Phanerozoic environmental changes in the Caucasus and adjacent areas: stratigraphy, fossil diversity, mass extinctions, sea-level fluctuations, and tectonics

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Sworn statement before a commissioner of oaths

Herewith, I state that none of the 21 papers included in the present thesis was submitted by me elsewhere for a PhD or any other degree. For 7 co-authored papers, a clarification of the relative contribution of each author is stated below.

Ruban, D.A. & Tyszka, J. 2005. Diversity dynamics and mass extinctions of the Early-Middle Jurassic foraminifers: A record from the Northwestern Caucasus. Palaeogeography, Palaeoclimatology, Palaeoecology. 222: 329-343.*
* corresponding author
Ruban: recalculation and interpretation of data on the Northwestern Caucasus, regional versus global comparisons, development of the general concept of the paper and writing its major parts
Tyszka (Institute of Geological Sciences, Cracow Research Center, Polish Academy of Sciences, Poland): data and interpretations on the Polish regions taken for a comparison, general editing of the paper and writing some of its parts (including those concerning the Polish regions and preparation of figures)

Ruban, D.A. & Yoshioka, S. 2005. Late Paleozoic - Early Mesozoic Tectonic Activity within the Donbass (Russian Platform). Trabajos de Geología. 25: 101-104.*
*corresponding author
Ruban: writing of the whole manuscript, including development of the main tectonic concept
Yoshioka (Kyushu University, Japan): general editing of the paper

Tawadros, E., Ruban, D. & Efendiyeva, M. 2006. Evolution of NE Africa and the Greater Caucasus: Common Patterns and Petroleum Potential. The Canadian Society of Petroleum Geologists, the Canadian Society of Exploration Geophysicists, the Canadian Well Logging Society Joint Convention. May 15-18, 2006. Calgary. P. 531-538. [extended abstract]*
*presenting author
Tawadros (petroleum consultant, Canada): characteristics of the Phanerozoic evolution of the Northeastern African basins; discussion of the comparison between NE Africa and the Greater Caucasus and their hydrocarbon potential; general editing of the paper (including figures)
Ruban: characteristics of the Phanerozoic evolution of the Greater Caucasus; discussion of the comparison between NE Africa and the Greater Caucasus and their hydrocarbon potential; preaparation of the key illustrations
Efendiyeva (Geological Institute, National Academy of Science, Azerbaijan): data on the Eastern Caucasus; data on the petroleum reserves of the Azerbaijan Hydrocarbon Province

Ruban, D.A., Al-Husseini, M.I. & Iwasaki, Y., 2007. Review of Middle East Paleozoic Plate Tectonics. GeoArabia. 12: 35-56.
no corresponding author is indicated; all authors contributed equally
Ruban: interpretations of data on the Caucasus and some other regions of the Middle East, comparison of available tectonic reconstructions, preparation of the initial draft of the manuscript, general editing of the paper

Al-Husseini (Gulf PetroLink, Bahrain): writing the chapters on some regions, preparation of the final draft of the manuscript, general editing of the paper, drawing the figures (with support of Gulf PetroLink technical staff)
Iwasaki (American Museum of Natural History, USA): interpretation of the Devonian paleogeography on the basis of trilobite analysis, general editing of the paper

Ruban, D.A. & van Loon, A.J. 2008. Possible pitfalls in the procedure for paleobiodiversity-dynamics analysis. Geologos. 14: 37-50.*
* corresponding author
Ruban: writing the whole manuscript and development of the main concepts
van Loon (Adam Mickiewicz University, Poland; Netherlands): addition of some general considerations on paleodiversity analysis, editing of the paper (with a preparation of the final draft) and drawing the final versions of its figures

Gutak, Ja.M., Tolokonnikova, Z.A. & Ruban, D.A. 2008. Bryozoan diversity in Southern Siberia at the Devonian-Carboniferous transition: new data confirm a resistivity to two mass extinctions. Palaeogeography, Palaeoclimatology, Palaeoecology. 264: 93-99.*
*corresponding author
Gutak (Kuzbass State Pedagogical Academy, Russia): providing the data for geological context of the paper (stratigraphy, paleoenvironmental interpretations)
Tolokonnikova (Kuzbass State Pedagogical Academy, Russia): providing the initial data on bryozoan stratigraphic ranges and taxonomy, preliminary drawing of some figures (except those demonstrating the main results)
Ruban: data recalculation and interpretation, regional versus global comparisons, discussion of data in paleoenvironmental context, writing the whole manuscript and drawing of figures

Zorina, S.O., Dzyuba, O.S., Shurygin, B.N. & Ruban, D.A. 2008. How global are the Jurassic-Cretaceous unconformities? Terra Nova. 20: 341-346.*
*corresponding author
Zorina (Central Research Institute of Geology of Industrial Minerals, Russia): data on the eastern part of the Russian Platform, discussion of the paper's general concept
Dzyuba (Trofimuk Institute of Petroleum Geology and Geophysics, Siberian Branch of Russian Academy of Sciences, Russia): data on West and East Siberia, discussion of the paper's general concept
Shurygin (Trofimuk Institute of Petroleum Geology and Geophysics, Siberian Branch of Russian Academy of Sciences, Russia): data on East and West Siberia, discussion of the paper's general concept
Ruban: development of the general concept of the paper, compilation and interpretation of data from many regions across the globe (including incorporation of data given by 3 other co-authors), discussion of eustatic curves, examination of possible causes of eustatic drops, writing the whole manuscript and drawing of all figures

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Dmitry A. Ruban

SUMMARY

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1. Introduction and main objectives

1.1. Geological setting of the studied regions

The Caucasus is a large region, which occupies the territory of southwestern Russia, and all of Georgia, Armenia, and Azerbaijan. It is dominated by two subparallel mountain chains, namely the Greater Caucasus (which includes the Main Caucasian Range with the highest European peak - Mt. Elbrus) and the Lesser Caucasus, which both stretch between the Black Sea and the Caspian Sea. The geological structure of the Caucasus is complicated (Fig. 1). It includes two main domains, also referred to as the Greater Caucasus and the Lesser Caucasus, which are divided by two Transcaucasian depressions, i.e., the Rioni Depression and the Kura Depression (Fig. 2). A large area, which lies to the north of the Caucasus, is the so-called Ciscaucasus, which is a young stable platform developed over Paleozoic structures. A famous Late Paleozoic coal-bearing basin, named the Donbass, forms something of a branch or offshoot, derived from the Ciscaucasus. This basin cuts off the Russian Platform, whose southern block is the so-called Ukrainian Massif with its eastern edge being known as the Rostov Dome (Fig. 3). The geological evolution of the Ciscaucasus, the Donbass basin, and the Rostov Dome was linked closely to that of the Caucasus.

1.2. General purpose of this study

The Caucasus is often regarded as a typical Alpine region, because the present-day Caucasian geological architecture is dominated by structures developed during the Alpine phases, i.e., during the Cenozoic (Ershov et al., 2003; Tawadros et al., 2006). Undoubtedly, the Caucasus is an important section in the Alpine active belt, which stretches from the Atlantic Ocean to Southeast Asia, but the evolution of the Caucasus and the Alps might have been connected even more strongly, especially in the Late Paleozoic-Early Mesozoic. The geology of the Caucasus provides a rich source of information on the Phanerozoic paleoenvironmental changes, which should not only be discussed in a regional framework, but also in the global context. The key position of the Caucasus, between the Alps and Carpathians in the west, the Iranian and Central Asian domains in the east, the Precambrian Russian Platform (craton) in the north, and the Turkish domains in the south, makes the Caucasus a very important region to discuss the changes in the regional paleoenvironments. These changes can help to enhance our understanding of the evolution of the whole Tethyan sector. Unfortunately, the data from the Caucasus are only rarely used in determining these large-scale geological constraints, and the Caucasian region appears to be largely ignored (with very few exceptions) in international geology. My study is aimed at providing some essential knowledge on the Phanerozoic record of the Caucasus and adjacent areas (the Donbass and the Rostov Dome).

The conclusions from the attempted studies were published in international journals. Twenty one articles are included in this thesis, and a number of other papers are in press, accepted or submitted.

1.3. Main objectives

- A compilation of litho-, bio-, and chronostratigraphic information in order to constrain the modern stratigraphic framework and to recognize spatial changes in the sedimentary architecture in the Caucasus and the Rostov Dome.
- A careful compilation of a vast amount of already published paleontological data on various fossil groups from the Paleozoic-Mesozoic of the Caucasus, including

brachiopods, bivalves, ammonoids, and foraminifers. These data are essential for further constraints of diversity dynamics.

- An examination of general trends in the Paleozoic-Mesozoic fossil diversity in the Caucasus with a special attention to brachiopods as the most diverse and the best studied group.
- A recognition of the regional signatures of the Frasnian/Famennian, Permian/Triassic, Triassic/Jurassic, Pliensbachian/Toarcian, and Jurassic/Cretaceous mass extinctions in the Caucasus.
- An evaluation of Paleozoic and Triassic-Jurassic transgressions, regressions, and changes in basin depth in the Caucasus on the basis of facies analysis and with the use of the constrained stratigraphic frameworks. Regional sea-level changes are to be brought in correspondence with the global eustatic curves. It is always important to evaluate possible relationships between fossil diversity and sea level changes.
- A development of new models of the Phanerozoic tectonic evolution of the Greater Caucasus and the Late Paleozoic-Triassic evolution of the Donbass. These regional models are made with respect to global plate tectonics, accounting for terrane displacements, activity of planetary shear zones, and continental breakups.
- New paleotectonic constraints help in the interpretation of the regionally-documented paleoenvironmental changes. On the other hand, the stratigraphical and paleontological conclusions make possible important improvements in the regional paleotectonic constraints.
- Interregional comparisons of the geological events are essential to discuss similarities and dissimilarities of the geological evolutionary processes and to establish geological analogues of the Caucasus and adjacent areas.

Although the entire Phanerozoic record of the Caucasus and adjacent areas is examined, I emphasize the Devonian-Jurassic time interval, whose stratigraphical, paleontological, and tectonic record is the richest, the most diverse, and, therefore, the most intriguing. The unusual Neogene record of the Rostov Dome is examined in detail (Ruban, 2005a).

2. Materials and methodological framework

2.1. Materials

All stratigraphical data used to perform my studies were collected during field studies in the Western and Central Caucasus (1996-2008), in the Donbass Basin (1996-2006), and in the Rostov Dome area (1996-2002). Field excursions in Azerbaijan (2007) and the Swiss Alps (2008) under the guidance of the local specialists also helped to strengthen an understanding of some geodynamic interpretations.

In the Caucasus, a number of sections and outcrops were investigated (Fig. 4). Among them are the Lipovyj section of the Early Toarcian deposits (probably accumulated on a rocky shore) (Fig. 5), the Bezymjannaja section of the transitional Aalenian-Bajocian crinoid limestones (limestones of this age were first found by the author in the Western Laba-Malka Area), the Khadzhokh-2 section of the condensed Callovian siliciclastics with an exceptionally abundant fossil assemblage, and many others (Fig. 6). Composite sections are delineated in order to summarize stratigraphic data from particular sets of sections and outcrops (e.g., Fig. 7). It is necessary to emphasize that an analysis of composite sections is crucial because of two reasons. Firstly, very few outcrops extend significantly over continuous sections in the Caucasus. Secondly, the very large size of the Caucasus area requires summarizing of the data. For example, Jurassic composite sections are constructed (Ruban, 2007a) for each particular area distinguished within the Caucasus, by lithologic peculiarities by Rostovtsev et al. (1992) Twelve suitable (i.e. good outcrops) sections of the

Upper Miocene deposits were investigated in the Rostov Dome area (Figs. 8, 9). The information deduced from these sections is presented and interpreted in papers included into this thesis (see list below).

Paleontological sampling was oriented mainly for stratigraphic purposes. Some representative samples are stored in my private collection (Appendix 1). An examination of stratigraphic ranges of some common invertebrate species permitted a re-evaluation of the age of the strata. For example, a comparison of the brachiopod assemblage from the crinoid limestones of the Dzhangurskaja Formation with the characteristic assemblages of Western Europe (Cariou & Hantzpergue, 1997) permitted me to confirm an Aalenian age of these beds, which had been questioned in earlier literature (Rostovtsev et al., 1992).

A careful compilation of already published paleontological data on various fossil groups (e.g., see appendices 5, 6) was supported by a critical examination of the hundreds of published sources on regional geology and paleontology to minimize uncertainties, misinterpretations (especially stratigraphic), and problems with taxa synonymy. All earlier-published data on fossils were enhanced where possible, some with the assistance of European and American specialists, who corrected taxonomic lists and justified the suprageneric taxonomy for some fossil groups (e.g., Y. Almeras, M. Bécaud, - Jurassic brachiopods; A.J. Boucot - suprageneric taxonomy of Permian-Jurassic brachiopods; M. Bécaud - Early Jurassic ammonites; N.M.M. Janssen, W. Riegraf - belemnites; A.A. Kasumzadeh, W. Schätz - Triassic bivalves). General regional paleontological overviews were considered together with publications based on case studies in order to avoid sampling errors. All datasets are aimed to be representative. Field and literature data are always incorporated as accurately as possible to reach their best confidence levels. For example, discovery of Middle Jurassic crinoid limestones in the Western Laba-Malka area and a discussion of their age permitted the re-positioning of some facies distributed widely within the Greater Caucasus at the regional stratigraphic scale (Ruban, 2007a). This enabled the timing of the local sea-level fall to be estimated.

2.2. Methods

The methodological framework of this study is multidisciplinary, and it comprises several steps of studies. Firstly, the regional (litho-, bio-, chrono-) stratigraphic frameworks are improved, particularly to bring these in line with the present chronostratigraphical developments of the International Commission on Stratigraphy (ICS) of the International Union of Geological Sciences. For example, a position of the base-Aalenian boundary was justified according to the GSSP (Global Stratotype Section and Point) in Fuentelsalz with data on ammonoids and foraminifers. The regional Upper Miocene stages of the Eastern Paratethys were replaced with the global stages approved by the ICS on the basis of correlation of absolute stage boundaries (Ruban, 2005a). The next step was the compilation of all available paleontological data. For this purpose, stratigraphic ranges of particular taxa were summarized in a series of datasheets (appendices 2, 3; see also Ruban, 2004, 2006a,b,d, 2007c, 2008; Ruban & Tyszka, 2005). This allows the establishment of a number of trends in the fossil diversity changes, and the documentation of mass extinctions (and also to hypothesize a new mass extinction, in the Aalenian). It is necesary to note, that theoretical background of fossil data in preparation for further quantitative analysis was considered (Ruban & van Loon, 2008). To address possible problems with sampling errors and taxa interpretations, the author visited collections of Triassic bivalves stored at the Geological Institute of the Azerbaijan National Academy of Sciences (Baku), where they are curated by A.A. Kasumzadeh. To reveal sea-level changes, datasets of composite lithologic sections are used (e.g., Ruban, 2007a). Facies analysis (see 6.1) was applied in order to interpret these data. Special attention was also paid to the paleotectonic reconstructions. Those already existing are mostly based on the outdated geosyncline paradigm and the so-called formation analysis (e.g., Laz'ko, 1975). In contrast, I attempt to apply modern concepts of plate tectonics and terrane analysis. Interregional comparisons of lithologic and paleontologic data (Ruban, 2007b,d) as well as tracing of the major unconformities (Ruban, 2007b) permitted me to recognize the key large-scale tectonic events, well-known and well-interpreted in the regions of Europe and Middle East, and, thus, to reach conclusions about their nature within the study areas for this thesis. For example, the mid-Permian unconformity is well traced in the Variscan structures of Europe, where it is known as a Saalian unconformity (in some localities, as a series of unconformities). Thus, a Saalian phase of tectonic activity can be hypothesized in the Greater Caucasus (Ruban, 2007b).

Some very specific methods used for the purposes of the present study are discussed in the relevant chapters below.

3. Stratigraphy

Regional litho-, bio-, and chronostratigraphy of the Caucasus and adjacent areas is improved in order to obtain a much improved stratigraphic framework and to permit precise interregional correlations.

3.1. Lithostratigraphy

Hundreds of formations are established in the Phanerozoic succession of the Caucasus and adjacent areas, but an especially complicated situation occurs with the Jurassic strata of the Caucasus and the Upper Miocene strata of the Rostov Dome. In the first case, formations were established originally in 36 particular areas for the Hettangian-Bathonian interval and in 26 areas for the Callovian-Tithonian interval. The author's re-examination of available data as well as his own field studies permitted some measure of updating of the knowledge of these lithological packages (Ruban, 2007a). In particular, an investigation of outcrops in the basin of the River Belaja led to the recognition of the so-called Bizhgon Member (Rostovtsev et al., 1992), composed of crinoid detrital pink-colored limestones, which was not established in the Western Laba-Malka area by previous studies. Investigation of faunal assemblages permitted me to date this member and to change its position in the regional lithostratigraphic scheme, which is important for further paleogeographical constraints. I also re-examined and documented in detail the stratotype section of the Kamennomostskaja Formation (Ruban, 2007b), i.e., the Khadzhokh-2 Section (Figs. 4, 6). This sheds a new light on a very uncertain description of this important section, which is one of a very few exposed Callovian sections in the Caucasus. Although this formation was established earlier, its re-examination confirms a striking lithological distinction from the under- and overlying sedimentary complexes to fulfill the ICS requirements (an angular unconformity at the base of this formation is traced to separate it from the Triassic flysch deposits; although lithologically heterogeous, this formation is characterized by a dominance of clastics in contrast to the overlying carbomates). The Upper Miocene strata represented by skeletal limestones cover the Rostov Dome entirely. However, no formations were defined there until now, except for the Janovskaja Formation. I established a set of new formations and suggested their precise correlation (Ruban, 2005a). All reference and other key Upper Miocene sections of the Rostov Dome were investigated (Figs. 8, 9), and as a result the Taganrogskaja, Rostovskaja, Donskaja, Merzhanovskaja and Aleksandrovskaja Formations were first recognized, and their logs were documented (Ruban, 2005a). Facies-based logs are yet to be published, although facies interpretations for each section were carried out. Establishing their spatial relationships provides a necessary clue to reveal the dynamics of past shorelines of the Paratethys Sea.

Three additional tasks related to lithostratigraphy were also resolved. First, a composite Paleozoic lithological section of the northern part of the Greater Caucasus was constructed, with an indication of the main sedimentary packages (Ruban, 2006a, 2007b). Secondly, the Triassic lithostratigraphy of the Western Caucasus was revised, taking into account previous constraints, new suggestions, and the field observations (Ruban, 2006b, 2008) (Fig. 7). Thirdly, four major unconformities are recognized in the Paleozoic-Mesozoic succession of the Greater Caucasus - in the Ordovician, mid-Permian, Triassic/Jurassic, and mid-Jurassic. These are described, correlated, and explained (Ruban, 2007b). The first three of them have clear analogs in adjacent regions of Europe and the Middle East. The Jurassic unconformities known from the Greater Caucasus are discussed in a very broad global context in order to trace the planetary-scale sedimentation breaks (Zorina et al., 2008). It thus seems that the Triassic/Jurassic unconformity is of global extent.

3.2. Biostratigraphy

The previous biostratigraphic subdivision of the Jurassic of the Caucasus based on ammonites was quite detailed, but required updating because of numerous corrections to the Jurassic time scale during the two past decades. In order to resolve this important task, I re-examine biostratigraphic data and provide an improved version of the inferred biozonation (Ruban, 2006c, 2007a). New data permit new determinations of the positions of the Aalenian/Bathonian and Tithonian/Berriasian boundaries in the regional record. The validity of ammonite zones is confirmed, but they are also compared with the data on other fossil groups like brachiopods and foraminifera. Ammonite- and foraminifera-based biostratigraphic units are correlated and also justified according to a regional lithostratigraphic subdivisional scheme (Ruban & Tyszka, 2005). A totally new biostratigraphic scheme is developed for the Upper Miocene deposits of the Rostov Dome (Ruban, 2005a). An abundance of bivalve remains permits identification of the principal bioevents (first and last occurrences) and enabled me to outline the *Tapes vitalianus* Interval Zone, the *Cerastoderma fittoni-C. subfittoni* Total Ranges Zone, the *Congeria panticapaea* Interval Zone, the *Congeria amygdaloides navicula* Total Range Zone, and the *Monodacna pseudocattilus-Prosodacna schirvanica* Interval Zone. These new units are brought into correspondence with the new formations noted above in section 3.1. The biostratigraphic units established in the Upper Miocene of the Rostov Dome are local, although they can serve as startpoints for a further definition of bivalve-based biozones of the entire Eastern Paratethys. In all cases, mentioned above, potential effects of fossil resedimentation and reworking (in terms of the present taphonomic concepts) were accounted for as accurately as possible.

3.3. Chronostratigraphy

The chronostratigraphic subdivisions used for the Caucasus and the Rostov Dome are updated according to the newest developments and recommendations of the International Commission on Stratigraphy. A three-fold subdivision of the Permian is traced in the Western Caucasus (Ruban, 2007b). A clear distinction of the Norian and the Rhaetian stages in the regional record is confirmed (Ruban, 2008). The justified Jurassic chronostratigraphic framework is extended to the entire Caucasus (Ruban, 2006c, 2007a). Additionally, globally-recognized stages are traced within the Upper Miocene strata of the Rostov Dome (2005a). A correlation of local biozones to such global stages as the Serravallian, the Tortonian, and the Messinian is based on the available absolute ages of their boundaries (for global stages, very precise dates provided at the Global Stratotype Sections and Points (GSSP's) are considered).

The improvement in the regional stratigraphic framework detailed above provides a comprehensive basis for further discussion of data and results of their interpretations, in the global context.

4. Fossil diversity

4.1. Specific methods

Taxonomic diversity is analyzed for the entire marine fauna of the Caucasus and its counterparts, and specifically for the particular fossil groups, including brachiopods, bivalves, ammonites, belemnites, and foraminifers. The data used to measure the changes in the fossil diversity of the Caucasus are compiled from numerous available sources. After a compilation, they have been examined critically and improved according to the current taxonomy. A total of about 1000 valid species are considered. Special attention is paid to the Triassic and Jurassic periods, which were characterized by the most marked richness of the local faunas, although the Paleozoic record is not omitted. For the purposes of this study, a number of standard and new methods are used. The general principles of paleobiodiversity studies are outlined by Ruban & van Loon (2008), who give the main techniques and possible solutions to the common problems. The standard methods include quantitative analyses of total diversity dynamics, and changes in the number of originated/appeared and extinct/disappeared taxa. To make a clear distinction between originations-extinctions and appearances-disappearances, it is crucial to take into account the probable influences of interruptions in taxa stratigraphic ranges. These temporal gaps are brought into correspondence with the so-called Lazarus-effect. I propose a way, if not to minimize it, to at least account for it as accurately as possible (Ruban & Tyszka, 2005; Ruban & van Loon, 2008). The initial calculation of the total diversity or number of appearances and disappearances is followed by the same calculation, but with data hypothesizing the probable presence of a taxon at a time of its registered gap in the regional record. Thus, the highest probable value (HPV) of diversity indices is evaluated. A measurement of the HPV for the diversity of the Early-Middle Jurassic foraminifers of the Northwestern Caucasus indicates its large dimensions. At the same time, this does not affect significantly the data interpretation nor trends in diversity measured without accounting for the Lazarus-effect.

Two special indices are proposed to investigate the evolutionary rates of fossil groups (Ruban & Tyszka, 2005; Ruban, 2006d, 2007c, 2008; Gutak et al., 2008). The first R-method is a simple calculation of the Jaccard's similarity for successive faunistic assemblages, where the number of common taxa in two comparable intervals is related to their whole diversity. The result shows a rapidity of changes in the assemblage composition through the geological time interval. The Rst-method is based on a calculation of the so-called pair-correlation between successive or non-successive assemblages. For each of the latter, the presence of higher-ranked taxa is indicated by the number of lower-ranked taxa, by which the former are represented in a given assemblage. This permits an evaluation of the rate of transformation of the taxonomic diversity structure. It shows the changes in the controls of lower-ranked taxa (e.g., species) by those that are higher-ranked (e.g., genera). This new method sesms to be a powerful tool to document the fundamental changes, re-organizations, and turnovers in faunistic evolution. Moreover, its application to non-successive assemblages (e.g., a comparison of the Cambrian and the Jurassic assemblages) may provide some important clues to the understanding of the overall fossil evolution. To test the new Rst-method with data from any region outside the Caucasus is crucial to weigh up its efficacy and probable limits. For this purpose, I chose the Devonian bryozoans of Southern Siberia (it might have been connected with the Greater Caucasus by the chain of Kazakh terranes). Similarly informative conclusions are made (Gutak et al., 2008). Moreover, a triplicated calculation of the Rst indices (for species-genera, genera-family, and species-family taxonomic levels) reveals transformations in both generic and familial controls of the whole diversity for the studied group.

4.2. Triassic biota

The Triassic fossil record is best preserved in the Western Caucasus (Appendix 2). The total marine biodiversity was quite low in the Early Triassic as a consequence of the Permian/Triassic mass extinction. However, a strong radiation, which can be judged as a regional "diversity explosion" occurred in the Anisian, when the number of species trebled (Ruban, 2006b). This was followed by a new drop in species numbers in the Ladinian. Then, a stepwise growth in the marine biodiversity set in, and reached a peak in the Late Triassic with its rich reefal assemblages. It is interesting to document a difference in the dynamics of particular fossil groups. Whereas the Anisian was a favorable time for the entire marine fauna, brachiopods and ammonoids declined sharply in the Ladinian, whereas the species diversity of bivalves and foraminifers descreased only a little. Ammonoid assemblages were very poor in the Carnian, when a strong repopulation of brachiopods, bivalves, and foraminifers began. Foraminifers declined somewhat in the Norian, which is characterized by the very high diversity of other groups, and also by an appearance of algae, corals, and sponges, not known from the older intervals. In contrast, no bivalves are found in the Rhaetian strata despite a high diversity of other marine organisms. This suggests an absence of any simple relationships between the diversity dynamics of the overall marine population and that of the particular groups of fossils. The R- and Rst-methods are used to reveal the evolutionary rates of the Triassic macrofauna (Ruban, 2008). Until the middle of the Late Triassic, they remained very high. Each younger assemblage differed from its predecessor fundamentally with a complete turnover in taxonomic composition. However, the Norian and the Rhaetian assemblages were much more similar. An analysis of non-successive assemblages leads to another intriguing observation. The Early Triassic marine macrofauna differed from that of the Anisian more strongly than from the younger Ladinian-Carnian. However, the Norian-Rhaetian assemblages were renewed significantly.

4.3. Jurassic biota

During the Jurassic, the number of bivalve, brachiopod, belemnite, ammonite, and foraminiferal species changed in a distinct way (Ruban & Tyszka, 2005; Ruban, 2006d, 2007a). The number of bivalve species remained low throughout the Early Jurassic. Then, it rose stepwise with a strong peak in the Callovian-Oxfordian. However, a rapid decline occurred in the Kimmeridgian-Tithonian (Ruban, 2006d, 2007a). The brachiopod diversity fluctuated strongly (Ruban, 2006a, 2007a). The peaks were reached in the Pliensbachian, the Bajocian, and the Oxfordian, whereas diversity minimums are registered in the Early Toarcian, the Early Aalenian, the Bathonian, and the Kimmeridgian. The number of ammonite species was the highest in the Late Toarcian, the Bajocian, and the Early Callovian (Ruban, 2007a). Belemnite assemblages remained highly diverse during the Pliensbachian-Bathonian, whereas they were limited before and after this time interval (Ruban, 2007a). Finally, radiations of foraminifers in the Pliensbachian, the Late Toarcian-Early Aalenian, and the Late Bajocian are registered (Ruban & Tyszka, 2005). The only fossil groups which demonstrated clear trends in total diversity changes throughout the Jurassic, were bivalves (a trend towards a diversification) and ammonites (a trend towards a decline).

4.4. Phanerozoic diversity of brachiopods

The especially detailed studies are addressed to brachiopods (Ruban, 2004, 2006a, 2007c), bivalves (Ruban, 2006d), and foraminifers (Ruban & Tyszka, 2005). Brachiopods are known from the entire Cambrian-Cretaceous interval of the northern part of the Greater Caucasus (Appendix 3). Their first radiation occurred in the mid-Cambrian (Ruban, 2006a). Then, they diversified in the Late Silurian (Ludlow-Přidoli)-Early Devonian. A somewhat stronger radiation took place throughout the Frasnian-Famennian. But the highest diversity of the Paleozoic brachiopods is recorded in the Late Permian (Lopingian), when dozens of taxa appeared. A new radiation occurred during the Triassic with the highest diversity observed at the Late Triassic-Early Jurassic interval (Ruban, 2006a). This trend was interrupted by the Ladinian event, when brachiopods disappeared from the regional record totally. Since the Middle Jurassic, the number of brachiopod taxa decreased, although this trend was interrupted by a few short-term peaks (Bajocian, Tithonian). During the Early Cretaceous, the Caucasian brachiopod assemblages remained impoverished. A detailed investigation of the Early-Middle Jurassic diversity dynamics of brachiopods permits me to conclude that fluctuations in their total diversity were induced by various combinations of origination and extinction rates (Ruban, 2004). In particular, the rise in species number during the Late Sinemurian-Early Pliensbachian occurred together with an acceleration of both origination and extinction rates, whereas a collapse of the origination rate seems to be no less responsible for the Late Pliensbachioan-Early Toarcian crisis than strengthening in the extinction number. Some very intriguing results are brought by the application of the Rst-method (Ruban, 2007c). A strong turnover in the structure of assemblages occurred during the Early Triassic-Anisian. However, it was much lower during the Late Triassic and in the Early-Middle Jurassic. An analysis of non-successive assemblages indicated a stability of the Late Triassic structure of taxonomic diversity. But it also shows clearly a significant similarity of the Pliensbachian, Toarcian, and Aalenian assemblages to those of the Early Triassic. Thus, the superfamilies, which dominated the species diversity in the Early Triassic, re-established their control since the Jurassic. This provides support to hypothesize a partial re-setting of the brachiopod evolution, which can be linked to the influence of mass extinctions (see below). Such a totally new conclusion is of great importance, because it gives a new view of the fossil resistance to environmental stress. An analysis of diversity dynamics of the Jurassic bivalves also suggests a complicated interaction of the origination and extinction rates (Ruban, 2006d). In particular, it appears that the strong Callovian diversification occurred thanks to an acceleration in the origination rate, whereas the extinction rate slowed somewhat in the Bathonian. The Rst-method indicates an intensification of turnovers in bivalve assemblages in the Early Jurassic, at the Bathonian-Callovian and Kimmeridgian-Tithonian transitions. An examination of the Early-Middle Jurassic foraminiferal assemblages indicates somewhat less intense fluctuations at the species and, especially, at the generic levels (Ruban & Tyszka, 2005). An interaction between origination and extinction rates can be viewed, for example, at the Late Toarcian-Early Aalenian diversity peak. A very low number of extinctions before the Late Toarcian coupled with a prominent acceleration of the origination rate during this interval led to a remarkable growth of the total diversity. Then, the extinction rate strengthened, but it was still recompensated by the number of originations. Thus, no changes in the total diversity occurred in the Early Aalenian. But both an increase in extinctions and a drop in originations led to the succeeding decline of foraminifers. This study also implies that different diversity dynamics between species and genera is possible for the same fossil group and the same time interval. The R- and Rst-methods suggest a low degree of transformation in the composition of assemblages. Even those relatively strong turnovers that occurred at the Pliensbachian-Toarcian and the Aalenian-Bajocian transitions were not so intense. Again, a difference between species and generic levels is observed.

4.5. Interregional comparisons of diversity trends

Quantitative evaluations of the fossil diversity in the Caucasus are compared with data from the other regions (the Swiss Alps, the Bakony Mountains, the Pieniny Klippen Belt) and considered against global constraints. Despite regional peculiarities in faunal evolution, general trends and events are recognized in the Caucasus (Ruban, 2004, 2006a,b,d, 2007c; Ruban & Tyszka, 2005), which suggests its exceptional importance for global biodiversity studies. Various factors were responsible for the regional diversity dynamics. These may be a growth of reefal communities in the Late Devonian, the Late Permian, the Late Triassic, and the Late Jurassic (Ruban, 2006a), abrupt basin deepening in the Ladinian (Ruban, 2006a), marine anoxia during the Early-Middle Jurassic (Ruban, 2004; Ruban &Tyszka, 2005), an onset of a major carbonate platform in the Callovian (Ruban, 2006d), a regional salinity crisis in the Kimmeridgian-Tithonian (Ruban, 2006a,d), changes in the paleotemperatures throughout the Jurassic (Ruban, 2006b), and some others. However, special attention is paid to the role of sea-level changes, which is discussed below. It is found that different fossil groups were not similar in their susceptibility and resistance to the influences of the above-mentioned factors. Foraminifers (Ruban & Tyszka, 2005) and bivalves (Ruban, 2006d) were more tolerant of oxygen depletion than brachiopods (Ruban, 2004). However, the latter were less affected by the regional salinity crisis than bivalves (Ruban, 2006d).

5.Mass extinctions

A number of mass extinctions are established in the Caucasus and studied in detail (Ruban, 2004, 2006a, 2007c, 2008; Ruban & Tyszka, 2005). Some other catastrophes (known elsewhere) are not documented, but their traces and possible consequences for the biotic evolution are discussed. The studied crises include those of the Frasnian/Famennian, Permian/Triassic, Triassic/Jurassic, Early Jurassic (Pliensbachian/Toarcian), and Jurassic/Cretaceous. The most detailed record is available to explore the Early Jurassic mass extinction, which seems to have been not less devastating than the representatives of the famous "Big Five" (Ruban & Tyszka, 2005). A potentially new mass extinction is also registered in the Aalenian.

5.1. Frasnian/Famennian mass extinction

The Frasnian/Famennian mass extinction appears to be the only event, which appeared globally, but did not stress the regional faunal evolution in the Caucasus. A radiation of brachiopods throughout the entire Late Devonian took place, although a turnover at the Frasnian/Famennian boundary is established (Ruban, 2006a). It is important to note, that the Famennian assemblage was dominated by cyrtospiriferids, which also diversified in some other regions during this age. The study of bryozoans from Southern Siberia, a region probably connected with the Greater Caucasus by a chain of the Kazakh terranes, demonstrates that this group was always a successful survivor from the Frasnian/Famennian catastrophe (Gutak et al., 2008).

5.2. Permian/Triassic mass extinction

The Permian/Triassic mass extinction led to an overall collapse of the regional faunas. The marine diversity remained diminished during the Early Triassic, and its full recovery was not completed even by the end of the Triassic (Ruban, 2006a, b). However, it is very important to note that this recovery started very early after the extinction peak. The presence of a characteristic brachiopod taxon, which indicates the base of the Triassic, is outlined as evidence for this (Ruban, 2006a). Moreover, the first bivalve assemblages were dominated by the well-known recovery taxa of *Claraia* (Ruban, 2006b). It appears that an acceleration of the evolutionary rates of the Triassic marine macrofauna of the Western Caucasus was another recovery pattern (Ruban, 2008).

