Helveticum, the Zdanice unit, thanks to the dominant presence of the flysch-type Zdanice–Hustopece Formation, is referred to as a part of the Carpathian Flysch belt.

The three sectors also differ in their structural architecture. The southernmost Waschberg sector is characterized by a rather steep internal structure (Matejka in Buday et al., 1961). The Zdanice sector displays a typical wedge-shaped geometry, and the northern Subsilesian sector, as documented by numerous wells, has an extremely low-angle flat structural pattern (Figure 20, shown on page 118). This is apparently related to the distribution of compressional and transpressional stresses along the strike of the Western Carpathian belt (Nemcok et al., 1998a). The amount of shortening accommodated by the sinistral strike-slip motion is the highest in the north–south-trending relatively steep Waschberg sector. Toward the north, the amount of transpressional strike-slip motion decreases, and the shortening is accommodated mainly by the normal forward thrusting, resulting in a flat structural grain of the thrust system.

The Silesian Unit

The Silesian thrust unit, positioned between the more external Subsilesian and the more internal Magura units, is the stratigraphically most complete unit of the Outer Western Carpathians (Figures 3; 17A, B). Its more or less continuous stratigraphic section extends from the Late Jurassic to the late Oligocene and possibly even to the early Miocene. As a distinct depositional site, the Silesian basin formed during the Late Jurassic and the Early Cretaceous rifting and extension of the European plate. Jurassic carbonate platforms marked the northwestern margin of the basin, whereas the Silesian ridge bordered the Silesian basin on the south, thus separating it from the more internal Magura depositional domain (Figure 5). Carbonates, deep-water shales, silicites, and turbidites were deposited during the divergent stage of the basin, whereas thick turbiditic flysch sequences were laid down during the convergent stage. The Silesian ridge functioned as the main source of clastics especially during the convergent synorogenic phase of the evolution of the Silesian basin. Both organic-rich source rocks and good reservoirs are present in the stratigraphic records of the Silesian unit, which, especially on the territory of Poland, proved to be the most prolific unit of the Western Carpathian Flysch belt.

The Silesian unit is most prominently developed in the Beskydy region of northern Moravia and western

Poland, where it forms the highest mountain peaks (more than 1300 m [4300 ft]). The continuation of the Silesian unit to the southwest into southern Moravia and the Alpine domain is not clear. Lower Cretaceous strata, similar to those of the Silesian unit, have been described from the Zdounky unit in central Moravia (Matejka in Buday et al., 1963) and from the Hauptzone of the Wienerwald (Vienna Forest) in Austria. However, the typical Silesian unit, as it is known from northern Moravia and western Poland, apparently originally ended at the southwest–northeast-trending Western Carpathian transfer zone, which separated the Alpine region from the more attenuated and differentiated Carpathian realm (Figure 5).

Three stratigraphically and structurally distinct subunits (facies), Godula, Baska (Matejka and Roth, 1949a, 1955), and Kelc (Elias, 1970, 1979), have been recognized in the Silesian unit of Moravia. They differ in the stratigraphic extent and overall thickness of their depositional sequences as well as in the proportion of detrital material derived from the marginal carbonate platforms. Although the dominant Godula facies makes up the bulk of the Silesian unit in Moravia and continues on in the territory of Poland, the areal extent of the other two subunits is limited. The Baska subunit represents the local marginal facies of the Silesian unit adjacent to the Stramberk carbonate platform, and the relatively thin Kelc subunit occupies the southwestern end of the Silesian basin. Only the Godula facies extends over the entire stratigraphic range from the Upper Jurassic to Oligocene, whereas the stratigraphic sequences of the Baska and the Kelc facies lack most of the Paleogene section, which most likely was tectonically detached during the thrusting. The structurally complex frontal zone of the Silesian thrust system, in addition to elements of the Baska and Kelc subunits, also comprises tectonic slivers of the incompetent Lower Cretaceous strata of the Godula subunit as well as the tectonically incorporated slivers of the Subsilesian unit. The recognition of various elements of these units and subunits is not always easy.

The Godula Subunit

The Godula subunit has been divided into numerous formations and members (Figure 17A, B). The oldest known strata of the Godula subunit, originally known as the Lower Tesin (Teschen in German; Cieszyn in Polish) Member (Hohenegger, 1861) and recently renamed to the Vendryne Formation (Elias et al., 2003), are represented by dark-brown calcareous shales with occasional thin beds of siltstones and limestones. Slump conglomerates (debris flows) with clasts of limestones (Ropice horizon of Mencik et al., 1983) occur in the upper part of the formation. Vasicek (1972a, b) assigned the Vendryne Formation, about 350–600 m (1100–2000 ft) thick, to the Oxfordian to Tithonian and possibly to the Berriasian(?) age. The Vendryne Formation is thus coeval with the Klentnice Formation of the Jurassic klippen in southern Moravia. Like the Klentnice Formation, it also contains an increased content of organic matter and, as such, may be considered as a source rock for hydrocarbons.

The Vendryne Formation passes upward into the Tesin (Teschen) Limestone, which, mainly in its lower part, is composed of pelagic thinly bedded micritic limestones alternating with calcareous shales. The presence of the *Calpionella alpina* allocates these strata to the uppermost Tithonian to the early Berriasian (Hanzlikova and Roth, 1964). Laterally, these no more than 20–30-m (66–100-ft)-thick pelagic limestones pass into a facies of detrital limestones with occasional cherts, debris flows, grain flows, and intercalations of calcareous shales. The grains and larger clasts of these detrital limestones are made of carbonate rocks derived from the Stramberk carbonate platform. In a broad sense, the Tesin Limestones might be an equivalent of the Ernsbrunn Limestones of the Outer Klippen Belt in southern Moravia (Figure 17A, B).

The following Valanginian to early Aptian Tesin (Teschen)–Hradiste Formation (sensu Matejka and Roth, 1954) consists of three distinct lithologies: (1) the calcareous shales with subordinate turbiditic sandstones, limestones, and pelosiderites in the lower part [the Hohenegger's (1861) Upper Tesin (Teschen) Member]; (2) the thick packets (tens of meters) of coarse turbiditic sandstones and conglomerates separated by thin layers of claystones in the middle part (the Hradiste Sandstone of Hohenegger, 1861); and (3) the calcareous shales with abundant lenses and concretions of pelosiderites in the upper part (the lower part of the Verovice shales, sensu Hohenegger, 1861). The coarse sandstones and conglomerates of the Hradiste Sandstone are composed predominantly of quartz grains and fragments of the Stramberk Limestones, apparently supplied mainly from the uplifted Stramberk platform in the northwest. To simplify the complex historically evolved nomenclature, Elias et al. (2003) suggested the use of the term Hradiste Formation for the entire complex of the Tesin (Teschen)–Hradiste Formation. For the sake of consistency with the stratigraphy of the Baska and Kelc subunits and the compatibility with numerous previous publications, in our article, we adhere to the traditional term Tesin–Hradiste Formation, sensu Matejka and Roth (1954) (Figure 17A, B).

Characteristic for the Tesin (Teschen)–Hradiste Formation is the presence of dikes, veins, lavas, pillow lavas, and pyroclastic rocks of the teschenite rift-related submarine alkalic, calc-alkalic, and basic volcanism. Smid (1962) and Smid and Mencik in Mencik et al. (1983) distinguished three groups of volcanic rocks: picrites, teschenites, and monchiquites. Hovorka and Spisiak (1988) associated the teschenite volcanism with a short-term rifting of the continental crust. Dostal and Owen (1998) pointed to similarities of these rocks with basalts, basanites, and nephelinites derived from the upper mantle. The volcanic activity peaked during the deposition of the lower part of the Tesin (Teschen)–Hradiste Formation in the early Berriasian to Hauterivian time, although teschenite volcanic rocks are sporadically found also in the underlying Tesin Limestone and the Vendryne Formation.

The Tesin (Teschen)–Hradiste Formation was deposited during the phase of rifting, extension, and further diversification of the Outer Carpathian depositional system. From the Magura basin, the Silesian depositional site was separated by the Silesian ridge (cordillera), which emerged as a horst between these two basins. As documented by the increase of the thickness of the Tesin–Hradiste Formation, from several hundred meters in the northwest to more than 1000 m (3300 ft) in the southeast. Most of the clastic material into the Tesin–Hradiste Formation was apparently supplied from the Silesian ridge. Despite the intensive rifting and extension, the Silesian basin remained underlain by a thinned continental crust.

The stratigraphy and fauna of the Lower Cretaceous sequences of the Silesian unit were studied by Uhlig (1883a, 1902) and recently by Eliasova (1962a), Hanzlikova (1965a), Vasicek (1971b, 1973, 1979, 1999), Housa (1975, 1978), Vasicek et al. (1994), Skupien and Vasicek (2002), and Elias et al. (2003).

In the 19th century, these iron-rich pelosiderites of the upper part of the Tesin–Hradiste Formation provided a raw material for the nascent iron and steel industry of Silesia (Roth and Matejka, 1953).

The overlying Verovice Formation (sensu Matejka and Roth, 1949a), several tens of meters to as much as 250 m (820 ft) thick, is composed predominantly of black organic-rich cherty shales with rusty iron oxide coatings, locally interbedded with thin beds of quartzitic sandstones and concretions of pelosiderites. Given the total absence of diagnostic fossils, the assignment of the Verovice Formation to the Aptian age is based solely on the determination of the ages of the underlying and overlying strata. The depositional environment of the Verovice Formation was extremely anoxic, favorable for preservation of organic matter. The high content of organic matter makes the Verovice Formation a potential source rock for hydrocarbons; however, its role in the petroleum systems of the Carpathians has yet to be better understood.

The deposition of the Verovice Formation in the Silesian basin and the coeval Rajnochovice Formation (Gault Flysch) in the Magura flysch marks a period of maximum deepening of the Outer Carpathian basins and the submergence of all sources of detrital material, both in the European platform and in the nascent Silesian ridge. According to Slaczka et al. (1999), the downwarping of the Silesian basin was most likely caused by the cooling effect of the underlying lithosphere, previously dominated by the high heat-flow regime of the rift stage.

The Verovice Formation passes upward into the 100–380-m (330–1250-ft)-thick Lhoty Formation assigned to the Albian (e.g., Hanzlikova, 1966). In its lower part, it consists of bioturbated shales with subordinate thinly bedded distal turbiditic sandstones, whose proportion increases upward. The upper part of the Lhoty Formation is characterized by the predominance of turbiditic sandstones and the occurrence of cherts (Mikuszowice Cherts of Szajnocha, 1884). The Lhoty Formation was still deposited in a deep-water uncompensated environment although, as indicated by the presence of bioturbations, less anoxic than that of the underlying Verovice Formation. Even further, a less restricted environment prevailed during the deposition of the overlying Cenomanian pelagic and hemipelagic red and green shales of the Mazak Formation (sensu Roth, 1980a), also known as the Variegated Godula Member (Zahalka and Koutek, 1927; Matejka and Roth, 1949c). These shaly deposits are interbedded with thick coarse turbiditic sandstones and conglomerates of the Ostravice Sandstone (Andrusov, 1933). The deposition of the variegated shales of the Mazak Formation assigned to the Cenomanian age (Hanzlikova, 1973) thus marks the change from the deep anoxic conditions of the Verovice and Lhoty formations into the more dynamic environment of the deep-water turbiditic flysch facies. The Ostravice Sandstone, interpreted by Elias (1995) as a prograding turbiditic subsea fan, represents the oldest clearly defined synorogenic sequence of the Outer Carpathians in Moravia and actually marks the transition from the passive-margin environment into an active-margin depositional setting (Figure 17A, B).

The overlying Godula Formation of the Cenomanian(?)–Turonian–Santonian age is a typical flysch sequence of alternating sandstones and shales with a variable proportion of these two main lithological components. The lithological subdivision of the Godula Formation has been elaborated by Burtan et al. (1937), Matejka (1949, 1952), Mencik et al. (1983), Elias (2000), and others. The lower part of the formation is typically represented by a turbiditic facies of thinly bedded glauconitic sandstones and shales followed in the middle part of the formation by a facies dominated by coarse glauconitic sandstones and conglomerates. The upper part of the Godula Formation is again represented by a facies of thinly interbedded glauconitic sandstones and shales, although coarse sandstones and conglomerates, such as those of the Malinowska skala Sandstone (Burtan et al., 1937; Elias, 2000) and of the Pustevny Sandstone (Mencik et al., 1983), occur locally. The vertical and horizontal variability in proportion of sandstones and shales reflects on the dynamic depositional environment of the Godula Formation dominated by prograding and shifting subsea fans. The overall thickness of the Godula Formation ranges from more than 3000 m (10,000 ft) at the southern proximal side of the Silesian basin adjacent to Silesian ridge to only a few hundred meters in the distal northern side of the basin facing the platform. The Godula Formation thus appears as a depositional wedge, whose thick part formed in a zone of maximum subsidence at the front of the emerging Silesian ridge, which supplied most of the clastic material into the Godula Formation. The sudden rise of the Silesian ridge was apparently caused by compressional stresses associated with the accelerated subduction of the Penninic–Pieninic ocean and the early collision of the Inner Carpathians with the fragmented margins of Europe in the early Late Cretaceous. The deposition of the Godula Formation in the Silesian basin thus marks the beginning of the convergency and the concomitant formation of the foreland depositional regime in the Outer Carpathian realm.

In the Campanian–Maastrichtian, the foreland depositional system progressed further onto the European foreland, where the Waschberg–Zdanice–Subsilesian basin came into existence. In the Silesian basin, the Istebna Formation of the Campanian to Maastrichtian to Danian(?) age (Liebus and Uhlig, 1902; Hanzlikova, 1972b) evolved from the underlying Godula Formation. The presence of conglomerates and erosional surfaces at the boundary with the underlying Godula Formation was, by some geologists (Zahalka and Koutek, 1927), mistakenly taken as evidence of a transgressive character of the Istebna Formation. The 1000–1200-m (3300–4000-ft)-thick Istebna Formation is represented by a typical flysch facies of alternating sequences (70–200 m; 230–660 ft) of arkosic sandstones, slump conglomerates, and sand flows (fluxoturbidites) with equally thick sequences of dark shales. The common presence of pelocarbonate concretions, authigenic siderite, pyrite, and organic matter indicates that the depositional environment of the Istebna Formation was anoxic. An abundance of gravity-driven slide conglomerates and sandstones testifies to an increased tectonic activity along the slopes of the rising Silesian ridge. Further north away from the Silesian ridge, the proximal facies of the Istebna Formation passes into more distal turbiditic facies. Moreover, according to the senior author of this account, the Frydek Formation of the Subsilesian basin may represent the most distal facies of the Istebna Formation fan system. The highly dynamic depositional environment of the Istebna Formation was apparently related to the further rise of the Silesian ridge during the Laramide orogeny in the Late Cretaceous to the early Paleocene. As documented by the depositional record both in the Silesian and Magura basins (Figure 17A, B), the Laramide uplifting and inversion of crustal blocks in the Outer Carpathian basin and in the European foreland had a great impact on the evolution of the entire Carpathian region.

Sporadically found rhyolitic and andesitic tuffitic rocks in the Istebna Formation and also in the underlying Godula Formation indicate that the volcanic activity in the Silesian basin continued into the Turonian–Maastrichtian.

The overlying Paleocene to Eocene Roznov Formation (Elias, 2002), formerly known as the Submenilitic Formation, about 800 m (2600 ft) thick, is composed predominantly of hemipelagic variegated, red, green, and gray shales with subordinate thin sandstone beds. Encased into this monotonous sequence, thick packets (as much as 150 m [500 ft]) of coarse sands (Ciezkowice Sandstone) are present, which may be interpreted as subsea channel fills and levee deposits of the basin floor. Numerous oil fields in the Polish sector of the Silesian unit are reservoired in these channelized sands (Karnkowski and Ozimkowski, 1998; Dziadzio et al., 2006).

The upper part of the Roznov Formation is marked by the presence of the Sheshory Marls (previously known as *Globigerina* Marls), composed of laminated brown siltic marls alternating with greenish pelagic clays. They represent a transition into the lower Oligocene Menilitic Formation, composed of dark-gray and brown silicified shales, cherts, marls, and micritic limestones. Lithologically, the Menilitic Formation of the Silesian unit is comparable with its equivalents in the Waschberg–Zdanice–Subsilesian unit, including its subdivision into the Subchert and Chert members, the Dynow Marls, and the Sitborice Member. Locally, in the zone adjacent to the front of the Magura nappe, thick beds of greenish glauconitic sandstones are found in the Menilitic Formation. They are comparable with the Kliwa Sandstones of the Polish Carpathians, where they represent a significant oil-bearing reservoir. However, the association of these sandstones with either the Silesian or the Fore-Magura units in the structurally complex zone at the front of the Magura nappe remains uncertain. The organic-rich Menilitic Formation of the

Silesian unit, if properly buried, may represent an important source rock for hydrocarbons.

The highest strata of the Godula subunit are represented by the late Oligocene to early Miocene Krosno Formation (Tietze, 1889). It is a typical flysch facies of alternating turbiditic sandstones, shales, and occasional conglomerates. As in the Zdanice–Hustopece Formation of the Waschberg–Zdanice–Subsilesian unit, with which the Krosno Formation is comparable, the proportion of sandstones and shales varies both laterally and vertically from the predominantly shaly to the predominantly sandy facies. On the territory of Moravia, the Krosno Formation is more than 1000 m (3300 ft) thick.

The overall maximum thickness of the Godula subunit may reach 4000 m (13,000 ft) on its southern proximal side but would decrease significantly toward the more distal northern side of the Silesian unit. These variations in thickness are caused mainly by the wedge-shaped Godula and Istebna formations, whose fast subsiding depocenters were adjacent to the emerging Silesian cordillera in the south.

The Baska Subunit

Unlike the dominant and widespread Godula subunit, the Baska subunit is restricted to a relatively smaller area of the Palkovice Hills and the Stramberk klippen (Figure 3). It consists of several tectonic imbricates comprising tectonic klippen detached from the Tithonian to Valanginian carbonate platform and the younger Hauterivian to Paleocene strata characterized by the abundant presence of clastic material derived from this carbonate platform as well. The overall stratigraphic thickness of the Baska subunit does not exceed 1600–1800 m (5200–5900 ft) (Figure 17A, B). The carbonate platform formed on an elevated block of the rifted margins of the European plate in the Tithonian and possibly lasted into the Berriasian–Valanginian (Figure 5). Two main types of limestones, the whitish gray Stramberk Limestones of Tithonian age and the red-brown and green Koprivnice Limestones with the uppermost Tithonian to Valanginian fossils, have been recognized both in the tectonic klippen and the carbonate debris found in the Valanginian–Cenomanian Kotouc facies and the Chlebovice Member of the Tesin–Hradiste Formation (Figure 17A, B). The Stramberk Limestones, as much as 350 m (1100 ft) thick, represent the typical platform assemblage of grainstones, reefal framestones, boundstones, and detrital slope deposits. They display many similarities with the Ernsbrunn Limestones of the Outer Klippen in southern Moravia. The younger Koprivnice Limestones, made of olistoliths, brecciated limestones, and marls, are thought to be associated with the last stages of the platform growth, its exposure, and partial destruction. The large klippen of the Stramberk and Koprivnice limestones exposed in several large active and inactive quarries are well-known paleontological sites. The rich fauna was studied by Zittel (1868, 1873), Blaschke (1911), Spath (1933), Housa (1961, 1975, 1976, 1978), Eliasova (1962b), Zitt (1974), Nekvasilova (1977), and others. As summarized by Mencik et al. (1983), more than 600 species, among them more than 50 species of ammonites, have been described from these localities.

The origin of some large carbonate blocks, e.g., the Kotouc Hill in Stramberk (more than 1000 m [3300 ft] in diameter), became a subject of controversy. Matejka and Roth (1955), Eliasova (1962b), Housa (1976), and Mencik et al. (1983) interpreted the large carbonate blocks as tectonic klippen detached from the carbonate platform during thrusting. According to Elias and Stranik (1963), Elias (1979), and Elias and Eliasova (1984, 1986), the large bodies (klippen) in Stramberk formed by accretion of bigger and smaller blocks of Stramberk and Koprivnice limestones derived from the disintegrated platform and redeposited into the younger strata of the Kotouc facies and the Chlebovice Member. Neither of these interpretations can fully explain the character of these chaotic deposits. On one hand, it is difficult to accept that all the large bodies of the very clean limestones, up to several hundred meters in diameter, formed by the accretion of the carbonate detritus eroded from an older platform and redeposited in the younger strata, as proposed by Elias and Stranik (1963), Elias (1979), and others. However, it is also unlikely that all the carbonate blocks represent tectonic slivers with preserved stratigraphic succession, as advocated by Housa (1976). In our opinion, a more plausible interpretation may be found somewhere between these extreme views. Like any other carbonate platform, the Jurassic Stramberk platform apparently had a flat plateau dominated by the sedimentation of carbonate grainstones, elevated rims with coral reefs, and slopes composed predominantly of debris derived from the edge of the carbonate platform (e.g., Eliasova, 1962b). Gravity slides and turbidity currents transported smaller and bigger debris detached from the edges of the platform farther into the adjacent basin. The tectonic slivers of the Baska facies may thus comprise both tectonically detached pieces of the original carbonate platform, including the coeval slope and the more distal talus deposits, as well as debris derived from the exposed platform into the younger strata. During the tectonic transport, the rigid pieces of the carbonate platform may have been further separated from the softer, less competent rocks present in the slopes of the platform. This may have resulted in the formation of a melange in which the larger blocks are true tectonic fragments (klippen) of the original platform, whereas the smaller blocks and debris are part of the coeval platform talus or even clastic material redeposited from the emergent and disintegrated platform into some younger strata.

The carbonate debris from the Stramberk and Koprivnice limestones are prominently distributed in the Valanginian to Albian strata of the Tesin–Hradiste Formation, i.e., in the lower part of the Kotouc facies and the Chlebovice Member, which represents the upper (Albian to Cenomanian) part of the Tesin–Hradiste Formation. In the Baska subunit, the Tesin–Hradiste Formation is only 150–500 m (500–1600 ft) thick as compared to the 500–1200-m (1600–4000-ft)-thick equivalent of this formation in the Godula subunit. To a lesser degree, the material derived from the Jurassic to Lower Cretaceous carbonate platform is also present in the Baska (Albian–Cenomanian) and the Palkovice (Coniacian–Danian) flysch formations. The Baska Formation, 250–350 m (800–1100 ft) thick, consists of sandstones, allodapic limestones with spongilitic cherts, and greenish gray shales. The Palkovice Formation, about 500 m (1600 ft) thick, is the youngest known stratigraphic member of the Baska unit. It is composed of black shales alternating with thick-bedded sandstones and conglomerates containing pebbles, cobbles, and blocks of the Stramberk Limestones. The presence of clasts of Stramberk Limestones in the Baska and Palkovice formations indicates that the Stramberk carbonate platform located at the western margin of the Silesian basin remained at least partly exposed until the Late Cretaceous. At that time, the remnants of the platform were gradually inundated by the sea and possibly became a part of the Subsilesian depositional system.

According to Michalik and Sotak (1990), an erosional event at the Jurassic–Cretaceous transition apparently partly destroyed the Jurassic carbonate buildups, including the platforms in the Waschberg sector and the Stramberk platform, which formed on the elevated blocks of the rifted Outer Carpathian depositional realm (Figure 5A, B). Abundant clasts from the Jurassic platforms are found elsewhere in the Cretaceous and Paleogene conglomerates of the Waschberg–Zdanice–Subsilesian unit and the Magura flysch (e.g., Picha et al., 1966; Elias, 1998).

Elias (1998) situated the original Tithonian to Valanginian Stramberk platform on a hypothetical Baska ridge, allegedly situated between the Silesian and Subsilesian basins. Because the Subsilesian basin in Moravia came into existence much later, in the Turonian, it seems to be more prudent to locate the carbonate platform more generally on the northern rifted margin of the Silesian basin without implying an existence of the Baska ridge. The similarly positioned Andrychow ridge between the Silesian and Subsilesian–Skole units in the Polish sector of the Western Carpathians (Ksiazkiewicz, 1960) apparently emerged in the Aptian (Golonka et al., 2000; Slaczka et al., 2006).

In the Stramberk area, the Baska subunit is thrusted over the Subsilesian unit and again over the autochthonous strata of the Neogene foredeep and the Hercynian basement. In the vicinity of Stramberk, more than 100 wells penetrated the Baska subunit and the Subsilesian unit of the Carpathian belt and drilled into the subthrust Neogene foredeep, in which some clastic reservoirs were developed into a gas storage.

The Kelc Subunit

The Kelc subunit, as defined by Elias (1970, 1979), represents the southwestern marginal part of the Silesian basin. It comprises strata of the Valanginian to Paleocene age subdivided into several formations (Figure 17A, B). The Valanginian to Aptian (Tesin–Hradiste and Verovice formations) are identical with their stratigraphic equivalents in the Godula subunit. The Albian Jasenice Formation, corresponding to the Lhoty formations in the Godula subunit, is characterized by the predominance of gray and green spotted shales with subordinate turbiditic sandstones and micritic limestones. Its thickness ranges between 100 and 200 m (330 and 660 ft). The overlying Nemetice Formation of Albian to Cenomanian age is composed of hemipelagic variegated (gray, green, and red) shales with occasional thin beds of sandstones. It is followed by the Cenomanian to Santonian(?) Dub Formation (Hanzlikova and Matejka, 1958), represented by a proximal facies of coarse- to fine-grained sandstones, calcareous sandy shales, and slump bodies of conglomerates with blocks of the Stramberk Limestones. Lithologically similar to the Dub Formation are the Turonian to lower Senonian deposits at Stary Jicin (Stranik et al., 1997) and also the younger Campanian to Paleocene Milotice Formation (Elias, 1979) and the Kojetin Formation mentioned by Mencik et al. (1983). They consist of gray and greenish calcareous sandy shales, calcareous sandstones, and slump conglomerates. The thickness of these various lithologic units is about 300 m (1000 ft).

The Kelc subunit, typically only less than 1000 m (3300 ft) thick, may be interpreted as marginal facies of the Silesian unit formed in a zone of limited subsidence and limited impact of the Stramberk carbonate platform. Its lower members bear similarities with the Lower Cretaceous strata of the Zdounky unit. The areal extent and the structural position of the Kelc subunit remain uncertain. Some doubts even exist about the recognition of the Kelc subunit as a separate entity in the Silesian unit.

The External Units Occurring in Front and Below the Magura Nappe

This is a widely defined group of tectonostratigraphic units, into which we include various minor tectonic thrust sheets and tectonic slivers, such as the Cejc–Zajeci unit, the Zdounky unit, and the Fore-Magura unit, typically distributed along the frontal edges of the Magura nappe in Moravia, as well as the units exposed in tectonic slivers and windows (window units) in the Magura nappe in Moravia, Slovakia, and Poland. Stratigraphically, these units bear many similarities with the adjacent external units of the Flysch belt, e.g., the Waschberg–Zdanice–Subsilesian and Silesian units, but differ facially. Some of these units contain a higher proportion of coarser clastics, including slump conglomerates and olistoliths, which indicate a depositional proximity to a tectonically active source. They are interpreted as being deposited on the northern and western side of the Silesian cordillera (Figure 5). During thrusting, the various strata of these units, including some older Jurassic and Early Cretaceous carbonate members, were detached and tectonically transported at the edge of the Magura frontal nappe or tectonically piled below the Magura nappe as duplexes and occasionally exposed in tectonic windows. The surface appearances of some of these units in Moravia are associated with the late orogenic transpressional faults. Although small in the areal extent, these units provide important information about the character of the Outer Carpathian depositional system as well as about the range of overthrusting of the out-of-sequence Magura nappe over the external domains of the Flysch belt.

The Cejc–Zajeci Unit

The Cejc–Zajeci unit is exposed along the eastern margin of the Zdanice unit and western edge of the Magura nappe in southern Moravia (Figure 3). From the Zdanice unit, into which it was originally included, the Cejc–Zajeci unit differs by a wider presence of the Upper Cretaceous strata (Campanian–Maastrichtian), by the absence of typical Menilitic cherts, and by the abundant occurrence of thick, discontinuous bodies (several meters to tens of meters) of coarse sandstone and conglomerates in the Nemcice (Submenilitic) Formation (Picha et al., 1968). The sandstones contain a high proportion of the organodetritic material, mainly fragments of lithothamnia. The discontinuous bodies of coarse clastics are interpreted as either submarine slumps or submarine channel fills and overbank deposits distributed at the mouth of submarine canyons. Their distribution, as well as the northwest-directed paleocurrent markings, suggests that these coarser clastics were supplied from the Silesian ridge, which separated the outer Waschberg–Zdanice–Subsilesian and Silesian basins from the inner Magura depositional system (Figure 5C, D). The conglomerates thus provide information about the composition and the geological history of the Silesian ridge (Picha et al., 1966). As a source of clastics for both the Magura unit and the external units, the Silesian ridge has been active since the Late Cretaceous. The influx of the coarse clastic material into the Cejc–Zajeci unit peaked in the Paleocene to middle Eocene and then gradually decreased and, in the late Eocene, when the Silesian cordillera submerged, ceased entirely. The channelized sandstones of the Cejc–Zajeci unit bear similarities with the Paleogene Ciezkowice Sandstones of the Silesian unit of northern Moravia and western Poland.

The structural position of the Cejc–Zajeci unit was established by the deep well Kobyli-1 (Appendix 1; Figure 20, shown on page 118, section DD′). At the depth of 702 m (2303 ft), the Cejc–Zajeci unit is thrusted over two stacked thrust sheets of the Zdanice unit, which, at the depth of 3135 m (10,285 ft), are thrusted over the autochthonous Jurassic strata of the foreland plate. Incorporated into the base of the lowest thrust sheet is a tectonic sliver (18 m; 59 ft) of the lower Miocene strata tentatively compared with the Krepice Formation of the Pouzdrany unit (Picha et al., 1971). On the southeastern side, the Cejc–Zajeci unit is covered by the Miocene strata of the Vienna basin (Figure 3). Small erosional remnants of the Magura flysch sandstones found elsewhere on the top of the Cejc–Zajeci unit indicate that the Magura nappe was originally thrusted over the Cejc–Zajeci unit. The contact between these two units was further modified by the strike-slip faulting related to the opening of the pull-apart Vienna basin in the middle Miocene.

Picha and Hanzlikova (1965) described loose blocks of Jurassic radiolaritic limestones and marls at two localities near Pritluky and Zajeci and interpreted them as weathered tectonic klippen incorporated into the frontal zone of the Cejc–Zajeci unit. Z. Stranik et al. (1982, personal communication) reinterpreted these occurrences of Jurassic rocks as olistoliths. They are comparable with other small Jurassic klippen, such as Cetechovice and Lukovecek, found in the Solan

Formation at the edges of major thrust units of the Magura nappe and recently interpreted as olistoliths.

We understand the Cejc–Zajeci unit as a marginal pericordilleran facies of the Waschberg–Zdanice–Subsilesian depositional system, adjacent to the southern end of the Silesian ridge, which separated this external zone from the Magura depositional realm and shed coarse clastics to both sides (Figure 6C). Cicha et al. (1964) tentatively correlated the Cejc–Zajeci unit with the Zdounky unit in central Moravia.

