POLSKA AKADEMIA NAUK • KOMITET NAUK GEOLOGICZNYCH

WYDAWNICTWO NAUKOWE PWN • WARSZAWA

acta geologica polonica

Vol. 44, No. 3-4 Warszawa 1994

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Jurassic tectonic events in south-eastern cratonic Poland

ABSTRACT: The Meta-Carpathian Arch, that comprised cratonic areas of SE Poland, usually experienced in the Mesozoic much less subsidence, or showed uplift, with regard to the Polish part of the Central European Basin situated farther north. In marked contrast, the areas of the Meta-Carpathian Arch were affected by strong subsidence in Middle and Late Jurassic time, and the peri-Carpathian segment of the Polish Rift began to develop in the Middle Jurassic as a southern extention of that rift, the existence of which north of the Holy Cross Lineament dates back to the Triassic or Permian. A stronger attenuation of continental crust can be recognized south of the Holy Cross Transfer Fault, which bounded the peri-Carpathian segment to the north; this is a feature consistent with the extention of the Polish Rift into the domain of the Carpathian Tethys. The Jurassic tectonic events controlled in the studied areas the formation of the three transgressive-regressive tectono-stratigraphic units: the COK Sequence (Callovian, Oxfordian and Lower Kimmeridgian, upper boundary near the top of the Hypselocyclum Zone); the LUK Sequence (Lower and Upper Kimmeridgian, upper boundary within the Eudoxus Zone); and the KVB Sequence (topmost Kimmeridgian, Volgian and Lower Berriasian). Coeval tectonic events, corresponding to those recorded in cratonic Poland, can be recognized in different parts of Europe, particularly in the Carpathians. The alternating phases of relative uplift and subsidence, experienced by the areas of the Meta-Carpathian Arch that flanked the Central European Basin on the south, can be attributed to fluctuations of intraplate stresses. The peculiar behavior of the Cracow region in Mesozoic and Miocene time, as an area never affected by strong subsidence, is thought to have resulted from the presence of Variscan granitoids at its depths. A development of the Polish Rift Basin in agreement with models assuming simple shear on lithospheric scale is suggested.

INTRODUCTION

In the larger context of Mesozoic tectonics, some peculiar tectonic events took place in the cratonic areas of south-eastern Poland: the Meta-Carpathian Arch was strongly downwarped, and the peri-Carpathian segment of the Polish

Rift Basin began to develop. These, and some other related topics, will be a subject of the present paper.

The interpretations put forward in this paper are based on analytical data found in numerous publications, *a.o.* in maps and sections which could not here be reproduced in detail. Accordingly, the purpose of some of the strongly generalized or smoothed maps and sections presented in this paper is to illustrate, rather than to document, these interpretations; and, in some cases (*e.g.* in Text-figs 2, 15 and 17), some spatial relationships have been more or less distorted to make the illustrations more readable. On the other hand, care has been taken to date geologic events with the greatest possible precision, with reference to ammonite zones, subzones and horizons. In this context it is pertinent to note that some biostratigraphic data found in earlier papers have been reinterpreted, or restated in terms of a different ammonite zonation.

Only some of the subdivisions of the Callovian, Oxfordian, Kimmeridgian and Volgian Stages, applied in this paper (Text-fig. 1), need short comment. A Zone of *Macrocephalites macrocephalus* or a Zone of *M. typicus* have usually been distinguished in the Lower Callovian of Poland beneath a Calloviense Zone of different range (Siemiątkowska-Giżejewska 1974, Giżejewska & Matyja 1978, Kopik 1979, Giżejewska 1981, Dayczak-Calikowska & Moryc 1988); this subdivision of the Lower Callovian differs from that recently proposed by Callomon & *al.* (1988). The zonal subdivision of the Oxfordian here applied is that currently in use in Submediterranean regions, but the boundary between the Middle and Upper Oxfordian Substages is taken at the base of the Bimammatum Zone. The subdivision of the Hypselocyclum Zone is that established by Atrops (1982), and the Uhlandi Subzone is distinguished within the Divisum Zone following Marques & Olóriz (1992). The interval between the Divisum and Eudoxus Zones will be assigned to the Acanthicum Zone (Ziegler 1962), and not to the Mutabilis Zone (as *in* Kutek 1968), because the stratigraphic range of the English Mutabilis Zone does not strictly correspond to this interval (Birkelund & *al.* 1983). The subdivisions of the Volgian Stage are those used by Kutek (1994) and Kutek & Zeiss (1994). The Scythicus Zone is the highest zone of the Volgian that can be recognized in cratonic Poland, no ammonite being present in the overlying Jurassic sediments that are developed in Purbeck-type facies. It is also worth of note that the Lower/Middle Volgian boundary roughly corresponds to the Middle/Upper Tithonian boundary (Kutek & Zeiss 1988, 1994; Kutek 1994), and that some high portions of the Volgian (chiefly the Upper Volgian Substage) are of Berriasian, and thus of Cretaceous, age (Hoedemaeker 1987, Kutek 1994).

Stages	Substages	Zones	Subzones	Horizons
VOLGIAN (pars)	Middle (pars)	Scythicus	Zarajskensis	Zarajskensis Regularis
			Scythicus	Scythicus Quenstedti
	Lower	Tenuicostata Pseudoscythica Sokolovi Klimovi		
KIMMERIDGIAN	Upper	Autissiodorensis Eudoxus Acanthicum		
	Lower	Divisum	Uhlandi	
		Hypselocyclum	Lothari	Perayensis Semistriatum Hypselocyclum Discoidale
			Hippolytense	
		Platynota		
OXFORDIAN	Upper	Planula Bimammatum		
	Middle	Bifurcatus Transversarium Plicatilis		
	Lower	Cordatum Mariae		
CALLOVIAN	Upper	Lamberti Athleta		
	Middle	Coronatum Jason		
	Lower	"Calloviense" "Macrocephalus"		

Fig. 1

Ammonite zonation of the Callovian, Oxfordian, Kimmeridgian, and Volgian Stages, as used in the present paper

MAJOR TECTONIC UNITS

A few major tectonic units and structures are pertinent to the problems discussed in this paper. These are the Mid-Polish Anticlinorium and the Polish Rift Basin, the Central European Basin, the Meta-Carpathian Arch, and the Holy Cross Lineament.

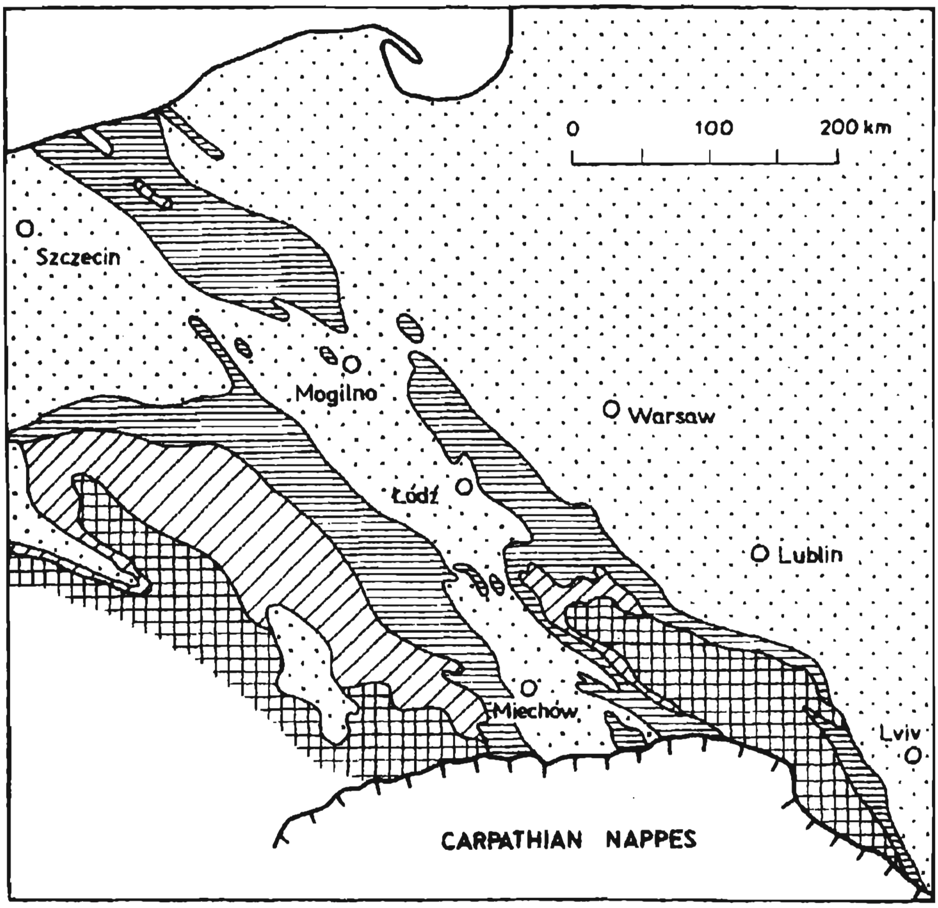


Fig. 2

Geological map of cratonic Poland and adjoining areas, without Cenozoic deposits (*chiefly after* Jaskowiak *& al.* 1968, *and* Pożaryski 1979; *strongly simplified*)

Pre-Permian deposits — cross-hachured, Permian and Triassic — oblique lines, Jurassic — horizontal lines, Cretaceous — stippled

THE MID-POLISH ANTICLINORIUM AND THE POLISH RIFT BASIN

The Mid-Polish Anticlinorium extends from the Baltic Sea south-east across Poland, plunging beneath the Carpathian nappes in south-eastern Poland and the western Ukraine (Text-figs 2–4). This anticlinorium is a result of the Laramide inversion of the Polish Rift in latest Cretaceous and Paleocene time (KUTEK & GŁAZEK 1972; POŻARYSKI & BROCHWICZ-LEWIŃSKI 1978, 1979; ZIEGLER 1990; DADLEZ 1993; DADLEZ & POKORSKI 1993).

Two synclinoria are developed along the Mid-Polish Anticlinorium, the Szczecin — Mogilno-Łódź — Miechów Synclinorium on its south-western, and the Pomerania — Warsaw — Lublin — Lviv Synclinorium on its north-eastern side (Text-fig. 2). These synclinoria are essentially post-depositional tectonic units, having been individualized as a result of the Laramide uplift of the

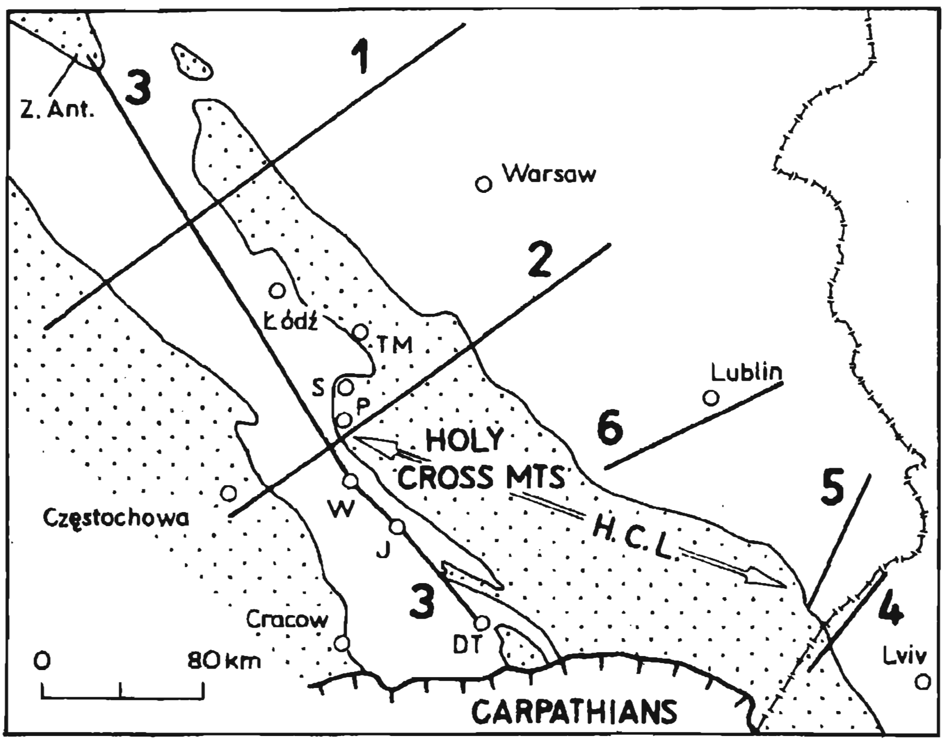


Fig. 3
Location map (stippled are pre-Permian rocks of the Mid-Polish Anticlinorium and of the Cracow-Silesian and Fore-Sudetic Monoclines)

1-6 — Section lines (*see* Text-figs 4-5, 10 and 14)

Z. Ant. — Zalesie Anticline, TM — Tomaszów Mazowiecki, S — Sulejów, P — Przedbórz, W — Włoszczowa, J — Jędrzejów, DT — Dąbrowa Tarnowska; H.C.L. — Holy Cross Lineament

Mid-Polish Anticlinorium. Mesozoic sediments up to the topmost Cretaceous (Maastrichtian) are preserved in both synclinoria. The belt of Jurassic, Triassic and Permian deposits, developed south-west of the Szczecin — Mogilno-Łódź — Miechów Synclinorium, belongs to the Fore-Sudetic and Cracow-Silesian Monoclines.

The existence of the north-western part of the Polish Rift situated north of the Holy Cross Lineament (Text-fig. 3) dates back to the Permian or Triassic (KUTEK & GŁAZEK 1972; POŻARYSKI & BROCHWICZ-LEWIŃSKI 1978, 1979; DADLEZ 1993; DADLEZ & POKORSKI 1993). South-east of this lineament it began to develop as a distinct rift structure in the Jurassic (KUTEK 1989). Subsidence was not uniform along the axis of the Polish Rift so that, for instance, Permian to Cretaceous sediments about 10 km thick accumulated in its Kuyavy segment (Text-fig. 5), whereas south of the Holy Cross Lineament their thickness amounted only to 2–3 km. This, in conjunction with a much more uniform uplift of the order of 2–3 km during the inversion of the Polish Rift (KUTEK & GŁAZEK 1972, DADLEZ & POKORSKI 1993) resulted in that Paleozoic and even latest Precambrian rocks came to the surface south of the Holy Cross Lineament, whereas farther north-west Triassic and Jurassic deposits are still preserved in the Mid-Polish Anticlinorium (Text-figs 2–4). In the northern Kuyavy region, where the preserved thicknesses of Upper Cretaceous sediments are well above 2 km in some places in the adjacent Łódź-Mogilno Synclinorium (JASKOWIAK-SCHOENEICHOWA & KRASSOWSKA 1988), even Lower Cretaceous deposits were not removed by erosion from the axial part of the Mid-Polish Anticlinorium (Text-figs 2–3 and 5).

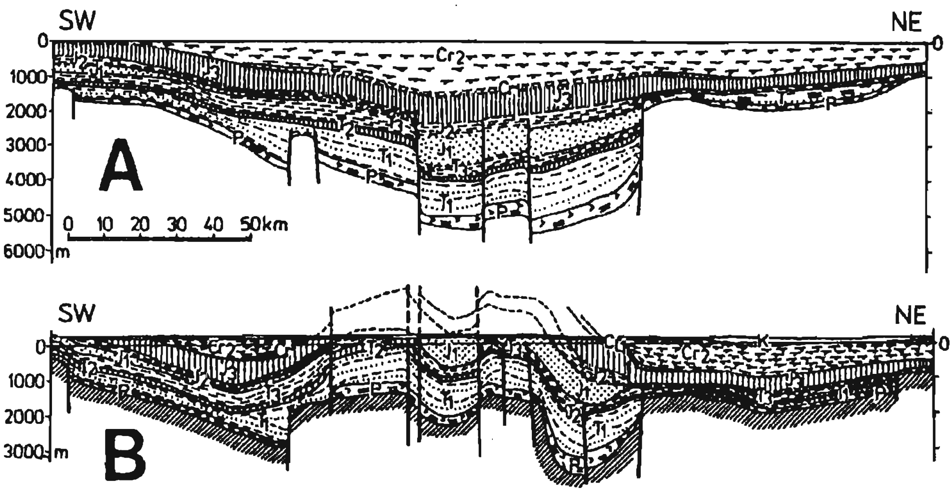


Fig. 4. Sections across the northern part of the Holy Cross Mts, approximately along line *2* in Text-figure 3 (*redrawn from* POŻARYSKI & BROCHWICZ-LEWIŃSKI 1969)

A — Paleotectonic cross-section through the Polish Rift Basin; B — Cross-section showing present-day tectonic structures

A general increase of thickness of Permian and Mesozoic sediments, in particular those of Cretaceous and Upper Jurassic age, can be recognized towards the axis of the Mid-Polish Anticlinorium (Text-figs 2–5). This allowed to distinguish a broad trough, which was called the Danish-Polish Furrow by POŻARYSKI (1957), and was also termed a peri-cratonic depression (POŻARYSKI 1977); the designation the Polish, or Mid-Polish Trough is now in current use (POŻARYSKI & BROCHWICZ-LEWIŃSKI 1978; DADLEZ 1989, 1993, 1994; ZIEGLER 1990; DADLEZ & POKORSKI 1973; DADLEZ & *al.* 1994). The broadly conceived Polish Trough comprised areas now occupied by the Mid-Polish Anticlinorium, but also by its bordering synclinoria. The Polish Trough was markedly affected by syndepositional faulting, this leading to the formation of grabens or halfgrabens in particular segments of the basin (Text-figs 4–5).

The axial part of the discussed basin, comprising well-developed graben structures and narrow, elongated depocenters, was sometimes separated out as an aulacogen (*e.g.* POŻARYSKI & KUTEK 1976, POŻARYSKI 1977); this term was usually used in a broad sense, referring to what can simply be called a continental rift. The zone of the present-day Mid-Polish Anticlinorium (Text-figs 2–5) coincided with grabens and depocenters of Permian or Mesozoic age (POŻARYSKI 1977; POŻARYSKI & BROCHWICZ-LEWIŃSKI 1978, 1970; DADLEZ 1989), and thus was situated within the limits of the "aulacogen". However, marked syndepositional faulting can also be recognized beyond this anticlinorium (POŻARYSKI 1977; POŻARYSKI & BROCHWICZ-LEWIŃSKI 1978, 1979; DADLEZ 1994; *comp.* Text-figs 2–5). This, and the asymmetric development of fault-structures as half-grabens, would make it a futile work to delineate distinctly a graben zone within the broader Polish Trough. It is also pertinent to note that both the designations (Mid-) Polish Trough and (Mid-) Polish

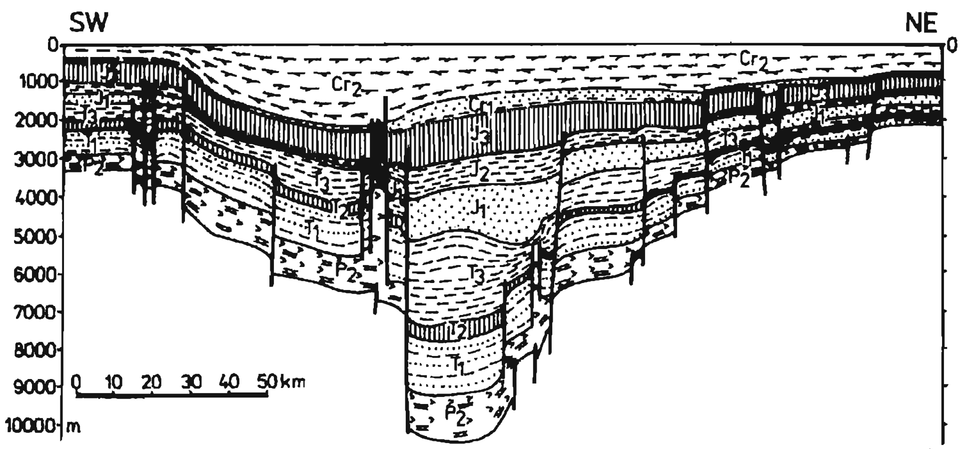


Fig. 5. Paleotectonic cross-section through the Polish Rift Basin, across the Kuyavy region, approximately along line *1* in Text-figure 3 (*redrawn from* POŻARYSKI & BROCHWICZ-LEWIŃSKI 1969)

Aulacogen were being used in a narrow and a broad sense, referring to a zone of grabens and depocenters, or to a broader elongated basin (*comp.* POŻARYSKI 1977; POŻARYSKI & KUTEK 1976; POŻARYSKI & BROCHWICZ-LEWIŃSKI 1978, 1979; POŻARYSKI & ŻYTKO 1980; DADLEZ 1989, 1993, 1994; DADLEZ & POKORSKI 1993; DADLEZ & *al.* 1994).

