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GEODYNAMIC EVOLUTION OF THE OROGEN: THE WEST CARPATHIANS AND OUACHITAS CASE STUDY

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Abstract: Twelve time interval maps have been presented which depict the plate tectonic configuration, paleogeography and lithofacies for the circum-Carpathian area from the Late Carboniferous through Neogene and for the circum-Ouachita region from Late Cambrian through Early Permian.

The following geodynamic evolutionary stages can be distinguished in these two orogens: Stage I – rifting of terranes off the major continent, forming oceanic basins (Triassic–Early Cretaceous in the Carpathian region, Cambrian–Devonian in the Ouachita region); Stage II – formation of subduction zones along the active margin, partial closing of oceanic basin, development of deep-water flysch basin associate with this rifting on the platform (passive margin) with the attenuated continental crust (Late Cretaceous–Paleocene in the Carpathian region, Early Carboniferous in the Ouachitas); Stage III – collision, perhaps terrane–continent, with the accompanying convergence of two large continents, development of accretionary prisms, Eocene–Early Miocene time in the Carpathian region, Late Carboniferous in the Ouachitas; and Stage IV – postcollisional, (Miocene–Present–future? in the Carpathians, Permian–Triassic in the Ouachitas). Both, Carpathians and Ouachitas are accretionary prisms formed in response to terrane-continent and continent-continent collision. The paleogeographic approach we have taken shows how these mountain belts were constructed through the orogenic cycle, which reflects complex plate tectonic processes. Carpathians and Ouachitas record complete and homologous Wilson cycle.

Key words: Plate tectonics, paleogeography, orogen, accretionary prism, Wilson cycle, Ouachita, Carpathians.

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INTRODUCTION

The aim of this paper is to compare the plate tectonic evolution and position of the major crustal elements of the Carpathians, and Ouachita Mountains within a global framework and to show the relationship between tectonic processes and sedimentary record during their orogenic cycles. The Carpathian and Ouachita Mts are geographically distant. The age of the formation of each orogen is also quite different. The Carpathians mountain belt formed during Mesozoic and Cenozoic time, while the Ouachitas formed in the Paleozoic; their tectonostratigraphic history, however, displays striking similarities.

Both orogens have been the subject of numerous classic sedimentological studies of the deep-water flysch deposits (Książkiewicz, 1954; Dżułyński & Ślączka, 1958; Dżułyński *et al.*, 1959; Dżułyński & Walton, 1965; Cline, 1960, 1970; Lowe, 1976, 1989; McBride, 1975; Morris, 1974, 1989; Moiola & Shanmugan, 1984; Pescatore & Ślączka, 1984; Shanmugan & Moiola, 1995; Picha, 1996). The Ouachita orogenic belt was also a subject of a tectonic synthesis from the point of view of Wilson cycle (Viele & Thomas, 1989). The history of the whole circum-Carpathian realm is more complex (e.g., see Golonka *et al.*, 2000, 2003a), if we concentrate however on the West Carpathians, the orogenic cycle is also clear and evident. Thus, these two orogens could provide a good example to compare the modern and ancient orogens.

Twelve time interval maps have been presented which depict the plate tectonic configuration, paleogeography and lithofacies for the circum-Carpathian region and adjacent areas from the Late Carboniferous through Neogene and for the circum-Ouachita region from the Late Cambrian through Early Permian.

The maps were constructed using the following defined steps:

1. Construction of the base maps using the plate tectonic model. These maps depict plate boundaries (sutures), plate position at the specific time and outline of present day coastlines.

2. Review of existing global and regional paleogeographic maps.

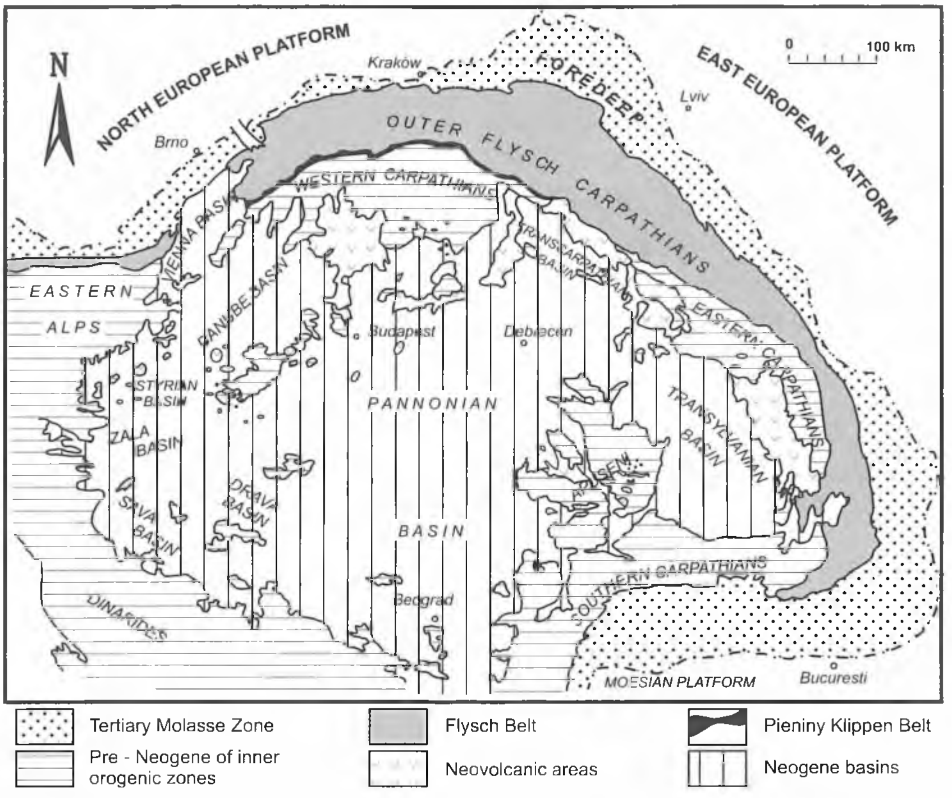

Fig. 1. Tectonic sketch map of the Alpine-Carpathian-Pannonian-Dinaride basin system (after Kováč *et al.*, 1998; simplified)

3. Posting of generalised facies and paleoenvironment database information on base maps.

4. Interpretation and final assembly of computer map files.

The maps were constructed using a plate tectonic model, which describes the relative motions between approximately 300 plates and terranes. This model was constructed using PLATES and PALEOMAP software (see Golonka *et al.*, 1994, 2000, 2003a) which integrate computer graphics and data management technology with a highly structured and quantitative description of tectonic relationships. The heart of this program is the rotation file, which is constantly updated, as new paleomagnetic data become available. Hot-spot volcanics serve as reference points for the calculation of paleolongitudes (Golonka & Bocharova, 2000). Ophiolites and deep-water sediments mark paleo-oceans, which were subducted and included into foldbelts. Magnetic data have been used to define paleolatitudinal position of continents and rotation of plates (see e.g. Van der Voo; 1993; Besse & Courtillot 1991; Krs *et al.*, 1996). An attempt has been also made to utilise the paleomagnetic date from minor plates and allochtonous terranes (see e.g. Patrascu *et al.*, 1992, 1993; Pechersky & Safronov 1993; Beck & Schermer 1994; Channell *et al.*, 1992, 1996; Mauritsch *et al.*, 1995, 1996; Feinberg *et al.*, 1996; Krs *et al.*, 1996; Márton and Martin 1996; Márton *et al.*, 1999, 2000; Haubold *et al.*, 1999; Muttoni *et al.*, 2000a, b; Grabowski, 2000). The nature of rotation indicated by paleomagnetism measured in sedimentary rocks in allochthonous terranes remains somewhat uncertain. It could be caused by the rotation of crustal (basement) elements, rotation of blocks separated by dextral faults (e.g. Márton *et al.*, 2000) or rotation of thrust sheets (e. g. Muttoni *et al.*, 2000b). Measurement in flysch deposits could also indicate the arrangement of magnetised grains (domains) by turbiditic currents. For example, the magnetic declination of Podhale Flysch in Poland records perhaps the sedimentological arrangement of grains (see the maps of sedimentological transport in the Carpathian flysch, *e.g.* Ksiazkiewicz, 1962) changed by the crustal rotation of the Inner Carpathian plate to the present position. The crustal rotation in a range of 20–30° agrees with the general geodynamic evolution of the area (Golonka *et al.*, 2000).

Information from several general and regional paleogeographic papers was filtered and utilized (e.g., Ronov et al., 1984, 1989; Dercourt *et al.*, 1986, 1993, 2000; Ziegler, 1988, 1989; Stampfli *et al.*, 1991, 1998, 2001; Stampfli, 2001; Kováč *et al.*, 1998; Plašienka, 1999; Neugebauer *et al.*, 2001; Golonka & Ford, 2000; Golonka *et al.*, 2000, 2003a). The authors of this paper also used unpublished maps and databases from the PALEOMAP group (University of Texas at Arlington), PLATES group (University of Texas at Austin), University of Chicago, Institute of Tectonics of Lithospheric Plates in Moscow, Robertson Research in Llandudno, Wales, and the Cambridge Arctic Shelf Programme. The plate and terrane separation was based on the PALEOMAP system (see Scotese & Lanford, 1995), with modifications in the Tethys area (Golonka *et al.*, 2000, 2003a). The calculated paleolatitudes and paleolongitudes were used to generate computer maps in the Mi-

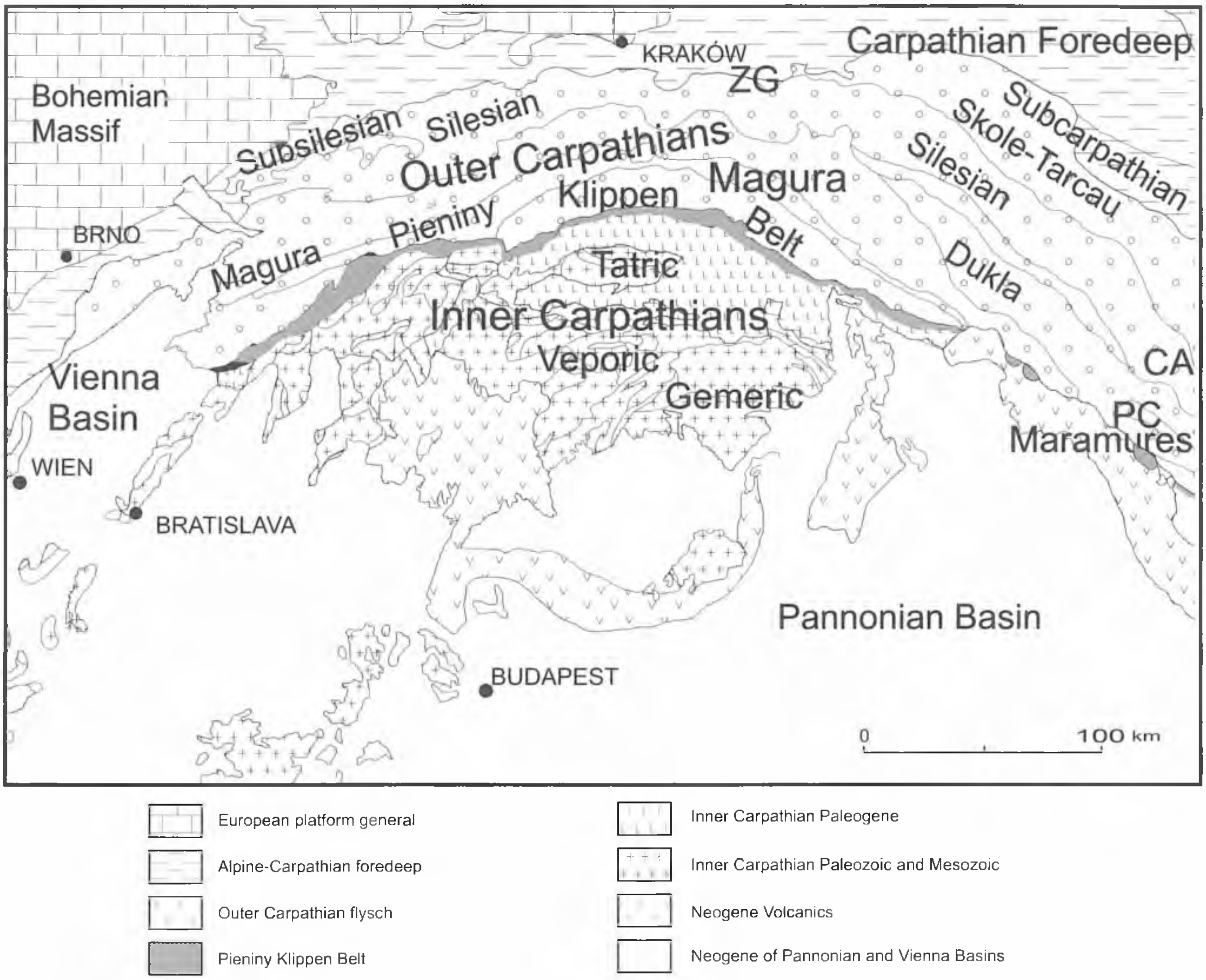

Fig. 2. Geological map of West Carpathians and adjacent areas. (Modified from Wessely and Liebl, 1996). Abbreviations: ZG – Zgłobice-Wieliczka Unit, CA – Charnahora-Audia Unit, PC – Porkulec-Convolute Unit

crostation design format using the equal area Molweide projection.