The Zdounky Unit

Tectonic slivers of the Zdounky unit are distributed along the northwestern edge of the Magura nappe in the Chriby Mountains of central Moravia (Figure 3). The largest surface exposure of the unit has been mapped near the village Zdounky; other smaller appearances have been found at Rostin, Cetechovice, and Bohuslavice near Kyjov. Small slivers of the Zdounky unit are also found incorporated into the frontal zone of the Magura nappe at Korycany and Stare Hute in the Chriby Mountains (Mencik and Pesl, 1958). The distinct lithological character of the Zdounky unit was recognized by Paul and Tausch (1899) and Dreger (1899), and further established by Uhlig (1903), Petraschek (1907), Pesl and Mencik (1956), Hanzlikova and Matejka (1962), Cicha et al. (1964), and Chmelik (1971).

The known strata of the Zdounky unit have been divided into the lower sequence (Lower Cretaceous–lower Oligocene) and the upper sequence (upper Oligocene) (Figure 17A, B). The lower sequence has sandstones and conglomerates [Berriasian–Barremian(?)] at the base, overlain by grayish calcareous shales and limestones. This is further followed by variegated shales with beds of organodetritic sandstones, in which Hanzlikova (in Chmelik, 1971) distinguished Aptian and Albian benthonic foraminifera. Greenish gray calcareous shales with foraminifera and radiolaria represent the Cenomanian. They are overlain by gray, green, and red shales with intercalations of organodetritic sandstones extending into the lower Oligocene. The strata of the lower sequence are strongly deformed; their depositional thickness would not exceed a few hundred meters. The upper sequence, 500–600 m (1600–2000 ft) thick, is represented by the Krosno facies of alternating proximal turbiditic sandstones with shales and occasional slump conglomerates that contain redeposited Eocene corals and echinoderms (Oppenheim, 1913).

The Berriasian to Cenomanian strata of the Zdounky unit display many similarities with the coeval strata of the Baska and Kelc facies of the Silesian unit; the Senonian to lower Oligocene strata, marked by the absence of the Menilitic Formation, are comparable with the Cejc–Zajeci unit of southern Moravia. Like in the Cejc–Zajeci unit, the presence of coarse organodetritic and siliciclastic sandstones and slump conglomerates indicates that the depositional site of the Zdounky unit was adjacent to the Silesian cordillera, which supplied the coarse clastics.

The tectonic slivers of the Zdounky unit in the Chriby Mountains are distributed along, and apparently structurally associated with, the late orogenic strike-slip fault, which outlines the frontal zone of the Magura nappe in southern Moravia and in the Chriby Mountains (Figure 3). This southwest–northeast-trending orogen-parallel fault is apparently a component of the Western Carpathian transfer zone. In the middle to late Miocene, the sinistral strike-slip motion in the transfer zone led to the opening of the pull-apart Vienna basin in northeastern Austria and southern Moravia and to the northeastern escape of the Western Carpathians.

The Fore-Magura Unit

The Fore-Magura unit appears as a narrow discontinuous band of tectonic slivers sandwiched between the Magura and the Silesian units in northern Moravia and western Poland (Figure 3). Its existence in the territory of Moravia was suggested by Burtan et al. (1937) and further confirmed by Hanzlikova et al. (1962), Pesl et al. (1964), Pesl (1967), Mencik (1973), Plicka (1978), Mencik et al. (1983), and Pesl and Hanzlikova (1983). The presence of the Fore-Magura unit below the Magura nappe was confirmed by the deep wells Jablunka-1 and Gottwaldov-2 (Appendix 1; Figure 20, section CC′ shown on page 118).

The known stratigraphic sequence of the Fore-Magura unit extends from the Campanian to the Oligocene (Figure 17A, B). The oldest Submenilitic Formation (Campanian to lower Oligocene), several hundred meters thick, is represented by variegated shales with sparse thin beds of sandstones. Thick bodies of sandstones and conglomerates in this formation are known from the vicinity of Bystrice pod Hostynem. The overlying Oligocene Menilitic Formation, several tens of meters to as much as 150 m (500 ft) thick, in its lower part consists of siliceous shales with fish scales, Dynow Marls with laminae of cherts, and discontinuous bodies of sandstones comparable with the Kliwa Sandstones of the Flysch belt in Poland, Ukraine, and Romania. The upper part of the formation, made by cyclical green and brown shales and subordinate sandstones, resembles the Sitborice Member of the Zdanice unit. The youngest Krosno Formation (500 m; 1600 ft) of the

Fore-Magura unit is characterized by an alternation of facies with a variable proportion of sandstones and shales. The predominantly sandstone facies with slump bodies of pebbly mudstones was named the Chvalcov beds by Pesl and Hanzlikova (1983). It may be correlated with the Pochodzita Sandstones (Burtan et al., 1937) of the Polish Carpathians.

The deposits of the Fore-Magura unit show many similarities with the coeval strata of the Waschberg–Zdanice–Subsilesian and Silesian units. However, the presence of discontinuous, apparently channelized, sandstones and conglomerates in the Submenilitic Formation indicates that the depositional site of the Fore-Magura unit was situated closer to the Silesian ridge, which remained the source of coarse clastics until the middle to the late Eocene (Figure 5).

Units of the Tectonic Windows

Eleven tectonic windows in the Magura flysch nappe have been recognized in the territory of Poland and Eastern Slovakia. In Moravia, only one tectonic window is present in the Magura nappe near Rajnochovice in the Hostyn Hills, but equivalents of the window units are known as minor thrust sheets and tectonic slivers in the Hostyn Hills and the Chriby and Beskydy mountains of Moravia.

In those windows, the Paleogene strata of the underlying window units (Sikora, 1970), such as the Obidowa Slopnice and Grybow (inclusive Smilno) units, were brought to the surface during the compressional thrusting in the early Miocene. For more information, see Slaczka et al. (2006). The known Eocene to Oligocene strata of these window units resemble the coeval strata of the external Silesian or Fore-Magura units. The existence of the window units below the Magura nappes indicates that in the Eocene and early Oligocene, the depositional realm of the external units extended far south, possibly beyond the present site of the Pieniny Klippen Belt (Figure 5C, D).

During compressional thrusting, the rigid Magura flysch nappes moving northward greatly distorted the incompetent, mostly shaly strata of these external domains by detaching and transporting some of them at their fronts and piling the others in tectonic duplexes at their bottom. The simultaneous movement of the crustal blocks underneath the thin-skinned structure apparently further complicated this tectonic process. The soft, external units thus served as a lubricant between the upper thin-skinned and the lower basement, involving structural levels of the Outer Carpathian thrust belt.

The existence of the low-gravity zone at the front of the Pieniny Klippen Belt might then be explained by the accumulation of light rocks of these external units below the Magura nappes in this zone (Figure 20, shown on page 118). Such an interpretation is supported by the results of the Oravska Polhora-FPJ-1 well in northwestern Slovakia (M. Zakovic, D. Bodis, M. Fendek, M. Potfaj, G. Gebauer, and J. Balint, 1989, personal communication; Potfaj, 2003), which, below the Magura nappe at the depth of 1323 m (4340 ft) to the final depth of 2417 m (7929 ft), penetrated the Obidowa–Slopnice or Grybow unit (Potfaj, 2003).

The Magura Group of Nappes

The Magura group of nappes is the dominant tectonostratigraphic unit of the Outer Western Carpathians. It forms a continuous belt along the Western Carpathian arc from the Vienna Forest in Austria to the Western Ukraine. Its southwestern part in northeastern Austria and southern Moravia is partly buried below the Neogene strata of the Vienna basin (Figure 3). The Rhenodanubian Flysch of the Eastern Alps is considered to be an equivalent of the Magura flysch in the Alpine region (Elias et al., 1990) (Figure 16).

Most of the known depositional sequences of the Magura flysch evolved during the Late Cretaceous and Paleogene convergent stage of the Tethyan–Alpine cycle. The Jurassic and Lower Cretaceous deposits of the passive continental margins are known only from the tectonic klippen Kurovice and Hluk and olistoliths (originally described as klippen) Cetechovice (Neumayer, 1870; Neumann, 1907; Chmelik, 1957) and Lukovecek (Uhlig, 1903; Rzehak, 1904; Oppenheimer, 1913; Rakus, 1987). They are represented by deep-water carbonates and marls, as well as by some flyschlike turbiditic deposits. The sparse occurrence of the older Jurassic and Lower Cretaceous deposits in the Magura nappe, however, may not be fully indicative of the actual distribution of these strata in their original depositional setting. The lower part of the Magura depositional sequence, deposited on an uneven substratum of a rifted basin, might have been decoupled from the upper part by the sole thrust (decollement) and left behind as an autochthon and possibly partly subducted with its substratum (Figure 6). Thus, only the upper part of the Magura depositional sequence, made predominantly by the Late Cretaceous to the early Oligocene synorogenic flysch deposits was integrated into the present Magura nappes. These synorogenic deposits are represented by deep-water turbiditic facies consisting predominantly of hemipelagic muds and a variety of gravity-driven deposits, including proximal and distal turbidites, debris flows, and occasional olistoliths. The proportion of shales and sandstones varies both vertically and laterally, reflecting on the dynamic and complex depositional environment of the subsea fans. As such, the deposits of the Magura unit may be considered as one of the most typical flysch facies in the entire Western Carpathian belt. The overall architecture of the Magura basin was constrained by the existence of the Silesian ridge on its northwestern side and the Czorsztyn ridge on its southeastern side (Figure 5). The Magura sedimentary basin thus consisted of a variety of depositional environments from the steep slopes of the Silesian and Czorsztyn ridges to the deep-water environment in the axial part of the basin (Figure 6). The Silesian ridge and, to a lesser extent, the Czorsztyn ridge supplied most of the coarse clastic material into the Magura basin.

Structurally, the Magura unit consists of numerous thrust sheets and imbricates. Because of a relatively competent stratigraphic section with a high proportion of sandstones, the internal structure of the individual thrust sheets is relatively simpler than that of the most external units of the Flysch belt.

Based on significant lithostratigraphic and structural differences, Matejka and Roth (1950) subdivided the complex Magura flysch in Moravia into three major units: the Raca, Bystrica, and Bile Karpaty–Krynica (Oravska Magura) units (Figures 3; 17A, B). The Bile Karpaty unit of southeastern Moravia is, on the surface, separated from the Krynica unit of northwestern Slovakia by a wide gap. However, an existence of some connection between these two segments underneath the Pieniny Klippen Belt cannot be excluded. According to Mencik (1969), the transverse segmentation of the Magura flysch units might be related to deep northwest–southeast-trending faults in the underlying subthrust platform.

The Raca Unit

The Raca unit is the most external and the most widespread unit of the Magura nappe system. It appears as a continuous belt along the entire length of the Western Carpathian Flysch belt (Figure 3). The original depositional site of this unit was on the southeastern side of the Silesian ridge, which separated the Magura flysch basin from the depositional realm of the external units (Figure 5A, B). The Raca unit comprises strata ranging in age from the Late Jurassic to the Oligocene (Figure 17A, B). Based on the lithological differences, Pesl and Krystek (1965) distinguished in the Raca unit six lithofacial zones: Hostyn, Tri kameny, Trnava–Staskov, Vsetin, Luhacovice, and Kycera.

The oldest known strata of the Raca subunit are found in the Kurovice tectonic klippe located at the frontal edge of the Magura nappe (Figure 3). They are represented by the deep-water Kurovice Limestones and the overlying Tlumacov Marls discordantly overlapped by conglomerates and breccias (Elias et al., 1996) (Figure 17A, B). The Kurovice Limestones (Glocker, 1840), as much as 150 m (500 ft) thick, are composed of medium- to thick-bedded and fine-grained limestones with sparse lenses of cherts. The common presence of graded bedding (in some beds, the fine-grained limestones even pass into marls) would indicate a deposition from the gravity-driven turbidity currents. Based on studies of foraminifera (Benesova et al., 1962, 1968) and the aptychi and calpionellids (Vasicek and Rehakova, 1994), the age of the Kurovice Limestones was established in a wide range from the Oxfordian to Tithonian [Berriasian(?)]. The Kurovice Limestones pass upward into the about 60-m (200-ft)-thick Tlumacov Marls (Elias and Eliasova, 1985), represented by flysch-type alternation of thinly to medium-bedded biomicritic limestones and marls. Based on the evaluation of aptychi and calpionellids (Vasicek and Rehakova, 1994) and nannofossils (Svabenicka et al., 1997), the Tlumacov Marls were dated as Berriasian to Valanginian. The conglomerates and breccias with intercalations of variegated shales, which overlap the Kurovice Limestones and Tlumacov Marls, are interpreted as debris flows and assigned to the Senonian (Svabenicka et al., 1997).

The oldest known flysch strata of the Raca unit, so far identified only in tectonic slivers at the fault plane along which the Tri kameny lithofacies zone is thrusted over the frontal Hostyn zone, are the black shales alternating with siliceous sandstones. These strata have been assigned to a wide stratigraphic range from the Hauterivian to the Cenomanian (Svabenicka et al., 1997) and called the Gault flysch, referring thus to the lithological and stratigraphical resemblance of these strata with the Lower Cretaceous Gault flysch of the Rhenodanubian Flysch belt in Eastern Alps. Stranik et al. (2004, personal communication) renamed these strata the Rajnochovice Formation after the village Rajnochovice in the Hostyn Hills, where the type locality of this formation is located in the banks of the Juhyna River.

The continuous succession of the Raca unit begins with the Kaumberg Formation, named after the type locality in the Vienna Forest (Wiennerwald) in Austria (Gotzinger, 1954). Formerly, these strata were denoted as the lower variegated beds (Matejka and Roth, 1949b). In the Polish Carpathians, the Kaumberg Formation corresponds to the Upper Cretaceous variegated beds, known also as the Malinowa Formation. The Kaumberg Formation is composed predominantly of red and green hemipelagic to pelagic shales with intercalations of siltstones and fine-grained sandstones. Typically, the

Kaumberg Formation, about 300 m (1000 ft) thick, is found at the base of individual thrust sheets in the Raca unit (Svabenicka et al., 1997). Its age ranges from the Cenomanian to the Campanian–Maastrichtian transition. The lower part of the formation is comparable with the variegated beds of the Mazak Formation of the Silesian unit (Figure 17A, B). The stratigraphic position of the Kaumberg Formation was well established in the deep well Jarosov-1 (Figure 7; Appendix 1), which penetrated 5540 m (18,175 ft) of the Raca unit and was thus instrumental in defining the stratigraphic succession of the Magura flysch especially at the deeper Cretaceous level (Hanzlikova, 1976).

The Solan Formation (Matejka and Roth, 1949b) of the Maastrichtian to Paleocene age represents a typical flysch sequence with a variable proportion of sandstones and shales. Based primarily on lithology, the Solan Formation has been subdivided into the Raztoka, Hostyn, and the Lukov members (V. Pesl, V. Cekan, J. Kolejka, M. Ruzicka, L. Rybarova, and V. Volsan, 1984, personal communication). The Raztoka Member of alternating shales and sandstones with abundant biotite, about 1200 m (4000 ft) thick, occurs typically in the Tri kameny lithofacial zone of the Raca unit. Within this member at the locality Ustgrun, the Cretaceous–Paleogene boundary was identified for the first time in the Carpathian Flysch belt (Svabenicka et al., 1997). The Hostyn Member, about 800 m (2600 ft) thick, is the stratigraphic equivalent of the Raztoka Member in the Hostyn lithofacial zone. It is made predominantly by thick-bedded calcareous sandstones with fragments of lithothamnia and nummulites. The uppermost Lukov Member, 200–800 m (660–2600 ft) thick, is composed of thick beds of sandstones and conglomerates interpreted as sand flows and debris flows of the proximal part of a subsea fan (Elias, 1963). The weathering-resistant coarse sandstones and conglomerates of the Lukov Member typically stand out as morphological ridges, which mark the trend of individual thrust sheets.

The general upward coarsening of the Solan Formation possibly reflects on a gradual progression of proximal turbiditic facies over the more distal parts of the deep-water turbidite fan. The overall thinning and fining of the Solan Formation toward the southeast, as well as the orientation of the paleocurrent markings and the distribution of slump conglomerates, indicate that the Solan Formation was sourced mainly from the Silesian cordillera located on the northern side of the Magura basin (Pesl and Krystek, 1965; Pesl, 1968) (Figure 6). The sandstones and conglomerates of the Solan Formation contain a high proportion of grains and pebbles of Jurassic limestones and fragments of lithothamnia apparently derived from the cordillera and its shallow shelf. In composition, the sandstones and conglomerates of the Solan Formation show many similarities with the channelized sandstones and conglomerates of the Istebna Formation of the Silesian unit deposited on the other (northern) side of the Silesian ridge.

The overlying (about 300-m [1000-ft]-thick) Beloveza Formation (Paul, 1869) was assigned to the Paleocene to middle Eocene (Hanzlikova in Matejka and Roth, 1956). It is composed predominantly of green and red shales with thin beds of sandstones. Occasional packets of arcosic sandstones, up to several tens of meters thick, may be interpreted as submarine channel fills similar to those found in the Paleocene to Eocene strata of the Silesian unit (Ciezkowice Sandstones) and in the Cejc–Zajeci unit both deposited on the opposite northwestern side of the Silesian ridge.

The youngest Zlin Formation (Zapletal, 1937) of the middle to late Eocene and possibly to the early Oligocene age is as much as 2500 m (8200 ft) thick in the Raca unit. Its lower part, overlying the Beloveza Formation, has been subdivided into several lithologically defined members, such as Krive, Rusava, Ujezd, Luhacovice, Vsetin, and Kycera (Pesl, 1968). The Krive Member, exposed at the front of the Magura nappe in the Valaske Mezirici area, is characterized by the predominance of sandstones with a significant biodetritic component. The Rusava Member, about 500 m (1600 ft) thick, is limited to the front of the Raca unit in the Hostyn Hills; the Ujezd and Luhacovice members, 500–700 m (1600–2300 ft) thick, are typically present in the more internal zones of the Raca unit. All these members, characterized by the presence of coarse sandstones and conglomerates, were laid down in the proximal part of the turbiditic fans close to the mouth of submarine canyons. The upper part of the Zlin Formation is represented by the Vsetin Member of the middle Eocene to the early Oligocene age. It is composed predominantly of calcareous shales and marls alternating with subordinate beds of fine- to medium-grained glauconitic sandstones, locally as much as 10 m (33 ft) thick. Typically, the Vsetin Member overlies the previously mentioned lower members of the Zlin Formation dominated by sandstones and conglomerates, but locally, it rests directly on the Beloveza Formation. The Vsetin Member represents a typical foredeep facies supplied predominantly from the uplifted, more internal zones of the Magura flysch basin rather than from the Silesian cordillera, which, in the late Eocene, was already submerged. The paleocurrent measurements in the Vsetin Member indicate a longitudinal transport (Pesl and Krystek, 1965). The Vsetin Member, as much as 2000 m (6600 ft) thick, is the most typical stratigraphic unit of the Zlin Formation and of the entire Magura unit. Its equivalent in the innermost part of the Raca unit is the Kycera Member, characterized by thick-bedded, muscovite-rich sandstones alternating with mostly calcareous shales.

The Bystrica Unit

The distinct character of the Bystrica unit was recognized by Kodym (1923), who called it the Bylnice nappe. The name Bystrica unit was introduced by Matejka and Roth (1949b). The Bystrica unit forms a relatively narrow belt between the Raca and the Bile Karpaty–Krynica (Oravska Magura) units along most of the Western Carpathian arc. Only in the Kysuca area, where the Bile Karpaty unit is missing, is the Bystrica unit in direct contact with the Pieniny Klippen Belt (Figure 3). It consists of several thrust sheets, which comprise the Solan, Beloveza, and Bystrica formations only (Figure 17A, B). It cannot be excluded, however, that some older members of the Bystrica unit not exposed on the surface might be present at a greater depth. The lowermost Solan Formation, whose known rudimentary thickness does not exceed 400 m (1300 ft), consists of coarse-grained arcosic sandstones (Matejka and Roth, 1949b). Vujta et al. (1991), however, included these presumably Solan sandstones into the overlying Beloveza Formation. The Beloveza Formation, more than 300 m (1000 ft) thick, of the late Paleocene to early Eocene age (Hanzlikova in Matejka and Roth, 1956) is composed of green and red shales with subordinate thin-bedded sandstones. The highest Bystrica Formation, as much as 1600 m (5200 ft) thick, is comparable with the Vsetin Member of the Zlin Formation in the Raca unit. It is represented by turbiditic sequences of calcareous shales, marls, and subordinate sandstones with nummulites. The individual turbiditic sequences commonly begin with glauconitic sandstones, which pass into laminated siltstones, calcareous shales, marls, and thin layers of pelagic shales. Characteristic for the Bystrica Formation and the Bystrica unit, in general, are the thick intervals of marls (up to several meters) known as the Lacko Marls in Poland. Based on foraminifera and nannofossils, the Bystrica Formation has been assigned to the Eocene (Hanzlikova, 1955; Bubik and Svabenicka in Vujta et al., 1991).

The Bile Karpaty–Krynica (Oravska Magura) Units

The Bile Karpaty–Krynica (Oravska Magura) units represent the innermost zone of the Magura flysch nappe system adjacent to the Pieniny Klippen Belt. It has two segments: the Bile Karpaty unit in the south and the Krynica unit, also known as the Oravska Magura unit, in the north (Figure 3). Only the southern Bile Karpaty segment is present on the territory of Moravia and, as such, a subject of our deliberation. Matejka and Roth (1956) divided the Bile Karpaty unit into the Hluk and Vlara subunits.

The Hluk Subunit. The Hluk subunit is found mainly in the southwestern part of the Bile Karpaty unit; toward the north, it ends, at least at the surface, at the Nezdenice fault. It comprises strata from the Barremian to the Eocene, which, based on lithological and micropaleontological criteria, are divided into several lithostratigraphic units: the Hluk and Kaumberg formations, Puchov Marls, and the Antoninek, Svodnice, Nivnice, and Kuzelov formations (Figure 17A, B).

The oldest, the Hluk Formation of the Barremian–Albian age (Vasicek, 1947; Hanzlikova in Matejka and Roth, 1956; Stranik et al., 1995; Svabenicka et al., 1997), is made by pelagic and hemipelagic black shales, marls, and deep-water turbiditic limestones with cherts. The shales prevail in the lower part of the Hluk Formation, and the limestones prevail in its upper part. Both lithologically and biostratigraphically, the Hluk Formation, more than 120 m (400 ft) thick, may be correlated with the Wolfpassing Member and the Bartberg Member of the Nord zone (Grun et al., 1972) and their stratigraphic equivalents in the Schottenhof zone in Wienerwald (Vienna Forest) in Austria.

The overlying Kaumberg Formation, which, in the Hluk facies, is about 100 m (330 ft) thick, is composed of green and red noncalcareous shales with intercalations of fine-grained laminated sandstones. The rich assemblages of agglutinated foraminifera indicate the Cenomanian to Maastrichtian age of the Kaumberg Formation (Svabenicka et al., 1997).

The Puchov Marls are composed of red calcareous shales and marls with rich calcareous and agglutinated benthic foraminiferal and nannofossil fauna, which indicate the Maastrichtian age (Hanzlikova, 1972a; Bubik and Svabenicka in Stranik et al., 1995; Svabenicka et al., 1997). On the surface and in the wells drilled in the Hluk gas field, the Puchov Marls appear as tectonic slivers, about 100 m (330 ft) thick. Their lithology and biofacies seem to be identical with the Puchov Marls of the Pieniny Klippen Belt and some deposits of the Hauptklippen zone of the Wienerwald (Vienna Forest) in Austria. The depositional and structural adherence of the Puchov Marls to the Bile Karpaty unit thus remains uncertain. These strata may be interpreted as a part of the Pieniny Klippen Belt or as an extension of the Hauptklippen zone of the Wienerwald tectonically incorporated into the Bile Karpaty flysch unit or even as an extension of the klippen belt depositional facies into the Magura realm (Bubik, 1995; Svabenicka et al., 1997).

Equally questionable is the structural position of the Antoninek Formation (Vujta et al., 1989) made by sandy limestones, detrital limestones, marls, and calcareous shales and, based on planktonic foraminifera and nannofossils, assigned to the Campanian–Maastrichtian age. Its known thickness is only a few tens of meters. The Antoninek Formation shows many similarities with the Sneznica Formation of the Pieniny Klippen Belt. Vujta et al. (1989) thus considered the Antoninek Formation to be a tectonic sliver of the klippen belt incorporated into the Bile Karpaty flysch unit.

The Svodnice Formation, as much as 1000 m (3300 ft) thick, was defined by Pesl (1968) and attributed to the Maastrichtian–Paleocene by Svabenicka and Bubik (1992). It is characterized by a facies of rhythmically alternating calcareous shales, subordinate sandstones, and fucoidal limestones, apparently deposited in a distal environment of turbiditic fans. Thicker beds of sandstones named Bzova Sandstones (Potfaj, 1993) occur in the upper part of the formation. Stranik in Elias et al. (1990) compared the Svodnice Formation with the Laab Formation of the Wienerwald (Vienna Forest) in Austria.

The overlying Nivnice Formation (Stranik et al., 1989), about 600 m (2000 ft) thick, is composed of brown-gray and green-gray calcareous shales alternating with subordinate, mostly thin-bedded sandstones and occasional debris flows. Well-preserved foraminiferal and nannoplankton fauna attributes the Nivnice Formation to the late Paleocene–early Eocene (Svabenicka and Bubik in Stranik et al., 1995).

The Kuzelov Formation defined by Stranik et al. (1989) is characterized by a predominance of greenish, brownish, and reddish shales with mostly thin-bedded laminated sandstones. Diagnostic for the Kuzelov Formation are the commonly found beds of pelocarbonates. The Kuzelov Formation is about 250 m (820 ft) thick, and its age ranges from the late Paleocene to the early Eocene (Svabenicka and Bubik in Stranik et al., 1995).

The structural position of the Eocene, the so-called "variegated strata," which occur as narrow bands in some outcrops and wells and are commonly tectonically combined with the Kaumberg Formation, is questionable. Lithologically and stratigraphically, they resemble the Beloveza Formation of the Raca and Bystrica units or the variegated shales of the Proc Formation in the Kopanice facies of the Pieniny Klippen Belt (Stranik et al., 1995).

The Vlara Subunit. The known stratigraphic extent of the Vlara subunit of the Bile Karpaty unit is limited to the Kaumberg, Javorina, Svodnice, and the Chabova formations (Figure 17A, B). The Kaumberg Formation (350 m; 1100 ft) does not differ from its equivalent in the Hluk subunit. Potfaj (1993) called the upper part of the formation, made by alternating thin-bedded sandstones and variegated shales, the Ondrasovec Member.

The Javorina Formation (750 m; 2500 ft) is composed predominantly of sandstones alternating with thin layers of silty shales. The sandstones are characterized by a high proportion of carbonate, mostly dolomitic detritus, apparently supplied from the klippen belt and the nappes of the Inner Carpathians. Because of the poorly preserved microfauna, the age determination of this formation is uncertain and varies in a wide range from the Campanian to Paleocene (Svabenicka, 1990). The Javorina Formation, apparently deposited in a proximal part of a subsea fan system, has no coeval lithologic equivalent in the more distal Hluk facies.

The Svodnice Formation, about 1200 m (4000 ft) thick, is comparable with its equivalent in the Hluk subunit. Its Maastrichtian(?) to Paleocene age has been confirmed by Hanzlikova in Stranik et al. (1989), Svabenicka (1990), and Potfaj (1993). Thick-bedded coarse sandstones with carbonate clasts alternating with thin layers of shales represent the youngest Chabova Formation. Like the Javorina Formation, the Chabova Formation was possibly deposited in a channelized proximal part of the subsea turbiditic fan system. The calcareous nannoplankton indicates the uppermost Paleocene to early Eocene age (Potfaj, 1993). The preserved part of the Chabova Formation is 150 m (500 ft) thick.

The Pieniny Klippen Belt

The Pieniny Klippen Belt is a narrow (only a few kilometers wide) zone of complex steeply dipping fanlike structure, which parallels the Carpathian arc from northeastern Austria to Romania and separates the Inner Carpathians from the Outer Carpathians (Figure 3). It is built of smaller and bigger blocks of weathering-resistant Jurassic and Lower Cretaceous, predominantly carbonate rocks (klippen) that protrude from the less resistant Upper Cretaceous and Paleogene shales, marls, and flyschlike strata. The contact between the klippen and surrounding rocks is mostly tectonic. Excluding the clasts in conglomerates, no pre-Mesozoic sedimentary strata or rocks have been found in the klippen belt. The complex stratigraphy and structure of the Pieniny Klippen Belt were studied by many authors, most prominently by Andrusov (1938) and Birkenmajer (1960).

Until the Late Cretaceous, the depositional history of the Pieniny Klippen Belt was similar to that of the Inner Carpathians. The Jurassic to Lower Cretaceous strata of the klippen belt were apparently laid down on both the attenuated continental crust of the European continental margins and the oceanic crust of the Penninic–Pieninic ocean, which formed during the Triassic and Jurassic rifting north of the Tatric realm of the Inner Carpathians. From the Magura depositional site, the Pieninic oceanic basin was at least partly separated by the Czorsztyn ridge (Figures 5, 6), whose character and function became a subject of discussion. Some authors, e.g., Sotak (1992), combined these two depositional realms into one, the Pieniny–Magura depositional basin, and assigned it to the northern Penninic zone (Valais domain).

In the Late Cretaceous and early Paleogene, the Penninic–Pieninic oceanic crust and lithosphere were gradually subducted, and the Inner Carpathians collided with the Czorsztyn ridge. In this process, the Jurassic to Lower Cretaceous strata of the klippen belt were decoupled from their subducting substratum and piled in front of the progressing Inner Carpathian thrust system, whereas the sedimentation of the Late Cretaceous and Paleogene clastic sequences continued over the top of the klippen belt structures. The appearance of flysch facies with exotic conglomerates would indicate that during the progression of deformation, internal ridges emerged occasionally in the klippen belt depositional realm.

Little direct evidence, at least at the surface, is present for the existence of the Penninic–Pieninic-derived oceanic units in the Inner Western Carpathians. The Upper Cretaceous to Paleogene flysch complexes exposed along the northern Tatric edge in the peri-klippen belt may represent the Penninic ocean in the Carpathians; however, their original basement is not known (Plasienka, 1995). According to several tectonic models, this basement, called Vahicum by Mahel (1981), was underthrusted beneath the Inner Carpathians. Plasienka (1995) suggested that some small tectonic units on the northern side of the Tatric zone of the Inner Carpathians, such as Borinka and Belice, comprise oceanic rock complexes and thus might be considered as parts of the Penninic–Pieninic oceanic realm.

Since the Campanian, the Pieniny Klippen Belt became part of the Outer Carpathian foreland depositional system and, as such, was further affected by the Paleogene and Neogene tectonic phases of the Alpine orogeny. It was the scraping and intense deformation of the Jurassic to Lower Cretaceous strata and simultaneous deposition of younger, Upper Cretaceous and Paleogene strata that resulted in the complex contacts between the older, Jurassic to Lower Cretaceous rigid klippen and their softer Upper Cretaceous and Paleogene siliciclastic depositional cover. The structural complexity of the klippen belt was further enhanced by the strike-slip faulting and backthrusting, which occurred during the latest stages of the Alpine orogeny (e.g., Birkenmajer, 1985).