The Polish Trough (in a broad sense), which is clearly a rift-type tectonic unit (*see* Text-figs 4–5), will be called *the Polish Rift Basin* in this paper. The designation "Polish Rift" will also be applied, especially in tectonophysical context, or when referring to distinct grabens, or their faulted borders.

A graben stage ranging from the Permian to the Early Cretaceous, and a Late Cretaceous downwarp stage, were distinguished by POŻARYSKI & BROCHWICZ-LEWIŃSKI (1978, 1979) in the development of the Polish Rift Basin. On a minor scale, however, this basin provides evidence for an alternation of several graben and downwarp stages. For instance, confined, fault-bounded sediments of Berriasian and Early Valanginian age, indicative of a rifting stage, are overlain by expanded Upper Valanginian sediments, testifying to a downwarp stage (KUTEK & MARCINOWSKI 1985). In most cases the alternation of such stages cannot be recognized in small-scale cross-sections, such as those presented in Text-figures 4–5.

THE CENTRAL EUROPEAN BASIN

The Central European Basin, which is also named the Northwest European Basin (ZIEGLER 1990), extends from the North Sea through North Germany into Poland (Text-fig. 6), where it encompasses the Polish Rift Basin. In the Kuyavy region in Central Poland, a continuous, or nearly continuous, Permo-Mesozoic sedimentary succession about 10 km thick accumulated in one of the depocenters of the Central European Basin (Text-fig. 5; *see also* POŻARYSKI & BROCHWICZ-LEWIŃSKI 1978, 1979).

A peculiar feature of the Polish part of the Central European Basin was that it was situated nearest to the Tethyan Domain. As a consequence, marine connections and routes of faunal migrations were established several times in the Mesozoic across Poland between the Central European Basin and basins of the Carpathian Tethys (*e.g.* in Roet and Muschelkalk times, in the Middle and Late Jurassic, in the Berriasian, Valanginian, Hauterivian and Albian, and in the Late Cretaceous). Those seaway connections, if sufficiently broad, resulted in extensive carbonate sedimentation in cratonic Poland (*e.g.* in Roet and Muschelkalk times, in the Late Jurassic and in the Late Cretaceous). A spectacular effect of the peri-Tethyan position of the Polish part of the Central European Basin is the extensive development of the Late Jurassic sponge lithofacies (KUTEK & *al.* 1984), a feature not found farther west in that basin but characteristically displayed, for instance, in the peri-Tethyan South-German Basin.

As a consequence of the position of the Polish part of the Central European Basin near the Tethys and south of the elevated Baltic Shield, the Mesozoic faunal assemblages of cratonic Poland usually reveal stronger Tethyan, and weaker Boreal, influences than, for instance, those of Britain, an area whose position favored connections with Arctic basins.

THE META-CARPATHIAN ARCH

The Meta-Carpathian Arch (*see* Text-fig. 6) separated structurally, and at some times also paleogeographically, the Central European Basin from basins of the Carpathian Domain in the Permian, Mesozoic and Tertiary. The designation itself was first used by NOWAK (1972) with regard to a belt of uplifts north of the Miocene foredeep of the Carpathians. It will be applied here in a broader sense with reference to tectonic patterns observable from the Permian to the Tertiary.

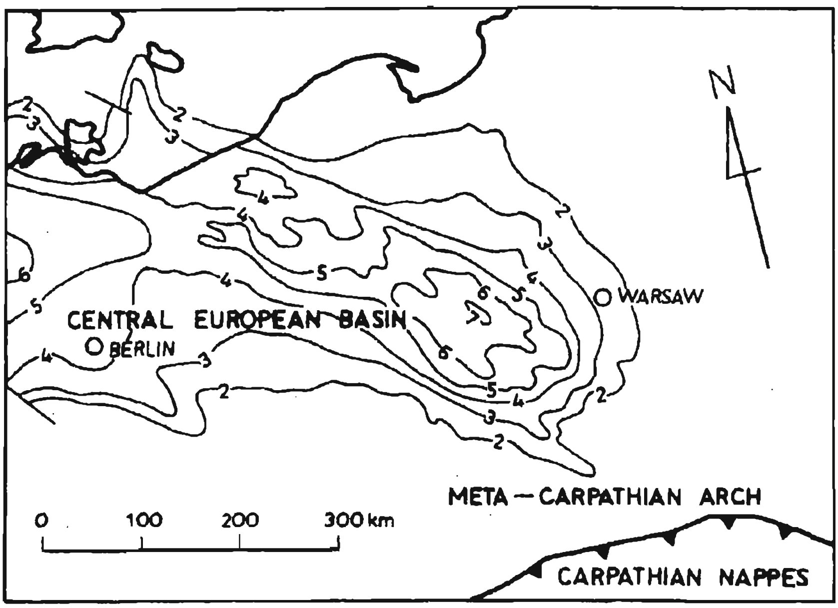


Fig. 6. Map showing the Meta-Carpathian Arch and the eastern part of the Central European Basin, outlined by isopachs (in 1000 m) of Permian to Cenozoic deposits (*chiefly based on* ZIEGLER 1990, Encl. 43)

Isopachs account for thicknesses of Cenozoic sediments in the Central European Basin, but thicknesses of Miocene deposits in the Carpathian Foredeep (amounting to well over 2000 m) are omitted

During most of Mesozoic time, south-eastern cratonic Poland (the area of the present-day foreland of the Polish Carpathians) was part of the northern slope of the Meta-Carpathian Arch. As a consequence, Permo-Mesozoic sedimentary successions are here markedly thinner, and reveal a higher degree of stratigraphic discontinuity, than the coeval successions deposited farther north in Poland, within the Central European Basin (Text-fig. 6). This can be clearly seen in maps published in several papers and atlases (*e.g.* ZNOSKO 1968, KUTEK & GŁAZEK 1972, CZERMIŃSKI & PAJCHLOWA 1975, SOKOŁOWSKI & TOMASZEWSKI 1987; and papers published in *Kwartalnik Geologiczny*, Vol. 32, No. 1, 1988).

Stratigraphic data relevant to the Meta-Carpathian Arch are chiefly provided by the sediments preserved in the synclinoria bordering the Mid-Polish Anticlinorium (Text-figs 2—3), especially by those of the Miechów and Łódź-Mogilno Synclinoria (Text-fig. 7). The restored cross-sections drawn along the latter synclinoria (Text-fig. 7), at some distance from the south-western border of Mid-Polish Anticlinorium (*see* Text-fig. 3), are strongly generalized and

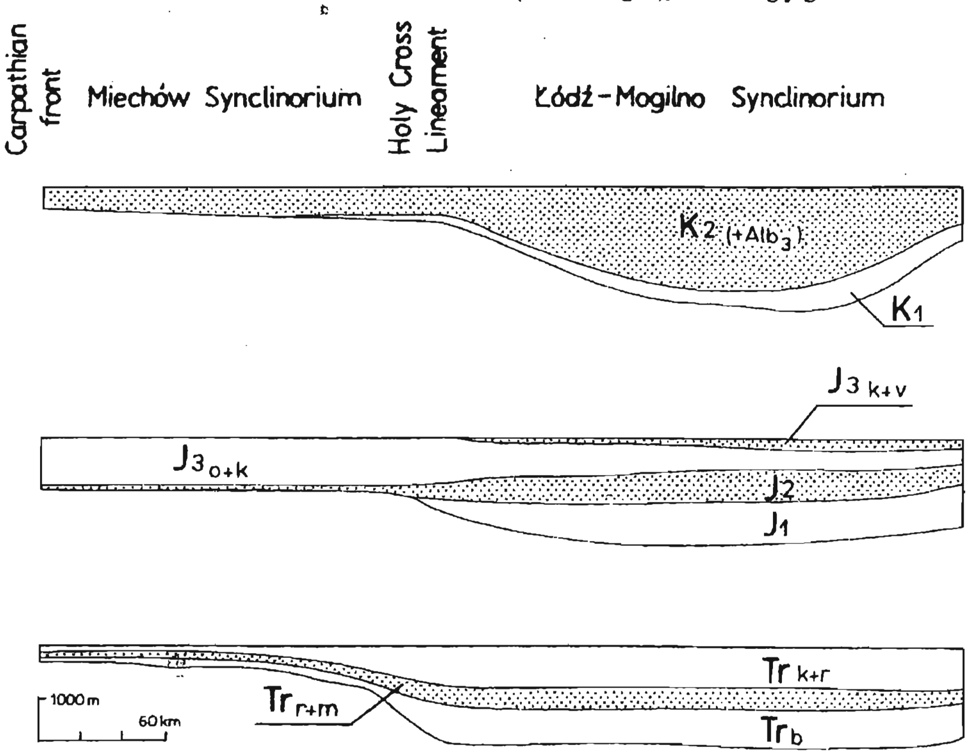


Fig. 7. Generalized cross-sections through the Triassic, Jurassic and Cretaceous deposits of the Miechów and Łódź-Mogilno Synclinoria, approximately along line *3* in Text-figure 3

ABBREVIATIONS: **Tr** — Triassic, **J** — Jurassic, **K** — Cretaceous; **b** — Buntsandstein, **r+m** — Roet and Muschelkalk, **k+r** — Keuper and Rhaetian, **o** — Oxfordian, **k** — Kimmeridgian, **v** — Volgian, **Alb$_3$** — Upper Albian smoothed. These sections clearly reveal the general decrease of thicknesses of Mesozoic deposits from the Central European Basin towards the Meta-Carpathian Arch.

In Triassic and Early Jurassic time, the axis of the Meta-Carpathian Arch was situated south of the present-day northern front of the Carpathians. This is especially well indicated by the Triassic/Jurassic unconformity in the middle part of the foreland of the Polish Carpathians, in the Miechów Synclinorium, where Jurassic sediments overstep southwards Upper, Middle and, near the Carpathians, Lower Triassic deposits (MORYC 1971). No Liassic sediments are preserved near the Polish Carpathians but farther north, in the Holy Cross Mts and the Cracow-Silesian Monocline, there occur Lower Liassic gravels, with maximum sizes of pebbles diminishing northwards (DADLEZ 1962). This, together with the estimated distance from the source area of the pebble material of about 300 km (UNRUG & CALIKOWSKI 1960), and some other regional data, suggests uplift in an area subsequently included in the domain of the Outer (Flysch) Carpathians.

In Central Poland, within the limits of the Central European Basin, Middle Jurassic (Aalenian to Callovian) deposits attain thicknesses amounting to about 1 km (DAYCZAK-CALIKOWSKA & MORYC 1988). The thickness of Middle Jurassic successions decreases dramatically towards the Carpathians (Text-fig. 7), and Aalenian and Bajocian deposits fall out of the successions (DAYCZAK-CALIKOWSKA & MORYC 1988). Near the Carpathians, Bathonian or Callovian, and locally even Lower Oxfordian or early Middle Oxfordian sediments rest directly on a pre-Jurassic substratum (DŻUŁYŃSKI 1950, GIŻEJEWSKA & WIECZOREK 1976, ZAPAŚNIK 1977, KUTEK *& al.* 1982, DAYCZAK-CALIKOWSKA & MORYC 1988). On the other hand, as revealed by boreholes, the thickness of the Middle Jurassic deposits that form part of the autochthonous sedimentary cover of the cratonic substratum of the Carpathian nappes increases south of the northern front of these nappes (KOSZARSKI 1985, DAYCZAK-CALIKOWSKA & MORYC 1988). From this it follows that towards the end of the Middle Jurassic the axis of the Meta-Carpathian Arch was situated near the zone now occupied by the Carpathian front, this heralding the unusual development of the area of the Carpathian foreland in the Late Jurassic.

In the western Ukraine, in the area now situated east of the Mid-Polish Anticlinorium, the axis of the Meta-Carpathian Arch was located north of the present-day front of the Carpathians. This is indicated by the presence in the Ukrainian foreland of the Carpathians of a thick sedimentary succession commencing with the Toarcian and including the whole Middle Jurassic (SANDLER 1969, MONKEVICH *& al.* 1985; *see* Text-figs 14—15 in this paper).

The area of the Meta-Carpathian Arch was strongly downwarped in the Late Jurassic, and ceased to develop as an arch. This problem will be discussed in detail in the following sections of this paper.

The Meta-Carpathian Arch re-appeared as an uplifted zone in the Early Cretaceous, with an axis situated that time in an unusually northern position in the region of the Holy Cross Mts. The Early Cretaceous Meta-Carpathian Arch, called in this case the Vistula Swell, was precisely outlined by CIEŚLIŃSKI (1976). In the Holy Cross Mts, on both sides of the Polish Anticlinorium, Albian (or, locally Cenomanian) sediments rest directly on Kimmeridgian deposits. The Neocomian deposits appear beneath the Albian ones to the north, and continuous Lower Cretaceous successions are developed in some parts of the Polish Lowland. To the south, in turn, Valanginian and Hauterivian deposits appear beneath Middle Cretaceous sediments in the southern part of the Lublin Synclinorium, and in the Lviv Synclinorium; on the western side of the Mid-Polish Anticlinorium, Valanginian deposits were encountered near the Carpathian front (CIEŚLIŃSKI 1976, GEROCH *& al.* 1972, KUTEK & GŁAZEK 1972, MONKEVICH *& al.* 1985, KUTEK *& al.* 1989). This distribution pattern of the pre-Albian Early Cretaceous deposits is largely due to the uplift of the Meta-Carpathian Arch in the Barremian and Aptian (*comp.* RACZYŃSKA 1979), but some data are also indicative or suggestive of uplifts in Late Tithonian to Early Berriasian time, and in Early Valanginian (pre-Platylenticeras and Polyptychites) time (KUTEK *& al.* 1989).

With the southwards directed Albian transgression, the usual tectonic pattern was re-established, and the axis of the Meta-Carpathian Arch shifted south of the present-day front of the Carpathians. The thickness of the Upper Cretaceous decreases from over 2500 m in the Mogilno-Łódź Synclinorium in Central Poland to less than 1000 m in the Miechów Synclinorium (JASKOWIAK-SCHOENEICHOWA & KRASSOWSKA 1988). The position of the axis of the Meta-Carpathian Arch in the Paleogene is indicated by the fact that Upper Cretaceous deposits, that are still present at the Carpathian border, were totally removed by pre-Miocene erosion from an area of the cratonic substratum of the Carpathian nappes situated but a short distance south of the northern front of these nappes (MORYC 1985).

From this review it follows that the Late Jurassic tectonic pattern observable in south-eastern cratonic Poland, and to some extent also the Middle Jurassic and Early Cretaceous ones, differ from the pattern usually displayed by that region in the Permian, the rest of the Mesozoic, and the Paleogene.

THE HOLY CROSS LINEAMENT

The Holy Cross Lineament (Text-fig. 3) separates, in the Paleozoic Core of the Holy Cross Mountains, the (southern) Kielce Zone from the (northern) Łysogóry Zone, which underwent a different structural and depositional development in Cambrian to Carboniferous time (KUTEK & GŁAZEK 1972). This indicates that this lineament corresponds to a tectonic discontinuity in the pre-Permian basement. The lineament extends farther WNW, where its presence is manifested at the junction of the Miechów and the Łódź-Mogilno Synclinoria by a set of Laramide anticlines (Text-fig. 2), and also by the thickness pattern of Permian and Mesozoic deposits (Text-figs 7–8). To ESE, an obvious manifestation of the Holy Cross Lineament can be found in the characteristic extention of the Mid-Polish Anticlinorium to the east, south of this lineament (Text-figs 2–3). The segment of the Mid-Polish Anticlinorium situated south of the Holy Cross Lineament, as well as the corresponding segment of the Polish Rift Basin, will be referred to as peri-Carpathian.

The Holy Cross Lineament usually manifested itself in the Mesozoic as the southern limit of areas of stronger subsidence. In the Holy Cross Mountains, for instance, the Buntsandstein, which is up to 400 m, but usually much less, thick south of the Holy Cross Lineament, attains a thickness of about 1000 m a short distance to the north of this lineament (SENKOWICZOWA 1970, KUTEK & GŁAZEK 1972, SZYPERKO-TELLER & MORYC 1988). Liassic sediments, in turn, which reach a thickness of about 1 km north of the lineament, are absent, or strongly reduced, on its southern side (KARASZEWSKI & KOPIK 1970, KUTEK & GŁAZEK 1972). At the junction of the Miechów and Łódź-Mogilno Synclinoria, a distinct gradient zone can be recognized along the Holy Cross Lineament (Text-fig. 8; *see* HAKENBERG 1980), with an increase of about 1.5 km to the north of the aggregated thickness of Permian, Triassic and Jurassic deposits despite the fact that, exceptionally, thicknesses of Upper Jurassic deposits increase south of the lineament (Text-figs 7 and 9–11).