5.3. Triassic/Jurassic mass extinction

The Triassic/Jurassic transition is interrupted by a hiatus in the Caucasus, and, thus, the relevant crisis cannot be documented directly (Ruban, 2007b). However, Ruban (2007c) suggests that the Rst-method applied for non-successive Triassic-Jurassic brachiopod assemblages permits one to investigate the possible influences of this mass extinction on the regional evolution of this fossil group. A comparison of the taxonomic diversity structure of the Rhaetian, Sinemurian, and Pliensbachian assemblages indicates their continued similarity. Such superfamilies as Spiriferinoidea and Zeillerioidea played an important role in both the Late Triassic and Early Jurassic assemblages and, thus, were not wiped out by the mass extinction. This conclusion contrasts with results from the same data, re-calculated from the Swiss Alps, where a significant turnover is registered at the Triassic-Jurassic transition (Ruban, 2007c). It is, however, interesting that the taxonomic diversity structure of the Caucasian brachiopods in the Early Jurassic resembled that in the Early Triassic. This permits me to hypothesize a partial resetting of the regional evolution of this group as a consequence of the Triassic/Jurassic event. These results underline, in general, that the new Rst-method can be a powerful means to explore the traces of catastrophes, even those misplaced from the regional record.

5.4. Pliensbachian/Toarcian mass extinction

The Early Jurassic (Pliensbachian/Toarcian) mass extinction is documented in the Caucasus with precision. It affected brachiopods (Ruban, 2004, 2006a, 2007c) and to a lesser extent, foraminifers (Ruban & Tyszka, 2005). The diversity analysis of bivalves, ammonites, and belemnites (Ruban, 2006d, 2007a) does not indicate any catastrophic patterns in the Early Jurassic. Brachiopods declined strongly already in the Late Pliensbachian (Ruban, 2004). This was preceded by their abnormal radiation. In the Early Toarcian, brachiopods disappeared entirely and no taxa are known from the relevant deposits. A repopulation began in the Middle Toarcian, but even the Late Toarcian diversification did not recompensate for the diversity loss at the Pliensbachian/Toarcian boundary. The results from the Rst-method suggest an intense turnover at this boundary (Ruban, 2007c). Moreover, this mass extinction led to a complete renovation of the taxonomic diversity structure. If the Pliensbachian assemblage is quite similar to that of the Rhaetian, a striking difference between the Toarcian and the Rhaetian assemblages is established. Surprisingly, a similarity of the Toarcian taxonomic diversity structure to that of the Early Triassic was noted, which suggests that the superfamilies which were important for species diversity after the Permian/Triassic catastrophe also found the post-Early Jurassic mass extinction conditions favorable. The total foraminiferal diversity decreased in the Early Toarcian by 1.8 times at the species level, but only by 1.2 times at the generic level (Ruban & Tyszka, 2005). The R-method indicates a strong turnover among species directly at the Pliensbachian/Toarcian boundary, whereas the same turnover among genera was somewhat delayed, occuring in the Middle Toarcian. The Rst-method permits me to document a very prominent turnover at the time of the mass extinction. The value of the Rst index is lowest in the Jurassic. Thus, my studies imply an evident occurrence of the Early Jurassic mass extinction in the Caucasus. The regional record provides evidence that oxygen depletion (related to the oceanic anoxia) was one of the probable explanations of this catastrophic event. A difference in the regional sea-level changes from those documented globally does not permit one to consider them as a possible trigger of the mass extinction, at least in the Caucasus (Ruban, 2004; Ruban & Tyszka, 2005).

5.5. Jurassic/Cretaceous mass extinction

The Jurassic/Cretaceous mass extinction stressed brachiopod assemblages strongly. Their total diversity decreased by about 10 times (Ruban, 2006a) with just a few species known from the Berriasian. Despite their recovery during the Valanginian-Hauterivian time interval, the pre-extinction diversity was never reached again. Thus, one may hypothesize that the Jurassic/Cretaceous mass extinction was a prelude to the final brachiopod collapse in the Northern Caucasus.

5.6. Aalenian event

In addition to these regional signatures of the well-known mass extinctions, it appears that the regional data provide evidence for a new mass extinction, which occurred in the Aalenian. Brachiopods almost disappeared in the Early Aalenian, but recovered rapidly in the Late Aalenian (Ruban, 2004). The total species diversity of foraminifera declined by 1.7 times in the Late Aalenian, whereas the total generic diversity decreased throughout the entire Aalenian (Ruban & Tyszka, 2005). The assemblage turnover was as large as that at the Pliensbachian/Toarcian boundary. The gradual recovery embraced the Early Bajocian, and it did not recompensate for the diveristy loss. Evidently, data from only one region is not enough to speculate about new mass extinctions. However, brachiopods collapsed during the Aalenian in the Bakony Mountains of Hungary and in the Swiss Alps, whereas foraminiferal assemblages were stressed in the Pieniny Klippen Belt of the Carpathians and probably in Spain. The likely cause of this event was also an oxygen depletion.

5.7. Geohistorical study of Mesozoic mass extinctions

Besides an analysis of regional data, I attempt a geohistorical investigation of the data published in the middle of the XIX century by A. d'Orbigny (Ruban, 2005b). Their quantitative assessment permitted me to conclude that almost all Mesozoic mass extinctions (Triassic/Jurassic, Jurassic/Cretaceous, Aptian, Cenomanian/Turonian, and Cretaceous/Paleogene) might have been documented already 150 years ago. Thus, despite a remarkable growth in the available paleontological information and the description of thousands of new species, the quality of the fossil record necessary to identify mass extinctions did not change significantly. This conclusion is very significant for our understanding of the completeness of the fossil record and its further changes.

6. Sea-level fluctuations

An investigation of sea-level changes is an important clue to the understanding of the regional Phanerozoic environmental changes and biotic evolution. The data available from the Caucasus reveal the regional transgressions, regressions, and basin deepening/shallowing events for the Cambrian-Jurassic time interval.

6.1. Facies analysis

All constraints are based on a careful facies analysis. Recognition of the general facies types is suitable for the attempted studies, and each facies is interpreted within a set of diverse geological information, which included lithology, sedimentary structures and textures, fossil assemblages, relationship with contemporary facies in adjacent areas, and relationship of facies in a stratigraphic succession etc. (Ruban, 2007a). Mixing or misinterpretation of facies due to similar lithological peculiarities is avoided. For example, a clear distinction between the Norian shelfal siliciclastics and the underlying Ladinian-Carnian flysch deposits is made despite their general similarity. Another example comes from the Early Jurassic, where marine and non-marine strata both containing abundant plant remains were distinguished. Geospatial analysis of facies, used to reveal the basin dynamics, takes into account possible deviations of trends observed in log or on a regional scale from some stereotypic assumptions. A very typical example is a deposition of evaporites in the Late Jurassic. Although evaporite sedimentation is often linked to sea-level lowstands, this was associated with a highstand (that occured just after a transgression maximum) in the Caucasus. Moreover, evaporitic sedimentation did not prevent the growth of coral reefs (a comparable situation is also known from the Miocene of the Mediterranean), and, thus, a carbonate platform became something like a substrate for the development of an evaporitic basin. Moreover, a consideration and a semi-quantitative or quantitative analysis of facies is done en-masse, which a priori diminishes the likelyhood of interpretation errors.

The sets of interpreted and compiled facies data (e.g., Ruban, 2007a) can be seen to be important. They may serve for further quantitative interpretations, which would permit to delineate particular surfaces (maximum flooding surfaces or sequence boundaries) or to discuss the sedimentary input and its influences on the basin dynamics. A global tracing of hiatuses (Zorina et al., 2008) suggests the importance of such studies for an evaluation of global eustatic changes and planetary-scale sedimentary evolution.

6.2. Paleozoic sea level

An analysis of distribution and a facies interpretation of the Paleozoic deposits in the Greater Caucasus permits the construction of paleogeographic frameworks for 9 time slices, and enables documentation of the principal changes of the shoreline (Ruban, 2007d). The Greater Caucasus was embraced by the sea in the Cambrian, which was followed by an Ordovician regression. A new transgression occurred in the Early Silurian and the shoreline along the northern border of the Greater Caucasus remained stable until the Late Devonian, whereas an opening of a new Paleotethys Ocean occurred in the south. A strong Famennian transgression led to the drowning of the entire region. The sea regressed in the Mississippian and occupied a restricted area until the Late Permian, when a new transgression took place. This general picture is detailed by the study of three principal Late Paleozoic transgressive episodes, which took place in the Lochkovian, the Frasnian-Famennian, and the Changhsingian (Ruban, 2007e). The second of them was the largest. All these episodes occurred at times of global sea-level rise, which implies their eustatic origin. Although regional tectonics did not affect them greatly, local tectonic activity may explain why evidence for the other global sea-level rises do not appear in the Greater Caucasus. An interesting observation is that these three transgressions coincided with important episodes of carbonate deposition.

6.3. Triassic sea level

The Triassic sea-level changes are reconstructed with precision (Ruban. 2008). A rapid transgression took place already in the Early Induan and the position of the shoreline remained stable until the Anisian, when its stepwise basinward shift took place. The next transgression, although smaller in its extent, began in the Ladinian, whereas a regressive episode is known from the mid-Carnian. A very strong trangression took place in the Early Norian with a peak in the middle of this stage, and a regressive episode embraced the Middle-Late Rhaetian. During the entire Triassic, the basin was, however, of a shallow-water character, with a unique exception. The Ladinian transgression coincided with a prominent deepening pulse. Both global eustasy and regional tectonic activity controlled these basin dynamics, and some major global sea-level changes are depicted in the regional record.

6.4. Jurassic sea level

The dynamics of the Caucasian basins in the Jurassic are reconstructed semi-quantitatively on the basis of a careful facies analysis in all particular areas of the region (Ruban, 2006d, 2007a). Some special attention is also paid to the Laba-Malka area (Ruban, 2004; Ruban & Tyszka, 2005). A stepwise transgression took place throughout the entire Late Sinemurian-Toarcian interval with a peak at the Toarcian-Aalenian transition. It was followed by a shorter regression in the Aalenian. Then, a rapid transgression occurred in the Early-Middle Bajocian to be followed by a longer regressive episode in the Late Bajocian-Bathonian. A major landward shoreline shift was realized during the Callovian-Oxfordian, and the Kimmeridgian was generally a time of maximum extent of marine environments, although interrupted by a short-term regressive episode. A profound regression occurred in the Tithonian. The changes in the average basin depth differed significantly. Three deepening pulses occurred in the Pliensbachian, the Late Aalenian, and the Late Bathonian. Despite a large transgression in the Late Jurassic, the Caucasian basins did not become deeper. These conclusions are confirmed by the specific study of the Greater Caucasus Basin (Ruban, 2007a). It is found that both eustasy and regional tectonic activity controlled the reported transgressions, regressions, and changes in the basin depth. The most interesting example of a dissimilarity of global and regional records comes from the Toarcian, where a significant delay of a transgression in comparison to the global and other regional records is documented.

6.5. General conclusions

The attempted studies of the sea-level changes and a comparison of their results with the global constraints permit some very important conclusions. First, it becomes evident that transgressions-regressions differed from the changes in the basin depth. This observation is of great methodological importance, because it makes urgent a constraint of two individual curves for every particular region. Secondly, both eustasy and tectonics are important factors for the dynamics of the basin shoreline and depth. It is interesting, that although the Greater Caucasus and also the entire Caucasus remained active regions throughout the analyzed Cambrian-Jurassic interval, the signatures of many globally-recognized eustatic events are clear from the studied territory. To discuss the importance of sedimentary input requires some further modelling. However, it appears that sedimentary input played a lesser role in its influence on transgressions/regressions (except in the case of deltaic systems) than on deepening/shallowing episodes. Finally, the role of sea-level changes in the fossil diversity dynamics is discussed (Ruban, 2004, 2006a, 2006d, 2007a, 2008; Ruban & Tyszka, 2005). Some positive relationships are found. All Paleozoic transgressions coincided with biotic radiations. The episodes of reefal growth in the Late Devonian, the Late Permian, the Late Triassic, and the Late Jurassic all corresponded to transgressions. An abrupt deepening of the basin in the Ladinian stressed the marine fauna and led to a total disappearance of brachiopods. However, the only fossil group, whose diversity changes were well connected to the interaction of transgressions and regressions was the ammonites. Responses of bivalve, brachiopod, belemnite, and foraminifer diversity were more complicated and did not demonstrate simple relationships with basin dynamics. The most surprising is the fact, that links between fossil diversity dynamics and eustatic changes are more evident on a global scale than on a regional level (Ruban, 2007a).

7. Tectonics

7.1. Paleozoic-Triassic terrane model

A critical revision of the available lithological, paleontological, and other kinds of geological data from the Caucasus and adjacent areas with regard to the modern plate tectonic reconstructions for the Paleozoic and Mesozoic allows one to reconsider the tectonic evolution of the study territory and to propose a totally new model. The latter is discussed by Ruban (2006c, 2007b,d,f) and Ruban et al. (2007). This model is based on two major observations, namely (1) a similarity of the Late Paleozoic sedimentary and fossil records of the Greater Caucasus and some Hunic terranes, including the Carnic Alps and the Bohemian Massif, and (2) evidence for an arc-arc collision in the Caucasus during the Middle Jurassic. The similarity of the geological histories of the Greater Caucasus, the Carnic Alps, and the Bohemian Massif suggests their mutual proximity in the Late Paleozoic. If this was so, the Greater Caucasus Terrane was a part in the chain of the Hunic terranes derived from the Gondwanan margin in the middle of the Silurian due to the opening of the Paleotethys Ocean.

In the Late Devonian, these terranes became anchored at the Laurussian margin in the Proto-Alpine area. This interpretation raises the question as to how the Greater Caucasus could have reached its present position far to the east. An appropriate explanation is given by a consideration of the major shear zone along the northern margin of the Paleotethys. Dextral displacements along this zone during the Late Paleozoic-Middle Triassic led to terrane stacking in the Proto-Alpine area. However, the direction of displacements along this zone changed radically from dextral to sinistral in the Middle-Late Triassic. This provided a mechanism to displace the Greater Caucasus Terrane to the east, where it collided with the Russian Platform. Such a scenario explains also a major unconformity documented at the Triassic-Jurassic transition (Ruban, 2006b). This shear zone was a part in the net of Intrapangaean shear zones, which stretches across Western Europe, eastern North America, South America, South Africa, and Australia. The situation at the northern Paleotethyan margin in the Late Paleozoic-Early Mesozoic was a bit similar to that at the western margin of North America in the Mesozoic-Cenozoic (Ruban, 2007f). Not only the Caucasus, but all of the southern periphery of the Russian Platform was influenced by the activity along the shear zone. Right-lateral displacements led to the derivation of the Ukrainian Block from the Russian Platform and opening of a rapidly-subsiding coal-bearing basin (Donbass) in between (Ruban & Yoshioka, 2005). A change to left-lateral displacements in the Middle Triassic resulted in the closure of this basin and a local deformation phase, which created the Donbass fold belt. This model is well supported by the most recent dating of the tectonic activity in the Donbass. Moreover, the Paleozoic geodynamics of the Greater Caucasus Terrane, which was a part of the Hunic Superterrane, and of the Lesser Caucasus Terrane, which was a part of the younger Cimmerian Superterrane, was linked closely to that of many other Middle Eastern terranes (Ruban et al., 2007). While the Greater Caucasus bears an affinity to the Pontides and probably the Alborz terranes, the geological history of the Lesser Caucasus was more linked to that of Central Iran and the Taurides.

7.2. Jurassic geodynamic reconstructions

Three Jurassic geodynamic reconstructions (for the Late Toarcian, the Early Bajocian, and the Midle Oxfordian) are constrained on the basis of a careful investigation of data from all Caucasian areas (Ruban, 2006c). Despite minor contrary details, they allow two very important conclusions. The first one concerns an arc-arc collision in the Middle Jurassic. There is evidence for a joining of the Northern and Southern Transcaucasian arcs since at least the Bajocian. This tectonic event may explain the mid-Jurassic major unconformity (Ruban, 2007b) and also shed light on a poorly defined mid-Jurassic orogeny hypothesized earlier, but which has remained unexplained. My reconstructions indicate the presence of a large Caucasian Sea connected with the Neotethys Ocean in the south and other seas to the west and the east by long seaways. This provides a much needed clue to explain the style of the biotic evolution in the Caucasus and its possible relationships to that in Europe and the Middle East.

7.3. Phanerozoic phases of the tectonic evolution of the Greater Caucasus

An overall examination of geological data from the Greater Caucasus permitted the identification of 7 phases in its tectonic evolution, namely the Gondwanan Phase (Cambrian-Ludlow), the Hunic Phase (Ludlow-Devonian), the Proto-Alpine Phase (Carboniferous-Middle Triassic), the Left-Shear Phase (Late Triassic-Earliest Jurassic), the Arc Phase (Jurassic-Eocene), the Paratethyan Phase (Oligocene-Miocene), and the Transcaucasus Phase (Pliocene-Recent) (Tawadros et al., 2006). These phases are also compared with those established in the Northeastern African basins in order to reveal some similar patterns, which permits one to outline some new perspectives for hydrocarbon exploration in the Greater Caucasus (Tawadros et al., 2006).

8. Concluding remarks

The attempted study in this thesis permits me to bring the understanding of the Phanerozoic history of the Caucasus and some adjacent areas to a new level of complexity. Stratigraphic constraints strengthen the precision of all further interpretations. An analysis of fossil diversity reveals the regional appearance of mass extinctions and other crises. An interpretation of the sea-level changes allows explanation of the regional biotic and entire geological evolution in the light of transgressions, regressions, and changes in basin depth. Tectonic constraints help to understand how all regional data can be interpreted in the context of the entire northern Paleo- (and Neotethyan) margin, and which inter-regional correlations seem to be the most promising. Thus, all these studies contribute to a comprehensive synthesis of the Phanerozoic environmental changes in the Caucasus and adjacent areas.

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Ruban, D.A. 2004. Diversity dynamics of Early-Middle Jurassic brachiopods of Caucasus, and the Pliensbachian-Toarcian mass extinction. Acta Palaeontologica Polonica. 49: 275-282.

(keywords: brachiopods, diversity, mass extinction, Jurassic, Caucasus)

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(keywords: seaway, basin, arc-arc collision, Jurassic, Caucasus)

Ruban, D.A. 2006d. Taxonomic diversity dynamics of the Jurassic bivalves in the Caucasus: regional trends and recognition of global patterns. Palaeogeography, Palaeoclimatology, Palaeoecology. 239: 63-74.

(keywords: bivalves, diversity, mass extinction, Jurassic, Caucasus)

Tawadros, E., Ruban, D. & Efendiyeva, M. 2006. Evolution of NE Africa and the Greater Caucasus: Common Patterns and Petroleum Potential. The Canadian Society of Petroleum Geologists, the Canadian Society of Exploration Geophysicists, the Canadian Well Logging Society Joint Convention. May 15-18, 2006. Calgary. P. 531-538. [extended abstract]

(keywords: terrane, shear zone, petroleum potential, NE Africa, Greater Caucasus)

Ruban, D.A. 2007a. Jurassic transgressions and regressions in the Caucasus (northern Neotethys Ocean) and their influences on the marine biodiversity. Palaeogeography, Palaeoclimatology, Palaeoecology. 251: 422-436.

(keywords: transgression, regression, eustasy, diversity, Jurassic, Caucasus)

Ruban, D.A. 2007b. Major Paleozoic-Mesozoic Unconformities in the Greater Caucasus and Their Tectonic Re-Interpretation: A Synthesis. GeoActa. 6: 91-102.

(keywords: unconformity, Paleozoic, Mesozoic, Greater Caucasus)

Ruban, D.A. 2007c. Taxonomic diversity structure of brachiopod associations at times of the early Mesozoic crises: evidence from the Northern Caucasus, Russia (northern Neotethys Ocean). Paleontological Research. 11: 349-358.

(keywords: brachiopods, diversity, Jurassic, mass extinction, Caucasus, Triassic)

Ruban, D.A. 2007d. Paleozoic palaeogeographic frameworks of the Greater Caucasus, a large Gondwana-derived terrane: consequences from the new tectonic model. Natura Nascosta. 34: 16-27.

(keywords: palaeogeography, terrane, carbonate platform, Paleozoic, Greater Caucasus)

Ruban, D.A. 2007e. Late Paleozoic Transgressions in the Greater Caucasus (Hun Superterrane, Northern Palaeotethys): Global Eustatic Control. Cadernos do Laboratorio Xeolóxico de Laxe. 2007. 32: 13-24.

(keywords: palaeogeography, Late Paleozoic, sea level, eustasy, Greater Caucasus)

Ruban, D.A. 2007f. The southwestern margin of Baltica in the Paleozoic-early Mesozoic: Its global context and North American analogue. Natura Nascosta. 35: 24-35.

(keywords: shear zone, Baltica, North America, Paleozoic, Mesozoic)

Ruban, D.A., Al-Husseini, M.I. & Iwasaki, Y., 2007. Review of Middle East Paleozoic Plate Tectonics. GeoArabia. 12: 35-56.

(keywords: terrane, Gondwana, Arabian Plate, Middle East)

Ruban, D.A. 2008. Evolutionary rates of the Triassic marine macrofauna and sea-level changes: evidences from the Northwestern Caucasus, Northern Neotethys (Russia). Palaeoworld. 17: 115-125.

(keywords: evolutionary rate, macrofauna, sea level, Triassic, Caucasus)

Ruban, D.A. & van Loon, A.J. 2008. Possible pitfalls in the procedure for paleobiodiversity-dynamics analysis. Geologos. 14: 37-50.

(keywords: diversity, extinction, Lazarus taxa)

Gutak, Ja.M., Tolokonnikova, Z.A. & Ruban, D.A. 2008. Bryozoan diversity in Southern Siberia at the Devonian-Carboniferous transition: new data confirm a resistivity to two mass extinctions. Palaeogeography, Palaeoclimatology, Palaeoecology. 264: 93-99.

(keywords: bryozoans, diversity, mass extinction, transgression, regression, basin depth, Southern Siberia, Paleozoic)

Zorina, S.O., Dzyuba, O.S., Shurygin, B.N. & Ruban, D.A. 2008. How global are the Jurassic-Cretaceous unconformities? Terra Nova. 20: 341-346.

(keywords: unconformity, eustasy, correlation, Jurassic, Cretaceous)

Suggestion of order in which papers should be read

The references are presented according to the order of subjects discussed in the summary. A relative importance of each paper for the particular subject is provided. Most of the papers, however, deal with several subjects, and these relationships can be deduced from the main text of this summary. For each subject detailed below, the papers are aligned along the course of the geologic time.

Stratigraphy

Ruban, D.A. 2007b. Major Paleozoic-Mesozoic Unconformities in the Greater Caucasus and Their Tectonic Re-Interpretation: A Synthesis. GeoActa. 6: 91-102.

Zorina, S.O., Dzyuba, O.S., Shurygin, B.N. & Ruban, D.A. 2008. How global are the Jurassic-Cretaceous unconformities? Terra Nova. 20: 341-346.

Ruban, D.A. 2005a. The Upper Miocene of the Rostov Dome (Eastern Paratethys): Implication of the chronostratigraphy and bivalvia-based biostratigraphy. Geološki anali Balkanskoga poluostrva. 66: 9-15.

Fossil diversity

Ruban, D.A. & van Loon, A.J. 2008. Possible pitfalls in the procedure for paleobiodiversity-dynamics analysis. Geologos. 14: 37-50.

Gutak, Ja.M., Tolokonnikova, Z.A. & Ruban, D.A. 2008. Bryozoan diversity in Southern Siberia at the Devonian-Carboniferous transition: new data confirm a resistivity to two mass extinctions. Palaeogeography, Palaeoclimatology, Palaeoecology. 264: 93-99.

Ruban, D.A. 2006b. Diversity dynamics of the Triassic marine biota in the Western Caucasus (Russia): A quantitative estimation and a comparison with the global patterns. Revue de Paléobiologie. 25: 699-708.

Ruban, D.A. 2006a. Diversity changes of the Brachiopods in the Northern Caucasus: a brief overview. Acta Geologica Hungarica. 49: 57-71.

Ruban, D.A. 2006d. Taxonomic diversity dynamics of the Jurassic bivalves in the Caucasus: regional trends and recognition of global patterns. Palaeogeography, Palaeoclimatology, Palaeoecology. 239: 63-74.

Mass extinctions

Ruban, D.A. 2007c. Taxonomic diversity structure of brachiopod associations at times of the early Mesozoic crises: evidence from the Northern Caucasus, Russia (northern Neotethys Ocean). Paleontological Research. 11: 349-358.

Ruban, D.A. 2004. Diversity dynamics of Early-Middle Jurassic brachiopods of Caucasus, and the Pliensbachian-Toarcian mass extinction. Acta Palaeontologica Polonica. 49: 275-282.

Ruban, D.A. & Tyszka, J. 2005. Diversity dynamics and mass extinctions of the Early-Middle Jurassic foraminifers: A record from the Northwestern Caucasus. Palaeogeography, Palaeoclimatology, Palaeoecology. 222: 329-343.

Ruban, D.A. 2005b. Mesozoic marine fossil diversity and mass extinctions: an experience with the middle XIX century paleontological data. Revue de Paléobiologie. 24: 287-290.

Sea-level fluctuations

Ruban, D.A. 2007d. Paleozoic palaeogeographic frameworks of the Greater Caucasus, a large Gondwana-derived terrane: consequences from the new tectonic model. Natura Nascosta. 34: 16-27.

Ruban, D.A. 2007e. Late Paleozoic Transgressions in the Greater Caucasus (Hun Superterrane, Northern Palaeotethys): Global Eustatic Control. Cadernos do Laboratorio Xeolóxico de Laxe. 2007. 32: 13-24.

Ruban, D.A. 2008. Evolutionary rates of the Triassic marine macrofauna and sea-level changes: evidences from the Northwestern Caucasus, Northern Neotethys (Russia). Palaeoworld. 17: 115-125.

Ruban, D.A. 2007a. Jurassic transgressions and regressions in the Caucasus (northern Neotethys Ocean) and their influences on the marine biodiversity. Palaeogeography, Palaeoclimatology, Palaeoecology. 251: 422-436.

Tectonics

Tawadros, E., Ruban, D. & Efendiyeva, M. 2006. Evolution of NE Africa and the Greater Caucasus: Common Patterns and Petroleum Potential. The Canadian Society of Petroleum Geologists, the Canadian Society of Exploration Geophysicists, the Canadian Well Logging Society Joint Convention. May 15-18, 2006. Calgary. P. 531-538. [extended abstract]

Ruban, D.A., Al-Husseini, M.I. & Iwasaki, Y., 2007. Review of Middle East Paleozoic Plate Tectonics. GeoArabia. 12: 35-56.

Ruban, D.A. & Yoshioka, S. 2005. Late Paleozoic – Early Mesozoic Tectonic Activity within the Donbass (Russian Platform). Trabajos de Geología. 25: 101-104.

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Cariou, E. & Hantzpergue, P. (Eds.). 1997. Biostratigraphie du Jrassique ouest-européen et méditerranéen: zonations parallèles et distribution des invertébrés et microfossiles. Bulletin du Centre Recherche Elf Exploration et Production. 17: 1-440.

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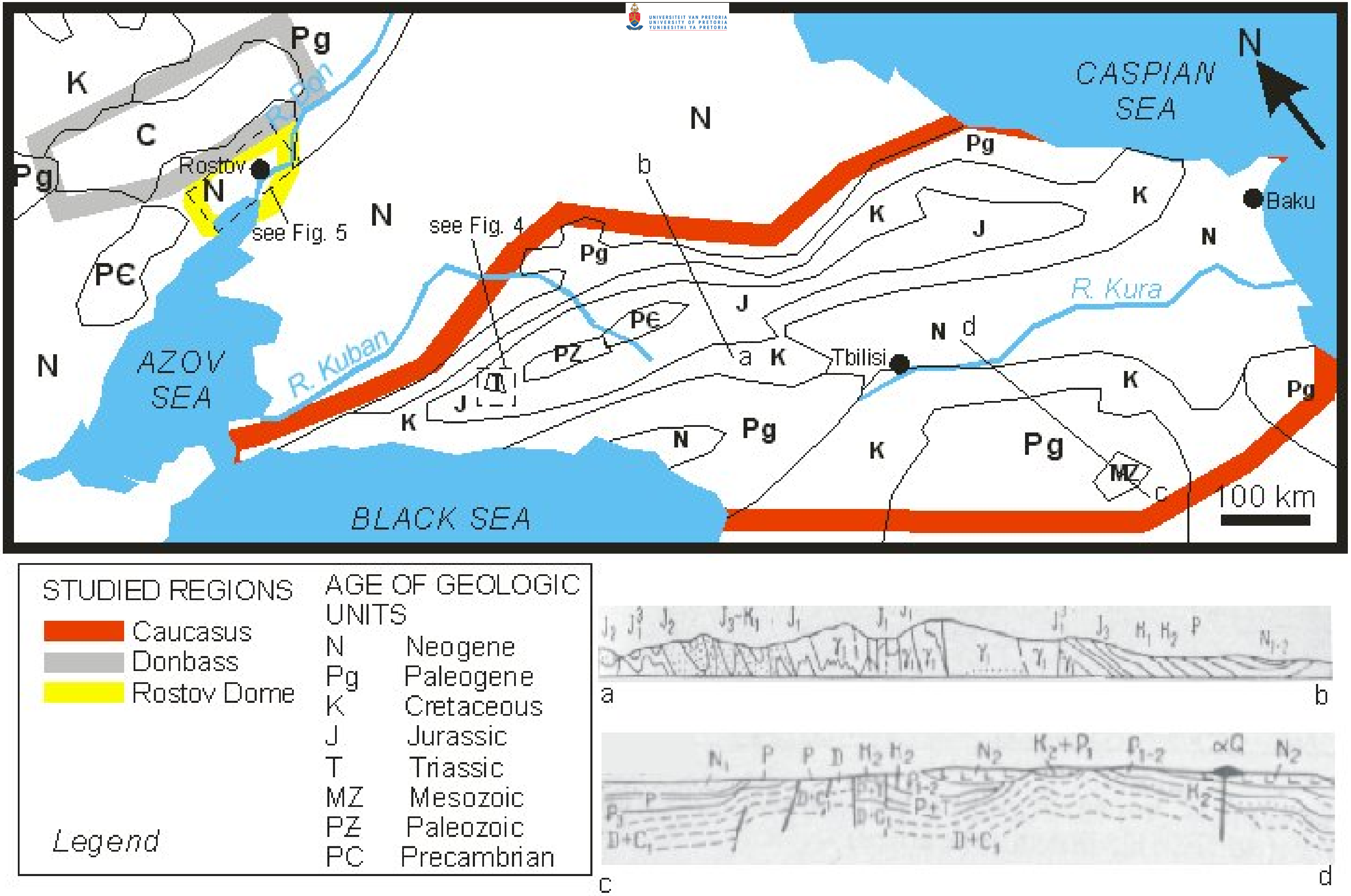
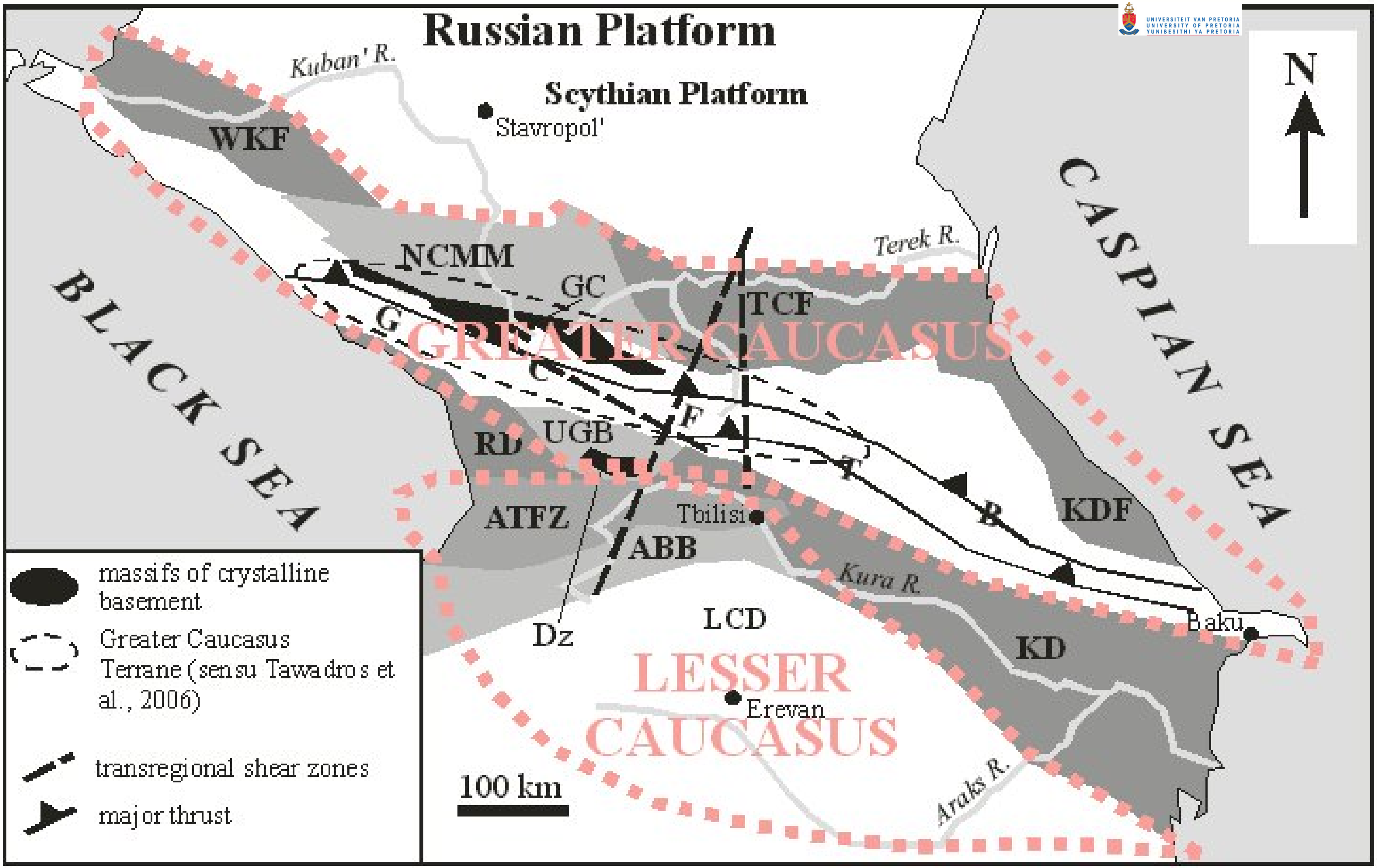
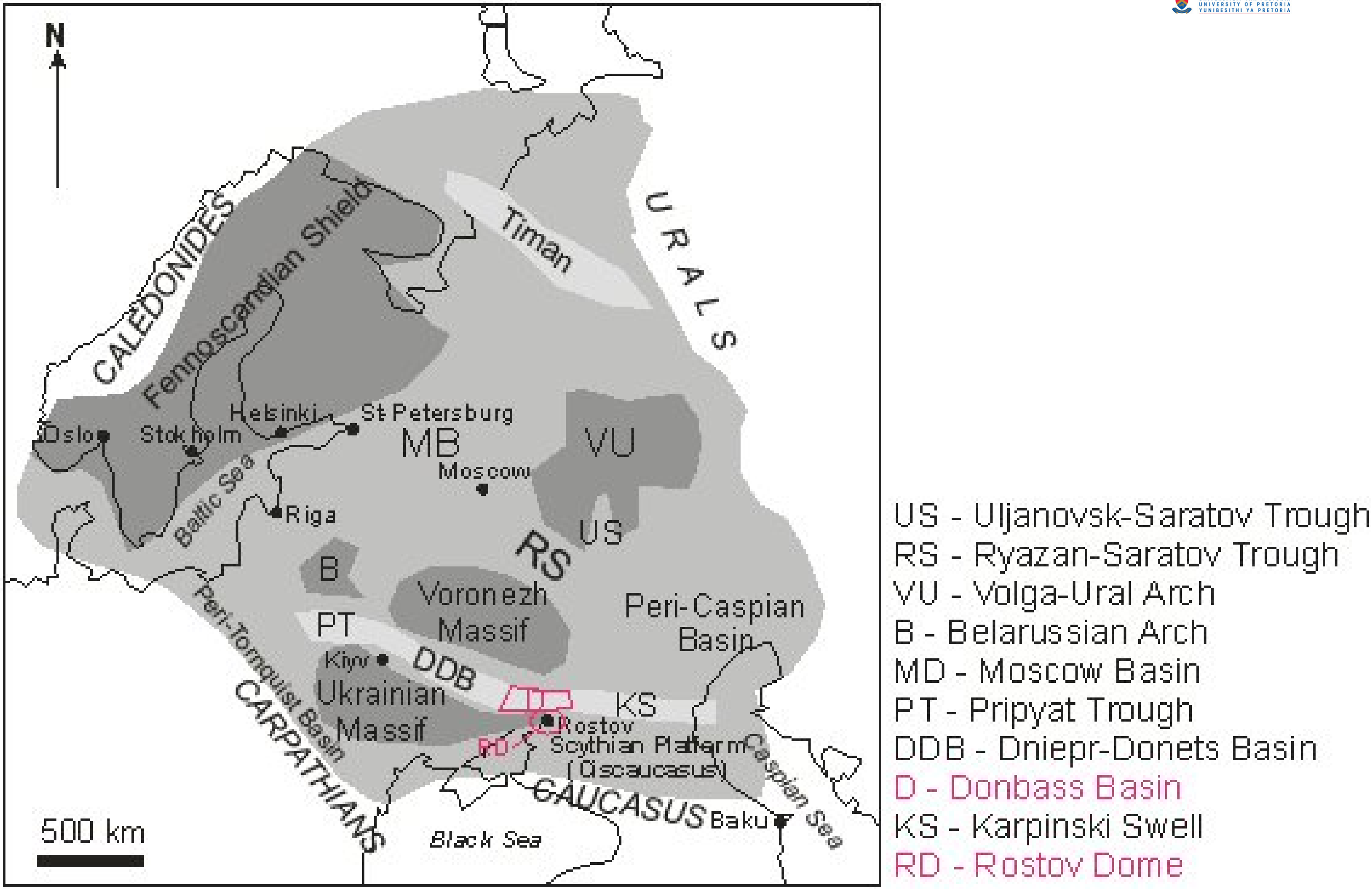


Figure 1. Geological map of the Caucasus and adjacent areas and a location of studied regions.