The Pieniny Klippen Belt is subdivided into the shallow-marine Czorsztyn and the deep-water Kysuca (Pieniny) units and a multitude of transient units, such as the Drietoma, Bosaca, Haligovce, Klape, and Manin units of the so-called peri-klippen belt (Andrusov, 1938). The Czorsztyn unit is characterized by the presence of Jurassic crinoidal and nodular limestones, the existence of an Early Cretaceous hiatus, and the presence of Albian to Maastrichtian variegated marly facies (couches rouges). It was deposited on an elevated site, interpreted either as a fragment of the continental crust (our view) or as a midoceanic ridge (Golonka et al., 2003). The Kysuca unit is distinguished by the presence of the Doggerian spotted marls and radiolarites, the Late Jurassic to Early Cretaceous pelagic cherty limestones, and the Late Cretaceous to Paleogene conglomeratic flysch sequences. The deep-water depositional environment of the Kysuca unit is commonly believed to be oceanic and a part of the Penninic–Pieninic ocean, which separated Apulia from the West European platform (Stampfli, 2001). The transient units of the periklippen belt are composed mainly of Cretaceous and, in part, Paleogene couches rouge marls and flysch sequences with frequent conglomerates.

The contact of the Pieniny Klippen Belt with the Inner Carpathians is tectonic. The klippen belt is steeply thrusted back over the Inner Carpathian units, including the Inner Carpathian Paleogene in the area north of Zilina. The extent of backthrusting of the klippen belt over the Inner Carpathian Paleogene typically ranges within a few kilometers but may reach as much as 8–10 km (4.8–6 mi) in the northern part of the Orava region (Chmelik in Mahel and Buday, eds., et al., 1968). Our interpretation of seismic line 2AT84 in the northern Orava region (Figure 20, section AA′, shown on page 118), suggests at least 12 km (7.2 mi) of tectonic transport of the klippen belt and the inner zone of the Magura unit over the projected edge of the Inner Carpathians. As evident from seismic data, the angle of the leading backthrust faults ranges between 30 and 45°. South of Zilina, the direct contact between the Pieniny Klippen Belt and the most external Tatric units of the Inner Carpathians is obscured by the presence of the peri-klippen belt.

The relationship of the Pieniny Klippen Belt to the inner units of the Magura flysch belt is complex. As demonstrated by the detailed fieldwork and some wells, e.g., Klanecnica-1 (Potfaj, 1993, 1998) and Lubina-1

(Lesko et al., 1978) near Stara Tura in Western Slovakia, the depositional relationship between various units of the Magura flysch and the klippen belt is obscured by enormous structural complexities of the contact zone. In the Bile Karpaty area, Stranik et al. (1989) included the Upper Cretaceous and Paleogene strata of the contact zone into the Kopanice facies. The older Cenomanian to Coniacian strata of this facies, represented by the variegated shales with subordinate sandstones, are comparable with the Kaumberg Formation of the Bile Karpaty unit of the Magura flysch. They are overlain by a turbiditic flysch sequence of alternating sandstones and calcareous shales with sparse intercalations of variegated shales and thin beds of limestones. The sandstones are characterized by a high proportion of clasts from carbonate rocks and by a high amount of calcitic cement. Foraminifers (E. Hanzlikova, 1989, personal communication) and nannofossils (Svabenicka in Stranik et al., 1989) would suggest the Campanian to Eocene age for these several-hundred-meter-thick flysch deposits, which show similarities with the Proc Formation of Eastern Slovakia (Lesko, 1960), and the Upper Cretaceous to Paleogene Jarmuta Formation of the Polish Carpathians (Horwitz and Rabowski, 1929). The whole Kopanice facies is comparable with the Kyjov facies (Stranik and Roth, 1959), the Lackovce facies (Pesl and Mencik, 1959), or the Inovce facies (Lesko, 1960) of Eastern Slovakia, where all these facies are attributed to the cover units of the Pieniny Klippen Belt. In our opinion, the Kopanice facies, as well as other facies of the contact zone, represent the Cretaceous to Paleogene cover of the Jurassic to Lower Cretaceous klippen of the Pieniny Klippen Belt. Considering the complex structure of the contact zone between the Pieniny Klippen Belt and the Magura unit, it is not inconceivable that at least the younger Upper Cretaceous strata known from the klippen belt might also be present in the deeper parts of the complex, imbricated structure of the Bile Karpaty and Krynica (Oravska Magura) units of the Magura flysch. The existence of tectonic slivers of Puchov Marls and Antoninek Limestones, the apparent members of the klippen belt, within the Bile Karpaty unit, would support such an interpretation.

The Inner Carpathian Paleogene

In the middle Eocene (Lutetian), the hinterland spreading and subsidence resulted in a marine transgression over the northern parts of the already deformed and consolidated Inner Carpathians and the formation of the Inner Carpathian Paleogene basin (Podhale basin in Poland) in which thick flysch sequences were laid down until the early Miocene (Figure 5D). The transgressive sequence began with conglomerates, followed by nummulitic limestones, shallow-marine sands, and minor coals. In the late Eocene and Oligocene, a rapid subsidence led to the deepening of the Inner Carpathian Paleogene basin and the deposition of deepwater shales with occasional submarine debris flows and slumps. Locally, e.g., in the Orava, Zilina, and Handlova regions, thin menilitic shales and cherts coeval with the lower Oligocene Menilitic Formation of the Outer Carpathians were laid down. Layers of sedimentary manganese ores are found in the Spis (Kisovce–Svabovce), Handlova, Rajec, and Orava regions (Picha, 1964a). These predominantly shaly strata were followed by an accumulation of turbiditic flysch deposits as much as 2000 m (6600 ft) thick characterized by an upward-increasing proportion of sandstones.

Because the Inner Carpathian Paleogene is not a subject of our study, we restrict our deliberation to a few notes and instead refer to some more comprehensive papers dealing with this subject, e.g., Radomski (1958), Picha (1964b), Chmelik (in Mahel and Buday, eds., et al., 1968), Marschalko (1978), Gross et al. (1984), Nemcok et al. (1996), and Janocko et al. (2006).

The Inner Carpathian Paleogene basin is asymmetrical with its deepest part adjacent to the Pieniny Klippen Belt and shallowing toward the south. The primary source of the clastic material was the actively moving frontal zone of deformation, which, at the time of deposition of the Inner Carpathian Paleogene in the late Eocene to Oligocene, possibly included the northern edge of the Inner Carpathians, the Pieniny Klippen Belt, and the inner zones of the Magura flysch. The overall architecture of the Inner Carpathian Paleogene thus resembled that of a foreland basin; only, it formed in the hinterland of the orogenic belt and did not prograde toward the foreland. Despite the different tectonic setting, the Inner Carpathian Paleogene is a mirror image of the Outer Carpathian late orogenic foreland basin, from which it was separated by the tectonically active Pieniny Klippen Belt and the already uplifted inner zone of the Magura flysch. In the Inner Carpathian Paleogene, only the zones adjacent to the Pieniny Klippen Belt are marginally folded and locally (e.g., in the Orava region) overthrusted by the Mesozoic rocks of the klippen belt. These deformations are at least partly related to the late orogenic transpressional strike-slip motion in the Pieniny Klippen Belt.

The Vienna Basin

The late orogenic to postorogenic Vienna basin is located in the transitional zone between the Eastern Alps and the Western Carpathians on the territory of southern Moravia, Slovakia, and northeastern Austria (Figures 1, 3). Its extent along the strike of the orogenic belt is confined to the limits of the northwest–southeast-trending Dyje–Thaya depression, whose tectonic history goes back at least to the Jurassic rifting and extension. The Vienna basin is superimposed on various nappe units of the Northern Calcareous Alps and the Inner Carpathians, the Pieniny Klippen Belt, and various Outer Carpathian units, namely, the Rhenodanubian and Magura flysch and the Waschberg–Zdanice external units. Geological maps of the basement of the Vienna basin were published by Kroll et al. (1993), Zimmer and Wessely (1996), and Arzmuller et al. (2006).

The depositional fill of the Vienna basin, as much as 5500 m (18,000 ft) thick, is composed mainly of molasse-type shallow-marine and nonmarine detrital deposits of the Eggenburgian (early Burdigalian) to the Pliocene age (Figure 15). The composition and distribution of various facies in the basin reveal the complex geotectonic history of the Vienna basin, which may be divided into three main stages: (1) the Eggenburgian to Ottnangian stage of synorogenic subsidence of the Carpathian thrust belt in the realm of the Dyje–Thaya depression; (2) the Karpatian to early Badenian tectonic transport and pull-apart extension along the orogen-parallel strike-slip faults, and (3) the late Badenian to Pliocene postorogenic subsidence. During its entire history, the Vienna basin was connected with the Pannonian Basin and, until the early Badenian, also with the Neogene foredeep (Jiricek, 1988). The structural and depositional history of the Vienna basin was discussed by Royden (1985), Tomek and Thon (1988), Wessely (1988), Jiricek and Seifert (1990), Fodor (1995), Seifert (1996), and Kovac (2000), among others. Our deliberations are limited only to some general aspects of the stratigraphy and structure of the Moravian part of the Vienna basin. For more information, we refer to the article by Arzmuller et al. (2006).

The Eggenburgian to Ottnangian (Burdigalian) Stage

The depositional history of the Vienna basin began with a marine transgression in the Eggenburgian. The basal conglomerates and sandstones (e.g., Chropov and Winterberg members) are followed by a monotonous sequence of calcareous shales and marls of the lower Luzice Formation. These 300–700-m (1000–2300-ft)-thick strata can be correlated with the coeval Sakvice Marls of the Zdanice unit preserved in the Sakvice depression above the subthrust Vranovice paleovalley and in the Kobyli lake and also with some deposits in the Carpathian foredeep. Two deltas have been recognized in the Eggenburgian depositional system. They prograde from the south-southwest into the deep basinal part of the Vienna basin located along the present Austrian–Slovak border. The boundary between the Eggenburgian and Ottnangian is locally, e.g., in the northern part of the basin, marked by an erosion and deposition of calcareous sands (Hodonin and Luzice sands), which locally rest directly on the Magura flysch basement. The Ottnangian calcareous shales and marls of the upper Luzice Formation, about 300 m (1000 ft) thick, show an upward trend of gradual shallowing and regression of the sea, apparently caused by the tectonic motion in the progressing belt. As a result of this tectonic activity, the piggyback transported Eggenburgian and Ottnangian strata of the Vienna basin were partly deformed and eroded.

The Karpatian to Early Badenian (Burdigalian–Langhian) Stage

The new marine transgression in the Karpatian surpassed the previous extent of the Vienna basin. Through the Sakvice depression in the Zdanice unit, the Vienna basin was connected with the foredeep. The southern part of the basin was dominated by the lacustrine and fluviatile deposits, which, farther north, passed into the deltaic and basinal deposits of the central depression, where the Karpatian strata reached their maximum thickness of about 1400 m (4600 ft). Generally, the Karpatian section of the Vienna basin is characterized by a large lithological variability. According to Jiricek (1988), the lower part of the Karpatian section is represented by the Jablonica Conglomerates, the Tynec Sands, and/or the Laksary schliers; the middle part is made of the Sastin Sands; and the upper part is made of the Zavod schliers and/or the Lab *ostracoda* beds (Figure 15). Kovac (2000) generalized the stratigraphic record and divided the Karpatian deposits of the Vienna basin in Slovakia into the lower part, the Laksary Formation, and the upper part, the Zavod Formation. The regression and deposition of brackish sediments with anhydrite in the Zavod Formation mark the end of the Karpatian. In southern Moravia, this was the time of the final thrusting of the outermost units of the Western Carpathian Flysch belt, including the overlying (piggyback) Eggenburgian to Karpatian strata of the Vienna basin, over the eastern part of the Neogene foredeep. However, because the rigid foreland plate of the Bohemian Massif increasingly inhibited the forelandward thrusting, the compressional stresses tended to be released by the northeast-directed strike-slip motion (escape tectonics). At the Karpatian–Badenian transition, the Vienna basin thus began to attain the character of a pull-apart basin dominated by the orogen-parallel strike-slip faults (Burchfiel and Royden, 1982; Royden, 1985).

The Badenian stage began with a new marine transgression and deposition of thick calcareous shales (tegels) of the Lanzhot Formation. The existence of analogous deposits in the foredeep would affirm to a renewed communication between these two depositional systems. The continuing pull-apart mechanism led to a further deepening of the Moravian central depression, which formed between the Steinberg and the Lanzhot–Hrusky fault systems, and which was filled with more than 4000 m (13,100 ft) of Neogene deposits. The subsidence of the depression peaked during the middle Badenian, when thick, variegated, predominantly shaly deposits of the Zizkov Member of the Hrusky Formation were laid down (Figure 15). A large, deltaic system with lobes extending into the central depression formed at the western side of the Vienna basin. Two smaller deltas have been recognized in the northeastern and southwestern ends of the basin. In the upper part of the Zizkov Member, shallow-marine Lab Sands were deposited in coastal areas, whereas bioherms of lithothamnion limestones formed on the elevations. The Lab Sands represent the most important hydrocarbon-bearing reservoir in the Vienna basin.

The Late Badenian (Serravalian) to Pannonian (Tortonian) Stage

In the late Badenian, because of the gradual decrease of orogenic stresses, the strike-slip fault system of the Vienna basin became gradually inactive. The subsidence, rather than the pull-apart extension, increasingly dominated the depositional regime after the late Badenian. The Vienna basin shallowed and was converted into a brackish embayment, which only sporadically communicated with the open sea. Following the deposition of the variegated fresh-water clays and sands at the Badenian–Sarmatian transition, the basin spread into the Hradiste graben. There, the Sarmatian deposits rested directly on the Magura flysch. The brackish sediments of the Bilovice Formation, as much as 800 m (2600 ft) thick, are represented by coarser clastics, lumachellas, and oolitic limestones in the coastal areas and calcareous clays and sands in the basinal facies.

During the Pannonian, the connection with the Pannonian Basin was further restricted, and the Vienna basin became a slightly brackish lake with marginal lagoons and swamps. The lowest Bzenec Formation consists of deltaic gravels and sands at its base and calcareous clays in its upper part. The Kyjov lignite seams (Figure 15) overlain by thick sands and variegated calcareous clays formed in the marginal marshes. The existence of a thick and widespread lignite bed in the lower part of the overlying Dubnany Formation indicates that the marshes and swamps expanded over most of the remnant Vienna basin. The lignites have been mined near Ratiskovice, Dubnany, Hodonin, and Mikulcice. The deposition of the lacustrine coaly clays, sands, and variegated shales of the fresh-water Gbely Formation concluded the depositional history in the Vienna basin at the territory of Moravia.

During the Pliocene–Holocene, fluviatile sediments accumulated in the valleys of the Morava and Dyje rivers, and lacustrine deposits several hundred meters thick were laid down in the Zohor–Plavecky Mikulas graben at the western side of the Male Karpaty Mountains.

Structure of the Vienna Basin

The internal architecture of the Vienna basin is complex. It is dominated by a system of grabens and horsts that formed and modified in response to the compressional and tensional stresses in the underlying Carpathian thrust belt and the subthrust foreland plate. During its dynamic structural history, the Vienna basin passed from the early stage of subsidence on the top of the progressing orogenic belt to the stage of the pull-apart extension and finally to the stage of postextensional subsidence.

The kinematic character of the strike-slip faults in the Vienna basin became a subject of discussions. According to Royden (1985) and Royden and Dovenyi (1988), the pull-apart extension of the Vienna basin occurred in the underlying thin-skinned thrust belt above a relatively shallow sole detachment of the thrust belt. Such an interpretation would be consistent with the low heat flow (Cermak, 1979) and only a limited thermal subsidence following the pull-apart extension. Seismic data, however, indicate that at least some of the major strike-slip faults, e.g., the Steinberg and Leopoldsdorf faults, do not sole out in the basal detachment of the thin-skinned thrust belt but rather continue deeper into the subthrust basement (Wessely, 1988). Lankreijer et al. (1995), based on their structural analysis, suggested that the structural pattern of the Vienna basin changes from the thin-skinned extension in the northwestern part to the whole lithospheric extension in the central and southeastern part of the basin. Picha (1996) suggested that the deep-rooted strike-slip faults in the Vienna basin may copy the fault planes of the preexisting normal faults formed at the time of the Jurassic and Early Cretaceous rifting and extension of the European plate (Figure 20, section DD′, shown on page 118). During the convergency, these and some other preexisting faults might have been reactivated as reverse faults and/or as strike-slip faults in the lower level of the orogenic system, which apparently existed below the thin-skinned thrust belt. Locally, these deep faults penetrated into the overlying thrust belt and created a new structural regime in the thin-skinned belt. Apparently, it was the late orogenic northeastward translation of the entire Carpathian system along the Western Carpathian transfer zone, during which the actively moving deep faults propagated into the overlying thin-skinned belt and opened the pull-apart structure of the Vienna basin. Picha (2002) compared the structural setting of the Vienna basin with that of the Albanian foredeep in the Adriatic region. Like the Vienna basin, the Albanian foredeep formed in a major northwest–southeast-trending strike-slip transfer zone, which existed between the Dinarides and the Apulian plate and provided a means of tectonic transport from the collision zone of Apulia with Europe toward the subduction zone of the Hellenic trench. Both the Vienna basin and the Albanian foredeep thus represent a specific type of pull-apart basin, associated with the escape strike-slip tectonics. They evolved on a thick continental crust devoid of magmatic centers and thermal anomalies.

Neovolcanics in the Outer Carpathian Thrust Belt

The only neovolcanic rocks known from the Outer Carpathian thrust belt in Moravia are the trachyandesites and alkalic basalts at Banov, Bojkovice, and Hrozenkov in the vicinity of Uhersky Brod (Figure 3). They have been investigated by Schmidt (1858), Tschermak (1858), and Klvana (1891), among others, and more recently by Krystek (1955), Shrbeny (1974), Prichystal (1993), Adamova et al. (1995), and Prichystal et al. (1998).

These neovolcanics appear as dikes and sills, several meters to several tens of meters thick in all partial units of the Magura flysch and locally in the Pieniny Klippen Belt. They cut through the thrust structures without being affected by any tectonic movement, thus clearly documenting their post-thrust origin. Radiometric analyses indicate the early Badenian (16.8 ± 0.6 Ma) age of these volcanic rocks. Their origin thus falls in the time of the late orogenic strike-slip faulting and opening of the pull-apart stage of the Vienna basin. The country rocks of the dikes and sills are affected by contact metamorphism. So far, no remnants of surface volcanism have been found, although at several localities near Banov and Nezdenice, volcanic breccias that apparently formed in volcanic vents are present. Associated with the volcanic activity are mineralogical occurrences of polymetallic ores and asphaltic organic materials.

The chemical composition, close to that of the typical alkalic volcanic rocks, indicates that these Outer Carpathian volcanics attain a transitional position between the alkalic neovolcanic rocks of the Bohemian Massif and of the calc-alkalic neovolcanic rocks of the Inner Carpathians in Slovakia. The strontium (Sr) isotopic ratio points to the mantle as a source of magma for these Outer Carpathian neovolcanics.

Rhyolite and rhyodacite tuffites are known from the Badenian strata of the Neogene foredeep and from the Eggenburgian strata of the Vienna basin in Moravia. These tuffitic rocks were apparently supplied from volcanoes located in the Inner Carpathians.

THE STRUCTURE OF THE OUTER WEST CARPATHIAN THRUST BELT

The Outer Carpathian thrust belt on the territory of Moravia is a thin-skinned structure composed entirely of sedimentary sequences. However, below this thin-skinned belt and the Neogene foredeep, deeper compressional structures exist involving both the crystalline basement and its sedimentary cover (Picha, 1996; Krejci et al., 2002) (Figure 20). As indicated by numerous wells and regional seismic lines, the thin-skinned Outer Carpathian thrust belt consists of numerous imbricates, duplexes, and thrust sheets (nappes). The internal architecture of these various thrust units is strongly controlled by the stratigraphy. Whereas the incompetent, predominantly shaly successions display a complex internal deformation marked by tight folds, imbricates, and duplexes, the thick competent sequences with a high proportion of sandstones or carbonates are less internally deformed and form larger scale folds and thrust sheets (nappes). Architectural differences also exist between the internal and external zones of the Carpathian thrust belt. Various thrust sheets of the external zones typically dip rather gently toward the hinterland; in the inner zones, as evident in the relatively consistent Magura nappe, the inclination of thrust units gradually increases. Some strata are even overturned and locally thrusted backward. The fanlike complex structure of the Pieniny Klippen Belt and of the adjacent zone of the Flysch belt is partly thrusted back over the edges of the Inner Carpathians (Figure 20, section AA′).

The major thrust faults bounding major tectonostratigraphic units might have initiated along the preexisting normal faults, which, during the divergent phase, separated various depocenters and intervening ridges in the Outer Carpathian depositional system. However, not every juxtaposition of lithologically distinct units should be interpreted as a major, tectonically induced separation of those units in their original depositional setting. Significant lithological changes between tectonically separated units, such as the different proportion of sands and shales, may as well be explained by lithological variations in the channelized depositional fan system.

Like in other thrust belts, the deformation of various units of the Outer Western Carpathians progressed from the hinterland toward the foreland. The regular sequential succession of the forelandward-heading thrust sheets is locally disjoined by an appearance of the out-of-sequence thrust imbricates and duplexes, such as the window units in the Magura complex nappe system. The Magura unit itself is a major out-of-sequence thrust sheet, which glides over the structures of the external units (Figure 20).

The timing of deformation and thrusting of various major units of the Outer Carpathians is typically established from the age of the strata in the footwall and the termination of the continuous sedimentation in the hanging wall. In the Pieniny Klippen Belt zone and the innermost Bile Karpaty unit of the Magura flysch in Moravia, the sedimentation ended in the Eocene; in the more external Raca subunit of the Magura flysch, the sedimentation continued until the early Oligocene; and in the most external Zdanice and Pouzdrany units, the uninterrupted sedimentation lasted into the early Miocene. In time, the deformation of the Outer Carpathians not only progressed from the internal to the external zones but also along the strike of the arcuate Carpathian belt from the west to the east (Buday et al., 1961). At its western end in northeastern Austria and southern Moravia, the frontal Pouzdrany and Waschberg–Zdanice units are thrusted over the lower Miocene (Eggenburgian to Karpatian) deposits of the Neogene foredeep; in northern Moravia, the lowermost middle Miocene (lower Badenian) strata are found below the edges of the thrust belt (Jurkova, 1976). In the Vrancea zone of the Eastern Carpathians, the thrusting has continued into the Pliocene–Quaternary(?).

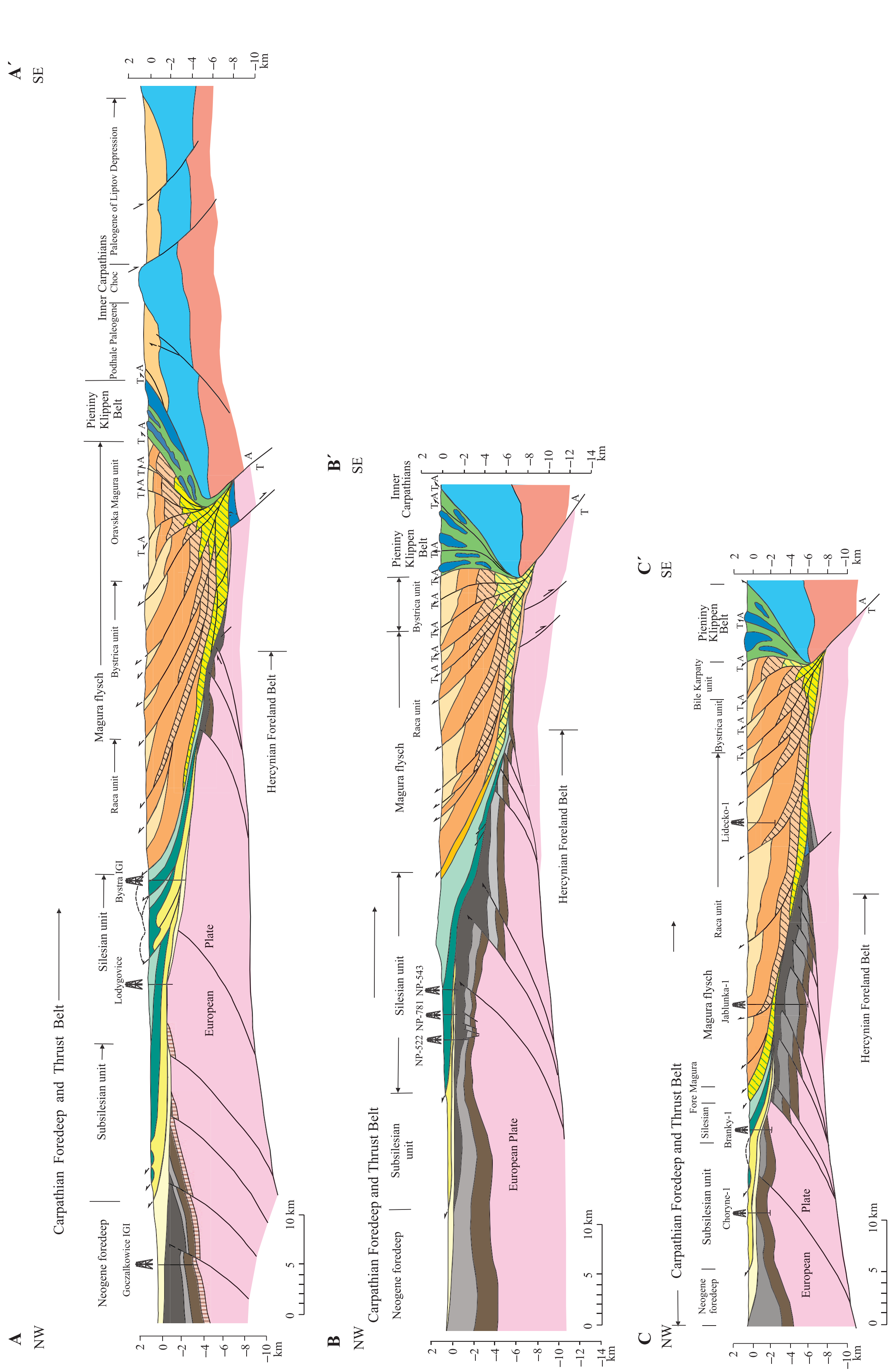
Strike-slip Faulting and Backthrusting

The thrusting of the Outer Western Carpathian units over the European foreland in Moravia, especially during the last stages of the orogenesis in the late early to middle Miocene, was associated with the southwest–northeast-trending left-lateral strike-slip motion related to the northeastward extrusion (escape) of the Carpathian–Pannonian block (Roth, 1980b; Tomek et al., 1987; Ratschbacher et al., 1991a). As a consequence of this lateral slip, the various units of the thrust belt progressed toward the foreland increasingly in an oblique direction, generating transtensional and transpressional stresses elsewhere in the thrust system. The presence of the transtensional stresses is well documented by the opening of the pull-apart stage of the Vienna basin in southern Moravia and northeastern Austria, whereas the existence of the transpressional folding, faulting, and backthrusting is best evident in the Pieniny Klippen Belt and the adjacent zone of the Magura nappe in Poland (Swidzinki, 1953) and Eastern Slovakia (Stranik, 1965). The stress measurements by Nemcok et al. (1998a, b) indicate that the proportion of shortening accommodated by the sinistral strike slip is the highest in the southwestern part of the Western Carpathian arc and decreases toward the north, where most of the shortening is accommodated by the frontal compression. Most of the strike-slip motion apparently occurred in the southwest–northeast-trending Western Carpathian transfer zone especially along the deep faults of the Pieniny Klippen Belt. The Western Carpathian transfer zone, which was a major factor in opening the Outer Carpathian depositional system during the divergent stage, in a reverse sense of the motion, also facilitated a great deal of the tectonic shortening and escape tectonics during the convergent stage (Figure 5).

The combination of normal forward thrusting and strike-slip motion resulted into an extremely complex structural pattern of the Pieniny Klippen Belt, which is partly thrusted back over the Inner Carpathians (Figure 20). The backthrusting of the Outer Carpathian units over the Inner Carpathian Paleogene most likely occurred during the last stages of the Alpine orogeny in the late early to middle Miocene and was apparently coeval with the opening of the pull-apart stage of the Vienna basin. The extension in one area was thus compensated by the compressional shortening in other parts of the interconnected orogenic system.