JURASSIC SEQUENCES AND THICKNESS PATTERNS

The thicknesses of the Upper Jurassic deposits of cratonic Poland (Text-fig. 9) increase southwards from up to about 850 m in the Polish Lowland, *i.e.* from the area occupied by the Central European Basin, to about 1300 m in the Carpathian foreland. It is significant, moreover, that complete

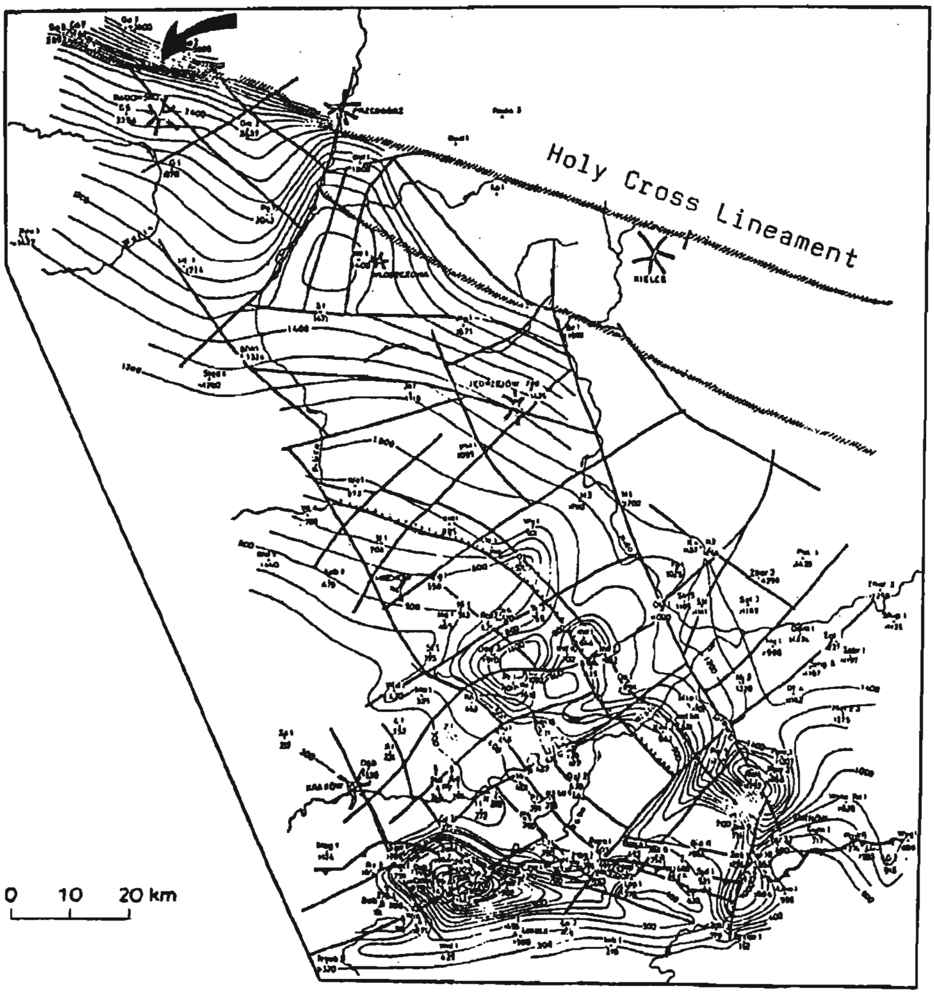


Fig. 8. Map of thicknesses of Permian to Jurassic deposits in the Miechów Synclinorium (*reproduced from* HAKENBERG 1980); arrow indicates a gradient zone coinciding with the Holy Cross Lineament

Upper Jurassic successions, including the Volgian, are preserved in Central Poland, whereas in the Miechów Synclinorium, near the Carpathians, the thicker Upper Jurassic succession encompasses only Oxfordian, Lower Kimmeridgian and early Upper Kimmeridgian deposits. Thus, the Upper Jurassic thickness pattern markedly contrasts with those displayed by the Permian, Triassic, Lower and Middle Jurassic, and also the Early and Upper Cretaceous deposits of cratonic Poland.

Upper Jurassic thicknesses of up to 1100 m were well documented in the SW margin of the Holy Cross Mts (Kutek 1968, 1969; Matyja 1977), and in boreholes in the region of Dąbrowa Tarnowska (Morycowa & Moryc 1976). Accordingly, several detailed stratigraphic interpretations presented in the following sections of this paper are largely based on data from these regions.

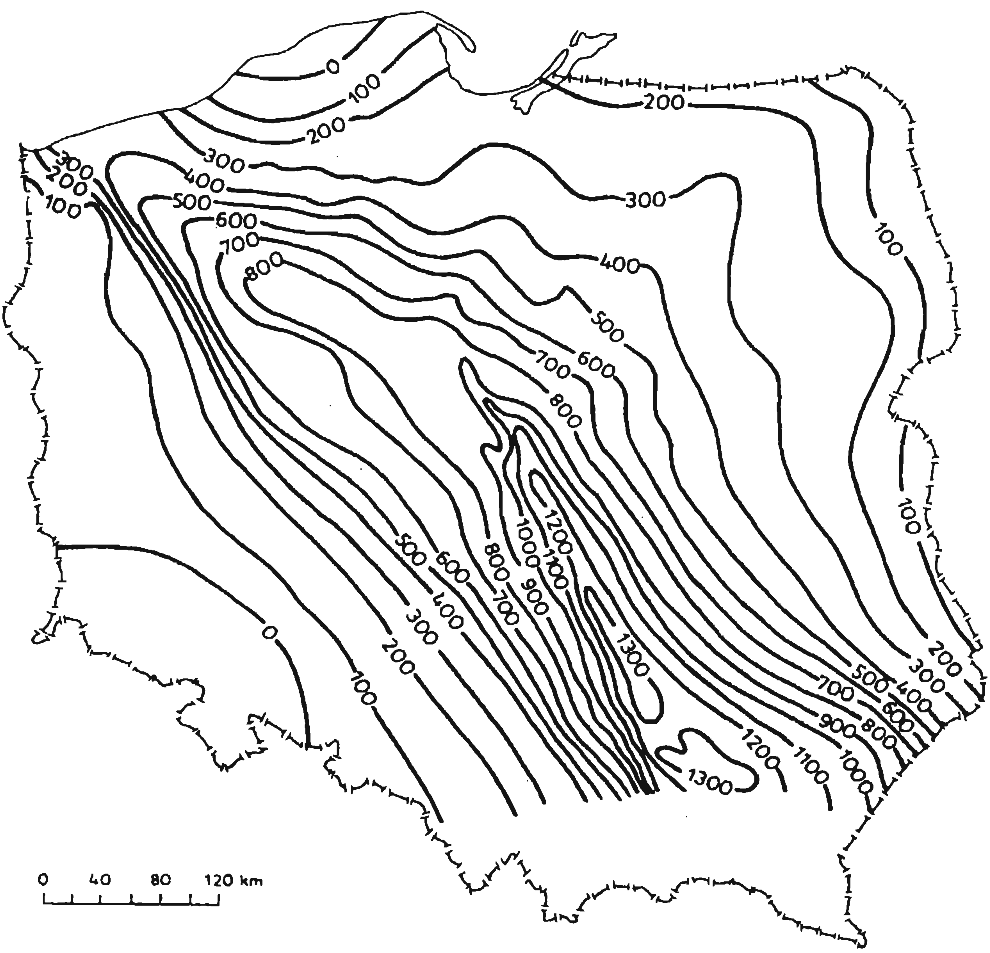


Fig. 9. Isopachs (restored in part) of Upper Jurassic deposits in cratonic Poland (*redrawn from* Niemczycka & Brochwicz-Lewiński 1988)

Text-figure 9 reproduces a paleotectonic map published by NIEMCZYCKA & BROCHWICZ-LEWIŃSKI (1988), with isopachs that refer in part to restored thicknesses of Upper Jurassic deposits, so that some of the suggested interpretations are disputable, particularly those pertinent to areas from which Upper Jurassic sediments have been totally, or largely, removed by erosion. Nonetheless, marked increase of thicknesses of Upper Jurassic deposits towards the Carpathians can be recognized unequivocally in the Miechów and Łódź-Mogilno Synclinoria, where these deposits are still covered by Cretaceous sediments (*comp.* Text-figs 2–3 *and* 10–11).

THE RELEVANT SEQUENCES

Tectonic events can be better interpreted in terms of allostratigraphic units, than with reference to chronostratigraphic units which, in the Jurassic, are based on biostratigraphic units. Three allostratigraphic units will be distinguished hereafter. These are transgressive-regressive units which will be called sequences; this term, however, should not imply a eustatic nature of these units. They can be termed tectono-stratigraphic units in accordance with the terminology discussed by VAIL & *al.* (1991).

The following sequences, which are chiefly developed in south-eastern and central regions of cratonic Poland, are distinguished: (1) the COK (Callovian — Oxfordian — Kimmeridgian) Sequence, which includes deposits representing the Callovian, the Oxfordian, and the Platynota and Hypselocyclum Zones of the Lower Kimmeridgian; (2) the LUK (Lower Kimmeridgian — Upper Kimmeridgian) Sequence, consisting of the "Coquina Formation", with a lower boundary situated near the base of the Divisum Zone and an upper boundary within the Eudoxus Zone; and (3) the KVB (Kimmeridgian — Volgian — Berriasian) Sequence, extending from the Eudoxus Zone up to the Lower Berriasian; it includes the shaly ammonite-bearing Pałuki Formation (DEMBOWSKA 1979), that ranges up to the Regularis Horizon of the Middle Volgian Scythicus Zone (KUTEK 1994), and the overlying Kcynia Formation, which consists of sediments developed in a Purbeck-type facies (DEMBOWSKA 1979, MAREK & RACZYŃSKA 1979, MAREK & *al.* 1989).

As major disconformities resulting from the extensive Early Callovian transgression, and regressive tendencies responsible for the development of Purbeck-type facies, are known from several regions of Europe, Poland included, the COK, LUK and KVB Sequences might be assigned to one supersequence ranging from the Lower Callovian to the Lower Berriasian. However, as suggested below, these three sequences developed in changing tectonic regimes.

No Jurassic sequences of pre-Callovian age are distinguished in this paper. However, it is pertinent to note that Bajocian deposits, which are not represented in parts of the Carpathian foreland, attain a thickness of nearly 800 m in Central Poland, within the Central European Basin (DADLEZ 1993).

Conversely, the Bathonian deposits, which are transgressive in several regions of cratonic Poland, display much less contrasted thicknesses, which are of the order of 100 m in Central Poland (DAYCZAK-CALIKOWSKA & MORYC 1988).

THE *COK* SEQUENCE

This sequence includes deepening-upward Callovian sediments, transgressive and terrigenic in parts, and shallowing-upward carbonates representing the Oxfordian and the Lower Kimmeridgian Platynota and Hyposelocyclum Zones. The thickness of the Oxfordian and Lower Kimmeridgian sediments of this sequence (Text-fig. 10) increases southwards from 350–375 m near the northern boundary of the Łódź-Mogilno Synclinorium to about 700 m near the Carpathians (KUTEK 1968, 1969; MROZEK 1975; DEMBOWSKA 1977; MATYJA 1977; MATYJA & *al.* 1985; WIERZBOWSKI & KUTEK 1990). At some places, these sediments attain thicknesses of about 800 m near the Carpathians (*comp.* MORYCOWA & MORYC 1976, NIEMCZYCKA & BROCHWICZ-LEWIŃSKI 1988).

The cross-section shown in Text-figure 10 has been drawn along the Miechów and Łódź-Mogilno Synclinoria, at some distance west of the Mid-Polish Anticlinorium (Text-fig. 3), from the region of Dąbrowa Tarnowska — Szczucin near the Carpathians (MORYCOWA & MORYC 1976), through the boreholes Jędrzejów, Węgleszyn and Włoszczowa in the middle and northern parts of the Miechów Synclinorium (WIERZBOWSKI & KUTEK 1990), and then through the southern part of the Łódź-Mogilno Synclinorium (MROZEK 1975, DEMBOWSKA 1977). In these regions the upper boundary of the COK Sequence coincides with the easily recognizable base of the overlying Coquina Formation. Near the northern boundary of the Łódź-Mogilno Synclinorium, in the Zalesie Anticline, the interval corresponding to the Oxfordian and to the Platynota and Hypselocyclum Zones was distinguished chiefly on biostratigraphic grounds (MATYJA & *al.* 1985).

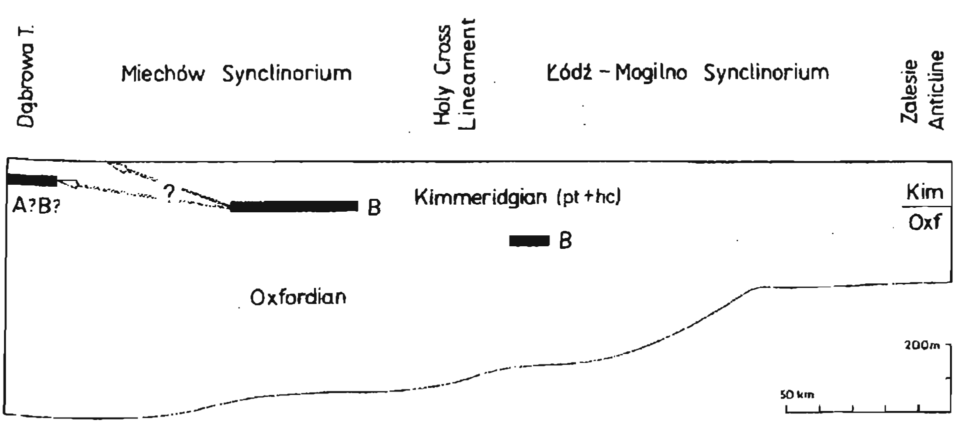


Fig. 10. Cross-section through Oxfordian and Kimmeridgian deposits of the *COK* Sequence in the Miechów and Łódź-Mogilno Synclinoria, approximately along line *3* in Text-figure 3; Callovian deposits of the sequence omitted

A, B — Marker horizons *A* and *B*; pt — Platynota Zone, hc — Hypselocyclum Zone

The discussed cross-section (Text-fig. 10), especially its part referring to the Łódź-Mogilno Synclinorium, is strongly smoothed.

The Callovian is transgressive on the Bathonian, or on pre-Jurassic rocks, for instance in the Cracow region and in the SW margin of the Holy Cross Mts; the Macrocephalus (or Typicus) Zone is missing in several sections (RÓŻYCKI 1953, SIEMIĄTKOWSKA-GIŻEJEWSKA 1974, DAYCZAK-CALIKOWSKA & MORYC 1988). In the Miechów and Łódź-Mogilno Synclinoria and their bordering zones, the Callovian is at best a little over 25 m thick, and thicknesses of a few metres are not uncommon (RÓŻYCKI 1953, SIEMIĄTKOWSKA-GIŻEJEWSKA 1974, DAYCZAK-CALIKOWSKA & MORYC 1988). In some regions, *e.g.* in that of Cracow, the Callovian sediments are thicker, and largely developed as sands and gravels, when infilling topographic lows, and much reduced and condensed on highs (DŻUŁYŃSKI 1950, GIŻEJEWSKA & WIECZOREK 1977, KUTEK & *al.* 1982). In general, the Callovian sedimentation did not keep pace with subsidence, and in some regions was strongly controlled by paleorelief, inherited or sustained by syndepositional faulting (KUTEK & *al.* 1982); hence the thicknesses of the Callovian do not provide a good measure of the magnitude of subsidence. For this reason the Callovian sediments of the COK Sequence have not been indicated in the discussed cross-section (Text-fig. 10). The subsidence, however, that led to the accumulation of the very thick Oxfordian sediments of the COK Sequence, commenced already in the Callovian.

In several regions, *e.g.* in the Cracow-Silesian Monocline and in the SW margin of the Holy Cross Mts, the Upper Callovian sediments are often strongly condensed; in particular, the Lamberti Zone is missing or strongly reduced. The basal Oxfordian Mariae Zone is also often extremely thin (RÓŻYCKI 1953, SIEMIĄTKOWSKA-GIŻEJEWSKA 1974, MATYJA 1977, KOPIK 1979). In terms of sequence stratigraphy, the condensed sections spanning the Callovian/Oxfordian junction may be interpreted as indicative of a peak of transgression (*see* LOUTIT & *al.* 1988). However, the stratigraphic picture is more complicated. Condensed Lower or Middle Callovian sediments appear in several sections, being overlain in some of them by expanded younger Callovian sediments. Moreover, *e.g.* in the region of Cracow, there occur expanded marly successions straddling the Callovian/Oxfordian junction (RÓŻYCKI 1953, SIEMIĄTKOWSKA-GIŻEJEWSKA 1974, KOPIK 1979, GIŻEJEWSKA 1981, KUTEK & *al.* 1982).

A carbonate ramp began to develop in the Oxfordian in eastern Poland, in the Lublin region (which belongs to the East European Craton). This ramp prograded to the west and south-west, and in the Early Kimmeridgian was transformed into a well-developed carbonate platform (KUTEK & *al.* 1984). The progradation of the ramp was an essentially diachronic process so that, as a consequence, a sponge megafacies is replaced by a shallower carbonate megafacies, typically rich in corals, nerineacean gastropods and *Diceras*, in the Bifurcatus Zone in the NE margin of the Holy Cross Mts (GUTOWSKI 1992), in the Planula Zone in their SW margin (KUTEK 1968, 1969; MATYJA & *al.* 1989), and at still higher levels in the Miechów Synclinorium (WIERZBOWSKI & KUTEK 1990). Heavily cross-bedded oolites and laminated micrites topping the carbonate platform are widely developed in the NE and SW margins of the Holy Cross Mts, as well as in the Miechów and Łódź-Mogilno Synclinoria (KUTEK 1968, 1969; MROZEK 1975; WIERZBOWSKI & KUTEK 1990; GUTOWSKI 1992).

In the regions concerned, the boundary between the Oxfordian and Kimmeridgian Stages is usually located in sediments poor in ammonites, and hence cannot be traced with desirable precision. However, two virtually isochronous, shaly-marly horizons, recognizable also in boreholes, provide some useful stratigraphic information. The first horizon is situated high in the Planula Zone and can be recognized in the Cracow-Silesian Monocline, in the western part of the Miechów Synclinorium, and in parts of the Łódź-Mogilno Synclinorium. This horizon, which will be called marker horizon *A*, corresponds *e.g.* to "unit *2*" of JAWOR (1980) in the part of the Miechów Synclinorium just east of Cracow, and to the "lower marly (*zmd*) unit" of KUTEK & *al.* (1977) in the Cracow-Silesian Monocline. The second horizon (marker horizon *B*), which is situated low in the Platynota Zone, thus near the Oxfordian/Kimmeridgian junction, can also be recognized in the north-eastern part of the Miechów Synclinorium, in the SW margin of the Holy Cross Mts, and farther north in the regions of Sulejów and Tomaszów Mazowiecki (KUTEK 1968, MATYJA & WIERZBOWSKI 1988, WIERZBOWSKI & KUTEK 1990), and is therefore of greater correlation value. This horizon corresponds *e.g.* to the "unit *4*" of JAWOR (1980), to the "middle marly (*zms*) unit" of KUTEK & *al.* (1977), to the "lowermost marly horizon" of KUTEK (1968) in the SW margin of the Holy Cross Mts, and to the "unit *XII*" of MROZEK (1975) in the southern part of the Łódź-Mogilno Synclinorium.

The marker horizon *B* permits to distinguish two intervals within the Upper Jurassic sediments of the COK Sequence, the interval *I* below this horizon, and the interval *II* above (*see* Text-fig. 10). The thickness of the interval *I*, which roughly corresponds to the Oxfordian, increases southwards from about 225 m in the Zalesie Anticline (Text-fig. 3) just north of the Łódź-Mogilno Synclinorium (MATYJA & *al.* 1985), through 350–500 m in the southern part of this synclinorium (MROZEK 1975) and 700–750 m in the south-western margin of the Holy Cross Mountains (KUTEK 1968, 1969; MATYJA 1977), to about the same value or 800 m in the Dąbrowa Tarnowska region near the Carpathians (MORYCOWA & MORYC 1976). Significantly, a contrasting tendency is displayed by the interval *II*, which includes the Lower Kimmeridgian Platynota and Hypselocyclum Zones (Text-fig. 10). Its thickness is usually between 100 m and 135 m in the south-western margin of the Holy Cross Mountains (KUTEK 1968), but north of the Holy Cross Lineament it amounts to 250–300 m in the southern part of the Łódź-Mogilno Synclinorium (MROZEK 1975) and about 300 m in the regions of Sulejów and Tomaszów Mazowiecki (MATYJA & WIERZBOWSKI 1988); still farther north, in the Zalesie Anticline, the thickness of the Lower Kimmeridgian can be estimated to be about 150 m (MATYJA & *al.* 1985). This signifies a return to the usual Mesozoic thickness pattern, with greater thicknesses north of the Holy Cross Lineament, in the Central European Basin.