Abbreviations: ABB - Arthvin-Bolnisi Block, ATFZ - Adjara-Thrialethian Fold Zone, Dz - Dzirula Massif, GC - Greater Caucasian Massif, GCFTB - Greater Caucasian Fold and Thrust Belt, KD - Kura Depression, KDF - Kusar-Divichian Foredeep, LCD - Lesser Caucasus Domain, NCMM - North Caucasian Marginal Massif, RD - Rioni Depression, TCF - Terek-Caspian Foredeep, UGB - Uplifted Georgian Block, WKF - West Kubanian Foredeep. Different patterns are used to differentiate the units.

Figure 2. Main tectonic units of the Caucasus (modified after Gamkrelidze, 1997).



US - Uljanovsk-Saratov Trough
RS - Ryazan-Saratov Trough
VU - Volga-Ural Arch
B - Belarussian Arch
MD - Moscow Basin
PT - Pripyat Trough
DDB - Dniepr-Donets Basin
D - Donbass Basin
KS - Karpinski Swell
RD - Rostov Dome

Figure 3. Main tectonic units of the Russian Platform (modified after Nikishin et al., 1996).

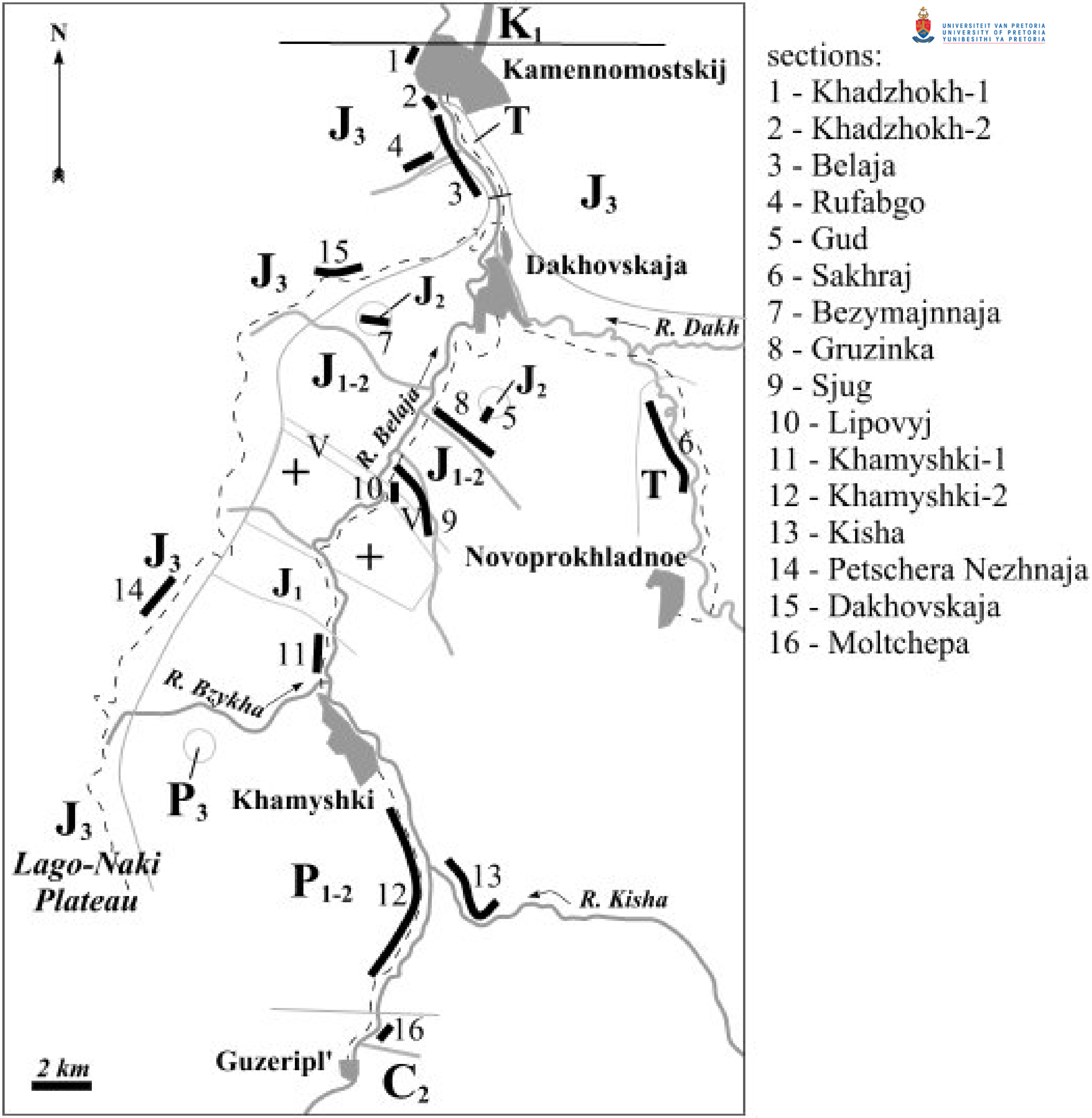


Figure 4. Location of sections in the Laba-Malka area.

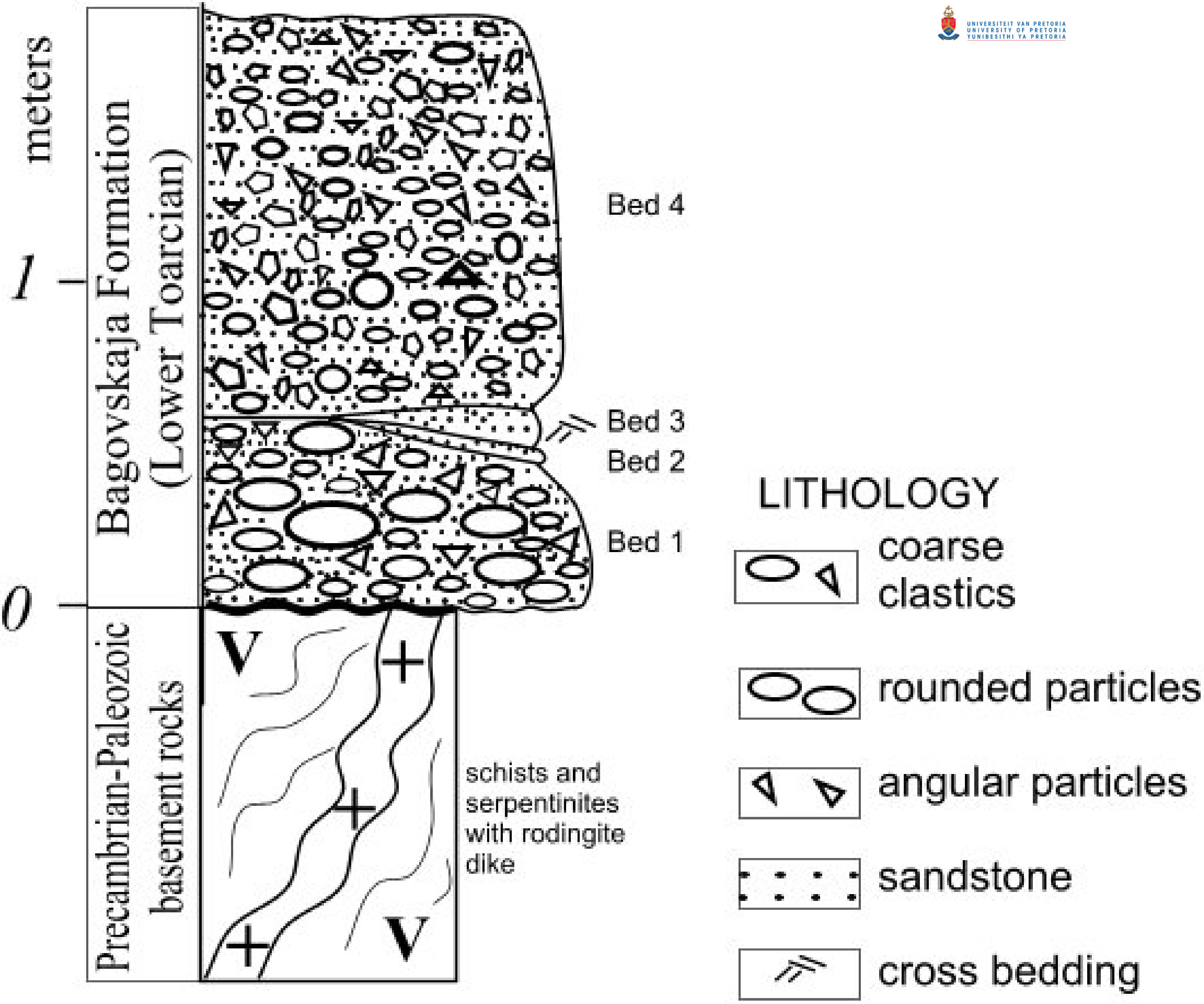


Figure 5. The Lipovyj section (see Fig. 4 for location).

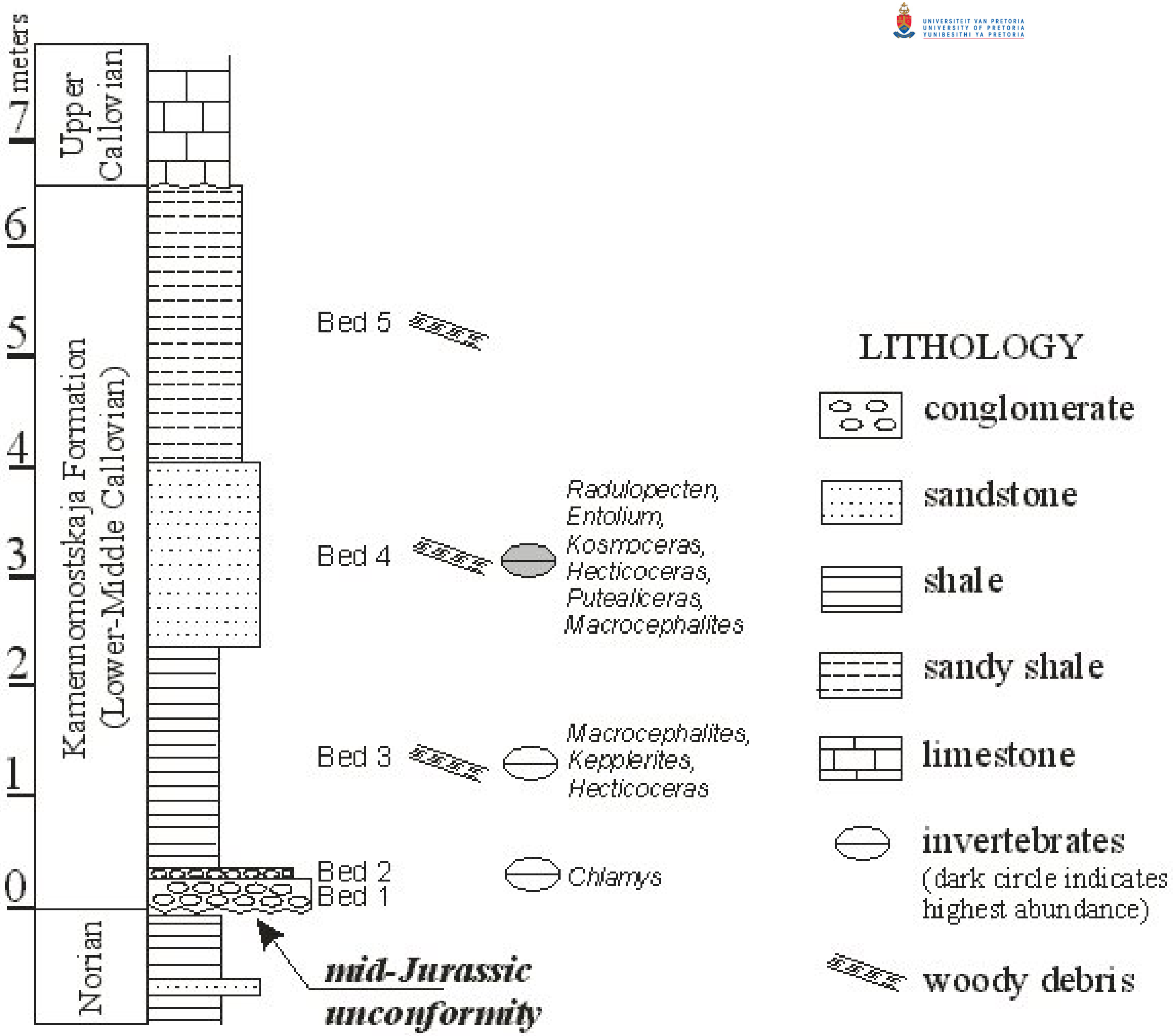


Figure 6. The Khadzhokh-2 section (see Fig. 4 for location) (modified from Ruban, 2007b).

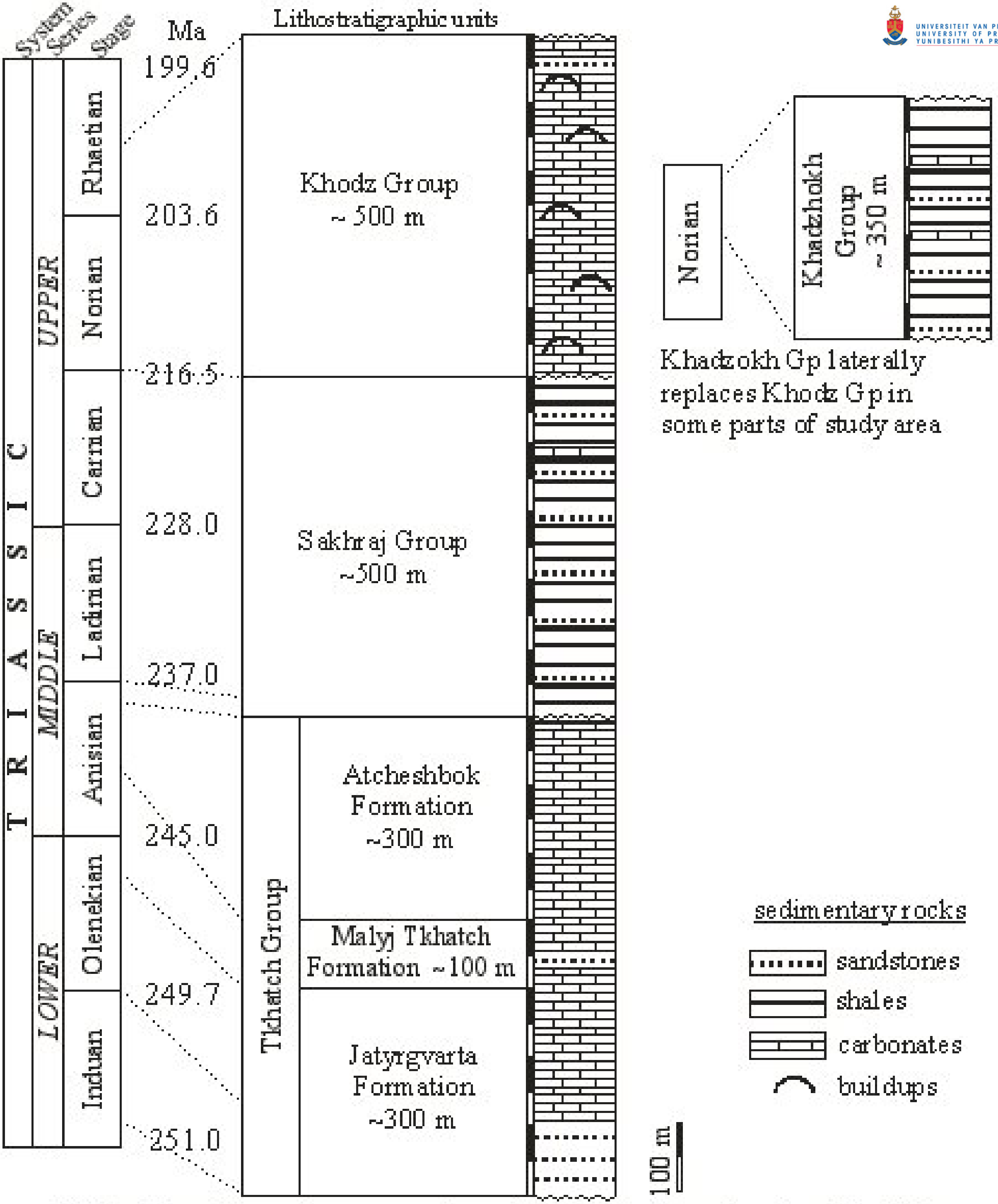


Figure 7. Stratigraphy and a composite section of the Triassic deposits of the western part of the Greater Caucasus (after Ruban, 2008; based on data from a number of sections and outcrops).

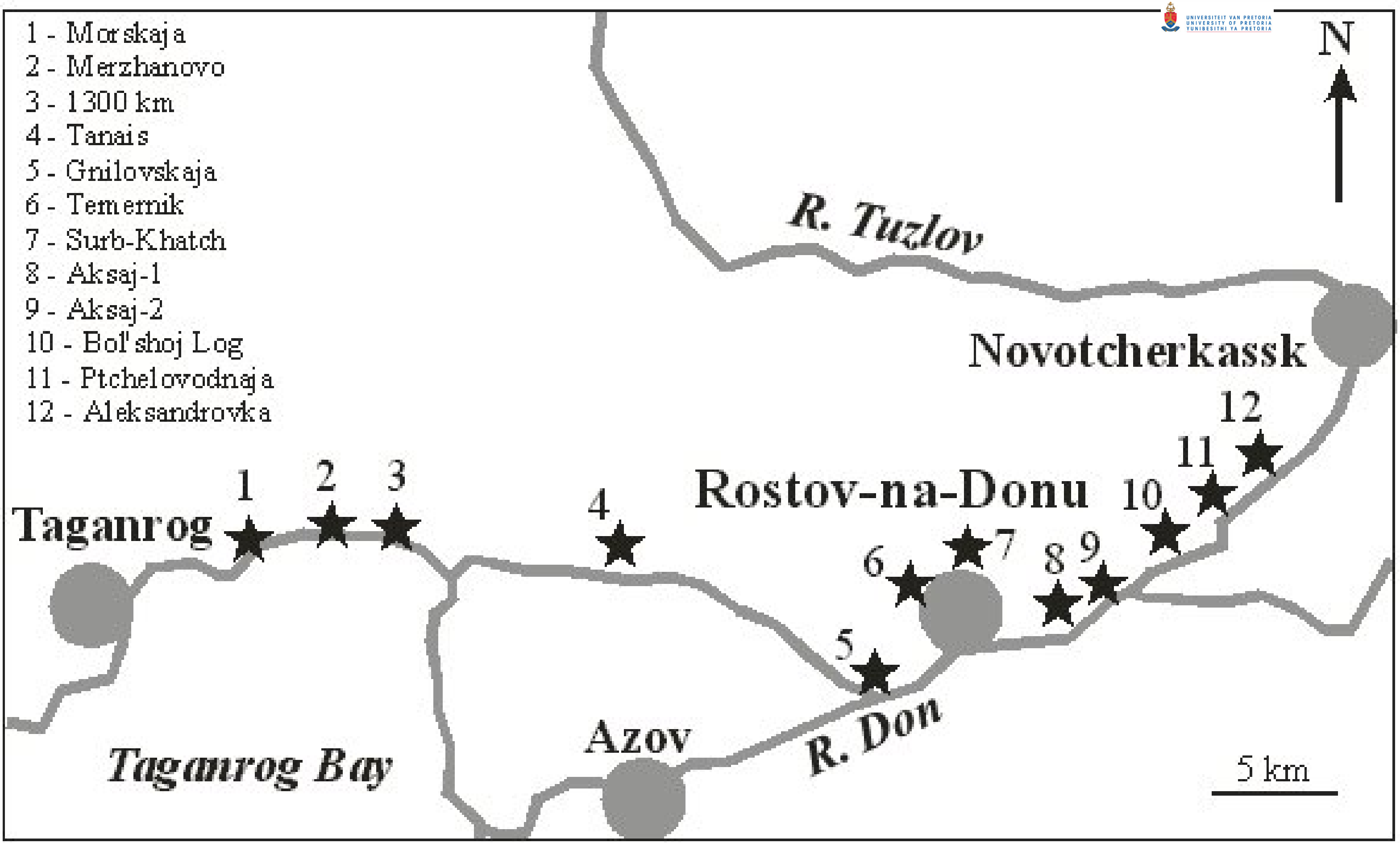


Figure 8. location of the studied Upper Miocene sections in the Rostov Dome.

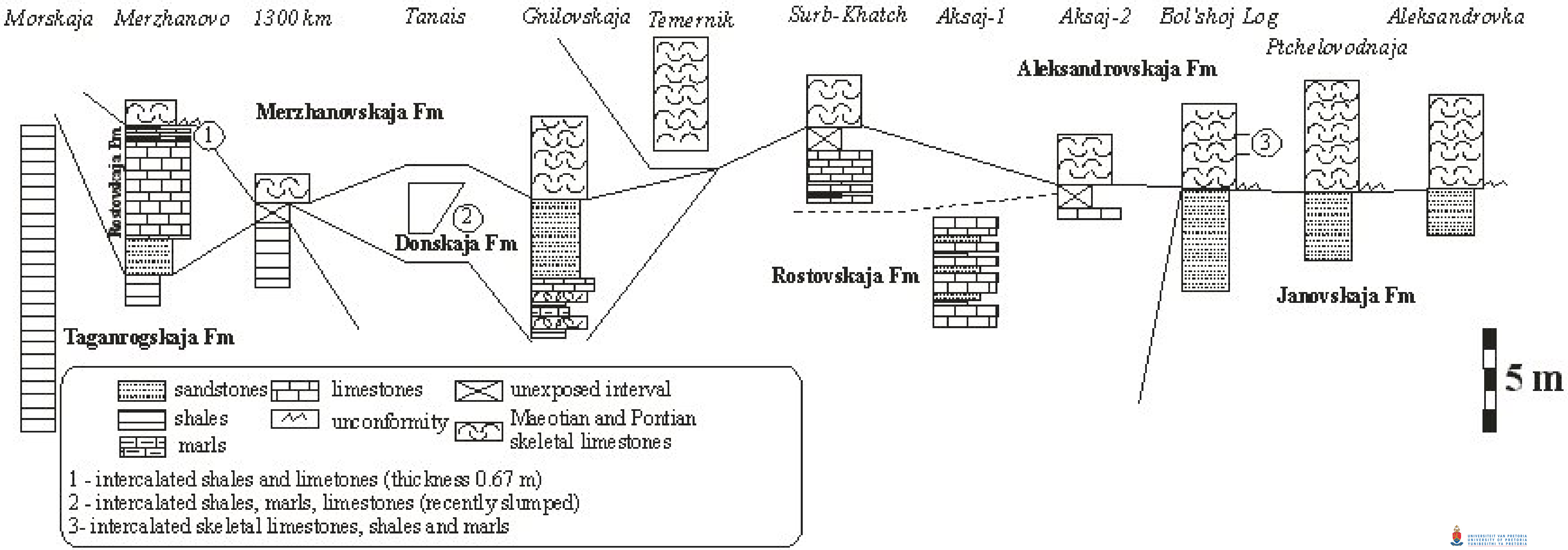


Figure 9. Correlation of the Upper Miocene sections of the Rostov Dome (after Ruban, 2005a). See Ruban (2005a) for precise formation ages.

Appendix 1. List of identified Jurassic bivalves from the Laba-Malka area (see Fig. 4 for section location).

BEZYMJANNAJA SECTION

Plagiostoma cf. *subrigidula* (Schlippe, 1888), *Plagiostoma* sp.

samples: BG-1, BG-2

lithostratigraphy: Bizhgon Member (crinoidal limestones)

age*: late Aalenian*

source: Ruban, D.A., Valentseva, D.R., Skljarov, V.V. & Khokhlatcheva, N.V. 2007. New data on the Aalenian-Bejocian stratigraphy in the River Belaja watershed (Northwestern Caucasus). In: Alekseev, A.S. (Ed.). Paleostrat-2007. MOIP, Moskva. pp. 22-23. (in Russian)

KHADZHOKH-2 SECTION

Chlamys (Chlamys) viminea (Sowerby, 1826), *Ctenostreon* sp., *Entolium demissum* (Phillips, 1829), *Entolium* sp., *?Entolium* sp., *?Pleuromya* sp., *?Pseudolimea* sp., *?Pseudolimea* cf. *Limea* sp., *Radulopecten (Radulopecten) subinaequicostatus* (Kasansky, 1909), *?Radulopecten (Radulopecten)* sp., *?Sublatachlamys* aff. *suchubika* Romanov, 1985.

samples: cl-03-X-01, cl-03-X-02, cl-03-X-03, cl-03-X-04, cl-03-X-05, cl-03-X-06, cl-03-X-07, cl-03-X-08, cl-03-X-09, cl-03-X-10, cl-03-X-11, cl-03-X-12, cl-03-X-14, cl-03-X-16, cl-03-X-20, cl-03-X-22, cl-03-X-24, cl-03-X-25

lithostratigraphy: Kamennomostskaja Formation (basal conglomerates and sandstones - beds 2 and 4)

age*: early-middle Callovian*

source: Ruban, D.A. 2005. New data on the Callovian fauna of the Northwestern Caucasus. Nautchnaja mysl' Kavkaza, Prilozhenie. 13: 138-140. (in Russian)

DAKHOVSKAJA SECTION

Entolium demissum (Phillips, 1829), *?Entolium* sp., *?Mytiloceramus* sp.

samples: cl-03-4500-01, cl-03-4500-03, cl-03-4500-04, cl-03-4500-05, cl-03-4500-06, cl-03-4500-07, cl-03-4500-08

lithostratigraphy: Gerpegemskaja Formation (lower part of carbonate succession)

age*: late Callovian*

source: Ruban, D.A. 2005. New data on the Callovian fauna of the Northwestern Caucasus. Nautchnaja mysl' Kavkaza, Prilozhenie. 13: 138-140. (in Russian)

Appendix 2. Compiled list of genera of the Triassic marine biota of the Western Caucasus (after Ruban, 2006b). Number of species in each genus in the Triassic stages is indicated.

Genera	Induan+Olenekian	Anisian	Ladinian	Carnian	Norian	Rhaetian
AMMONOIDS						
Acrochordiceras		3				
Aegeiceras		1				
Arcestes			1			3
Arpadites		1				
Badiotites			1			
Beyrichites		1				
Caucasites		2				
Cladiscites					2	2
Dieneroceras	1					
Flemingites	1					
Flexoptychites		1		1		
Gymnites		1				
Hollandites		3				
Japonites		1				
Joannites				1		
Juvavites					1	
Laboceras		3				
Leyophyllites		4				
Lobites					1	
Longobardites		1				
Megaphyllites		1			2	1
Mesocladiscites		1				
Monophyllites		1	3			
Nannites	1					
Owenites	3					
Paracladiscites						2
Paradanubites		2				
Paragoceras	1					
Parasageceras		1				
Parussuria	1					
Phyllocladiscites		2				
Pinacoceras					1	1
Placites					1	1
Proptychites	1					
Pseudosageceras	1					
Rhacophyllites					1	1
Smithoceras		1				
Sturia		2	1			

Subowenites	1					
Subvishnuites	1					
Wyomingites	1					
Xenodiscus			1			
ALGAE						
Lithotamnidium					1	
Spongiomorpha					4	
BRACHIOPODS						
"Rhynchonella"		1				
Abrekia	1					
Adygella				1	1	1
Adygelloides						1
Amphiclina					1	2
Ampliclinodonta					1	
Angustothyris		1				
Aulacothyropsis				1		4
Austriella					1	
Austrirhynchia						2
Balatonospira				1		
Bobukella				1	1	
Caucasorhynchia					1	1
Caucasothyris					1	
Coenothyris		2				
Costirhynchia		1		1		
Costispiriferina		1				
Crurirhynchia					1	
Crurithyris	1					
Cubanothyris					2	2
Decurtella		2				
Dinarispira		2				
Dioristella		1		1		
Euxinella					3	6
Fissirhynchia					1	1
Guseriplia						2
Holcorhynchella		1				
Koeveskallina		1				
Koninckina				1	1	
Laballa					2	3
Lepismatina						1
Lobothyris					2	
Majkopella						3
Mentzelia		1		2	1	2
Moisseievia					2	1
Neoretzia						3
Neowelerella	1					
Norella		1				
Oxycolpella					3	2
Pexidella		1			1	
Piarorhynchella		1				
Pseudocyrtina					1	

Pseudorugitella					2	2
Punctospirella		1				
Rhaetina				1	6	4
Rhimirhynchopsis					1	1
Robinsonella						1
Sinucosta		1			1	1
Spinolepismatina					1	
Sulcatinella		2				
Sulcatothyris				1		
Tetractinella		1				
Thecospira					1	
Thecospiropsis					1	
Triadithyris					2	1
Trigonirhynchella					1	2
Volirhynchia		2				
Wittenburgella					1	1
Worobievella					1	1
Zeilleria					3	6
Zugmayerella						1
BIVALVES						
Cassianella					1	
Claraia	4					
Daonella			2			
Halobia			1	5		
Hoernesia		1				
Indopecten					1	
Leda			1			
Limea		1				
Lyssochlamys				2		
Monotis					3	
Myophoria					1	
Mytilus		1				
Paleocardita					1	
Posidonia		1	1			
Pseudomonotis					1	
Schafhaeutlia		1				
Velopecten		1		1		
CORALS						
Astraeomorpha					2	1
Montlivaultia					1	
Rhabdophyllia					1	
Stephanocoenia					1	
Stylophyllopsis					2	
Thamnastraea					2	2
Thecosmilia					6	2
FORAMINIFERS						
"Frondicularia"						2
"Orthovertella"	1					
"Protonodosaria"	1					
"Tetrataxis"					1	1

Agathaminina					1	1
Ammobaculites		2		1	1	
Ammodiscus	1		2	1	1	
Angulodiscus					1	1
Arenovidalina		4				
Astacolus		2	1	3		
Auloconus					1	1
Aulotortus					3	3
Calcitornella		1	1			
Cornuloculina		1		1	1	1
Cornuspira				1		
Coronipora						1
Dentalina	3	5	6	4		
Diplotremmina		1		1		
Duostomina		1	1			
Duotaxis						1
Earlandia		1				
Galeanella					1	2
Gandinella	1					
Gaudryina				2		1
Gaudryinella						1
Glomospira		2				
Glomospirella		3				1
Hoyenella	1	1				2
Hyperammina		2		1		
Ichtyolaria				2		
Involutina					2	2
Labalina				1		
Lagena		1				
Lenticulina			4	3		2
Lingulina		2	4	1		
Marginulinopsis				1		
Meandrospira	2	3				
Miliolipora					1	1
Nodosaria	5	2		6		2
Nodosinella		1				
Ophthalmidium			1	2	3	4
Pachyphloides			1	2		
Pilammina		3				
Planiinvoluta					1	4
Pseudonodosaria		7	9	5		
Quinqueloculina			1			
Reophax			2			
Saccammina		1				
Semiinvoluta					1	1
Spiroplectammina		1		1		
Tetrataxis					1	
Tolypammina		3				
Triasina					1	
Trochammina		1		2	2	3

Trocholina		1	3	4
Turrispirillina				1
Vaginulina	1			
Vaginulinopsis	1			
SPONGES				
Hodsia			1	
Molengraffia			1	
Sahraja			1	

Appendix 3. Stratigraphic distribution of brachiopods in the Northern Caucasus (after Ruban, 2006a with modifications and additions). Number of species of each genera in the Triassic-Lower Cretaceous stages is indicated.