This paleogeographic approach shows how a mountain belt was constructed through the orogenic cycle, which reflects complex plate tectonic processes. Some unsolved questions and problems could be answered by a comparison between one and another orogen.

TECTONIC SETTING

WEST CARPATHIANS

The Carpathians form a great arc of mountains, which stretches more than 1 300 km from the Vienna Forest to the Iron Gate on the Danube (Fig. 1). On the west the Carpathians are linked with the eastern Alps and on the east pass into the Balkan chain. Traditionally the Carpathians are subdivided into the West and East Carpathians (Mahel, 1974). The West Carpathians consists of an older range known as the Inner or Central Carpathians and the younger one, known as the Outer or Flysch Carpathians (Mahel, 1974; Książkiewicz, 1977; Ślączka & Kaminski, 1998). At the boundary of these two ranges lies the Pieniny Klippen Belt (PKB) (Fig. 2). The Inner Carpathians are regarded as a prolongation of the Northern Calcareous Alps, and formed part of the Apulia plate in regional sense that is a promontory of the African plate (Picha, 1996). They are divided into the Tatric, Veporic and Gemeric nappes (Fig. 2) that are the prolongation of the Lower, Middle and Upper Austroalpine nappes respectively (Pescatore & Ślączka, 1984). The Inner Carpathians nappes contact along a Tertiary strike-slip boundary with Pieniny Klippen Belt (Fig. 2, 3). The Pieniny Klippen Belt (PKB) is composed of several successions (mainly deep and shallow-water limestones), covering a time span from the Early Jurassic to Paleogene (Golonka & Sikora, 1981; Birkenmajer, 1986). This strongly tectonized structure is a terrain of about 800 km long and 1 to 20 km wide, which stretches from Vienna on the West to the Maramures (or Poiana Botizii area, NE Romania) on the East (Fig. 2). The PKB is in the western part of the area thrust over the Outer Carpathian nappes (Fig. 3), in Poland and

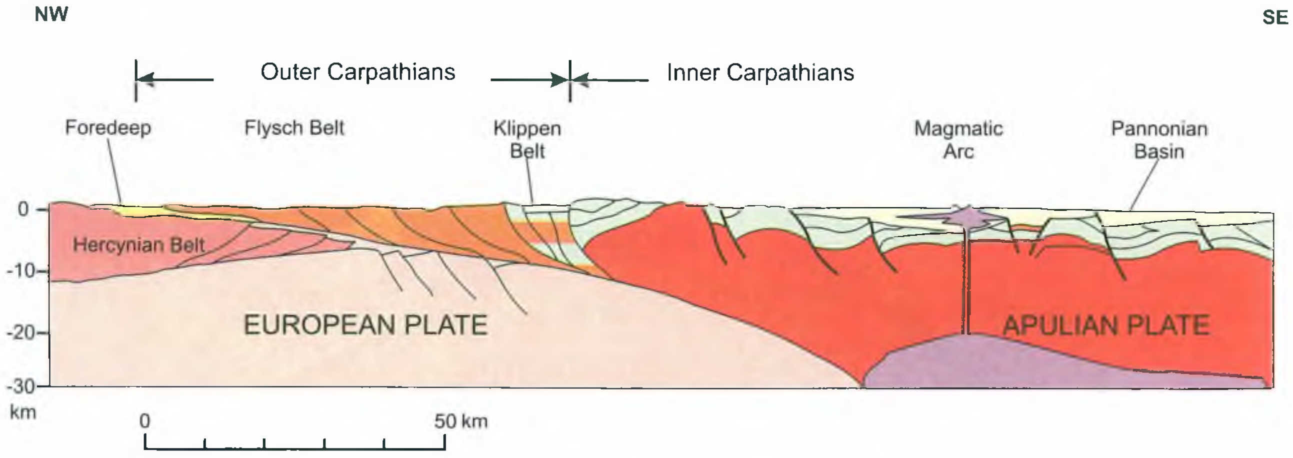

Fig. 3. Generalized cross-section across Carpathian-Pannonian region (Picha, 1996)

Eastern Slovakia is separated from the Outer Carpathians by a Miocene subvertical strike-slip fault (Birkenmajer, 1986). The Outer Carpathians are built up of a stack of nappes and thrustsheets changing along the Carpathians, built mainly of continual flysch sequences up to six kilometers thick representing the time span from the uppermost Jurassic up to the Lower Miocene. All the Outer Carpathian nappes are overthrust onto the southern part of the North European platform covered by the autochtonous Miocene deposits of the Carpathian Foredeep on the distance of 70 km, at least (Książkiewicz, 1977; Pescatore & Ślączka, 1984). During overthrusting movement the northern Carpathians nappes became uprooted from the basement and only their basinal parts were preserved (Fig. 3). A narrow zone of folded Miocene deposits was developed along the frontal Carpathian thrust i.e. and Zgłobice Wieliczka Unit in the Northern Carpathians (ZG – Fig. 2) and Subcarpathian (Borislav–Sambor–Rozniatov Unit) of the Ukrainian and Romanian parts of the Eastern Carpathians (Fig. 2; Książkiewicz, 1977; Pescatore & Ślączka, 1984; Kruglov, 1989).

The deep structure of the Polish Outer Carpathians and its basement, that is southern prolongation of the North European Platform, has been recognised by deep boreholes as well as by magnetotelluric, gravimetric, magnetic, geomagnetic, and deep seismic sounding profiles (Ślączka, 1976; Oszczypko *et al.*, 1989; Picha, 1996; Guterch *et al.*, 2001). Tens of deep (up to 4500 meters) boreholes, which reached the Carpathian substratum allowed, recognise depth of Carpathian thrust plane, its minimal range as well as character of substratum. Generally the thrust plane of the Carpathians deep slowly to the south in their western part. Seismic data provides similar value of the thrust plane depth. The position of the crust-mantle boundary (Moho) has been recognised along several seismic (Guterch *et al.*, 2001). The depth to the Moho discontinuity generally ranged from 30–40 km at the front of the central part of the Carpathians and increases to 50 km south of the Nowy Sącz. South of PKB this value decreases to 36–37 km. Depth of consolidated basement beneath Carpathian is situated at depth of 10–18 km. The thickness of the lithosphere in the Polish Carpathians varies from 160 km near Kraków to 100 km in the PKB.

OUACHITA FOLDBELT

The Ouachita Mountains in Oklahoma and Arkansas are a surface expression of the Ouachita belt (Fig. 4), which continues in the subsurface to the southwest, surfacing once again in the Marathon region of West Texas near the Mexican border (Arbenz, 1989; Viele & Thomas, 1989). The continuation of the foldbelt beneath the surface into Mexico is at least uncertain and speculative. Structural, lithofacies and Pb isotopic data (*e.g.* Dickinson & Lawton, 2001) do not support this continuation. To the east, the Ouachita orogenic belt converges with the southern Appalachian belt in the area of the central Mississippi uplift (beneath a thick Mesozoic cover). Our knowledge of this junction remains speculative (Thomas, 1977, 1989).

Structures in the exposed part of the Ouachita orogenic belt are folds and thrusts vergent toward the North American craton with major decollements thrusts visible especially along some parts of the orogenic front (Fig. 5). In most of the Ouachita Mountains, the structures trend east, but curve strongly to the southwest at the western end in Oklahoma and slightly southeastward at their eastern end in Arkansas.

Among the structures there are several anticlinoria such as the Broken Bow (BRU), Benton (BU) and Potato Hill (PH – Fig. 4) uplifts, where older rocks of the Ouachita facies are exposed (Arbenz, 1968, 1989; Viele & Thomas, 1989).

Generally all the exposed rocks in the Ouachita Mountains are strongly allochthonous and have been thrust as far as 80 kilometers northward from their former position. The overthrust surface is steep in the northern part of orogen, southward it is nearly horizontal. The major elevated zone can be distingushed here (e.g., zone drilled by the Hassie Hunt Carl Neely well on Fig. 5). The slope of the southwest surface is strongly dependent on the basement configuration (Lillie *et al.*, 1983; Golonka, 1988).

The orogenic belt faces northwestward and northward toward the craton where variable thicknesses of Paleozoic rocks lie on the Precambrian basement. Positive basement areas include: the Nashville dome, Ozark Uplift, the Arbuckle (AM) and Wichita Mountains (WM) (southwestern Oklahoma) and the Llano uplift of Central Texas (Fig. 4).

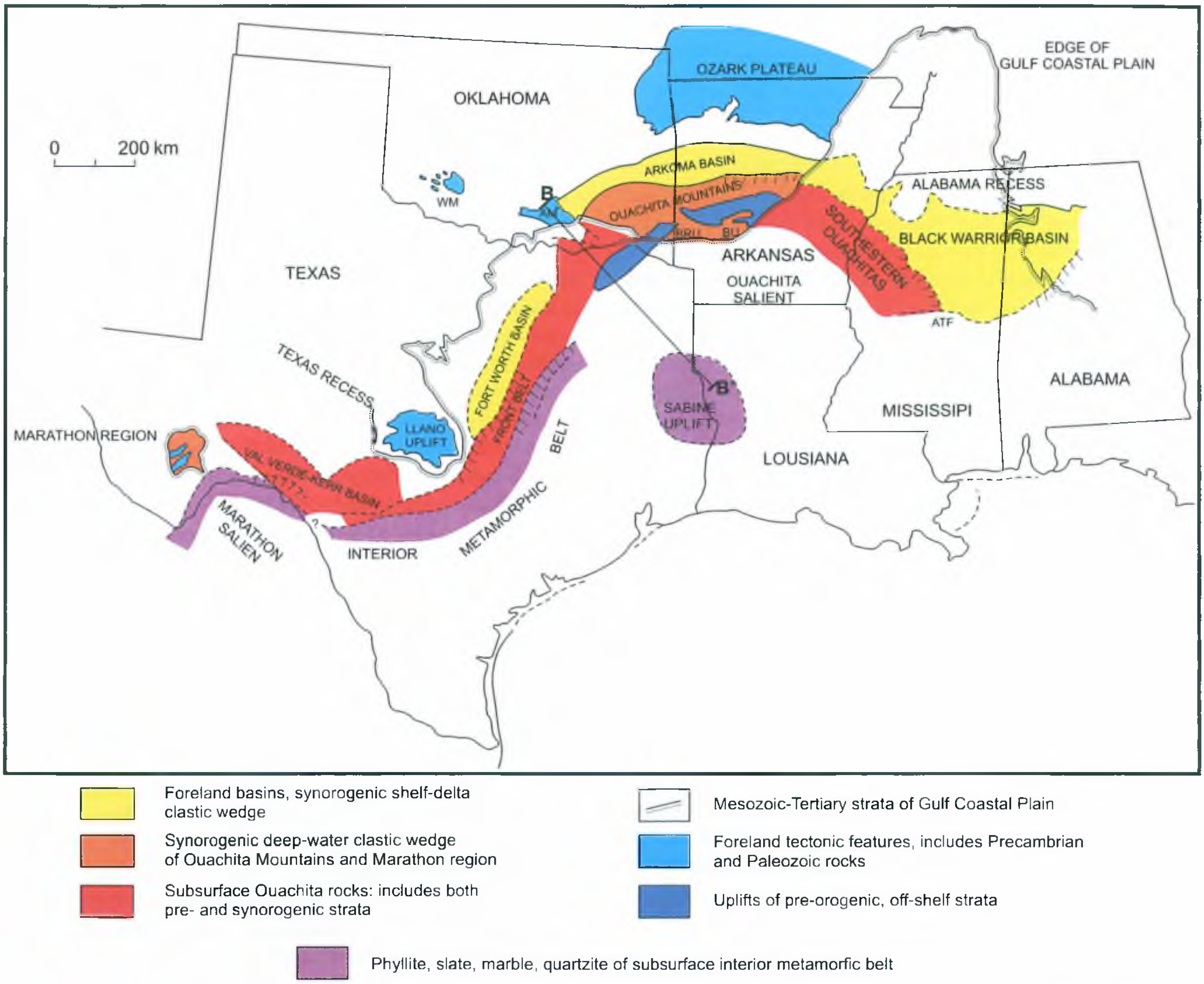

Fig. 4. Map of the Ouachita foldbelt (after Viele, 1989, Viele & Thomas, 1989, Arbenz, 1989, modified). B-B' – Location of cross-section (fig.5). Abbreviations: AM – Arbuckle mountains, ATF – Appalachians tectonic front, BRU – Broken Bow Uplift, BU – Benton Uplift, WM – Wichita Mountains

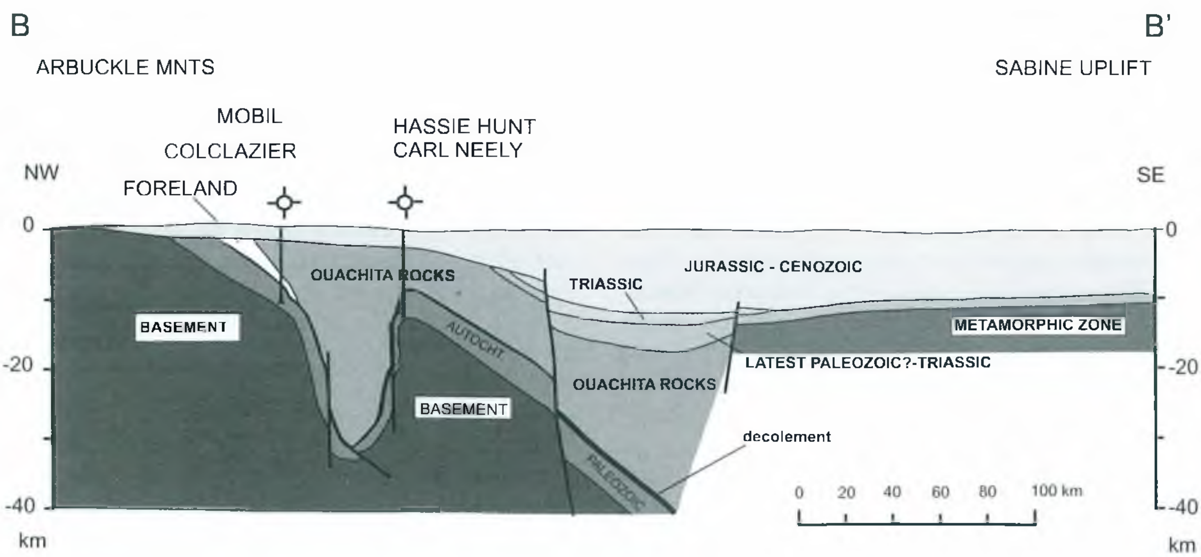

Fig. 5. Generalized cross section across Ouachita foldbelt between Arbuckle Mountains and Sabine Uplift, based partially on seismic reflection line GC36 (unpublished, courtesy of Mobil Exploration and Producing Technical Center, Dallas, Texas)

The negative areas are: the Black Warrior, Arkoma, Fort Worth and Val Verde basins (Viele & Thomas, 1989) (Fig. 4).