Deep Structures Below the Foredeep and the Thin-skinned Belt

Below the Neogene foredeep and the thin-skinned Outer Carpathian belt in Moravia, deeper compressional and extensional structures exist whose geometries are not related to the dominant structural trends of the overlying belt. The normal and reverse faults bounding these features, with a few exceptions, such as the strike-slip faults in the Vienna basin, do not continue into the thrust belt. The sole decollement of these deep


A
NW
A´
SE
Carpathian Foredeep and Thrust Belt
Neogene foredeep
Subsilesian unit
Silesian unit
Magura flysch
Raca unit
Bystrica unit
Oravska Magura unit
Pieniny Klippen Belt
Inner Carpathians
Podhale Paleogene
Choc
Paleogene of Liptov Depression
Goczalkowice IGI
Lodygovice
Bystra IGI
European
Plate
Hercynian Foredeep Belt
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Neogene foredeep
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Search

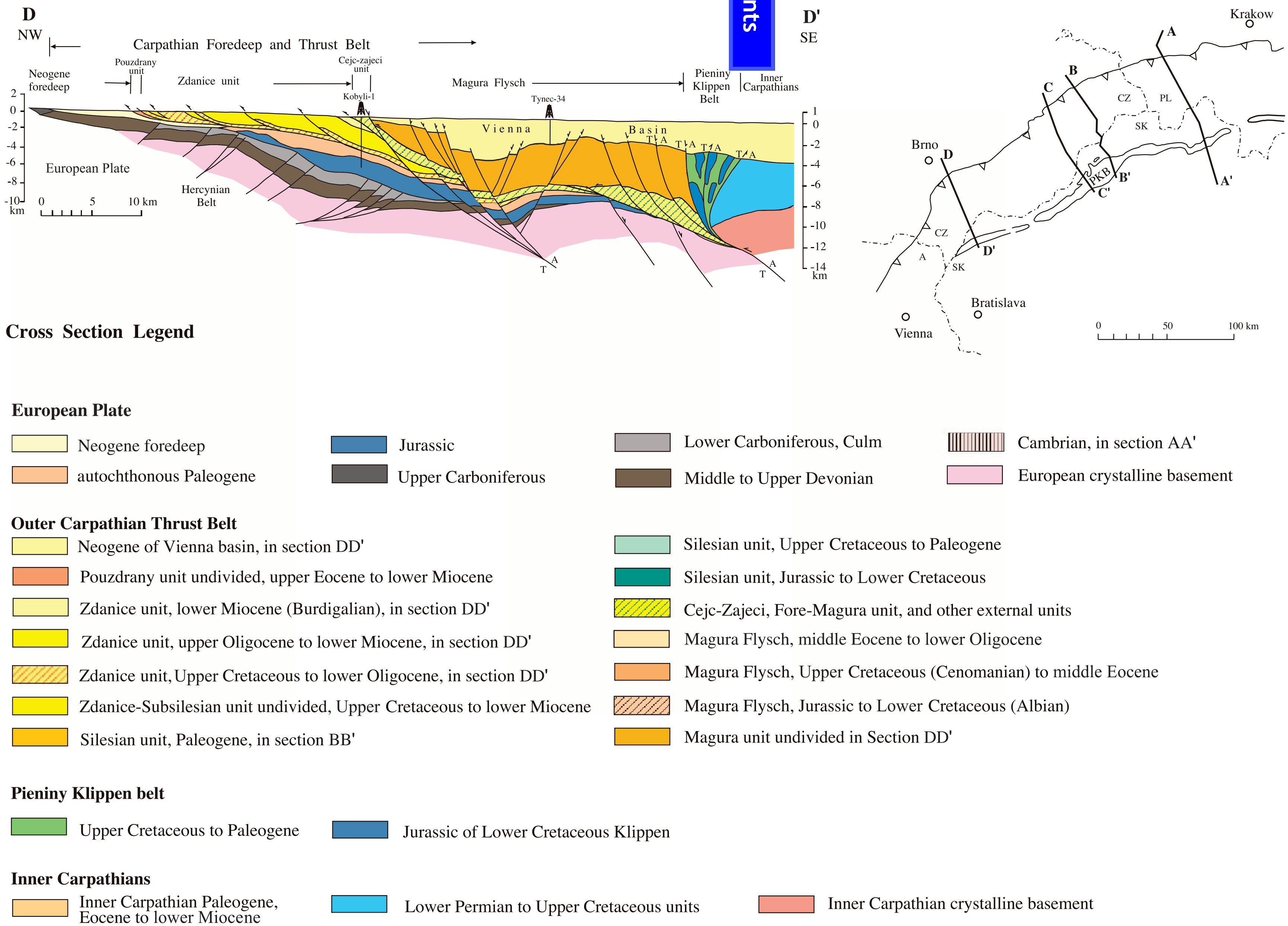

Figure 20. Four regional partly balanced cross sections through the Western Carpathian thrust belt and its foreland along the lines: AA′ Bielsko Biala–Low Tatra Mountains, BB′ Odra Hills–Manin, CC′ Hranice–Ilava, DD′ Brno–Kuty. Location of the lines is shown in Figures 3 and 7. The same vertical and horizontal scales apply. Stratigraphic records of some of the wells shown on cross sections CC′ and DD′ are reported in Appendix 1.

structures may exist at greater depths in the brittle to ductile transition zone or even at the base of the crust. Structural analyses indicate that these structures belong to several genetically diverse types. The oldest recognized, deep antiformal features are the Hercynian (late Paleozoic) anticlinal structures typically bounded by the southeastward-verging high-angle reverse faults, which typically terminate in the youngest Carboniferous (Stephanian) strata and do not continue into the overlying Mesozoic and Cenozoic strata of the European foreland plate. However, in areas without a significant autochthonous cover, which would document their age, the origin of these structures remains uncertain. They may as well be interpreted as Alpine structures associated with antithetic back-arc faulting in the foreland plate. An example of such a structure of questionable origin is the Orava structure evident on the seismic line 2AT/84, whose interpretation is presented at the cross section AA′ in Figure 20. The association of this structure with either the Hercynian or Alpine orogeny is difficult to ascertain from the existing data. In the cross section AA′, we tentatively interpret the structure as Hercynian but do not exclude its younger Alpine origin.

Within the Jurassic Dyje–Thaya depression of southern Moravia and northeastern Austria, deep antiformal structures exist, which are apparently generated by the rotation of crustal blocks during the Jurassic rifting and extension of the European plate. At least one of them, the large antiformal Tynec structure (Figure 20, section DD′), was reactivated during the late orogenic strike-slip faulting. The original Jurassic fault bounding the western side of this structure might have been a precursor of the major orogen-parallel Steinberg fault, along which the pull-apart depocenter of the Vienna basin opened in the early to middle Miocene. In the territory of Austria, some of these rift-related structures have been explored by deep drilling (Wessely, 1990). They represent the third depth level of exploration in the Vienna basin area, the first being the Vienna basin and the second the Carpathian thrust belt.

Some of the southeastward-dipping normal and reverse faults identified on the seismic lines in the European platform, along the western side of the Pieniny suture zone (Figure 20, sections AA′, BB′, and DD′), may also be genetically related to the Jurassic rifting. Typically, these faults do not propagate into the overlying nappes, although some of them might have been activated by transpressional stresses during the last stages of the Alpine orogeny.

Lastly, below the thrust belt and the Neogene foredeep, deeper compressional structures exist, whose origin is clearly related to the Alpine orogeny. Some of them may be associated with the Late Cretaceous–early Paleocene Laramide uplifting of the Carpathian foreland; the others may have been active during the late Paleogene to the middle Miocene convergency and lateral escape. Because of the compressional stresses, the previously attenuated European crust was compressed and thrusted back toward the foreland, thus forming a lower structural level of the Carpathian orogenic belt, as suggested among others by Roure et al. (1993), Roca et al. (1995), and Roure and Sassi (1995). One such recognized structure is the overthrust of plutonic rocks of the Brno massif over Cenomanian sediments near the town of Blansko (Krejci et al., 2002). The chain of antiformal structures, which parallels the frontal zone of the thrust belt and the foredeep in Moravia known as the Slavkov–Tesin ridge (Dlabac and Mencik, 1964), and the positive structure that underlies the Dolni Dunajovice gas field (Figure 26, shown on page 137) are possibly other examples of the young Alpine compressional tectonics in the foreland plate. The orientation of some of the foreland-type structures is not necessarily parallel with the Alpine–Carpathian structural trends; their directions might have been inherited from the preexisting structural trends in the European foreland. The Alpine foreland-type deformations apparently propagated far into the foreland. F. Chmelik (1977, personal communication) even anticipated that the Alpine deformations may continue westward as far as the eastern side of the Boskovice Furrow (graben) in central Moravia (Figure 3).

The compressional deformation of the underlying European plate in the Carpathians resembles the structural pattern of southern Apennines, where, below the thin-skinned Apenninic thrust belt, deeper parautochthonous structures exist in the Apulian plate at a depth of 4–6 km (2.4–3.6 mi). Some of these deep structures have been drilled, and significant accumulations of hydrocarbons, such as the Tempa Rosa, Monte Alpi, and Volturno fields, have been found in these deep prospects.

The origin and age of the various deep structures in the sub-Carpathian foreland in Moravia remain uncertain and deserve further attention. Some of these structures represent potential targets of exploration for hydrocarbons and, therefore, the timing of their formation remains possibly one of the most critical factors in any future evaluation of their hydrocarbon potential (Picha, 1996).

Tectonic Shortening

Restoration of orogenic belts to their original predeformation stage represents a complex task that is commonly addressed by construction of balanced cross sections and calculation of the rate of shortening by dividing the length of shortening by the time during which shortening occurred. Attempts also have been made to establish the shortening using other indicators such as the plate-tectonic reconstructions or the distance of the flexural wave migration in the foreland and the width of the orogenic wedge (DeCelles and DeCelles, 2001). The balanced cross sections, possibly the most useful tools in restoring the thrust belts, are normally valid only in the unmetamorphosed external zones of orogenic belts, e.g., the Outer Carpathians, and even in these relatively simple zones, small-scale compressional structures may produce significant shortening that hardly can be integrated into the balanced restorations.

A significant factor that has to be considered in restoring the Carpathian depositional and structural settings is the amount of erosion that occurred both prior to and during the orogeny. As the deformation and uplifting progressed gradually from the hinterland into the foreland of the thrust belt, the more internal, already uplifted and deformed units, such as the Magura flysch, shed clastic material into the more external zones of the thrust belt and eventually into the autochthonous foredeep. The present frontal edges of various thrust sheets thus do not fully represent the original extent of these units. In the Beskydy region of northern Moravia, for example, the frontal zones of the Magure nappe were removed by erosion, and the underlying Silesian nappe was exposed on the surface in a large half window (Figure 3).

To calculate the amount of shortening of the Outer Carpathians in Moravia, four partly balanced cross sections have been constructed (Figure 20). Their geometry is constrained by the combination of surface geology, well data, and interpretation of seismic lines. The amount of shortening can be more reliably established in the less deformed more competent units with a high proportion of sandstones, such as the Magura unit, whereas the shortening of the highly deformed, incompetent, predominantly shaly units, e.g., Subsilesian sector of the Waschberg–Zdanice–Subsilesian unit, can only be guessed. Numerous deep and shallow wells drilled in the Subsilesian unit of central and northern Moravia indicate that its internal structure is extremely complex. Detailed folding and faulting and the existence of duplexes known, for example, from the tectonic windows in the Beskydy region would suggest a very high rate of shortening, whereas an absence of a full stratigraphic section in other areas would indicate an existence of tectonic stretching. As the lowermost unit of the Outer Carpathian nappe stack in Moravia, the incompetent Subsilesian unit (sector) was squeezed and smeared along the base of the more competent Silesian and Magura nappes. In the process, even some Neogene foredeep strata were detached and integrated into the frontal zone of the thrust belt. The overall amount of shortening of the Subsilesian unit thus cannot be reliably established from the existing evidence.

The internal structure of the Silesian nappe, which, as indicated by seismic data, does not seem to continue far below the Magura nappe, is also complex. Locally, tectonic slices of crystalline basement rocks, Devonian and Carboniferous strata, as well as Neogene deposits of the foredeep are incorporated into the basal part of the Silesian unit. The Silesian unit comprises both massive, several-kilometers-thick, competent strata of the Upper Cretaceous Godula and Istebna flysch formations and the predominantly incompetent Upper Jurassic and Lower Cretaceous strata laid down on the rifted passive margins. During the deformation and tectonic transport, these two lithologically different sets of strata were locally decoupled, deformed, and thrusted disharmonically. The structural discrepancies between the competent upper and incompetent lower parts of the Silesian unit were originally explained by an existence of two separate nappes, the lower highly deformed and imbricated lower Tesin (Teschen) nappe and the upper Godula nappe with a relatively simple structural pattern (Mencik, 1966). According to some authors, e.g., Jurkova (1971), the complex disharmonic structure of the Silesian nappe resulted from two phases of deformation of the Silesian and Subsilesian units in terms of Stille's (e.g., 1936) concept of orogenic phases. We believe that the complexities of the internal structure of the Silesian unit can better be explained by the disharmonic folding, faulting, and decoupling of competent and incompetent strata during a continuous process of shortening and overthrusting of the Outer Carpathians onto the European foreland. During this process, the competent younger strata of the Silesian unit were locally thrusted over the older incompetent members of the unit, thus invoking the idea of an existence of separate nappes formed during the subsequent stages of the deformation.

The relatively competent Magura nappe displays a more regular internal structure, consisting of numerous imbricates and thrust sheets. Its relatively simpler internal structure enables a more reliable restoration and calculation of the range of the tectonic shortening. As a whole, however, the Magura unit is an out-of-sequence unit (e.g., Nemcok at al., 1998a) superimposed on the external units of the Outer Carpathian thrust stack. This further increases the amount of tectonic shortening of the whole Outer Carpathian thrust system.

Based on the interpretation of the well data and the regional seismic profiles, we tend to believe that the structural units of the Outer Carpathian belt (Subsilesian, Silesian, Magura, and the Pieniny Klippen Belt) do not continue far underneath the Inner Carpathians. The progressing stack of the Inner Carpathian nappes apparently scraped off most of the Late Cretaceous and Paleogene Outer Carpathian flysch strata and piled them at its front. Some of the older Upper Jurassic and Lower Cretaceous deposits of the Outer Carpathian basin might have been left behind on the pre-Jurassic substratum and, with it, partly subducted. In that sense, the Carpathian sector seems to differ from the Eastern Alps in Austria, where the Helvetic and Flysch units continue further below the nappes of the Calcareous Alps (e.g., Zimmer and Wessely, 1996). With respect to these differences, it is necessary to stress that the contact between the Inner and Outer Carpathians was greatly affected by the strike-slip motion in the Pieniny Klippen Belt.

The tectonic shortening in the Outer Carpathians is apparently not limited to the thin-skinned sedimentary cover but encompasses the decoupled basement and possibly the whole crust as well. The crustal-scale convergency of the previously attenuated and rifted crust of the passive margins and underthrusting of the oceanic crust, as proposed for the Inner Carpathians by Plasienka (1995), seems to explain better the geodynamic evolution of the Carpathian thrust belt without resorting to multiple subductions, whose existence is not evidenced by subduction-related Mesozoic and Paleogene volcanism.

The process of rifting and extension, which formed the Outer Carpathian passive continental margins, was reversed during the convergent phase. The various rift-apart crustal blocks, including the Silesian and Czorsztyn internal ridges, were compressed and accreted back to the European platform (Figure 6). In such a process, the original extensional normal faults might have been reactivated as thrust and reverse faults. Both during the extensional and compressional phase, the various crustal blocks were apparently decoupled from the mantle, and during the conversion, the depositional sequences of the thin-skinned belt were also decoupled from their substratum. Such an interpretation largely reduces the amount of subduction of the lower crust needed to compensate for the shortening in the thin-skinned thrust belt. This is compatible with a similar situation in Eastern Alps, where, according Helwig (1976), a significant subduction of the continental crust did not occur either.

However, in comparison with the Western Alps, where the advanced collision led to the inversion of the external massifs, the inversion of the European foreland below the Carpathian belt is minor, and major basement structures of the Aar Massif type are lacking (Roure et al., 1993). According to Ziegler and Roure (1996), the absence of major post-Paleocene compressional foreland structures beneath the Carpathians suggests only a low level of coupling between the orogenic wedge and its foreland.

Based on the restoration of balanced cross sections (Figures 6, 20), our calculations suggest that the minimum width of the Outer Western Carpathian depositional system, excluding the Pieniny Klippen Belt, in the time of its maximum extent in the late Cretaceous was at least 230 km (143 mi). Because the present width of the thin-skinned Outer Carpathian belt in Moravia, excluding the impact of erosion, is about 70 km (43 mi) (Figure 3), the minimum cumulative shortening would be about 160 km (99 mi). Thus, during 17 m.y. of convergency from the late Oligocene to early Badenian, the extent of the original depositional environment of the Outer Carpathians was reduced by about 70%. Such a shortening of the sedimentary sequences must then be compensated by an equal reduction of the basement. Our interpretations (Figure 6) indicate that the original depositional basement of the Outer Carpathians, including both rifted and attenuated continental and possibly some oceanic crust, was shortened 35% by compression and accretion to the foreland plate and an additional 35% by underthrusting the Carpathian plate and partial subduction into the asthenosphere (see discussion in the following section). The total shortening of the basement by 70% would thus compensate for about 70% of the shortening of its sedimentary cover accreted into the thin-skinned Outer Carpathian thrust belt. Considering about 17 m.y. of the duration of convergency from the middle Oligocene to the middle Miocene (Badenian) and the minimum amount of shortening (160 km; 99 mi), the rate of shortening of the Outer Carpathians in Moravia was about 9.4 mm/yr (0.37 in./yr). A similar rate of shortening of about 10.6 mm/yr (0.42 in./yr) may be calculated from the amount of shortening (from 235 km [146 mi] in the middle Oligocene to 55 km [34 mi] at the present time) anticipated by Roca et al. (1995) for the Polish sector of the Outer Western Carpathians. The rate of shortening of 7–10 mm/yr (0.27–0.39 in./yr) during the Eocene–Miocene determined from restored balanced cross sections in the North Alpine foreland basin by Pfiffner (1986) is also quite comparable.

The minimum anticipated width of the Outer Carpathian depositional system of about 230 to 250 km (about 142 to 155 mi) is thus comparable with the present width of the North Sea basin between Norway and the Shetland Islands. If inverted and thrusted, that part of the North Sea would possibly form a thrust belt of a width similar to that of the present Outer Western Carpathian belt.

The minimum ranges of shortening of the Outer Carpathians based on construction of balanced cross sections, however, differ substantially from much larger amounts of tectonic shortening anticipated by other authors, e.g., Roth (1987). Based on the mutual position of the Cretaceous paleomagnetic poles, he calculated that since the Cretaceous, the total north-south approach of the present-day North European and African plates amounted to 2200–2300 km (1375–1438 mi). The existing width of the Outer Carpathian belt (about 70 km [43 mi]) and the volume of sediments comprised in it would not allow for such an enormous shortening.

The Extent of the European Plate Underneath the Carpathian Belt

More than 100 yr ago, Suess (1875) proposed that the northwestern Europe was partly buried below the Alps and Carpathians, and somewhat later, Ampferer (1906) introduced the concept of crustal subduction. Zoubek (1948) and Stille (1953), among others, recognized that the crystalline basement of the European plate (Bohemian Massif) differs from the basement of the Inner Carpathians and postulated that during the Alpine convergency, the European plate was underthrusted below the Carpathian orogenic belt. More recently, the character of the deep contact of the Carpathian belt with the underlying platform has been examined by seismic (Tomek and Hall, 1993), seismotectonic (Schenk et al., 1994), magnetotelluric (Cerv et al., 1994), magnetic (Gnojek and Heinz, 1993), electromagnetic (Jankowski et al., 1985), and other methods. For more information on the subject, see chapters by Hrusecky et al. (2006), Nemcock et al. (2006), Pospisil and Adam (2006), and Pospisil et al. (2006) in this publication. Despite these intensive studies, the extent of the subthrust European plate below the Carpathians and the amount of its subduction have remained subjects of discussion. Some authors (e.g., Slaczka, 1975; Roth, 1977, 1980b; Dudek, 1980; Stranik et al., 1993) have suggested that the European plate extends to the axis of a distinct zone of gravity lows, which, in the Western Carpathians, roughly parallels the northwestern side of the Pienniny Klippen Belt from Hodonin to Namestovo and then continues northeastward toward the Nowy Sacz. The existence of such a low-gravity zone has been alternatively explained by a deep gravity minimum, deep faulting and fracturing in the Moho, and even by inhomogeneity in the top of the upper mantle. Interpretation of reflection seismic data, however, indicates that the gravity minimum is most likely caused by shallow factors such as the maximum thickness (10–12 km; 6–7.2 mi) of the accretionary wedge of the low-gravity flysch and molasse rocks (e.g., Ibrmajer, 1971; M. Krs and A. Sutora, 1974, personal communication; Pospisil and Filo, 1980; Tomek, 1982), and that the European platform continues underneath this zone farther east below the Pieniny Klippen Belt (Figure 6D). The gravity data indicate that the lower part of the accretionary wedge has lower gravity than its upper part. This can be explained by the piling of lighter, predominantly shaly duplexes of the external units of the Flysch belt, such as the Fore-Magura, Grybow, and Obidowa Slopnice, below the heavier thrust sheets of the Magura flysch in the low-gravity zone (Figure 20).

Another widely considered line of termination of the European plate is the peri-Pieniny lineament of Maska (in Buday et al., 1961; Svoboda, ed., et al., 1966; Pecova et al., 1979). Originally located in the Pieniny Klippen Belt, the peri-Pieniny lineament was later relocated into the western side of the Male Karpaty Mountains and called the Zahorie fault (O. Fusan, J. Ibrmajer, and J. Plancar, 1979, personal communication).

Other authors, e.g., Uhlig (1907), Roth (1978), and Lesko et al. (1980), suggested that the European platform extends farther east and south beyond the Pieniny Klippen Belt and below the Inner Carpathians. According to Roth (1978), the North European platform below the Inner Carpathians continues as far as the centers of the volcanic activity in central Slovakia. The continuation of the European platform below the Calcareous Alps–Inner Carpathians is at least partly documented by results of the deep wells Berndorf-1 and Aderklaa-UT1 in Austria (Wessely, 1990), which, below the edges of the Calcareous Alps, drilled into the Bohemian Massif. In addition, the reflection profiles from the Danube basin (western part of the Pannonian Basin) would indicate that the European plate may extend far below the Austroalpine and Penninic units of Alps to the Mihalyi high and the Raba line, about 130 km (80 mi) southeast from the frontal edge of the thrust belt (Szafian et al., 1999).

Our geological interpretation of regional seismic lines across the Outer Carpathians in Moravia and Slovakia shows that the European platform continues uninterrupted to the vicinity of the Pieniny Klippen Belt, where it is intersected by normal and reverse faults, which do not continue into the thrust belt (Figure 20). At least some of these faults might have originated during the Jurassic to Early Cretaceous rifting and extension of the European plate. Later, during the Alpine orogeny, some of these normal faults might have been inverted by compressional and transpressional stresses. We believe that these faults mark the break between the thick platform-type crust of the European plate and the rifted attenuated crust of the European continental margins. During the compressional orogeny, the edges of the thick European platform acted as a buttress, causing the piling of the rootless slices of the Flysch belt and the Pieniny Klippen Belt in the low-gravity zone. Given the significant component of strike-slip motion in the Pieniny Klippen Belt, it is likely that the pieces of the European crust juxtaposed on both sides of the klippen belt had moved laterally. Their present fit may thus differ from the original one especially prior to the late orogenic northeastward translation of the Western Carpathians.

The attenuated crust of the European continental margins, unless completely subducted, must continue beyond the Pieniny Klippen Belt underneath the Inner Carpathian nappes, possibly to the volcanic centers of central Slovakia, as proposed by Roth (1978). We anticipate that the European plate bends under the Inner Carpathian (Apulian) plate and descends at an angle of about 45° into the asthenosphere. At a depth of 80–100 km (50–60 mi), it partly melts and generates the subduction-type volcanic activity (Figure 6D). The European plate thus underplates the northwestern part of the Inner Carpathians in Slovakia, which may explain the relatively high mountainous relief of that region. Such an interpretation would suggest the existence of a significant crustal root down to depth of about 100 km (60 mi) below the Inner Carpathians, similar to the crustal root predicted for the Western Alps. According to Marchant and Stampfli (1997), the eclogized crustal root of the European plate overthrusted by the Adriatic crust and lithosphere extends below the Po plain.

Nemcok et al. (1998b, 2006) excluded the existence of a crustal root along the western part of the Carpathian arc. Instead, they proposed that the end of the subduction of the remnant oceanic flysch basin and the beginning of the collision were accompanied by the detachment of the subducting plate and by the occurrence of the break-off-related volcanism. The detachment began in the early Miocene in the western part of the Carpathian arc and propagated eastward to the Vrancea area of a presently active tear. Such an interpretation assumes the existence of a large oceanic domain in the Outer Carpathian Flysch basin. As previously discussed, the extent of the oceanic crust and lithosphere in the Outer Western Carpathian basin was apparently limited and confined mainly to the margins of the Penninic–Pieninic ocean south of the Czorsztyn ridge (Figure 6). The subduction of the limited oceanic realm would thus not compensate for the large shortening in the Outer Carpathian belt. Neither would the interpretation proposed by Nemcok et al. (1998b) satisfactorily explain the fate of the large portions of the continental lithosphere, which originally underlined the Outer Carpathian depositional system. These discussions indicate that the geodynamic reconstruction of the Carpathian region remains a challenging task, whose satisfactory solution will require additional geological and geophysical studies addressing both local and wider regional problems of the Alpine–Carpathian system of Europe.

Neotectonics and Topography of the Carpathian Region

In comparison with the Alps, the topographic elevation of the Carpathians above the sea is relatively low. According to Roure et al. (1993), the subdued relief of the Carpathians might be caused by a low level of collision (coupling) between the orogenic wedge and the foreland. Unlike the Alps, the Carpathians underwent two periods of subsidence and extension. The first one, marked by normal faulting and formation of the Inner Carpathian Paleogene basin on the top of the Mesozoic nappes, occurred in the middle Eocene to early Miocene in the northern part of the Inner Carpathian belt. The second, associated with the formation of the volcanic arc and the back-arc Pannonian Basin in the Miocene, affected southern zones of the Inner Carpathians.

The Outer Carpathian belt superimposed on the relatively stable European platform remained mostly unaffected by the postorogenic extension and subsidence. Only the southwestern part of the Carpathian belt, mainly confined to the extent of the Dyje–Thaya depression on the territory of northeastern Austria and southern Moravia (Figure 3), underwent a significant extension and subsidence most prominently marked by the opening of the Vienna basin. The formation of the Vienna basin during the Miocene was apparently related both to the transtensional tectonics in the Western Carpathian transfer zone and to the inherited weakness of the crust in the Dyje–Thaya Jurassic depression. The continuous subsidence of this area is documented by the accumulation of the Pliocene to Holocene alluvial deposits of the Dyje and Morava Rivers. The measurements of the horizontal dynamics in the contact zone between the Bohemian Massif and the Western Carpathians indicate that extension prevails in the southern part of the Carpathian foredeep, whereas compression dominates its northern part. The former is apparently associated with a wide-range subsidence of southern Moravia (Dyje–Thaya depression), the latter with the uplifting of the Beskydy and Jesenik Mountains of northern Moravia (Bucha and Blizkovsky, 1994).

Elsewhere outside the Vienna basin, the Flysch belt in southern Moravia is characterized by a low relief of

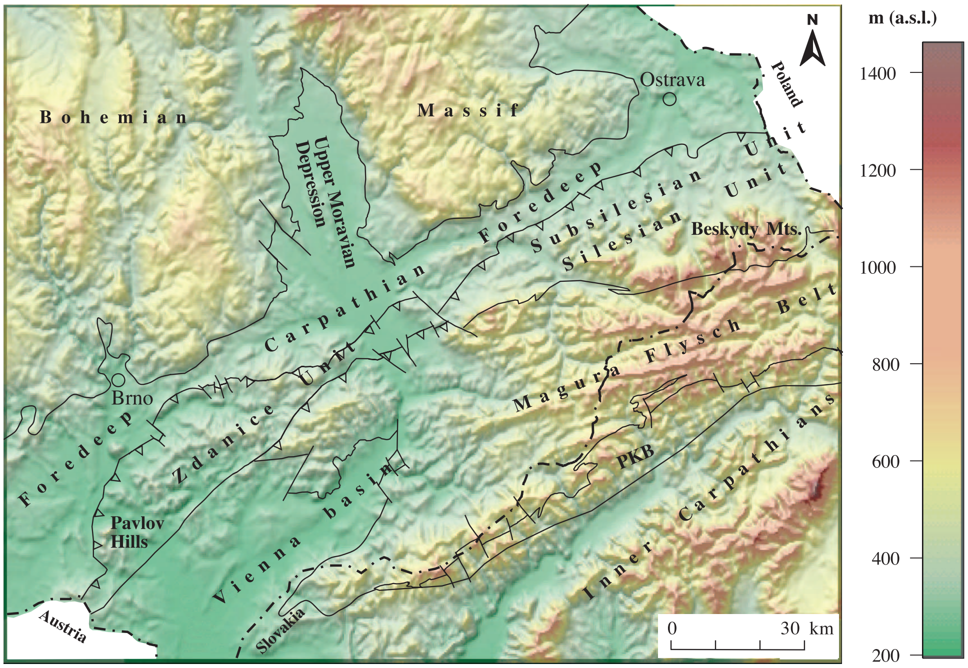

Figure 21. Topographic map of the Western Carpathians and their European foreland in Moravia.

rolling hills, from which only the carbonate klippen of Ernstbrunn and Pavlov Hills stand out morphologically. In central Moravia and especially in the Beskydy Mountains of northern Moravia, the relief of the Outer Carpathian Flysch belt is much more mountainous with the highest peak, the Lysa hora (Bold Mountain), reaching 1328 m (4356 ft). This relatively high relief might be explained by the combination of the crustal uplifting of the Beskydy Mountains (Figure 21) and the presence of thick, weathering-resistant Upper Cretaceous sandstone beds of the Godula and Istebna formations. The mountain ranges of the Flysch belt, in general, and the Magura flysch, in particular, trend mostly in the southwest–northeast direction, parallel with the thrust-related structures. Their topography is further enhanced by the alternation of packages of weathering more resistant sandstones and less resistant shales. The Pliocene to Holocene fluvial erosion dissected the uplifted ranges by deep northwest–southeast-running valleys.

The earthquake foci in the Western Carpathians, according to Bucha and Blizkovsky (1994), are confined to three main structural systems: (1) southwest–northeast-trending sliding surfaces between the crustal blocks; (2) northwest–southeast-trending faults in the underlying European platform; and (3) thrust planes and shears in the Carpathian thrust belt. The first two systems are apparently related to two primary structural trends, known from the Bohemian Massif, and expressed also by the southwest–northeast-directed Western Carpathian transfer zone and the northwest–southeast-trending Jurassic Dyje–Thaya depression, respectively. Most of the recorded earthquake activity is associated with the peri-Pieniny lineament of Maska (in Buday et al., 1961). The links between neotectonics and the continental topography in the Pannonian Basin and the Carpathian orogenic arc have become a subject of intensive studies (Bada et al., 2001; Cloetingh et al., 2002).

The Pleistocene continental glaciation reached into the northernmost part of the Carpathian foredeep in Moravia. The permafrost, to the depth of several tens of meters, formed elsewhere in the periglacial regions. The unstable slopes of the flysch mountain ranges are predisposed to extensive mass wasting and landsliding.

MAJOR DEPOSITIONAL SEQUENCES OF THE OUTER WEST CARPATHIANS

To better comprehend the geological history of the Outer Carpathians and their foreland and to integrate it into the evolution of the broader Tethyan–Carpathian system, attempts have been made to define the main depositional sequences and to relate them to major tectonostratigraphic events, such as rifting, major transgressions, episodic progradation of the Carpathian orogenic belt, and tectonic mobilization of the cratonic foreland (e.g., Picha and Stranik, 1999; Slaczka et al., 2006).

Based on such criteria, we have divided the Jurassic to Neogene strata of the Outer Western Carpathians in Moravia into six main sequences related to the critical tectonic and depositional events in the evolution of the Outer Carpathians (Figure 17A, B). The term sequence is used here in a broad sense, without any reference to the terminology of the sequence stratigraphy.

1) The oldest, Middle Jurassic (Dogger) to the Early Cretaceous [Berriasian–Valanginian(?)] sequence encompasses the oldest strata of the Tethyan–Alpine tectonic cycle found in the European foreland and in the Outer Carpathians in Moravia and northeastern Austria. Their depositional system was controlled by the Middle Jurassic rifting of the European platform, followed by the encroachment of the Tethyan Sea and development of the passive continental margins in the outer zone of the Carpathian system. The rifting of the European foreland in Moravia and northeastern Austria was predominantly confined to the northwest–southeast-trending Dyje–Thaya depression (Figure 5). The synrift phase was marked by deposition of the clastic Gresten Formation, followed by the development of carbonate platforms and basins on the rifted continental margins. A chain of carbonate buildups, known from the tectonic klippen such as Ernsbrunn, Pavlov Hills, and Stramberk, evolved along the shallow northwestern side of the rift basin (Figure 5). Included into this oldest sequence are the autochthonous Jurassic strata of the foreland in Moravia and northeastern Austria, the Outer Klippen of the Waschberg–Zdanice unit, the Jurassic strata of the Silesian unit including the Stramberk klippe, the Kurovice klippe of the Magura unit, and the Jurassic to Lower Cretaceous klippen of the Pieniny Klippen Belt. Reworked rocks of this sequence are commonly found elsewhere in the conglomerates of the Flysch belt.