CAMBRIAN

Acrotreta gerassimovi Lermontova - Middle Cambrian

DEVONIAN

Cingulodermus ex gr. *superstes* (Barrande), *Clorinda pseudolinguifera* (Kozlowski) and *Janius* ex gr. *irbitensis* (Tschernyshova) - Lochkovian

Ivdelina (Procerulina) ex gr. *procerula* (Barrande) - Pragian

Atrypa posturalica Markowskii, *Gypidula comis* Owen, *Hypothyridina cuboides* Sowerby, *Spinatrypa* ex gr. *bifidaeformis* Tschernyshova - Frasnian

Cyrtospirifer verneuili (Murchison), *Cyrtospirifer* cf. *calcaratus* Sowerby, *Cyrtospirifer* cf. *archiaci* Murchison, *Cyrtospirifer* cf. *postarchiaci* Nalivkin, *Isopoma brachyptycta* (Schnur), *Productus* sp., *Productella* ex gr. *subaculeata* (Murchison), *Productella calva* var. *multispinosa* Sokolskaya, *Productella calva* var. *koscharica* Nalivkin, *Pugnax janischevskii* Rozman, *Rhipidiorhynchus* ex gr. *livonicus* Buch - Famennian

CARBONIFEROUS

Spirifer cf. *distans* Sowerby - Mississippian

PERMIAN

Anidanthus, Cathaysia, Caucasoproductus, Chonetella, Compressoproductus, Crurithyris, Denticuliphoria, Derbyia, Dielasma, Edriosteges, Enteletella, Enteletes, Gefonia, Gerassimovia, Haydenella, Hemiptychina, Heterelasmina, Hybostenoscisma, Krotovia, Labaella, Labaia, Lamnimargus, Leptodus, Licharewiconcha, Licharewina, Linoproductus, Marginifera, Martinia, Nikitinia, Notothyris, Permianella, Permophricodothyris, Phricodothyris, Prelissorhynchia, Probolionia, Prorichthofenia, Pseudowellerella, Rhipidomella, Richthofenia, Rostranteris, Scacchinella, Semibrachithyrina, Spinomarginifera, Stenoscisma, Stenoscismatoidea, Strophalosia, Strophalosiina, Tectarea, Terebratuloidea, Tethysiella, Transennatia, Tyloplecta, Uncinunellina, Urushtenia, Wellerella - Changhsingian

TRIASSIC

GENERA	I+O	An	La	Ca	No	Rh
"Rhynchonella"		1				
Abrekia	1					
Adygella				1	1	1
Adygelloides						1
Amphiclina					1	2
Ampliclinodonta					1	
Angustothyris		1				
Aulacothyropsis				1		4
Austriella					1	
Austrirhynchia						2
Balatonospira				1		
Bobukella				1	1	
Caucasorhynchia					1	1
Caucasothyris					1	
Coenothyris		2				
Costirhynchia		1		1		

Costispiriferina		1			
Crurirhynchia				1	
Crurithyris	1				
Cubanothyris				2	2
Decurtella		2			
Dinarispira		2			
Dioristella		1	1		
Euxinella				3	6
Fissirhynchia				1	1
Guseriplia					2
Holcorhynchella		1			
Koeveskallina		1			
Koninckina			1	1	
Laballa				2	3
Lepismatina					1
Lobothyris				2	
Majkopella					3
Mentzelia		1	2	1	2
Moisseievia				3	
Neoretzia					3
Neowelerella	1				
Norella		1			
Oxycolpella				3	2
Pexidella		1		1	
Piarorhynchella		1			
Pseudocyrtina				1	
Pseudorugitella				2	2
Punctospirella		1			
Rhaetina			1	6	4
Rhimirhynchopsis				1	1
Robinsonella					1
Sinucosta		1		1	1
Spinolepismatina				1	
Sulcatinella		2			
Sulcatothyris			1		
Tetractinella		1			
Thecospira				1	
Thecospiropsis				1	
Triadithyris				2	1
Trigonirhynchella				1	2
Volirhynchia		2			
Wittenburgella				1	1
Worobievella				1	1
Zeilleria				3	6
Zugmayerella					1

stage abbreviations: In – Induan, Ol – Olenekian, An – Anisian, La – Ladinian, Ca – Carnian, No – Norian, Rh – Rhaetian.

JURASSIC

GENERA	**Si**	**Pl**	**To**	**Aa**	**Bj**	**Bt**	**Cl**	**Ox**	**Km**	**Tt**
"Rhynchonella"	2	1	2	2	2					
"Spiriferina"	1	1								

"Terebratula"							1	1		2
Acanthorhynchia								1		
Acanthothyris				1	3					
Aromasithyris							1	2	1	
Aulacothyris		2	1				2			
Bodrakella		1								
Calcirhynchia	1	1								
Calvirhynchia					1					
Capillirhynchia				1			1			
Cardinirhynchia							1			
Caucasella							1			
Caucasorhynchia		1								
Cererithyris					1					
Cheirothyris										1
Cincta		2								
Cirpa		1								
Colosia								2	1	
Cryptorhynchia					1					
Cuersithyris		1								
Cuneirhynchia		4								
Curtirhynchia			1							
Digonella		1						1		
Disculina		1								
Dorsoplicathyris							2			
Ferrythyris							1			
Flabellirhynchia		1	1							
Furcirhynchia	1			1	1					
Gibbirhynchia	1	2	1							
Goniothyris							1		1	1
Grandirhynchia			1	1	1					
Gusarella							1			
Heimia					1					
Homoeorhynchia	1	2	1							
Ismenia										2
Ivanoviella							3	1		
Juralina								5	1	2
Lacunosella							1	3	2	1
Linguithyris	1				1					
Liospiriferina	2	4	1							
Loboidothyris					1			1		
Lobothyris		3	1	2	2					
Lophothyris									1	
Monsardithyris					1					
Monticlarella									1	2
Morrisithyris					1		1			
Nucleata								2	1	1
Paruirhynchia				1						
Piarorhynchia	1	4	1							
Postepithyris									1	
Praemonticlarella			1							
Prionorhynchia	1	1	1							
Pseudogibbirhynchia			2	2	2					

Ptyctorhynchia			1							
Ptyctothyris					1		3			
Quadratirhynchia					2					
Rhactorhynchia					3					
Rhynchonelloidea				1						
Rhynchonelloidella							1			
Rimirhynchia		1								
Rudirhynchia	2									
Rugitela					1					
Scalpellirhynchia	1									
Securina	1	1								
Sellithyris								1	1	
Septaliphoria							2	2		6
Somalirhynchia								1		
Sphaeroidothyris					2					
Spiriferina	4	3								
Squamirhynchia		1								
Stolmorhynchia				4	2		1			
Striirhynchia					1					
Stroudithyris				1						
Tchegemithyris							2			
Terebratula								1		
Tetrarhynchia	1				1					
Thurmanella							1			
Torquirhynchia								1	2	1
Trichorhynchia				2	1					
Tropeothyris										9
Tubithyris					2					
Wattonithyris					1					
Weberithyris										3
Zeilleria	15	14		3	4		2	1		

stage abbreviations: Si – Sinemurian, Pl – Pliensbachian, To – Toarcian, Aa – Aalenian, Bj – Bajocian, Bt – Bathonian, Cl – Callovian, Ox – Oxfordian, Km – Kimmeridgian, Tt – Tithonian.

LOWER CRETACEOUS

GENERA	**Be**	**Va**	**Ha**	**Ba**	**Ap**	**Al**
Belbekella	1		1	4	1	
Belothyris			1	2	2	
Orbirhynchia				1	1	
Peregrinella			1			
Praelongithyris					1	
Psilothyris		2				
Sellithyris		1	3			
Sulcirhynchia	1					
Symphythyris			1			
Terebrataliopsis		1				
Weberithyris		1				

stage abbreviations: Be – Berriasian, Va – Valangianian, Ha – Hauterivian, Ba – Barremian, Ap – Aptian, Al – Albian.

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PALAEO

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Jurassic transgressions and regressions in the Caucasus (northern Neotethys Ocean) and their influences on the marine biodiversity

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Abstract

In the Jurassic, the Caucasus, presently located in the southwest of Russia, Georgia, Armenia and Azerbaijan, was located on the northern active margin of the Neotethys Ocean. Facies interpretation in all 62 areas, distinguished by differences in facies, allows to semi-quantitatively evaluate Jurassic regional transgressions and regressions for this region. Major transgressive regressive cycles took place in the Hettangian–Aalenian, Bajocian–Bathonian and Callovian–Tithonian. Each transgression was more extensive than the previous. The same cycles are established in the Greater Caucasus Basin. Deep-marine environments were common in the Pliensbachian, late Aalenian and late Bathonian, whereas they were very restricted in the Late Jurassic. The Jurassic transgressions and regressions in the Caucasus coincided with the proposed global eustatic changes. However, some differences were caused by the regional tectonic activity. Although transgressions and regressions cause some changes in marine biodiversity, it seems that only ammonites might have been directly influenced by them. Diversity of bivalves, brachiopods and belemnites was driven by other factors. However, global changes in marine biodiversity were more closely related to the eustatic fluctuations than it was documented on a regional scale.

Keywords: Transgression; Regression; Eustasy; Biodiversity; Jurassic; Caucasus

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1. Introduction

Global sea level fluctuated strongly during the Jurassic (Hallam, 1978; Haq et al., 1987; Hallam, 1988, 1992, 2001; Haq and Al-Qahtani, 2005). Special attention has been paid to the intriguing question of how these changes as well as regional transgressions and regressions influenced marine biodiversity (Wiedman, 1973; Hallam, 1975, 1977; Jablonski, 1980; Lehmann, 1981; Gygi, 1986; Hallam, 1987; McRoberts and Aberhan, 1997; Hallam and Wignall, 1999; O'Dogherty et al., 2000; Sandoval et al., 2001a,b; Smith, 2001; Sarti, 2003; Ruban, 2004; Aberhan et al., 2005; Ruban and Tyszka, 2005; Ruban, 2006a). However, this question is not yet fully answered, and the influence is still poorly understood.

In this paper, I will focus on Jurassic transgressions and regressions in the Caucasus, a large region, stretching along the southern periphery of the Russian Platform and embracing the territory of southwestern Russia, Georgia, Armenia and Azerbaijan (Fig. 1). The few previous studies addressed to regional transgressions and regressions suggested that they were somewhat different from those observed globally and elsewhere in Europe

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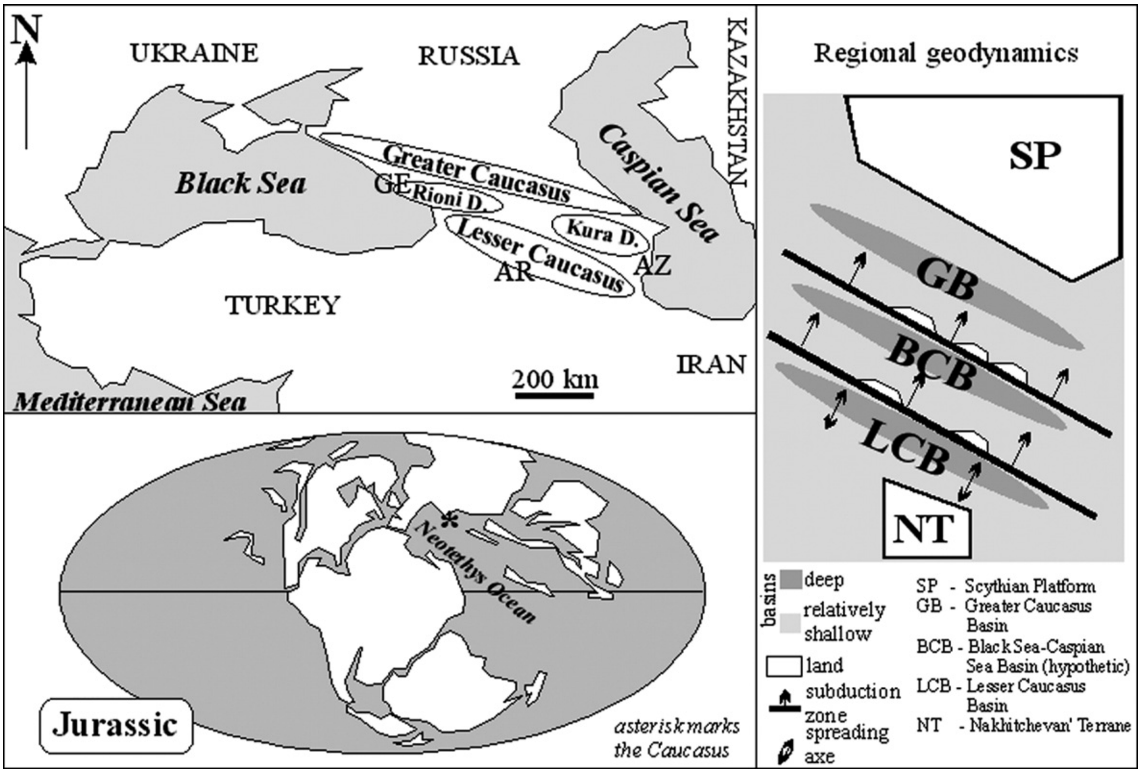

Fig. 1. Geographical location of the studied region. GE — Georgia, AR — Armenia, AZ — Azerbaijan. Palaeogeographical map is simplified after Scotese (2004). Geodynamic sketch is modified from Lordkipanidze et al. (1984), who used the geological and palaeomagnetic evidences.

(Ruban, 2004; Ruban and Tyszka, 2005; Ruban, 2006a). The Late Jurassic history of the Caucasus is marked by two remarkable events — the development of a carbonate platform with the growth of carbonate buildups (Rostovtsev et al., 1992; Kuznetsov, 1993; Martin-Garin et al., 2002; Akhmedov et al., 2003; Ruban, 2005a, 2006a; Tawadros et al., 2006) and a salinity crisis (Jasamanov, 1978; Kuznetsov, 1993; Ruban, 2006a; Tawadros et al., 2006). These events are interpreted as controlled by the basin dynamics and climate, and to document their relationships with transgressions and/or regressions is intriguing.

In this paper, an attempt is made to detail and to semi-quantitatively evaluate Jurassic transgressions and regressions in the Caucasus. The study area may serve as a test region to investigate their influences on the marine biodiversity. Previous studies have suggested that sea-level changes did not cause mass extinction among brachiopods (Ruban, 2004) and foraminifers (Ruban and Tyszka, 2005), and they did not control bivalve diversity (Ruban, 2006a). Here the influences of the Jurassic transgressions and regressions on ammonites, bivalves, brachiopods and belemnites are considered.

2. Geological setting

The Caucasus is a large elongated region consisting of three principal segments, the Greater Caucasus, the Transcaucasian depressions (the Kura Depression and the Rioni Depression), and the Lesser Caucasus (Fig. 1). During the Jurassic, the Caucasus was located on the northern margin of the Neotethys Ocean (Gamkrelidze, 1986; Dercourt et al., 2000; Stampfli and Borel, 2002; Golonka, 2004; Tawadros et al., 2006; Ruban, 2006d). Accretion of minor terranes along the southern margin of the Russian Platform (also known as the Scythian Platform) resulted in the development of several marine basins, separated by island arcs (Lordkipanidze et al., 1984; Ershov et al., 2003; Efendiyeva and Ruban, 2005; Tawadros et al., 2006; Ruban, 2006d) (Fig. 1). The geometry of these basins changed during the Jurassic (Ruban, 2006d), but the evaluation of the changes and their precise delineation have not yet been realized.

The Jurassic stratigraphy of the Caucasus was comprehensively presented by Prosorovskaya (1979) and Rostovtsev et al. (1985, 1992). Ruban (2003, 2006b) and Ruban and Pugatchev (2006) revised the regional stratigraphy of the Western Caucasus and proposed a new framework (Fig. 2), incorporating recent developments in Jurassic chronostratigraphy (Gradstein et al., 2004; Gradstein and Ogg, 2005, 2006), including the ratification of the Sinemurian, Aalenian and Bajocian Global Standard Sections and Points (Pavia and Enay, 1997; Cresta et al., 2001; Bloos and Page, 2002), and biostratigraphy (Cariou and Hantzpergue, 1997). The Callovian Stage in the Caucasus is traditionally ascribed to the Upper Jurassic (Rostovtsev et al., 1992; Ali-Zadeh, 2004) which contradicts to the present International Stratigraphic Chart (Gradstein et al., 2004). The palynological study of Gaetani et al. (2005) suggested a late Bathonian age for the lower part of the Kamennomostskaja Formation. If so, the termination of the major Bathonian hiatus should be reconsidered throughout the

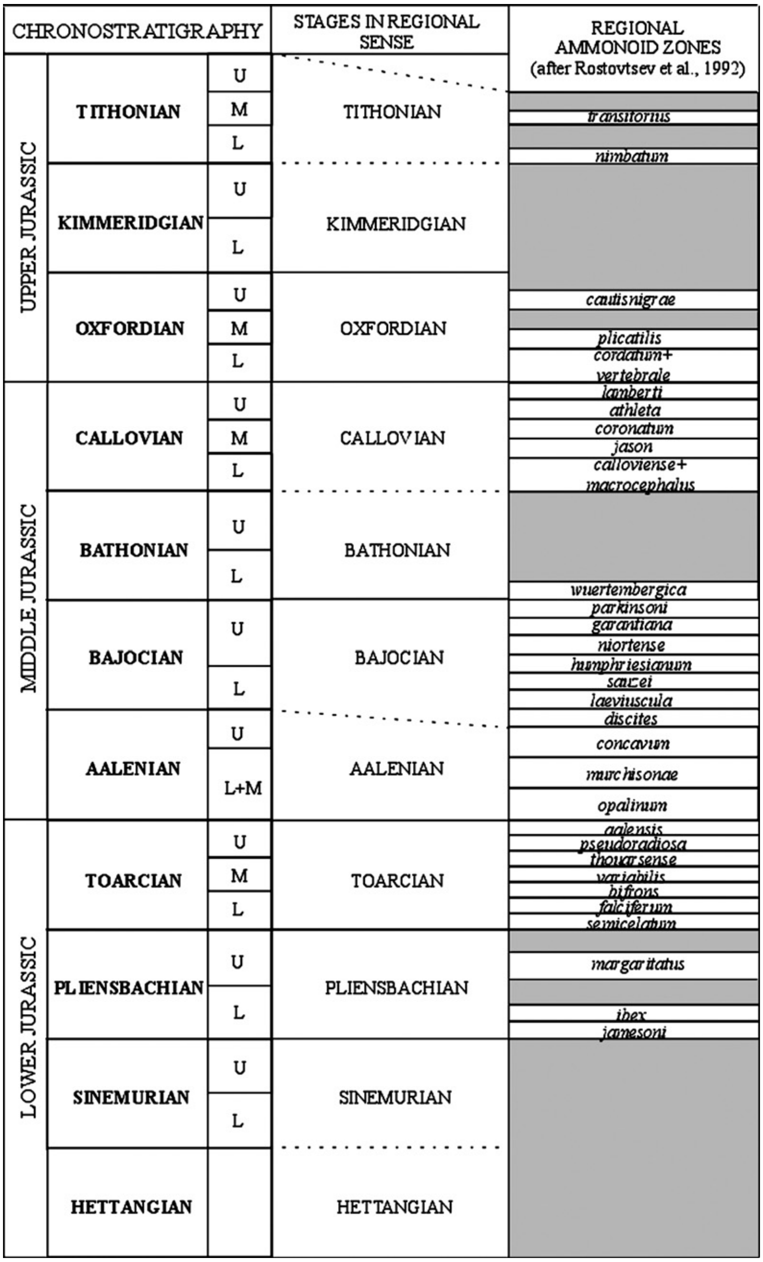

Fig. 2. Stratigraphic scale of the Jurassic used in the Caucasus (after Ruban, 2006d). Abbreviations: L — Lower, M — Middle, U — Upper. Unzoned intervals are shaded as gray. Dashed line marks uncertainty in the boundary definition. Regional ammonite zonation does not correspond at this scale to the showed chronostratigraphy (it seems to be impossible to correlate them at now), but only to the stages in regional sense.

entire Caucasus, which may lead to a significant revision of the Middle Jurassic stratigraphy of the region. However, in this paper I prefer the widely accepted early–middle Callovian age of the Kamennomostskaja Formation, an age supported by ammonites, bivalves and belemnites (Prosorovskaya, 1979; Rostovtsev et al., 1992; Ruban, 2005b).

The Jurassic deposits, lithologically quite variable, outcrop in hundreds, if not thousands, of sections across the Caucasus. The territory of the Caucasus is subdivided into several dozen areas which are traditionally called "zones" (Rostovtsev et al., 1992) distinguished by differences in facies (Fig. 3). However, the term "zone" should be abandoned to avoid confusion with biostratigraphic zones.

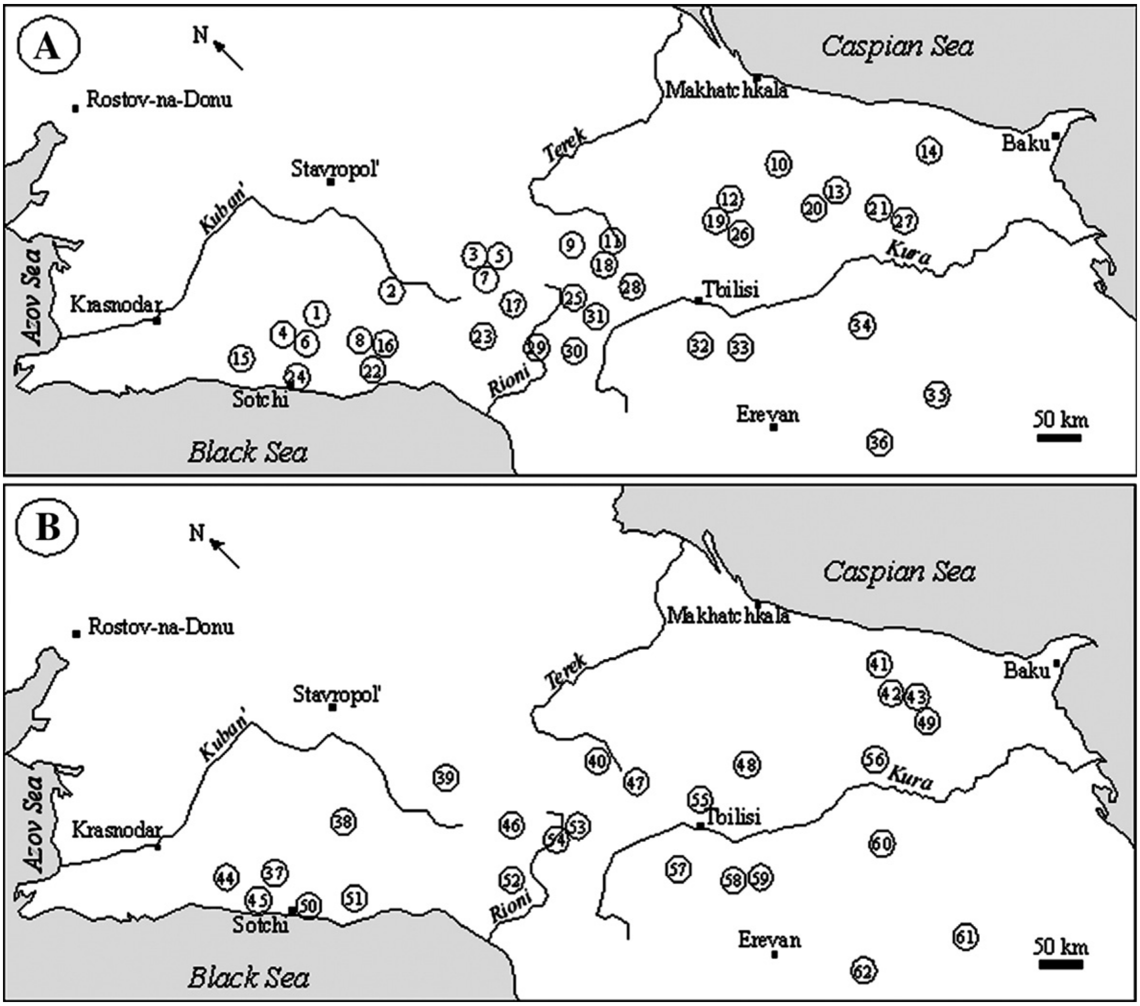

Fig. 3. Location of the Jurassic areas (marked by circles) in the Caucasus (after Rostovtsev et al., 1992). A — Hettangian–Bathonian areas (1–36), B — Callovian–Tithonian areas (37–62). Areas ("subzones" and "regions" of Rostovtsev et al. (1992) are mentioned here as areas): 1 — Western Labino–Malkinskaja, 2 — Central Labino–Malkinskaja, 3 — Eastern Labino–Malkinskaja, 4 — Western Pshikish–Tyrnyauzskaja, 5 — Eastern Pshikish–Tyrnyauzskaja, 6 — Northern Arkhyz–Guzeripl'skaja, 7 — Eastern Arkhyz–Guzeripl'skaja, 8 — Southern Arkhyz–Guzeripl'skaja, 9 — Digoro–Osetinskaja, 10 — Agwali–Khivskaja, 11 — Western Bokovogo Khrebta, 12 — Central Bokovogo Khrebta, 13 — Eastern Bokovogo Khrebta, 14 — Southeastern Bokovogo Khrebta, 15 — Gojtkhsko–Atchishkhinskaja, 16 — Severoabkhazskaja, 17 — Svanetskaja, 18 — Western Glavnogo Khrebta, 19 — Central Glavnogo Khrebta, 20 — Tfanskaja, 21 — Durudzhinskaja, 22 — Western Gagra–Dzhavskaja, 23 — Eastern Gagra–Dzhavskaja, 24 — Amuksko–Lazarevskaja, 25 — Sakaojskaja, 26 — Shakrianskaja, 27 — Vandamskaja, 28 — Kakhetino–Letchkhumskaja, 29 — Tskhenistskali–Okribskaja, 30 — Southwestern Dzirul'skaja, 31 — Northeastern Dzirul'skaja, 32 — Loksko–Khramskaja, 33 — Alaverdskaja, 34 — Shamkhorsko–Karabakhskaja, 35 — Kafanskaja, 36 — Araksinskaja; 37 — Lago–Nakskaja, 38 — Labinskaja, 39 — Malkinskaja, 40 — Kabardino–Dagestanskaja, 41 — Jugo–Vostotchnogo Dagestana, 42 — Sudurskaja, 43 — Shakhdagskaja, 44 — Abino–Gunajskaja, 45 — Novorossijsko–Lazarevskaja, 46 — Svanetsko–Verkhneratchinskaja, 47 — Liakhvi–Aragvinskaja, 48 — Kakhetinskaja, 49 — Dibrarskaja, 50 — Akhtsu–Katsyrkha, 51 — Dzhirkhva–Akhibokhskaja, 52 — Tkvartcheli–Okribskaja, 53 — Ratchinskaja, 54 — Tsessi–Kortinskaja, 55 — Iori–Tsitelitskarojskaja, 56 — Vandamskaja, 57 — Khramskaja, 58 — Lalvarskaja, 59 — Idzhevanskaja, 60 — Dashkesano–Karabakhskaja, 61 — Kafanskaja, 62 — Nakhitchevanskaja.

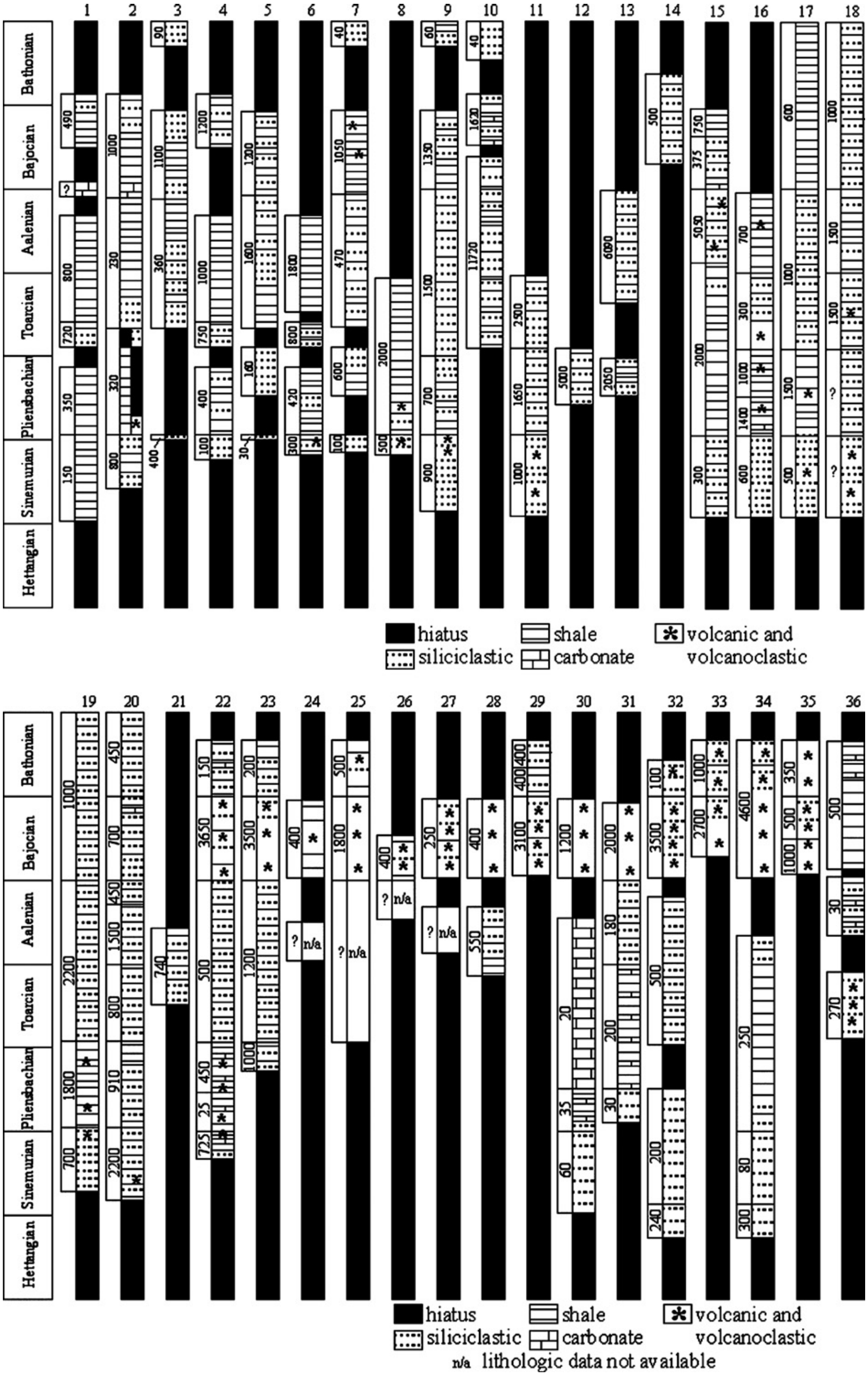

Fig. 4. Composite lithologic sections of the Caucasian areas: Hettangian–Bathonian (based on field observations and data of Rostovtsev et al., 1992). Location of areas — see Fig. 3. Dominating sedimentary rocks are shown. Maximum thickness (meters) is indicated to the left of each column.

I propose therefore to use the usual term “area”. Composite lithologic sections have been drawn for each of these areas (Figs. 4, 5). In general, siliciclastics (up to 10,000 m thick) dominate the Lower–Middle Jurassic successions, whereas carbonates (up to 3000 m thick) prevail in the Upper Jurassic succession (Tsejsler, 1977; Prosorovskaya,

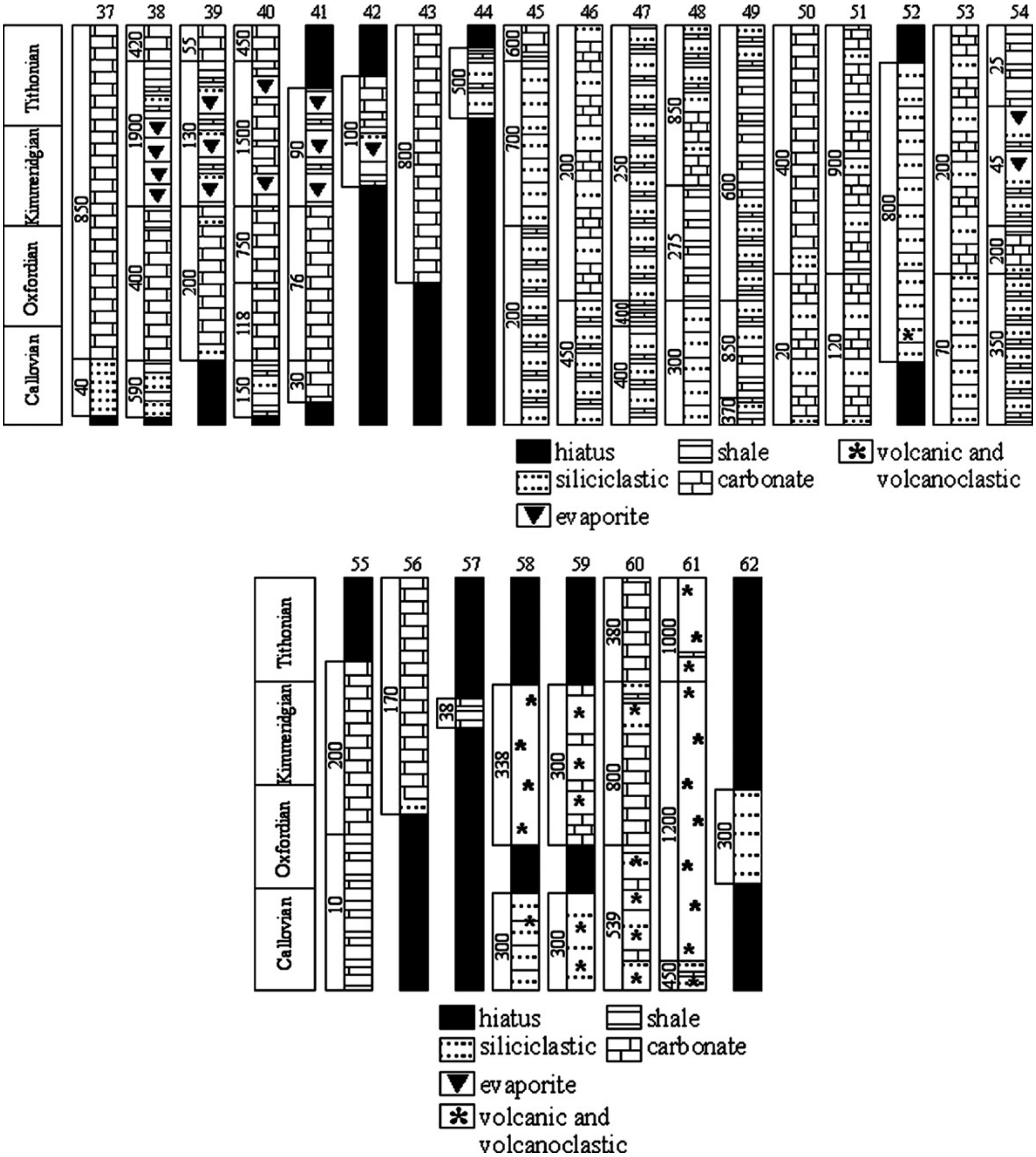

Fig. 5. Composite lithologic sections of the Caucasian areas: Callovian–Tithonian (based on field observations and data of Rostovtsev et al., 1992). Location of areas — see Fig. 3. Dominating sedimentary rocks are shown. Maximum thickness (meters) is indicated to the left of each column.

1979; Rostovtsev et al., 1992; Tawadros et al., 2006). In the Lesser Caucasus, volcanics and volcanoclastics are abundant (Prosorovskaya, 1979; Rostovtsev et al., 1985, 1992). Palaeobiogeographically, the Caucasus belonged to the Tethyan Subrealm until the Middle Jurassic, when it became a part of the Tethyan Realm (Westermann, 2000).