The implied change in subsidence pattern is also reflected in facies distribution. In the SW margin of the Holy Cross Mts the lithofacies with corals, nerineaceans and *Diceras* does not extend above the marker horizon *B*, and extensive lithosomes of shallow-water oolites appear but a short distance above this horizon (KUTEK 1968, 1969). To the north, in the regions of Sulejów and Tomaszów Mazowiecki (Text-fig. 3), carbonates with corals, nerineaceans and *Diceras* occur below and above marker horizon *B*, attaining thicknesses of up to 150 m above this horizon. On the other hand, in general, oolite sedimentation is porly expressed in the higher portion of the Lower Kimmeridgian of these regions (BARCZYK 1961, WIECZOREK 1975, MATYJA & WIERZBOWSKI 1988). An intermediate succession of facies is situated in the northernmost part of the SW margin of the Holy Cross Mts, in the zone of the Holy Cross Lineament (Text-fig. 3). Here, the sediments with corals do not extend above the marker horizon *B*, but the main occurrences of oolites appear fairly high above this horizon, in the Hypselocyclum Zone (KUTEK 1968).

The bathymetric contrast between the sediments of the coral lithofacies and the overlying sediments is not great, as indicated by the fact that the former sediments are topped by, and alternate with, oolites in some sections (KUTEK 1968, 1969; MATYJA & WIERZBOWSKI 1988). Hence, the increase in thickness of the interval *II* to the north cannot be interpreted as a result of mere infilling of a pre-existing space of accomodation, but has to be explained in terms of subsidence.

The highest portions of the COK Sequence from the borehole sections of the region of Dąbrowa Tarnowska — Szczucin near the Carpathians, which have hitherto yielded no ammonites

(MORYCOWA & MORYC 1976), are difficult to correlate with the sections of the SW margin of the Holy Cross Mts, much better documented biostratigraphically (KUTEK 1968). With marked contrast to the latter sections, the lithofacies with corals extends near the Carpathians up to the base of the overlying Coquina Formation. Two possibilities of correlation can be suggested. First, the shaly layer "*C-c*" ("*A?B?*" in Text-figure 10) distinguished by MORYCOWA & MORYC (1976) may be interpreted as corresponding to the marker horizon *B*. This would imply a marked contrast in vertical facies distribution between the two regions, and a thickness of interval *II* of about 100 m near the Carpathians. However, the layer "*C-c*" may be correlated on still better ground with the marker horizon *A* well developed near the Carpathians in the western part of the Miechów Synclinorium (JAWOR 1980, unit *2*). This would imply that in the region of Dąbrowa Tarnowska — Szczucin some high portions of the COK Sequence were removed by erosion prior to the accumulation of the overlying Coquina Formation. This interpretation finds some support in the presence of quartz material in the basal part of the Coquina Formation near the Carpathians (MORYCOWA & MORYC 1976), a feature not found in the sections along the SW margin of the Holy Cross Mts. The latter interpretation would imply that the primary thickness of the COK Sequence was well over 800 m in the region of Dąbrowa Tarnowska — Szczucin.

The following conclusions can be drawn from the data outlined above. The sediments of the interval *I* corresponding approximately to the Oxfordian increase in thickness to the south, towards the present-day Carpathian front, whereas an increase in thickness to the north is displayed by the sediments of interval *II*, that represents the Platynota and Hypselocyclum Zones of the Lower Kimmeridgian (Text-fig. 10). This implies a change of subsidence patterns, from stronger subsidence in the south in the Oxfordian, to stronger subsidence to the north in the Early Kimmeridgian. From this it follows that the formation of the COK Sequence, which as a transgressive-regressive unit is an entity, took place in two successive, distinct tectonic regimes.

Of course, the change of tectonic regime reflected in the thickness patterns discussed above needs not to take place exactly at the level of marker horizon *B*. However, there are independent data suggesting a change of tectonic regime near the Oxfordian/Kimmeridgian boundary. Good evidence is provided by the Cracow-Silesian Monocline (the "Polish Jura Chain" of some authors) of Late Oxfordian syndepositional faulting, associated with spectacular occurrences of turbidites, debris flows and submarine slumps in the Bimammatum and Planula Zones (KUTEK & *al.* 1990), and some data from the Cracow region suggest syndepositional faulting in the Callovian, Early Oxfordian and early Middle Oxfordian, up to the Plicatilis Zone (KUTEK & *al.* 1982). More generally, Oxfordian deposits often reveal dramatic lateral changes in lithofacies in the Cracow-Silesian Monocline, the Miechów Synclinorium and the south-western margin of the Holy Cross Mountains, whereas the Lower Kimmeridgian sediments included in interval *II* display a smooth pattern of laterally extensive lithosomes (DŻUŁYŃSKI 1952, KUTEK 1968, KUTEK & *al.* 1977, MATYJA 1977, WIERZBOWSKI & KUTEK 1990). This is suggestive of fault-controlled sedimentation in the Oxfordian, and of absence of pronounced fault activity in the Early Kimmeridgian.

A broad marine connection with the Carpathian Domain of the Tethys resulted from the Callovian to Early Kimmeridgian downwarping of the Meta-Carpathian Arch, which had obvious biogeographic consequences. Mixed faunas, composed of ammonites of both Tethyan and Boreal origin (Macrocephalitidae, Kosmoceratidae, Reineckeiidae, Cardioceratidae, and Haploceratаceae) occur in the Callovian of cratonic Poland (RÓŻYCKI 1953, GIŻEJEWSKA & MATYJA 1978, KOPIK 1979). The ammonite faunas found in the

Oxfordian as well as in the Lower Kimmeridgian Platynota and Hypselocyclum Zones north of the Carpathians are of Submediterranean type (KUTEK & *al.* 1984). On the other hand, the Cardioceratidae penetrated in the Middle Oxfordian as far south as to the Czorsztyn Ridge, a tectonic unit included in the Pieniny Klippen Belt of the Carpathians (KUTEK & WIERZBOWSKI 1986, KUTEK 1990).

THE *LUK* SEQUENCE (THE COQUINA FORMATION)

This sequence comprises a formation largely consisting of coquinas, which is developed in several regions of south-eastern and central cratonic Poland (KUTEK 1968, DEMBOWSKA 1979). This formation, which has not yet been named in a strictly formal way, will be referred to as the Coquina Formation. The base of this formation is usually situated a little beneath the base of the Divisum Zone (*see below*). In the south-western margin of the Holy Cross Mountains and in the Miechów Synclinorium the Coquina Formation is

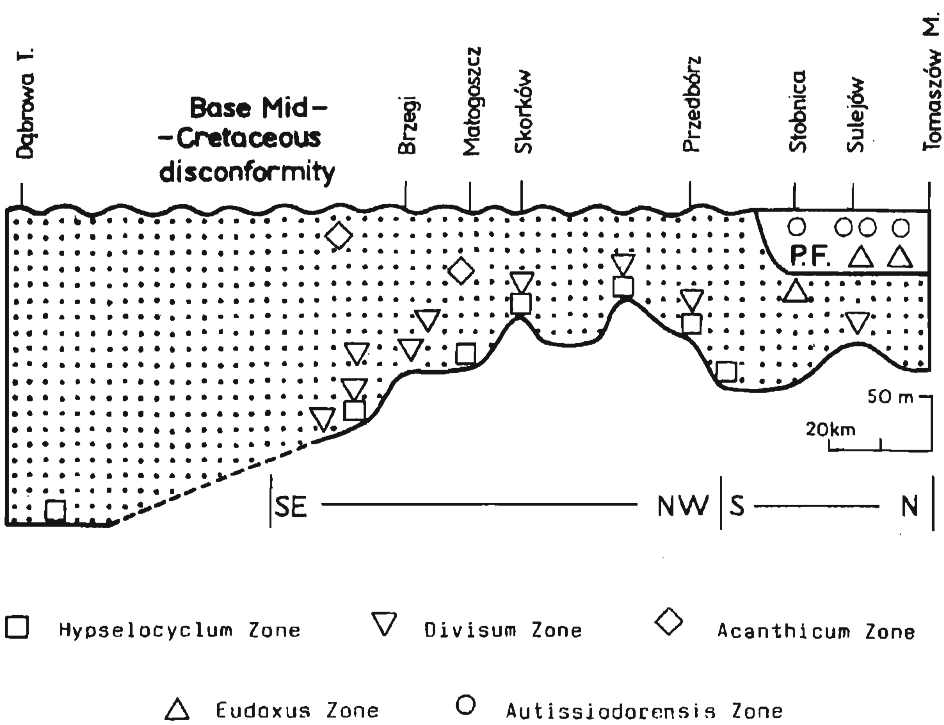


Fig. 11. Cross-section through the Coquina Formation along the western and south-western margins of the Holy Cross Mts and Dąbrowa Tarnowska region (*see* Text-fig. 3); indicated are ammonite occurrences of different age; P.F. — Pałuki Formation

incomplete, being disconformably overlain by the Albian or Cenomanian. In these regions (Text-fig. 11), the youngest ammonites found in the Coquina Formation represent the Acanthicum Zone (KUTEK 1968). North of the Holy Cross Lineament, in the regions of Sulejów and Tomaszów Mazowiecki at the western margin of the Mid-Polish Anticlinorium (*see* Text-fig. 3), and in parts of the Łódź-Mogilno Synclinorium, the Coquina Formation is conformably overlain by the Pałuki Formation; here ammonites indicative of the Eudoxus Zone have been found both in the topmost parts of Coquina Formation, and in the basal parts of the Pałuki Formation (KUTEK 1961, 1994; KUTEK & ZEISS 1994). An important point is that the Coquina Formation, which is only 50-150 m, and in most cases less than 100 m, thick north of the Holy Cross Lineament (MROZEK 1975, DEMBOWSKA 1977, MATYJA & WIERZBOWSKI 1988), increases southwards in thickness to over 200 m in the south-western margin of the Holy Cross Mountains (KUTEK 1968, PSZCZÓŁKOWSKI 1970), and to about 300 m near the Carpathians (MORYCOWA & MORYC 1976), despite the fact that the formation represents a more restricted stratigraphic interval in the south (*see* Text-fig. 11).

The thicknesses of the Coquina Formation indicated in Text-figure 11 are those displayed in the regions of Tomaszów Mazowiecki and Sulejów, in the SW margin of the Holy Cross Mts, and in the region of Dąbrowa Tarnowska — Szczucin. In the regions, where the Coquina Formation is topped by the base-Mid-Cretaceous disconformity/unconformity, the preserved thicknesses of this formation depend in part on the extent of pre-Albian (or pre-Cenomanian) erosion (KUTEK 1968, PSZCZÓŁKOWSKI 1970). For instance, the increase in thickness of the Coquina Formation just north of Przedbórz (Text-fig. 11) is due to the fact that additional strata, belonging to this formation, appear beneath the Albian to the north (*see* KUTEK 1968, Table 2).

The Coquina Formation (and the LUK Sequence) includes: the "Kimmeridgian" (Formation *D*) of MORYCOWA & MORYC (1976) in the region of Dąbrowa Tarnowska — Szczucin; the Skorków, Brzegi, Staniewice and Top Lumachelles as well as the Upper Platy Limestones and Top Clays of KUTEK (1986) in the SW margin of the Holy Cross Mts; the Stobnica Beds of KUTEK (1962) in the regions of Sulejów and Tomaszów Mazowiecki; and the units *XXIV* and *XXV* of MROZEK (1975) as well as the "Lower Kimmeridgian" of DEMBOWSKA (1977) in the southern part of the Łódź-Mogilno Synclinorium. The formation here called the Coquina Formation was distinguished as the Formation *V* by DEMBOWSKA (1979) in Central Poland. The Coquina Formation is also well developed in the NE margin of the Holy Cross Mts (*see* DĄBROWSKA 1983, KUTEK 1983, GUTOWSKI 1992).

Coquinas comprising *Lopha* (= *Alectryonia*) as the distinctive constituent, from 1 to 30 m thick, are usually developed in the lower part of the Coquina Formation; in the SW margin of the Holy Cross Mts, the *Lopha* coquinas are often preceded by a thin layer of biomicrites (local Polish designation: *grab limestones*), or by oolite lithosomes. All these are open-marine sediments, in contrast with the highest sediments of the underlying COK Sequence, which are largely developed as laminated micrites indicative of a more restricted sedimentary environment (KUTEK 1968, 1969; WIERZBOWSKI & KUTEK 1990; GUTOWSKI 1992). The bulk of the Coquina Formation consists of *Nanogyra* coquinas alternating with micritic limestones and shales. Detrital varieties of coquinas predominate in the upper part of the Coquina Formation, and quartz material, cross-bedding and indices of erosion appear in its highest portions, *e.g.* in the Top Lumachelle in the SW margin of the Holy Cross Mts, and in the Malenia Coquina in their NE margin (KUTEK 1961, 1968, 1969, 1983; MROZEK 1975; DĄBROWSKA 1983; GUTOWSKI 1992). Hence, the Coquina Formation can be interpreted as a transgressive-regressive unit.

The best biostratigraphic data relative to the base of the Coquina Formation are those from the SW margin of the Holy Cross Mts. They were interpreted by KUTEK (1968) in accordance with GEYER's (1961) concept of the Hypselocyclum and Divisum Zones but they can be re-interpreted stratigraphically, and in part also taxonomically, applying the biostratigraphic subdivision of the Lower Kimmeridgian Substage established by ATROPS (1982). Four successive assemblages of ammonites can be recognized in the lower part of the Coquina Formation, in stratigraphic positions outlined diagrammatically in Text-figure 12*A* (for details *see* Table 2 *in* KUTEK 1968).

Assemblage A: *Rasenia (Prorasenia), Rasenia (Eurasenia), Ataxioceras*, a.o. *A. (Parataxioceras)* cf. *lothari* (OPPEL), no *Crussoliceras* — Lothari Subzone, probably Hypselocyclum Horizon (*comp.* Text-fig. 1). The assemblage occurs in biomicrites underlying *Lopha* (=*Alectryonia*) coquinas where these attain considerable thicknesses. Ammonite occurrences *7, 37, 40, 41* and *43* in KUTEK (1968).

Assemblage B: *Rasenia (Eurasenia), Ataxioceras* and *Crussoliceras* — Semistriatum Horizon (lowest Divisum Zone *sensu* GEYER), indicated by the co-occurrence of *Ataxioceras* and *Crussoliceras*. *Lopha* coquinas; Ammonite occurrences *9, 17*, and *18* in KUTEK (1968).

Assemblage C: *Crussoliceras*, but no *Ataxioceras*, nor *Rasenia* — either Perayensis Horizon, or lowest Divisum Zone *sensu* ATROPS. Ammonite occurrences *16* and *42* at high levels within *Lopha* coquinas (KUTEK 1968).

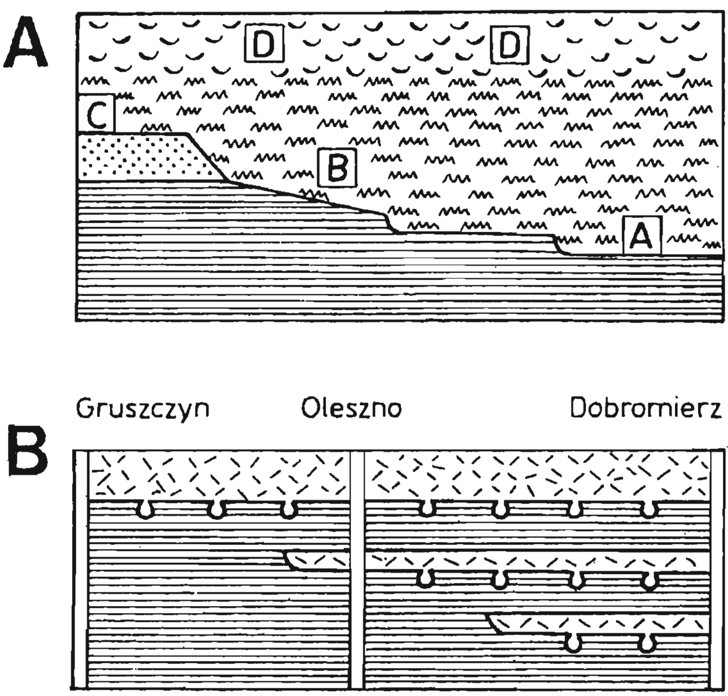


Fig. 12. Cartoons displaying data relevant to the interpretation of the basal part of the Coquina Formation (not to scale)

A: Micritic limestones, laminated in part — continuous lines; oolite limestones — stippled; *Lopha* (= *Alectryonia*) coquinas and associated sediments — wavy hachures; *Nanogyra* coquinas and associated sediments — comma-like hachures; letters *A, B, C* and *D* refer to ammonite assemblages discussed in text

B: Laminated micritic limestones — continuous lines; biodetrital and biomicritic limestones, and *Lopha* (= *Alectryonia*) coquinas — disordered hachures; for other explanations see text

Assemblage D: *Crussoliceras* and *Aspidoceras uhlandi* (OPPEL) — Uhlandi Subzone of the Divisum Zone. Ammonite occurrences *3, 4, 5, 6, 8, 13, 14, 15, 36, 38*, and *39* above *Lopha* coquinas (KUTEK 1968).

From the biostratigraphic and lithostratigraphic data provided by the SW margin of the Holy Cross Mts (KUTEK 1968) the conclusion can be drawn that the base of the Coquina Formation, if taken in all sections at the base of *Lopha* coquinas or associated biomicrites, is diachronic in an interval of about 25 m, the base of the formation being usually situated beneath the base of the Divisum Zone *sensu* ATROPS (*see* Text-fig. 11*A*). This diachronism, however, vanishes in part if some oolite lithosomes, which underlie *Lopha* coquinas in some sections, but appear to belong to the same sedimentary tract as such coquinas occurring at low levels in the Coquina Formation (and comprising sometimes oolitic intercalations) in other sections, are also included in this formation (*see* Text-fig. 11*A*). Nevertheless, the generalized base of the Coquina Formation would still display some diachronism. In this context, it is significant that in some sections (*e.g.* at Oleszno and in the Dobromierz Anticline) thin biodetrital or biomicrtitic intercalations appear beneath the top of laminated micrites (the "Shaly Limestones" of KUTEK, 1968) that underlie the bulk of the Coquina Formation. This suggests that the generalized base of the Coquina Formation is composed of several biodetrital or biomicritic layers, isochronic but of limited lateral extention, which appear at different levels and in most cases peter out within micritic sediments (Text-fig. 12*B*). Significantly, hardgrounds expressed *e.g.* by bivalve borings (GRUSZCZYŃSKI 1979, 1986), can be associated with anyone of these successive, closely spaced levels. From this it follows that hardgrounds, a large number of which have been recognized in the Lower Kimmeridgian of the SW margin of the Holy Cross Mts (KAŻMIERCZAK & PSZCZÓŁKOWSKI 1968; KUTEK 1969; GRUSZCZYŃSKI 1979, 1986) do not provide a useful tool for stratigraphic correlations over large distances.

From the data outlined above it can be concluded that no considerable stratigraphic hiatuses are connected with the base of the Coquina Formation in the SW margin of the Holy Cross Mts. However, as remarked, a significant gap may occur at the base of the Coquina Formation in the region of Dąbrowa Tarnowska — Szczucin near the Carpathians. Occurrences of *Ataxioceras* in the basal part of the Coquina Formation in the latter region (MORYCOWA & MORYC 1976), and also in the NE margin of the Holy Cross Mts (GUTOWSKI 1992), indicate that the base of this formation is situated beneath the base of the Divisum Zone in both these regions.