2) The Early Cretaceous (Neocomian to Albian) sequence began with a primary change of the structural pattern of the Outer Western Carpathian region, which became dominated by the dextral motion along the Western Carpathian transform zone (Figure 5). Whereas the area west of this zone was uplifted and partly eroded, the sedimentation continued uninterrupted in the rest of the Tethyan continental margins east of this zone. The continuing rifting and extension resulted in a further differentiation of the Outer Carpathian depositional realm into a system of basins and ridges. In the Silesian basin, the further rifting and extension of the attenuated crust was accompanied by the occurrence of the teschenite volcanism. The Silesian ridge separated the Silesian basin from the Magura basin, which evolved on a highly attenuated continental and, according to some views, even oceanic crust (Figure 6). The Magura basin was then separated from the oceanic realm of the Pieniny Klippen Belt by the Czorsztyn ridge. The extension and crustal thinning in the Outer Carpathian realm peaked during the Early Cretaceous. In the Aptian–Albian, a widespread anoxic event, marked by the deposition of deep-water black shales, silicites, and turbidites, occurred in the entire Tethyan system. In the Western Carpathians, this anoxic event is evidenced by the deposition of the Verovice and Lhoty formations in the Silesian basin and the Rajnochovice Formation and Gault flysch in the Magura and the Rhenodanubian Flysch, respectively.

3) The Late Cretaceous (Cenomanian) to the late Eocene (Bartonian) sequence reflects the continuous deformation of the Inner Carpathians and their collision with the Czorsztyn ridge and other fragments of the European plate. In the Outer Carpathian realm, this was marked by a gradual transition from a divergent to a convergent phase and development of a progradational foreland basin. The compressional stresses at the Albian–Cenomanian transition (formerly the Austrian orogeny) caused the sudden rise of the intrabasinal ridges, which shed big amounts of clastics into the adjacent deep-water flysch basins. In the Cenomanian, this process was combined with the global transgression (high water stand), during which the epicontinental shallow seas spread over vast territories of the continents, including the Carpathian foreland in Moravia. In the Campanian–Mastrichtian, the Alpine–Carpathian seaway progressed over the European foreland to form the Waschberg–Zdanice–Subsilesian basin (Figure 5). At the Cretaceous–Paleogene transition, during the so-called Laramide orogeny, the European foreland adjacent to the Outer Carpathian basin was uplifted, deeply weathered, and eroded. The magnitude of the uplifting and erosion is well documented by the incision of two large paleovalleys in southern Moravia. The hemipelagic variegated (commonly red) muds, derived mainly from the weathered platform, were deposited elsewhere in the external units of the Outer Carpathians. At the same time, deep-water turbiditic sedimentation prevailed in the Silesian and Magura flysch basins supplied with clastics from the Silesian cordillera. Characteristic for the Paleocene to middle Eocene variegated muddy deposits of the external units is the occasional presence of thick, channelized sands, which are significant reservoirs, especially in the Silesian unit of Poland.

The Late Cretaceous to late Eocene sequence represents a large part of the depositional stack of the Magura flysch (Kaumberg, Solan, Beloveza, and Zlin formations), of the Silesian unit (Godula, Istebna, and

Roznov formations of the Godula subunit and their equivalents in the Kelc and Baska subunits), and of the Waschberg–Zdanice–Subsilesian units (Nemcice and Frydlant formations) (Figure 17A, B). In addition, the lower part of the autochthonous fill of the Nesvacilka paleovalley containing the questionable Late Cretaceous to Paleocene fauna might belong to this sequence. The Late Cretaceous to late Eocene sequence of the Outer Carpathians also encompasses most of the Helveticum and the Rhenodanubian Flysch of the Austrian, Bavarian, and Swiss Alps.

4) The late Eocene (Priabonian) to early Oligocene (Rupelian) sequence is associated with the extensive late Eocene marine transgression over the European foreland, deepening of the external Carpathian basins, submergence of inner ridges, and the termination of sedimentation in the Magura flysch. Euxinic facies developed in the Carpathians and throughout the entire Tethyan region from the Western Mediterranean to the Black and Caspian Seas region and beyond.

In the Alpine sector, the late Eocene transgressive event marks the formation of the Molasse basin with the deposition of a variety of lagoonal and marine facies, including the Lattorfian Fish shales and Rupelian bituminous marls of the lower marine Molasse in Austria and Bavaria (Bachmann and Muller, 1991; Wagner, 1996). The Menilitic cherts, typical for the Carpathians, are not found in the Alpine Molasse. The more advanced collision and thickening of the crust in the Alpine region apparently prevented more substantial deepening of the Molasse basin needed for upwelling and accumulation of silica-forming organisms, such as diatomites (Figure 22A).

In the Moravian sector of the Western Carpathians, the late Eocene marine transgression reached into the most external zones of the Pouzdrany and the Waschberg–Zdanice–Subsilesian basins, and the paleovalleys and submarine canyons were filled with hemipelagic deposits, which, according to Picha and Stranik (1999), exemplify the continuation of the lower marine Molasse of the Alps into the Carpathian realm (Figure 22A). The late Eocene to early Oligocene sequence in Moravia extends from the autochthonous fill of the Vranovice and Nesvacilka paleovalleys into the Pouzdrany and Uhercice formations, and the Boudky Marls of the Pouzdrany unit, the *Globigerina* (Sheshory) Marls, and the Menilitic Formation of the Waschberg–Zdanice–Subsilesian and Silesian units and their equivalents in the Fore-Magura and Cejc Zajeci units to the youngest strata of the Zlin Formation in the outer zone of the Magura flysch (Figure 17A, B). In the territory of Slovakia and Poland, equivalents of these strata are also found in the Grybow, Obidowa Slopnice, and other window units. In the Inner Carpathian Paleogene, the Lutetian transgressive shallow-marine facies was followed by the deep-water shaly deposits, including some Menilitic cherts and manganese ores (Picha, 1964a). Equivalents of these strata are also present in the Hungarian Paleogene basin, known as the euxinic Tard Clay (Tari et al., 1993).

The early Oligocene, Menilitic cherts, and other organic-rich shales and marls and their equivalents elsewhere in the Tethyan region represent a significant correlation horizon. They were deposited in a stagnant environment with a very limited influx of coarser clastics both from the submerged orogenic belt and from the foreland. Typically, these deposits contain a high content of organic matter, which makes them one of the most important sources of hydrocarbons in the entire Tethyan–Alpine system from Western Europe to the Caucasus, the Caspian region (Maykop Formation), the Himalayas, and Southeast Asia.

5) The late Oligocene to early Miocene (Chattian–Aquitanian, Egerian) sequence was deposited in a system of foredeeps, which formed in the late Oligocene to early Miocene throughout the Carpathian orogenic system. Their formation was linked to the uplifting and deformation of the inner zones of the Outer Carpathians, e.g., the Pieniny Klippen Belt and the Magura flysch, and to the flexural downbending of the foreland crust (Figures 6C, 22B). The sequence is dominated by the deep-water turbiditic Krosno-type flysch facies, supplied with clastic material predominantly from the folded and uplifted inner zones of the Magura flysch. In Moravia, these deposits are represented by the Krepice Formation of the Pouzdrany unit and its equivalents, the Eisenschussige clays and sands (Grill, 1962) of the Waschberg sector in Austria, the Zdanice–Hustopece Formation of the Waschberg–Zdanice–Subsilesian unit (the Auspitzer Mergel in Austria; Grill, 1962), and the Krosno Formation of the Silesian unit. Similarities also exist between the Zdanice–Hustopece Formation of southern Moravia and the deep-water turbidite facies of the Puchkirchen beds of the Bavarian Molasse (Bachmann and Muller, 1991). Further west in Switzerland, this turbiditic facies gradually passes into nonmarine deposits of the lower fresh-water Molasse (Schlunegger et al., 1997) (Figure 22B). In the Inner Carpathians, a major deep-water flysch basin, characterized by the turbiditic sedimentation and transport of the clastic material from the uplifted zones of the Magura flysch and the Pieniny Klippen Belt, evolved in the late Oligocene to early Miocene. Its deepest and tectonically most active part was adjacent to the Pieniny Klippen Belt zone. The Krosno-type flysch deposits of the late Oligocene to early Miocene

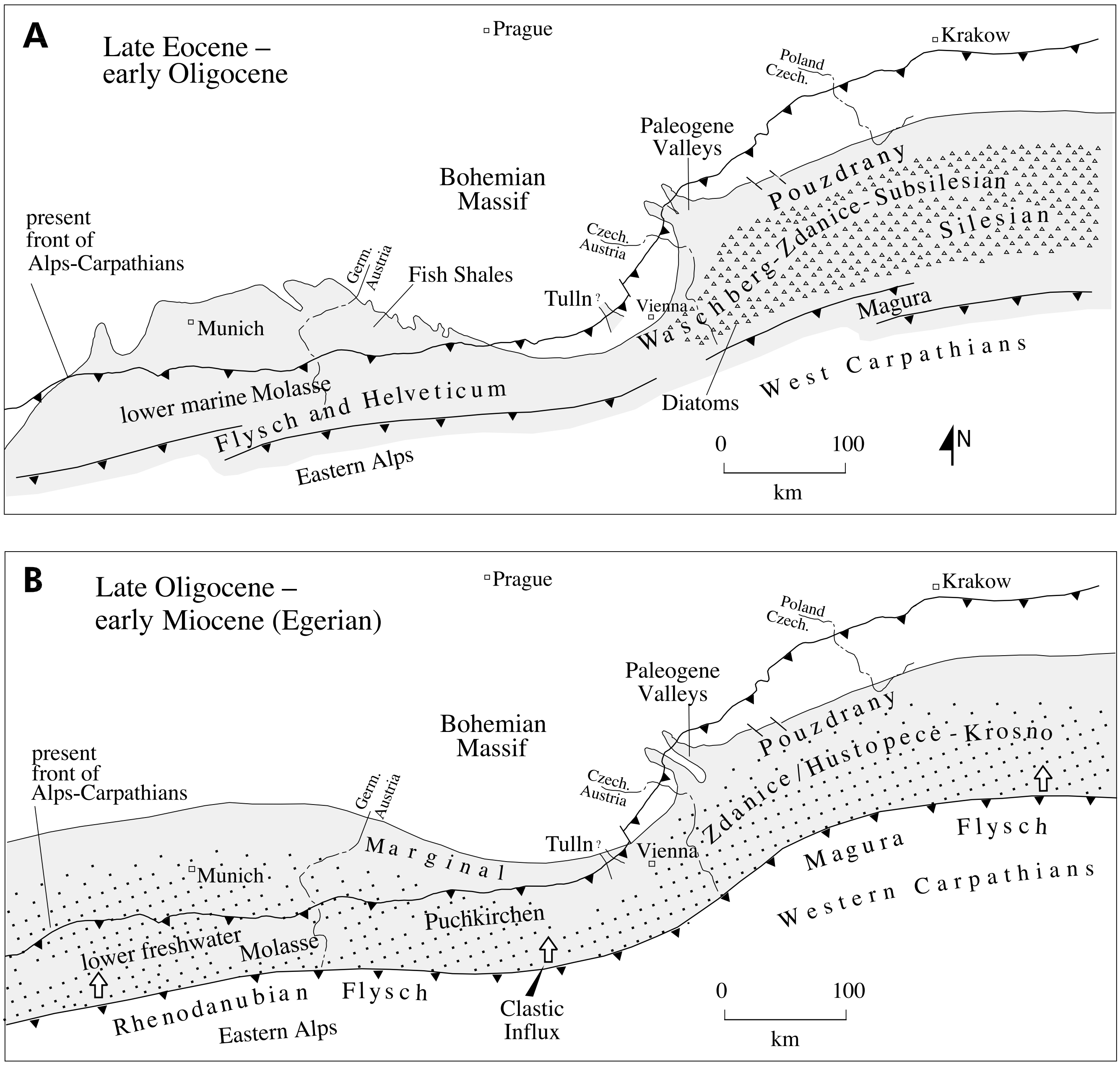

Figure 22. Correlation of (A) late Eocene to early Oligocene and (B) late Oligocene to early Miocene strata in the Eastern Alps and Western Carpathians (Picha and Stranik, 1999, reprinted with permission).

age, typically present in the external zones of the Outer Carpathians and the Inner Carpathian Paleogene basin, represent the youngest deep-water flysch deposits found anywhere in the Western Carpathians.

6) The early Miocene (Eggenburgian) to Pliocene sequence represents the youngest stage of development of the Tethyan–Alpine system. It is related to final phases of folding and thrusting of the Outer Carpathians and the development of a system of the molasse-type foredeeps on the European foreland, referred to as the Paratethys (Laskerev, 1924). Within the Carpathian thrust belt, intramontane basins, such as the pull-apart Vienna basin and the back-arc Pannonian Basin, formed as an outcome of the late orogenic strike-slip (escape) tectonics and back-arc extension and subsidence. The molasse-type, shallow-marine, and continental deposits of this sequence reflect both on the tectonic activity of the Carpathian orogenic belt and the global sea level oscillations. In Moravia, this sequence is represented by the whole depositional fill of the Neogene foredeep and the Vienna basin and also by the youngest members

of the Waschberg–Zdanice–Subsilesian unit, such as the Eggenburgian Sakvice Marls, the Ottnangian and Karpatian Pavlovice and Laa formations, and their equivalents in the Pouzdrany unit (Figure 17A, B). All these strata may be correlated with the deposits of the upper marine and upper nonmarine Molasse of Bavaria and Austria and with other coeval strata of Paratethys elsewhere.

THE HYDROCARBON RESOURCES OF THE CARPATHIAN THRUST BELT AND ITS FORELAND IN MORAVIA

The first attempts to find hydrocarbons on the territory of Moravia go back to the turn of the 19th century. In 1899–1900, J. May drilled the first well to 450.7 m (1478.6 ft) and the second well in 1902 to 645.2 m (2116.7 ft) in the Flysch belt near Bohuslavice nad Vlarou without any apparent economic success, although the presence of oil and natural gas was confirmed. In 1900, the first well drilled in the Vienna basin near Hodonin to a depth of 217 m (711 ft) encountered noncommercial shows of oil and gas. Natural gas produced from a well was, for the first time, used as a fuel in a sugar mill in Slavkov (Austerlitz) in 1908. The first commercial oil in the Vienna basin was found in 1914 at Gbely (now in the territory of Slovakia) at the depth of 164 m (538 ft). In 1919, oil was discovered in two horizons, Sarmatian and upper Badenian, near the town Hodonin. This was followed by discoveries of the Vacenovice field in the Vienna basin and the Zatcany and Sokolnice fields in the shallow part of the Carpathian foredeep. The production of oil in the former Czechoslovak Republic reached 10,110 t in 1920 and 22,796 t in 1930. The first official report about the production of natural gas (242,300 m^3; 8,556,742 ft^3) was issued in 1930. During World War II, the intensified exploration resulted in discoveries of the oil and gas fields Bilovice, Zizkov, Luzice, Tynec, Kostice, and Breclav in the Vienna basin.

After the Second World War, a major exploration effort by the newly established state oil company Moravske naftove doly led to discoveries of several small and midsized fields, e.g., Brodske, Poddvorov, and Mutenice, and, in 1959, of the largest oil and gas reservoir in the Moravian part of the Vienna basin, the Hrusky field. The Hrusky field and its extension, the Josefov field, held initial reserves of 15.5 million bbl of oil and 90 bcf of gas. The Vienna basin remained the main producer of hydrocarbons in Czechoslovakia until the end of 1970. In the 1960s, a pioneering deep drilling project, conducted initially by the Czechoslovak Geological Survey, revealed the presence of hydrocarbons in various stratigraphic horizons in the foreland plate of the Bohemian Massif underneath the thin-skinned Carpathian thrust belt in Moravia. The exploration activity gradually moved from the Vienna basin to these deeper subthrust plays, where several fields, such as Kostelany, Damborice, Uhrice, Zdanice, and Zarosice, have been found to date. For more information on some of these fields, see the account by Kostelnicek et al. (2006). Following the development of some of these fields, the production of oil in the Czech Republic increased to more than 6000 bbl/day (more than 315,000 t/yr) in 2002. The remaining undrilled potential of the subthrust play in Moravia is still substantial, especially at greater depth. Most recently, the application of the new seismic techniques, such as 3-D survey, enabled the reassessment of the previously explored plays in the Vienna basin and led to additional discoveries of oil and gas even in this mature area.

To date, about 5.5 million t (38.4 million bbl) of oil and 9 billion m^3 (320 bcf) of natural gas have been produced from all the fields in the territory of Moravia. An additional 7.5 billion m^3 (265 bcf) of coal gas from the degasification of the coalfields was produced in the Moravian part of the Upper Silesian basin by 1998. However, several attempts to produce economic quantities of coalbed methane from the coal seams of the Upper Silesian basin, in the territory of both the Czech Republic and Poland, so far have not been successful. Some of the depleted oil and gas fields, e.g., Dunajovice, Hrusky, Bojanovice, and Uhrice, have been converted into gas storages.

Practically all the oil and gas fields discovered at the territory of Moravia are listed in Tables 2–4 (shown on pages 136, 139, and 143, respectively), and their locations are shown in Figure 25, shown on page 135.

Numerous papers and internal reports have been written about various aspects of the hydrocarbon prospectivity of Moravia; however, only a fraction of them are available in the public domain and published in English. Among them, the recent publications by Durica et al. (1986), Krejci (1993), Krejci et al. (1994, 1996), Ciprys et al. (1995), Dorman (1995), Brzobohaty et al. (1996), Francu et al. (1996), Picha (1996), and Picha and Peters (1998) seem to be the most informative and relatively easily accessible to readers.

THE HYDROCARBON SYSTEMS OF THE CARPATHIAN THRUST BELT AND ITS FORELAND IN MORAVIA

In Moravia, hydrocarbons have been found in the Neogene foredeep, the Vienna basin, the Flysch belt, and the subthrust European platform. Their genesis is related to a variety of petroleum systems, which exist in this relatively small but geologically complex area. As defined by Demaison and Huizinga (1994), among others, the concept of hydrocarbon systems integrates all components involved in the generation, migration, accumulation, and preservation of hydrocarbons. In principle, that includes source rocks, reservoirs, seals, and traps.

The Source Rocks

In the territory of Moravia and lower Austria, potential source rocks are found both in the European foreland plate and in the Carpathian thrust belt.

The Source Rocks in the European Foreland Plate

In the European foreland plate, organic-rich deposits occur in the Middle Devonian, Upper Carboniferous, Jurassic, and Paleogene (Figures 23, 24).

The generative potential of the Paleozoic source rocks, such as the Devonian carbonates (e.g., the Lazanky Limestones) and the Carboniferous shales, is still little understood. Some Carboniferous rocks, especially those associated with the coal-bearing sequences of the Upper Silesian basin, have a fair source rock potential. On average, they contain 1.2 wt.% of the predominantly gas-prone type III kerogen (Krejci et al., 1994). The Carboniferous coal beds evidently sourced some accumulations of gases in northern Moravia. However, the generative potential of these mostly postmature Paleozoic rocks seems to be limited to areas where the source rocks did not pass into the generative window prior to the emplacement of the Carpathian nappes and the formation of the Neogene foredeep (Figure 23). According to Lafargue et al. (1994), the possibility that the organic-rich Paleozoic rocks may have sourced some potential deep subthrust plays in the Carpathians should not be dismissed entirely.

The Upper Jurassic (Malmian) organic-rich Mikulov marls found in the Jurassic Dyje–Thaya depression of southern Moravia and northeastern Austria (Figures 23, 24) represent a world-class source rock whose generative potential has been well established (Ladwein, 1988; Krejci et al., 1994; Francu et al., 1996; Picha and Peters, 1998). These marls, as much as 1500 m (4900 ft) thick, were deposited in a marine basinal restricted environment with a high rate of accumulation of the predominantly algal-derived type II organic matter. The total organic carbon (TOC) of these rocks ranges from 0.2 to more than 10.0 wt.% and averages around 1.9 wt.% (Ladwein et al., 1991; Krejci et al., 1994). These Jurassic organic-rich rocks sourced most of the oils in the Vienna basin (Ladwein, 1988) and in the subthrust foreland plate of southern Moravia and northeastern Austria (Ciprys et al., 1995; Francu et al., 1996; Zimmer and Wessely 1996; Picha and Peters, 1998).

The Jurassic black shales and marls are also locally found in the Polish and Ukrainian part of the Carpathian foreland. High organic contents (2.0–3.0 wt.% of TOC) and hydrogen index values between 100 and 400 mg hydrocarbons (HC)/g TOC in these potential source rocks have been reported, e.g., by Lafargue et al. (1994).

The organic-rich shales and marls of the autochthonous Paleogene represent the second most important source rock in Moravia (Figures 23, 24). They are found predominantly in the Paleogene fill, as much as 1500 m (4900 ft) thick, of the two paleovalleys and submarine canyons (Picha 1979a, 1996). The TOC content in these rocks ranges between 1 and 9 wt.% and averages around 3 wt.% (Brzobohaty, 1993; Krejci et al., 1994). The organic matter is composed mostly of the terrigenous gas-prone type III kerogen with a varying amount of the oil-prone type II kerogen of algal and bacterial origin, whose proportion, however, may increase further down the slope away from the river estuaries. The Paleogene source rocks apparently sourced the Krumvir and Karlin gas fields in southern Moravia and possibly the oil fields Kostelany, Krasna, and Dolni Lomna in central and northern Moravia (Figure 25, shown on page 135). The organic-rich autochthonous Paleogene deposits of southern Moravia are partly coeval with the organic-rich Menilitic shales of the Carpathian flysch belt and the Lattorfian (Priabonian) Fish shales (Fischschiefer) of the Alpine Molasse in Bavaria and Austria. The Fish shales, as much as 100 m (330 ft) thick, contain 1–8 wt.% TOC dominated by the type II kerogen. They pass into a generative window at a depth of 4000 m (13,000 ft), mainly in the zone buried below the Alpine thrust belt (Schmidt and Erdogan, 1996; Wagner, 1996; Zimmer and Wessely, 1996). The Fish shales are considered to be the main source rocks for the oil and gas fields in the Alpine Molasse of Austria. Until the end of 1992, the cumulative production from those fields amounted to 56 million bbl of oil and 12.5 billion m^3 (462 bcf) of gas (Wagner, 1996).

The organic-rich autochthonous and parautochthonous marginal Paleogene strata extend from the Alpine Molasse of Bavaria and Austria to southern Moravia, where they fill two paleovalleys, and then possibly below the Flysch belt to the Black Sea and further east through the Caucasus to the Caspian Sea. The potential existence of these organic-rich deposits below the Outer Carpathian belt may enhance the prospectivity of the entire Carpathian region. The hydrocarbons generated from these rocks could have

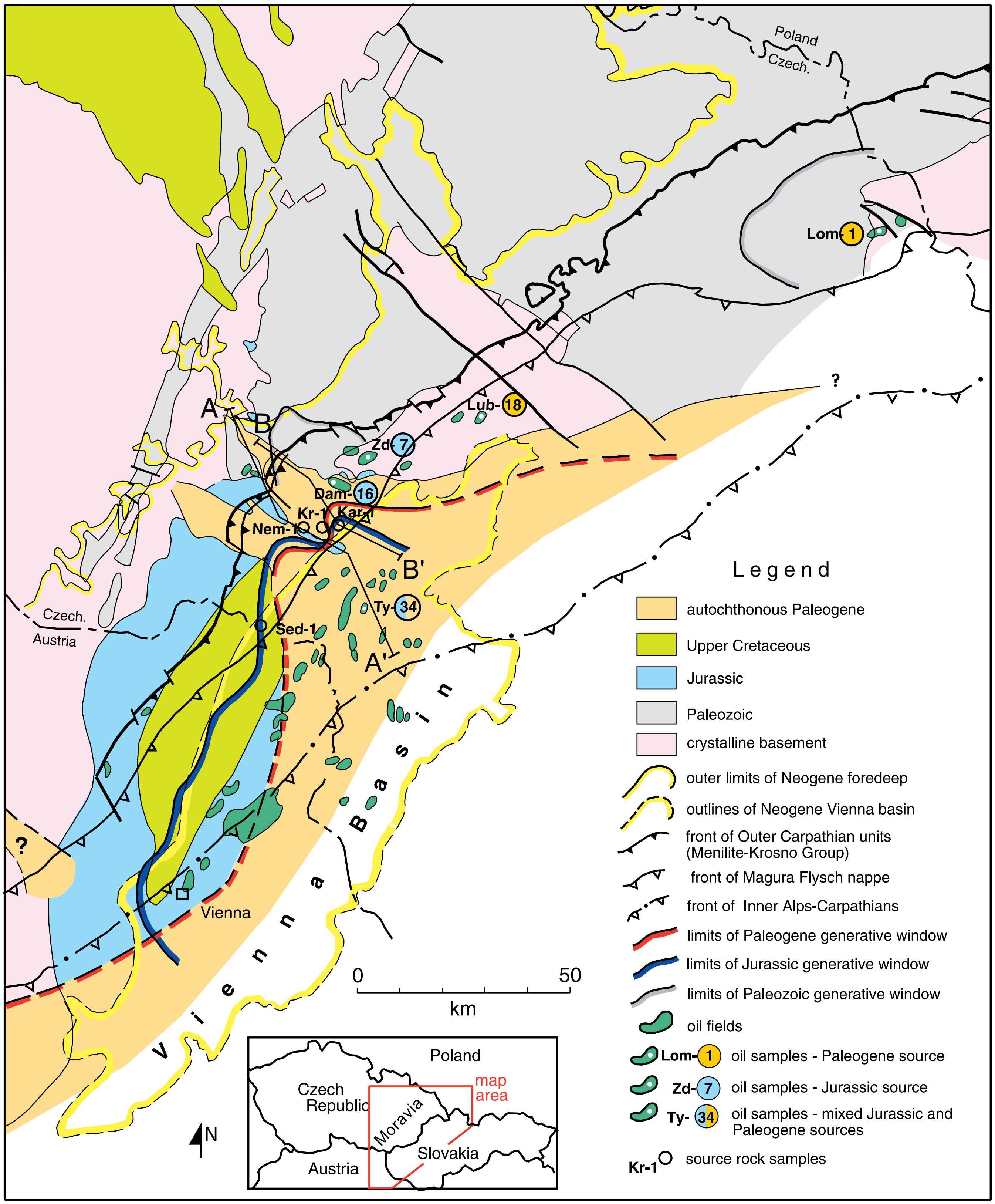

Figure 23. Pre-Neogene subcrop geological map of the European foreland plate in Moravia and northeastern Austria showing the distribution of Paleozoic, Jurassic, and Paleogene strata with potential source rocks. Oil samples (e.g., Lom-1) analyzed for biomarkers (modified from Picha and Peters, 1998).

migrated updip into potential reservoirs both in the foreland subthrust plate and possibly along the faults into the thrust belt and successor basins (Picha, 1996).

The Neogene foredeep in Moravia has practically no organic-rich source rocks; moreover, the organic material in the Neogene deposits is immature. As reported by Kotarba et al. (1987) from the Polish part of the foredeep, the organic matter is mainly of terrestrial origin (type III) and shows only a feeble degree of maturation.

The Source Rocks in the Carpathian Thrust Belt

Two potential source rocks in the Western Carpathian thrust belt are related to two prominent euxinic events in the Tethyan region: (1) the Lower Cretaceous (Aptian–Albian) event marking the time of the maximum divergency in the Outer Carpathian depositional realm and (2) the lower Oligocene euxinic event related to the transgression over the foreland and the Inner Carpathians and the deepening of the foreland basins combined with a period of global cooling and proliferation of siliceous phyto- and zooplankton (Figure 17A, B).

The Lower Cretaceous (Aptian–Albian) black shales and cherts are prominently developed in the Silesian basin, but their equivalents are also found in the Magura and Rhenodanubian Flysch (Rajnochovice Formation and Gault Flysch, respectively) and in the external Carpathian units of eastern Poland and Ukraine (Koltun et al., 1998). All these rocks are organic rich, but because of the low hydrogen index, their generative potential seems to be limited. According to Lafargue et al. (1994), the average TOC value of the Lower Cretaceous black shales is 2.5 wt.%, and their average hydrogen index is 150 mg HC/g TOC. Likewise, ten Haven et al. (1993), in a sample from the Polish Carpathians, report a high TOC value of 3.47 wt.% but a low hydrogen index of only 53 mg HC/g TOC. According to Slaczka (1996) and Koltun et al. (1998), at least in some areas, the Lower Cretaceous shales passed into the generative window prior to the thrusting and may thus be considered as potential source rocks.

The early Oligocene Menilitic shales and cherts of the Waschberg–Zdanice–Subsilesian, Silesian, and other external units have long been recognized as the most important source rock in the entire Carpathian belt. They are partly coeval with the autochthonous marginal Paleogene deposits of the subthrust plate, but unlike the autochthonous Paleogene, the Menilitic shales were detached, incorporated into complex structures of the thrust belt, and tectonically transported over the foreland (Figure 19). In Moravia, the organic-rich laminated siliceous shales, cherts, marls, and micritic limestones of the Menilitic Formation are only about 20 m (66 ft) thick. The organic matter present in these rocks is predominantly marine, with only a limited terrigenous input. According to ten Haven et al. (1993), five extracts of the Menilitic shales from the Polish Carpathians contain 2.15–22.15 wt.% of TOC, with hydrogen indices from 300 to 650 mg HC/g TOC. Lafargue et al. (1994) published similar results, 2–20 wt.% TOC (average 6%) with 300–600 mg HC/g TOC, from Menilitic shales in Poland, Ukraine, and Romania. Of all the studied source rocks in Moravia, the Menilitic shales have the highest liptinite contents (Francu et al., 1996). If buried into the oil window, they would generate significant amounts of oil.

Geochemical studies by Koltun et al. (1998) indicate that in the prolific Boryslav–Pokuttya unit of the Ukrainian Carpathians, the oil window for the Menilitic Formation (assuming the geothermal gradient 24°C/km) lies at a depth between 4.2 and 5.8 km (2.6 and 3.6 mi) with the peak oil generation at 5.3 km (3.3 mi). The maximum depth at which oil (with <50° API) may occur is 6.9 km (4.2 mi), whereas wet gas may be found to the depth of 7.1 km (4.4 mi).

The organic-rich rocks of the Menilitic Formation in the Waschberg–Zdanice–Subsilesian and Silesian units in Moravia remain mostly immature. However, lateral migration of hydrocarbons from the more mature rocks of the Menilitic Formation present in thrust sheets and duplexes of external units underneath the Magura nappe is a real possibility. The biomarker analyses by Picha and Peters (1998) confirmed that the Paleogene organic-rich rocks of the Menilitic Formation or of the autochthonous Paleogene sourced oils in the Lubna (Kostelany), and Dolni Lomna fields central and northern Moravia (Figures 23, 25 [shown on page 135]).

Migration

The petroleum systems of the Carpathian region in Moravia are primarily associated with the Jurassic and Paleogene source rocks, the latter including both the autochthonous Paleogene of the foreland plate and the Menilitic Formation of the thrust belt. The oils from the Vienna basin (Tynec-34 well) appear to be sourced from both the Jurassic and Paleogene source rocks (Picha and Peters, 1998).