3. Materials and methods

The method of transgression and regression evaluation used in this study is somewhat similar to that proposed by Ruban (2006a,b,c) and earlier by Hallam and Wignall (1999), Peters and Foote (2001), Smith (2001), and Crampton et al. (2003). Transgressions and regressions

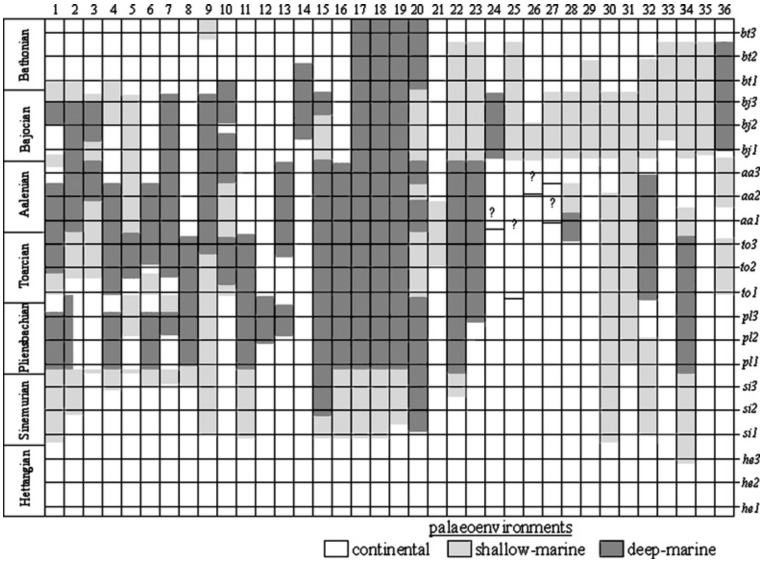

Fig. 6. Interpreted palaeoenvironments of the Hettangian–Bathonian. Location of areas — see Fig. 3. Time slices used to evaluate sea-level changes are shown by horizontal lines and indexes from the right side.

are defined as land- and seaward migrations of the shorelines respectively (Catuneanu, 2006; cf. Veeken, 2006). They should be distinguished from deepenings and shallowings, which describe water depth in the basin.

The first step in my study is to interpret the facies for each area, based on information from Rostovtsev et al. (1992) and personal field observations (Figs. 6, 7). Interpretations made earlier by Ruban (2006a) were verified and slightly revised. Three main types of palaeoenvironments were defined: continental, shallow-marine and deep-marine. Continental palaeoenvironments are marked by a hiatus or rarely by continental deposits usually comprising sandstones and shales with abundant floral remains and lacking any marine fauna. Shallow-marine palaeoenvironments are dominated by siliciclastics or carbonates with benthonic shelfal fauna. Deep-marine palaeoenvironments are marked by laminated dark-coloured shales and turbidites common with submarine slumps.

The second step is to calculate the number of areas with a particular type of palaeoenvironments for each of the time slices. In this paper I consider three time slices for each stage (Figs. 6, 7) rather than the single time slice per stage used by Ruban (2006a).

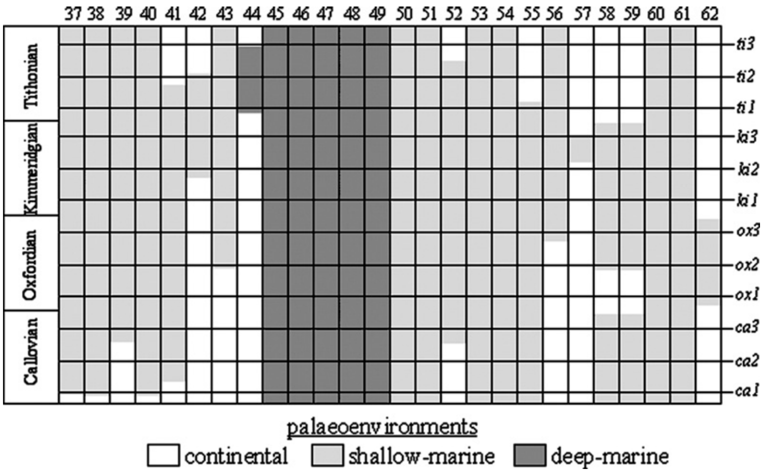

Fig. 7. Interpreted palaeoenvironments of the Callovian–Tithonian. Location of areas — see Fig. 3. Time slices used to evaluate sea-level changes are shown by horizontal lines and indices from the right side.

The third step was to evaluate semi-quantitatively the transgressions and regressions with TR-index (named ISL by Ruban, 2006a):

$$\mathrm{TR} = (s + d)/c,$$

where s, d, and c are the number of areas for each time slice with shallow-marine, deep-marine and continental palaeoenvironments respectively. Lower values of this index indicate regression, whereas higher values indicate transgression.

Additionally, the semi-quantitative evaluation of changes in the average basin slope angle, DS-index is attempted:

$$\mathrm{DS} = d/s.$$

However, this index cannot be used to document changes of the maximum water depth of the basin. The latter may be recorded by the appearance of deep-marine palaeoenvironments even in the unique area. The DS-index characterizes the extent of deep-marine palaeoenvironments in palaeogeographical space in relation to the extent of shallow-marine environments. Lower values of the index do not indicate that the basin was shallower at this time interval than previously when values were higher. Rather, they indicate that the deep-marine palaeoenvironments became restricted in a little amount of areas. The DS-index measures the average basin slope angle. Therefore, it is necessary to take into consideration another curve, which demonstrates the maximum water depth of the basin. This was not made herein since the Sinemurian deep-marine palaeoenvironments existed at least in one area of the Caucasus.

The regional tectonic processes were perhaps the principal control of the regional transgressions and regressions. In the Jurassic, the Caucasus consisted of several basins different from one another in origin, tectonic regime, and general "geometry" (Lordkipanidze et al., 1984; Ershov et al., 2003; Efendiyeva and Ruban, 2005; Tawadros et al., 2006; Ruban, 2006d). This makes difficult an interpretation of the constrained TR- and DS-curves, as they are attributed to the entire Caucasus. With the suggestions by Rostovtsev et al. (1992) and Ruban (2006d) it is possible to establish the areas, which belong to the Greater Caucasus Basin in the Jurassic. This permits to calculate TR-index and DS-index for this particular basin. Unfortunately, our knowledge on the other Caucasian basins remains limited, and we cannot attempt the same for them.

The absolute age and duration of stages mentioned in this paper are based on the time scale of Gradstein et al. (2004).

4. Regional transgressions and regressions

Three Jurassic transgressive–regressive cycles are recognized in the Caucasus (Fig. 8). The first Jurassic transgressive–regressive cycle embraced the Hettangian–Aalenian interval, lasting 28 m.y. After major Hettangian hiatus, which can be traced across the entire Caucasus, a gradual transgression began. In the early Toarcian, a small regressive episode is known in the Caucasus, which was followed by a significant transgression. A remarkable, but short-term, regional regression occurred in the Aalenian. The second transgressive–regressive cycle was shorter and it embraced the Bajocian–Bathonian interval, lasting 6.9 m.y. The sea rapidly transgressed at the beginning of the Bajocian and reached a maximum territory in the middle–late Bajocian. In the Bathonian, the marine basin was restricted to a size similar to that of the Sinemurian, and at the end of the Bathonian, sedimentation was terminated within most areas of the Caucasus. This time interval corresponded to the second major regional hiatus. The third transgressive–regressive cycle embraced the Callovian–Tithonian interval, lasting 19.2 m.y. A gradual transgression took place during the Callovian and Oxfordian, and minor regressive episode was documented in the early Oxfordian. The peak of transgression took place in the late Oxfordian–Kimmeridgian. A significant short-term regression occurred in the early Kimmeridgian, but during the middle and late Kimmeridgian the sea had the same extent as in the late Oxfordian. A gradual regression occurred in the Tithonian, although the sea still covered a large region at the end of the Tithonian. The Callovian–Late Jurassic transgression explains the development of the wide carbonate rimmed shelf with the growth of carbonate buildups (Rostovtsev et al., 1992; Kuznetsov, 1993; Martin-Garin et al., 2002; Akhmedov et al., 2003; Ruban, 2005a, 2006a; Tawadros et al., 2006). Very shallow lagoonal environments were common in the Kimmeridgian–early Tithonian, where evaporties or varicoloured shales were deposited (Tsejsler, 1977; Jasamanov, 1978; Rostovtsev et al., 1992; Kuznetsov, 1993), which corresponded to the beginning of the end-Jurassic regression. In the Andean region, Late Jurassic evaporitic deposition also occurred during a regressive episode (Legaretta and Uliana, 1996; Hallam, 2001). The same event took place in Northeastern Africa (Tawadros, 2001, pers. comm. 2006), Arabia (Sharland et al., 2001) and Germany (Stratigraphische Tabelle von Deutschland, 2002).

In the Greater Caucasus basin, the same transgressive–regressive cycles have been established (Fig. 8). Only minor differences in the TR-pattern between the entire Caucasus and the Greater Caucasus Basin are found. The Late Plienbachian transgression was larger

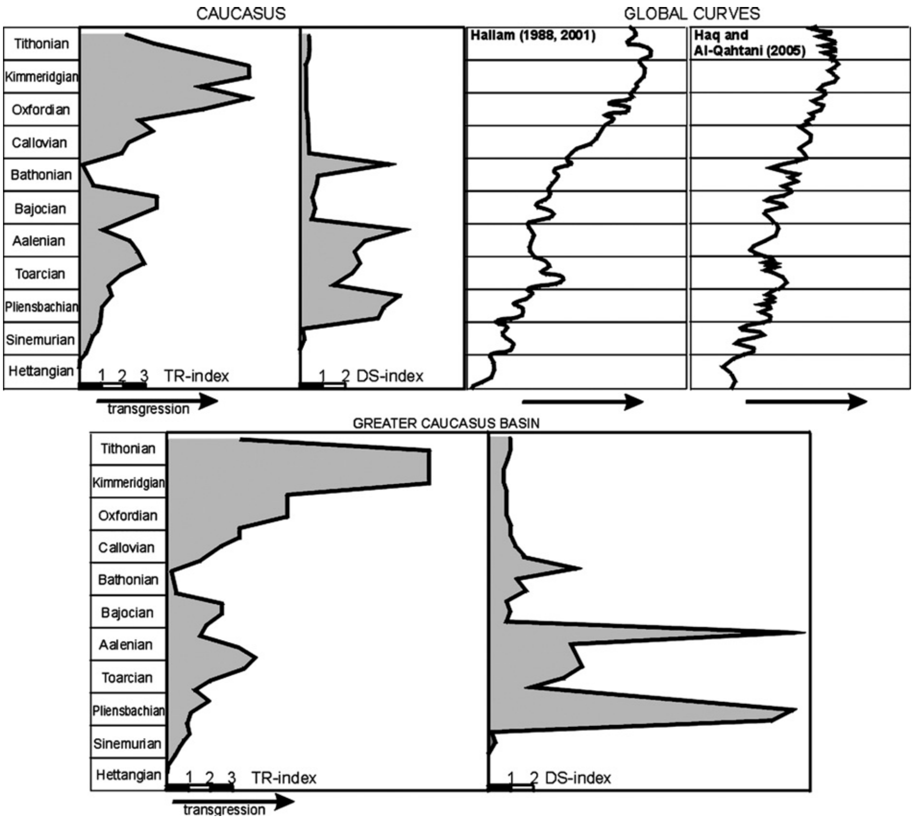

Fig. 8. Regional transgressions and regressions and the global Jurassic curves.

in the latter, whereas the Bajocian transgression was less. Also the regressions in the early Oxfordian and the early Kimmeridgian were not present within the Greater Caucasus Basin.

In contrast to the transgressive–regressive pattern, the changes in the average basin slope angle appeared as few short-term pulses both in the entire Caucasus and in the Greater Caucasus Basin (Fig. 8). The first pulse occurred in the Pliensbachian, the second pulse in the late Aalenian, and the last in the late Bathonian. During the Late Jurassic the Caucasus was characterized by the relatively small number of areas with deep-marine palaeoenvironments in contrast to the Early and Middle Jurassic.

The regionally documented shoreline migrations might have been controlled by both global eustatic fluctuations and regional tectonics as deduced from the present sequence stratigraphic models (Catuneanu, 2006). The eustatic changes in the Jurassic might have been caused by plate tectonics, plume activity and glacioeustasy (Hallam, 1992, 2001; Veeken, 2006). A comparison of the curve by Hallam (1988), updated in Hallam (2001), and the curve of Haq et al. (1987), updated recently by Haq and Al-Qahtani (2005), with the curve of the Caucasian transgressions and regressions (Fig. 8), suggests that their general trends correspond quite well. However, nothing major appeared globally in the Bathonian, when a major regression took place in the Caucasus. The relationships between the global eustatic fluctuations and the regionally documented transgressions and regressions were always complicated because the eustasy is not a unique factor of the regional shoreline migrations. McGowran (2005) even questions how trustable are our global sea-level reconstructions based on the regional studies.

The regional tectonic activity was potentially the main factor, which controlled the Jurassic transgressions and regressions in the Caucasus. The opening and extension of the new marine basins, originated in the beginning of the Jurassic, dominated until the early Aalenian (Ershov et al., 2003), provoked a regional transgression. At the same time, a subsidence of the southern margin of the Russian Platform (Ershov et al., 2003) additionally contributed to the latter. The late Aalenian regression might have been a result of the “orogeny” hypothesized by Ershov et al. (2003). The Bajocian transgression had the same mechanism as that in the Early Jurassic. The major Bathonian regression was a result of the other phase of the mid-Jurassic “orogeny” (Ershov et al., 2003) or it

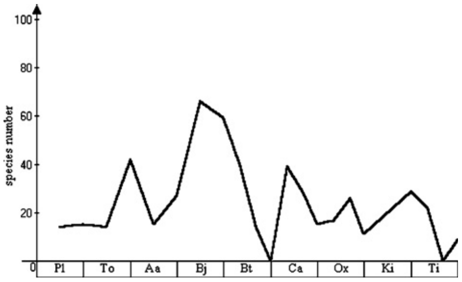

Fig. 9. Ammonite diversity dynamics in the Caucasus (data from Rostovtsev et al., 1992).

was a consequence of the arc–arc collision proposed by Ruban (2006d). The Callovian–Late Jurassic transgression was so large because of continuing basin extension and subsidence of its margins (Ershov et al., 2003). As for the end-Jurassic regression, it may be linked to the regional compressional event and partial uplift of basin margins (Ershov et al., 2003).

5. Transgressions and regressions and marine biodiversity in the Caucasus

Transgressions and regressions might have been an important factor which drove the changes in taxonomic diversity of the marine fauna. Four fossil groups are considered here: ammonites, bivalves, brachiopods and belemnites as they were the principal contributors to the marine biodiversity. The total number of species exceeds 1200 (see review papers by Makridin and Kamyshan, 1964; Prosorovskaya, 1993a,b; Rostovtsev et al., 1992; Topchishvili et al., 2005; Ruban, 2004, 2005b, 2006a,c). Below, a comparison between the changes of the total species number (Figs. 9–12), and transgressions and regressions (Fig. 8) is shown for each of these groups.

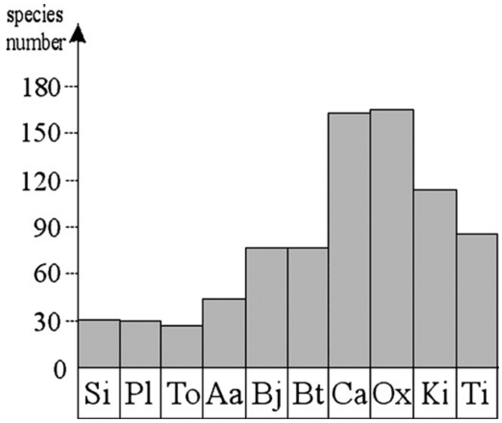

Fig. 10. Bivalve diversity dynamics in the Caucasus (after Ruban, 2006a).

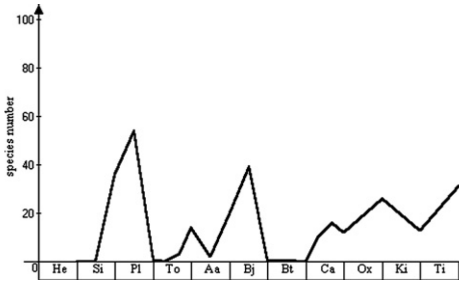

Fig. 11. Brachiopod diversity dynamics in the Northern Caucasus (data from Makridin and Kamyshan, 1964; Rostovtsev et al., 1992; Prosorovskaya, 1993a,b; Ruban, 2004, 2006c). The data and curve are attributed to the Northern Caucasus only, but they seem to be representative for the entire Caucasus.

5.1. Ammonites

The taxonomic diversity of the Caucasian ammonites fluctuated strongly during the Jurassic (Fig. 9). An absolute maximum was reached in the Bajocian, while the early Aalenian, Bathonian, late Callovian, late Oxfordian and middle Tithonian are characterized by significant diversity drops. Overall, a slight decline in ammonite diversity can be documented between the Early–Middle Jurassic and the Late Jurassic.

The ammonite diversity changes (Fig. 9) coincided with the transgressions and regressions (Fig. 8). Diversity rises corresponded to transgressive episodes, and falls to regressions. However, the coincidence of the overall transgression, documented for the entire Jurassic, contrasts with the long-term, slight species decline. This may be explained by the abrupt change from basins with a wide distribution of deep-marine conditions in the Early–Middle Jurassic to shallow-marine in the Late Jurassic. Ammonites were stenotypic organisms (Sandoval et al., 2001a). However, the other explanation of the Late Jurassic diversity decline may be related to the restriction of connection between the boreal and

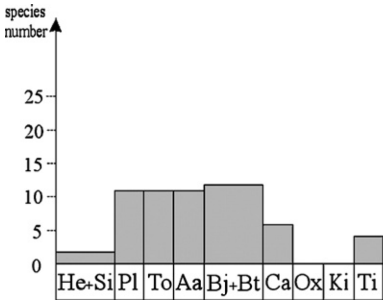

Fig. 12. Belemnite diversity dynamics in the Caucasus (data from Rostovtsev et al., 1992; Topchishvili et al., 2005; Ruban, 2005b).

temperate–tropical realms through the Russian Platform (Dercourt et al., 2000; Rogov et al., 2006).

5.2. Bivalves

The taxonomic diversity of the Caucasian bivalves was low in the Early Jurassic (Fig. 10). After a small decline in the Toarcian, it increased rapidly and a maximum was reached in the Callovian–Oxfordian. Then a gradual diversity drop took place, but the total species number in the Tithonian was greater than in Bajocian–Bathonian.

There are no direct relations between the regional bivalves diversity (Fig. 10) and transgressions and regressions (Fig. 8) (Ruban, 2006a). For example, the total species number rose both in the Callovian, when transgression began, in the Aalenian, when regression took place, and it did not fall in the Bathonian, when major regression occurred. However, the bivalve diversity was much higher during that time interval when shallow-marine conditions dominated. Thus, it is difficult to consider transgressions and regressions as a factor which influenced bivalve diversity, and many other factors were at least as significant.

5.3. Brachiopods

The taxonomic diversity of the North Caucasian brachiopods changed rapidly during the Jurassic (Fig. 11). Peaks occurred in the early Pliensbachian, early Bajocian, Oxfordian and Tithonian, while significant drops took place in the late Pliensbachian–early Toarcian, early Aalenian, late Bajocian–Bathonian and Kimmeridgian. No overall changes in the total species number are found.

There are no direct links between the changes of the brachiopod diversity (Fig. 11) and sea level (Fig. 8). For example, both the early Pliensbachian diversification and the late Pliensbachian decline occurred during gradual transgression. Only the regression in the Bathonian led to the demise of brachiopods. But even in this case, there is uncertainty: the diversity decline began in the late Bajocian, slightly before the regional regression had begun. The brachiopod diversity in the Northern Caucasus was controlled by other factors than transgressions and regressions.

5.4. Belemnites

The taxonomic diversity of the Caucasian belemnites did not change significantly during the Jurassic (Fig. 12). It remained at the same level during the Pliensbachian–Bathonian. A remarkable disappearance of belemnites occurred in the Oxfordian–Kimmeridgian interval. The same disappearance is also known in the English Kimmeridgian (Wignall, pers. comm. 2006). The observed patterns (Fig. 12) cannot be explained by transgressions and regressions (Fig. 8), and it seems that other factors were at least as important. The dominance of shallow-marine conditions does not explain the total absence of belemnites in the Oxfordian–Kimmeridgian because there were some deep-marine areas during this time interval.

5.5. Brief synthesis

Comparison between regional transgressions and regressions and diversity dynamics of four fossil groups suggests that only ammonites were influenced directly by the former, although even for this fossil group other factors might have been even more important (see above). These other factors were more significant for bivalves, brachiopods and belemnites. A detailed study of ammonite diversity and sea-level changes in the Cordillera Bética (south of Spain) suggests that they were correlated (O'Dogherty et al., 2000; Sandoval et al., 2001a,b). The same conclusion was reached with the Caucasian data.

6. Discussion

6.1. Jurassic transgressions and regressions in the Caucasus and selected Peri-Tethyan and Neotethyan regions

Jurassic transgressions and regressions, reconstructed in the Caucasus, are compared to those established in some Peri-Tethyan and Neotethyan regions. In Western Europe, two major transgressive–regressive cycles are established in the Jurassic (Jacquin and de Graciansky, 1998; Jacquin et al., 1998). The first was the Ligure Cycle which started in the Late Triassic, and the peak was reached in the middle Toarcian, when a rapid regression occurred. The Aalenian–Bajocian transition is marked by a widespread unconformity. The succeeding North Sea Cycle started in the Bajocian, and the stepwise transgression reached its maximum in the Kimmeridgian. However, minor regressive episodes occurred in the Bathonian and Oxfordian. Since the Tithonian, a regression took place, and this cycle ended in the beginning of the Early Cretaceous. Although transgression and regressions in the basins of the Western Europe are rather similar to those of the Caucasus (Fig. 8), significant differences are evident. There is no such difference between the Hettangian–Bathonian and Callovian–Tithonian intervals in Western Europe as it is found in the Caucasus. The principal boundary between cycles in Western Europe is the regional unconformity at the Aalenian–Bajocian transition, while in the Caucasus the most remarkable hiatus occurred in the Bathonian. The latest Triassic–earliest Jurassic transgression was rapid in Western Europe, while it appeared later and more gradually in the Caucasus. The peak of the Early Jurassic transgression was reached a little later in the Caucasus.

Smith (2001) measured the outcrop area of the terrestrial/fluviatile, marine unfossiliferous and marine fossiliferous sedimentary rocks which outcrop in England and France and concluded that the sea transgressed from the Late Triassic until the Middle Jurassic. A minor regressive episode was established in the Aalenian. The second transgression occurred in the Bajocian–Bathonian although a regression took place in the Callovian. After the next transgression in the Late Jurassic, a remarkable regression occurred in the Berriasian. Such changes in England and France only partly correspond to the changes recorded in the Caucasus (Fig. 8). The Hettangian–Aalenian records are similar, while the Bathonian regression, which took place in the Caucasus, is not recovered in England and France, and the Callovian regression, though not so large as the Bathonian regression in the Caucasus, has no analogue in England and France.

Wignall et al. (2005) reported the earliest Toarcian regression in Western Europe, which is comparable to that in the Caucasus.

Guillocheau et al. (2000) recognized Carnian–Toarcian, Aalenian–lower Bathonian, lower Bathonian–Oxfordian, and Kimmeridgian–lower/upper Berriasian boundary cycles in the Paris Basin. These cycles are difficult to trace in the Caucasus at all.

The Stratigraphische Tabelle von Deutschland (2002) presents a detailed overview of the Jurassic formations and facies established in Germany. Shallow-marine facies dominated in the German basins during the Jurassic, and no pelagic facies have been identified. Regressive episodes, documented by the high number and wider extent of local hiatuses, occurred in the Hettangian, late Sinemurian, middle Toarcian, late Aalenian–middle Bajocian, Bathonian–early Callovian, late Callovian, and Kimmeridgian–Tithonian. Only the first of these has an analogue in the Caucasus. Some other regressions in the German basins, such as late Aalenian–middle Bajocian, Bathonian–early Callovian, Kimmeridgian–Tithonian, only partly corresponded to the regressions documented in the Caucasus (Fig. 8).

Surlyk (2003) developed a curve for East Greenland, which demonstrates the shoreline migration. Thus, this is essentially a transgressive–regressive curve. According to it, a general weak-regressive trend in the Toarcian–Bajocian changed to the prominent transgressive trend in the Bathonian–Kimmeridgian. Among the second-order events, the most remarkable were the early Bathonian, middle Callovian–early Oxfordian, and Kimmeridgian transgressions as well as the regressions at the Toarcian–Aalenian transition, in the late Bajocian, and in the mid-Oxfordian. The general trends documented in East Greenland are analogous to those in the Caucasus (Fig. 8). However, the second-order events appear to be incomparable, except the Kimmeridgian transgression, which is evident both in East Greenland and the Caucasus.

On the Arabian Plate, transgressions occurred in the early–middle Toarcian, early Bajocian, early Bathonian, Callovian–Oxfordian, middle Kimmeridgian, and late Tithonian with a maximum in the Early Cretaceous, while regressions took place in the late Toarcian, late Bajocian, late Bathonian, late Oxfordian–early Kimmeridgian, and late Kimmeridgian–middle Tithonian (Sharland et al., 2001). Only a few of these episodes had direct analogues in the Caucasus (Fig. 8). Therefore, transgressions and regressions documented in the latter and in Arabia differed somewhat.

Available data and their interpretations (Schandelmeier and Reynolds, 1997; Tawadros, 2001, pers. comm. 2006; Guiraud et al., 2005) allow recognition of chronology of transgressions and regressions in northern and northeastern Africa. Transgression occurred during the Early Jurassic. The Toarcian–Aalenian transition is marked by an unconformity which seems to be a result of regression. Then sea transgressed, although the late Callovian and late Tithonian are marked by regressive episodes. The peak of transgression was reached in the Kimmeridgian. Such sea-level changes in northern Africa do not correspond well to the changes documented in the Caucasus (Fig. 8).

Consequently, Jurassic transgressions and regressions in the Caucasus were only partly similar to those recorded in other regions. This may be explained by differences in the tectonic history of those regions.

6.2. Global changes in the Jurassic biodiversity and sea level

A problem is the low resolution of the Jurassic global marine biodiversity curve. The most reliable data of Peters and Foote (2001) provide the maximum and minimum numbers of marine genera for the Early, Middle and Late Jurassic, but even these would allow recognition of the links with the eustatic changes. A biodiversity curve of Newman (2001) is a bit more detailed. The average number of Early Jurassic marine genera is 1046. It increased up to 1425 genera in the Middle Jurassic, and then slightly rose again up to 1446 genera in the Late Jurassic. Comparison of these numbers with the global sea-level changes (Haq et al., 1987; Hallam, 1988, 2001; Haq and Al-Qahtani, 2005) (Fig. 8) shows that global marine biodiversity increased together with eustatic rises during the Jurassic.

A comparison of the global generic diversity dynamics of Jurassic bivalves (Miller and Sepkoski, 1988) with the eustatic changes (Haq et al., 1987; Hallam, 1988, 2001; Haq and Al-Qahtani, 2005) suggests a close relation. A rapid eustatic rise in the Early Jurassic provoked a significant radiation of bivalves. The next radiation, which occurred in the late Middle Jurassic, evidently coincided with the Bajocian–Callovian transgression. The diversity peak was reached in the Late Jurassic at the same time when sea level was the highest. The Tithonian eustatic fall resulted in a bivalve decline. Earlier, Hallam (1977) also found that eustatic changes significantly controlled the global bivalve diversity.

A comparison between the global belemnite diversity changes (Doyle and Bennett, 1995) and sea-level fluctuations (Haq et al., 1987; Hallam, 1988, 2001; Haq and Al-Qahtani, 2005) suggests that all three increases in diversity of belemnites, occurring in the Pliensbachian–Toarcian, Bajocian and Callovian–Oxfordian, corresponded to global eustatic rises. However, belemnites declined in a stepwise pattern during the Jurassic while sea level rose. Therefore, the links between the Jurassic sea-level changes, global marine biodiversity and diversity of particular fossil groups are evident on a global scale.

7. Conclusions

Three transgressive–regressive cycles have been established in the entire Caucasus, namely the Hettangian–Aalenian, Bajocian–Bathonian and Callovian–Tithonian cycles. Each transgression was more extensive than the previous. The same cycles have been established for the Greater Caucasus Basin. The Jurassic transgressions and regressions documented in the Caucasus correspond generally to the global eustatic fluctuations recorded by Haq et al. (1987), Hallam (1988, 2001), and Haq and Al-Qahtani (2005). The regional tectonic activity was another important control of the regional transgressions and regressions. The Caucasian transgressions and regressions only partly corresponded to those established in some Peri-Tethyan and Neotethyan regions.

Jurassic transgressions and regressions influenced the marine biodiversity in the Caucasus. However, direct relationships between them are obvious for the ammonites only, in contrast to bivalves, brachiopods and belemnites. On a global scale, marine biodiversity corresponded well to the eustatic changes.

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The palaeogeographic outlines of the Caucasus in the Jurassic: The Caucasian Sea and the Neotethys Ocean

Dmitry A. Ruban

Abstract. The Caucasian Sea, fringing the northern margin of the Jurassic Neotethys Ocean, largely covered the Caucasus. Continental, shallow-marine and deep-marine palaeoenvironments delineate palaeogeographic outlines for three significant time slices: the Late Toarcian, the Early Bajocian and the Middle Oxfordian. These new palaeogeographic outlines of the Caucasus and adjacent territories match the Neotethys Ocean reconstructions. In the Late Toarcian, the Caucasian Sea embraced the Greater Caucasus Basin and the Black Sea – Caspian Sea Basin, which were divided by the Northern Transcaucasian Arc; it opened to the Neotethys Ocean which covered the Exterior Caucasian Basin. In the Early Bajocian, the Caucasian Sea only embraced the Greater Caucasus Basin; it opened the epicontinental seas of the Russian Platform, connecting them with the Neotethys Ocean by straits between islands of the Transcaucasian Arc. In the Middle Oxfordian, the Caucasian Sea which further embraced the Greater Caucasus Basin had its outer shelf fringed by carbonate build-ups. The connection between the Russian Platform shallow sea and the Neotethys Ocean was maintained. In the course of the Jurassic, a seaway developed along the northern margin of the Neotethys, of which the Caucasian Sea became a significant part.

Key words: sea, seaway, basin, arc, Jurassic, Caucasus, Neotethys.

Апстракт. Кавкаско море захватало је северни обод јурског Неотетиског океана и великим делом је прекривало Кавказ. Континенталне, плитководне и дубоководне палеосредине оцртавају палеогеографске оквире три значајна временска раздобља: горњи тоар, доњи бајес и средњи оксфорд. Ове нове палеогеографске границе Кавказа и суседних области уклапају се у реконструкцију Неотетиског океана. У горњем тоару Кавкаско море је обухватало Велики Кавкаски басен и Црно море – Каспијски морски басен, који су били раздвојени Северним транскавкаским луком који се отварао према Неотетиском океану који је прекривао спољашњи Кавкаски басен. За време доњег бајеса, Кавкаско море је захватало само Велики Кавкаски басен; оно је било отворено према епиконтиненталном мору Руске платформе повезујући га са Неотетиским океаном земљоузима између острва Транскавкаског лука. Током средњег оксфорда стварале су се карбонатне наслаге по ободу спољашњег шелфа Кавкаског мора, које је и даље захватало Велики Кавкаски басен. Одржавала се веза између плитководне Руске платформе и Неотетиског океана. У току јуре постојао је морски пролаз дуж северног обода Неотетиса, где је Кавкаско море заузимало његов значајни део.

Кључне речи: море, морски пролаз, басен, лук, јура, Кавказ, Неотетис.

Introduction

The Caucasus stretches over about 1000 km between the Black and Caspian seas (Fig. 1). In the Jurassic, it was located on the northern margin of the Neotethys Ocean, forming a “key” transition between western and central parts of the Northern Neotethys (Stampfli & Borel, 2002; Golonka, 2004). Not only palaeogeographically and palaeotectonically, but also palaeobiogeographically, the Caucasus was an important region. Westermann (2000) after Uhlig (1911) have defined the Mediterran-Caucasian Subrealm of the Mesozoic Tethyan Realm.

In spite of its importance, the Caucasusian Jurassic palaeogeography is still poorly known. Previous publications are often only available in Russian and/or lack

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the incorporation of modern palaeogeographic and palaeotectonic concepts. Outdated "formation" analysis or geosynclinal theory are the basis of many studies. To date, plate-tectoinc and terrane analysis of the Caucasus still remains sporadic and schematic. In many Russian reconstructions, the Caucasus was viewed as an isolated region and its border often delineated by the boundaries of the former USSR. To avoid misunderstanding, which is inevitable when dealing with a high amount of the sufficiently reliable sources, in this paper only a few Russian works have been considered. The first one is a book by Jasamanov (1978), who presented general palaeogeographic information on the Caucasus for each of the Jurassic stages, while the second is a review by Lordkipanidze *et al*. (1984), who presented the most acceptable palaeotectonic reconstructions, based on palaeomagnetic data. Tectonic models proposed by Ershov *et al*. (2003) were also employed.

Fig. 1. Geographic location of the studied region. GE – Georgia, AR – Armenia, AZ – Azerbaijan. The position of the Caucasus in the Jurassic is shown on the palaeogeographic map, strongly simplified after Scotese (2004).

Thus, in any somewhat more detailed palaeogeographic reconstruction of a larger portion of the Jurassic Northern Neotethys, the Caucasus remained a blank space. The general target of this article is to initiate a discussion on the highlighted topic. Simplified Jurassic palaeogeographic outlines of the Caucasus are proposed and discussed. It should be emphasized that this attempt is based on personal field investigations, as well as a revision of the available and trustwarthy data from recent studies of the entire Jurassic Neothethys (Stampfli & Borel, 2002; Golonka, 2004).

Geologic setting

The Caucasus consists of three main segments: (1) the Greater Caucasus, (2) the Lesser Caucasus (or the Transcaucasus) and (3) the Kura-Rioni Depression (or the Rioni Depression and the Kura Depression, also called the Transcaucasian Depressions) (Fig. 1). Their tectonic settings have been briefly overviewed by Saintot & Angelier (2002), Allen *et al*. (2003) and Ershov *et al*. (2003).

The Jurassic deposits, widely distributed within the Caucasus, vary in distinct areas. Their stratigraphy has been reviewed by Rostovtsev *et al*. (1992). The stratigraphic scale used in the Caucasus was revised by the author according to new developments in the Jurassic chronostratigraphy, using ammonoids, brachiopods (for detail see Ruban, 2003), foraminifers and marker horizons (Fig. 2). Stratigraphic suggestions from both International Commission on Stratigraphy and the Groupe Français d'Étude du Jurassique (Cariou & Hantzpergue, 1997) were taken account in doing this. A correspondence between the chronstratigraphic stages and substages (after Gradstein *et al*., 2004) and stages in the regional sense (after Rostovtsev *et al*., 1992) was established. The precise revision of the regional ammonoid-based zonation is a task for further special studies. It is also necessary to note that traditionally the Callovian stage in the Caucasus is attached to the Upper Jurassic (Rostovtsev *et al*., 1992), in contrast to the present scale, recommended by the International Commission on Stratigraphy (Gradstein *et al*., 2004).