The top of the Coquina Formation is usually taken at the top of the highest occurrences of *Nanogyra* coquinas. However, borehole data from the region of Tomaszów Mazowiecki and the Łódź-Mogilno Synclinorium (KUTEK & ZEISS 1994; *and* unpublished material of the present writer) indicate that the highest layers of such coquinas appear in different sections in the midst of the Eudoxus Zone, but not exactly at the same levels. From this it follows that the upper boundary of the Coquina Formation is again a composite boundary, composed of coquina layers of different lateral extent.

The assemblage *A* from the basal part of the Coquina Formation in the SW margin of the Holy Cross Mts, including *Ataxioceras* and *Rasenia*, does not differ biogeographically from earlier ammonite assemblages, occurring in the Hypselocyclum Zone within the COK Sequence (KUTEK 1968, KUTEK & *al.* 1984). The genus *Rasenia* does not extend into the Divisum Zone in Polish sections, which contain an ammonite fauna chiefly composed of *Crussoliceras* and *Aspidoceras*. This gives a Mediterranean touch to this fauna. The still poorly known ammonite faunas from the Acanthicum Zone and the lower Eudoxus Zone contain *Aspidoceras*, and species of *Aulacostephanus* known *a.o.* from the Submediterranean region of South Germany (KUTEK 1961, 1962, 1968; KUTEK & ZEISS 1994).

THE *KVB* SEQUENCE

This sequence consists of the Pałuki Formation, which extends from the upper Eudoxus Zone to the Regularis Horizon of the Middle Volgian Scythicus Zone (KUTEK 1994, KUTEK & ZEISS 1994), and of the Kcynia Formation, which ranges from the Zarajskensis Horizon of the Scythicus Zone (KUTEK 1994) up to, and including, the Lower Berriasian (DEMBOWSKA 1979, MAREK & RACZYŃSKA 1979, MAREK & *al.* 1989). Shales predominate in the Pałuki Formation, which is usually very rich in ammonites. The Kcynia Formation chiefly consists of a facies complex of the Purbeck type, comprising evaporitic, brackish and fresh-water sediments (MAREK & *al.* 1989). In Central Poland the Kcynia Formation is overlain, unconformably in part (Text-fig. 13), by Upper Berriasian marine sediments of the Rogoźno Formation (for more information on stratigraphy and facies of the lowermost Cretaceous of cratonic Poland, *see* MAREK & *al.* 1989).

In the region of Tomaszów Mazowiecki, the appearance of *Corbula* limestones with a low diversity fauna near the base of the Kcynia Formation, in the Zarajskensis Horizon, can be regarded as marking the onset of sedimentation in Purbeck-type conditions (KUTEK 1994). In this context it is worth of note that the Zarajskensis Horizon can be correlated with some part of the Upper Tithonian Calpionellid Zone *A*; thus the Zarajskensis Horizon is of Late, but not of early Late Tithonian, age (*comp.* KUTEK 1994).

No significant biostratigraphic hiatus is connected with the base of the Pałuki Formation in the Łódź-Mogilno Synclinorium and in the adjacent margin of the Mid-Polish Anticlinorium. In the NE margin of the Holy Cross Mts, however, the Krzyżanowice Nerineacean Limestones (DĄBROWSKA 1983) can be interpreted as a transgressive sediment. These limestones, which can be biostratigraphically dated as belonging to the Eudoxus or Autissiodorensis Zone, correspond to some level low in the Pałuki Formation.

The Pałuki and Kcynia Formations are preserved in Central Poland and have no counterparts in southern cratonic Poland, in the Miechów Synclinorium and its bordering zones, where Middle Cretaceous sediments rest directly on the Kimmeridgian or Oxfordian (Text-fig. 11).

The Pałuki Formation attains a thickness of about 150 m in the region of Tomaszów Mazowiecki, both its Upper Kimmeridgian and Lower Volgian portions being about 60 m thick, and its Middle Volgian portion (Quenstedti to Regularis Horizons) about 35 m thick. In the Łódź-Mogilno Synclinorium, the thicknesses of the Pałuki Formation ("Upper Kimmeridgian", "Lower Volgian" and "Middle Volgian" of DEMBOWSKA 1977), if complete, are usually between 100 and 200 m. These thicknesses show no systematic variation that could be interpreted as relevant to the development of the Meta-Carpathian Arch. The thicknesses of the Kcynia Formation are highly variable, in part as a result of halokinesis; they may amount to about 300 m in some sections (DEMBOWSKA & MAREK 1975).

There are some lines of evidence indicative or suggestive of a re-activation of the Meta-Carpathian Arch during a part of the timespan corresponding to the formation of the KVB (latest Kimmeridgian to Early Berriasian) Sequence. First, in Central Poland (*see* Text-fig. 13), the Upper Berriasian marine sediments overstep unconformably to the south successively older deposits of the KVB Sequence (DEMBOWSKA & MAREK 1975, KUTEK & *al.* 1989). Second, the accumulation of the Purbeck-type sediments of the Kcynia Formation in Central Poland, in the area of the Central European Basin, can be explained as

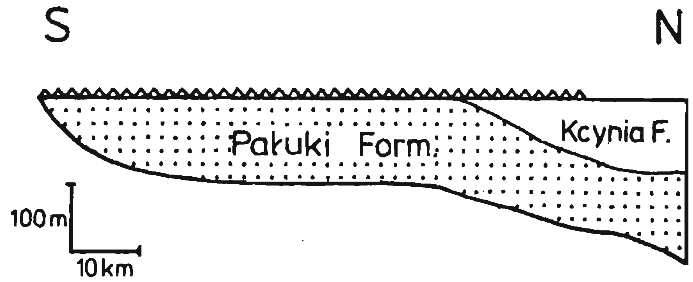


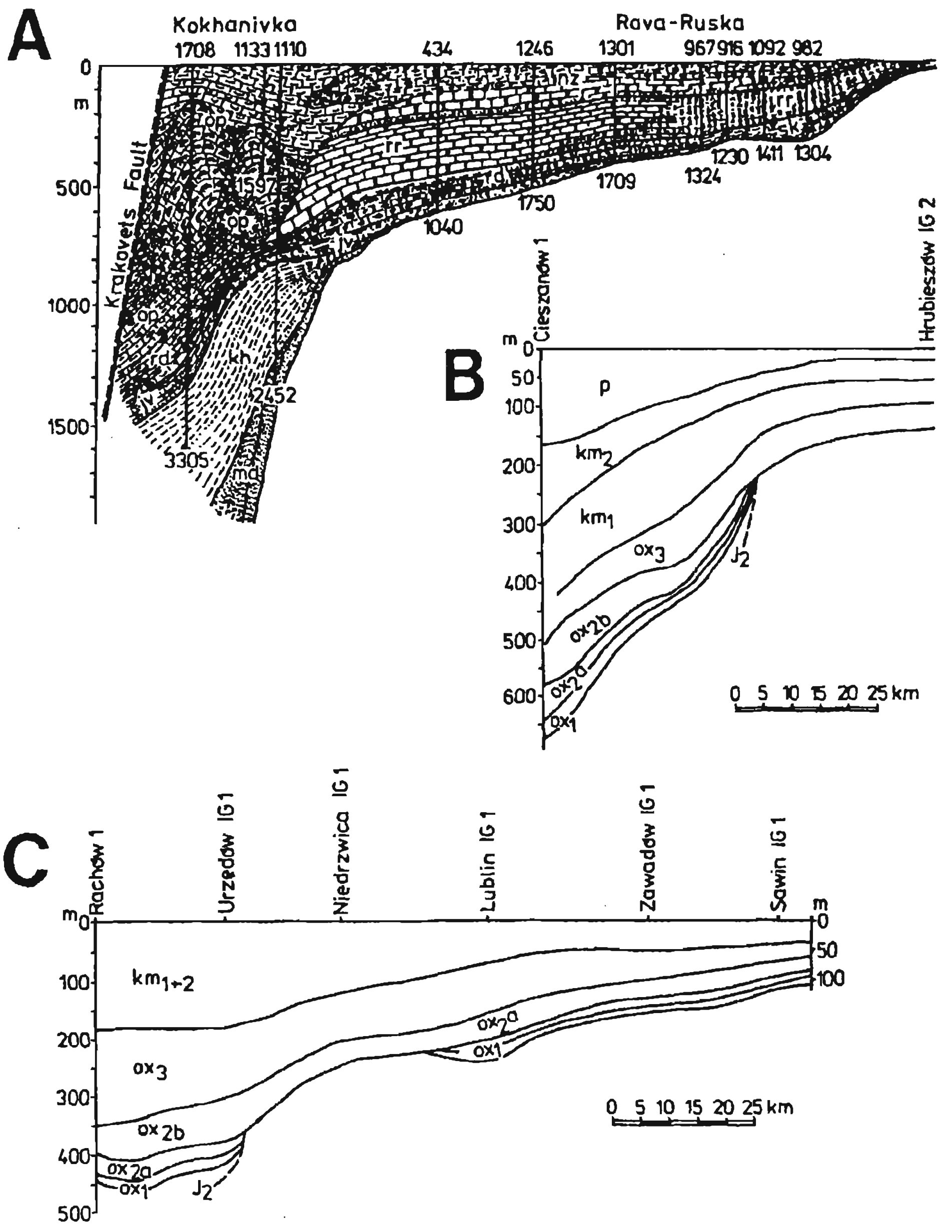
Fig. 13. Schematic cross-section showing the relationship of the unconformity developed at the base of the Upper Berriasian (and younger Cretaceous deposits in the south) to Lower Berriasian and end-Jurassic deposits in the Łódź-Mogilno Synclinorium; for other explanations see text

a result of the uplift of the Meta-Carpathian Arch hindering water circulation with, and then closing marine connection with the Tethys (KUTEK & *al.* 1989). It follows from the latter interpretation that the arch was subjected to uplift during an interval ranging from the Middle Volgian (Late Tithonian) to the Early Berriasian.

Some biogeographic data are also relevent to the discussed problem. The ammonites occurring in the upper Eudoxus Zone in the basal part of the Pałuki Formation still reveal Submediterranean affinities, but the ammonite faunas in the bulk of this formation (in the Autissiodorensis Zone of the Upper Kimmeridgian, in the Lower Volgian, and in the Middle Volgian up to the Regularis Horizon) are of Subboreal type (KUTEK & ZEISS 1974, 1994; KUTEK 1994). A peculiar Subboreal evolutionary lineage leading from *Discosphinctoides* and *Virgataxioceras* (Autissiodorensis Zone) through *Ilowaiskya* and *Pseudovirgatites* (Lower Volgian) to *Zaraiskites* (Scythicus Zone, Middle Volgian) can be recognized in the Pałuki Formation; and the representatives of *Amoeboceras* and *Aulacostephanus*, which are common in the Polish Autissiodorensis Zone, have also no counterparts in Submediterranean or Mediterranean regions. This is suggestive of a paleogeographic separation of the Polish part of the Central European Basin with regard to the Tethys in latest Kimmeridgian and Volgian time. On the other hand, however, Tethyan ammonites appearing in the Pałuki Formation at some levels up to the Middle Volgian indicate that some faunal migrations from the Tethys were still possible. It was suggested by the present Author (KUTEK 1990) that this incomplete faunal separation was a result of the filtering effect of a barrier formed by carbonate shoales of the Štramberk (=Stramberg) Limestone, that were developed in parts of the Carpathian Domain and adjacent regions of cratonic Europe in latest Kimmeridgian and Tithonian time. This poses the problem of the tectonic setting in which the sedimentation of the Štramberk carbonates took place, a problem that will be discussed at some length in another section of this paper.

THE JURASSIC EAST OF THE MID-POLISH ANTICLINORIUM

The Coquina Formation (the LUK Sequence) can be recognized in the north-eastern margin of the Holy Cross Mountains, and the underlying Upper Jurassic (Oxfordian and Lower Kimmeridgian, up to and including the



Hypselocyclum Zone), as well as Callovian deposits can be assigned to the COK Sequence. In that margin the Coquina Formation, with but a few exceptions (*e.g.* in the Annopol Anticline and the region of Iłża), is incomplete because of a base-Neocomian or base-Albian unconformity, and there are no reliable data concerning thicknesses of the COK Sequence. In the Annopol Anticline (in the Rachów borehole; *see* Text-fig. 14*C*), which is situated north of, and near the Holy Cross Lineament (Text-fig. 3), the combined thickness of the COK and LUK Sequences is about 500 m (NIEMCZYCKA 1976, GUTOWSKI 1992).

In most parts of the Lublin Synclinorium, and also in the western Ukraine, the COK and LUK Sequences cannot be recognized because of dramatic changes in facies (expressed *e.g.* by the appearance of dolomites, anhydrites and redbeds), and extreme paucity of reliable biostratigraphic data; in those regions, with but one exception (MORYC & WIŚNIOWSKA 1965), no ammonites seem to have been found above the Lower Oxfordian in Upper Jurassic deposits, which are almost exclusively known from boreholes (SANDLER 1969, NIEMCZYCKA 1976, MONKEIVICH & *al.* 1985). Nevertheless, it can be recognized with confidence that the thickness of Upper Jurassic deposits increases towards the Mid-Polish Anticlinorium, and that Middle Jurassic deposits appear, or dramatically increase in thickness at or near the eastern margin of the anticlinorium (Text-figs 14–15).

It is also evident that the thickness of Upper Jurassic deposits increases from the north-west to the south-east. In the Lublin Synclinorium, north of the Holy Cross Lineament, an Upper Jurassic succession composed of Oxfordian and almost complete Kimmeridgian deposits attains a thickness of about 400 m near the north-eastern margin of the Holy Cross Mountains in the region of Iłża (in the borehole Bąkowa-*IG-1*, NIEMCZYCKA 1976), and farther south, in the Annopol Anticline, as mentioned above, a succession consisting of Oxfordian, Lower Kimmeridgian and incomplete Upper Kimmeridgian deposits is *c.* 500 m thick (Text-fig. 14*C*). In the zone of the Holy Cross Lineament, still in Poland, the Upper Jurassic successions, including Tithonian ("Portlandian") deposits, attain thicknesses of over 650 m (Text-fig. 14*B*), and in the western

Fig. 14

A — Paleotectonic cross-section through Jurassic deposits of the western Ukraine, along line *4* in Text-figure 3 (*reproduced from* MONKEVICH & *al.* 1985)

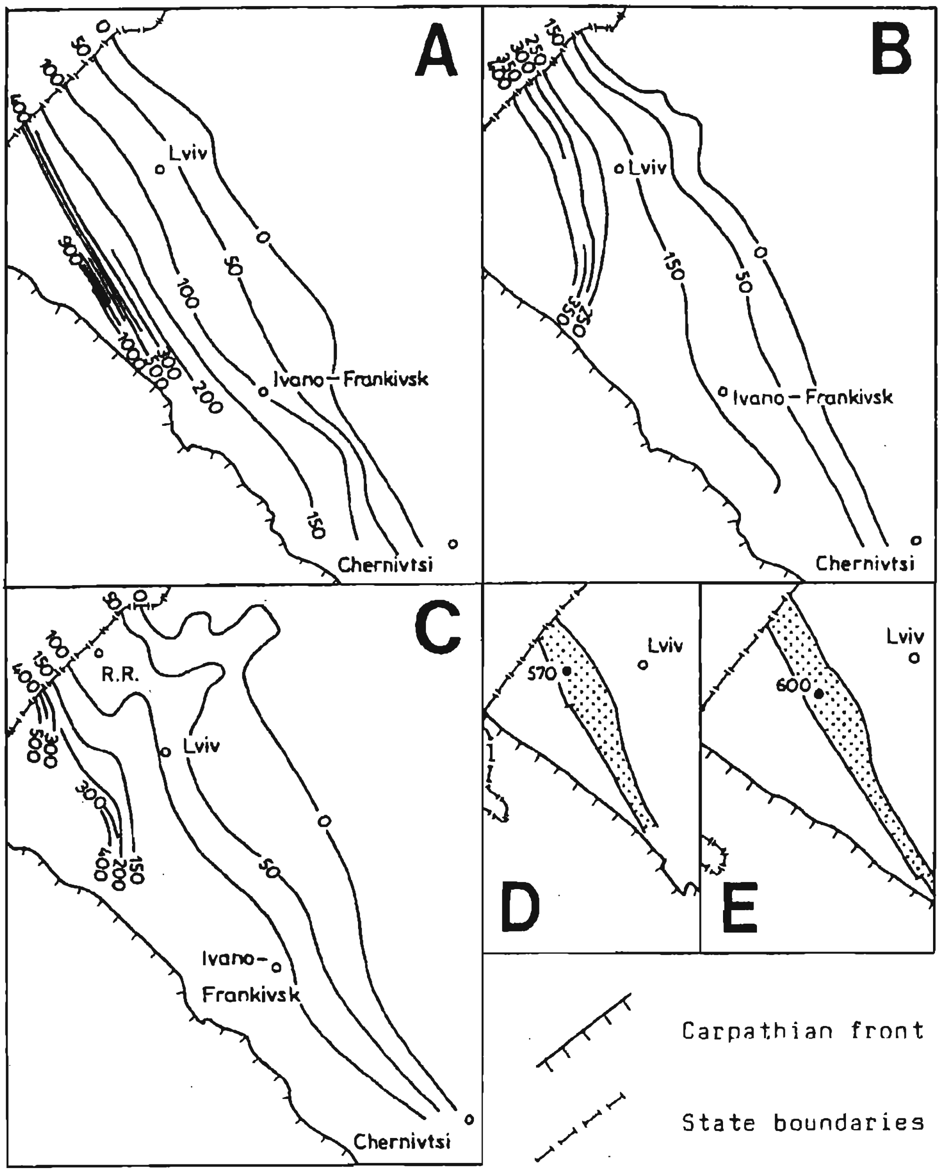
ABBREVIATIONS: **md** — Medinichi Formation, Toarcian-Aalenian; **kh** — Kokhanivka F., Bajocian-Bathonian; **jv** — Yavoriv F., Callovian; **rd** — Rudki F., Oxfordian; **sk** — Sokal F., Oxfordian; **rr** — Rava Ruska F., Kimmeridgian; **nz** — Nizhniv F., Tithonian; **op** — Oparsk F., Kimmeridgian and Tithonian

B-C — Paleotectonic cross-sections through Upper Jurassic deposits of the Lublin region, along line *5* (for *B*) and *6* (for *C*) in Text-figure 3 (*redrawn from* NIEMCZYCKA 1976)

ABBREVIATIONS: **ox** — Oxfordian, **km** — Kimmeridgian, **p** — "Portlandian"

Ukraine the thickness of Upper Jurassic successions, also composed of Oxfordian, Kimmeridgian and Tithonian deposits, amounts to over 1300 m (Text-figs 14*A* and 15*A*–*B*).