The Arbuckle and Wichita structures are foreland structures related to the Ouachita orogeny. Generally there are two uplifted zones – Arbuckle and Wichita Mountains with the exposed Precambrian and Lower Cambrian crystalline rocks, separated by the Ardmore and Anadarko basins (Ham, 1950, 1969). They are characterized by a system of faults trending east and southeast, as well as uplifts and grabens. Structural relief from uplift to basin floor is as much as 9 km (Lillie *et al.*, 1983; Golonka, 1988). The faults are high angle, often overthrust. These structures continue southward beneath the overthrust, allochthonous Ouachita orogenic belt. Based on seismic and well data there are two uplifted zones, separated by a depression zone (Golonka, 1988) (Fig. 4). The northern uplift is the continuation of the Arbuckle Mountains. It is bordered on the north by the Anadarko basin. The depression zone is related to the Fort Worth and Ardmore basins. The axis of the uplift trends northeast through Oklahoma and Arkansas beneath the Potato Hills.

GEODYNAMIC EVOLUTIONARY STAGES

Four main geodynamic stages can be distinguished during the evolution of the both, the Carpathians and Ouachitas regions. These stages are marked by the general changes in the development of basins and orogens (Table 1 and 2). The stages usually can be divided into more local substages.

STAGE I – SYNRIFT

This stage is characterized by rifting of terranes off the major continent and formation of oceanic type of basins (Table 2).

The West Carpathians (Triassic/Jurassic–Early Cretaceous) ± 135 My

The basement of most of the plates that play an important role in the Mesozoic–Cenozoic evolution of the circum-Carpathian area, was formed during the Late Paleozoic collisional events (Golonka *et al.*, 2000, a). Moesia, Eastern Alps (EA), Inner Carpathians (IC), Tisa (Ti – Fig. 7A) and adjacent terranes, were sutured to the Eurasian (Laurasian) arm of Pangea (Figs. 6, 7A), while Adria and adjacent terranes were situated near the Gondwana (African) arm. The equatorial position of the Bohemian Massif, adjacent to the Carpathian plates, agrees with the global Pangean model of Golonka (2000, 2002). The Paleotethys Ocean was located south of Eurasia and was established in southern and central Europe during Permian–Triassic time. The Meliata–Halstatt Ocean (Me – Fig. 7B) formed as a result of rifting of the Tisa terrane from Eurasia during Triassic time. The Vardar (Va)–Transylvanian (Tr) Ocean separated the Tisa (Bihor–Apuseni) block from the Moesian–Eastern European Platform (Sandulescu, 1988; Sandulescu & Visarion, 2000). There is a possibility of existence of the embayment of Vardar–Transylvanian oceanic zone between Inner Carpathian and European Platform (Golonka *et al.*, 2000, 2003a). Pelagic limestones of Triassic pebbles in the exotic sequences in the Pieniny Klippen Belt (Birkenmajer, 1988; Birkenmajer *et al.*, 1990) and Magura Unit (Soták, 1986) could have originated in this embayment. The embayment position and its relation to the other parts of Tethys, Vardar Ocean and Meliata–Halstatt Ocean(PD, Fig. 7B) remain quite speculative.

The Triassic shallow water limestones, marls and dolomites with numerous reefs (Kiessling *et al.*, 1999) prevailed in the Inner Carpathians platform areas and were underlayed by the Upper Permian and Lower Triassic clastic deposits. The sedimentary sequence rests on metamorphic and granitic rocks of the Late Paleozoic age.

The above mentioned period of sedimentation was followed by the Jurassic opening of the Ligurian–Penninic–Pieniny Klippen Belt/Magura Ocean (Figs. 7B, C). This opening resulted in the rifting of Alpine–Inner Carpathian terranes off Eurasia. Several basins separated by carbonate platforms developed in the Carpathian region. A local rift with the andesitic volcanics developed also along the southern margin of the North European platform (Ślączka, 1998). Stampfli (2001) recently postulates a single Penninic Ocean (Pe, Fig. 7B) separating Apulia and Eastern Alps blocks from Eurasia. We proposed a similar model for the Pieniny Klippen Belt Ocean in the Carpathians. The orientation of the Pieniny Klippen Belt Ocean was SW–NE (see discussion in Golonka & Krobicki, 2001). This ocean was divided into the northwestern (Magura basin, Mg – Fig. 7C) and southeastern (Pieniny Basin – Fig. 7B, C) basins by the mid-oceanic Czorsztyn Ridge (CR – Fig. 7C). The deepest part of the southeastern basin is documented by extremely deep water Jurassic–Early Cretaceous deposits (pelagic limestones and radiolarites) of Złatna unit (Sikora, 1971; Golonka & Sikora, 1981) later described also as Ultrapieninic unit or Vahicum (e.g., Plašienka, 1999). The transitional slope sequences between deepest basinal units and ridge units are known as Pieniny, Branisko (Kysuca), Niedzica and Czertezik successions (Fig. 8). The shallowest ridge sequences are known as the Czorsztyn Succession. Dark Lower Jurassic shales (Table 1) in this succession are followed by Middle Jurassic–earliest Cretaceous crinoidal and nodular limestones and Cretaceous variegated marls. Sedimentation of pelagic limestones, marls, cherts, cherty limestones and some syn-rift turbiditic deposits occurred during this period in the basinal areas, whereas on the uplifted parts more shallow, calcareous sediments developed. The deepest part of the northwestern basin is represented by extremely deep-water condensed Jurassic–Early Cretaceous deposits (pelagic cherty limestones and radiolarites) of the southern Magura (or Grajcarek or Hulina) unit (Golonka & Sikora, 1981; Birkenmajer, 1986; Golonka *et al.*, 2000, 2003a). The paleogeographic extent of the Magura Basin remains somewhat enigmatic and speculative. Also speculative is existence of oceanic crust below the whole Magura basin. The transitional slope sequence is known from some outcrops located north of the Czorsztyn Ridge (such as Zawiasy and Stare Bystre in Poland) (Golonka & Sikora, 1981). Ridge sequences as well as transitional slope sequences are also called Oravicum (e.g., Plašienka, 1999).

Table 1

Comparative stratigraphic chart

OUACHITAS

Time		Ouachita	Arbuckle
Stage IV (290-225 Ma)	Triassic	Clastic red beds of Wagle Mills Formation	Clastic red beds of Wagle Mills Formation
	Permian	Fluvial-deltaic and continental strata	Fluvial-deltaic and continental strata
Stage III (325-290 Ma)	Late Carboniferous (325-290)	Poorly sorted sandstones and dark gray shales, sometimes limestones (Atoka Fm). Olistostromes with enormous blocks of limestone, chert, and black shale; black shales and \ thin-bedded sandstones (Johns Valley). Flysch and olistostromes (Jackfork sandstone)	Poorly sorted sandstones and dark gray shales, sometimes limestones (Atoka Fm). Limestone, shale and sandstone (Wapanucka limestone)
Stage II (365-325 Ma)		Flysch with tuffs, shales and cherts (Stanley shale). Black shales, cherts (Woodford chert)	Black shales and sandstones (Springer Fm.) dark-colored Caney shale. Sycamore limestone. Black shales, cherts (Woodford chert)
Stage I (500-365 Ma)	Devonian	Black shales, cherty limestones, cherts (Arkansas Novaculite)	Black shales, cherts (Woodford chert)
	Silurian	Turbiditic sandstones, siltstones and shales (Missouri Mountain Shale, Blaylock Sandstone)	Limestones and shales (Hunton)
	Ordovician	Turbiditic sandstones and limestones, graptolitic shales, cherts, coarse-grained submarine slides (Blaylock, Polk Creek, Bigfork, Womble, Blakely, Mazarn, Crystal Mountain, Collier)	Limestones, dolomites, shales, cherts (Hunton, Sylvan, Viola, Simpson, Arbuckle)
	Cambrian	Collier graptolitic shake with lenses and beds of limestones of a distal turbiditic character	Arbuckle Group, Honey Creek limestones, Reagan transgressive sandstone

CARPATHIANS

Time		Magura-Pieniny	Outer subbasins
Stage IV (20-0 Ma)	Plio-Quat.	Fluvial and continental strata	Fluvial and continental strata
	M. Miocene	Continental shales and sandstones with marine intercalations	Brackish sandstones, shales and limestones. Marine, fine grained and clastic sediments; limestones, evaporites including salt, locally olistostromes
Stage III (55-20 Ma)	Early Miocene	Medium- and thin-bedded sandstones (Malcov Fm.)	Basal conglomerates passing upwards to brackish and marine sandstones, shales and marls, locally salt. Olistostromes. Sandstones, shales and gray marls (Krosno beds)
	Oligocene	Medium- and thin-bedded sandstones (Malcov Fm.) Flysch, marls, proximal turbidites (Magura Fm.)	Sandstones, shales and gray marls (Krosno beds), olistostromes. Dark brown bituminous shales and cherts (Menilite Shales, locally sandstone submarine fans (Mszanka, Cergowa, Kliwa sandstones), olistostromes
	Eocene	*Globigerina* marls Flysch, Variegated shales, proximal turbidites (Magura Fm.)	*Globigerina* marls Flysch, Variegated shales, proximal turbidites (Ciężkowice sandstone)
Stage II (100-55 Ma)	Paleocene	Flysch (Ropianka),variegated shales, marls	Thick bedded, coarse-grained turbidites and fluxoturbidites (Istebna, Ciężkowice), marls (Węglówka), variegated shales.
	L. Cretaceous	Flysch (Ropianka), calcareous turbidites marls, radiolarites, green and red shales	Thick bedded, coarse-grained turbidites and fluxoturbidites (Godula, Istebna), flysch (Ropianka), marls, (Węglówka, Frydek)
Stage I (235-100 Ma)	E. Cretaceous	Black shales, turbiditic sandstones, marls, pelagic cherty limestones, cherts (radiolarites), nodular limestones	Black shales, carbonate turbidites (Cieszyn, Sinaia), turbiditic sandstones, coarse-grained submarine fans, olistostromes (Bucegi-Soymul)
	Late Jurassic	Pelagic cherty limestones, cherts (radiolarites), nodular limestones	Carbonate turbidites, black shales and marls
	Middle Jurassic	Crinoidal limestones, cherts (radiolarites)	
	Early Jurassic	Dark shales, pelagic limestones, turbiditic sandstones and limestones	
	Triassic	Pelagic limestones	

Table 2

Tectonostratigraphic comparison

OUACHITAS				CARPATHIANS			
Time		Tectonic events and elements	Sedimentary style	Time		Tectonic events and elements	Sedimentary style
Paleozoic–Triasic	Perm.Triassic (290-225 Ma)	Thrusting and inversion, Final formation of Arkoma foreland basin. Formation of post-orogenic extensional basins	Marine and continental molasse. Volcanics. Continental red beds	Mesozoic–Cenozoic	M. Miocene-Future(20-? Ma)	Thrusting and inversion, Final formation of Carpathian foredeep. Formation of Pannonian extensional basin	Marine and continental molasse. Volcanics
	Late Carboniferous (325-290 Ma)	Convergence of South America and North America, Sabine - Eurasia collision, thrusting and inversion with the continuing formation of accretionary prisms. Incipient formation of Arkoma foreland basin	Flysch, olistostromes. Marine and continental molasse. Volcanics		Eocene–Early Miocene (55-20 Ma)	Convergence of Africa and Eurasia, Alcapa - Eurasia collision, thrusting and inversion with continuing formation of accretionary prisms. Incipient formation of Carpathian foredeep	Flysch, olistostromes, pelagic shales, carbonates and biogenic siliceous deposits. Marine and continental molasse. Volcanics. Carbonates on ridges
	Early Carboniferous (365-325 Ma)	Subduction along the active margin of the Ouachita ocean. Enigmatic Sabine terrane moving north, partial closing of oceanic basin. Development of main flysch basin on the platform (passive margin Ma) with the attenuated crust and beginning of formation of accretionary prisms	Flysch, olistostromes, pelagic shales, carbonates and biogenic siliceous deposits. Volcanics		Late Cretaceous–Paleocene (100- 55 Ma)	Subduction zones along the active margin of the Magura-Pieniny ocean, Inner Carpathian terrane moving north, partial closing of oceanic basin, nappes with northward polarity in the Inner Carpathians, development of main flysch basins on the platform (passive margin) with attenuated crust and beginning of formation of accretionary prisms	Flysch, olistostromes, pelagic pelagic shales, carbonates and biogenic siliceous deposits. Volcanics
	Cambrian–Devonian (500-365 Ma)	Rifting of Precordilleran terrane off North America. Formation of the Ouachita ocean	Passive margin, deep water turbidites, pelagic shales, carbonates and biogenic siliceous deposit, volcanics? Carbonates on surrounding platform		Triassic–Early Cretaceous (235-100 Ma)	Rifting of Eastern Alps, Inner Carpathians terranes off Eurasia. Formation of the Pieniny-Magura, Outer Carpathian oceans and basins	Passive margin, deep water turbidites, pelagic shales, carbonates and biogenic siliceous deposits, volcanics. Carbonates on surrounding platform