Jurassic lithostratigraphy of the Caucasus has been reviewed in detail by Rostovtsev *et al*. (1992). In general, two major sedimentary complexes are identified. The Sinemurian-Bathonian complex comprises argillaceous and clastic deposits with a total thickness upto 10000 m. The Callovian-Tithonian complex is represented chiefly by carbonates (thickness up to 3000 m) and also evaporites in the upper part. The accumulation of the Late Jurassic deposits was connected with the evolution of a large carbonate platform rimmed by carbonate buildups (Kuznetsov, 1993; Akhmedov *et al*., 2003; Ruban, 2005). In some areas (especially in the Lesser Caucasus), substantial amounts of volcanoclastic deposits are present. Two major regional hiatuses encompass the Hettangian-Early Sinemurian and the Bathonian.

In the Jurassic, the Caucasus was located in the central part of the northern margin of the Neotethys Ocean (Fig. 1) (Stampfli & Borel, 2002; Golonka, 2004). Tectonic activity resulted from the dynamics between the terranes, which contacted with each other, and also with the larger Eurasian Plate. Several parallel subduction and spreading zones were located in this territory (Lordkipanidze *et al*., 1984; Ershov *et al*., 2003), although a precise interpretation of the Jurassic geodynamics in this region has not been made yet and many questions remain open.

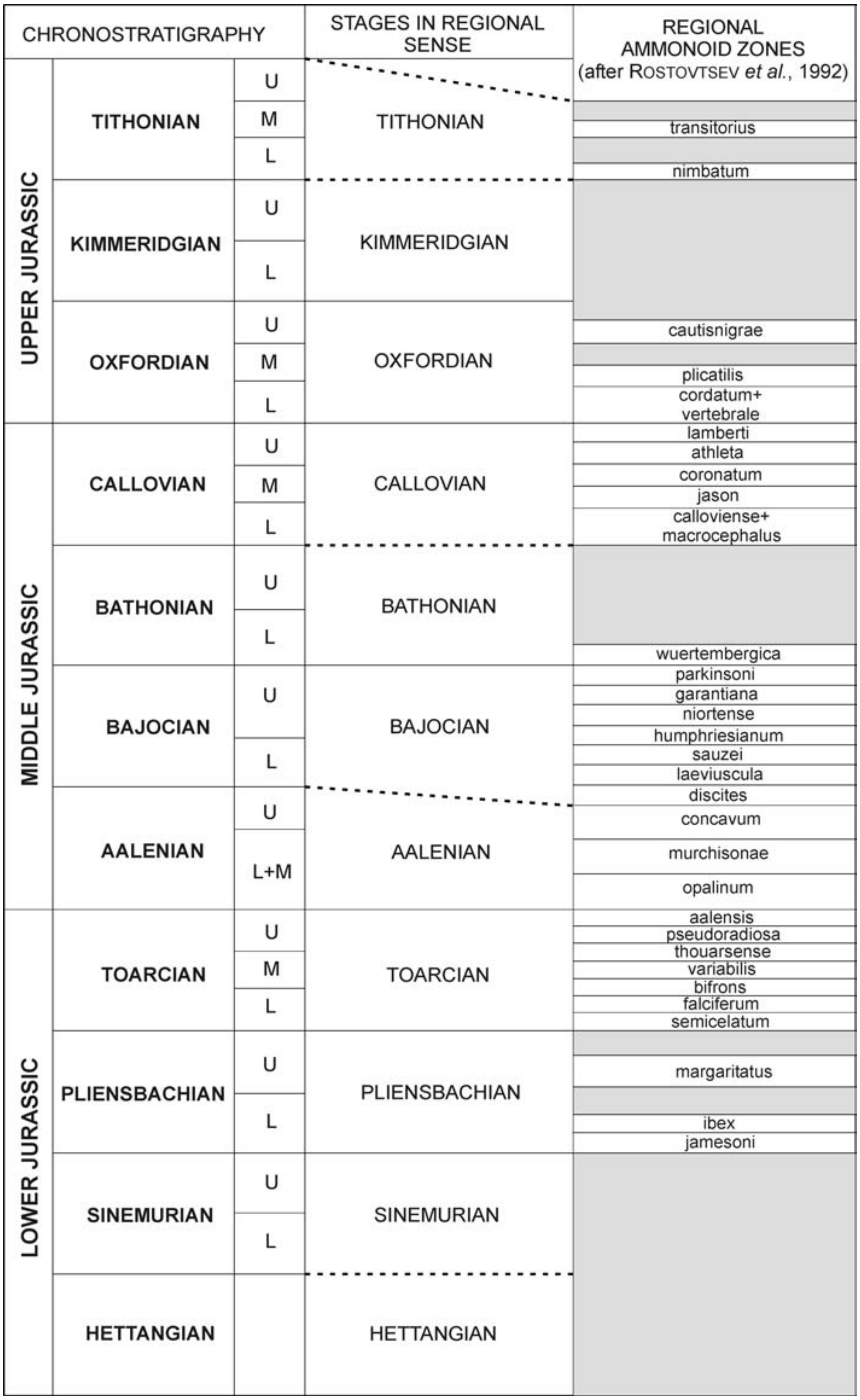

Fig. 2. Corrected stratigraphic scale of the Jurassic used in the Caucasus. Abbreviations: L – Lower, M – Middle, U – Upper. Unzoned intervals are shaded as gray. Dashed lines mark uncertainty in the boundary definition. Regional ammonoid zonation does not correspond on this scale to the shown chronostratigraphy (it seems to be impossible to correlate them at present), but only to the stages in a regional sense. The Callovian *macrocephalus* and calloviense regional zones, and the Oxfordian *vertebrale* and *cordatum* regional zones are evidently not separated in the regional ammonoid succession.

In the Early-Middle Jurassic, the Caucasus was located in a subtropical to moderate humid zone. The Toarcian palaeotemperatures are estimated as 15–20°C; in the Early Aalenian, they decreased to 5–15°C, but in the Late Aalenian, the temperatures increased again to 20–25°C, and apparently constant until the end of the Jurassic (Jasamanov, 1978). After the beginning of the Callovian, the climate became subtropical to tropical and semi-humid. In the Late Kimmeridgian-Tithonian, evaporites were accumulated (Jasamanov, 1978; Rostovtsev *et al.*, 1992), which indicated arid conditions. In the Early-Middle Jurassic, dysoxic to anoxic palaeoenvironments were typical for the Caucasian basins (Ruban, 2004; Efendiyeva & Ruban, 2005; Ruban & Tyszka, 2005). The palaeobiogeographic position of the Caucasus is uncertain. While Dommergues (1987) places it in the Euro-Boreal domain for the Early Jurassic, Westermann (2000) includes it into the Tethyan Realm. An analysis of brachiopods suggests a rather transitional position (Ruban, 2003).

Methods

Essentially, this study relies on palaeoenvironmental interpretation, realized in the same way as described by Ruban (2006). The territory of the Caucasus is subdivided into several dozens of particular areas, which are traditionally called “zones”. They are distinguished by the facies composition of the Jurassic succession. A total of 36 “zones” delineate the Hettangian-Bathonian interval (Fig. 3A), and 26 the Callovian-Tithonian interval (Fig. 3B) (Rostovtsev *et al.*, 1992). A palaeoenvironmental interpretation for all formations in each “zone” was made. The comprehensive information of Rostovtsev *et al.* (1992) and personal field observations in the Labino-Malkinskaya (see also Efendiyeva & Ruban, 2005; Ruban & Tyszka, 2005), Lago-Nakskaja and Labinskaja “zones” were used.

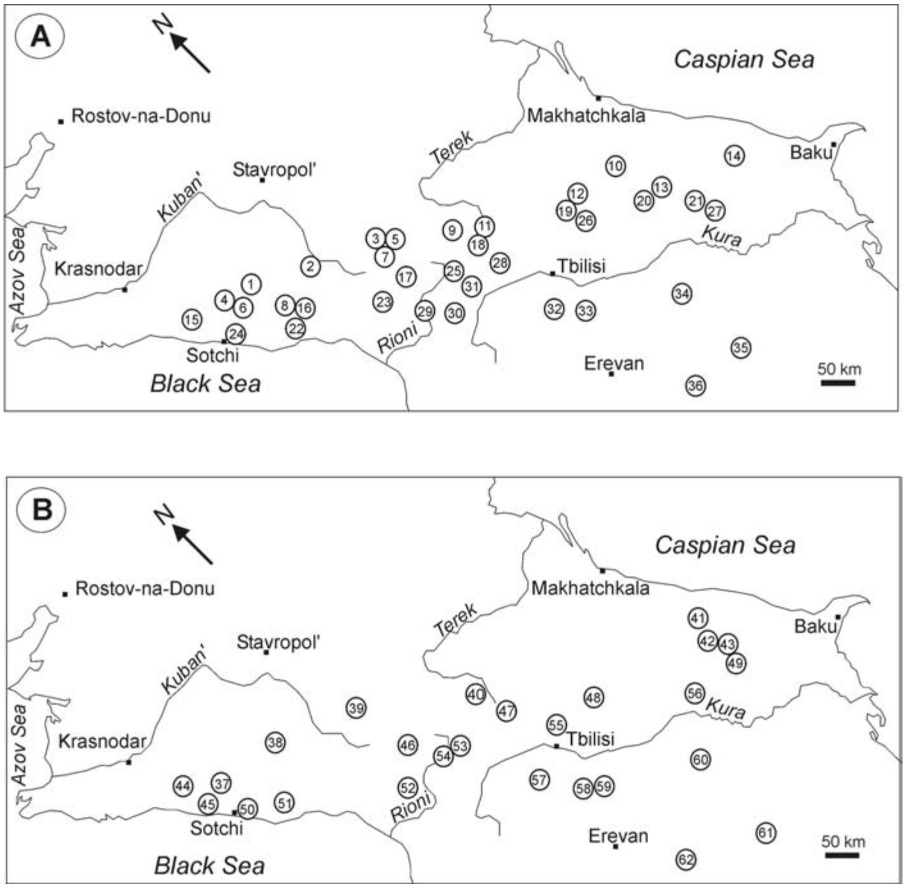

Fig. 3. Location of the Jurassic "zones" (marked by circles) in the Caucasus (after ROSTOVTSEV *et al.*, 1992). A, Hettangian-Bathonian "zones" (1–36); B, Callovian-Tithonian "zones" (37–62). "Zones" ("subzones" and regions of ROSTOVTSEV *et al.* (1992) are mentioned here as "zones"): 1, Western Labino-Malkinskaja; 2, Central Labino-Malkinskaja; 3, Eastern Labino-Malkinskaja; 4, Western Pshikish-Tyrnyauzskaja; 5, Eastern Pshikish-Tyrnyauzskaja; 6, Northern Arkhyz-Guzeripl'skaja; 7, Eastern Arkhyz-Guzeripl'skaja; 8, Southern Arkhyz-Guzeripl'skaja; 9, Digoro-Osetinskaja; 10, Agwali-Khivskaja; 11, Western Bokovogo Khrebta; 12, Central Bokovogo Khrebta; 13, Eastern Bokovogo Khrebta; 14, Southeastern Bokovogo Khrebta; 15, Gojtkhsko-Atchishkhinskaja; 16, Severoabkhazskaja; 17, Svanetskaja; 18, Western Glavnogo Khrebta; 19, Central Glavnogo Khrebta; 20, Tfanskaja; 21, Durudzhinskaja; 22, Western Gagra-Dzhavskaja; 23, Eastern Gagra-Dzhavskaja; 24, Amuksko-Lazarevskaja; 25, Sakaojskaja; 26, Shakrianskaja; 27, Vandamskaja; 28, Kakhetino-Letchkhumskaja; 29, Tskhenistskali-Okribskaja; 30, Southwestern Dzirul'skaja; 31, Northeastern Dzirul'skaja; 32, Loksko-Khramskaja; 33, Alaverdskaja; 34, Shamkhorsko-Karabakhskaja; 35, Kafanskaja; 36, Araksinskaja; 37, Lago-Nakskaja; 38, Labinskaja; 39, Malkinskaja; 40, Kabardino-Dagestanskaja; 41, Jugo-Vostotchnogo Dagestana; 42, Sudurskaja; 43, Shakhdagskaja; 44, Abino-Gunajskaja: 45, Novorossijsko-Lazarevskaja, 46, Svanetsko-Verkhneratchinskaja; 47, Liakhvi-Aragvinskaja; 48, Kakhetinskaja; 49, Dibrarskaja; 50, Akhtsu-Katsyrkha; 51, Dzhirkhva-Akhibokhskaja; 52, Tkvartcheli-Okribskaja; 53, Ratchinskaja; 54, Tsessi-Kortinskaja; 55, Iori-Tsitelitskarojskaja; 56, Vandamskaja; 57, Khramskaja; 58, Lalvarskaja; 59, Idzhevanskaja; 60, Dashkesano-Karabakhskaja; 61, Kafanskaja; 62, Nakhitchevanskaja.

Three main types of the palaeoenvironments were distinguished in general: continental, shallow-marine and deep-marine. Continental palaeoenvironments were usually documented by the hiatuses, while rarely by the subaeral deposits. Shallow-marine palaeoenvironments were interpreted by the presence of clastic or carbonate deposits, similar to those usually accumulated at a seashore or on a shelf. Deep-marine palaeoenvironments were traced mostly by the slope deposits (e.g., turbidites). In addition to lithology, also fossils, including plant remains, as well as sedimentological criteria, such as submarine slumps, concretions, etc., were used to determine the palaeoenvironments.

Special attention was paid to three time slices: the Late Toarcian, the Early Bajocian and the Middle Oxfordian, which all correspond to important phases in the evolution of the Caucasus. In the Late Toarcian, all the principal basins of the Caucasus were formed completely. The Early Bajocian and the Middle Oxfordian correspond to the time invervals after something like reorganizations of the Caucasian basins occurred, each following major regressions.

Maps showing the variety of the palaeoenvironments during these time slices were drawn for the Caucasus (Figs. 4A, 5A, 6A). They are attached to the present-day geography of the studied region. Therefore, the next step was to take into consideration the palaeotectonic reconstructions. In this paper, the reconstructions of LORDKIPANIDZE *et al*. (1984) were preferred, because they are based on reliable palaeomagnetic data. Additionally, the results of ERSHOV *et al.* (2003) were considered. Analyzing the composed maps of the palaeoenvironment distribution, attempt were made to recognize palaeogeographic elements (basins, arcs) highlighted by LORDKIPANIDZE *et al*. (1984), and, when necessary, correct their location. Then the verified and corrected information from the Caucasus was incorporated into the reconctructions for the entire Neotethys made by STAMPFLI & BOREL (2002) and GOLONKA (2004). Additionally, reconstructions made for the Pliensbachian by MEISTER & STAMPFLI (2000) became very helpful.

The final result, a set of the palaeogeographic sketches delineates what was the outline of the Caucasus at each of the studied time slices (Figs. 4B, 5B, 6B). They embrace the whole territory of the Caucasus and adjacent regions, including the Pontides, Moesia, Iranian terranes and the southern periphery of the Eurasia continent. Although these sketch-maps remain at a relatively low resolution and the position of landmasses (i.e., continents and islands) is schematic, they may help to fill the gap in our knowledge of the Jurassic palaeogeography of the Caucasus.

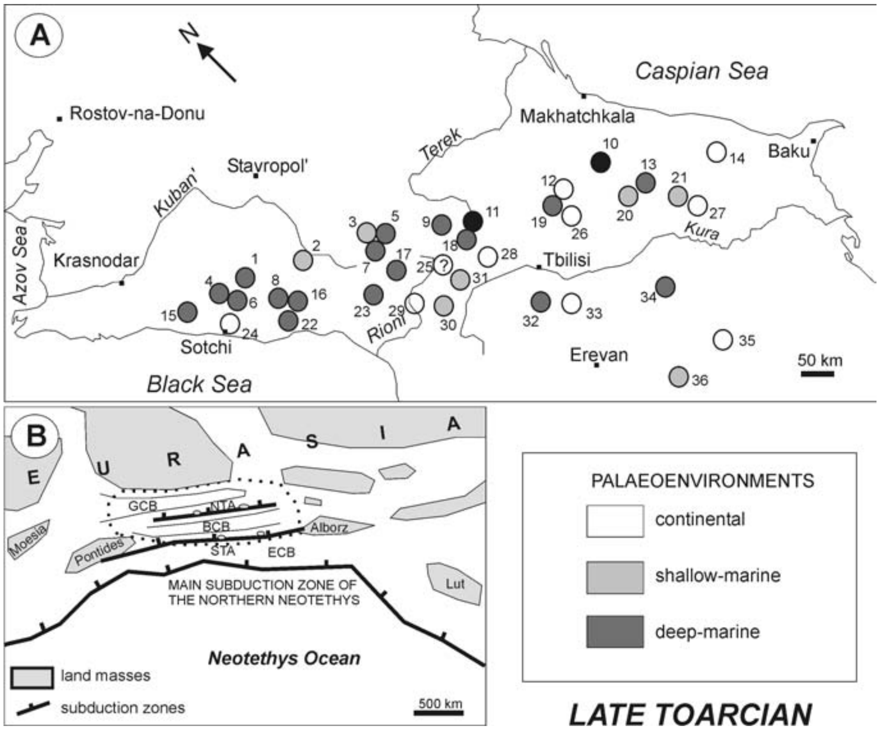

Fig. 4. The Late Toarcian palaeoenvironments (A) and the palaeogeographic outline (B) of the Caucasus (explanation of "zones" – Fig. 3A). "?" marks uncertainty in the interpretation of the continental palaeoenvironments, because of the doubtful establishment of hiatus. GCB, Greater Caucasus Basin, BCB, Black Sea – Caspian Sea Basin, ECB, Exterior Caucasian Basin, NTA, Northern Transcaucasian Arc, STA, Southern Caucasian Arc. The dotted line bounds the Caucasian Sea.

Reconstructions of the Jurassic outlines of the Caucasus

The Late Toarcian (~ 177 Ma)

Marine palaeoenvironments prevailed over most of the Caucasus in the Late Toarcian (Fig. 4A). In its northern part, dominating deep-marine environments trace the elongated basin, which may evidently correspond to the Greater Caucasus Basin of LORDKIPANIDZE *et al*. (1984). Perhaps its western part was the widest and deepest. Sporadic shallow-water environments to the south support the idea of the presence of the Northern Transcaucasian Arc (LORDKIPANIDZE *et al.,* 1984), related to the subduction zone. Moreover, there is no sound evidence for the presence of a large landmass there, as this is usually imagined (e.g., JASAMANOV, 1978). Presumably, only small islands might have been related to this arc.

Another deep basin is weakly delineated southwards, which may be related to the Black Sea – Caspian Sea Basin of LORDKIPANIDZE *et al*. (1984). Shallow-water environments to the south of it correspond potentially to the Southern Transcaucasian Arc, i.e. another subduction zone. This arc is considered as the eastern edge of the Pontide structure (LORDKIPANIDZE *et al.,* 1984). In our palaeoenvironmental interpretation, there is no evidence to recognize the Lesser Caucasus Strait of the Tethys and the Nakhitchevan' Block, which were shown by LORDKIPANIDZE *et al*. (1984). Another basin, with the proposed name "the Exterior Caucasian Basin", might have been located between the Southern Transcaucasian Arc and the main subduction zone of the Northern Neotethys.

In the Late Toarcian outline of the Caucasus (Fig. 4B), a large sea, for which the name Caucasian Sea is proposed, opens towards the Neotethys Ocean. Wide straits between the landmasses to the west and east of this region entered this sea. The Caucasian Sea embraced two sedimentary basins, divided by a submarine mountain range, united perhaps to the west. Possibly, two archipelagoes consisting of very small islands which formed the Northern and Southern Transcaucasian Arcs characterized this sea. The boundary between the Caucasian Sea and the Neotethys Ocean stretched along the Southern Transcaucasian Arc. Our sketch-map suggests that the Exterior Caucasian Basin was embraced by the Neotethys Ocean.

The Early Bajocian (~ 171 Ma)

The Early Bajocian times were characterized by laterally variable palaeoenvironments within the Caucasus (Fig. 5A). Deep-marine environments trace the Greater Caucasus Basin, while shallow-water and continental

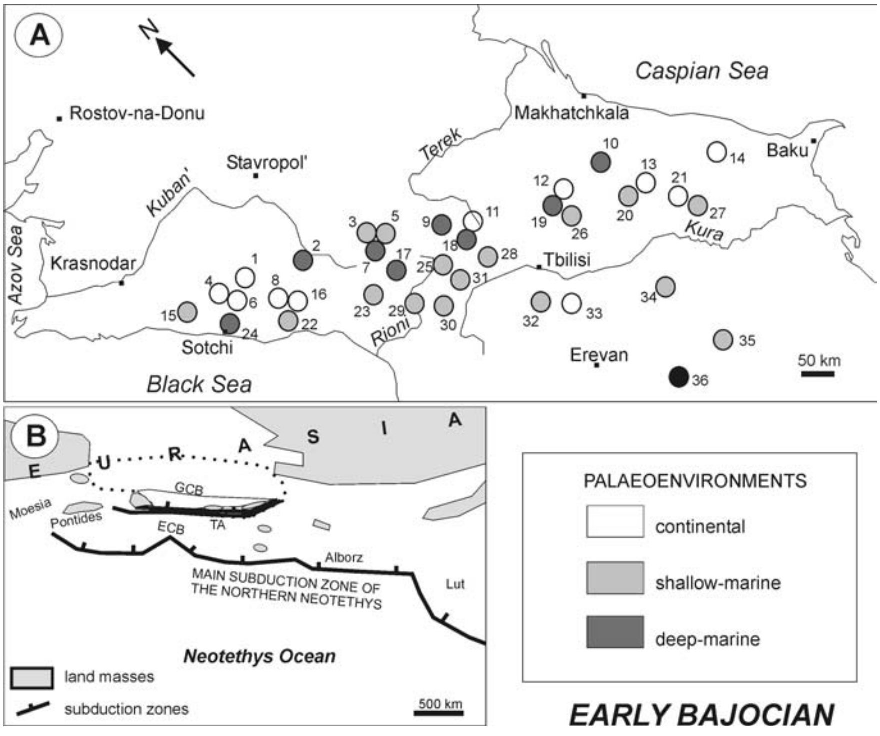

Fig. 5. The Early Bajocian palaeoenvironments (A) and the palaeogeographic outline (B) of the Caucasus (explanation of "zones" – Fig. 3A). GCB, Greater Caucasus Basin; ECB, Exterior Caucasian Basin; TA, Transcaucasian Arc. The dotted line bounds the Caucasian Sea.

environments delineate an island arc to the south, including relatively large islands. Intriguing was the landmass in the western part of the studied territory, where several continental deposits suggest large islands, which appeared as the result of the collision between Northern and Southern Transcaucasian Arcs, which closed the Black Sea – Caspian Sea Basin, generating a single Transcaucasian Arc. Palaeomagnetic data that highlight the presence of the Black Sea – Caspian Sea Basin in the Middle Jurassic appear doubtful (Lordkipanidze *et al*. 1984). The Exterior Caucasian Basin was located between the Transcaucasian Arc and the main subduction zone of the Northern Neotethys. Some islands can be locally evidenced from continental palaeoenvironments.

The Early Bajocian outline of the Caucasus is presented in Fig. 5B. The studied territory was occupied by the Caucasian Sea. It was isolated from the Neotethys Ocean by the island archipelago of the Transcaucasian Arc. Connection between the sea and ocean was realized by straits between these islands, as well as landmasses, located to the west. From the north, the Caucasian Sea was opened to the large, but shallow interior sea, occuping a waste area of the Russian Platform. Only one sedimentary basin was embraced by this sea. The transgression resulted in the appearance of a very large shelf to the north of this basin, and the structure of the sea in the Early Bajocian was characterized by a strong asymmetry. The boundary between the Caucasian Sea and the Neotethys Ocean stretched along the Transcaucasian Arc. Our sketch-map suggests that the Exterior Caucasian Basin was embraced by the Neotethys Ocean. The islands occurring there might have been of volcanic origin and, therefore, related to the wide belt of intense magmatism to the north of the main subduction zone of the Northern Neotethys.

The Middle Oxfordian (~ 158 Ma)

During the Middle Oxfordian, the Caucasus was dominated by shallow-marine palaeoenvironments (Fig. 6A). Marine environments trace the Greater Caucasus Basin. The composed map does not permit the idea of Lordkipanidze *et al*. (1984) about the complete separation of the Western and Eastern Subbasins and the presence of island between them, to be supported. We observed deep-marine environments in the western, central and eastern parts of the Greater Caucasus Basin. Nevertheless, the existence of islands at the western and eastern edges of the latter, hypothesized by Lorkipanidze *et al*. (1984) and also by Golonka (2004), is confirmed by our results, because continental palaeoenvironments were interpreted for those areas. Another island (or a chain of islands), delineated by the continental environments to the south, may be related to the Transcaucasian Arc. In contrast to Lordkipanidze *et al*. (1984), no evidence for the presence of the Northern and Southern Transcaucasian Arcs, separated by the Black Sea – Caspian Sea Basin, was found. Therefore, it is hypothesized that in the Middle Oxfordian, a unique arc existed, as it was already in the Early Bajocian. However, this arc migrated southwards in comparison with the earlier time slices. Shallow-marine environments in the south of the studied territory are attributed to the Exterior Caucasian Basin.

The Middle Oxfordian outline of the Caucasus is presented in Fig. 6B. The studied territory was oc-

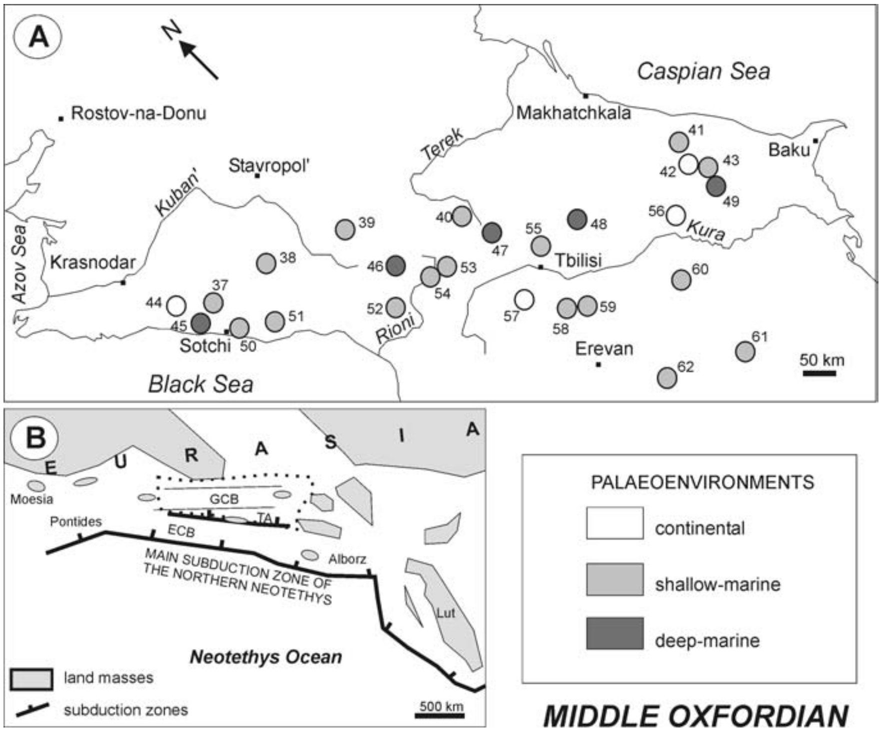

Fig. 6. The Middle Oxfordian palaeoenvironments (A) and the palaeogeographic outline (B) of the Caucasus (for an explanation of the "zones", see Fig. 3B). GCB, Greater Caucasus Basin; ECB, Exterior Caucasian Basin; TA, Transcaucasian Arc. Dotted line bounds the Caucasian Sea.

cupied by the Caucasian Sea. It was only a little isolated from the Neotethys Ocean by the above mentioned island and submarine mountain range of the Transcaucasian Arc. Straits between landmasses to the west and east of this sea also connected it with the Neotethys Ocean. From the north, the Caucasian Sea opened into the interior sea, as in the Early Bajocian, but its area was diminished. Only one sedimentary basin was embraced by this sea. A large shelf existed to the north-east of this basin. Thus, the sea basin in the Middle Oxfordian was characterized by strong asymmetry in its eastern part, but it was quite symmetric in its western part. The boundary between the Caucasian Sea and the Neotethys Ocean stretched along the Transcaucasian Arc. Our sketch-map suggests that the Exterior Caucasian Basin was embraced by the Neotethys Ocean.

A distinctive feature of the Late Jurassic of the Caucasian basins was the wide distribution of carbonate buildups (Jasamanov, 1978; Khain, 1962; Lordkipanidze *et al.*, 1984; Kuznetsov, 1993; Martin-Garin *et al.*, 2002; Rostovtsev *et al.*, 1992; Akhmedov *et al.*, 2003; Cecca *et al.*, 2005; Ruban, 2005). This coincided with the reef growth documented on the entire northern margin of the Neotethys Ocean (Kiessling *et al.*, 1999; Leinfelder *et al.*, 2002; Martin-Garin *et al.*, 2002; Olivier *et al.*, 2004; Cecca *et al.*, 2005). The term "carbonate buildups" is preferred to that of "reefs", as they are traditionally called in Russian literature (e.g., Jasamanov, 1978; Khain, 1962; Rostovtsev *et al.*, 1992). Schmid *et al.* (2001) mentioned the Caucasian buildups as mounds. The carbonate buildups are concentrated around the deepest parts of the Greater Caucasus Basin (Fig. 7). It is suggested that to the north, they developed on the outer shelf periphery, connected to the stable landmass of the Russian Platform, while in the south, they occupied the narrow outer shelf of the Transcaucasian Arc. However, some buildups were also found crossing the basin, suggesting atolls, isolated or in groups, characterizing the Late Jurassic Caucasian Sea and Exterior Caucasian Basin. In general, the distribution of the carbonate buildups was tectonically controlled (Khain, 1962; Akhmedov *et al.,* 2003).

Discussion

The presented palaeogeographic sketch maps suggest that during the Jurassic, the Caucasian Sea was located between the Eurasian landmass and large and little islands (Figs. 4B, 5B, 6B). A string of large islands located west- and eastwards were the result of accretion of small terranes along the subducted margin of the northern Neotethys. Straits between these small landmasses made a connection with the Caucasian Sea possible. Together they were able to form an important seaway that stretched along the southern periphery of Eurasia. The tectonic origin of this Exterior Caucasian seaway is very different from thopse of the well-known Hispanic Corridor and the Viking Corridor, the results of break-up of continents (Hallam, 1983; Smith & Tipper, 1986; Riccardi, 1991; Westermann, 1993; Aberhan, 2001). It also differed from the other seaways, lsuch as the Cretaceous Western Interior Seaway

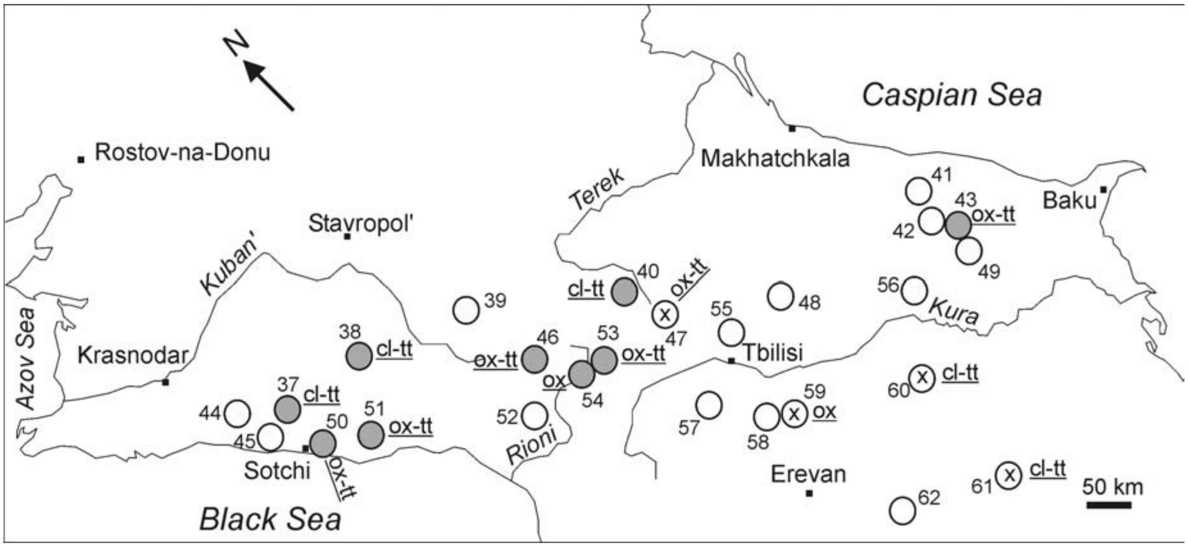

Fig. 7. The location of the Callovian-Tithonian carbonate buildups in the "zones" of the Caucasus (data was extracted from ROSTOVTSEV *et al*. (1992); for the western part of the Caucasus, it was supported by personal field observations) (for an explanation of the "zones", see Fig. 3B). "Zones" where carbonate buildups were evidently documents are highlighted as gray, while "zones" where only coral founds are known are marked by x. The age of buildup- or coral-bearing deposits is indicated.

in North America (REYNOLDS & DOLLY, 1983; SAGEMAN & ARTHUR, 1994; ROBERTS & KIRSCHBAUM, 1995; WHITE *et al.,* 2001, 2002).

The western branches of this seaway included the oceanic basins of the Western Neotethys, such as the Meliata, Maliac, Pindos and Vardar, as well as the Alpine Tethys which opened during the Jurassic (STAMPFLI & BOREL, 2002; BROWN & ROBERTSON, 2004; GOLONKA, 2004). The central part of the seaway consisted of straits separating the blocks of Moesia, Rhodope and Western Pontides. Further east, the seaway communicated with the small Izmir-Ankara Ocean (STAMPFLI & BOREL, 2002). It was directly connected with the Caucasian Sea. The eastern branches of the mentioned seaway extended as straits between the Alborz, South Caspian, Aghdarband, Herat and other terranes of the central part of the northern Neotethyan margin (GOLONKA, 2004). STAMPFLI & BOREL (2002) additionally placed the so-called South Caspian Ocean eastwards of the Caucasus, which seems to be a fragment of the seaway. The latter ended in two branches, as is suggested from the palaeoreconstructions of GOLONKA (2004). Northwards, the seaway connected the basin between the Turan, Herat and Pamirs landmasses, while southwards it led directly to the Neotethys Ocean.

Conclusions

This study of the Jurassic palaeogeography of the Caucasus allows the formulation of some important conclusions:

1) the Caucasian Sea occupied most, although not all, of the studied area during the entire Jurassic;

2) in the Late Toarcian, the Caucasian Sea embraced most of the Caucasus, including the Greater Caucasus Basin and the Black Sea – Caspian Sea Basin, and was opened to the Neotethys Ocean, which covered the Exterior Caucasian Basin;

3) in the Early Bajocian, the Caucasian Sea comprised the Greater Caucasus Basin, it opened to the epicontinental seas of the Russian Platform, and it was connected with the Neotethys Ocean by the straits between islands of the Transcaucasian Arc;

4) in the Middle Oxfordian, the Caucasian Sea also covered the Greater Caucasus Basin and was open to both the epicontinental sea of the Russian Platform and the Neotethys Ocean;

5) during the Jurassic, the Caucasus was included in the long seaway, which stretched along the northern margin of the Neotethys.