Another important point is that in the western Ukraine, on the eastern side of the Mid-Polish Anticlinorium, Toarcian and Middle Jurassic (Aalenian to Callovian) deposits attain thicknesses amounting to about or over 1000 m



(Text-figs 14*A* and 15*C*, *D*, *E*). The combined thickness of Middle Jurassic (Aalenian to Callovian) deposits is still remarkable just north of the Polish-Ukrainian border (DAYCZAK-CALIKOWSKA & MORYC 1988). This is in marked contrast with the western side of the Mid-Polish Anticlinorium near the Carpathians, where Middle Jurassic successions are often composed of Bathonian and Callovian deposits a few dozen meters thick at most, and also with most Middle Jurassic successions of the Lublin Synclinorium, which are usually very thin, and often exclusively composed of Callovian deposits (NIEMCZYCKA 1983, DAYCZAK-CALIKOWSKA & MORYC 1988). Remarkably thicker and more complete Middle Jurassic successions appear in the westernmost parts of the Lublin Synclinorium near, or at the north-eastern margin of the Holy Cross Mountains (thus, of the Mid-Polish Anticlinorium), which zone clearly corresponded in Middle Jurassic time to the north-eastern border of the Polish Rift (*comp.* Text-fig. 4).

In the cross-sections and maps (Text-figs 14—15) here reproduced from papers by SANDLER (1969), NIEMCZYCKA (1976) and MONKEVICH & *al.* (1985) the stratigraphic interpretations have been left unchanged. However, considerable errors can be involved in some of these interpretations, especially concerning the Upper Jurassic, and significant undetected stratigraphic hiatuses can be expected to occur in some parts of the sections. A minor correction could be introduced into the western part of the northernmost section (Text-fig. 14*C*), where the boundary between the "Lower" and "Upper" Kimmeridgian corresponds to the base of the Coquina Formation, and thus approximately to the boundary between the Lower Kimmeridgian Hypselocyclum and Divisum Zones. These stratigraphic uncertainties preclude a precise dating of some geologic events, but they cast no doubt on the existence in Middle and Late Jurassic time of faulted rift-borders clearly defined by the appearance, or dramatic increase in thickness, of Middle and Upper Jurassic deposits at or near the eastern margins of the Mid-Polish Anticlinorium.

Carbonates rich in nerineacean gastropods, which overlie deposits of undoubted Kimmeridgian age in the southernmost Lublin region, were assigned by NIEMCZYCKA (1976) to the Tithonian ("Portlandian"). The bulk of these carbonates most probably belong to the Tithonian, but their lowest part may be of latest Kimmeridgian age; in the NE margin of the Holy Cross Mts, near Iłża, a similar facies development is displayed by the Krzyżanowice Nerineacean Limestones, which directly overlie the Coquina Formation, and which Late Kimmeridgian age (Eudoxus or Autissiodorensis Zone) is firmly established by the presence of *Amoeboceras (Nannocardioceras)* cf. *anglicum* (SALFELD), reported by DĄBROWSKA (1957). No calpionellids could be found in the "Portlandian" carbonates of the Lublin region (NIEMCZYCKA, *pers. comm.*), which suggests that they do not range up into the Upper Tithonian. The same can be said about the Tithonian of the western Ukraine, where no calpionellids seem to have hitherto been recognized with certainty.

Fig. 15

A-B-C — Isopachs of Jurassic deposits of the western Ukraine (*simplified after* SANDLER 1969): A — Kimmeridgian and Tithonian, B — Callovian and Oxfordian, C — Bajocian and Bathonian

R.R. — Rava Ruska

D-E — Distribution of deposits of the Kokhanivka Formation, Bajocian-Bathonian (D) and Medinichi Formation, Toarcian-Aalenian (E) in the western Ukraine (*simplified after* MONKEVICH & *al.* 1985)

The Tithonian (and latest Kimmeridgian?) carbonates of the southern Lublin region and of the western Ukraine can be regarded as equivalent in facies, and to some extent also in age, to the shallow-water carbonates of the Štramberk Limestone of the Outer Carpathians.

ADDITIONAL REMARKS

The thicknesses of the COK, LUK and KVB Sequences are believed to provide a reliable measure of subsidence even without any corrections for bathymetry and compaction. With regard to bathymetry, no substantial errors are involved as all these sequences attain considerable thicknesses, and are topped by shallow-marine sediments. On the other hand, there are reasons to believe that the Oxfordian sediments in the studied areas, which are chiefly developed as limestones, were not subjected to considerable compaction. The same can be said about a part of the Kimmeridgian sediments. Admittedly, the shaly Pałuki Formation included in the KVB Sequence undoubtedly underwent strong compaction; however, none of the conclusions put forward in this paper depend on an estimation of the uncompacted thickness of that formation.

The following data indicate that the formation of the discussed sequences was chiefly controlled by tectonic factors, and not by eustasy. The sediments included in these sequences attain a remarkable thickness of over 1000 m, so that the accomodation space for these sediments could not have been produced by rise in sea-level alone. For instance, the Oxfordian sediments attain a thickness of about 700 m or more near the Carpathians, whereas the Oxfordian rise in sea-level, as estimated by HAQ & *al.* (1988), was less than 50 m. More significantly, a dramatic change of subsidence and uplift pattern took place in cratonic Poland in the Jurassic: the Meta-Carpathian Arch, which was still uplifted with respect to the Polish part of the Central European Basin in the early Middle Jurassic, was subsequently transformed into a region where Oxfordian and Kimmeridgian sediments attained their greatest thicknesses in Poland (*see* Text-figs 7 and 10–11).

A concomitant change in sediment source areas also took place. For instance, Liassic, but also Early Callovian gravels of southern provenance occur in southern cratonic Poland where carbonate sedimentation was established in the Oxfordian. By contrast, supply of terrigenic material continued in the Oxfordian from the Ukrainian Massif into adjacent regions of the western Ukraine and south-eastern Poland, and from the Baltic Shield into northern Poland (DADLEZ & KOPIK 1970, KUTEK & GŁAZEK 1972, SANDLER 1969, NIEMCZYCKA 1976, MONKEVICH & *al.* 1985).

THE PERI-CARPATHIAN SEGMENT OF THE POLISH RIFT BASIN

THE AGE OF THE PERI-CARPATHIAN RIFT SEGMENT

There is good evidence indicating that the existence of the segments of the Polish Rift situated in Central and northern Poland, north of the Holy Cross Lineament (*see* Text-figs 3—5), dates back to the Triassic or Late Permian (KUTEK & GŁAZEK 1972; POŻARYSKI & KUTEK 1976; POŻARYSKI & BROCHWICZ-LEWIŃSKI 1978, 1979; DADLEZ 1989, 1993; DADLEZ & POKORSKI 1993; DADLEZ & *al.* 1994). It can also be demonstrated that the area now occupied by the peri-Carpathian segment of the Mid-Polish Anticlinorium never constituted an individualized, uplifted tectonic or topographic unit in Permian and Mesozoic time (KUTEK & GŁAZEK 1972). On the other hand, there is evidence indicating that a rift-size graben did not yet exist south of the Holy Cross Lineament before the Middle Jurassic. This evidence, which was briefly outlined in an earlier paper by the present Author (KUTEK 1989), is presented below.

The Permian and Mesozoic sediments still preserved within the margins of the peri-Carpathian segment of the Mid-Polish Anticlinorium are not suggestive of a rift basin accomodating thick sedimentary successions of pre-Middle Jurassic age.

On the eastern side of that segment of the Mid-Polish Anticlinorium the Jurassic deposits rest directly on pre-Permian (usually early Paleozoic) rocks (POŻARYSKI 1979; *see also* Text-figs 2 and 14—15). In the western Ukraine the oldest Jurassic deposits are usually those included in the Medinichi Formation of Toarcian-Aalenian age (MONKEVICH & *al.* 1985). Locally, still older Jurassic sediments, to which a Hettangian-Pliensbachian age was attributed on palynological data, were encountered beneath the Medinichi Formation (MONKEVICH & *al.* 1985); this biostratigraphic interpretation, however, was not clearly confirmed in a later review edited by GARETSKY (1990). The earliest Jurassic deposits occurring in the southernmost Lublin region, just north of the Ukrainian border, were assigned to the Aalenian (DECZKOWSKI & FRANCZYK 1988, DAYCZAK-CALIKOWSKA & MORYC 1988).

In the western, corresponding margin of the Mid-Polish Anticlinorium (in that facing the Miechów Synclinorium), the Triassic, or Permian-Triassic successions display variable thicknesses which do not amount to more than a few hundred meters. Middle Jurassic successions, comprising only Bathonian and Callovian sediments in most sections, are usually extremely thin, and the highly incomplete Liassic sediments are preserved only in a few parts of that margin (KARNKOWSKI & OŁTUSZYK 1968, MORYC 1971, KUTEK & GŁAZEK 1972,

GIŻEJEWSKA 1974, WAGNER 1988, SZYPERKO-TELLER & MORYC 1988, GAJEWSKA 1988a, b, DECZKOWSKI & FRANCZYK 1988, DAYCZAK-CALIKOWSKA & MORYC 1988).

The south-western margin of the peri-Carpathian segment of the Mid-Polish Anticlinorium coincides with the Rzeszów Lineament (POŻARYSKI 1971, KUTEK & GŁAZEK 1972). A set of small Laramide anticlines, developed *en échelon* along the SW margin of the Holy Cross Mts, can be interpreted as a result of strike-slip movements in the basement along the Rzeszów Lineament (although the formation of these anticlines was probably also influenced by anisotropy induced in the pre-Permian basement by WNW trending Variscan structures). Narrow belts of Permian and Triassic sediments in the pre-Jurassic subcrop (*comp.* Text-fig. 16*A*), and relics of Liassic sediments, are aligned along the Rzeszów Lineament (KARNKOWSKI & OLTUSZYK 1968; PAWŁOWSKA 1962, 1970; MORYC 1971; KUTEK & GŁAZEK 1972; SZYPERKO-TELLER & MORYC 1988). This indicates an early age of the Rzeszów Lineament, and may be suggestive of the development of a narrow graben, or a half-graben, as early as in the Permian or Triassic. From this, however, it does not follow that a rift-size structure already existed in pre-Middle Jurassic time south of the Holy Cross Lineament, in the whole area now occupied by the peri-Carpathian segment of the Mid-Polish Anticlinorium.

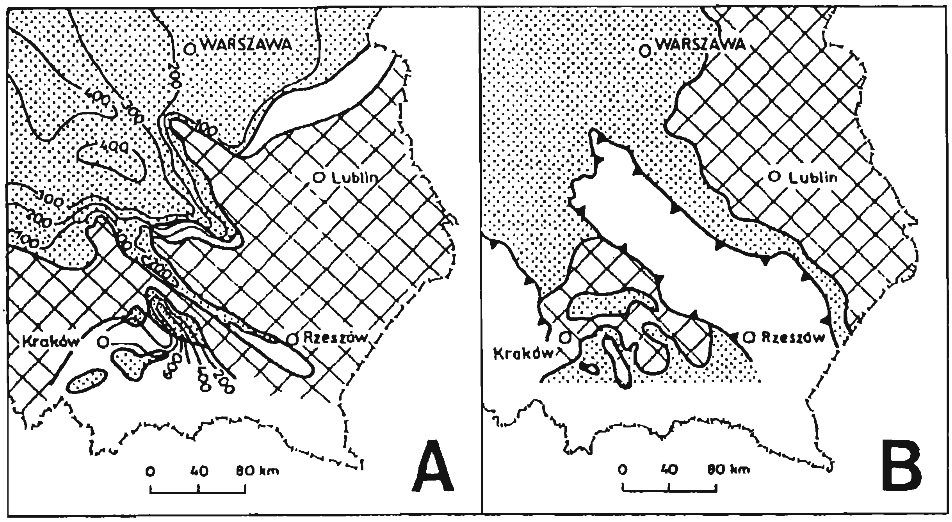


Fig. 16

A — Map of paleothicknesses of the Lower Buntsandstein in SE cratonic Poland (*simplified after* SZYPERKO-TELLER & MORYC 1988)

Stippled are areas with still preserved deposits, cross-hachured areas with non-deposition of Lower Buntsandstein

B — Distribution of Lower Bathonian deposits in SE cratonic Poland (*modified after* DAYCZAK-CALIKOWSKA & MORYC 1988)

Stippled are areas with preserved, cross-hachured areas devoid of, Lower Bathonian deposits; barbed lines outline areas of post-Cretaceous erosion

In this context, significant are also the highly irregular or patchy distribution and thickness patterns, displayed by Permian, Triassic, Liassic and Middle Jurassic sediments in the Miechów Synclinorium (KARNKOWSKI & OŁTUSZYK 1968, MORYC 1971, MORAWSKA 1979, HAKENBERG 1980, WAGNER 1988, SZYPERKO-TELLER & MORYC 1988, GAJEWSKA 1988a, b, DECZKOWSKI & FRANCZYK 1988, DAYCZAK-CALIKOWSKA & MORYC 1988). In marked contrast to the rift-related pattern revealed by Upper Jurassic deposits, the former patterns (*see* Text-figs 8 and 16) do not indicate an early existence of the peri-Carpathian segment of the Polish Rift, and they are suggestive of diffused (non-polarized) crustal extention. Especially significant from this point of view are several small depocenters of the Buntsandstein, fault-bounded in part, extending in different directions and accomodating thicknesses well over 500 m (KARNKOWSKI & OŁTUSZYK 1968, MORYC 1971, SZYPERKO-TELLER & MORYC 1988); they are but indistinctly indicated in the small-scale map reproduced in Text-figure 16*A*.

Significant is also the development of the area of the Lublin Synclinorium, situated north of the eastern part of the peri-Carpathian segment of the Mid-Polish Anticlinorium (Text-figs 2–3, 9, 14 and 16). The Upper Jurassic and Cretaceous deposits preserved in this synclinorium display thickness patterns parallel or subparallel to the Mid-Polish Anticlinorium, but most parts of this region are devoid of Permian, Triassic, Liassic and Middle Jurassic (pre-Bathonian) sediments (NIEMCZYCKA 1976, ŻELICHOWSKI & KOZŁOWSKI 1983, WAGNER 1988, SZYPERKO-TELLER & MORYC 1988, GAJEWSKA 1988a, b, DECZKOWSKI & FRANCZYK 1988, DAYCZAK-CALIKOWSKA & MORYC 1988).

From these data it can be concluded that a distinct rift-size structure had not been developed south of the Holy Cross Lineament before the Middle Jurassic. Thus, the question arises when such a structure began to develop. The following data are relevant to this question.

The dramatic increase of the thicknesses of Upper Jurassic deposits in the Ukrainian margin of the Mid-Polish Anticlinorium (*see* Text-figs 14*A* and 15*A*, *B*), and the clearly rift-related thickness patterns of the Upper Jurassic deposits in the Lviv and Miechów Synclinoria (Text-figs 9, 14, 15*A*, *B* and 17), unequivocally indicate that a rift basin was already well developed south of the Holy Cross Lineament in the Late Jurassic. More precise stratigraphic data (*see* Text-figs 14*A*, *B* and 17) indicate that it was so already in the Oxfordian. As previously remarked, the small thicknesses of the deepening-upwards Callovian sediments in southern Poland, which were strongly influenced by topographic variation in some regions, do not provide a good measure of subsidence. However, these sediments together with the Oxfordian ones belong to the same trasgressive-regressive tectonic cycle. This suggests that a rift basin existed south of the Holy Cross Lineament already in the Callovian.

A pronounced rift-border is clearly defined in the marginal zone of the Mid-Polish Anticlinorium in the western Ukraine by thick Middle Jurassic (and Toarcian) sediments that rest directly on Paleozoic rocks (*see* Text-figs 14 and 15*C*, *D*, *E*). The Ukrainian data also indicate that this border commenced to develop during the accumulation of the Medinichi Formation of Toarcian-Aalenian age (*see* Text-figs 14*A* and 15*E*). On the other hand, the Middle Jurassic successions are very thin and incomplete, in the western margin of the Mid-Polish Anticlinorium adjacent to the Miechów Synclinorium, where even the distribution pattern of Bathonian sediments (*see* Text-fig. 16*B*) is not suggestive of the existence of a rift in the close vicinity. From all these data it can be concluded that the peri-Carpathian segment of the Polish Rift Basin developed successively from the east (from the Ukrainian side) to the west, and that a westward spreading of subsidence and faulting from the initial rift zone allowed the widening of this segment during Middle Jurassic and Oxfordian time (*see* Text-fig. 18).

ASYMMETRY OF THE PERI-CARPATHIAN SEGMENT OF THE POLISH RIFT BASIN

This asymmetry is expressed by the fact that Toarcian and Middle Jurassic deposits, about 1000 m thick, are followed by Upper Jurassic (Oxfordian to Tithonian) deposits, well over 1000 m thick (Text-fig. 14*A*), in the Ukrainian margin of the peri-Carpathian segment of the Mid-Polish Anticlinorium, whereas in its corresponding western margin the Upper Jurassic (Oxfordian to Upper Kimmeridgian) deposits, up to 1100 or 1300 m thick, are underlain by very thin Middle Jurassic sediments; however, the Oxfordian deposits appear to be thicker on the western side of that anticlinorium (*comp.* Text-figs 10, 14*A* and 17). Furthermore, good evidence for a prominent down-to-the-west fault-zone is provided by the Ukrainian margin of the Mid-Polish Anticlinorium (*see* Text-figs 14*A* and 15), whereas on its western side no corresponding symmetric (down-to-the-east) fault zone of similar magnitude can be recognized.

Laramide folding and paucity of deep boreholes, as well as the scarcity of continuous Jurassic sections not covered by Quaternary sediments in the SW margin of the Holy Cross Mts, make it difficult to assess syndepositional faulting in the zone of the Rzeszów Lineament (corresponding to the SW margin of the Mid-Polish Anticlinorium, and including the SW Mesozoic margin of the Holy Cross Mts). Some outcrop data from the SW margin of the Holy Cross Mts (KUTEK 1968, KUTEK & GŁAZEK 1972, MATYJA 1977), combined with those from the deep borehole Brzegi-*IG-1* (WIERZBOWSKI & KUTEK 1990), are suggestive of a syndepositional fault active in the Oxfordian, but its possible magnitude may only be of the order of 100–150 m. Despite the fact that the Upper Jurassic deposits are over 1000 m thick in the SW margin of the peri-Carpathian segment of the Mid-Polish Anticlinorium, no available data pertinent to this margin are suggestive of Middle or Late Jurassic syndepositional faulting of a magnitude comparable with that displayed by the Ukrainian rift-border.