During the Late Jurassic (Fig. 7C) the southern part of the North European Platform, north of the Pieniny/Magura realm, started to be rifted and small basins (e.g., proto-Silesian Basin in the Western Carpathians, Sl – Fig. 7C), with black, mainly redeposited marls (?Kimmeridgian–Tithonian) were created (Pescatore & Ślączka, 1984). The Western Carpathian Silesian Basin probably extended in the Eastern Carpathian Sinaia or "black flysch" Basin (Sandulescu, 1984; Kräutner, 1996: Kräutner & Krstic, 2000). The rifting in the eastern Carpathian was accompanied by

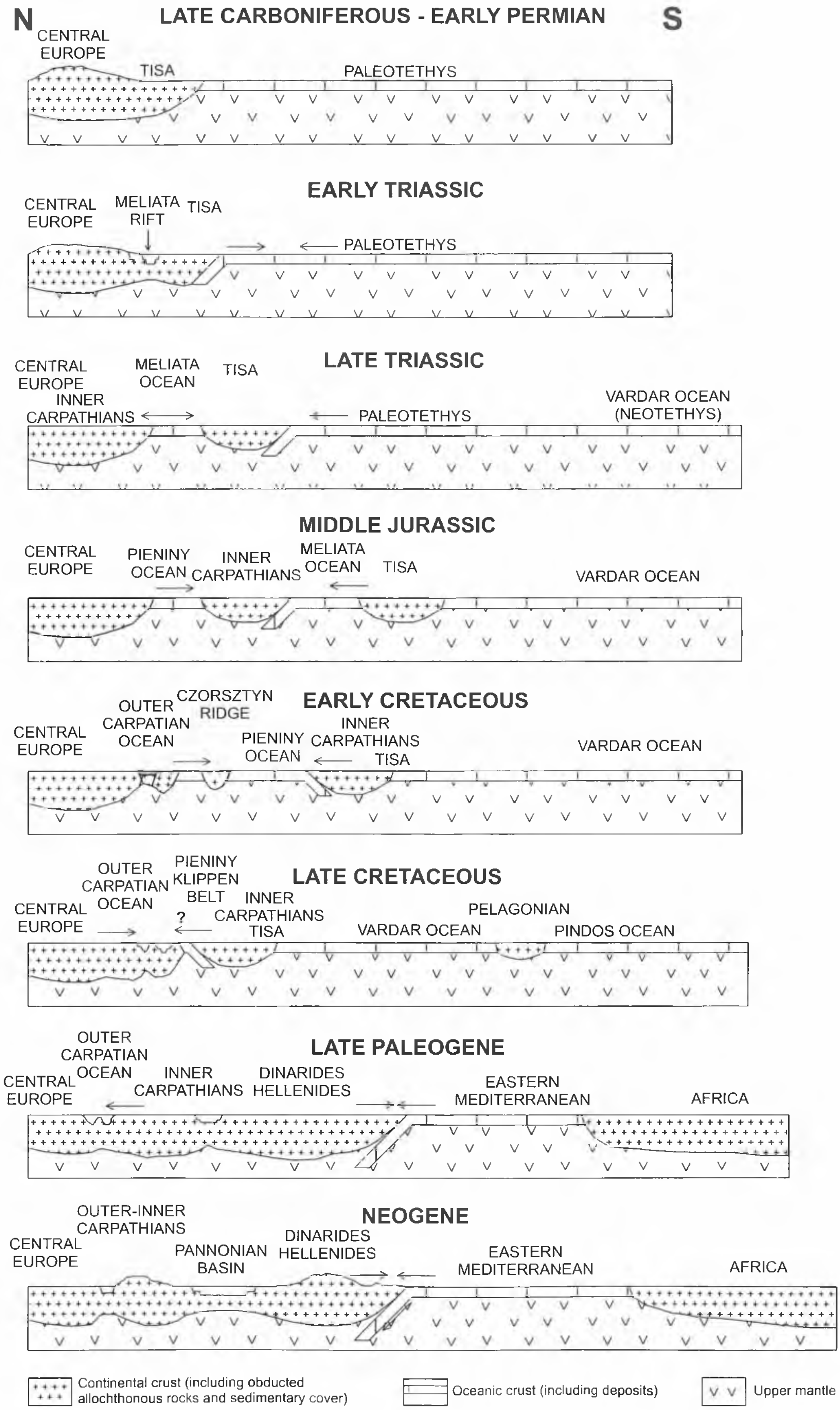

Fig. 6. Highly schematic (not to scale) plate tectonic profiles (Modified from Golonka *et al.*, 2000). Central Europe – Carpathians – Greece

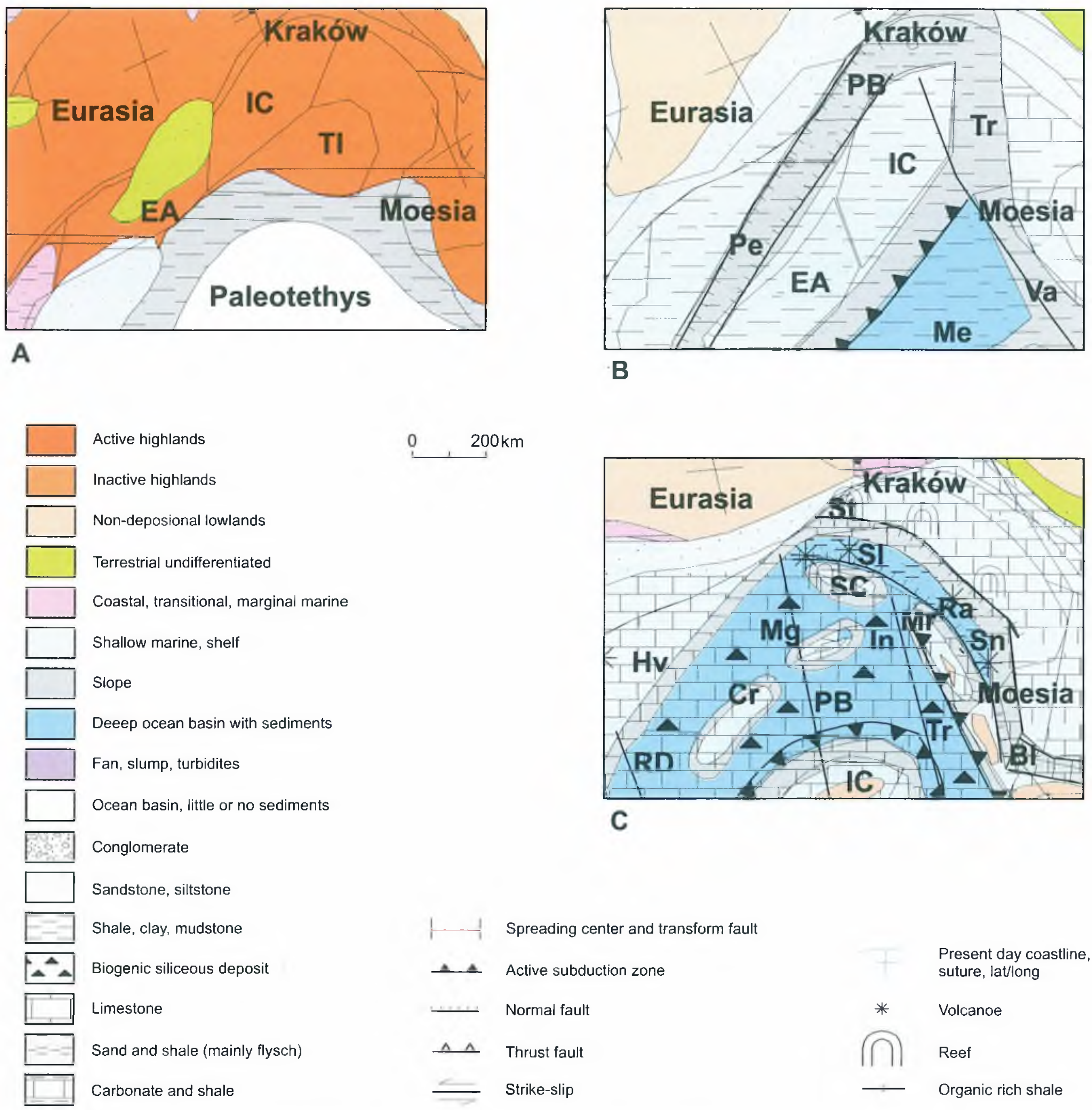

Fig. 7. Paleogeography and lithofacies of the circum-Carpathian area (Modified from Golonka *et al.*, 2000, 2003a). A – Late Carboniferous, B – Early Jurassic, C – Late Jurassic–Early Cretaceous. Abbreviations: Bl – Balkans, Cr – Czorsztyn ridge, Du – Dukla, EA – Eastern Alps, Hv – Helvetic, IC – Inner Carpathians, In – Inacovce-Kricevo basin, Mg – Magura basin, Mr – Marmarosh, Pe – Ligurian-Penninic Ocean, PB – Pieniny Basin, Ra – Rakhov basin, RD – Rheno-Danubian basin, SC – Silesian ridge (cordillera), Sl – Silesian basin, Sn – Sinaia basin, St – Štramberk olistolith, Va – Vardar Ocean, Ti – Tisa, Tr – Transilvanian Ocean

the extrusion of diabase-melaphyre volcanics (Lashkevitsch *et al.*, 1995; Golonka *et al.*, 2000). These basins indicate the beginning of the Outer Carpathians realm development. The opening is related to the closing of Pieniny-Magura (North and South Penninic) basin. In the Carpathian region subduction developed at the end of Jurassic –beginning of the Cretaceous (Figs. 6, 7C) along the southern margin of the narrowing basin north of the approaching Inner Carpathian and began to consume the Pieniny Klippen Belt Ocean (Birkenmajer, 1986).

The rapid supply of shallow-water calcareous material to the new-born basins could be an effect of the strong tectonoeustatic sea-level fluctuations known from that time. Black sediments mark the beginning of an euxinic cycle in the Outer Carpathian basin that lasted until Albian. The black marls pass gradually upwards into calcareous turbidites (Cieszyn limestones – Sinaia beds) which created several submarine fans (Table 1). Occurrence of deep-water microfauna indicates that subsidence of the basins must have been quite rapid (Poprawa *et al.*, 2002a, b). During the