Further studies are necessary to verify and detalize the very simple palaeogeographic reconstructions proposed in this paper. A significant task is the collection of data on the carbonate buildups, which has already been made for the Azerbaijanian part of the Caucasus (AKHMEDOV *et al*., 2003). These data will help to delineate the Late Jurassic carbonate platform. Special attention should also be paid to the high-resolution palaeotectonic interpretations.

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Резиме

Палеогеографски оквири Кавказа у Јури: Кавкаско море и Неотетиски океан

Кавказ представља пространу издужену област између Црног мора и Каспијског језера, која обухвата југоисточну Русију и целу Грузију, Јерменију и Азербејџан. Ова област се за време јуре налазила на северном ободу Неотетиског океана и представљала је "главни" прелаз између западног и средњег дела северног Неотетиса. Ранија тумачења њене палеогеографије су застарела и често заснована на погрешним схватањима. Овај рад је покушај да се реконструишу палеогеографски оквири Кавказа у три временска раздобља – горњи тоар, доњи бајес и средњи оксфорд. Територија Кавказа је подељена на 62 посебне области, које се традиционално називају "зонама". За сваку зону је дато тумачење палеосредине свих формација и то континентална, плитководна и дубоководна палеосредина за поменута временска раздобља. Анализом карата распореда палеосредина утврђени су палеогеографски елементи података за Кавказ у реконструкције читавог Неотетиса. Као коначни резултат добијен је скуп палеогеографских скица које дају контуру Кавказа из свих проучаваних раздобља.

У горњем тоару (око 177 Ма), пространо море, за које се предлаже назив Кавкаско море, отварало се према Неотетиском океану. У ово море су залазили широки мореузи између копна са западне и источне стране ове области. Кавкаско море је обухватало два седиментациона басена, раздвојена подводним планинским ланцем, који су се можда спајали на западу. То море су вероватно карактерисала два архипелага врло малих острва која су образовала северни и јужни транскавкаски лук. Граница између Кавкаског мора и Неотетиског океана протезала се дуж јужног транскавкаског лука. Спољашњи, јужни кавкаски басен био је опкољен Неотетиским океаном.

У доњем бајесу (око 171 Ма) проучавану територију је заузимало Кавкаско море. Оно је било одвојено од Неотетиског океана архипелагом транскавкаског лука. Веза између мора и океана остваривала се мореузима између острва и копна на западу.

На северу се Кавкаско море отварало према великом али плитком унутрашњем мору које је заузимало пространу област руске платформе. То унутрашње море је обухватало само један седиментациони басен. Резултат трансгресије је формирање великог шелфа на северу басена, а структура мора се у доњем бајесу карактерисала великом асиметријом. Граница између Кавкаског мора и Неотетиског океана, као и у претходном случају, протезала се дуж транскавкаског лука. Спољни басен је био обухваћен Неотетиским океаном. Острва која су тамо постојала могла су бити вулканског порекла, па су према томе у вези са широким појасом интензивног магматизма северно од главне зоне подвлачења северног Неотетиса.

У средњем оксфорду (око 158 Ма) је проучавана територија још увек била под Кавкаским морем. Била је само мало изолована од Тетиског океана поменутим подморским планинским ланцем транскавкаског лука. Мореузи између копна западно и источно од овог мора такође су га повезивала са Неотетиским океаном. Са севера се Кавкаско море отварало према унутрашњем мору као и у доњем бајесу, али се његова површина смањила. Море је обухватало само један седиментациони басен. Велики шелф је постојао на североистоку басена. Тако се морски басен у средњем оксфорду карактерисао великом асиметријом у источном делу, али му је западни део био доста симетричан. Граница између Кавкаског мора и Неотетиског океана протезала се дуж транскавкаског лука. Схематска карта указује да је спољашњи кавкаски басен био обухваћен Неотетиским океаном.

Широка распрострањеност карбонатних наслага представља значајну карактеристику кавкаских басена горње јуре. Током јуре је створен морски пролаз дуж обода Неотетиса, а Кавкаско море је постало његов значајни део. Западни огранци овог морског пролаза обухватили су океанске басене западног Неотетиса као што су Мелиата, Малиак, Пинд и Вардар као и алпски Тетис. Средњи део морског пролаза састојао се од мореуза који су раздвајали блокове Мезије, Родопа и западних Понтида. Даље на исток морски пролаз је био у вези са малим Измирско-анкарским океаном, који је био директно повезан са Кавкаским морем. Источни огранци морског пролаза настављали су се у мореузе између Алборза, јужног Каспија, Агдарбанда, Херата и других терана средњег дела северног обода Неотетиса. Такозвани Јужни каспијски океан источно од Кавказа је по свој прилици представљао део басена између туранског, хератског и памирског копна, док је на југу водио директно у Неотетиски океан.

ГЕОЛОШКИ АНАЛИ БАЛКАНСКОГА ПОЛУОСТРВА
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Trans-border (east Serbia/west Bulgaria) correlation of the Jurassic sediments: main Jurassic paleogeographic units

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Abstract. In the region across the Serbian/Bulgarian state border, there are individualized 5 Jurassic paleogeographic units (from West to East): (1) the Thracian Massif Unit without Jurassic sediments; (2) the Lužnica-Koniavo Unit – partially with Liassic in Grsten facies and with deep water Middle Callovian–Kimmeridgian (*p. p.*) sediments of the type "*ammonitico rosso*", and Upper Kimmeridgian–Tithonian siliciclastics flysch; (3) The Getic Unit subdivided into two subunits – the Western Getic Sub-Uni – without Lower Jurassic sediments and the Eastern Getic Sub-Unit with Lower Jurassic continental and marine sediments, which are followed in both sub-units by carbonate platform limestones (type Stramberk); (4) the Infra (Sub)-Getic Unit – with relatively deep water Liassic and Dogger sediments (the Dogger – of type "black shales with *Bossitra alpina*") and Middle Callovian–Tithonian – of type "*ammonitico rosso*"; (5) the Danubian Unit – with shallow water Liassic, Dogger and Malm (Miroč–Vrška Čuka Zone, deep water Dogger and Malm (Donjomilanovačko–Novokoritska Zone).

Key Words: Jurassic, paleogeographic units, south-eastern Serbia, western Bulgaria.

Апстракт: У подручју српско-бугарске државне границе издвојено је пет јурских палеогеографских јединица (од запада ка истоку): 1. Тракијски масив без јурских седимената; 2. Лужница–Кониаво – делимично са лијасом развијеним у грестенској фацији и са дубоководним седиментима средњег каловеја–кимерица (*p. p.*) типа "*ammonitico rosso*" и силикокластичним флишом горњег кимерица–титона; 3. Гетикум, подељен на Западногетску подјединицу без доњојурских седимената и Источногетску подјединицу са доњојурским континенталним и маринским седиментима после којих у обе подјединице следе кречњаци карбонатске платформе (типа Stramberk); 4. Инфра(суб)гетикум са релативно дубоководним седиментима лијаса и догера (догер типа "црних глинаца са *Bossitra alpina*") и средњег каловеја–титона (типа "*ammonitico rosso*"); 5. Данубијска јединица са плитководним лијасом, догером и малмом (зона Мироч–Вршка Чука), дубоководним догером и малмом (Доњомилановачка–новокоритска зона).

Кључне речи: јура, палеогеографске јединице, југоисточна Србија, западна Бугарска.

Introduction

During the Springer of 2005, a Serbian–Bulgarian team commenced bilateral research with the aim of making an effort to unify the views of Bulgarian and the Serbian geologists concerning the geology of the Jurassic on both sides of the Bulgarian/Serbian border. For the beginning, an attempt will be made to unify our opinions on the main paleogeographic units and subsequently new research on the lithostratigraphy and the correlation of Jurassic sediments from both side of the border will be performed.

During the Jurassic, from the Romanian Carpathians, the following main paleogeographic units can be prolonged in eastern Serbia: Thracian Massif Unit, Lužnica–Koniavo Unit, Getic, Infra (Sub)-Getic, Danubian. They

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are more or less accepted by the Serbian geologists, but are not used in Bulgaria. These units, which are strictly paleogeographic and differ from the present day tectonic units, have a bipartite structure, built of two ensembles of beds – lower (the Lower Jurassic up to the Lower Callovian) and upper (the Middle Callovian–Tithonian). From the differences in these two parts, the main paleogeographic units in the studied region were reconstructed.

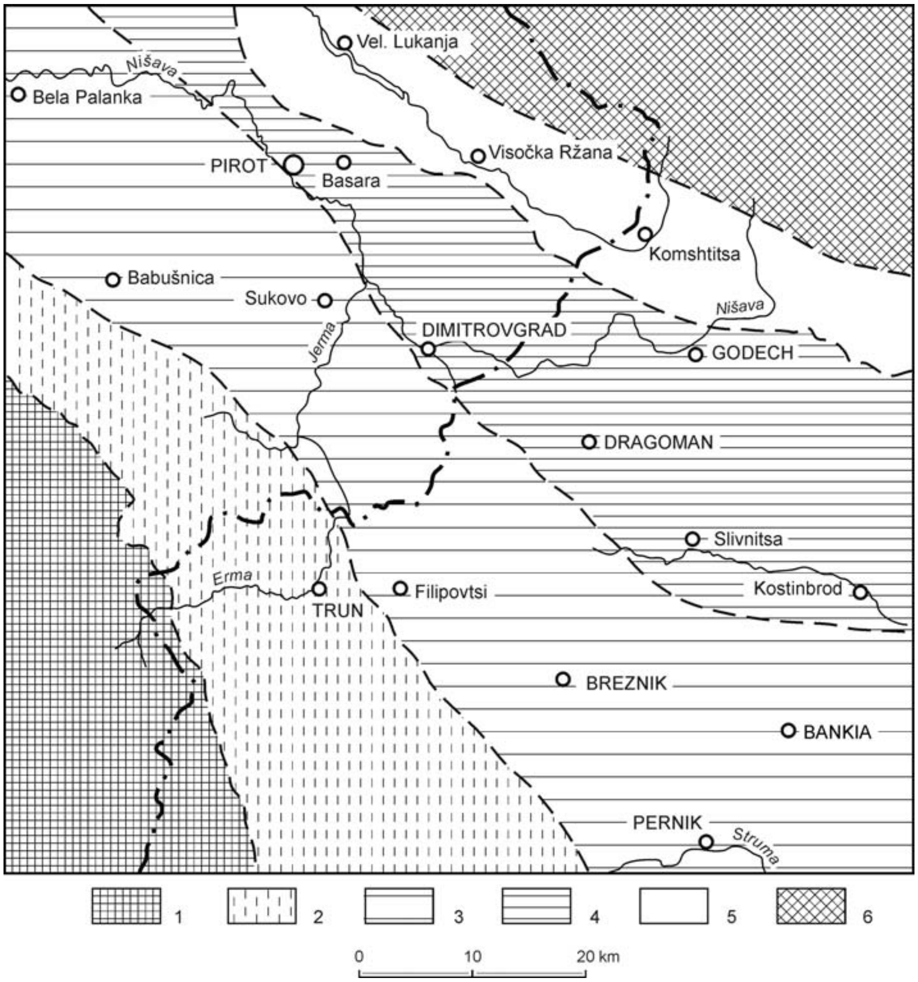

Fig. 1. Main Jurassic paleogeographic units in south-eastern Serbia and western Bulgaria. **1**, Thracian Massif Unit; **2**, Lužnica–Koniavo Unit; Getic Unit: **3**, Western Getic Sub-Unit; **4**, Eastern Getic Sub-Unit; **5**, Infra (Sub)-Getic Unit; **6**, Danubian Unit.

The Thracian Massif Unit (Fig. 1)

The name Thracian Massif Unit is used for the paleogeographic unit which unifies the Rhodope Massif, the Serbo-Macedonian Massif and the Srbsko-Makedonska Masa (Dimitrijević, 1992), etc. During the Jurassic, the Thracian Massif Unit played the role of source area and was never covered by sea water, and hence there are not marine sediments on it. For this unit (Andjelković *et al*., 1996) used the name of Moravska Zone.

The Lužnica–Koniavo Unit (Figs. 1, 2)

A characteristic for the Jurassic of this paleogeographic unit is the almost complete lack of Lower Jurassic sediments and the presence of Upper Jurassic deep water deposits. Lower Jurassic deposits exist only in the Svetlya paleograben in western Bulgaria. They are represented (Fig. 2) in the base by continental clays and sandstones (of Gresten facies), covered by Middle and Upper Liassic bioclastic and sandy limestones (of type Gresten – *sensu lato*). Apart from the territory of the Svetlya graben, during the Early Jurassic, the terrain of the Lužnica–Koniavo Unit represented dry land. During the Middle Jurassic (Dodekova *et al*., 1984), the whole area was covered by shallow marine waters, and sandstones and higher bio- and lithoclastic limestones were sedimented. Only in the western part of the territory of Bulgaria were black shales deposited. With the Middle Callovian started a relative subsidence of the terrain and the formation of "*ammonitico rosso*" type sediments, and since the Late Kimmeridgian, the deposition of flysch type alternation of argillites/marls and graded bedding sandstones – Niš-Troyan Flysch Trough (Basin) (Nachev, 1976), Lužnica Flysch, or Ruj Flysch (Dimitrijević & Dimitrijević, 1987) started. As a whole, the terrain is noted as Supra Getikum (Dimitrijević, 1992), or Supragetic units (Sandulescu & Dimitrescu, 2004). This unit is noted (Andjelković *et al*., 1996) as the tectonic structure Lužnička nappe (K-I) of the Karpatikum. In Bulgaria (Tchoumatchenko, 2002) the name Jurassic Kraishtides is used for this unit. To avoid discordance between the meaning included by the different authors, the most neutral term of Lužnica–Koniavo Unit is used here.

The Getic Unit (Figs. 1, 2)

The paleogeographic unit with the same name is well known in the Romanian East and South Carpathian (Sandulescu & Dimitrescu, 2004; etc.), as well as in Serbia (Dimitrijević, 1992; Kraütner &

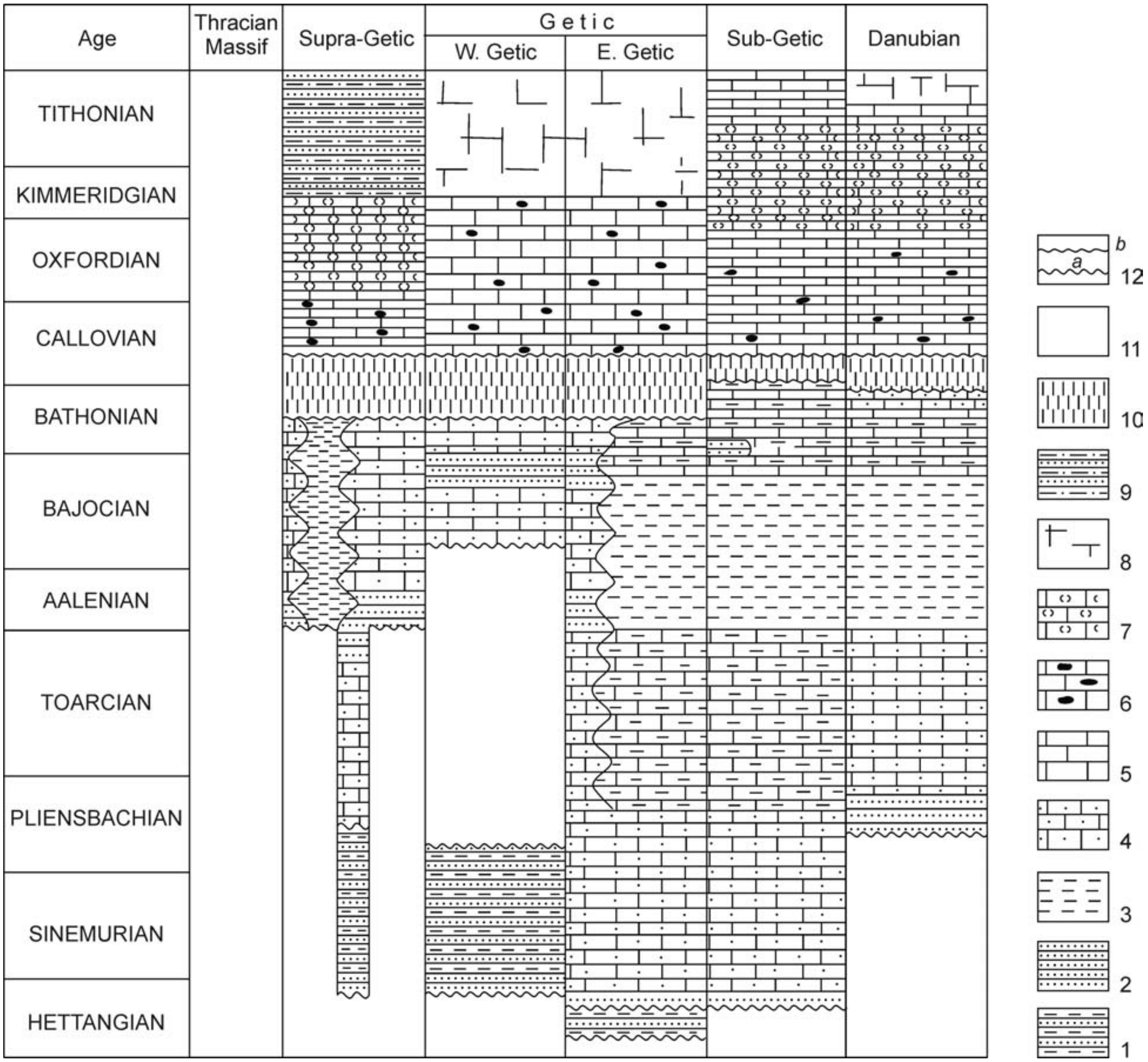

Fig. 2. Stratigraphical section across the main Jurassic paleogeographic units in south-eastern Serbia and western Bulgaria. **1**, Continental clays and sandstones (Gresten facies); **2**, sandstones; **3**, black shales; **4**, sandy and bioclastic limestones; **5**, micritic limestones; **6**, micritic limestones with chert concretions; **7**, nodular and lithoclastic limestones (type "*ammonitico rosso*"); **8**, thick bedded limestones (type "*Stramberk*")*;* **9**, flysch type alternation of sandstone and clay; **10**, interval with submarine break in the sedimentation; **11**, interval with aerial break in the sedimentation (dry land conditions); **12**, stratigraphic boundaries: a, transgressive; b, connected with submarine break in the sedimentation.

Krstić, 2003). In the Bulgarian literature, this paleogeographic unit is known as Dragoman Jurassic Horst (Sapunov *et al.*, 1985). As a Jurassic paleogegraphic unit, the Getic finished in the region south of the town of Pernik on Bulgarian territory, where it is "cut" by the Thracian Massif Unit (its Rhodope part).

More uniform for the Getic Unit is the upper part (beds), which consist of Callovian–Upper Jurassic–Lower Cretaceous thick bedded carbonate platform limestones, in many places with coral reefs. However the lower part – the Jurassic sediments lying below these thick bedded bioclastic limestones is different in different parts of the Getic Unit. This permited the division of the Getic Unit into two sub-units: the Western Getic Sub-Unit with Middle Jurassic marine sediments and the Eastern Getic Sub-Unit with Lower and Middle Jurassic in the lower part.

The Western Getic Sub-Unit (Figs. 1, 2)

The Western-Getic Sub-Unit is situated to the East of the Lužnica–Koniavo Unit. It is structured by two parts. The lower part is built up of Middle Jurassic shallow water sandstones and sandy and bioclastic limestones, which lie locally on continental Lower Jurassic clays and sandstones of Gresten facies. The upper part is built up of relatively thick bedded carbonate platform limestones, in the lower part with concretions of white chert and in the upper part by chert-free thick bedded shallow water bioclastic limestones.

The Western Getic Sub-Unit represents the Karpatikum in the territory of Serbia and enters into the following tectonic units (Andjelković *et al.*, 1996): K-II – Gornjačko-Suvoplaninska nappe, K-IV – Kučajsko-Svrljiška nappe.

The Eastern Getic Sub-Unit (Figs. 1, 2)

The characteristic for this paleogeographic unit (e.g. the section of Berende Izvor in western Bulgaria) is the presence of Lower Jurassic sediments upwards from the base: continental clay and sandstones, marine sandstones, bioclastic limestones and marls, interbedded by clayey limestones, and of Middle Jurassic black shales with *Bositra alpina* (facies well known in the Alps), followed by clayey limestones and marls, caped by a thin bed of sandy, crinoidal limestones. To West (e.g. near the town Slivnitsa), the Liassic is represented by irony red limestones, and the Dogger by sandstones and bioclastic limestones, similar to those in the Western Getic Sub-Unit. The upper parts are also similar to those of the Western Getic Sub-Unit – thick bedded limestones, in the base with concretions of white chert, and capped by chert free limestones. These sediments build the Vidlič Mountain in Serbia and are individualized as Vidlič Scale (Kraütner & Krstić, 2003), or Vidlička nappe (K-VII) (Andjelković *et al.*, 1996). In Bulgaria it is part of the Dragoman paleo-horst (Sapunov *et al.*, 1985).

The Infra (Sub)-Getic Unit (Figs. 1, 2)

The Infra (Sub)-Getic Unit, with relatively deeper water sediments is situated to the east of the Getic Unit. It is built also built up of two parts: lower and upper. The sedimentation (Fig. 2) of the lower part started during the Late Hettangian with marine sandstones, continued with bioclastic limestones, followed by marls, intercalated by clayey limestones. The Middle Jurassic is represented by black shales with *Bossitra alpina*, followed by marls and clayey limestones, similar to those in the lower part of the Eastern Getic Sub-Unit. The Middle Callovian–Tithonian sediments are represented by lithoclastic and nodular grey and red limestones, similar to the facies “*ammonitico rosso superiore*”, well-known in the Alps. In Bulgaria, the Infra-Getic paleogeographic Unit is known as Izdremets Paleograben (Sapunov *et al.*, 1985). In the Serbian literature it is known as Infra Getikum (Dimitrijević, 1992), Dobrodolsko-Grliška nappe (K-VIII) of the Karpatikum and the Staroplaninsko-Porečka Unit (Andjelković *et al.*, 1996), or Upper Danubian (Kraütner & Krstić, 2003).

The Danubian Unit (Figs. 1, 2)

The terrain of the Danubian Unit is situated to the East of the Infra (Sub)-Getic Unit. The Lower Jurassic is represented by shallow water brecco-conglomerates, sandstones, clays, clayey limestones with bivalves and sparite limestones with crinoids. The Middle Jurassic in the region of Danubian Unit is represented by two facies: (a) the Klaus facies (red nodular and ferruginous limestones of Upper Bajocian, Bathonian and Lower Callovian age – stratigrapical condensation; (b) laterally it passed to black shales with *Bositra alpina*. The Callovian–Upper Kimeridgian is developed in the facies “*ammonitico rosso*” (Donjomilanovačko–Novokoritska region in Serbia and Mihaylovgrad Paleo-Graben in Bulgaria). The Oxfordian–Berriasian in Serbia is built of deep water sediments: radiolarites and limestones with cherts.

In the Bulgarian literature (Sapunov *et al.*, 1988), this Unit is known as the Vratsa Jurassic Horst; in the Serbian literature it is individualized as Milanovačko-Novokoritska Unit (Andjelković *et al.*, 1996), the Danubikum (Dimitrijević, 1992), the Lower Danubian Units (Kraütner & Krstić, 2003), etc.

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Резиме

Упоређење седмената главних јурских палеогеографских јединица у граничној области источне Србије и западне Бугарске

На бугарским и српским геолошким картама (нпр. геолошке карте Србије и Бугарске у размери 1:100000, недавно објављене у обе земље) "нормално" се види да захватају терене само до државне границе. Са обе стране границе геолошке карте су потпуно различите и веома је тешко утврдити односе геолошких јединица. Због тога је једна мешовито српско-бугарска екипа геолога у пролеће 2005. године започела истраживања у области државне границе у циљу усаглашавања гледишта о геологији јуре са обе стране бугарско-српске границе. За почетак покушавамо да усагласимо наша мишљења о главним палеогеографским јединицама, а затим да упоредимо јурске седименте на обе стране границе.

Главне палеогеографске јединице из времена јуре које се протежу од румунских Карпата у источној Србији су: Тракијски масив, Лужница–Кониаво, Гетикум, Инфра(суб)гетикум и Данубијска јединица. Њих мање или више прихватају српски геолози, али се та подела не примењује у Бугарској. Највећа разлика је у појму "јужни Карпати". У Бугарској, према BONCHEV-у (1936, 1938) до новијих радова DABOVSKI *et al*. (2002) и NACHEV & NACHEV (2003), "јужним Карпатима" су се називали само доњокредни и горњокредни седименти северозападе Бугарске на Крајинским височима западно од вароши Кула – зона простирања синајског флиша. За друге седименте који прелазе српско-бугарску границу, а који се настављају на јурске јединице румунских Карпата, користе се бугарски термини само до државне границе. Настојаћемо да пратимо палеогеографске јединице које су откривене са обе стране границе и назваћемо их по предности њиховог обележавања.

Те јединице су строго палеогеографске, али палеогеографија је предодређена јурском тектоником у овом делу Балканског полуострва. Ове палеогеографске јединице имају двоструку структуру, односно изграђене су од две групе слојева – доња (доња јура до доњег каловеја) и горња (средњи каловеј–титон). На основу разлика ова два дела реконструисали смо главне палеогеографске јединице у проучаваној области. Шта је карактеристично за јуру ових јединица? Јединица Тракијског масива је без јурских седимената; јединица Лужница–Кониаво делимично садржи лијас у грестенској фацији и дубоководне средњокеловејске––кимериџске (*p. p.*) седименте типа "*ammonitico rosso*" и горњокимериџско–титонски силикокластични флиш; Гетска јединица је подељена на Западногетску подјединицу са доњојурским седиментима и Источногетску подјединицу са доњојурским континенталним и морским седиментима, праћена у обе подјединице кречњацима карбонатне платформе (типа Страмберк); Инфра(суб)гетска јединица садржи релативно дубоководне лијаске и догерске седименте (догер типа "црних глинаца са *Bossitra alpina*") и средњокеловејско–титонске типа "*ammonitico rosso*"; Данубијска јединица садржи плитководни лијас, догер и малм (зона Мироч–Вршка Чука) и дубоководни догер и малм у Доњомилановачко–новокоритској зони.

ГЕОЛОШКИ АНАЛИ БАЛКАНСКОГА ПОЛУОСТРВА ANNALES GÉOLOGIQUES DE LA PÉNINSULE BALKANIQUE	67	19–33	БЕОГРАД, децембар 2006 BELGRADE, December 2006

Tran-sborder (south-east Serbia/west Bulgaria) correlations of the Jurassic sediments: Infra-Getic Unit

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Abstract. The Infra-Getic Unit is a palaeogeographic unit, predestined by palaeotectonics. From the point of view of geological heritage, it represents a geosites framework. For the purpose of the correlation, the Serbian sections of Lukanja, Bogorodica Monastery, Rosomač and Senokos, as well as the Bulgarian sections of Komshtitsa, Gintsi, and Stanyantsi were used. The Jurassic sediments of the Infra-Getic Unit crop out on the southern slops of the Stara Planina Mountain in east Serbia and west Bulgaria. The Lower Jurassic started with continental and continental-marine sediments (clays and sandstones) (Lukanja clastics and Lukanja coal beds in Serbia and the Tuden Formation in Bulgaria) and continue with Lukanja quartz sandstones (Serbia) and the Kostina Formation (Bulgaria). These sediments are covered by Lukanja brachiopod beds and Lukanja limestones (Serbia) and the Romanov Dol, Ravna and Dolni Loukovit Members of the Ozirovo Formation (Bulgaria) predominantly consist of bioclastic limestones. The sedimentations follow with Lukanja belemnites-gryphaea beds (marls and clayey limestones), which in Bulgaria correspond to the Bukorovtsi Member (also marls and clayey limestones) of the Ozirovo Formation. The Middle Jurassic sedimentation started with black shales with *Bossitra alpine*. These sediments are individualized in Serbia as Senokos aleurolites and clays and in Bulgaria they are known as the Etropole Formation. In Serbia the section continues with sandstones called Vodenički sandstones of Bajocian age, known in Bulgaria as the Dobrogled Member of the Polaten Formation. However, in Bulgaria, the age is Upper Bajocian–Lower Bathonian, and it cover the marls of the lower member (Gornobelotintsi Member) of the Bov Formation and is covered by the upper member – alternation of marls and clayey limestones – the Verenitsa Member of the Bov Formation. The Vodenički sandstones–Dobrogled Member which ended their distribution in the section of Komshtitsa, to the east (in the Gintsi section), they are not represented – build a body of sandstones, a prodelta coming from the west to the east. The Bov Formation corresponds to the Senokos ammonite beds in east Serbia. The upper boundary of the Senokos ammonite beds and of the Bov Formation is sharp. It is covered by grey limestones of the Yavorec Formation in Bulgaria and by the Kamenica limestones in eastern Serbia. They are covered by grey or red nodular/lithoclastic limestones ("*ammonitico rosso*" type) of the Gintsi Formation in Bulgaria and the Pokrovenik ammonitic (*acanthicum*) limestones in Serbia. The Jurassic section in the Infra-Getic ended with grey micritic and lithoclastic limestones, which belong to the Rosomač and Rsovci limestones in east Serbia and to the Glozhene Formation in Bulgaria.

Key words: Jurassic, Infra-Getic, correlations, lithostratigraphic units, south-eastern Serbia, western Bulgaria.

Апстракт. Инфрагетска јединица је палеотектонски условљена палеоеографска јединица а са становишта геолошког наслеђа представља подручје геолошких објеката. У циљу упоређења анализирани су профили Лукање, Манастира Богородице, Росомача и Сенокоса у Србији и профили Комштице, Гинци и Стањанци у Бугарској. Јурски седименти Инфрагетске јединице су откривени на јужним падинама Старе Планине у источној Србији и западној Бугарској. Доња јура почиње са континенталним и континентално-маринским седиментима (глинци и пешчари) (Лукањски кластити и лукањски слојеви угља у Србији

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и Туден формација у Бугарској) и наставља се Лукањским кварцним пешчарима (Србија) и Костином формацијом (Бугарска). Ови седименти су прекривени Лукањским брахиоподским слојевима и Лукањским кречњацима (Србија) и члановима Романов Дол, Равна и Долни Луковит Озировске формације (Бугарска) ицграђени претежно од биокластичних кречњака. Седиментација се наставља Лукањским белемнитско-грифејским слојевима (лапорци и глиновити пешчари) којима у Бугарској одговара Букуровачки члан (такође лапорци и глиновити пешчари) Озировске формације. Средњојурска седиментација почиње црним глинцима са *Bossitra alpina*, седиментима који су у Србији познати као Сенокоски алевролити и глине, а у Бугарској као Етрополска формација. У Србији се профил наставља Воденичким пешчарима бајеске старости, који су у Бугарској познати као Доброгледски члан Полатенске формације где су горњобајеске–доњобатске старости. Ови пешчари прекривају лапорце члана Бов формације (Горњобелотиначки члан), а преко њих лежи горњи члан исте формације (Веренички члан) изграђен од смене лапораца и глиновитих кречњака. Воденички пешчари, односно Доброгледски члан, завршава се у профилу Комштице, даље према истоку (у профилу Гинци) они нису развијени. Бовска формација одговара Сенокосним амонитским слојевима у источној Србији. Горња граница Сенокосних амонитских слојева и Бовске формације је оштра; прекривена је сивим кречњацима Јаворечке формације у Бугарској, односно Каменичким кречњацима у источној Србији. Преко њих леже сиви и црвени квргави или литокластични кречњаци (типа *ammonitico rosso*) Гинци формације у Бугарској и Покровенички акантички кречњаци у источној Србији. Јурски профил у Инфра-генетикуму завршава се сивим микритским и литокластичним кречњацима Росомача и Рсоваца у источној Србији и Гложенској формацији у Бугарској.

Кључне речи: Јура, Инфра-гетик, упоређење, литостратиграфске јединице, југоисточна Србија, западна Бугарска.

Introduction

In this paper we expose our essay to make correlations accross the Serbian/Bulgarian state border of the existing in the published literature Jurassic formal lithostratigraphic units in the framework of the Infra-Getic paleotectonic and paleogeographic unit (Fig. 1). This unit is known in the Serbian literature presumably as the Staroplaninska–Porečka units (ANDJELKOVIĆ *et al*., 1996) and as the Izdremets Jurassic paleograben (SAPUNOV *et al*., 1986, etc.).

Substratum

The substratum of the Jurassic sediments in the studied area of the Infra (Sub)-Getic consists of Triassic rocks. In the Serbian Bogorodica, Rosomač and Senokos sections, the substratum consists of redish aleurolites, marls to argillites with concretions of sphaero-siderites, inter-bedded by sandstones, from 5 up to 100 m thick. They are called the Senokos red series (ANDJELKOVIĆ, 1996, p. 78) (Pl. 1, Fig. 2). These sediments cross the state border near the village Komshtitsa and continue to the east up to the Gintsi village. They are the Bulgaria Komshtitsa Formation in the Bulgaria (TRONKOV, 1969). The problem of the age is controversial because of the lack of characteristic fossils: in Serbia two opinions exist: (1) that of ANDJELKOVIĆ *et al*. (1996, p. 78, etc.), after which the Senokos red series is with the Late Triassic age; (2) this of UROŠEVIĆ & RADULOVIĆ (1990), after which they are Rhaeto–Liassic. ANDJELKOVIĆ (1996, p. 78) considered the Senokos Formation as Upper Rhaetian because they were formed under a dry and hot climate, while Jurassic sediments were formed under humid conditions. In Bulgaria, the Komshtitsa Formation, after TRONKOV (1993, p. 170) is connected by a progressive lithologic passage with the Carnian Russinovdol Formation and for that reason it is considered as Carnian–Norian.

In the section of Velika Lukanja, the Jurassic substratum is represented by 2 m of thick red breccia limestones – the Jelovica limestones (Pl. 1, Fig. 1), of Late Raetian age (ANDJELKOVIĆ *et al*., 1996, p. 78).

In the out crops near the villages Stanyantsi, Berende Izvor, Tuden, etc. in Bulgaria, the Jurassic substratum is built up of grey Middle Triassic limestones of the Iskar Carbonate Group.

Lower Jurassic (Figs. 2, 3)

In the vicinities of the villages of Velika Lukanja (Serbia) and Stanyantsi (Bulgaria), the Lower Jurassic sedimentation started by a continental sedimentation (Fig. 2). These sediments are called, in Serbia, the Lukanja clastics and Lukanja coal beds (ANDJELKOVIĆ, 1996, p. 84–86) and, in Bulgaria, the Tuden Formation (SAPUNOV *et al*., 1990). They are covered by marine sandstones.

The Lukanja clastics, (2–120 m thick), (known also as „podinski nivo”, ANDJELKOVIĆ, 1958, pp. 13–14) lie with discordance on different Triassic rocks. The Lukanja clastics are built up of conglomerates and sandstones. The conglomerates are with quartz pebbles and cement of silica, rarely of clay. The sandstones predominantly consist of quartz and silica or clayey cement. (Pl. 1, Fig. 4).

The Lukanja coal beds (8–150 m thick) started with fine grained quartz sandstones are gradually intercalated by clay and clayey sandstones with coal beds (Pl. 1, Fig. 3).