Oils and gases generated from the deeply buried source rocks in the subthrust European plate migrated both laterally into the foreland, where they supplied several oil and gas fields, and vertically through the thrust belt into the Miocene reservoirs in the Vienna basin (Figure 24). The lateral migration was probably helped by the existence of impermeable shaly deposits at the base of the Carpathian thrust belt, which

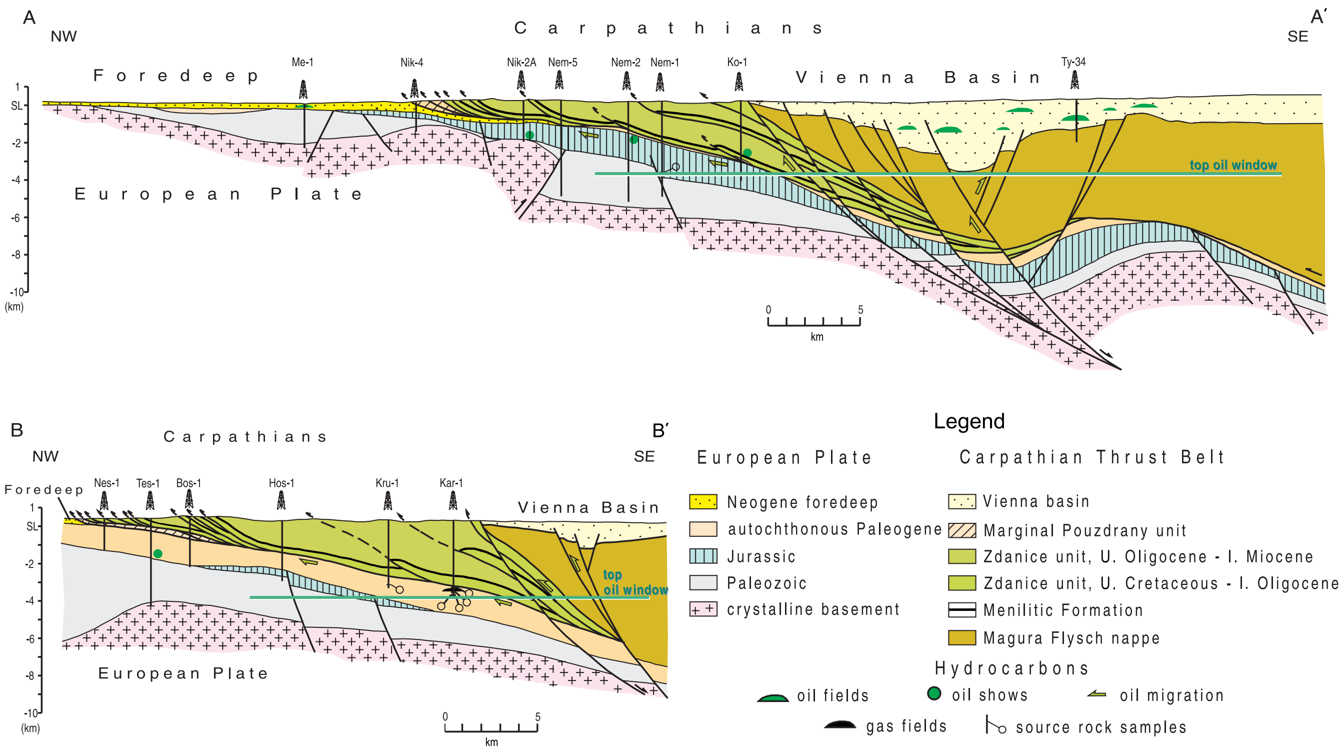

Figure 24. Cross sections AA′ and BB′ (location in Figure 23) through the West European plate, the overlying Outer Western Carpathian thrust belt, and the successor Vienna basin showing the top of the generative window for the organic-rich source rocks in the Jurassic and Paleogene strata of the subthrust plate and the Menilitic Formation of the thrust belt. With the exception of deep zones below the thick Magura nappe, the Menilitic organic-rich shales and cherts of the Zdanice unit remain mostly immature in the frontal zones of the Flysch belt in Moravia. The cross section BB′ passes through the Paleogene fill of the Nesvacilka paleovalley and submarine canyon. In the Vienna basin area, some of the Jurassic rift-related faults were apparently reactivated as transtensional faults during the pull-apart stage of the evolution of the basin in the middle Miocene and served as conduits for migration of hydrocarbons from the Jurassic and Paleogene source rocks of the subthrust plate through the thick flysch sequences of the thrust belt into the Neogene reservoirs of the Vienna basin. Modified after Picha and Peters (1998). Stratigraphic records of some of the wells shown on cross sections are reported in Appendix 1.

acted as a regional seal and prevented vertical escape and dispersal of hydrocarbons. The distribution of hydrocarbon occurrences in the foreland plate and the Neogene foredeep indicates that the tectonically controlled Paleogene valleys may have served as conduits for migration of hydrocarbons from the deeper zones of the Carpathian system into the foreland (Picha, 1979a). The vertical migration from the deep Jurassic and Paleogene source rocks in the subthrust plate through the thick flysch belt into reservoirs in the Neogene Vienna basin was most likely enabled by the continuing activity of the late orogenic wrench faults, such as the Steinberg fault. The strike-slip motion along these orogen-parallel deep faults started the pull-apart stage of the Vienna basin. In a transtensional regime, these faults may have served as conduits for the vertical hydrocarbon migration, whereas in the transpressional regime, these strike-slip faults may have actually sealed some of the reservoirs. The proper timing of generation and migration of hydrocarbons with respect to the tectonic regime on these faults seems to be critical for the accumulation and preservation of hydrocarbons in the reservoirs (Picha and Peters, 1998).

Some short-range migration from mature source rocks, e.g., the organic-rich Menilitic shales, into nearby reservoirs, e.g., the Kliwa Sandstones or the Paleocene–Eocene channelized sandstones, apparently also occurred in the Carpathian thrust belt. According to Lafargue et al. (1994), this type of migration was a significant factor in the deeper parts of the Eastern Carpathian thrust belt in Ukraine and Romania.

The Reservoirs

In Moravia, proven reservoirs have been found in the European foreland plate, the Neogene foredeep, the Carpathian thrust belt, and the Vienna basin. In the

European foreland plate, oils are reservoired in the fractured and weathered surface of the Precambrian granitic rocks, in the Devonian to Carboniferous carbonates, in the Jurassic clastics and carbonates, and in the clastics of the autochthonous Paleogene and the Neogene foredeep (Figure 9). Numerous sandstones are present in the Carpathian Flysch belt, which might have good reservoir properties, but so far, no significant accumulations of hydrocarbons have been found in these rocks in Moravia. Potentially, the Paleocene to Eocene channelized sandstones enclosed in the impermeable claystones in the external units of the Carpathian thrust belt seem to be the most promising targets. Similar sandstones (Ciezkowice Sandstones) are proven producers in the Polish Carpathians. Good, predominantly clastic reservoirs exist in the Vienna basin, of which the middle Miocene (Badenian and, to a lesser degree, Sarmatian) sands and algal (lithothamnian) limestones are the most prolific.

The character of these various reservoirs is further discussed in the context of the next section on the various types of hydrocarbon plays.

THE HYDROCARBON PROVINCES, PLAYS, AND FIELDS

During the long history of exploration, several distinct hydrocarbon plays have been identified in the Neogene foredeep, the autochthonous European plate, the thin-skinned Carpathian thrust belt, and the Vienna basin. Numerous oil and gas fields have been discovered in those various settings in the territory of Moravia. They are listed in Tables 2, 3, and 4 (shown on pages 136, 139, and 143, respectively) and their locations are shown in Figure 25.

The Neogene Foredeep

In comparison with the Molasse Basin of Bavaria and Austria, the Neogene foredeep in Moravia is relatively shallow. The average thickness of the Neogene strata ranges within a few hundred meters. Besides the biogenic gas, the foredeep does not have any generative potential of its own. The small- and middle-sized oil and gas fields (Figure 25; Table 2), found in the clastic reservoirs of the Neogene foredeep, were sourced predominantly from the older, organic-rich, and deeply buried formations. In northern Moravia, the gas fields were supplied from the underlying coal-bearing Upper Carboniferous strata; in central and southern Moravia, the hydrocarbons migrated into the reservoirs in the foredeep from the Jurassic and Paleogene source rocks buried deeply below the Carpathian thrust belt.

The largest field found in the Neogene foredeep in Moravia is the Dolni Dunajovice gas field, discovered in 1973. It is an antiformal structure associated with the Vestonice fault, which cuts through the crystalline basement and Jurassic strata into the lowermost Miocene deposits of the foredeep (Figure 26). The fault is interpreted as an antithetic Neogene fault related to flexural downwarping of the foredeep (Adamek, 1979) or as a foreland-type compressional and possibly transpressional structure associated with the propagation of compressional orogenic stresses in the foreland plate, the interpretation we prefer. The gas is reservoired in Eggenburgian basal sandstones draping over the basement structure and sealed by the overlying Eggenburgian Marls. More than 810 million m^3 (28.5 bcf) of gas had been produced from the field prior to its conversion into a gas storage in 1989. For more information on the Dolni Dunajovice field, see Kostelnicek et al. (2006). On the border with Austria, another gas field, the Alt Prerau–Novy Prerov, was found in clastic rocks of the lower Miocene and produced from the Austrian side.

The very shallow (50 m; 160 ft) oil fields Menin and Zatcany (Figure 25) contain heavy biodegraded oils in the Karpatian and Badenian sands. The fields are situated at the margin of the Nesvacilka paleovalley, which apparently served as a migration path from the generative centers below the Carpathian belt into the foreland. The heavy oils from the Menin–Zatcany fields were partly exploited by the surface mining operations.

Hydrocarbons have been also found in the coastal and lagoonal Karpatian and possibly Ottnangian (Salaj, 1996) sands deposited at the slopes of the crystalline basement elevations in central Moravia. These reservoirs are hydrodynamically connected with the fractured reservoirs in the crystalline basement rocks. Commercial reserves of hydrocarbons in these Karpatian sands have been found in the Zdanice, Mourinov, Korycany, Kostelany (Lubna) and Rusava fields. Small, noncommercial occurrences of hydrocarbons in the Karpatian strata not associated with the reservoirs in the crystalline basement rocks have been encountered in numerous wells, e.g., Kozusice-4, Mikulov-2, Jezov-3, and Rostin-2.

In northern Moravia, the gas fields in the Neogene foredeep are linked to the migration of gases from the underlying Upper Carboniferous coal-bearing strata of the Upper Silesian coal basin. The largest of them, the Horni Zukov field, has produced more than 1.1 billion m^3 (39 bcf) of gas. It is followed by the Pribor–Klokocov field (1.0 billion m^3; 35 bcf), the Choryne field (640 million m^3; 21 bcf), and the Stramberk field (600 million m^3; 21 bcf)

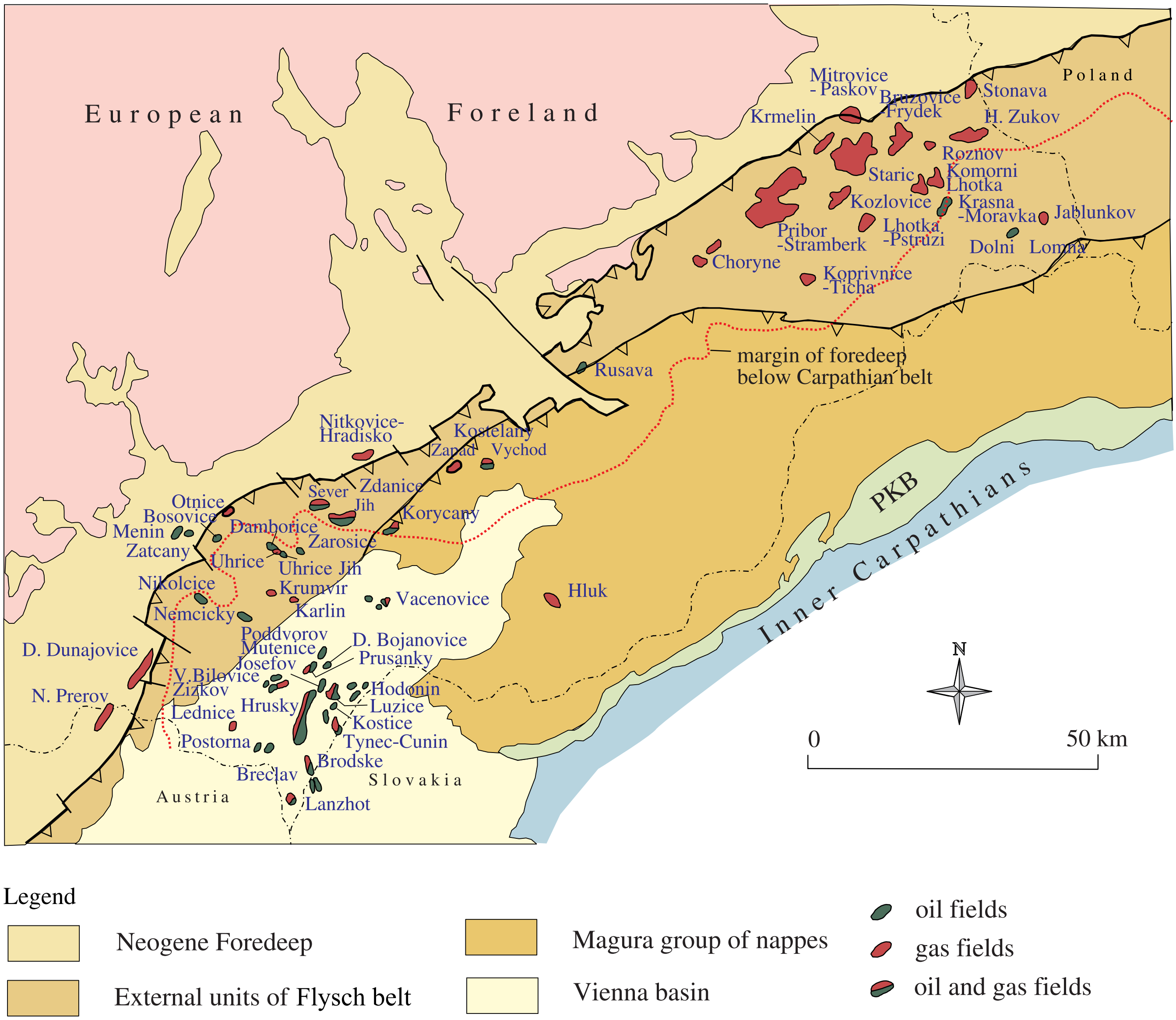

Figure 25. Map of oil and gas fields in the Neogene foredeep, the Western Carpathian thrust belt, the Vienna basin, and the European subthrust plate in Moravia. PKB = Pieniny Klippen Belt.

(Table 2). With the exception of the Horni Zukov field, which is reservoired in the lower Badenian, all the other fields mentioned above are located in the Karpatian reservoirs. The Horni Zukov and Stramberk gas fields have been converted into the gas storages, with potential capacity of 400 and 420 million m^3/yr (14.1 and 14.8 bcf/yr), respectively. Natural gas was also found in the Karpatian strata resting marginally on the Devonian to Lower Carboniferous carbonates in the Krasna oil field.

The Neogene foredeep, although heavily drilled, may still have some additional hydrocarbon potential, especially in the subthrust zone around the basement elevations and in the fault-related traps. In addition, the depositional fill and the slopes of the Bludovice and Detmarovice paleovalleys in the Upper Silesian basin have not been fully explored for gas.

The Carpathian Flysch Belt

The hydrocarbon potential of the Carpathian Flysch belt in the territory of Moravia is very limited (Table 2). However, the imposition of the wedge-shaped flysch belt over the European platform was critical for the

Table 2. Oil and gas fields in the Carpathian Neogene foredeep and the Flysch belt in Moravia. Location of fields in Figure 25.

Field Name	Year of Discovery	Age of Reservoirs	Type of Reservoir	Porosity (%)	Field Type	Production Status	Initial Reserves Oil (million bbl)	Gas (bcf)
Carpathian foredeep								
Bruzovice–Frydek	1952	Miocene	clastic	15–20	gas	coal degasation		5
		Carboniferous	clastic	7–15				
Choryne	1908	Karpatian	clastic	15–20	gas	conservation		21
		Carboniferous	clastic	7–15				
Dolni Dunajovice	1973	Eggenburgian	clastic	23–37	gas	gas storage		35.5
Horni Zukov	1915	lower Badenian	clastic	15–20	gas	gas storage		39.1
		Carboniferous	clastic	7–15				
Komorni Lhotka	2001	Miocene	clastic	15–20	gas	development ongoing		3.7
		Carboniferous	clastic	7–15				
Kozlovice–Lhotka	1975	Silesian unit	clastic	10–15	gas	development ongoing		0.9
		Miocene	clastic	15–20				
		Carboniferous	clastic	7–15				
Krmelin	1958	Miocene	clastic	15–20	gas	coal degasation		1.5
		Carboniferous	clastic	7–15				
Menin–Zatcany	1944	Badenian	clastic	30	oil	producing	1	
	1930	Karpatian	clastic	25–30				
Mitrovice–Paskov	1909	Miocene	clastic	15–20	gas	coal degasation		10
		Carboniferous	clastic	7–15				
Novy Prerov	1986	lower Miocene	clastic	18–25	gas	producing		0.76
Pribor Jih (Stramberk)	1965	Karpatian	clastic	15–20	gas	gas storage		21
		Carboniferous	clastic	7–15				
Pribor–Klokocov	1908	Karpatian	clastic	15–20	gas	conservation		35
	1912	Carboniferous	clastic	7–15				
Stonava	1952	Miocene	clastic	15–20	gas	coal degasation		1.5
		Carboniferous	clastic	7–15				
Staric–Liskovec–Sviadnov	1913	Miocene	clastic	15–20	gas	coal degasation		7
		Carboniferous	clastic	7–15				
Carpathian Flysch belt								
Hluk	1943	Bile Karpaty Flysch	clastic	14–24	gas	abandoned		0.4
Lhotka–Pstruzi	1975	Silesian unit	clastic	10–15	gas	producing		6
Roznov	1983	Subsilesian unit	clastic	10–15	gas	development ongoing		0.3
Koprivnice–Ticha	1982	Silesian unit	clastic	10–15	gas	producing		17.6
		Miocene	clastic	15–20				
		Carboniferous	clastic	7–15				

generation of hydrocarbons from source rocks located in the underlying platform. The predominantly impermeable rocks of the flysch belt also provide an important regional seal.

The marginal Pouzdrany unit, as presently known, does not have any economic hydrocarbon potential. The Oligocene organic-rich rocks of this unit are immature; however, the tectonic slivers of this unit buried deeper below the Outer Carpathian thrust belt, together with the remnants of the autochthonous Paleogene, may represent hydrocarbon source rocks.

Likewise, the Waschberg–Zdanice–Subsilesian unit has practically no hydrocarbon potential. Numerous wells, drilled both in the southern Zdanice sector and the northern Subsilesian sector, encountered practically no noticeable signs that oil and/or gas might be present in commercial quantities in these units. Some rocks of the Menilitic Formation are organic rich but remain mostly immature. Only limited parts of the Zdanice–Subsilesian unit buried deeper below the Silesian and Magura nappes may have generated some hydrocarbons. However, the impermeable strata of the lower members of this

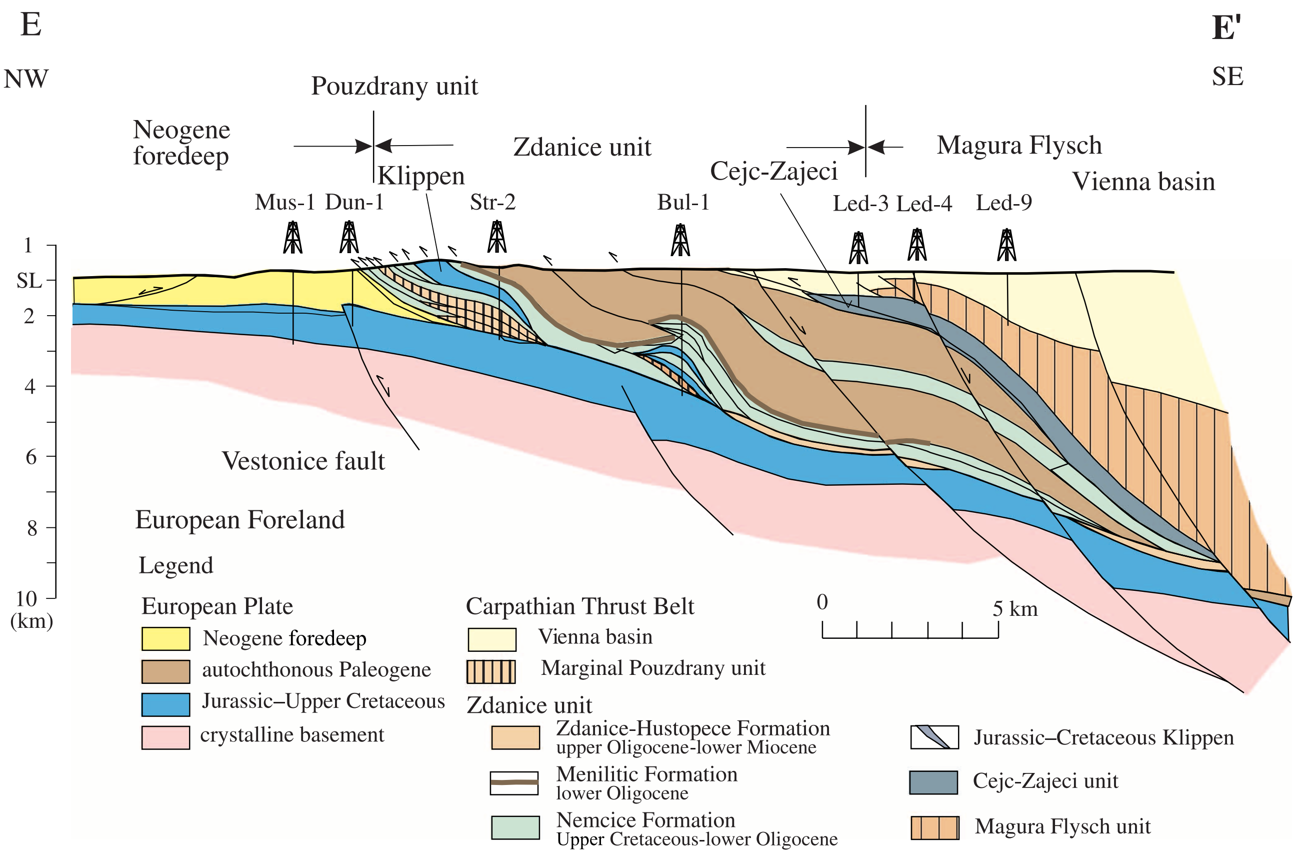

Figure 26. Cross section EE′ (location in Figures 3, 7) through the Neogene foredeep, the frontal units of the Western Carpathian thrust belt, the Vienna basin, and the underlying European plate in southern Moravia. Note the structural position of the Jurassic–Cretaceous klippen in the Waschberg–Zdanice unit and the relationship of the allochthonous Pouzdrany unit to the autochthonous Paleogene. The Vestonice fault is interpreted as a compressional or transpressional structure. It may have originated as a normal fault during the Jurassic rifting and was activated again as a reverse fault during the Alpine convergency in the early Miocene (Eggenburgian). The Dolni–Dunajovice gas field is reservoired in the lower Miocene sands draping over the antiformal structure associated with the hanging wall of the Vestonice fault. Modified from Picha and Stranik (1999). Stratigraphic records of some wells shown on the cross section are reported in Appendix 1.

unit represent an excellent regional seal, critical for the lateral migration of hydrocarbons from the deep generative kitchens into the reservoir in the foreland and their preservation in the otherwise open reservoirs, such as the fragmented and weathered crystalline basement rocks.

Some Paleocene to Eocene sandstones and conglomerates of the Nemcice (Submenilitic) Formation in the Cejc–Zajeci unit possess good porosities and may be considered as reservoir rocks. The thickness of the individual sand bodies ranges from several meters up to several tens of meters; however, their lateral extent seems to be limited. These coarse clastics enclosed in the impermeable shaly deposits may actually represent the deep-water channels similar to those explored in the North Sea (e.g., Alba and Frigg fields) and elsewhere along the present continental margins. Recognition of these bodies at depth and outlining of closures in the structurally complex (duplex-type) Cejc–Zajeci thrust unit, however, would require 3-D seismic survey. A shallow well drilled into the Eocene sands west of the village of Kobyli encountered several gas shows (88% methane and 9% nitrogen) (Picha et al., 1968). The deep Kobyli-1 well drilled nearby, however, did not register any hydrocarbon shows in the entire Cejc–Zajeci unit (Picha et al., 1971). The Paleocene to Eocene sandstones in the Cejc–Zajeci unit bear many similarities with the Ciezkowice Sandstones of the Silesian unit in Poland.

The Silesian unit has been the most significant oil producer in the Polish Western Carpathians. It contains both significant source rocks in the Lower Cretaceous strata and the lower Oligocene Menilitic Formation and numerous reservoir sandstones. Among them, the Paleogene Ciezkowice Sandstones are the most prolific (Karnkowski and Ozimkowski, 1998; Dziadzio et al., 2006). In numerous smaller fields, they produced almost half of the total oil recovered in the entire Polish Carpathians (Wdowiarz, 1985; Slaczka, 1996). Additional reservoirs are present in the Lower Cretaceous Tesin–Hradiste, Verovice, and Lhoty formations, the Upper Cretaceous Istebna Formation (locally called Czarnorzeki beds), and in the Oligocene Kliwa Sandstones and the Krosno Formation (Karnkowski, 1993).

In the territory of Moravia, so far, only two small accumulations of gas have been found in the Silesian unit (Table 2). The Roznov-1 well encountered 45 gas-bearing horizons, 0.8–11.4 m (2.6–37.4 ft) thick, in the Silesian and Subsilesian units. In the Lhotka–Pstruzi field, in addition to the subthrust Karpatian and Carboniferous reservoirs, accumulations of natural gas have also been found in the overlying Silesian unit. Despite these poor results, the Silesian unit possibly remains one of the more promising subjects of exploration in the Flysch belt of Moravia.

As documented by numerous wells, the hydrocarbon potential of the Magura flysch in Moravia seems to be very limited. It has only minor source rocks in the Zlin Formation of the Raca unit and the Hluk facies of the Bile Karpaty unit, and poor reservoirs elsewhere. The porosity of sandstones ranges from 7 to 12% in the unweathered and more than 12% in the weathered rocks (Ondra and Hanak, 1992). Besides seepages and oil and gas shows in wells, only two small fields have been found so far in the Magura flysch (Table 2; Figure 25). A small gas field, Hluk, located in the Bile Karpaty unit at a depth of 200–600 m (660–2000 ft), has been exploited since 1943. In the abandoned Vacenovice field (Table 4, shown on page 143), oil and gas were produced from the weathered surface of the Paleogene flysch deposits underneath the depositional fill of the Vienna basin. On the positive side, the Magura nappe represents a significant regional seal.

In our opinion, the duplexes of Paleogene strata of the external units buried below the major Silesian and Magura nappes remain the most promising targets of exploration in the entire Western Carpathian Flysch belt.

The Subthrust Plays

The hydrocarbon potential of the European plate was greatly enhanced by its burial below the accretionary wedge of the Carpathian thrust belt. The overthrusting by the thrust belt not only enhanced the maturation of source rocks and generation of hydrocarbons but also improved the trapping mechanism by providing a good regional seal (Picha, 1996). The increasing thickness of the thrust belt toward southeast, however, limits the extent of the subthrust plate accessible to exploration. Considering the economic and technical feasibility of drilling to a depth of 6–7 km (3.7–4.3 mi), the width of the drillable subthrust zone in Moravia typically would not exceed 50 km (31 mi) (Figure 27). On the positive side, the relatively low heat flow in the Carpathian foreland (Cermak, 1975, 1979) allows for liquid hydrocarbons to be generated and preserved at depths of more than 5 km (3 mi) and dry gas at depths of 9 km (5.6 mi) (Stranik et al., 1993). Thus, both oil and gas may be present in the deep subthrust plays (Picha, 1996).

Hydrocarbons have been found in the Paleogene, Jurassic, and Paleozoic strata and the crystalline basement of the subthrust European plate (Table 3; Figure 25).

The Nesvacilka and Vranovice Paleovalleys and Their Paleogene Fill

The large Nesvacilka and Vranovice paleovalleys represent a complex hydrocarbon province with multiple potential hydrocarbon plays. So far, only the more thoroughly explored Nesvacilka paleovalley yielded commercial discoveries. The Nesvacilka paleovalley is located in a broader structure of the northwest–southeast-trending Nesvacilka graben, whose origin is most likely related to the Jurassic rifting (Figure 13). The Paleogene fill not only holds source rock and reservoirs but also functions as a regional seal for hydrocarbon accumulations located in the fractured crystalline rocks and the Paleozoic and Jurassic carbonate and clastic reservoirs situated along the northeastern more gently dipping side of the Nesvacilka graben and the northern side of the Nesvacilka paleovalley (Picha, 1996) (Figure 12) (see also the account by Kostelnicek et al., 2006). Last but not least, the paleovalleys apparently also served as conduits for migration of hydrocarbons from the generative kitchen located deeper below the thrust belt into shallower reservoirs in the foreland.

The autochthonous Paleogene fill of the large Nesvacilka and Vranovice paleovalleys possesses all the main components of a petroleum system: organic-rich source rocks, deep-water channelized reservoirs enclosed in impermeable rocks, and proper depth of burial below the thickening wedge of the thrust belt (Figure 28). The best reservoirs are the meandering channels of subsea fans similar to those explored for hydrocarbons in deep-water turbiditic deposits elsewhere along the passive continental margins of South America, Africa and Gulf of Mexico; only the

Table 3. Oil and gas fields in the European platform underneath the Outer Carpathian belt in Moravia. Location of fields in Figure 25.

Field Name	Year of Discovery	Age of Reservoir	Type of Reservoir	Porosity (%)	Field Type	Production Status	Initial Reserves Oil (million bbl)	Gas (bcf)
Bosovice	1990	Paleogene	clastic	15–25	oil	development ongoing		
Damborice	1986	Paleogene	clastic	15–23	oil and gas	producing	12	5.5
		Jurassic	carbonate	10				
		Jurassic	clastic	15–25				
Dolni Lomna	1985	Silesian unit	clastic	10–15	oil	development ongoing	0.15	
		crystalline	paragneiss	3–5				
Karlin	1991	Paleogene	clastic	15–20	gas	producing	0.15	3
Korycany	1978	lower Miocene	clastic	15–18	oil and gas	producing	0.2	3
		crystalline	granodiorite	3–5	oil and gas	producing		
Kostelany Vychod (East)	1968	lower Miocene	clastic	15	oil and gas	producing	1	30
		crystalline	granite	3–5				
Kostelany Zapad (West)	1971	lower Miocene	clastic	15	gas	producing		24.9
		crystalline	granite	3–5				
Krumvir	1997	Paleogene	clastic	15–25	gas and oil	producing	0.3	2.7
Krasna–Moravka	1980	lower Miocene	clastic	15–20	oil and gas	development ongoing	0.3	0.73
		crystalline	carbonate	4				
Nemcicky	1978	Devonian	carbonate	3–5	oil and gas	producing(?)	0.01	0.03
Nikolcice	1969	Paleogene	clastic	15–23	oil and gas	producing		0.01
Nitkovice–Hradisko	1971	Devonian	carbonate	3–5	oil and gas	producing		
Otnice	1994	Paleogene	clastic	15–25	gas	development ongoing		1.5
Rusava	1976	lower Miocene	clastic	12	gas	abandoned		0.1
		crystalline	gabbro	2–5				
Uhrice	1982	Paleogene	clastic	15–23	oil and gas	gas storage	1.5	7.9
		Jurassic	carbonate	10				
		Jurassic	clastic	15–25				
	1978	Devonian	carbonate	3–5	oil and gas			
Uhrice Jih (South)	2001	Jurassic	clastic	21	oil and gas	producing	2.5	2.3
Zarosice	2001	Jurassic	carbonate	10	oil and gas	producing	3.8	3.5
Zdanice Jih–Zdanice	1984	lower Miocene	clastic	25	oil and gas	producing	3.5	4.5
Sever–Kloboucky	1973	crystalline	diorite–tonalite	3–6				

dimensions of those features in Moravia seem to be smaller.