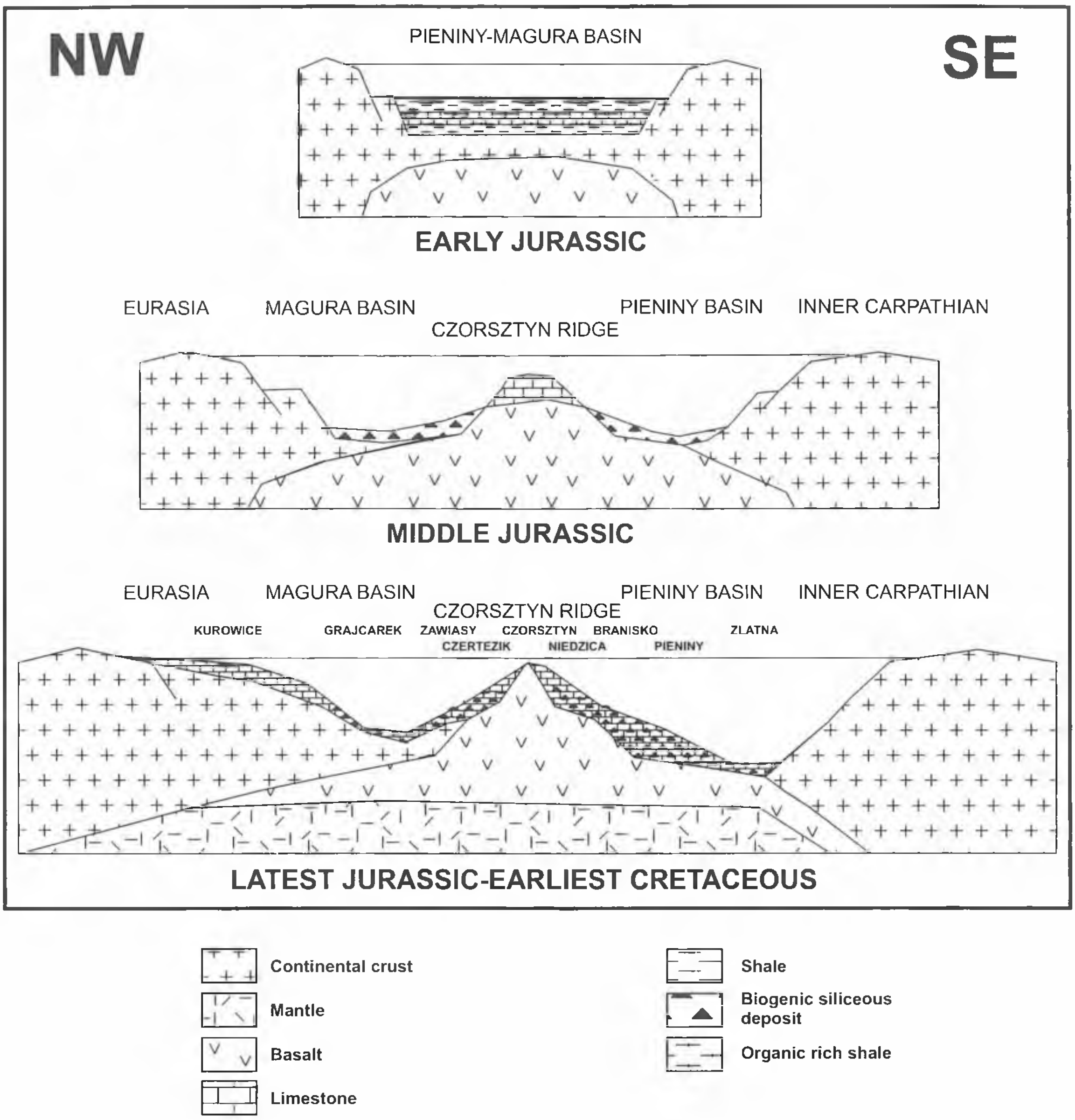

Fig. 8. Highly schematic (not to scale) profiles showing the evolution of the Pieniny Klippen Belt-Magura ocean during Jurassic–earliest Cretaceous time

early part of the Cretaceous the calcareous turbidites gave way to black calcareous shales and thin sandstones passing upwards into black, commonly siliceous shales. This type of sediments is already known also from the other Outer Carpathians basins. During the Hauterivian, Barremian and Aptian several coarse-grained submarine fans developed. The supply of clastic material was probably connected with the Early Cretaceous uplift known from the Bohemian Massif. The main phase of the hypabyssal intrusions and extrusions alkaline magma (mesocratic teschenites and leucocratic syenites) was connected with the late period of the North European Platform rifting. However it can not be excluded that magmatic eruptions could have started earlier (Smulikowski, 1980), as in the case of the Eastern Carpathians. Lack of changes in foraminifers assemblages through the Early Cretaceous time suggest lack of pronounced changes of depth of basins that corresponded generally to *Recurvoides* zone of Haig. It implies generally continuos tectonic subsidence of the basins during that period. This subsidence was equal to the rate of sedimentation.

In the early Albian within the black shales, widespread turbiditic sedimentation started, that can be connected with a compressional period very pronounced in the Eastern Carpathians. In that part of the Carpathian domain the compressional movement started during the Aptian and Albian and the inner part of the Carpathians was folded, nappes formed and in front of moving nappes coarse-grained sediments (Bucegi – Soymul Conglomerates) and olistostromes developed (Kruglov, 1989, 2001; Sandulescu, 1984, 1988).

The Ouachita foldbelt (Cambrian–Devonian) ± 130 Ma

There are two models explaining the formation of the Ouachita basin. In one model (e.g., Keller and Cebull, 1973) an ocean with oceanic crust was opened during early Paleo-

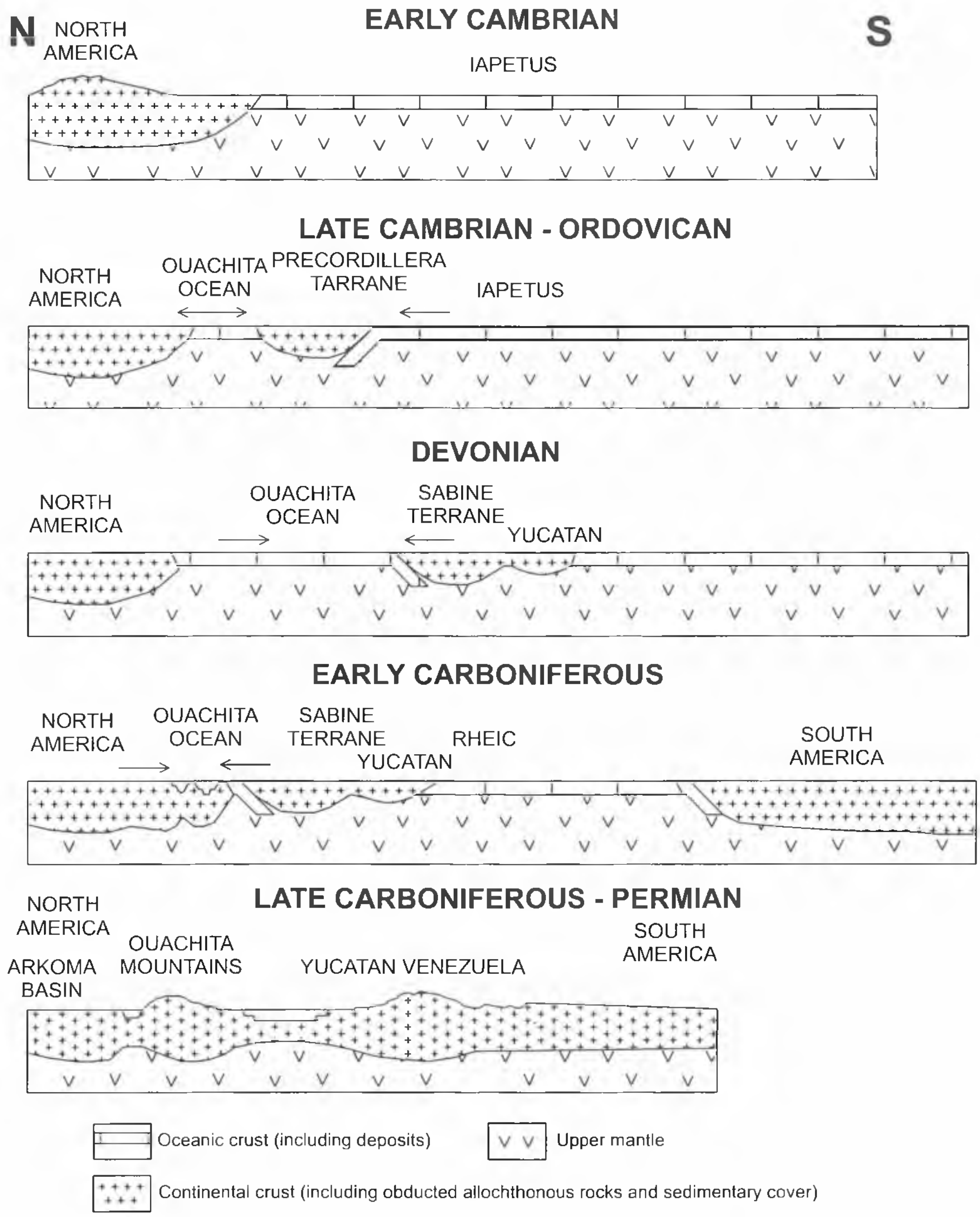

Fig. 9. Highly schematic (not to scale) plate tectonic profiles (modified from Golonka and Ślączka, 2000). North America-Ouachitas-Yucatan-South America

zoic time. According to Thomas & Astini (1999) the Argentine Precordillera was rifted from the Ouachita embayment of Laurentia during Cambrian time. This model is preferred by the authors (Figs. 9, 10A, B; note that North American continent is rotated, see Walker *et al.*, 1995, and for example Austin, Texas is north of Little Rock, Arkansas). It agrees well with the global Phanerozoic plate tectonic maps (Golonka, 2000, 2002). Also the lithostratigraphic sequences (Tables 1, 2) support the geodynamic evolution of the origin from rift through oceanic passive margin, accretionary prism related to basin closure to thrusting and inversion. In the other model (*e.g.*, Arbenz, 1989) the Ouachita basin was separated from the main (Iapetus-Rheic) ocean by poorly defined terranes. Present day Sabine uplift may constitute the remnant of these terranes. Viele & Thomas (1989) stated that Carboniferous volcanics formed a part of the upper (southern plate). The fragments of these southern terranes are perhaps included into the Inner Ouachita Foldbelt in Texas.

According to Thomas & Astini (1999) the distribution of synrift rocks and structures indicate diachronous rifting events along the margin of Laurentia during Cambrian time. The Ouachita rocks are younger than rocks known from the Blue Ridge area in Appalachians. The time of rifting and formation of the Alabama–Oklahoma transform fault (Thomas, 1991) could be connected with an emplacement of the igneous rocks in the Southern Oklahoma. These rocks, as young as 525±25 Ma, are overlain by Upper Cambrian sandstones. For the Late Cambrian beginning of sedimentation in the Ouachita basin (Table 1) we have somewhat arbitrarily assumed time around 500 Ma according to the Paleozoic time scale by Golonka & Kiessling, (2002).

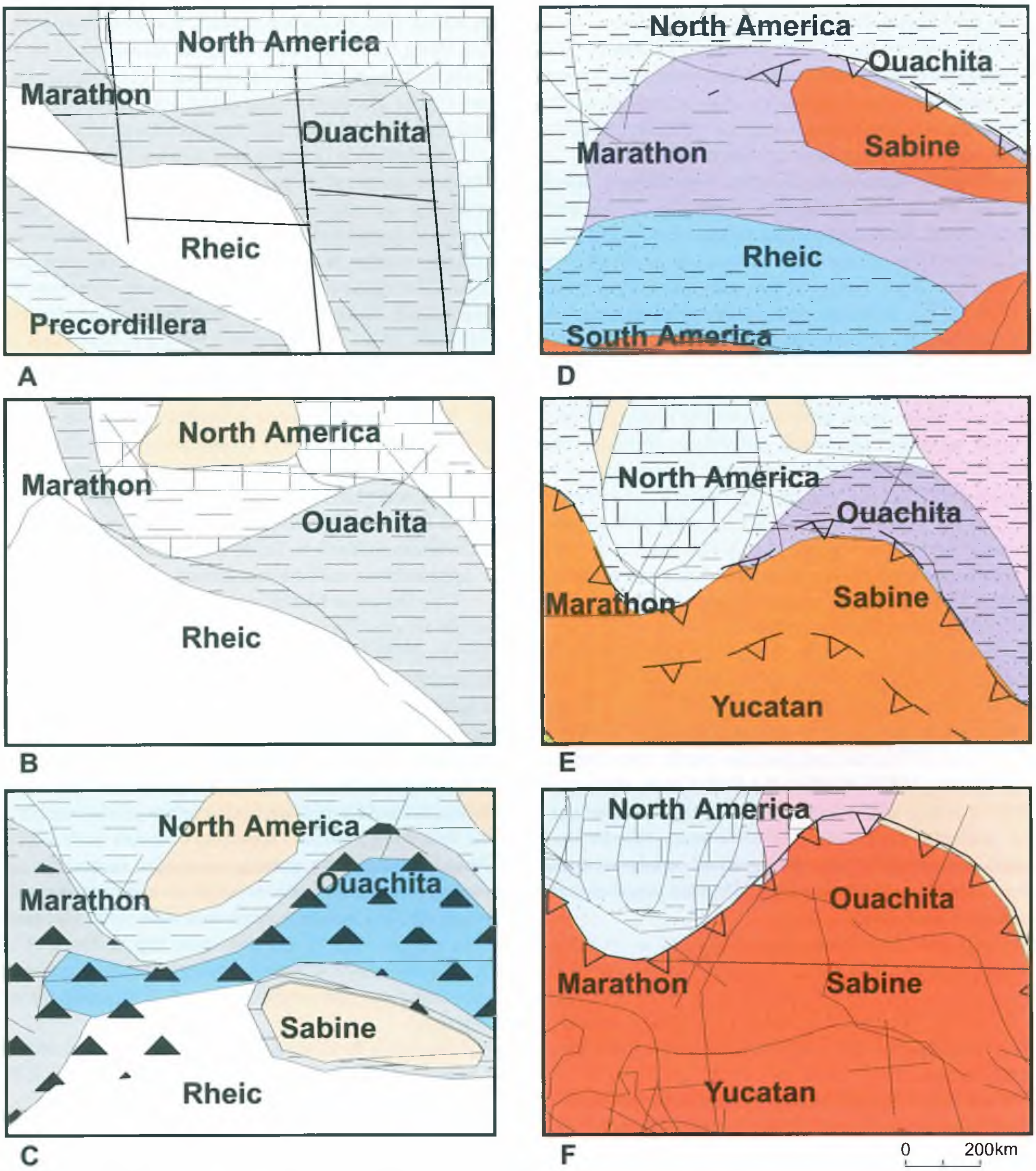

Fig. 10. Paleogeography and lithofacies of the circum-Ouachita area. (modified from Golonka, 2000, Golonka & Ford, 2000. Note: North American (Laurentia) continent is rotated (see Walker *et al.*, 1995). Spreading and transform faults from Thomas & Astini, 1999, Thomas *et al.*, 2002). A – Late Cambrian, B – Early Ordovician, C – Late Devonian, D – Early Carboniferous, E – Late Carboniferous, F – Early Permian. Legend as on Fig. 7

Two different facies were developed in the Early Paleozoic (Table 1): a deep off-shelf water passive margin facies in the Ouachita orogenic belt and a predominantly shallow water cratonic facies in the Ouachita foreland (Ham, 1950, 1959, 1969; Cline, 1960, 1970; Cline *et al.*, 1959; Flawn *et al.*, 1961; Lowe, 1985, 1989; McBride, 1975; Viele & Thomas, 1989).