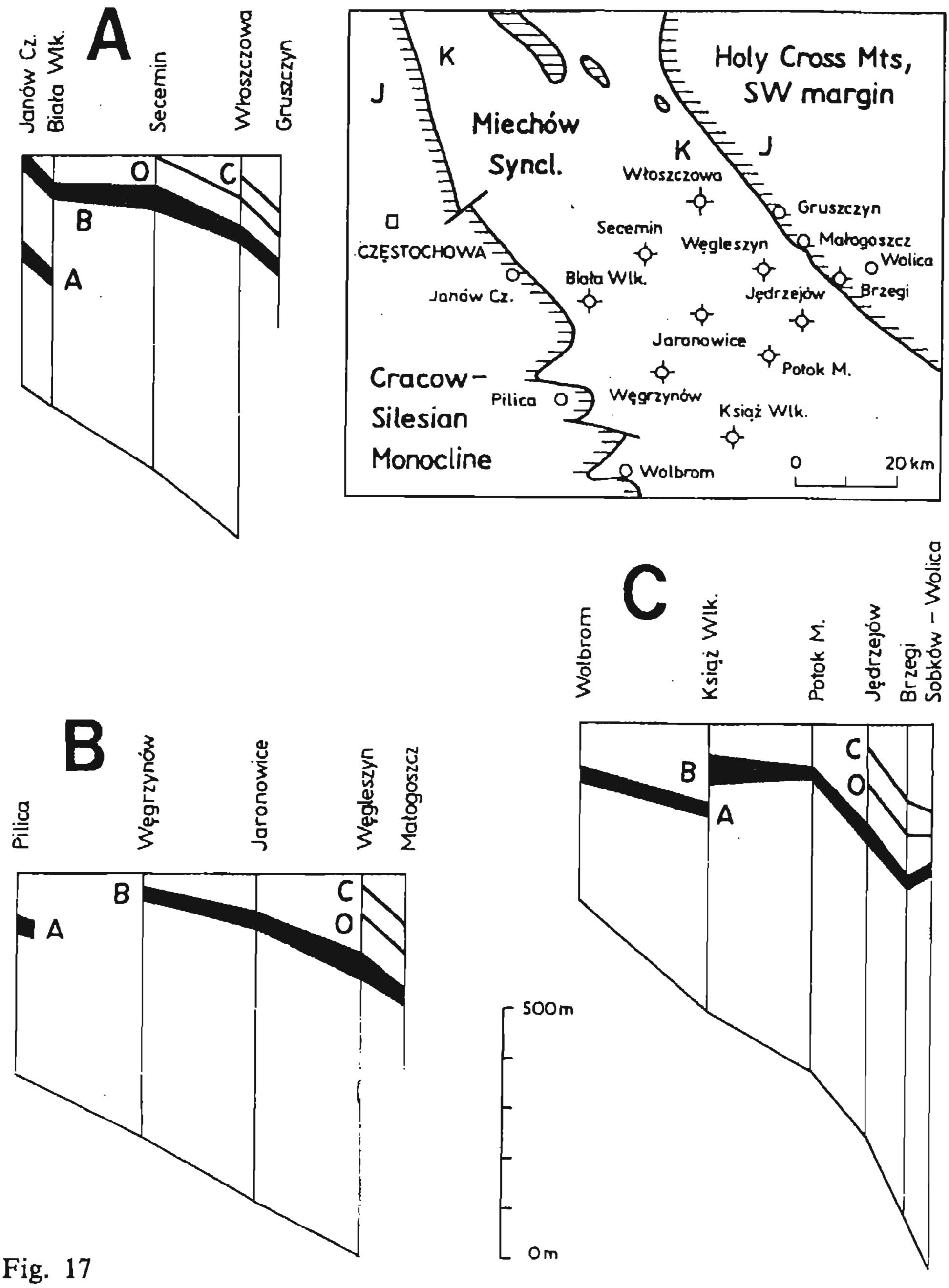


Fig. 17

A-C — Cross-sections through Upper Jurassic deposits in the middle and northern parts of the Miechów Synclinorium (*see* inserted map; *simplified after* Wierzbowski & Kutek 1990); marker horizons *A* and *B* are indicated; O — main onkolite (*see* Kutek 1968); C — base of the Coquina Formation

Data from boreholes in the middle and northern parts of the Miechów Synclinorium, combined with biostratigraphic data from outcrops of the Cracow-Silesian Monocline and the SW margin of the Holy Cross Mts (WIERZBOWSKI & KUTEK 1990), allow to recognize that the thickness increase of Upper Jurassic deposits in that synclinorium towards the Mid-Polish Anticlinorium is due in part to the younging of these deposits to the north-east, but also to the thickness increase of Oxfordian deposits (*see* Text-fig. 17). The latter increase is roughly gradual, which does not preclude the existence of syndepositional faults between the widely spaced boreholes, but which would hardly be consistent with a concept of magnitude of such faults of the order of several hundred meters. Roughly gradual increase of Oxfordian sediments to the east or north-east can also be recognized in the south-western part of the Miechów Synclinorium, where a down-to-the-SW fault of a magnitude of about 100 m, affecting Oxfordian and Lower Kimmeridgian deposits, but probably of post-Jurassic age, was recognized by JAWOR (1980).

Syndepositional faults with magnitudes to about 100 m, and different orientation (down-to-the-SW, but also down-to-the-north) could be recognized in Oxfordian deposits of the Cracow-Silesian Monocline (KUTEK & *al.* 1990), *e.g.* east of Częstochowa and west of Wolbrom (*see* Text-fig. 17*D*). Finally, fairly large tilted blocks, comprising Carboniferous, Permian, Triassic and Jurassic deposits, and bounded by down-to-the-SW faults, are developed in the Cracow-Silesian Monocline, and in the eastern part of the Upper Silesian Coal Basin. These structures can easily be recognized in the map of DOKTOROWICZ-HREBNICKI (1960), but their exact age is unknown; they may largely be a result of post-Jurassic tectonic events.

The data outlined above allow to conclude that the peri-Carpathian segment of the Polish Rift Basin developed in Middle and Upper Jurassic time approximately as a half-graben, with the largest faults, and the thickest Middle Jurassic syn-rift sediments at, or near its eastern, that is the Ukrainian border.

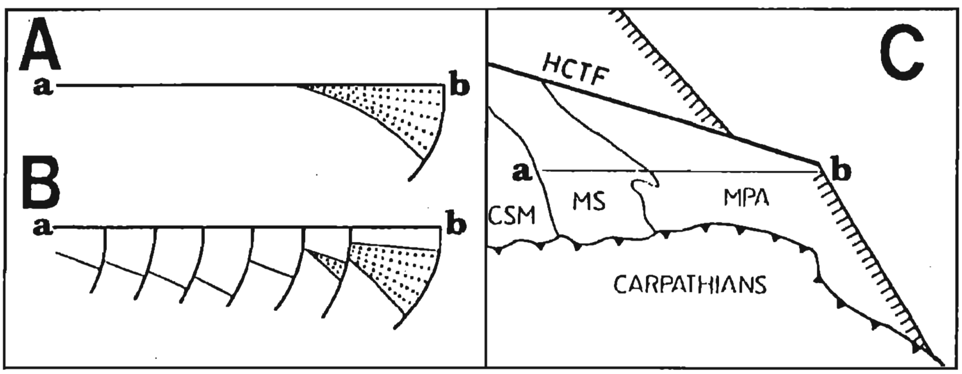


Fig. 18

A-B — Paleotectonic cross-sections through the peri-Carpathian segment of the Polish Rift Basin, approximately at the end of the Bathonian (A) and Oxfordian (B); stippled are thick accumulations of Middle Jurassic (and Toarcian) deposits

C — Map outlining the peri-Carpathian segment of the Polish Rift Basin between the Holy Cross Transfer Fault (*HCTF*) and the Carpathian front

Indicated are the pronounced eastern borders of the Polish Rift (*barbed lines*) north and south of the Holy Cross Transfer Fault, and the present-day boundaries between the Mid-Polish Anticlinorium (*MPA*), the Miechów Synclinorium (*MS*), and the Cracow-Silesian Monocline (*CSM*); **a-b** — section line of *A* and *B*

This half-graben included areas now occupied by the Mid-Polish Anticlinorium and the Miechów Synclinorium, and possibly also parts of the Cracow-Silesian Monocline (*see* Text-fig. 18).

The primary purpose of the cross-sections and the map presented herein (Text-fig. 18) is to show the asymmetric shape of the peri-Carpathian segment of the Polish Rift Basin, and its general tectonic setting. The detailed interpretations involved in these sections have been inspired by the model of asymmetric rifting of WERNICKE (1985, Fig. 3). This model can successfully explain the great thicknesses of Middle Jurassic syn-rift sediments at the Ukrainian rift border, strongly contrasted with the extremely reduced Middle Jurassic successions of the Miechów Synclinorium, and also the fact that Oxfordian sediments attain thicknesses on the western side of the Mid-Polish Anticlinorium, markedly greater than on its Ukrainian side. However, these are but tentative interpretations, as Jurassic deposits have largely been removed by erosion from the peri-Carpathian segment of the Mid-Polish Anticlinorium, and as little is known about Jurassic syndepositional faulting in the areas situated farther to the west. Note also that the thin Callovian and Oxfordian sediments occurring east of the Ukrainian faulted border of the rift-graben (*see* Text-fig. 14*A*) have not been indicated, and thus not interpreted, in the cross-section shown in Text-figure 18*B*.

THE HOLY CROSS LINEAMENT AS A TRANSFER FAULT

The Holy Cross Lineament, that separated the peri-Carpathian segment of the Polish Rift Basin from its portions situated farther north, coincides with a discontinuity in the pre-Permian basement, which is marked by the boundary of the Paleozoic Kielce and Łysogóry Zones in the Holy Cross Mountains. Moreover, this lineament is expressed as a rather narrow zone in the shape of the Mid-Polish Anticlinorium (Text-figs 2–3), and in the gradient zone of the Permo-Mesozoic thickness pattern at the junction of the Miechów and Łódź-Mogilno Synclinoria (*see* Text-fig. 8). Therefore, the Holy Cross Lineament should be regarded as a transfer fault separating rift segments, rather than a diffuse accomodation zone (*comp.* ROSENDAHL 1989, ROSENDAHL & *al.* 1992). The latter designation, however, seems to be applicable to the zones that separate from each other the segments of the Polish Rift Basin developed north of the Holy Cross Lineament (*see* DADLEZ 1994). The Holy Cross Lineament, when acting as a transfer fault, will be referred to as the Holy Cross (Mountains) Transfer Fault.

The easternmost part of the peri-Carpathian segment of the Polish Rift Basin was bounded to the north by the easternmost part of the Holy Cross Lineament (the Holy Cross Transfer Fault), situated east of the Holy Cross Mts (*see* Text-figs 2–3 and 18). Hence, it can be expected that Jurassic and

Cretaceous sediments, accumulated in the eastern part of the peri-Carpathian segment of the Polish Rift, were thicker than farther north across the Holy Cross Transfer Fault, in the southern part of the Lublin region. The thickness and distribution patterns of Upper Jurassic and Cretaceous deposits of this region (NIEMCZYCKA 1976, 1983; MAREK 1983b; KRASSOWSKA 1983) appear to confirm this supposition.

Jurassic deposits crop out in the north-eastern and south-western Mesozoic margins of the Holy Cross Mts; hence, these deposits were usually compared as concerns Jurassic facies and paleogeography (*e.g.* KUTEK 1969, MATYJA & *al.* 1989). However, these margins are situated north and south of the Holy Cross Lineament (the Holy Cross Transfer Fault), and therefore belong to different segments of the Polish Rift basin (*see* Text-figs 2–3 and 18). Hence, such a comparison may be misleading in a tectonic context. For instance, Upper Jurassic sediments attain a thickness of about 500 m in the north-eastern, and of about 1100 m in the south-western margin of the Holy Cross Mts, but the thickness of Upper Jurassic sediments is still greater, than in the latter margin, in the corresponding eastern margin of the peri-Carpathian segment of the Mid-Polish Anticlinorium, in the western Ukraine.

The Mid-Polish Anticlinorium extends in a characteristic fashion eastwards south of the Holy Cross Lineament, and the eastern margins of the Mid-Polish Anticlinorium, situated north and south of this lineament, correspond to distinct rift borders (*see* Text-figs 2–4, 14–15 and 18). This means a corresponding eastward extention of the peri-Carpathian segment of the Polish Rift Basin, which thus occupied an area broader than the segment of this basin situated north of the Holy Cross Lineament. This, in turn, suggests that the development of the Polish Rift Basin was connected with stronger crustal extention south, than north, of the Holy Cross Transfer Fault. In this context it is pertinent to note that backstripping procedures applied to Upper Jurassic (Oxfordian-Kimmeridgian) sediments that accumulated in the Polish Rift Basin north of the Holy Cross Transfer Fault, also suggest an increase of crustal extention to the south (*see* DADLEZ & *al.* 1994b).

As the Polish Rift broadens south of the Holy Cross Transfer Fault, it can be expected that it acted as a strike-slip fault in Jurassic, and possibly also in Cretaceous time. Consequently, some structural features in the Holy Cross Mountains of uncertain age, which were interpreted as a manifestation of Variscan strike-slip movements (POŻARYSKI & *al.* 1992, POŻARYSKI & TOMCZYK 1993), can be reinterpreted as an expression of Mesozoic strike-slip movements.

Deep seismic soundings allowed to recognize that in the Holy Cross Mountains the thickness of the Earth crust decreases abruptly southwards across the Holy Cross Lineament, in two seismic profiles, from 43-45 to 36-38 km, and from 52 to 44 km (for details *see* GUTERCH & *al.* 1976, 1984, 1986a,b; GRAD 1980; DADLEZ & *al.* 1994a). This difference in crust thickness was usually interpreted as due to pre-Mesozoic tectonic events. However, as summarized by ZIEGLER (1990), the position of the Moho, as recognized in Western and Central Europe, usually reveals much better correspondence with post-Variscan, than with earlier, tectonic events. Hence, it can be concluded that the decrease in thickness of the crust south of the Holy Cross Lineament was, in part at least, a result of stronger Mesozoic crustal extention south, than north, of this lineament during the formation of the Polish Rift Basin.

It is pertinent to note that deep seismic soundings revealed a thinner crust beneath the Miechów Synclinorium, and the western part of the peri-Carpathian segment of the Mid-Polish Anticlinorium, than beneath its eastern part, straddling the border of Poland and the Ukraine (*comp.* GUTERCH & *al.* 1984, Figs 3, 10–11; 1986a, Figs 12–13; 1986b, Fig. 32; 1991, Fig. 2/VIII). This is of interest, as models of asymmetric rifting assuming simple shear on lithospheric scale predict a shift of zones of greatest crustal attenuation with regard to axial or proximal (initial) zones of rift grabens (WERNICKE 1985, COWARD 1986).

THE SOUTH-EASTERN EXTENTION OF THE POLISH RIFT BASIN

The Mid-Polish Anticlinorium plunges beneath the Carpathian nappes in south-eastern Poland and the western Ukraine (*see* Text-fig. 2). This suggests that the Polish Rift Basin, which gave rise to this anticlinorium, extended southwards into the Carpathian Domain in Jurassic and Cretaceous time.

The problem of the extention of the Polish Rift (aulacogen) into the Carpathian Domain ("geosyncline") was discussed by POŻARYSKI & ŻYTKO (1981). According to these authors, an extention of the Polish Rift Basin can be sought in the Outer Dacidian Trough (Rift), recognized in the Carpathians of Romania (SANDULESCU 1984, 1988, 1989a, b, 1990). This trough began to develop in an intracontinental setting, and was underlain by strongly attenuated continental crust; it was separated to the west from the oceanic domain of the Transylvanides by a strip of a relatively thick continental crust. The Outer Dacidian Trough was subjected to Laramide orogenic tectonism. This might coincide with the Laramide inversion of the Polish Rift Basin, that gave rise to the Mid-Polish Anticlinorium. The Polish Rift Basin, however, was underlain by thicker continental crust and, as a consequence, the Mid-Polish Anticlinorium is an intraplate tectogene of non-orogenic type.

A different correlation of the tectonic zones of Romania and those of the cratonic foreland of the Polish Carpathians was proposed by SANDULESCU (1984). In this context, however, it is worth of note that geological data from Romania (POŻARYSKI & ŻYTKO 1981; SANDULESCU 1984, 1989a) contradict the concept of a continuous, post-Variscan "Danish-Polish-Dobrudjan aulacogene" of GŁAZEK & *al.* (1973).

In any case, the broadening of the Polish Rift Basin, and the slight attenuation of the continental crust, south of the Holy Cross Transfer Fault, is consistent with the concept of a southward extention of the Polish Rift basin into the Carpathian Domain, which parts were underlain by oceanic or attenuated continental crust. In this context it is to note that a set of Laramide anticlines (indicated in a generalized fashion in Text-fig. 2) is developed in the south-eastern corner of the Miechów Synclinorium, extending the Mid-Polish Anticlinorium to the west. This is perhaps also an indirect manifestation of the broadening of the Polish Rift Basin to the south.

According to the interpretation put forward in this paper, the Polish Rift abuted against the Holy Cross Lineament, and did not extend farther south, before the Middle Jurassic (or Toarcian). Thus, it was a "blind", and not a "penetrating" rift (*see* MILANOVSKI 1992). This was a situation reminiscent of what is now found in the North Sea area, where the North Sea Rift System does not extend into mainland Europe. From this, however, it does not follow that the formation of the peri-Carpathian segment of the Polish Rift was, from a genetic point of view, the result of a southward propagation of this rift across the Holy Cross Lineament. More probably, this segment was formed as a result of a northward propagation of a rift from the Carpathian Domain into the cratonic areas of the western Ukraine and south-eastern Poland. This interpretation implies that the latter rift joined the Polish Rift, and amalgamated with it along the Holy Cross Lineament, in Middle Jurassic time.

In this context it is significant that the faulted rift-border developed in the western Ukraine does not extend north of the Holy Cross Lineament, where the faulted eastern border of the Polish Rift is situated farther west (*see* Text-figs 2–4, 14–15 and 18), and that the cratonic area of the western Ukraine was uplifted before its inclusion in the peri-Carpathian segment of the Polish Rift Basin in Toarcian and Middle Jurassic time.

The discussed interpretation is also consistent with the early commencement of rifting, in pre-Middle Jurassic time, in the Carpathian Domain within the northern margin of the Tethys (BIRKENMAJER 1986, 1987; SANDULESCU 1990).

A SUMMARY AND CONCLUDING REMARKS

RELATIONSHIPS TO THE CARPATHIAN DOMAIN IN MIDDLE JURASSIC TO KIMMERIDGIAN TIME

The strong downwarping of the Meta-Carpathian Arch, that commenced in the Middle Jurassic and continued in the Late Jurassic (Oxfordian and Kimmeridgian), resulted in the establishment of a broad marine connection across Poland between the Carpathian Tethys and the Central European Basin, and most parts of cratonic Poland were then transformed into what can be called a shelf-embayment of the Tethys. As a consequence, extensive carbonate successions of Oxfordian and Kimmeridgian age developed in cratonic Poland, including the characteristic peri-Tethyan sponge megafacies; and, as previously remarked, strong Mediterranean or Submediterranean influences are expressed by the Callovian, Oxfordian and Kimmeridgian ammonite faunas of cratonic Poland.

The foundering of peri-Tethyan regions of cratonic Europe, which ceased to act as sediment source areas, resulted in establishment of clastic-starved conditions, and the development of condensed successions at the Callovian/Oxfordian junction, in southern cratonic Poland. Subsequently, continuing subsidence, coupled with increasing rates of carbonate sedimentation, resulted in the deposition of thickening upwards successions of the Oxfordian Stage. In the Miechów Synclinorium, the Cracow Monocline and the SW margin of the Holy Cross Mts, the Lower Oxfordian is usually a few meters, the Middle Oxfordian about 100 m, and the Upper Oxfordian a few hundred meters thick (RÓŻYCKI 1953; KUTEK 1968; KUTEK & *al.* 1977, 1990; MATYJA 1977; WIERZBOWSKI & KUTEK 1990). This is an example of what was called by CALLOMON (1964) the Oxfordian Tilt. This designation refers to the fact that in several peri-Tethyan regions, *e.g.* in south Germany and in southern Poland, much reduced early Oxfordian sediments are followed by markedly thicker younger sediments of the Oxfordian Stage, whereas farther north in Europe, *e.g.* in England, and also in Pomerania in Poland, expanded successions straddling the Callovian/Oxfordian junction are followed by mederately thick, or incomplete higher Oxfordian sediments (*see* DEMBOWSKA & MALINOWSKA 1976, DAYCZAK-CALIKOWSKA 1977).