The hydrocarbon potential of the autochthonous Paleogene fill in the Nesvacilka paleovalley has been only partly explored (Brzobohaty, 1993; Brzobohaty et al., 1996). Numerous gas shows and some oil shows have been encountered in several wells, e.g., Bosovice-1, Hosteradky-1, Milesovice-1, Tesany-1, Nasedlovice-1, and Susice-1, but so far, only two smaller gas-condensate-oil fields, Krumvir and Karlin, have been found and put into production. The Krumvir-2 well, drilled to the total depth of 3600 m (11,811 ft), encountered a 30-m (100-ft) section of good-quality gas- and oil-bearing sand with 15–25% porosity. An initial drillstem test flowed 750 bbl of oil/day of 30° API, low-sulfur crude oil against a restricted 0.22-in. (5.5-mm) choke. High-pressure, gas-bearing sand horizons have also been encountered in the Karlin-1 well located further east of the Krumvir field (Benada and Blazej, 1991). The deeper zones of the Nesvacilka valley fill with a higher proportion of sandstones remain unexplored. Some measurements, however, indicate that the quality of these reservoirs may deteriorate with the increasing depth. The Vranovice paleovalley has been explored only marginally,

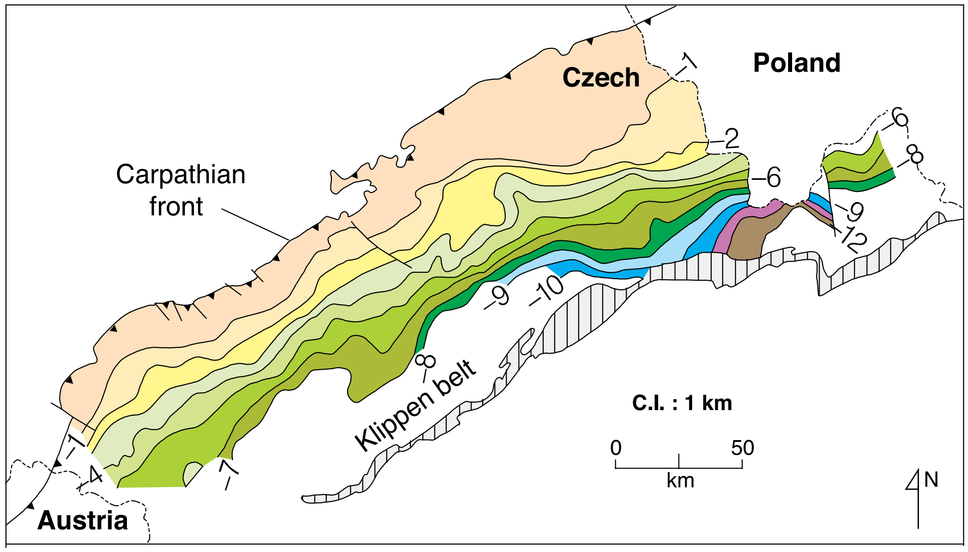

Figure 27. Depth (in km) to the base of the Outer Carpathian thrust belt in Moravia. Modified after Picha (1996).

without any apparent success. Application of new technologies, such as the 3-D seismics, may provide a tool for better assessment of the geometry and quality of reservoirs and a lowering of the exploration risk. We believe that small- and middle-sized oil and gas-condensate fields can be found in the predominantly stratigraphic traps, especially in the deeper parts of the paleovalley fill. Moreover, existence of channelized subsea fans may be anticipated at the mouths of these large submarine canyons. Presently, these Paleogene fan deposits are either deeply buried below the Carpathian thin-skinned nappes or detached and integrated into the allochthonous belt itself. If accessible by drilling, these potential fans may yet become targets of a future exploration.

As such, the autochthonous Paleogene remains an interesting object of the subthrust exploration in southern Moravia. In many aspects, including the reservoirs and generation and migration of hydrocarbons, the Paleogene play of southern Moravia resembles the hydrocarbon habitat of the Molasse Basin in Austria and Bavaria. The structural setting of these two hydrocarbon provinces, however, is different; it is predominantly a subthrust play in Moravia, whereas in

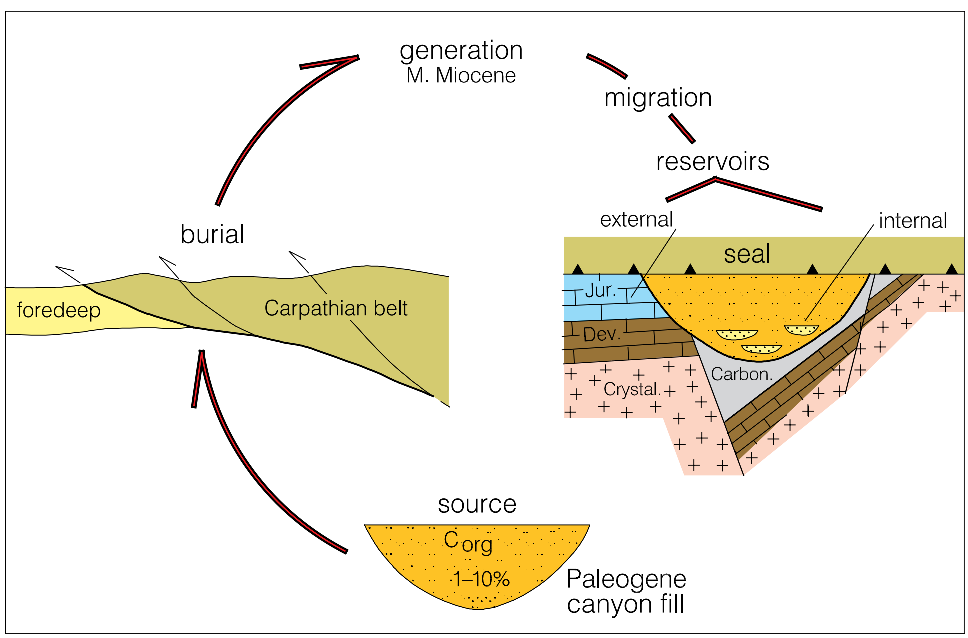

Figure 28. Hydrocarbon habitat of the Nesvacilka graben and the Paleogene canyon system (Picha, 1996).

Austria and Bavaria, the productive Paleogene to Neogene reservoirs found in the Molasse Basin are located in front of the Alpine thrust belt.

The Jurassic Play

By far, the largest subthrust oil and gas accumulations in Moravia have been found in the erosionally isolated blocks and pinnacles of Jurassic strata in the northeastern slopes of the Nesvacilka graben and the Nesvacilka paleovalley, buried below the autochthonous Paleogene deposits and the Zdanice nappe (Figures 7, 12, 13; Table 3). The hydrocarbons are reservoired in the clastic rocks of the basal Gresten Formation (Damborice oil field, Uhrice gas field, and Uhrice South oil and gas field) or in the Vranovice Limestones and dolomites (Zarosice oil field). The porosities of the Gresten sandstones may reach as much as 25%. Additional hydrocarbons have been found in adjacent Paleozoic carbonates and clastics. The Damborice field, holding about 30 million bbl of oil in place, represents the largest subthrust discovery in the Czech Republic; more than 70% of the total present production in the country comes from this field. The Uhrice gas field with initial reserves of 224 million m^3 (7.9 bcf) of natural gas has been exploited and converted into a gas storage. All these fields were sourced from the Jurassic organic-rich Mikulov Marls. Because no generative kitchen is present in the proximity of these fields, the hydrocarbons must have migrated laterally updip from the zones of the subthrust plate buried deeper below the Carpathian thrust belt and the Vienna basin. Similar prospects in erosional blocks and pinnacles may also exist in the Vranovice paleovalley, which has not been explored for this type of play. Shows of hydrocarbons in Jurassic strata have been also encountered in the deep wells Dolni Dunajovice-1, Hrusovany-1, Nikolcice-1, Kobyli-1, and Nove Mlyny-2.

Within the Dyje–Thaya depression, large subthrust antiformal structures, such as Tynec, Holic, and Lednice, have been identified on seismic data. Estimated depths to the tops of these structures range from 4000 m (13,000 ft) for the Lednice structure to 6000–7000 m (19,700–23,000 ft) for the Tynec and Holic structures. These structures apparently formed as tilted blocks and horsts during the Jurassic rifting, and at least some of them, e.g., Tynec structure, were further activated during the late orogenic strike-slip faulting. In southern Moravia, these structures are situated in the generative zone of the Jurassic and Paleogene source rocks. The Tynec deep structure (Figure 20, section DD′) underlies the shallow oil and gas fields in the Vienna basin. Biomarker studies (Picha and Peters, 1998) indicate that the oil from the Tynec-34 well has a mixed Jurassic and Paleogene signature. Based on the interpretation of limited seismic data, the size of the Tynec structure has been estimated to be about 70 km^2 (27 mi^2). The seismically better defined Holic structure, located further northeast mostly on the territory of Slovakia, may cover as much as 90 km^2 (35 mi^2). The deep Lednice (16 km^2; 6.2 mi^2) structure interpreted from seismic, geomagnetic, and gravity data (Ciprys and Thon, 1990) is located below the western margin of the Vienna basin in southern Moravia. Potential reservoirs in these structures may be found in the Paleogene clastics (channels and fans), Jurassic clastics and carbonates, Devonian dolomites, and the Precambrian crystalline rocks. All these rocks are proven reservoirs in the shallower zones of the subthrust platform. The thrust belt with predominantly shaly sequences on its bottom provides an excellent seal. With respect to a great depth of these structures, most likely, gas and condensate rather than oil may be found in these prospects.

In the territory of Austria, the gas fields Klement and Hoflein and the oil field Grunau have been found in Jurassic and Cretaceous clastic and carbonate reservoirs in the shallower zone of the foreland plate below the Waschberg and Flysch belt nappes (Grun, 1984; Zimmer and Wessely, 1996). Very deep, large, presumably rift-related structures, such as Maustreng–Zistersdorf and Aderklaa, have been partly explored by deep drilling without any commercial success (Wessely, 1990; Milan and Sauer, 1996).

The presence of potential reservoirs in all these deep structures is considered to be the most critical risk factor. However, because of their large size and location in the generative kitchen for hydrocarbons, these structures remain tempting exploration targets.

Fractured Malmian Marls, as much as 1500 m (4900 ft) thick, if overpressured, could also be considered as unconventional producers of natural gas (tight-gas play). Gas generation apparently still continues in these organic-rich source rocks.

The Jurassic play, especially in combination with the Nesvacilka and Vranovice paleovalleys and their impermeable Paleogene fill, is undoubtedly the most promising subthrust exploration play in Moravia. The chances of finding additional fields in this play remain quite high.

The Paleozoic Play

Only small accumulations of hydrocarbons have been found in the Paleozoic sequences: the Middle Devonian clastics, the Upper Devonian to the Lower Carboniferous karstified and fractured limestones and dolomites, and the Lower Carboniferous Culm deposits (Thonova and Benada, 1990). In northern Moravia, a small Krasna field (Figure 25), reservoired in the Devonian carbonates and crystalline basement rocks, has been developed and put on production. Other noncommercial Paleozoic subthrust accumulations of hydrocarbons have been encountered in the wells at Nitkovice (small gas field), Slavkov, Nemcicky, and Drazuvky, all located in central Moravia. Some of these occurrences are just minor extensions of larger fields, hydrodynamically connected with other more productive main reservoirs, e.g., the crystalline basement rocks (Zdanice field) or Jurassic clastics and carbonates (Damborice field).

In the Upper Silesian basin, coal gases accumulated in elevations of the Carboniferous paleorelief. They are reservoired in the weathered surface of the Upper Carboniferous rocks, as well as in the overlying Miocene deposits of the Carpathian foredeep. Additional coal gases are found dissolved in waters of the lower Badenian detritic fill of the Bludovice and Detmarovice erosional depressions.

The Paleozoic strata of the Carpathian foreland were deformed during the Hercynian orogeny. Foreland-type compressional structures of the frontal zone of the southwest–northeast-trending Hercynian belt in Moravia extend far below the Carpathian thrust belt (Figure 7). These Hercynian structures with potential reservoirs in the Devonian and Carboniferous clastics and carbonates did exist prior to the development of the Mesozoic Tethyan rifted margins and the Carpathian thrust belt. Unless disrupted by erosion and further tectonism, these structures, buried and possibly sealed by the impermeable rocks of the Neogene foredeep and the Carpathian thrust belt, may have become receptacles for hydrocarbons generated during the Carpathian tectogenesis (Picha, 1996). The reservoir properties of the Paleozoic rocks may actually improve eastward toward the stable foreland of the Hercynian belt, which, during the Devonian to Lower Carboniferous, was the site of a shallow-marine carbonate sedimentation favorable for generation and preservation of porosities. At least some of the Hercynian compressional structures might have been reactivated during the Late Cretaceous–Paleogene (Laramide) and/or the Miocene late orogenic mobilization of the European foreland plate.

Despite the lack of commercial discoveries to date, it is still possible that accumulation of hydrocarbons may still be found in Paleozoic structures deeply buried below the Carpathian thrust belt. Some interpretations of seismic data indicate a potential existence of Devonian and Carboniferous reefal limestones and dolomites in the subthrust plate at depths greater than 5000 m (16,400 ft). However, any further exploration of the Paleozoic potential will require application of more advanced seismic techniques combined with a thorough structural and stratigraphic analysis of the potential prospects.

Elevations of the Crystalline Basement

Several oil and gas fields, such as Zdanice, Kloboucky, Lubna, Kostelany, Krasna, and Dolni Lomna, have been found in elevations of the crystalline basement (buried hills) below the Neogene foredeep and the frontal zones of the thin-skinned Carpathian thrust belt in Moravia (Figure 8; Table 3). These accumulations are reservoired in the weathered and fractured surface zones of granitic rocks and in the overlying lower Miocene (Karpatian) sands of the Neogene foredeep sealed by the impermeable deposits of the Zdanice–Subsilesian nappe (Krejci, 1993; Blizkovsky et al., 1994). The Precambrian crystalline rocks have low porosities, on average 2–4%; however, the thickness of the oil-saturated zone may reach several tens up to several hundred meters. The overlying lower Miocene sands are only a few tens of meters thick; their porosities range from 15 to 25%. These relatively shallow, less than 1500-m (4900 ft)-deep oil accumulations are charged by a long-range migration from the Jurassic and Paleogene organic-rich source rocks in the deeper zones of the European platform and the Carpathian thrust belt.

The largest of these crystalline basement-reservoired accumulations, the Zdanice field located in central Moravia (Figure 25), holds more than 35 million bbl of 19–33° API oil in place, which, however, is difficult to extract (Krejci, 1993; Buchta and Dohnal, 1996). In 1998, the entire production from the Zdanice elevation was only about 300 bbl/day. Chances are that the rate of production may improve by application of more advanced technologies, such as horizontal drilling or steam flooding.

The Vienna Basin

The Vienna basin, superimposed on the Carpathian thrust belt in southern Moravia and northeastern Austria (Figures 1, 3), has been one of the most prolific producers of hydrocarbons in the entire Carpathian region. More than 800 million bbl of oil has been produced from the numerous fields of the Vienna basin, mainly in the territory of Austria. The majority of this production has come from the giant Matzen field (Fuchs and Hamilton, 2006).

The oils of the Vienna basin were sourced predominantly from the Jurassic organic-rich marls and shales and, to a lesser degree, from the Paleogene strata, both found in the underlying European platform (Ladwein, 1988; Ladwein et al., 1991; Picha and Peters, 1998). The vertical migration through the Flysch belt into the reservoirs in the Vienna basin was apparently enabled by the continuing tectonic activity along the major wrench faults (Figure 24, section AA′). The potential source rocks in the Vienna basin, although locally buried to a depth of 4 km (2.4 mi), would remain mostly immature. Numerous structural and stratigraphic traps were created during the complex structural and depositional history of the Vienna basin. The middle Miocene (Badenian) sands represent the best reservoirs, but oil and gas accumulations have been also found in the lower Miocene and Sarmatian deposits.

Numerous smaller fields have been found in the Moravian part of the Vienna basin (Table 4; Figure 25). The Hrusky–Josefov field, with initial reserves of 15.5 million bbl of oil and 90 bcf of gas, is the largest. It has been partly converted into a gas storage. The geology and hydrocarbon potential of the Vienna basin is more thoroughly discussed by Seifert (1996), Arzmuller et al. (2006), and Fuchs and Hamilton (2006).

The Key Role of the Dyje–Thaya Depression in the Formation of Petroleum Systems

The existence of the Dyje–Thaya depression and its evolution during the Mesozoic and Cenozoic times apparently was a key factor in the development of several petroleum systems in southern Moravia and

Table 4. Oil and gas fields in the Moravian part of the Vienna basin. Location of fields in Figure 25.

Field Name	Year of Discovery	Age of Reservoir	Type of Reservoir	Porosity (%)	Field Type	Production Status	Initial Reserves Oil (million bbl)	Gas (bcf)
Breclav	1946	Badenian	clastic	15	oil and gas	producing	0.25	1.9
Brodske	1951	Badenian	clastic	20	oil	producing	2.5	1.9
Dolni	1989	Sarmatian	carbonates	15–25		producing		
Bojanovice		Badenian	clastics	20–25	oil and gas	gas storage		
Hodonin	1920	Sarmatian	clastic	15–20	oil and gas	producing	2.5	0.7
		Badenian	clastic	20				
		lower Miocene	clastic	18–20				
Hrusky–Josefov	1959	Sarmatian	clastic	15–25	oil and gas	producing	15.5	90
		Badenian	clastic	20				
		Karpatian	clastic	18–20				
Kostice	1953	Sarmatian	clastic	15–25	oil and gas		2.5	2
		Badenian	clastic	20–25				
Lanzhot	1957	Sarmatian	clastic	20	oil and gas	conservation	1	4.5
		Badenian	clastic	15				
Lednice–Valtice	1979	Badenian	clastic	20	gas	producing		3
		Badenian	carbonates	5–10				
Luzice	1921	Sarmatian	clastic	15–20	oil and gas	producing	2.75	10
	1944	Badenian	clastic	20				
		lower Miocene	clastic	18–20				
Poddvorov–Mutenice	1951	Sarmatian	clastic	20–25	oil and gas	producing	3.5	30
		Badenian	clastic	20–25				
Postorna	2001	Badenian	clastic	15	oil	exploration	0.45	
Prusanky	2001	Sarmatian	clastic	20–25	gas	producing		3.5
Tynec–Cunin	1959	Raca unit	clastic	15–20	oil and gas	producing	1.4	3.6
	1945	lower Miocene	clastic	15–18				
Vacenovice	1930	Sarmatian	clastic	20	oil and gas	abandoned	0.4	0.6
		Raca unit	clastic	15–20				
Velke Bilovice–Zizkov–Podivin	1944	Sarmatian	clastic	20	oil and gas	producing	0.4	30.8
		Badenian	clastic	20–25				

northeastern Austria. The existing evidence indicates that this depression was formed or at least activated during the Jurassic rifting and repeatedly uplifted in the Lower Cretaceous and in the Late Cretaceous–early Paleogene. Prior to the Jurassic rifting, the area of the depression was situated in the frontal zone of the late Paleozoic Hercynian orogenic belt. The potential compressional Hercynian structures, still possibly evident on some regional sections (Figure 20, section DD′), might have been reactivated during the subsequent tectonic events. During the Laramide uplifting of the European foreland at the Cretaceous–Paleogene transition, two large paleovalleys were cut inside the depression and later filled with the deep-water Paleogene deposits. In the early Miocene, the depression was mostly overthrust by the thin-skinned Carpathian belt. The Jurassic and Late Cretaceous strata were partly detached from the autochthonous cover of the European platform and integrated into the Waschberg sector of the thrust belt as the Outer klippen of Ernsbrunn and Pavlov Hills. Finally, in the middle Miocene, the pull-apart phase of the Vienna basin formed along the orogen-parallel strike-slip faults within the confines of the depression. Any interpretation of structural history and resulting structural architecture of the area, especially at deeper levels below the Carpathian thrust belt, thus remains a daunting task requiring a four-dimensional analysis (including timing).

Within the structural limits of the depression, thick, organic-rich marls accumulated in the Late Jurassic, and the Paleogene organic-rich deposits filled two paleovalleys and submarine canyons. After being buried below the thrust belt and the Vienna basin, these source rocks matured, and the Dyje–Thaya depression became a world-class generative center, from which the hydrocarbons migrated both vertically and horizontally into various reservoirs in the Vienna basin and the subthrust foreland, respectively. More than 850 million bbl of oil and 4 trillion cubic feet of gas have been produced from various stratigraphic and structural levels in the realm of this depression. With respect to its limited size, the Jurassic Dyje–Thaya depression may be regarded as one of the most productive provinces in the world. Although heavily drilled to the depth of about 3000 m (10,000 ft), the recent discoveries, especially in the subthrust settings, indicate that the Dyje–Thaya depression still remains one of the most promising places for exploration in the entire Carpathian region.

CONCLUDING REMARKS

We have attempted to present a comprehensive and up-to-date account on the geology and the hydrocarbon potential of the Carpathian thrust belt and its foreland in the territory of the Czech Republic. Some readers may find it too detailed, others too short and superficial. In our account, we emphasized those aspects of geology that are specific for the region, critical for integration of local geology into a wider framework of the Alpine–Carpathian region, and important for understanding the hydrocarbon potential. However, we have also included more detailed description of stratigraphy, including logs of numerous deep wells, which may be useful to future students of Carpathian geology. Several new concepts and original ideas, e.g., about the functioning of the Western Carpathian transfer zone, the existence of the Dyje–Thaya depression, and the foreland depositional setting, have been broadly formulated in this chapter for the first time.

In our deliberation, we tried to be objective and present a balanced picture of interpretations and ideas. Given the enormous amount of published and unpublished data, we did not have a choice but to be selective. Unintentionally, we may have omitted some important papers or interpretation, and we apologize for it. Despite of these challenges, we hope that the article will provide useful information to geologists interested in Carpathian geology and enhance further discussions and additional work on some of the controversial issues. Last but not least, we anticipate that the comprehensive account on the hydrocarbon systems may encourage further exploration for oil and natural gas in this prolific and still promising region.

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APPENDIX 1: SELECTED DEEP WELLS IN THE MORAVIAN CARPATHIANS (CZECH REPUBLIC)

Branky - 1 (Bra-1)

0-378 m	Silesian and Subsilesian units	
-------------	basal overthrust of the Flysch nappes	
-748 m	lower Miocene (Karpatian) of the Carpathian foredeep	
-1625 m	Lower Carboniferous (Visean) - Culm facies	
-2372 m	Devonian (Frasnian) to Lower Carboniferous (Visean) - Carbonate facies	
-2427 m	Devonian (Givetian) - basal clastics	
-2540 m	crystalline basement (metamorphites)	

Bulhary - 1 (Bul-1)

0-32 m	Badenian Vienna basin (transgression)	
-1192 m	Zdanice-Hustopece formations (Waschberg sector)	
-2295 m	Menilitic and Nemcice formations (imbricated) (Waschberg sector)	
-2391 m	Mikulov Marls, Jurassic	
-------------	overthrust	
-2688 m	Nemcice Formation (Waschberg sector)	
-2706 m	Mikulov Marls, Jurassic with incorporated Boudky Marls of the Pouzdrany unit	
-------------	overthrust	
-3119 m	Nemcice Formation (Waschberg sector)	
-------------	overthrust	
-3126 m	Pouzdrany Marls of the Pouzdrany unit	
-------------	overthrust	
-3231 m	Nemcice Formation (Waschberg sector)	
-------------	basal overthrust of the Flysch nappes	
-3500 m	Mikulov Marls (autochthonous Jurassic)	

Bystrice Pod Hostynem - 1 (Bys-1)

0-400 m	Fore-Magura unit	
	-47 m	Krosno Formation
	-400 m	Submenilitic Formation
-------------	overthrust	
-691 m	Silesian unit	
	-639 m	Krosno Formation
	-691 m	Roznov Formation
---------------	overthrust	
-1002 m	Silesian(?) or Waschberg - Zdanice - Subsilesian units(?) (Zdanice sector)	
	-915 m	Zdanice-Hustopece Formation
	-1002 m	Nemcice Formation
----------	basal overthrust of the Flysch nappes	
-1242 m	lower Miocene (Karpatian) of the Carpathian foredeep	
-1897 m	Lower Carboniferous (Culm facies)	
-2600 m	Devonian (Frasnian) - Lower Carboniferous (carbonate facies)	

Choryne - 9 (Cho-9)

-914 m	Lower Carboniferous	
-1595.9 m	Devonian	
	-961.7 m	Lisen Formation (Krtiny Limestones)
	-1471 m	Macocha Formation (Lazanky and Vilemovice limestones)
	-1595.9 m	Basal clastics (Devonian)
-1711.6 m	crystalline basement (metamorphite)	

D. Lomna - 3 (D.Lom-3)

0-1639 m	Silesian unit (Godula subunit)	
-1919 m	Subsilesian unit (Submenilitic Formation)	
-1962 m	lower Miocene (Karpatian)	
-2151 m	crystalline basement (metamorphite)	

Gottwaldov - 2 (Got-1)

0-3893 m	Raca unit of the Magura Group of nappes	
	-1340 m	Vsetin Member of Zlin Formation
	-1735 m	Ujezd Member of Zlin Formation
	-1980 m	Beloveza Formation
	-3835 m	Solan Formation
	-3893 m	Kaumberg Formation
-------------	overthrust of the Magura Group of nappes	
-4215 m	Fore-Magura unit	
-------------	basal overthrust of the Flysch nappes	
-4725 m	lower Miocene (Karpatian) of the Carpathian foredeep	
-4850 m	crystalline basement (metamorphite)	

Hnojnik - 1 (Hnoj-1)

0-284 m	Silesian unit	

-890 m	Subsilesian unit	
	basal overthrust of the Flysch nappes	
-942 m	lower Miocene (Karpatian) of the Carpathian foredeep	
-1100.5 m	Upper Carboniferous (Namurian A), Ostrava Formation	
-1100.5 m	Lower Carboniferous (Toumaisian) to Upper Carboniferous (Namurian), Culm facies	
-1214 m	Devonian (Eifelian) to Lower Carboniferous (Visean), Carbonate facies	

Holesov - 1 (Hol-1)

0-272 m	Silesian unit (Roznov Formation)	
-------------	basal overthrust of the Flysch nappes	
-728 m	lower Miocene (Karpatian) of the Carpathian foredeep	
-907 m	Macocha Formation (Devonian)	
-924 m	crystalline basement (Diorite)	
	overthrust	
-940 m	Macocha Formation (Devonian)	
-995 m	basal clastics (Devonian)	
-1050 m	crystalline basement (Diorite)	

Hrusovany - 1 (Hr-1)

-561 m	Carpathian foredeep	
	-169 m	Badenian
	- 490 m	Karpatian
	- 534 m	Eggenburgian to Ottnangian
- 742.5 m	- 561 m	basal clastics, Eggenburgian
- 816 m	autochthonous Mesozoic, Altenmarkt, and Nikolcice formations	
	crystalline basement (Granodiorite)	

Jablunka - 1 (Jab-1)

0-2505 m	Raca unit of the Magura Group of nappes	
	-100 m	Beloveza Formation
	-1276 m	Solan Formation
	-2505 m	Kaumberg Formation
-----------	overthrust of the Magura Group of nappes	
-2895 m	Fore-Magura unit	
-----------	basal overthrust of the Flysch nappes	
-3950 m	Upper Carboniferous (Namurian A) - Ostrava Formation	
-5405 m	Lower to Upper Carboniferous (Visean - Namurian A) - Culm facies	
	-5283 m	Hradec-Kyjovice Formation
	-5405 m	Moravice Formation
-6278 m	Devonian to Lower Carboniferous (Givetian - Visean) - carbonate facies	
	-5836 m	Lisen Formation (Krtiny and Hady - Ricky Limestones)
	-6278 m	Macocha Formation (Lazanky Limestones), Givetian
-6318 m	basal clastics	
-6506 m	crystalline basement (metamorphites with weins of Granite)	

Jablunkov - 1 (Jabl-1)

0-1190 m	Silesian unit	
	-96 m	Krosno Formation
	-515 m	Menilitic Formation
	-294 m	Roznov Formation
-----------	overthrust	
	-619 m	Krosno Formation
	-725 m	Roznov Formation
-----------	overthrust	
	-100 m	Lhoty Formation
	-1190 m	Mazak Formation
-----------	-2091 m	basal overthrust of Flysch nappes (melange [109 m] from Silesian and Subsilesian units)
-2278 m	lower Miocene (Karpatian) of the Carpathian foredeep	
-2345 m	Lower Carboniferous - Culm facies (Hradec - Kyjovice Formation)	
-3002 m	Devonian to Lower Carboniferous	
	-2793 m	Lisen Formation (Krtiny Limestones and Drazovice Limestones)
	-3002 m	Macocha Formation (Lazanky and Vilemovice limestones.)
-3156 m	basal clastics (Old Red) - Devonian	
-3200 m	crystalline basement (Diorite)	

Janovice - 2 (Ja-2)

0-964.7 m	Silesian and Subsilesian units	
-----------	basal overthrust of Flysch nappes	
-969.2 m	lower Miocene (Karpatian) of the Carpathian foredeep	
-1485 m	Upper Cretaceous (Namurian A), Ostrava Formation	
-1694.5 m	Lower Carboniferous (Visean) to Upper Carboniferous (Namurian), Hradec-Kyjovice Formation (Culm)	

Karlin - 1 (Kar-1)

-3133 m	Zdanice-Hustopece Formation	
-3196 m	Submenilitic Formation	

----------	overthrust (Zdanice unit)		
-3985 m	Upper Cretaceous and autochthonous Paleogene		

Kobyli - 1 (Kob-1)