The oldest rock of the deep water or so-called Ouachita facies is the Collier shale of Late Cambrian–Early Ordovician age (Arbenz, 1989; Lowe, 1989). The bulk of the formation is bluish-black, graptolitic shale with lenses and beds of limestones of a distal turbiditic character. The next unit is the Crystal Mountain sandstone, which overlies the Collier shale. Light gray, well-sorted, fine to medium grained, quartz cemented, thick-bedded quartz sandstones with minor shales prevail here. The sandstones are proximal turbidites. According to Lowe (1989) they indicate deposition from submarine sediment gravity flows including both high- and low-density turbidity currents. The Mazarn shale conformably overlies the Crystal Mountain sandstone. It is chiefly dark-gray, laminated, graptolitic shale with numerous thin interbeded distal turbidites, both quartzose as well as carbonate types. According to Lowe (1989) they were accumulated under low energy deep-water conditions.

The Blakely sandstone (Middle Ordovician) which overlies the Mazarn shale consists of massive to thin bedded, well-sorted, fine-grained quartz sandstones and varicolored silts and shales deposited by sandy high-density turbiditic currents with debris flows and submarine slides (Lowe, 1989). The Womble shale conformably overlies the Blakely sandstone. It consists principally of gray-black graptolitic slaty hemipelagic shale (Lowe, 1989; Gleason *et al.*, 2002). Numerous thin-bedded calcareous distal turbidites occur within the lower part, whereas the upper part contains some turbiditic siltstones, phosphatic breccias, and gray-black, well-sorted, pelletal turbiditic limestones.

The Bigfork Chert of Middle Ordovician age conformably overlies the Womble shale. This resistant unit consists of gray and black, thin-bedded, highly fractured chert with minor interbeded clay shales, siliceous shales and siliceous limestones. The limestones appear to be turbidites whereas the other rocks are pelagic in origin.

Conformably overlying the Bigfork Chert is the Polk Creek shale of Late Ordovician age (Lowe, 1989). This thin shale unit is black, highly fissile, graptolitic, and carbonaceous with a few thin beds of calcareous chert. The Blaylock sandstone of Silurian and possible partly Late Ordovician age (Lowe, 1989) overlies the Polk Creek shale conformably or with a stratigraphic break. The formation consists of subequal proportions of olive-gray, thin-bedded, laminated, feldspathic, very fine-grained sandstones and siltstones as well as greenish-gray shales that together comprise a shaly flysch facies.

The Missouri Mountain shale (Silurian) is in places transitional into the underlying Blaylock sandstone. Varicolored shales (or slates) comprise the bulk of the unit. Near the top are thin interbeded turbidites composed of white, fine to medium-grained well-rounded, silica-cemented quartz sandstones.

The next stratigraphic unit is the Arkansas novaculite, which includes rocks of both Devonian and lowermost Mississippian age (Fig. 10C). This distinctive unit consists chiefly of novaculite, a light-colored, extremely fine-grained, homogenous, highly fractured siliceous rock similar to cherts but characterized by a dominance of quartz rather than chalcedony. The upper and lower members contain laminated beds of rounded and angular quartz grains as well as intraformational breccias of black chert, phosphate and coarse quartz grains. Thin, graded, laminated siliceous beds and black shales characterize a middle member. According to Lowe (1989) breccias within the novaculite indicate erosion, while Viele & Thomas (1989) argue for deep-marine environment of deposition.

The cratonic or "Arbuckle" (Table 1) facies overlies Precambrian and Cambrian igneous rocks. The older sedimentary unit is the Timbered Hills Group of Late Cambrian age. It is composed of the Honey Creek Limestone and the Reagan Sandstone. The Reagan is a basal transgressive sandstone that incorporated much weathered material from underlying igneous rocks during its deposition. The Honey Creek consist chiefly of coarsely to very coarsely crystalline, highly glauconitic fragmental limestone. Thin intercalated mid-grained limestone and sandstone is frequent.

The Arbuckle Group which overlies the Timbered Hills Group is composed of more than 2000 m (in southern Oklahoma) of carbonates primarily limestones with approximately 15 percent dolomites. All Arbuckle carbonates are typically similar in lithology, composed primarily of limestones (fine-grained, stromatolitic, oolitic, or organodetritic limestones). The differentiation of the units into formations and their mapping is based largely on variations in the amount of quartz sand present and also on color and fossil content. Locally a distinction of dolomite units is also possible. While several formation names are used in the Ozark Uplift/Arkoma Basin area (Gasconade, Roubidoux, Jefferson City, Cotter and Powell formations) the Arbuckle Group is undivided in the wells. Similar shallow-water carbonate facies of Ordovician age are known in the foreland of Ouachita foldbelt in Texas as the Ellenburger Group and in Appalachians as the Knox Group.

The Simpson Group unconformable overlies the Arbuckle Group. The thickness of the Simpson ranges from 1000 to 2000 feet. The basal formation of the Simpson is limestone and dolomitic limestone. The lower part is cream to brown dolomite, very finely sucrose in appearance, the upper is limestone and usually fossiliferous. The Oil Creek Formation consists of gray-green and black shales, white to cream dense to fine crystalline limestone and about 100 feet thickness sand (important hydrocarbon producing horizon). The following McLish Formation contains more limestone. These limestones are usually ostracodal, frequently contain coarse brown to gray oolites set in a white to creamy matrix, frequently they become sandy and dolomitic. Occasionally some maroon, brown or olive shale can be found. The uppermost Bromide Formation consists mostly of green shales with relatively thin white to cream, frequently ostracodal limestones.

The Viola limestone (Upper Ordovician) conformably overlies the Simpson Group. The lower part is dark brown to almost black and very cherty. There are some beds of almost solid brown cherts. The upper part is white to pink, coarsely crystalline and contains few fossils. The Sylvan Shale represents the uppermost part of Ordovician; it is a light green and gray, flaky, splintery, soft shale containing graptolites.

The Hunton Formation, which overlies the Sylvan Shale, is of Ordovician–Silurian age. It consists of marls, dense marly thin-bedded limestones and shales. The Woodford chert (or shale) of Late Devonian and may be Early Mississippian age unconformably overlies the Hunton Formation. It consists of alternating beds of black sapropelic papery shale with phosphate nodules and black cherts.

Among the lower Paleozoic rocks of the Arbuckle facies, the Arbuckle Group and Simpson Group merit special consideration. Both of them are the potential petroleum producing horizons, and additionally some shales within the Simpson Group are potential source rocks. Deposition of graptolitic shales, pelagic, limestones, flysch deposits, marls, cherts, and cherty limestones occurred during this period.

STAGE II (POSTRIFT – EARLY COLLISIONAL)

This stage is characterized by formation of subduction zones along the active margin, partial closing of oceanic basin, development of main flysch basins on the platform (passive margin) with the attenuated crust and beginning of formation of accretionary prisms (Table 2).

The West Carpathians (Late Cretaceous–Paleocene) ± 55 My

In the beginning of the second stage, during the Cenomanian and Turonian, compression embraced the Inner Carpathians (IC) and several nappes with northward polarity developed. Subduction consumed the major part of the Pieniny Klippen Belt Ocean (Fig. 6). Cherty limestones gave way to marls and flysch deposits. With the development of the Inner Carpathian nappes the fore-arc basin was formed between the uplifted part of the IC terrane (so-called Andrusov Ridge) and the subduction zone. The flysch of the Klappe and Złatna (Fig.8) succession was formed in this area. Behind the ridge the Manin succession (Mn – Fig. 11A) was deposited within the back-arc basin. As an effect of these movements the Inner Carpathians and Alps jointed with the Adria plate and the Alcapa terrane was created. In the Cenomanian period, subsidence was faster than the sedimentation rate (Poprawa *et al.*, 2002a, b) and uniform, deep-water pelagic sedimentation of radiolarites, green and red shales embraced a greater part of the Outer Carpathians basins.

In the Outer Carpathians during this stage several ridges have been uplifted as an effect of the orogenic process. These ridges distinctly separated several subbasins, namely Magura, Porkulec-Convolute, Dukla-Fore-Magura, Silesian, Charnahora-Audia, Skole-Tarcau subbasins (Figs. 2, 11A). More outer subbasins (Skole, Silesian, Dukla-Fore Magura) reached diagonally the northern margin of the Outer Carpathians and successively terminated towards the west (Fig. 11A). From uplifted areas, situated within the Outer Carpathian realm as well as along its northern margin, enormous amount of clastic material was transported by various turbdity currents. This material filled the Outer Carpathian basins. (Dżułyński & Ślączka, 1956; Książkiewicz, 1968). Each basin had the specific type of clastic deposits, and sedimentation commenced in different time. It is interesting to note that this sedimentation started earlier in the outer subbasin (Skole-Tarcau subbasin) and migrated diachronously toward the inner subbasins (Bieda *et al.*, 1963; Ślączka & Kaminski, 1998). In the Skole-Tarcau basin (Sk, Tc – Fig. 11A) sedimentation started during the Turonian and ended in the Paleocene and deposits were represented by calcareous turbidites (Siliceous Marls) and thin- to thick-bedded turbidites (*Inoceramus*-Ropianka Beds). In western part of the area these turbidites were terminated during late Turonian/Coniacian by slump deposits. In the Silesian basin sedimentation started during the Late Turonian–Early Coniacian and lasted up to the Early Eocene being mainly represented by thick bedded, coarse-grained tur-

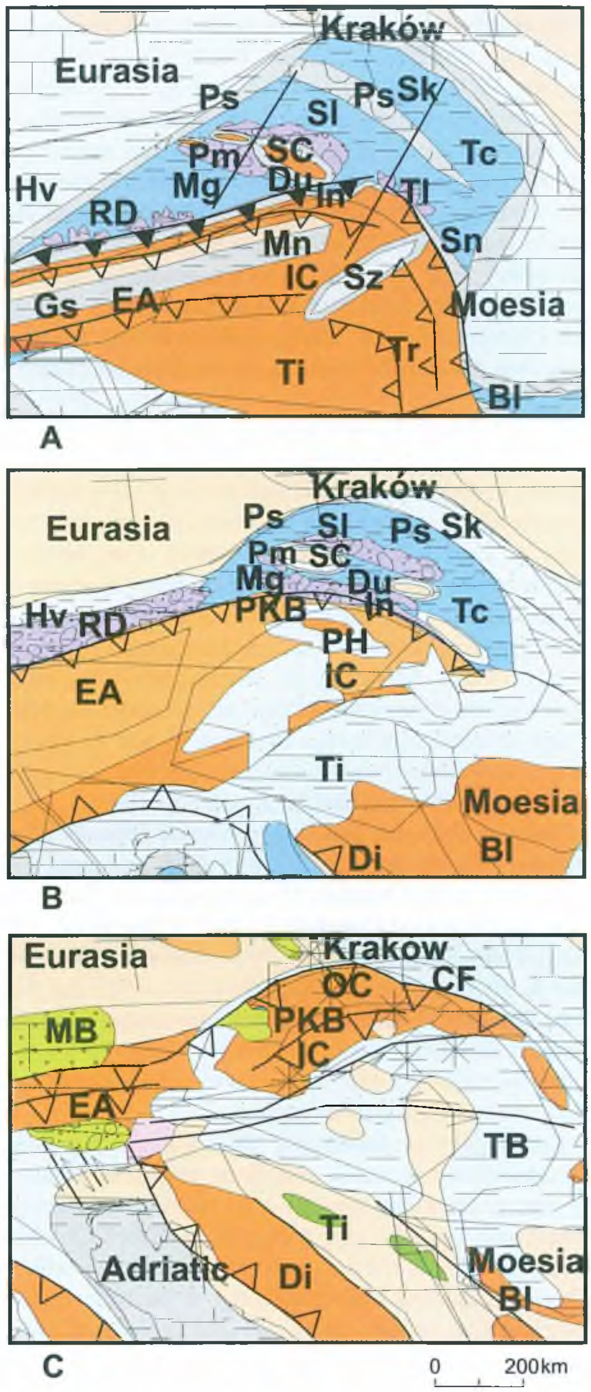

Fig. 11. Paleogeography and lithofacies of the circum-Carpathian area (Modified according to Golonka *et al.*, 2000). A – Late Cretaceous – earliest Paleogene, B – Oligocene, C – Miocene. Legend as on Fig. 7. Abbreviations: Bl – Balkans, Du – Dukla, EA – Eastern Alps, CF – Carpathian Foredeep, Gs – Gosau, Hv – Helvetic, IC – Inner Carpathians, In – Inacovce-Kricevo, MB – Molasse Basin, Mg – Magura basin, Mn – Manin, Mr – Marmarosh, PB – Pannonian Basin, PKB – Pieniny Klippen Belt, Pm – Fore-Magura, Ps – Subsilesian, Ra – Rakhov, RD – Rheno-Danubian basin, Rh – Rhodopes, SC – Silesian ridge (cordillera), Sl – Silesian basin, Sk – Skole, Sn – Sinaia basin, SZ – Szolnok, TB – Transilvanian Basin, Tc – Tarcau, Ti – Tisa, Tl – Teleajen, Tr – Transilvanian nappes

bidites and fluxoturbidites (Godula Beds, Istebna Beds and Ciężkowice Sandstone). In the Dukla (Du – Fig. 11A) (Ślączka, 1971) and Magura (Mg – Fig. 11A) subbasins sedimentation commenced during the Campanian and lasted till Paleocene (Oszczypko, 1992, 1998) and medium and thin-bedded, medium grained turbidites (*Inoceramus*-Ropianka Beds *s.l.*) prevailed there. On the underwater ridge that divided the Skole and Silesian subbasins (Subsilesian sedimentary area, Ps – Fig. 11A) a sequence of red, green marls (Węglówka Marls) of Senonian to Eocene and gray marls (Senonian Frydek Marls) were deposited. Small turbiditic fans developed locally in the small basins (Table 1).