The foundering of the cratonic area of south-eastern Poland enabled the development of the COK and LUK Sequences, with thicknesses increasing towards the Carpathian Domain; the strongest subsidence took place in the Oxfordian, as indicated by the Oxfordian deposits of the COK Sequence, which thicknesses amounted to about 700 m at least. On the other hand, the Toarcian — Middle Jurassic succession that is developed in the Ukrainian border-zone of the Polish Rift (*see* Text-figs 14*A* and 15*C*, *D*), includes marine shales of Bajocian and Bathonian age, which attain a thickness of at least 570 m (MONKEVICH & *al.* 1985). Significantly, the combined age of all these deposits coincides with that of the radiolarian shales deposited in the Carpathian Domain. In several successions of the Carpathians (and also in those of the Austroalpine Zone of the Alps), the occurrences of radiolarian cherts are restricted to an interval ranging from the Callovian to the Kimmeridgian, or only to the Oxfordian; moreover, in some successions of the Pieniny Klippen Belt, and in the Križna Succession of the Tatra Mountains, a Bajocian and/or Bathonian age was attributed with more or less confidence to the lowest occurrences of radiolarian cherts (BIRKENMAJER 1977, LEFELD 1988).

The Carpathian occurrences of radiolarian cherts are indicative or suggestive of deepening of depositional zones, but indirectly also of stretching of continental crust (and development of oceanic crust?), and thus of an extentional tectonic regime. In this context, the foundering of the Meta-Carpathian segment of the Polish Rift Basin can be interpreted as a response of the cratonic areas of south-eastern Poland and the western Ukraine to extentional tectonics in the Carpathian Domain.

The peri-Carpathian segment of the Polish Rift Basin occupied an area thrown down south and south-west of the Holy Cross Transfer Fault, and the Ukrainian faulted rift-border (Text-fig. 18). Hence, in terms of the old geosynclinal concept, the formation of this segment could be interpreted as a failed attempt to include that area into the Carpathian geosyncline. However, successful rifts did develop farther south in the Carpathian Domain in the Jurassic (BIRKENMAJER 1986, 1988), marking the commencement of a set of tectonic events leading to the transformation of this domain that culminated in the orogeny of the Outer Carpathians in the Tertiary.

The formation of the peri-Carpathian segment of the Polish Rift Basin also coincides with what was called by ZIEGLER (1982, 1990) the polarization of the European rift system, initiated in the Middle Jurassic.

RELATIONSHIPS TO THE CARPATHIAN DOMAIN IN KIMMERIDGIAN TO EARLY BERRIASIAN TIME

The KVB Sequence of cratonic Poland, which consists of the Late Kimmeridgian to Middle Volgian (mid-Upper Tithonian) Pałuki Formation, and the Middle Volgian to Early Berriasian Kcynia Formation, has its base located in the Eudoxus Zone (*see* Text-fig. 1). Interestingly, in south-eastern France a boundary between sequences of "tectono-eustatic" nature, *i.e.* controlled by changes in basin volume and physiography, is also located in the Eudoxus Zone (DROMART & *al.* 1993). In the North Sea area a major transgressive pulse, coupled with the onset of Late Cimmerian tectonism, also corresponds to the Eudoxus Zone (BOLDY & BREALEY 1990).

Two lithostratigraphic units of the Outer Carpathians of Poland and Moravia are pertinent to the problem here discussed. These are the Cieszyn Beds (the oldest beds involved in the flysch nappes of the Polish Outer Carpathians), and the Štramberk Limestone (a collective name for shallow-marine carbonates of latest Jurassic — earliest Cretaceous age).

The Cieszyn Beds are subdivided into the Lower Cieszyn Shales, the Cieszyn Limestones, and the Upper Cieszyn Shales. The Lower Cieszyn Shales, in which slumps, graded bedding and large quantities of blocks of the Štramberk Limestone appear up-section, are sometimes classified as preflysch;

the Cieszyn Limestones display features of a carbonate flysch (PESZAT 1967, KSIĄŻKIEWICZ 1971, NOWAK 1976, ŚLĄCZKA 1976). The Cieszyn Limestones were assigned to the Tithonian and Berriasian, whereas the Lower Cieszyn Shales to the Upper Kimmeridgian and Tithonian (NOWAK 1976). However, a Late Kimmeridgian age of the lower portion of the Lower Cieszyn Shales was indicated only by aptychi and microfossils, and the stratigraphic position of their (detached) base could not be recognized with desirable precision in terms of ammonite zonation.

The bulk of the carbonates of the Štramberk Limestone, which are chiefly known from exotic material, is of Tithonian age, but they extend up into the Berriasian (GEROCH & MORYCOWA 1966, OLÓRIZ & TAVERA 1982). An exotic block of the Štramberk Limestone, containing *Hybonoticeras hybonotum* (OPPEL) and found in the Lower Cieszyn Shales (NOWAK 1976), indicates that the Štramberk Limestone ranges down at least to the Hybonotum Zone (the lowest zone of the Tithonian); the same conclusion was reached independently by OLÓRIZ & TAVERA (1982). However, the exact age of the base of the Štramberk Limestone remains unknown (HOUŠA 1977). It was pointed out by the present Author (KUTEK 1990) that the development of a Subboreal ammonite fauna in cratonic Poland in the latest Kimmeridgian and Volgian may be explained as an effect of a barrier formed by shallow-water carbonates of the Štramberk Limestone at the southern margin of cratonic Europe. As that fauna is already clearly individualized in the Autissiodorensis Zone (KUTEK & ZEISS 1994), this interpretation implies that the Štramberk Limestone ranges down at least to a level corresponding to the base of the Autissiodorensis Zone. A still lower position of the base of the Štramberk carbonates within the Eudoxus Zone is quite possible, but this supposition cannot be confirmed by available biostratigraphic evidence.

The age of the base of the Štramberk carbonates is of special interest because there are indications that they were transgressive in several parts of the domain of the Outer Carpathians (KSIĄŻKIEWICZ 1956, MIŠIK 1974). In Central Poland, as previously stated, evidence for transgressive tendencies in the latest Kimmeridgian is provided by the Pałuki Formation, and by the Krzyżanowice Nerineacean Limestones of the north-eastern margin of the Holy Cross Mountains. Unfortunately, there are no precise data concerning the nature and age of the base of the Štramberk-type carbonates developed in the western Ukraine, and in the southern Lublin region in Poland. However, it can be suggested that the sediments with latest Kimmeridgian ammonites encountered in one borehole near the Polish—Ukrainian border (MORYC & WIŚNIOWSKA 1965) are indicative of a transgressive pulse affecting those regions.

In any case, the stratigraphic range of the Štramberk Limestone, and the combined range of the Lower Cieszyn Shales and the Cieszyn Limestones, correspond largely, if not exactly, to that of the KVB Sequence of cratonic

Poland. Latest Kimmeridgian transgressive tendencies can be recognized in the Outer Carpathian Domain as well as in cratonic Poland, where the Meta-Carpathian Arch experienced uplift concomitant with the accumulation, farther north, of Purbeck-type sediments of the Kcynia Formation in the Late Tithonian and Early Berriasian. In the Carpathian Domain, the depositional environments of the Štramberk Limestone were contrasted with those of the Cieszyn Beds, the latter evolving towards conditions of flysch-type sedimentation. An explanation for these coinciding events will be proposed in the following section of this paper.

ALTERNATION OF INTERPLATE STRESSES

The alternating patterns of subsidence and uplift in the studied areas of cratonic Poland can be explained with reference to the model developed by CLOETINGH and his co-authors (*see e.g.* CLOETINGH 1991, 1992; CLOETINGH & *al.* 1985, 1990; KOOI & CLOETINGH 1989; CLOETINGH & KOOI 1989). In accordance with this model, fluctuating intraplate stresses, acting on a supraregional scale, can markedly modify the long-term evolution of a sedimentary basin caused by thermal subsidence. Compressional stresses cause relative uplift of the basin flanks and (additional) subsidence of the basin center; an unconformity is thus produced. On the other hand, increases in the level of tensional stresses induce widening of the basin, lowering of its flanks and (relative) uplift of its center; this results in onlap and transgression in the margins of the basin (*see* Text-fig. 19*B*).

Thicker Oxfordian sediments corresponding to the interval *I* of the Upper Jurassic portion of the COK Sequence accumulated in the area of the Meta-Carpathian Arch than farther north, within the Central European Basin (Text-fig. 10); in the former area, the Oxfordian sedimentation was preceded by a Callovian transgression. The implied subsidence pattern is indicative of a tensional stress regime (*see* Text-fig. 19, *COK/I*). Conversely, the Lower Kimmeridgian sediments corresponding to the interval *II* of the COK Sequence attain greater thicknesses within the Central European Basin than farther south, near the Carpathians (Text-fig. 10), where a discontinuity may exist between the COK Sequence and the Coquina Formation of the overlying LUK Sequence. The latter subsidence (and uplift) pattern is indicative of compressional stresses (*see* Text-fig. 19, *COK/II*).

Because of the incomplete stratigraphic record in the area of the Meta-Carpathian Arch (the absence of the topmost sediments of the LUK Sequence, and the total absence of sediments of the KVB Sequence; Text-fig. 11), the events pertinent to the development of the LUK and KVB Sequences are illustrated herein (Text-fig. 19, *LUK* and *KVB*) in an oversimplified fashion. The subsidence pattern implied by the sediments of the LUK Sequence (the Coquina Formation), which thicken from the Central European Basin towards the flanking Meta-Carpathian Arch (Text-fig. 11), is again

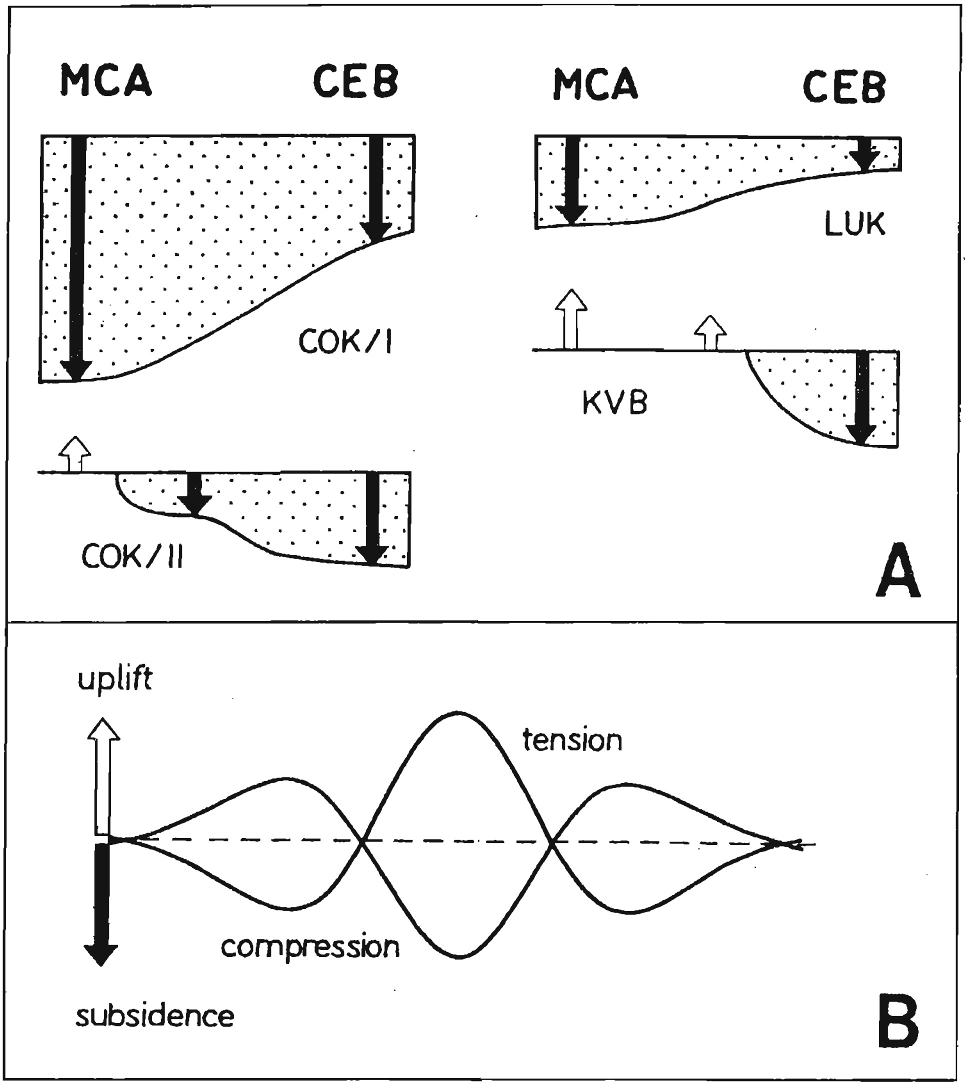


Fig. 19

A — Schematic illustrations showing subsidence and uplift within the Meta-Carpathian Arch (*MCA*) and the Polish part of the Central European Basin (*CEB*) during the formation of the COK Sequence (intervals *I* and *II*), and the LUK and KVB Sequences (not strictly to scale)

B — Schematic illustration of stress-induced subsidence/uplift of a basin (*after* CLOETINGH & KOOI 1992) indicative of an extentional stress regime (*see* Text-fig. 19, *LUK*). However, a compressional episode may be responsible for the regressive tendencies revealed by the topmost portions of the LUK Sequence. The Late Tithonian and Early Berriasian uplift of the Meta-Carpathian Arch, concomitant with the accumulation of the sediments of the Kcynia Formation within the Central European Basin, indicates a compressional stress regime (*see* Text-fig. 19, *KVB*). On the other hand, the transgressive tendencies, corresponding to an earlier period of the KVB interval, are suggestive of tensional stresses.

The sedimentation of the Štramberk carbonates and the Cieszyn Beds in the Carpathian Domain can also be interpreted with reference to the same model. The transgressive nature of the early Štramberk carbonates suggests their accumulation in a tensional stress regime. This, however, could be coupled with an increase of submarine relief caused by extentional faulting. Good evidence for syndepositional faulting, affecting Štramberk-type carbonates near the Carpathian Domain, can be found in the western Ukraine (*see* Text-figs 14*A* and 15*A*). Afterwards, increasing compressional stresses could induce uplift (relative in part) of depositional zones of the Štramberk carbonates, and intensify the subsidence of the basin accomodating the Cieszyn Beds, thus enhancing flysch-type sedimentation in this basin. This also explains the large quantities of Štramberk material redeposited in the Cieszyn Basin.

Finally, it is pertinent to note that the fairly large vertical motions of the area of the Carpathian foreland in the Jurassic, especially during its Oxfordian subsidence, imply high levels of intra-plate stresses (*see e.g.* CLOETINGH 1991, 1992; CLOETINGH & *al.* 1990; KOOI & CLOETINGH 1989). This is in agreement with the independent evidence for Jurassic rifting and crustal attenuation in that area. However, it should be borne in mind that the transition from the Meta-Carpathian Arch to the Central European Basin broadly coincides with the northern boundary of the peri-Carpathian segment of the Polish Rift Basin. Hence, the contrasting subsidence and uplift patterns here discussed (Text-fig. 19) can also be explained, in part, assuming a different (and shifting through time?) location, in different segments of the Polish Rift Basin, of zones of greatest stretching at crustal and lithospheric mantle levels, and of the resultant zones of thermal expansion and contraction.

PECULIARITY OF THE CRACOW REGION

In the context of Miocene events, the region of Cracow was called the "Cracow bolt" (NEY 1968) because the foredeep of the Carpathians, filled with Miocene sediments, is extremely narrow and shallow in this region. The Cracow region is also characterized by extremely reduced Permo-Triassic successions. Middle Jurassic deposits rest directly on pre-Permian rocks in several sections, and Oxfordian deposits are usually directly overlain by Upper Cretaceous ones. The Oxfordian deposits, which thickness, if complete, can be estimated to be about 300 m (*see* JAWOR 1980), may account for more than the half of the total thicknesses of Permo-Mesozoic successions in several sections.

It is of interest that the thickness of the Oxfordian, which amounts to about 300 m in the region of Cracow, increases to over 400 m farther north in the Cracow-Silesian Monocline, in the region of Częstochowa (KUTEK & *al.* 1990). This is consistent with the usually smaller thicknesses revealed by the Mesozoic deposits of the Cracow region, but this also indicates a thickness distribution reversed with regard to that displayed, farther east in the Miechów Synclinorium, by the Oxfordian deposits which thicken to the south towards the Carpathians (*see* Text-fig. 10).

This anomaly can be explained as follows. As remarked by MAHEL (1983), the regions of the Carpatho-Balkan domains, that had been affected by Variscan granitoid plutonism, developed in the Mesozoic as positive ("geanticlinal") units, whereas rifts giving rise to deep troughs developed in crustal regions devoid of such a plutonism. Similarly, the Mid-Polish Anticlinorium, which resulted from the inversion of the Polish Rift Basin, occupies a zone not affected by any Variscan plutonism. In the Cracow region, on the contrary, several manifestations of Variscan plutonism have been recognized (*see* BUKOWY 1994). Hence, it can be assumed that the presence of low-density granitoid bodies in the Cracow region hindered strong subsidence throughout post-Variscan time. As a consequence, abnormally thin, or incomplete, Mesozoic and Miocene successions developed in that region.

A MODEL FOR THE POLISH RIFT BASIN

The peri-Carpathian segment of the Polish Rift Basin developed in Mid-Jurassic time as an asymmetric structure, bounded on its north-eastern, Ukrainian side by a pronounced fault zone (*see* Text-figs 14*A*, 15 and 18). Also other segments of the Polish Rift Basin usually reveal faults of greater magnitude on their north-eastern side (POŻARYSKI & BROCHWICZ-LEWIŃSKI 1978, 1979; *see* Text-fig. 4). Still more significantly, the Mesozoic tectonic development of the areas situated east of the Polish Rift Basin displayed relatively little regional variation, whereas several broad tectonic highs and lows developed on the western side of the rift basin (*see e.g.* SZYPERKO-TELLER & MORYC 1988, GAJEWSKA 1988b, DECZKOWSKI & FRANCZYK 1988, DAYCZAK-CALIKOWSKA & MORYC 1988, DADLEZ 1989). Moreover, the Earth crust is usually thinner in Poland west of, than beneath, the Mid-Polish Anticlinorium (GUTERCH & *al.* 1986a, b, 1991); this is indicative of a shift of the greatest crustal attenuation with regard to the graben zone of the Polish Rift Basin. All these are features that are consistent with, and can be explained by, models of rifting that assume simple shear on lithospheric scale (*see e.g.* WERNICKE 1985, COWARD 1986, FAVRE & STAMPFLI 1992).

An explanation for the different tectonic development in the Mesozoic of the areas situated west and east of the Polish Rift Basin should not be sought only in the fact that the latter areas belonged to the ancient East European Craton. In this context, it is pertinent to note that examples are known of areas previously included in one craton, which subsequently experienced different development, in agreement with simple-shear models, on opposite sides of evolving rifts (FAVRE & STAMPFLI 1992, LIGHT & *al.* 1993).

Therefore, it can be suggested that simple-shear models can be successfully applied to explain the development of the Polish Rift Basin. However, the comprehensive problem but outlined herein is beyond the scope of this paper.

Acknowledgements

Helpful suggestions kindly offered by Professor A. RADWAŃSKI, the Editor of the *Acta Geologica Polonica*, allowed to improve substantially the design of this paper, and the included illustrations, which is acknowledged with gratitude.

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