0-702 m	Cejc - Zajeci unit, Nemcice Formation (Upper Cretaceous to lower Eocene)		
----------	overthrust		
-3113 m	Waschberg-Zdanice-Subsilesian units (Zdanice sector)		
		-1444 m	Zdanice-Hustopece Formation (Egerian)
		-1504 m	Menilitic Formation (Oligocene)
		-1607 m	Nemcice Formation (Paleocene - Eocene)
----------	overthrust		
		-2897 m	Zdanice-Hustopece Formation (Egerian)
		-3113 m	Nemcice Formation (Paleocene - Eocene)
----------	overthrust		
-3131 m	Pouzdrany unit, Krepice Formation(?) (Oligocene-Miocene),		
----------	overthrust		
-3135 m	Waschberg-Zdanice-Subsilesian units (Zdanice sector)		
----------	basal overthrust of the Flysch nappes		
-4351 m	autochthonous Mesozoic (Mikulov Marls)		

Korycany - 4 (Kor-4)

0-1638 m	Raca unit of the Magura Group of nappes		
		-256 m	Zlin and Beloveza formations
		-312 m	Lukov Member of the Solan Formation
		-1638 m	Raztoky Member of the Solan Formation
-----------	overthrust of the Magura Group of nappes		
-1812 m	Zdounky unit (Upper Cretaceous to Eocene)		
-----------	basal overthrust of the Flysch nappes		
-1834 m	lower Miocene (Karpatian) of the Carpathian foredeep		
-2250 m	crystalline basement (Granodiorite)		

Kozlovice - 1 (SV-1)

-411 m	Silesian unit
-837 m	autochthonous Karpatian
-1076 m	Ostrava Formation (Upper Carboniferous)
-1421 m	Petrikovice Member (Upper Carboniferous)
-1422 m	Stur Member (Upper Carboniferous)
-1880 m	Culm facies
-2274 m	carbonate facies (Devonian)
-2283 m	crystalline basement

Krasna - 1 (Kra-1)

0-1707.5 m	Silesian unit (Godula subunit)		
		-1613 m	Godula Formation (Turonian - Santonian)
		-1707.5 m	Godula Lhoty Formation (Albian)
-------------	overthrust		
-1712.5 m	basal clastics (Devonian)		
-1812.5 m	crystalline basement (metamorphite)		
------------	overthrust (up to 1847 m tectonic breccia zone with Devonian to Upper Carboniferous and Cenomanian rocks		
-1969 m	Upper Carboniferous (Namurian) - Ostrava Formation		
-2074 m	Lower Carboniferous (Visean) - Culm facies		
		-2001 m	Hradec - Kyjovice Formation
		-2050 m	Moravice Formation
		-2074 m	Lisen Formation
-2190 m	Devonian to Lower Carboniferous - carbonate facies		
		-2079 m	Hady - Ricky Limestone (Visean)
		-2199 m	Macocha Formation (Givetian)
-2211 m	Devionian (Givetian) - basal clastics		
-2561 m	crystalline basement (metamorphites with diorite intrusions)		

Lubna - 2 (Lub-2)

0-350 m	Raca unit of the Magura Group of nappes		
		-80 m	Solan Formation (Lukov Member)
		-350 m	Solan Formation (Raztoky Member)
----------	overthrust of the Magura Group of nappes		
-426 m	Upper Cretaceous (Maastrichtian)		
----------	overthrust		
-1140 m	Waschberg - Zdanice - Subsilesian units (Zdanice sector)		
		-972 m	Zdanice - Hustopece Formation
		-1140 m	Nemcice Formation
----------	overthrust of the Zdanice unit		
-1870 m	Pouzdrany unit (Oligocene to Miocene strata)		
----------	basal overthrust of the Flysch nappes		
-1928 m	crystalline basement (Granite)		

Menin - 1 (Me-1)

0-49 m	middle Miocene (lower Badenian) of the Carpathian foredeep		
-410 m	Devonian (Givetian - Frasnian) - Macocha Formation		
-2100 m	basal clastics (Cambrian(?), Devonian)		

Mikulov - 1 (Mik-1)

0-1700 m	lower Miocene of the Carpathian foredeep		
	-1100 m	Karpatian, Laa Formation	
	-1700 m	Eggenburgian - Ottnangian, (basal clastics,146 m)	
-2520 m	autochthonous Mesozoic		
	-1990 m	Kurdejov Limestones (Kimmeridgian-Tithonian)	
	-2410 m	Mikulov Marls (Oxfordian-Kimmeridgian)	
	-2478 m	Vranovice Member (Callovian-Oxfordian)	
	-2520 m	Nikolcice Member (Callovian)	
-2589 m	crystalline basement (Granite)		

Musov - 2 (Mus-2)

0-973 m	lower Miocene of the Carpathian foredeep		
	-890.3 m	Karpatian (Laa Formation)	
	-973 m	Eggenburgian (Vestonice Member)	
-1895 m	autochthonous Mesozoic		
	-1185 m	Kurdejov Limestones (Kimmeridgian-Tithonian)	
	-1647 m	Mikulov Marls (Oxfordian-Kimmeridgian)	
	-1814 m	Vranovice Member (Callovian-Oxfordian)	
	-1895 m	Gresten Formation (bazal clastics)	
-1910.25 m	crystalline basement (Quarzdiorite)		

Nasedlovice - 1

0-2200 m	Zdanice unit of the Magura Group of nappes		
	-2200 m	Zdanice-Hustopece Formation	
	-2245 m	Menilitic Formation	
	-2465 m	Submenilitic Formation	
	-3190 m	autochthonous Paleogene	
	-3560 m	Carboniferous (Culm facies)	
	-4099 m	Carbonate facies (Devonian, Carboniferous)	
	-3792 m	Lisen Formation	
	-4367 m	Macocha Formation	
	-4367 m	basal clastics (Old Red)	
	-4450 m	crystalline basement	

Nikolcice - 2A (Nik-2A)

0-1025 m	Waschberg - Zdanice - Subsilesian units (Zdanice sector)		
	-775 m	Zdanice - Hustopece Formation (Egerian)	
	-853 m	Menilitic Formation (Oligocene)	
	-1025 m	Nemcice Formation (Paleocene - Oligocene)	
-----------	basal overthrust of the Flysch nappes		
-1265 m	autochthonous Paleogene, Nesvacilka Formation with clastics at the base (1201–1265 m)		
-2225 m	autochthonous Mesozoic		
	-1560 m	Kurdejov Limestones (Kimmeridgian)	
	-1936 m	Mikulov Marl (Oxfordian to Kimmeridgian)	
	-2137 m	Vranovice Member (Callovian to Oxfordian)	
	-2225 m	Nikolcice Member (Callovian)	
-2308 m	crystalline basement (Granodiorite)		

Osvetimany - 1 (Osv-1)

0-2278 m	Raca unit of the Magura Group of nappes		
	-130 m	Luhacovice Member of the Zlin Formation (Eocene)	
-----------	overthrust		
	-310 m	Lukov Member of the Solan Formation	
	-520 m	Raztoky Member of the Solan Formation	
-----------	overthrust		
	-1230 m	Zlin Formation (Eocene)	
	-1450 m	Beloveza Formation (Paleocene to Eocene)	
	-1660 m	Lukov Member of the Solan Formation (Paleocene)	
	-2278 m	Raztoky Member of the Solan Formation (Upper Cretaceous to Paleocene)	
-2480 m	Zdounky unit		
-----------	basal overthrust of the Flysch nappes		
-2518 m	autochthonous Paleogene		
-2820 m	crystalline basement (Granitoid)		

Potstat - 1 (Pot-1)

0-2866.5 m	Lower Carboniferous (Visean), Culm facies (Moravice Formation)		
-3741 m	Devonian (Givetian) to Lower Carboniferous (Visean), Carbonate facies (Macocha Formation)		

	-3221.5 m	Vilemovice Limestones	
------------	overthrust		
-3741 m	Devonian to Lower Carboniferous (Visean), Carbonate facies (Lisen Formation)		
	-3658 m	Ricky Limestones	
	-3741 m	Krtiny Limestones	
-4100 m	Devonian (Frasnian to Famenian) Limestones		

Roznov - 1 (Roz-1)

0-437.5 m	Verovice Formation (Aptian), Silesian unit
-2062 m	Hradiste Formation (Aptian-Valanginian), Silesian unit
------------	overthrust of the Silesian unit
-2240 m	Frydek Formation (Campanian), Subsilesian unit
-2544 m	Frydland Formation (Eocene-Paleocene), Subsilesian unit
------------	overthrust
-2809 m	Frydek Formation (Maastrichtian-Campanian), Subsilesian unit
-2872 m	Frydland Formation (Paleocene), Subsilesian unit
------------	overthrust
-2948 m	Frydek Formation (Upper Cretaceous), Subsilesian unit
-2989 m	Frydland Formation (Oligocene–Eocene), Subsilesian unit
------------	basal overthrust of the Flysch nappes
-3015.5 m	lower Miocene (Karpatian), Neogene foredeep

Sedlec - 1 (Sed-1)

0-3320 m	Waschberg - Zdanice - Subsilesian units (Waschberg sector):	
	-1530 m	Zdanice-Hustopece Formation
	-1680 m	Nemcice Formation
------------	overthrust	
-3320 m	Waschberg - Zdanice - Subsilesian units (Waschberg sector)	
	-3025 m	Zdanice-Hustopece Formation
	-3149 m	Menilitic Formation
	-3320 m	Nemcice Formation
	overthrust	
-3489 m	Pouzdrany unit	
-4765 m	autochthonous Mesozoic	
	-3600 m	Glauconitic sand Formation Upper Cretaceous
	-3860 m	Kurdejov Limestone
	-4720 m	Mikulov Marls
	-4765 m	Gresten Formation
-4850 m	crystalline basement (Granitoids)	

Slusovice - 1 (Slu-1)

0-3080 m	Raca unit of the Magura Group of nappes	
	-785 m	Zlin Formation
	-1717 m	Beloveza Formation
	-2900 m	Lukov Member of the Solan Formation
	-3080 m	Raztoky Member of the Solan Formation
-----------	basal overthrust of the Magura Group of nappes	
-3785 m	lower Miocene (Karpatian?)	
-3850 m	carbonate rocks (Devonian - Lower Carboniferous)	
-3877 m	basal clastics (Old Red) - Devonian	
-4000 m	crystalline basement (Metamorphites)	

Strachotin - 2 (Str-2)

0-805 m	Waschberg - Zdanice - Subsilesian units (Waschberg sector)	
	-100 m	Zdanice-Hustopece Formation
	-400 m	Nemcice Formation
	-485 m	Upper Cretacious (Palava and Klement formations)
	-660 m	Ernstbrunn Limestone
	-805 m	Klentnice Formation
------------	overthrust	
-1675 m	Pouzdrany unit (Pouzdrany and Boudky Marls with slivers of Jurassic rocks)	
------------	basal overthrust of the Flysch nappes	
-3075 m	autochthonous Mesozoic	
	-1880 m	Glauconitic sand Formation (Upper Cretaceous)
	-2208 m	Kurdejov Limestones
	-2750 m	Mikulov Marls
	-2910 m	Vranovice Carbonates
	-2997 m	Nikolcice Formation
	-3075 m	Gresten Formation
-3147 m	crystalline basement (Granitoids)	

Stupava - 1 (Stup-1)

0-1969 m	Raca unit of the Magura Group of nappes	
	-205 m	Beloveza Formation

	-365 m	Lukov Member
	-1245 m	Raztoky Member
-----------	overthrust	
	-182.05 m	Lukov Member of the Solan Formation
	1965 m	Roztoky Member of the Solan Formation
-----------	overthrust of the Magura Group of nappes	
-2156 m	Zdounky unit	
	-2065 m	Lower sekvence (Albian)
	-2156 m	Upper sekvence (Eocene)
-----------	basal overthrust of the Flysch nappes	
-2431 m	lower Miocene (Karpatian) of the Carpathian foredeep	
-2500 m	crystalline basement (Granodiorite)	

Susice - 1 (SU-1)

0-215 m	Neogene of the Vienna Basin	
-3525 m	Raca unit of the Magura group of nappes	
	-2540 m	Zlin Formation
	-2590 m	Beloveza Formation
	-3320 m	Solan Formation
	-3525 m	Kaumberg Formation
	overthrust of the Magura Group of nappes	
-3670 m	Zdounky unit	
-----------	basal overthrust of the Flysch nappes	
-3760 m	autochthonous Paleogene	
-3810 m	crystalline basement	

Tesany - 1 (Te-1)

0-465 m	Waschberg - Zdanice - Subsilesian units (Zdanice sector)
-----------	overthrust
-610 m	Pouzdrany unit
-----------	basal overthrust of the Flysch nappes
-1905 m	autochthonous Paleogene (Nesvacilka Formation) with basal clastics (505 m)
-2980 m	Lower Carboniferous (Culm facies)
-4090 m	Devonian to Early Carboniferous (Carbonate facies)
-4290 m	basal clastics (Old Red) - Devonian
-4500 m	crystalline basement (Granodiorite)

Uhrice - 1 (Uh-1)

- 2140 m	Waschberg-Zdanice-Subsilesian units, Zdanice sector	
	-1978 m	Zdanice-Hustopece Formation
	-2002 m	Menilitic Formation
	-2140 m	Nemcice Formation
-----------	basal overthrust of the Flysch nappes	
-2763 m	autochthonous Paleogene, Nesvacilka Formation with clastics at the base (2697–2763 m)	
-3043 m	Upper Carboniferous	
-3574 m	Devonian	
-3831 m	Devonian, basal clastics	
-3960 m	crystalline basement, metamophites	

Valasske Mezirici - 1 (VM-1)

0-1169 m	Silesian unit
basal overthrust of the Flysch nappes	
-2948 m	carbonate facies (Devonian, Carboniferous)
-2980 m	basal clastic (Old Red)
-3036 m	crystalline basement

Vranovice - 1 (VRA-1)

0–841 m	Waschberg - Zdanice - Subsilesian units (Zdanice sector)	
	-600 m	Zdanice - Hustopece Formation (Egerian)
	-632 m	Menilitic Formation (Oligocene)
	-841 m	Nemcice Formation (Paleocene - lower Oligocene)
-----------	basal overthrust of the Flysch nappes	
-1172 m	autochthonous Paleogene - Nesvacilka Formation	
-1565 m	autochthonous Mesozoic	
	-1355 m	Mikulov Marls (Oxfordian - Kimmeridgian)
	-1450 m	Vranovice Member (Callovian - Oxfordian)
	-1565 m	Nikolcice Member a Gresten Formation
-1750 m	crystalline basement (Granodiorite)	

Zdanice - 13 (Zd-13)

0-929 m	Waschberg - Zdanice - Subsilesian units (Zdanice sector)	
	-650 m	Zdanice-Hustopece Formation (Egerian)
	-730 m	Menilitic Formation (Oligocene)
	-929 m	Nemcice Formation (Paleocene - lower Oligocene)

	-----------	basal overthrust of the Flysch nappes
	-975 m	lower Miocene of the Carpathian foredeep
	-1054 m	crystalline basement (Plutonic rocks)
Zarosice - 1 (Za-1)		
	0-1509 m	Waschberg - Zdanice - Subsilesian units (Zdanice sector)
		-1257 m Zdanice - Hustopece Formation (Oligocene - lower Miocene)
		-1509 m Nemcice Formation (Paleocene - Eocene)
		basal overthrust of the Flysch nappes
	-1873 m	autochthonous Paleogene (Nesvacilka Formation)
	-2274 m	Carboniferous (Visean - Namurian) - Ostrava Formation
	-2330 m	Lower Carboniferous (Tournaisian - Visean), Culm facies (Hradec-Kyjovice Formation)
	-2760 m	Devonian (Givetian) to Lower Carboniferous (Visean), carbonate facies (Macocha Formation)
	-2867 m	Devonian - basal clastics (Old Red).

APPENDIX 2: AUTHOR'S NAMES IN ASCII AND WITH DIACRITICS

Author's names not containing diacritics are not listed.

ASCII	With Diacritics
Adamek, J.	Adámek, J.
Adamova, M.	Adamová, M.
Arzmuller, G.	Arzmüller, G.
Baldi, T.	Báldi, T.
Baldi-Beke, M.	Báldi-Beke, M.
Benesova, E.	Benešová, E.
Beranek, B.	Beránek, B.
Blazej, J.	Blažej, J.
Blizkovsky, M.	Blížkovský, M.
Bodis, D.	Bodiš, D.
Bohacek, Z.	Boháček, Z.
Boucek, B.	Bouček, B.
Brzobohaty, J.	Brzobohatý, J.
Brzobohaty, R.	Brzobohatý, R.
Bubik, M.	Bubík, M.
Cahelova, J.	Cahelová, J.
Caslavsky, J.	Čáslavský, J.
Cech, S.	Čech, S.
Cejchan, P.	Čejchan, P.
Cekan, J.	Čekan, J.
Cermak, V.	Čermák, V.
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Chlupac, I.	Chlupáč, I.
Chmelik, F.	Chmelík, F.
Cizek, P.	Čížek, P.
Ctyroka, J.	Čtyroká, J.
Ctyroky, P.	Čtyroký, P.
Dlabac, M.	Dlabač, M.
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Durica, D.	Ďurica, D.
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Dvorak, J.	Dvořák, J.
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Francu, J.	Franců, J.
Gasparikova, V.	Gašparíková, V.
Gotzinger, G.	Götzinger, G.
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Jiricek, R.	Jiříček, R.
Jurasova, F.	Jurášová, F.
Jurkova, A.	Jurková, A.
Juttner, K.	Jüttner, K.
Kalasek, J.	Kalášek, J.
Klvana, J.	Klvaňa, J.
Kohler, E.	Köhler, E.
Konecny, V.	Konečný, V.
Korab, T.	Koráb, T.
Kovac, M.	Kováč, M.
Krejci, J. J.	Krejčí, J. J.
Krejci, O.	Krejčí, O.
Krhovsky, J.	Krhovský, J.
Kroll, A.	Kröll, A.
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Kulmannova, A.	Kulmanová, A.
Lesko, B.	Leško, B.
Mahel, M.	Mahel', M.
Malkovsky, M.	Malkovský, M.
Maly, L.	Malý, L.
Martinek, K.	Martínek, K.
Maska, M.	Maška, M.
Matejka, A.	Matějka, A.
Mencik, E.	Menčík, E.
Michalik, J.	Michalík, J.
Minarilova, D.	Minaříková, D.
Molcikova, V.	Molčíková, V.
Muller, C.	Müller, C.

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Mysik, M.	Mišík, M.	Ruzicka, M.	Růžička, M.
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Nemcok, M.	Němčok, M.	Schroder, B.	Schröder, B.
Nemec, F.	Němec, F.	Sedlak, J.	Sedlák, J.
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Obstova, V.	Obstová, V.	Shrbeny, O.	Shrbený, O.
Pagac, I.	Pagáč, I.	Sikula, J.	Šikula, J.
Palensky, P.	Pálenský, P.	Slaczka, A.	Ślączka, A.
Pecova, J.	Pečová, J.	Smid, B.	Šmíd, B.
Peslova, H.	Peslová, H.	Sotak, J.	Soták, J.
Picha, F.	Pícha, F.	Spisiak, J.	Spišiak, J.
Pitonak, P.	Pitoňák, P.	Stankova, E.	Staňková, E.
Plancar, J.	Plančár, J.	Stastny, M.	Št’astný, M.
Plasienka, D.	Plašienka, D.	Stranik, Z.	Stráník, Z.
Plicka, M.	Plička, M.	Svabenicka, L.	Švábenická, L.
Pokorny, V.	Pokorný, V.	Svancara, J.	Švancara, J.
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APPENDIX 3: GEOGRAPHICAL AND GEOLOGICAL TERMS IN ASCII AND WITH CZECH DIACRITICS

Terms not containing Czech diacritics are not listed.

ASCII	**With Diacritics**
Antoninek Formation	Antonínek Formation
Banov	Bánov
Baska Formation, facies, unit	Baška Formation, facies, unit
Baska ridge	Baška ridge
Beloveza Formation	Beloveža Formation
Bile (Biele in Slovak) Karpaty Mountains	Bílé (Biele in Slovak) Karpaty Mountains
Bilovice Formation, field	Bílovice Formation, field
Bohuslavice nad Vlarou	Bohuslavice nad Vlárou
Bosaca unit	Bošáca unit
Bosovice-1 well, field	Bošovice-1 well, field
Breclav field	Břeclav field
Brezina shale	Březina shale
Brezno Formation	Březno Formation
Brodske field	Brodské field
Bruzovice–Frydek field	Bruzovice–Frýdek field
Bystrice nad Olsi	Bystřice nad Olší
Bystrice pod Hostynem	Bystřice pod Hostýnem
Bystrice pod Hostynem-1 well	Bystřice pod Hostýnem-1 well
Bzova Sandstone	Bzová Sandstone
Cejc–Zajeci unit	Čejč–Zaječí unit
Chabova Formation	Chabová Formation

Choryne-9 well, field	Choryně-9 well, field
Chriby Mountains	Chřiby Mountains
Chvalcov Member	Chvalčov Member
Damborice field	Dambořice field
Damborice Group	Dambořice Group
Damborice-1 well	Dambořice-1 well
Detmarovice paleovalley	Dětmarovice paleovalley
Divaky Formation	Diváky Formation
Dolni Bojanovice	Dolní Bojanovice
Dolni Dunajovice-1 well, field	Dolní Bojanovice-1 well, field
Dolni Lomna-3 well, field	Dolní Lomná-3 well, field
Dolni Tesice	Dolní Těšice
Drazovice Limestone	Dražovice Limestone
Drazuvky	Dražůvky
Dubnany Formation	Dubňany Formation
Frydek Formation	Frýdek Formation
Frydlant Formation	Frýdlant Formation
Hady	Hády
Hady–Ricky Limestones	Hády–Říčky Limestones
Handlova	Handlová
Hnojnik-1 well	Hnojník-1 well
Hodonin	Hodonín
Hodonin field	Hodonín field
Holesov fault	Holešov fault
Holesov-1 well	Holešov-1 well
Holic structure	Holíč structure
Horni Zukov field	Horní Žukov field
Hosteradky-1 well	Hostěrádky-1 well
Hostyn Hills	Hostýn Hills
Hostyn Member, zone	Hostýn Member, zone
Hradiste Formation, graben	Hradiště Formation, graben
Hrusky Formation, field	Hrušky Formation, field
Hrusky–Josefov field	Hrušky–Josefov field
Hrusovany Limestones	Hrušovany Limestones
Hrusovany-1 well	Hrušovany-1 well
Hustopece Marls	Hustopeče Marls
Istebna Formation	Istebná Formation
Ivan Member	Iváň Member
Jablunka	Jablůnka
Jablunka-1 well	Jablůnka-1 well
Jarosov	Jarošov
Jarosov-1 well	Jarošov-1 well
Javorina Formation	Javořina Formation
Jesenik	Jeseník
Jezov-3 well	Ježov-3 well
Juhyne River	Juhyně River
Karlin-1 well, field	Karlín-1 well, field
Karvina Formation	Karviná Formation
Kelc subunit, facies	Kelč subunit, facies
Kisovce–Svabovce	Kišovce–Švábovce
Klanecnica	Klanečnica
Kloboucky field	Kloboučky field
Klokocov Member	Klokočov Member
Koberice Gypsum	Kobeřice Gypsum
Kobyli	Kobylí
Kobyli-1 well	Kobylí-1 well
Kojetin Formation	Kojetín Formation
Komorni Lhotka field	Komorní Lhotka field
Koprivnice Limestones	Kopřivnice Limestones

Koprivnice–Ticha field	Kopřivnice–Tichá field
Korycany-4 well, field	Koryčany-4 well, field
Kostelany Vychod field	Kostelany Východ field
Kostelany Zapad field	Kostelany Západ field
Kotouc facies	Kotouč facies
Kotouc Hills	Kotouč Hills
Kozusice-4 well	Kožušice-4 well
Krasna-1 well, field, elevation	Krásná-1 well, field, elevation
Krasna–Moravka field	Krásná–Morávka field
Krepice Formation	Křepice Formation
Krive Member	Křivé Member
Krmelin field	Krmelín field
Krtiny limestones	Křtiny limestones
Krumvir field	Krumvíř field
Kurdejov Limestones	Kurdějov Limestones
Kurim	Kuřim
Kuty	Kúty
Kuzelov Formation	Kuželov Formation
Kycera Member, zone	Kyčera Member, zone
Lab sands, ostracoda beds	Láb sands, ostracoda beds
Laksary Formation, schlier	Lakšáry Formation, schlier
Lanzhot Formation, field	Lanžhot Formation, field
Lanzhot–Hrusky fault	Hrušky–Lanžhot fault
Lazanky Limestones	Lažánky Limestones
Lhotka–Pstruzi field	Lhotka–Pstruží field
Lisen Formation	Líšeň Formation
Lower Tesin Member	Lower Těšín Member
Lubna-2 well, field, massif, elevation	Lubná-2 well, field, massif, elevation
Luhacovice Member, zone	Luhačovice Member, zone
Lukovecek klippe	Lukoveček klippe
Lulec Conglomerates	Luleč Conglomerates
Luzice Formation, sands, field	Lužice Formation, sands, field
Luzice (Lusatian) fault	Lužice (Lusatian) fault
Lysa hora Mountain	Lysá hora Mountain
Malesovice Formation	Malešovice Formation
Manin unit	Manín unit
Mazak Formation	Mazák Formation
Menin-1 well	Měnín-1 well
Menin–Zatcany field	Měnín–Žatčany field
Mikulcice	Mikulčice
Milesovice-1 well	Milešovice-1 well
Moravske naftove doly, oil company	Moravské naftové doly, oil company
Mourinov field	Mouřínov field
Mur–Murtz–Zilina line	Mur–Murtz–Žilina line
Musov Member	Mušov Member
Musov-2 well	Mušov-2 well
Mutenice field	Mutěnice field
Nasedlovice-1 well	Násedlovice-1 well
Nemcice Formation	Němčice Formation
Nemcicky	Němčičky
Nemcicky-1, 2 wells, field	Němčičky-1, 2 wells, field
Nemetice Formation	Němetice Formation
Nesvacilka Formation, paleovalley, graben	Nesvačilka Formation, paleovalley, graben
Nesvacilka-1, 2, 3 wells	Nesvačilka-1, 2, 3 wells
Nikolcice Member	Nikolčice Member
Nikolcice-1, 2A wells, field	Nikolčice-1, 2A wells, field
Nitkovice	Nítkovice
Nitkovice–Hradisko field	Nítkovice–Hradisko field
Nizky Jesenik Mountains	Nízký Jeseník Mountains

Nova Hora	Nová Hora
Nove Mlyny limestones	Nové Mlýny limestones
Nove Mlyny-1 well	Nové Mlýny-1 well
Novy Prerov field	Nový Přerov field
Novy Prerov Member	Nový Přerov Member
Ondrasovec Member	Ondrášovec Member
Oravska Magura	Oravská Magura
Oravska Polhora	Oravská Polhora
Orlova fold	Orlová fold
Ostrava–Karvina ridge	Ostrava–Karviná ridge
Osvetimany-1 well	Osvětimany-1 well
Palava Formation	Pálava Formation
Palkovice Formation, Hills	Pálkovice Formation, Hills
Pasohlavky Limestones	Pasohlávky Limestones
Podvorov–Mutenice field	Podvorov–Mutěnice field
Postorna field	Poštorná field
Potstat-1 well	Potštát-1 well
Pouzdrany Formation, Marls, unit	Pouzdřany Formation, Marls, unit
Pribor Jih (Stramberk) field	Příbor Jih (Štramberk) field
Pribor–Klokocov field	Příbor–Klokočov field
Pribor–Tesin ridge	Příbor–Těšín ridge
Pritluky	Přítluky
Proc Formation	Proč Formation
Prusanky field	Prušánky field
Puchov Marls	Púchov Marls
Raca unit	Rača unit
Racice Conglomerates	Račice Conglomerates
Rajec	Rájec
Ratiskovice	Ratíškovice
Raztoka Member	Ráztoka Member
Rostin	Roštín
Rostin-2 well	Roštín-2 well
Roznov Formation, area	Rožnov Formation, area
Roznov-1 well, field	Rožnov-1 well, field
Rozstani Shales	Rozstání Shales
Sakvice Marls, depression	Šakvice Marls, depression
Sastin sands	Šaštín sands
Sitborice Member	Šitbořice Member
Slavkov–Tesin ridge	Slavkov–Těšín ridge
Slusovice-1 well	Slušovice-1 well
Sneznica Formation	Sněžnica Formation
Solan Formation	Soláň Formation
Spis	Spiš
Stara Tura	Stará Turá
Stare Hute	Staré Hutě
Staric–Liskovec–Sviadnov field	Staříč–Lískovec–Sviadnov field
Stary Jicin	Starý Jičín
Sternberk–Benesov suture	Šternberk–Benešov suture
Strachotin-2 well	Strachotín-2 well
Stramberk klipppe, reef, platform, field	Štramberk klippe, reef, platform, field
Stramberk Limestones	Štramberk Limestones
Stranska skala	Stránská skála
Straz Sandstones	Stráž Sandstones
Susice-1 well	Sušice-1 well
Svedske Sance	Švédské Šance
Tesany Formation	Těšany Formation
Tesany-1 well	Těšany-1 well
Tesin–Hradiste Formation	Těšín–Hradiště Formation
Tesin nappe	Těšín nappe

Tesin (Teschen) Limestones	Těšín (Teschen) Limestones
Tlumacov Marls	Tlumačov Marls
Tri Kameny zone	Tři Kameny zone
Trnava–Staskov zone	Trnava–Staškov zone
Tynec–Cunin field	Týnec–Cunín field
Tynec structure, field	Týnec structure, field
Tynec-34 well	Týnec-34 well
Uhercice Formation	Uherčice Formation
Uhersky Brod	Uherský Brod
Uhrice Jih (South) field	Uhřice Jih (South) field
Uhrice-1 well, field	Uhřice-1 well, field
Ujezd Member	Újezd Member
Ustgrun	Ustgruň
Vah line	Váh line
Valasske Mezirici	Valašské Meziříčí
Valasske Mezirici-1 well	Valašské Meziříčí-1 well
Velke Bilovice–Zizkov–Podivin fields	Velké Bílovice–Žižkov–Podivín fields
Vendryne Formation	Vendryně Formation
Verona–Semmering–Vah fault	Verona–Semmering–Váh fault
Verovice Formation, Shales	Veřovice Formation, Shales
Vestonice Formation, fault	Věstonice Formation, fault
Vilemovice Limestones	Vilémovice Limestones
Vlara subunit	Vlára subunit
Vsetin Member, zone	Vsetín Member, zone
Waschberg–Zdanice–Subsilesian units, nappes	Waschberg–Ždánice–Subsilesian units, nappes
Zahori fault	Záhoří fault
Zajeci	Zaječí
Zarosice-1 well, field	Žarošice-1 well, field
Zatcany field	Žatčany field
Zavod Formation, schlier	Závod Formation, schlier
Zdanice–Hustopece Formation	Ždánice–Hustopeče Formation
Zdanice Jih–Zdanice Sever–Kloboucky field	Ždánice Jih–Ždánice Sever–Klobouky field
Zdanice Sandstone, unit, nappe, massif, elevation, basin	Ždánice Sandstone, unit, nappe, elevation, basin
Zdanice-13 well, field	Ždánice-13 well, field
Zerotice Member	Žerotice Member
Zilina	Žilina
Zizkov Member, field	Žižkov Member, field
Zlin Formation	Zlín Formation
Zohor–Plavec graben	Zohor–Plaveč graben

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