Ouachita foldbelt (Early Carboniferous) ± 40–45 My

During Early Carboniferous (Figs 11, 10D) time the Ouachita basin became a narrowing trough with the flysch basins receiving vast amount of clastics (Cline *et al.*, 1959; Briggs & Roeder, 1975; Arbenz, 1989). The subduction probably took place along the southern margin of this narrowing basin north of the approaching Inner Ouachitas (or of the enigmatic Sabine terrane) and began to consume the Ouachita Ocean. According to Arbenz (1989), the earliest evidence of the constriction of the Ouachita depositional basin and of extension in the northern region is from the Lower Carboniferous (Meramecian). The northern margin supplied the clastic material for flysch deposits.

The Mississippian and Pennsylvanian deposits of predominately flysch character are very thick and most widespread rocks in the Ouachita Mountains region (Table 1) as well as in the buried Ouachita foldbelt (Cline *et al.*, 1959; Flawn *et al.*, 1961; McBride, 1975; Morris, 1974, 1989; Viele & Thomas, 1989). These deposits overly the widespread Woodford Chert (Shale) of the Devonian–Early Mississippian age described in the previous chapter. They are divided into the Stanley Shale, Jackfor Sandstone, Johns Valley Shale and Atoka Formations.

The Stanley is dominantly olive-gray to black shale with subordinate proximal and distal turbidites which are very fine-grained and generally feldspathic sandstones at the base and near the top. Along the southern Ouachitas, sandstones become thicker, and more numerous; tuffs and tuffaceous sandstones are prominent at the base; and disturbed bedding, impure cherts, and siliceous shales are rare. Along the Frontal Ouachitas, a thinner Stanley section has less turbidites at the base and only a minor tuff interval, but more intervals of impure chert and siliceous shale.

The Jackfork consists of gray-black shale and rhythmically interbeded whitish-gray, very fine-grained, quartz-rich turbidite packets. Olistostrome type beds dominate in the Frontal Ouachitas of Arkansas, proximal turbidites in the Southern Ouachitas, and thinner bedded distal turbidites and black siliceous shales in the Ouachitas of Oklahoma. The proximal turbidites are massive, ridge-forming quartz-rich, scour-and-fill units with few or no shale interbeds. Distal turbidites have partial to complete Bouma sequences, better developed sole markings, greater evidence of trace fossils as well as slope or shock-induced disruptive structures.

The Mississippian and Pennsylvanian rocks of the cratonic or "Arbuckle" facies are divided into the Caney, Springer, Wapanucka and Atoka units (Table 1). The Caney shale consists of dark-colored shale, hard and siliceous in places but soft and flaky elsewhere. The hard, blue Sycamore limestone can be present at the base of Caney.

The Springer Formation consists in the main of barren black shales with hard ferruginous and calcareous concretions and contains several sandstone members important for oil production. The Wapanucka limestone consists of limestone, shale and sandstone. The limestone beds are predominantly dark gray, commonly siliceous, cherty, locally fossiliferous and oolitic. The shale is gray, clayey, calcareous and weatherstained. The sandstones and siltstones contain abundant argillaceous material, calcite and glauconite.

STAGE III – LATE COLLISIONAL

This stage is characterized by convergence of two large continents, terrane–continent collision, with the continuing formation of accretionary prisms (Table 2).

West Carpathians (Eocene–Early Miocene) 35 My

In the circum-Carpathian region the Adria-Alcapa (Inner Carpathians) terranes continued their northward or NE movement during Eocene–Early Miocene time (Golonka *et al.*, 2000) (Figs. 6, 11 B, C) Their oblique collision with the North European plate led to the development of the accretionary prism of Outer Carpathians. During the compressional stage interbasinal ridges (bulges) were reactivated. Flysch still continued to be deposited in the subbasins. Numerous olistostromes were formed during this time (Ślączka & Oszczypko, 1987).

The process of migration of clastic facies from inner to the outer part began, connected with development and migration of accretionary prisms. In the inner part of the Outer Carpathians (Magura subbasin, Mg – Fig. 11A, B) the migration of clastic facies and development of accretionary prisms started already in Early Eocene and lasted till Oligocene (Oszczypko, 1998). In this subbasin, varied lithofacies developed. The lateral differences in lithofacies across the basin allowed dividing the Magura subbasin into several sedimentary domains. During the Early Eocene commenced the sedimentation of thick bedded, coarse grained turbidites (Magura Fm.) in form of several submarine fans. In more distal parts medium- and thin-bedded sandstones and shales developed, passing farther towards the north into variegated shales. These lithofacies migrated in time across the Magura subbasin towards the north. Sedimentation within the Magura basin terminated generally by accretionary wedges represented by medium- and thin bedded sandstones (Malcov Fm.) during the Oligocene (Table 1).

Within the more outer parts of the Carpathians realm, from Dukla (Du) to Skole (Sk – Fig. 11A, B) subbasins evidence of migration of depocenter appeared at the Eocene/Oligocene boundary (Ślączka, 1969) as an effect of compressional movements. As the consequence of these movements, the bottom of the basins started to deform and initial anticlines locally developed, slump, and coarse grained sediments were locally deposited and volcanic activity increased. Deep-marine connection with Tethys Sea was closed and euxinic conditions developed (Ślączka, 1969). In more external basins (Dukla, Silesian, Subsilesian, and Skole; Du, Si, Ps, Sk – Fig. 11A, B) green and gray shales with thin- and medium- bedded sandstones with intercalations of red shales prevailed during Early Eocene. The sedimentation of thick- bedded turbidites continued during the Early Eocene, in Silesian subbasin only. Small turbiditic fans developed only locally. This type of sedimentation gave way during the Late Eocene to green shales and yellowish *Globigerina* marls marking a period of unification of sedimentation.

The Oligocene sequences commenced with dark brown bituminous shales and cherts (Menilite Shales) with locally developed sandstone submarine fans or a system of fans up to several kilometers long. The main ones were Mszanka and Cergowa Sandstones in Dukla subbasin and Kliwa sandstones in Skole–Tarcau subbasin. The upper boundary of the bituminous shales is progressively younger towards the north and shales pass gradually upwards into a sequence of micaceous, calcareous sandstones and grey marls (Krosno beds), and they thin upward. The lower part of the sequence is generally represented by a complex of thick-bedded sandstones that pass upwards into series of medium- to thin-bedded sandstones and grey marls, that terminated sedimentation of the whole flysch sequence in the Outer Carpathians and can be considered as post flysch deposits. The boundaries between these lithofacies are diachronous across the basins, older in the south and younger in the north. Also cessation of deposition is diachronous across the Carpathians due to migration of tectonic activity and formation of trailling imbricate folds and/or accretionary prisms generally from the south to the north. During the sedimentation of the Krosno beds (Table 1) several slump deposits with blocks of shallow water, Paleogene limestones up to tens of meters long and smaller blocks of metamorphic and igneous rocks were deposited from intrabasinal cordilleras (Ślączka, 1969). With the final phase of tectonic movement, in front of advancing nappes and/or accretional wedges (prisms), huge (up to kilometers in size) slumps (olistostromes) with material derived from approaching nappes, developed (Ślączka & Oszczypko, 1987).

During the third stage Africa converged with Eurasia (Table 2). A direct collision of the supercontinents never happened, but their convergence lead to the collision of intervening terranes leading to the formation of the Alpine-Carpathian orogenic system. The Miocene tectonic movements caused final folding of the basins fill and created several imbricate nappes which generally reflect the original basin margin configurations (Figs. 6, 11C). During overthrusting the outer, marginal part of the advanced nappes was uplifted whereas in the inner part sedimentation persisted in the remnant piggy-back basin. From that, uplifted part of the nappes big olistolites glided down into the adjacent, more distal basins. That uplifted, marginal part of nappes probably had a character of accretionary prisms. The nappes became detached from the basement and were thrust northward in the west and eastward onto the North European platform with its Miocene cover. Overthrusting movements migrated along the Carpathians from the west towards the east. The Outer Carpathian allochthonous rocks, as a result of Miocene tectonic movements, have been overthrust onto the platform for a distance of 50 to more than 100 km. In response to overriding of the platform by the Carpathians, a peripheral foreland basin formed along the moving orogenic front (Oszczypko & Ślączka, 1985, 1989; Oszczypko, 1998); the basin extended along the Carpathian folded zone for at least tens of kilometers. Part of the autochtonous Miocene deposits was detached from basement and included into the accretionary prism. This Carpathian foreland basin (CF – Fig. 11C) is an equivalent of the Arkoma Basin in the Ouachita region.

In front of the advancing Carpathians nappes the inner part (in the eastern part also the marginal Subcarpathian–Borislav part) of the flysch basin, started to downwarp and tectonic depression formed during the Early Miocene. Thick molasse deposits filled up this depression. Generally they started with shallow water basal conglomerates passing upwards to brackish and marine sandstones, shales and marls, locally with olistostromes containing the material derived from the advancing Carpathians. In the eastern Carpathians there were periodically conditions favorable for precipitation of salt.

Ouachita foldbelt (Late Carboniferous) ± 35 My

In the circum-Ouachita region the Inner Ouachitas and Sabine terrane continued their northward movement during Carboniferous time (Wickham *et al.*, 1976; Thomas, 1976; Ross, 1979; Thomas & Viele, 1983; Viele & Thomas, 1989). According to Viele & Thomas (1989) the Sabine composite terrane is a collage of tectonic elements including not only Sabine uplift, but also Yucatan and Coahuila platforms. In this case the Yucatan platform would represent the stable crustal element of Pan-African origin. The Sabine uplift contains volcanics related perhaps to the south dipping subduction. The Sabine terrane collision with the North American plate led to the development of the accretionary wedge of Ouachita Mountains. During the compressional stage flysch still continued to be deposited. Olistostromes of the Maumelle Chaotic Zone formed during this time The main collisional activity occurred during Late Carboniferous–Early Permian in Ouachita Mountains (Figs. 9, 10E, F). The flysch deposits passed upwards into molasse.

The Late Carboniferous Johns Valley shale is characterized by a thin distinctive zone of olistostromes with enormous blocks of limestone, chert, and black shale in the Central Ouachitas at Oklahoma. Eastward into Arkansas, the deformational zones are replaced by gray-black shales and rhythmically interbeded very fine-grained, matrix-rich, thin-bedded sandstones that constitute a distal turbidite facies. Some slump zones also occur in the southern Ouachitas.

The lower part of the Atoka formation is probably the most nearly classic flysch sequence in the Ouachitas. It consists of a thick sequence of alternating interbeds of dark gray shale and gray-brown very fine-grained, laminated, graded subfeldspathic lithic wacke turbidites with common hieroglyphs. Upward, shallow-water sedimentary structures and coal seams indicate basinal filling from the north.

The Atoka group of the foreland is more molasse in character than flysch. The environment of deposition seems to have been a relatively shallow fluvial-deltaic to nearshore (swamp, marsh with minor marine influence). Some limestone beds occur especially in the lower member, but the bulk of formation consists of predominately poorly sorted sandstones and dark gray shales. The facies boundary between Atokan rocks of the Ouachita Mountains and those of the foreland of the Arkoma Basin and Arbuckle Mountains is sometimes hard to define. Probably due to basinal filling from the north uniform facies formed during upper Atokan time.

This major tectonic activity was caused by the collision of the Inner Ouachitas and Sabine terrane with North America with the accompanying convergence of Laurasia on the north and Gondwana (composed of Africa and South America, including the Yucatan promontory) on the south. To the north of the Ouachitas and partly beneath them was the Northern America platform with its autochthonous cover. As a result of Carboniferous–Permian tectonic movements, Ouachita allochthonous rocks have been overthrust onto the platform for a distance of 50 to more than 100 km. As a result of Ouachitas overriding the platform, the peripheral foreland Arkoma Basin formed along the moving orogenic front. The Ouachita foldbelt was a part of the central Pangean mountain range, which extended from West Texas to Poland (Keller & Hatcher, 1999; Golonka, 2000, 2002).

STAGE IV – POSTCOLLISIONAL

West Carpathians (Middle-Late Miocene) ± 20 My

During and after the main orogenic phase and the suture of the continents, an initial rifting or back-arc extention system was initiated in Early Miocene time behind the Carpathian arc in the Pannonian Basin. Extension in the Alpine-Carpathian system continued during the Miocene to Pliocene, forming horst and grabens within the orogen as well as in its foreland.

At the end of the Early Miocene the initial foreland basin became overthrust by the Carpathians and a new, more external foreland basin, developed. This basin and its depocenter migrated outwardly and eastward, contemporary with the advancing Carpathians nappes. As a result, the Neogene deposits show clear diachrony in the foreland area. In the west sedimentation terminated already during the Middle Miocene and lasted till Pliocene in the east. Marine, fine grained and clastic sedimentation of the Carpathian and foreland provenance prevailed, with a break during the Middle Miocene, when younger salinity crisis appeared. Locally, olistostromes were deposited (Ślączka & Oszczypko, 1987) with material derived from the Carpathians as well as from the inner margin of the molasse deposits. During the Middle Miocene part of the northern Carpathians collapsed also and sea invaded the already eroded Carpathians.

The thrusting of the Carpathian orogene on the platform was diachronous and migrated from the west towards east; where it terminated during the Pliocene. Part of the autochtonous Miocene deposits was detached from basement and formed narrow folded zone in front of the Carpathians. Finally the Outer Carpathian allochthonous rocks overthrust the southern margin of the Eurasian plate, for a distance of 50 to more than 100 km. At the end of the Miocene begun the vertical, neotectonic movements that last up today. Further development of these processes could be predicted in the future.

Ouachita foldbelt (Permian–Triassic) ± 65 My

During and after the main orogenic phase and the suturing of the continents an initial rifting system was initiated in Permian?–Triassic time in Ouachitas (Woods & Addington, 1973). The basin (equivalent of the Pannonian Basin) developed south the Ouachitas arc (Arbenz, 1989). Subsidence is revealed by Upper Carboniferous and Permian fluvial-deltaic and continental strata beneath southern Arkansas (Lillie *et al.*, 1983). The rift system was strongly developed during the Triassic and the rifts contain Triassic clastic red beds of Eagle Mills Formation. According to Vile & Thomas (1989) the rifting within the Sabine plate marked the onset of the next Wilson cycle, the Mesozoic opening of the Gulf of Mexico. Transgressive Jurassic, Cretaceous and Cenozoic deposits of the Gulf Coast Basin partly covered the Ouachita region. The Sabine terrane as well as possible latest Paleozoic–Triassic basinal deposits could be hidden below the thick (locally over 10 km) Jurassic evaporites and Upper Jurassic, Cretaceous and Cenozoic marine deposits (Woods & Addington, 1973; Lillie *et al.*, 1983).

DISCUSSION

The synrift stage of both orogens lasted approximately 130 to 135 million years. During this stage basin of partially or fully oceanic character developed. The passive margin type of sediments (table 1) were deposited within these basins. They were characterized by deep-water turbidites, pelagic shales and carbonates with biogenic siliceous deposits and volcanics. Carbonate sedimentation prevailed on surrounding platforms. The biogenic siliceous deposits are represented by radiolarites and Maiolica in the Pieniny Klippen Belt and by cherts and novaculites in the Ouachitas. These sediments are very similar in both realms, especially the resemblance between Maiolica and Novaculite is striking. In the Carpathian Mountains the synrift deposits are often strongly deformed and located in the suture zone (Pieniny Klippen Belt) between two colliding plates. In the Ouachita Mts this suture zone is hidden below the Mesozoic sediments. It is possible that the similar melange-type structure like Pieniny Klippe Belt exists also somewhere between the metamorphosed zone/Sabine Uplift are and the basement of the North American craton.

The Carpathians and Ouachitas are generally characterized by the stacks of nappes and thrustsheets built mainly of continual flysch sequences. These stacks are of allochtonous character and thrust over the cratonic foreland. The foreland is dipping gradually we have however the uplift of the basement in the inner part of the Ouachita allochton. This uplift is visible on the fig 5 and well documented by wells and seismic profiles (Golonka, 1988). An enigmatic basement uplift exists despite the general south dip of the European platform under NW Carpathians that may be caused by the geothermal uplift of the astenosphere, replacing delaminated lithosphere or by mantle plumes, or by the basement-involved thrust faults (Golonka *et al.*, 2003b).

The existence and character of this uplift is a subject of the planned deep drilling under the International Continental Scientific Drilling Program ICDP).

The postrift–early collisional stage lasted around 55 my in the West Carpathians and 40–45 my in the Ouachita foldbelt. The main bulk of the flysch sediments containing also olistostromes, pelagic shales, carbonates, biogenic siliceous deposits and volcanics was deposited in the deep water basins during the beginning of formation of accretionary prisms. Distal and proximal turbidites, which form the thousands of meters of flysch were formed in the similar conditions, between converging plates with the significant role of the subduction process. The clastic material was supplied from these converging plates as well as from the ridges within the flysch basin. Several of these ridges have been recognized within the Carpathian realm (Golonka *et al.*, 2000), the role of the hypotethical ridges within the Ouachita domain is still speculative.

The formation of the accretionary prisms continued during the third – late collisional stage of the development of two orogens. This stage lasted 35 million years. The plates collision caused the final folding of the basins fills and created imbricated nappes. From the uplifted parts of the nappes large olistolites glided down into the adjacent, more distal basins. The olistostromes are known from the many parts of the Carpathian realm, within the Ouachitas they are mainly known from the Maumelle Chaotic zone. The flysch deposits were formed in the piggy-back type remnant basins as well as in the incipient foreland basins in both realms. The flysch turned gradually into molasse. The allochtonous rocks have been overthrust onto the platforms for a distance of 50 to more than 100 kilometers.

The postcollisional stage lasted 65 my in the Ouachitas and only 20 My in the Carpathians. It means that the Wilson cycle has been completed in the Ouachita foldbelt and could still continue for tens of millions years in the circum-Carpathian realm. The future collision of Africa and Europe is possible following the known pattern of Gondwana-Laurussia collision during the Late Paleozoic. The Pannonian Basin and the inner part of the Carpathian are characterized by large volcanic intrusion accompanying the molasse type of deposits. Volcanics are known also from the Sabine uplift. The existence of large volcanic areas is also possible within the Late Paleozoic deposits under the Mesozoic cover south of the Ouachita Mountains in Louisiana and Texas.

CONCLUSIONS

The Ouachita orogenic belt displays features of the complete Paleozoic Wilson cycle, from rifting (Late Cambrian) through formation of the oceanic type of basin with passive continental margin (Ordovician), formation of an active margin with subduction zone (Devonian) to the collision of terranes with a continent (Carboniferous), continent-continent collision with the formation of the foldbelt and foreland basin (Late Carboniferous–Permian).

The West Carpathian orogenic belt shows the features of a Mesozoic–Cenozoic Wilson cycle from rifting (Triassic) through formation of the oceanic type of basin with continental passive margin (Jurassic), formation of an active margin with subduction zone (Cretaceous), and the collision of terranes with a continent (Cenozoic). The full continent-continent collision (Eurasia-Africa) remains to be completed in the future. While the Pieniny Klippen Belt Basin and Magura Basin were closing, the Silesian and subsequent subbasins were opening. These two processes were closely related. This kind of development is still unknown from the Ouachitas. There is a possibility, however, of some other similarities as well.

Both orogens are characterized by thin-skinned tectonics. Sedimentation is similar with characteristic deposits like cherty limestones-novaculites, flysch, olistostromes, and molasse in the later stages. Internides-externides relation exists in both orogens: in the Ouachita foldbelt the externides are covered with Mesozoic sediments and are known only from wells. The both orogens have similar foreland features – Carpathian foredeep and Arkoma basin in front of Ouachitas. The Ouachita basement configuration warrants further investigation of the subject.

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Streszczenie

ZACHODNIE KARPATY I GÓRY OUACHITA: STUDIA PORÓWNAWCZE NAD EWOLUCJĄ GEODYNAMICZNĄ

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Praca przedstawia porównanie ewolucji Karpat i Gór Ouachita w Ameryce Północnej z punktu widzenia tektoniki płyt i pozycji głównych elementów skorupy Ziemi w obu regionach na tle paleogeografii globalnej. Pokazuje ona związki pomiędzy procesami tektonicznymi a sedymentacją w basenach w różnych etapach cyklu orogenicznego.

Karpaty i Góry Ouachita są odległe geograficznie, jak również różnią się wiekiem uformowania orogenu, Karpaty utworzyły się w mezozoiku i kenozoiku podczas gdy Ouachita powstały w paleozoiku, jednak ich tektonostratygraficzna historia wykazuje uderzające podobieństwa.

Skonstruowano dwanaście map pokazujących konfigurację tektoniki płyt, paleogeografię i litofacje od późnego karbonu po neogen dla regionu wokółkarpackiego i od kambru po wczesny perm dla regionu Ouachita. To paleogeograficzne ujęcie pozwala pokazać jak pas górski został utworzony podczas cyklu orogenicznego który odzwierciedla skomplikowany proces geotektoniczny. Niektóre dotychczas niewyjaśnione pytania i problemy mogą być rozwiązane poprzez porównanie obu orogenów.

Karpaty tworzą łuk górski ciągnący się od Lasku Wiedeńskiego po Żelazne Wrota na Dunaju (Fig. 1). Karpaty Zachodnie składają się z pasma starszego znanego jako Karpaty Wewnętrzne lub Centralne i młodszego znanego jako Karpaty Zewnętrzne lub fliszowe (Mahel, 1974; Książkiewicz, 1977). Pomiędzy nimi znajduje się pieniński pas skałkowy (Fig. 2) zbudowany z sukcesji wieku jura–paleogen. Na zachodzie jest on nasunięty na Karpaty Zewnętrzne, w Polsce i wschodniej Słowacji jest oddzielony od nich uskokiem przesuwczym. Karpat Zewnętrzne są zbudowane z płaszczowin i nasunięć, złożonych głównie z sekwencji fliszowych wieku jura–dolny miocen nasuniętych na platformę północnoeuropejską. Płaszczowiny północnokarpackie zostały odkłute od podłoża w procesie orogenicznym.

Góry Ouachita (Fig. 4), odsłaniają się na powierzchni w Oklahomie i Arkansas oraz w zachodnim Teksasie, większa część orogenu jest ukryta pod pokrywą osadową (Arbenz, 1989; Viele & Thomas, 1989). Na wschodzie Ouachita łączą się z Appalachami pod pokrywą osadową w środkowym Mississippi (Thomas, 1977, 1989).

Pas fałdowy Ouachita składa się z fałdów i nasunięć z wergencją w kierunku kratonu północnoamerykańskiego. Starsze utwory są odsłonięte w antyklinoriach takich jak Broken Bow (BRU), Benton (BU) and Potato Hill (PH – Fig. 4). Góry Ouachita są allochtonem nasuniętym co najmniej 80 kilometrów na swoje podłoże kratoniczne w kierunku północnym. Podniesione strefy mogą być wyróżnione w podłożu kratonicznym (Fig. 5) (Lillie *et al.*, 1983; Golonka, 1988).

W obu orogenach można wyróżnić cztery główne etapy rozwoju odzwierciedlające cykle Wilsona. Etap pierwszy charakteryzuje się tworzeniem się ryftów, oderwaniem się teranów od głównych kontynentów i tworzeniem basenów typu oceanicznego. W Karpatach Zachodnich etap ten trwał od przełomu triasu–jury po wczesną kredę (Fig. 6, 7). W basenie Ouachita trwał on od kambr po dewon (Fig. 9, 10).

Etap drugi charakteryzuje się formowaniem stref subdukcji wzdłuż krawędzi aktywnych basenów, częściowym zamykaniem się basenów oceanicznych, rozwojem basenów fliszowych ze skorupą typu przejściowego, związanych z ryftingiem na platformie (krawędź pasywna). Etap ten trwał od późnej kredy po paleocen w regionie karpackim (Fig. 6, 11A), we wczesnym karbonie w regionie Ouachita (Fig. 9, 10A).

Etap trzeci jest etapem kolizyjnym. Prawdopodobnie miała tu miejsce kolizja teranów z kontynentem z towarzyszącą konwergencją dwóch wielkich płyt kontynentalnych (Afryka i Eurazja w przypadku Karpat, Południowa Ameryka i Północna Ameryka w przypadku Ouachity). Etap ten charakteryzuje się rozwojem pryzmy akrecyjnej. Trwał on od eocenu po wczesny miocen w Karpatach (Fig. 6, 11), w ciągu późnego karbonu w rejonie Ouachita (Fig. 9, 10E).

Etap czwarty jest etapem postkolizyjnym. Trwa on od miocenu po dzień dzisiejszy (i prawdopodobnie będzie się kontynuował w przyszłości) w Karpatach (Fig. 6, 11C), zaś od permu po trias w rejonie Ouachita (Fig. 9, 11F) W etapie tym formują się ostatecznie orogeny, tworzą baseny przedgórskie i zagórskie. W przypadku Gór Ouachita według Viele i Thomasa (1989) ryfting triasowy oznacza rozpoczęcie się nowego cyklu Wilsona związanego z otwarciem się zatoki meksykańskiej. Tego rodzaju proces w Karpatach może nastąpić w przyszłości.

Oba orogeny charakteryzują się tektoniką płaszczowinową typy thin-skin (naskórkowa). Sedymentacja w obu przypadkach jest podobna z charakterystycznymi osadami jak wapienie rogowcowe – nowakulity, flisz, olistostromy, czy molasy w późniejszym etapie. Sytuacja – eksternidy-internidy istnieje w obu orogenach, w przypadku Ouachity internidy są pokryte osadami mezozoicznymi i znane tylko z wierceń. Oba orogeny maję podobne przedgórza – zapadlisko przedkarpackie i basen Arkoma u przedpola Gór Ouachita.