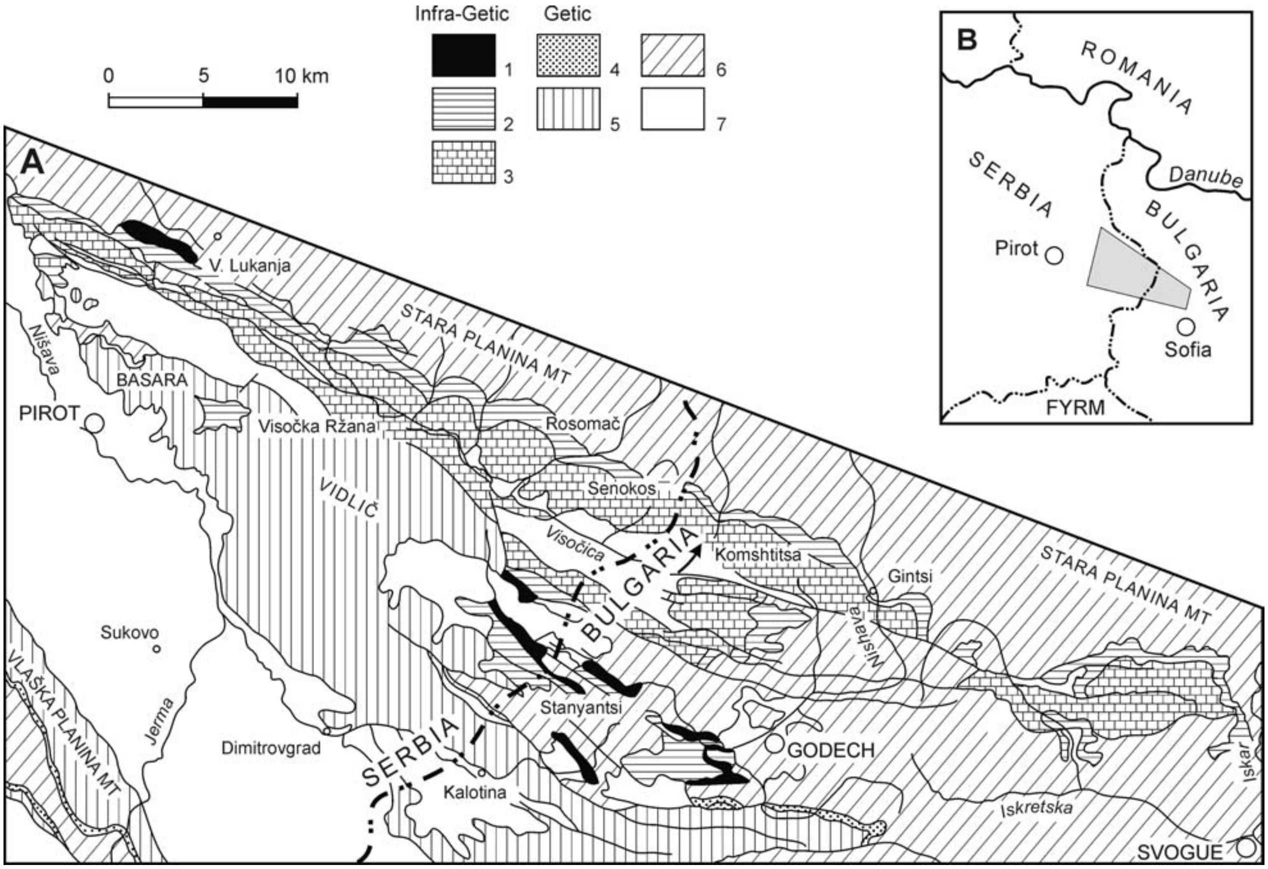

Fig. 1. Geological map of the studied region (simplified, after Kraütner & Krstić, 1993) with the location of the studied sections. **A**. Infra-Getic: 1, Lower Jurassic continental and continental/marine: sandstones and clays; 2, Lower and Middle Jurassic marine sandstones, bioclastic limestones, marls, black shales with *Bossitra alpina*, clayey limestones and marls; 3, Middle Callovian–Tithonian micritic, nodular and/or lithoclastic limestones; Getic: 4, Middle Jurassic sandstones, sandy and bioclastic limestones; 5, Middle Callovian–Tithonian reef and subreef limestones; 6, substratum; 7, cover; **B**. Location of the studied region.

The Tuden Formation (about 30 m thick) consists predominantly of clays, intercalated with sandstones. The sections started with grey to black clays and coal shales; the higher sections are intercalated with fine grained sandstones, often laminated (Pl. 3, Fig. 1).

Marine sandstones (Fig. 3) also lie on continental Lukanja coal beds and the Tuden Formation, as well as directly on Triassic sediments. In East Serbia, they are known as Lukanja quartzy sandstones (Andjelković *et al*., 1996, pp. 86–87) and in Bulgaria as the Kostina Formation (Sapunov *in* Sapunov *et al*., 1967) (Pl. 3, Fig. 2)

The Lukanja quartzy sandstones (Pl. 1, Fig. 5) (2.8 m thick in Rosomač, 8 m in Senokos and up to 120 m in the Mala Lukanja River) are built of coarse to middle

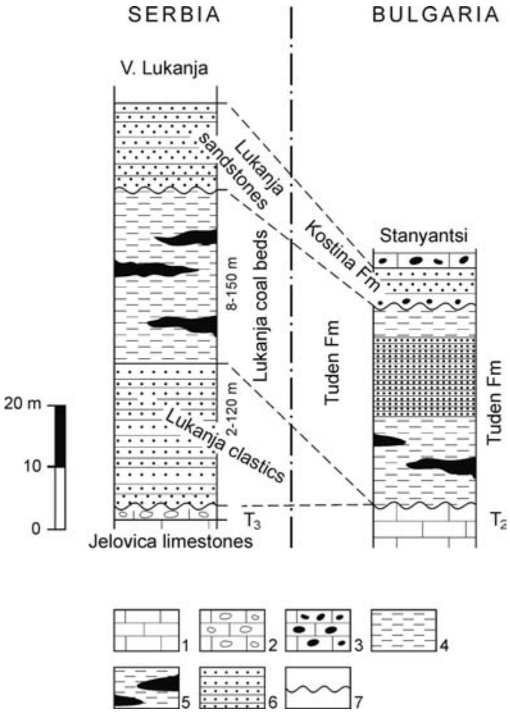

Fig. 2. Simplified columnar sections of the continental and continental-marine Lower Jurassic sediments of the Infra-Getic domain: Velika Lukanja (Pirot)–Stanyantsi (Godech) area. 1, micritic limestones (Middle Triassic); 2, red brecciated limestones = Senokos red series (Upper Rhaetian); Lower Jurassic: 3, black to grey limestones with quartz pebbles; 4, clays; 5, clays with coal beds and/or coal clays; 6, sandstones; 7, transgressive boundary.

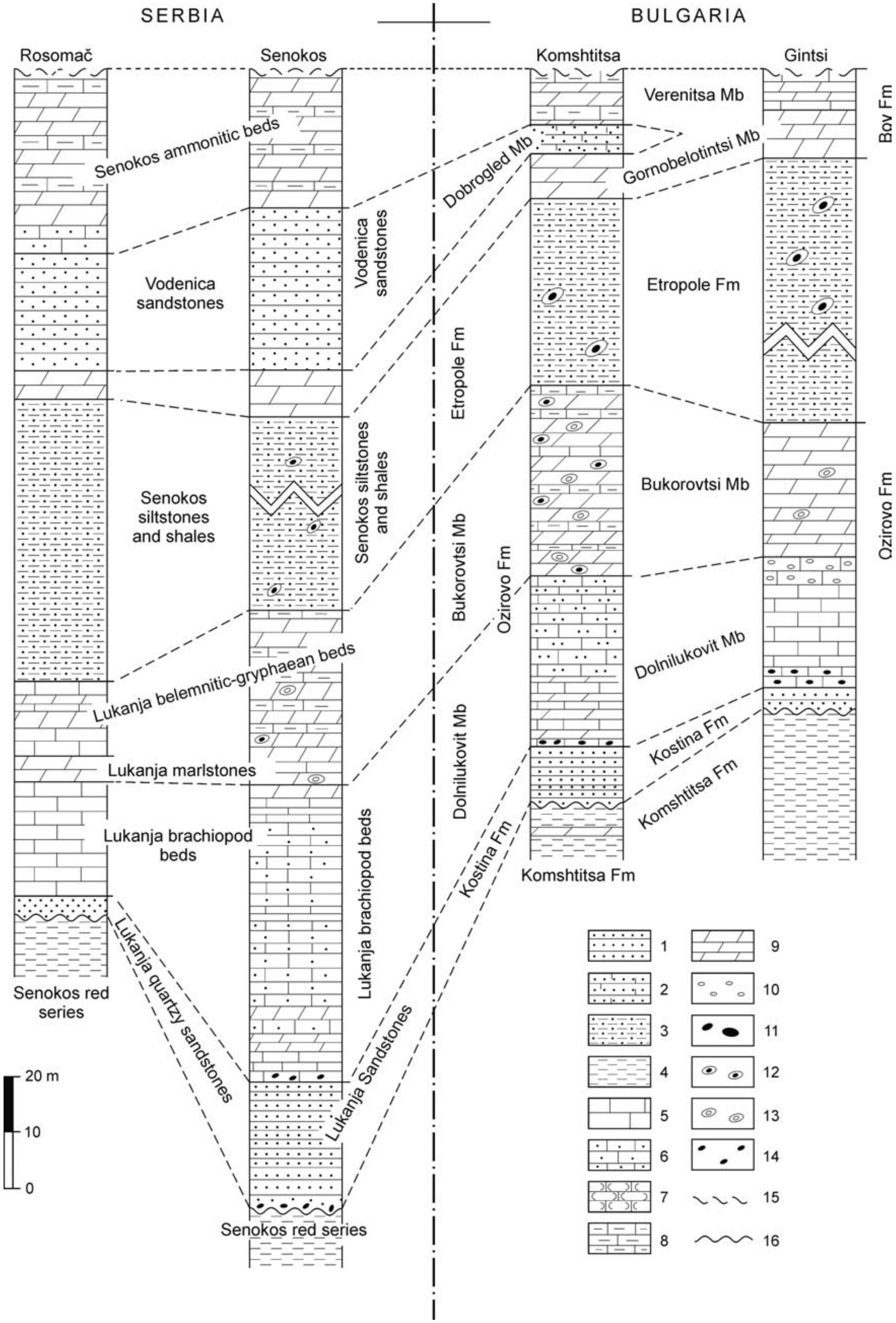

Fig. 3. Simplified columnar sections of the Lower and Middle Jurassic sediments of the Infra-Getic domain in the Pirot-Godech area. **1**, sandstones; **2**, calcareous sandstones; **3**, black shales and aleurolites with *Bossitra alpina*; **4**, clays; **5**, micritic limestones; **6**, sandy and/or bioclastic limestones; **7**, lithoclastic and/or nodular limestones (type "*ammonitico rosso*"); **8**, clayey limestones; **9**, marls; **10**, leptochloritic ooids; **11**, chert nodules; **12**, sideritic concretions; **13**, phosphoritic concretions; **14**, conglomeratic pebbles; **15**, transgressive boundary; **16**, boundary, connectected with submarine gap in the sedimentation.

grained quartz with silica cement and a transition to quartzite of Early Liassic age.

The Kostina Formation (Pl. 3, Fig. 3) lies directly on the Komshtitsa Formation in the vicinities of the villages Komshtitsa and Gintsi and on the Tuden Formation near the village of Stanyantsi. It consists of coarse to middle grained sandstones, from 3.5 m (near Stanyantsi village) up to 12 m thick (near Komshtitsa village). Its age is Middle Hettangian.

The Lower Jurassic sections continue with calcareous sedimentation. In east Serbia, the Lukanja brachiopod beds developed, which, in western Bulgaria, correspond to the Dolniloukovit Member of the Ozirovo Formation (Pl. 3, Figs. 4, 5).

The Lukanja brachiopod beds (ANDJELKOVIĆ, 1958, p. 15) near Senokos village started with dark grey to black crinoidal limestones (1.5 m thick), which contain many well-rounded quartz pebbles (Pl. 1, Fig. 6), as well as many brachiopods and bivalves. The same limestones with quartz pebbles and fossils also crop out in the Bulgarian section of Komshtitsa. In Senokos above them crop out marls (4–5 m) with rare interbeds of clayey limestones (Pl. 1, Fig. 7) (of Hettangian–Sinemurian age). Analogous sediments are individualized in Bulgaria as the Ravna Member of the Ozirovo Formation. The largest part of the Lukanja brachiopod beds is structured by sandy and bioclastic dark grey bituminous limestones, with many brachiopods, bivalves and belemnites (Pl. 1, Fig. 8). Near Senokos village, they were separated by ANDJELKOVĆ & MITROVIĆ-PETROVIĆ (1992) as the Senokos beds. The thickness of the Lukanja brachiopod beds is 40–45 m.

The Bulgarian Dolni Lukovit Member of the Ozirovo Formation (SAPUNOV, 1983) is presented by sandy and/or bioclastic (predominantly crinoidal) dark grey limestones (Pl. 3, Figs. 4, 5), containing many brachiopods, bivalves and belemnites. They are between 20–40 m thick and are of Carixian–Domerian (*p. p.*) age.

The Lower Jurassic sedimentation finished with rocks, individualized in east Serbia as the Lukanja marlstones and Lukanja belemnitic-gryphaean beds and in Bulgaria as the Bukorovtsi Member of the Ozirovo Formation.

The Lukanja marlstones (Upper Pliensbachian) consist of grey, laminated marls, clays, aleurolites and thin bedded clayey sandstones, with many small belemnites. They are covered by the Lukanja belemnitic-gryphaean beds (Upper Pliensbachian–Toarcian) (ANDJELKOVIĆ, 1958), built up of thin bedded sandy marls and clays. Within them, two parts are individualized: lower – belemnitic-brachiopod (with a predominance of small belemnites and brachiopods) and upper – belemnitic-gryphaean parts (with many large belemnites and *Gryphaea*). They contain many sideritic and phosphoritic concretions.

The Bulgarian Bukorovtsi Member is represented by grey silty marls, interbedded by thin (10–15 cm thick) beds of clayey limestones with many sideritic and phosphoritic concretions (Pl. 3, Fig. 6). They contain many belemnites and large bivalves (aequipectens and gryphaeas). They are not subdivided into different parts and encompass the Domerian and the Toarcian.

Middle Jurassic (Fig. 3)

The Middle Jurassic sediments, in south-east Serbia, are subdivided into the following lithostratigraphic units: Senokos siltstones and shales (Aalenian), Vodenica sandstones (Middle Bajocian) and Senokos ammonitic beds (Upper Bajocian, Bathonian and Lower Callovian), and in western Bulgaria into: the Etropole Formation (Aalenian–Bajocian) and the Bov Formation (Upper Bajocian–Upper Bathonian), intercalated by the Dobrogled Member of the Polaten Formation (Bathonian, lower part). The Senokos ammonitic beds and the Etropole Formation are similar to the “black shales with *Bossitra alpine*” from the Alpes.

The Senokos silstones and shales (ANDJELKOVIĆ, 1958) are structured by dark grey to black aleurolitic argillites and marly sandstones with phosphoritic, sideritic and calcareous concretions (Pl. 2, Fig. 1). Near Senokos and Rosomač villages, they are 50–70 m thick. The upper boundary with the Senokos ammonitic beds represents a transition. In Bulgaria, the Etropole Formation is analogous (SAPUNOV *in* SAPUNOV *et al*., 1967). It is built up of dark grey to black shales, generally aleurolitic with sideritic and rare phosphoritic concretions (Pl. 3, Figs. 7, 8). Near Komshtitsa village it is about 30 m thick and encompasses the Aalenian up to the lower part of the Upper Bajocian.

The Vodenica sandstones (ANDJELKOVIĆ, 1958, p. 20, 21), about 40 m thick, encompasses coarse grained quartz sandstones of red and reddish colour, thick bedded in the base, upwards becoming thin bedded (Pl. 2, Fig. 2); they also contain intercalations of microconglomerates.

The Dobrogled Member of the Polaten Formation (SAPUNOV *et al*., 1993) is about 4 m thick in the section of Komshtitsa (Bulgaria). This lithostratigraphic unit is represented by yellow to brown thick bedded limy sandstones in alternation with thin bedded calcareous limestones (Pl. 4, Fig. 1). Its age is Lower Bathonian.

Between the Senokos siltstones and shales and the Vodenica sandstones, as well as between the Etropole and the Bov Formation, the boundary is connected with a progressive transition. In the section of Komshtitsa (Bulgaria), between them grey-greenish silty marls with rare sideritic concretions, greenish marls developed, which are the horizontal prolongation of the lower member – Gornobelotintsi Member (Pl. 3, Fig. 8) of the Bov Formation. Such a lithostratigraphic unit is not individualized in the sections of south-east Serbia.

Above the Vodenica sandstones, in the vicinities of the villages Senokos and Rosomač, lies the Senokos ammonitic beds (ANDJELKOVIĆ *et al*., 1996, pp. 124–125), represented by grey-greenish sandy marls and clayey marls, rich in ammonites, which in the upper part become an alternation between grey-greenish aleuritic marls and clayey limestones with *Zoophycos,* about 30–40 m thick (Pl. 2, Figs. 3, 4). The lower boundary represents a passage effectuated by 3–4 m thick aleuritic limestones with many muscovite flakes. From the upper part, in marls and clayey limestones, ANDJELKOVIĆ *et al*. (1996, p. 128) cited *Macrocephalites macrocephalus*, *Oxycerites neumayeri* and *Hecticoceras haugi*. From the Lower Callovian (thickness 0.75 m); higher, also in marls and clayey limestones (thickness 0.50 m) were found the Middle Callovian ammonites *Hecticoceras haugi*, *Oxycerites tilli, Hecticoceras pompecki*, etc.

In Bulgaria, this unit corresponds to the Verenitsa Member of the Bov Formation (TCHOUMATCHENCO, 1978), represented by an alternation between clayey-silty limestones and thin beds of silty marls, about 8 m thick containing *Zoophycos* sp. indet of Late Bathonian age (Pl. 4, Fig. 2). From the uppermost part were collected *Rhopaloteuthis gillieroni* and *Homoeoplanulites homoemorphus*, which prove the middle part of the Upper Bathonian (*Oppelia (Oxycerites) aspidoides* Zone).

To west of Rosomača, in the valley of Jelovica River, Middle–Upper Callovian sandy limestones and sandstones crop out.

Middle Callovian–Tithonian (Fig. 4)

In the base, near the villages of Senokos and Rosomač, crop out the Kamenica limestones (ANDJELKOVIĆ *et al*., 1996, p. 133) represented in the base by brecciated limestones, which are covered by micritic, well bedded limestones, on the beds surfaces of which there are lumachelles of ammonites. These limestones are Lower and Upper (*p. p.*) Oxfordian.

To west of Rosomač village crop out sub reef sediments known as the Ržana limestones. They are represented by grey, well bedded limestones, 10 m thick, which contain many bivalves, gastropods, bryozoans, brachiopods, sponges, etc.

To the Kamenica limestones, in the vicinities of the villages of Komshtitsa, Gintsi, etc. in western Bulgaria, correspond the micritic limestones of the Late Callovian–Oxfordian–Middle Kimmeridgian (*p. p.*) Javorets Formation (Pl. 4, Figs. 3, 4) (NIKOLOV & SAPUNOV, 1970; TCHOUMATCHENCO *et al*., 2001) which consists of grey, predominantly micritic, medium to thin bedded limestones with concretions of black to dark grey chert. The hickness is about 20 m.

The Late Jurassic section continues with the Pokrovenik acanthicum limestones (Pl. 2, Figs. 5–7) in Serbia and the Gintsi Formation (Pl. 4, Figs. 5, 6) in Bulgaria.

The Pokrovenik acanthicum limestones (ANDJELKOVIĆ, 1958; ANDJELKOVIĆ *et al*., 1996, p. 139–142) (of "*ammonitico rosso*" type) consist of red, redish to grey, thin bedded limestones, which contain many lithoclasts and bioclasts (ammonites). On the basis of the ammonites, the Pokrovenik limestones are subdivided in three parts: Lower acanthicum limestones with red to reddish

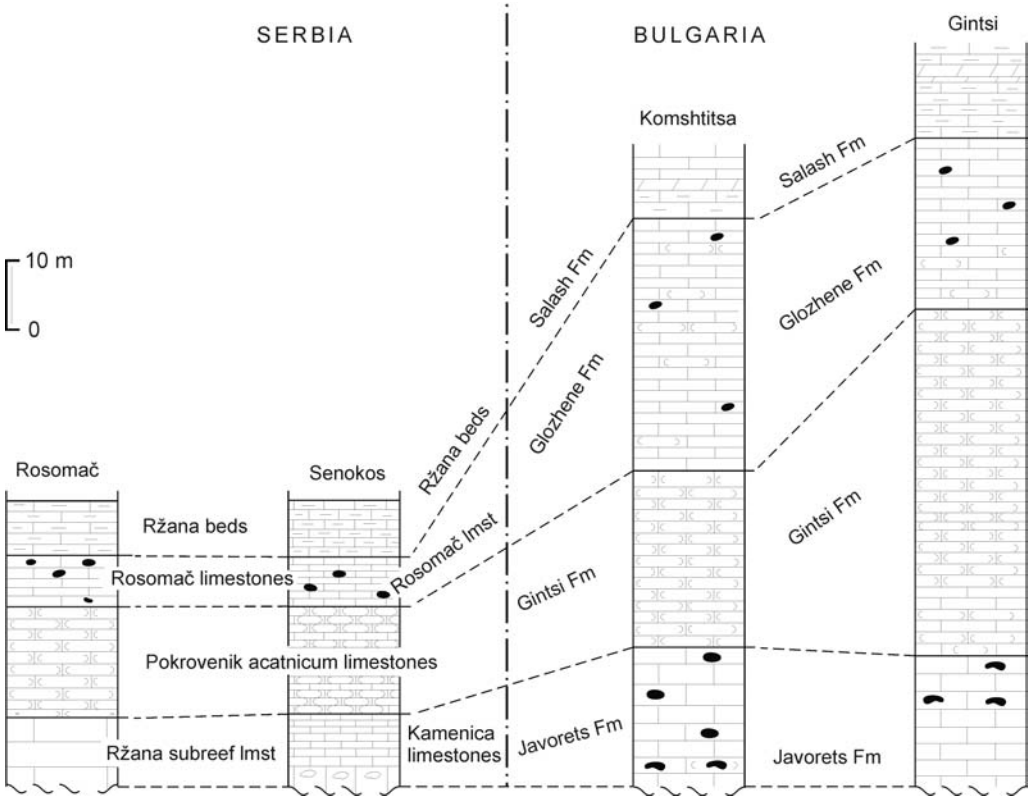

Fig. 4. Simplified columnar sections of the Middle Callovian–Upper Jurassic sediments of the Infra-Getic domain in the Pirot-Godech area. Legend as for Fig. 3.

nodular limestones, 3–5 m thick; Middle acanthicum limestones (7–20 m thick) with reddish limestones, intercalated by grey limestones, with ammonite fossils; Upper acanthicum limestones (3–10 m) with well bedded clayey, nodular and/or lithoclastic limestones, grey-redish to red, rich in ammonites.

The Pokrovenik acanthicum limestones are of Kimmeridgian–Early Tithonian age.

The Gintsi Formation in Bulgaria is analogous to the Pokrovenik limestones in Serbia. They are lithoclastic and/or nodular, red or grey limestones, with marly cement, of Middle Kimmeridgian (upper part)–Middle Tithonian (upper part) age, in the Komshtitsa section 29.15 m thick and in Gintsi section about 40 m.

The Rosomač limestones cover the reddish Lower Tithonian Pokrovenik limestones. The Rosomač limestones are represented by grey well stratificated limestones, containing dark grey to black interbeds of chert. They are Middle Tithonian–Berriasian (*p. p.*) (Pl. 2, Fig. 8). These limestones are known in Bulgaria as the Glozhene Formation (Pl. 4, Fig. 7).

The Jurassic sediments in east Serbia are covered by the Berriasian–Lower Barremian Ržana beds, consisting of slaty, grey biomicritic limestones with intercalations of chert nodules, situated in the Berriasian parts in 5 levels. Their analogous in Bulgaria are the clayey limestones of the Salash Formation (Nikolov & Tzankov, 1971) (Pl. 4, Fig. 8).

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The authors wish to thank Professor Jovan Jankičević (Belgrade University) for his useful comments and suggestions. The reviewers also deserve thanks for helpful comments which improved this manuscript. The research was supported by the Ministry of Science and Environmental Protection of the Republic of Serbia, Projects No. 146023 and 146009.

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Резиме

Упоређење јурских седимената Инфра-гетске јединице у граничној области југоисточне Србије и западне Бугарске

Инфра-гетска палеогеографска јединица условљена палеотектоником налази се између Данубијске и Гетске палеогеографске јединице. У српској литератури је позната као Инфра-гетикум, а у бугарској као јурски Издремечки палеоров. Циљ овог рада је упоређење односа јурских седимената са обе стране српско–бугарске границе. У том циљу су коришћени профили Лукање, Богородичиног манастира, села Росомач и Сенокос (околина Пирота, југоисточна Србија) и профили у атарима села Комштица, Гинци и Стањанци (Годечки срез у западној Бугарској). Јурски седименти у Инфра-гетској јединици су откривени на јужним падинама Старе Планине. Леже трансгресивно преко тријаске подине црвенкастих алевролита, лапораца и аргилита са прослојцима пешчара. Називају се Сенокосна црвена серија у Србији, а у западној Бугарској формација Комштица. Јурске наслаге почињу континенталним и континентално-маринским седиментима (глинци и пешчари) (Лукањски кластити и лукањски угљени слојеви у Србији и Туденска формација у Бугарској), настављају се лукањским кварцним пешчарима (Србија) и кварцним пешчарима Костинске формације (Бугарска). Ове седименте покривају Лукањски брахиоподски слојеви (биокластични кречњаци) и Лукањски кречњаци (Србија) и чланови Озировске формације: Романов Дол (кречњаци са кварцним шљунком), Равна (кречњаци и глинци до лапорци) и Долни Луковит (биокластични кречњаци) у којима преовлађују биокластични кречњаци (Бугарска). Седиментација се наставља Лукањским белемнитско-грифејским слојевима (лапорци и глиновити пешчари) којима у Бугарској одговара Букоровачки члан (такође од лапораца и глиновитих кречњака са квргама фосфорита и сидерита) Озировске формације. Средњојурска седиментација је почела црним алевритским глинцима са ситним шкољкама (*Bossitra alpina*) и крупним белемнитима. Ови седименти су утврђени у Србији као Сенокосни алевролити и глинци, а у Бугарској као Етрополска формација. Ова фација је у Алпима позната као "црни глинац са *Bossitra alpina*". У Србији се профил наставља Воденичким пешчарима бајеске старости, а у Бугарској Доброгледским чланом Полатенске формације. У Бугарској, они су горњобајеске–доњобатске старости, прекривају лапорце доњег члана (Горњобелотиначки члан) Бовске формације, а леже испод горњег Вереничког члана, представљених сменом лапораца и глиновитих кречњака, Бовске формације. Воденички пешчари, односно Доброгледски члан, завршавају у профилу Комштице. Бугарска Бовска формација одговара Сенокосним амонитским слојевима у источној Србији. Горња граница амонитских слојева Сенокоса и Бовске формације је оштра, местимично ерозиона (неправилна), прекривена амонитском бречом са мноштвом *Macrocephalites*–а. У Бугарској, навише следе сиви кречњаци са рожначким квргама Јаворечке формације, а у источној Србији Каменички кречњаци. Изнад њих су сиви или црвени квргави, односно литокластични кречњаци (тип "*ammonitico rosso*") Гинци формације у Бугарској и Покровенички амонитски (акантикум) кречњаци у Србији. Јурски профил Инфра-гетикума се завршава сивим микритским и литокластичним кречњацима који у источној Србији припадају Росомачким и Рсовачким кречњацима, а у Бугарској – Гложен формацији. Закључак је да се литостратиграфске јединице, које су посебно издвојили разни аутори у источној Србији и западној Бугарској, могу поредити и њихов однос доводити у везу са обе стране државне границе.

Plate 1

Serbia

Fig. 1. Jelovica limestones, Triassic, Velika Lukanja.
Fig. 2. Senokos red series, Upper Triassic, Senokos.
Fig. 3. Lukanja coal beds, Lower Jurassic, Velika Lukanja.
Fig. 4. Lukanja clastics, Lower Jurassic, Velika Lukanja.
Fig. 5. Lukanja Quartz sandstones, Lower Jurassic, Senokos.
Fig. 6. Base of the Lukanja brachiopod beds, Lower Jurassic, Senokos.
Fig. 7. Lukanja brachiopod beds, detail, Lower Jurassic, Senokos.
Fig. 8. Lukanja brachiopod beds, general view, Lower Jurassic, Senokos.


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Plate 2

Serbia

Fig. 1. Senokos siltstones and shales, Middle Jurassic, Senokos.
Fig. 2. Vodenica sandstones, Middle Jurassic, Senokos.
Fig. 3. Senokos ammonitic beds, Middle Jurassic, Senokos.
Fig. 4. Senokos ammonitic beds, Middle Jurassic, Senokos.
Fig. 5. Pokrovenik acanthicum limestones, Upper Jurassic, Rosomač.
Fig. 6. Pokrovenik acanthicum limestones, detail, Upper Jurassic, Rosomač.
Fig. 7. Pokrovenik acanthicum limestones, Upper Jurassic, Rosomač.
Fig. 8. Rosomač limestones, Upper Jurassic, Rosomač.


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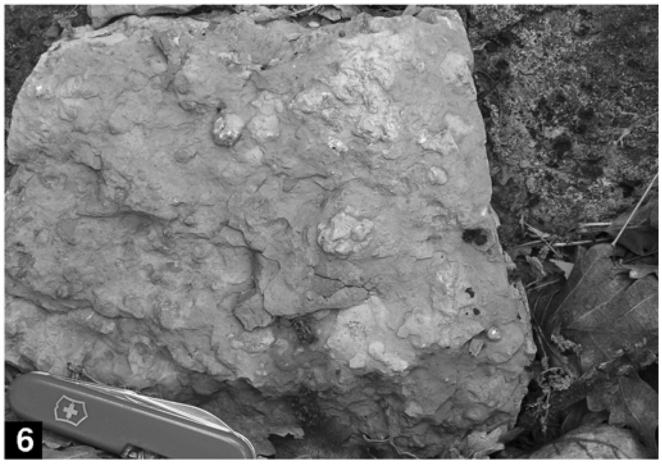
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PLATE 3

Bulgaria

Fig. 1. Touden Formation (Lower Jurassic) and the boundary with the Middle Triassic limestones, Stanyantsi.
Fig. 2. Boundary between the Komshtitsa Formation (Upper Triassic) and the Kostina Formation (Lower Jurassic), Komshtitsa.
Fig. 3. Kostina Formation, Lower Jurassic, Visochica River, Komshtitsa.
Fig. 4. Dolni Loukovit Member, Ozirovo Formation, Lower Jurassic, Komshtitsa.
Fig. 5. Dolni Loukovit Member (detail), Ozirovo Formatin, Lower Jurassic, Komshtitsa.
Fig. 6. Bukorovtsi Member, Ozirovo Formation, Lower Jurassic, Komshtitsa.
Fig. 7. Etropole Formation, Middle Jurassic, Barlya.
Fig. 8. The boundary between the Etropole Formation and the Bov Formation (Gornobelotintsi Member), Middle Jurassic, Barlya.


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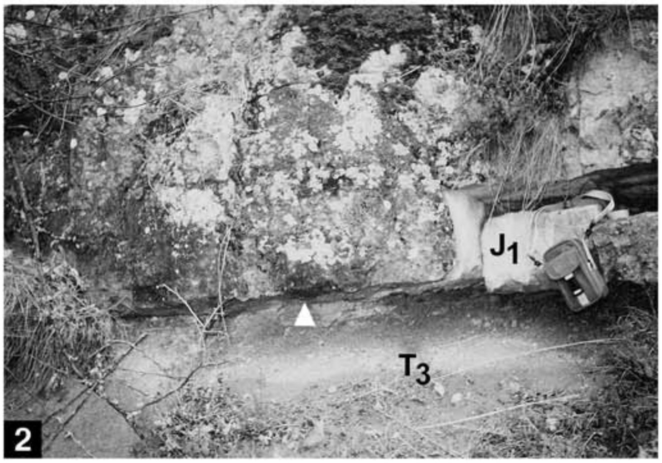
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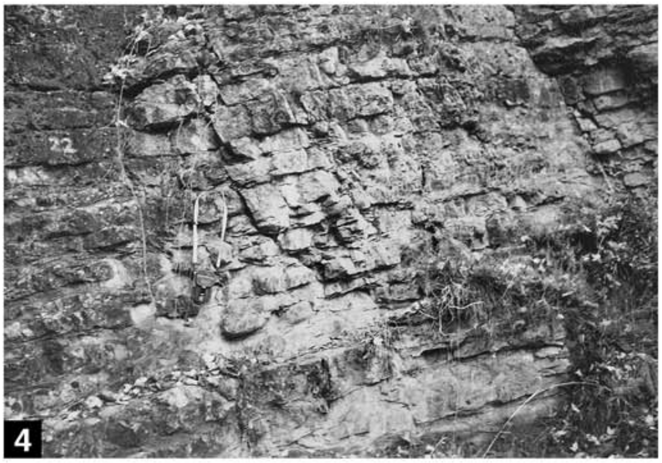
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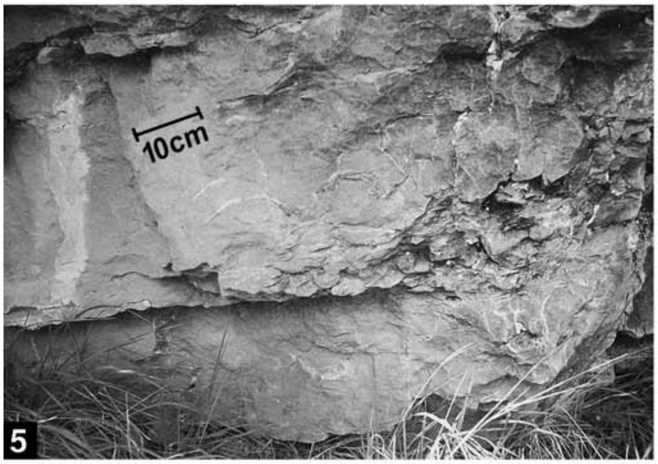
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PLATE 4

Bulgaria

Fig. 1. Dobrogled Member, Polaten Formation, Lower Bathonian, Barlya.
Fig. 2. Verenitsa Member, Bov Formation, Middle Jurassic, Barlya.
Fig. 3. Yavorets Formation, Callovian–Oxfordian, Barlya.
Fig. 4. Yavorets Formation, Callovian–Oxfordian, Barlya.
Fig. 5. Gintsi Formation, Upper Jurassic, Barlya.
Fig. 6. Gintsi Formation (detail), Upper Jurassic, Barlya.
Fig. 7. Glozhene Formation, Upper Jurassic, Barlya.
Fig. 8. Salash Formation, Lower Cretaceous, Barlya.


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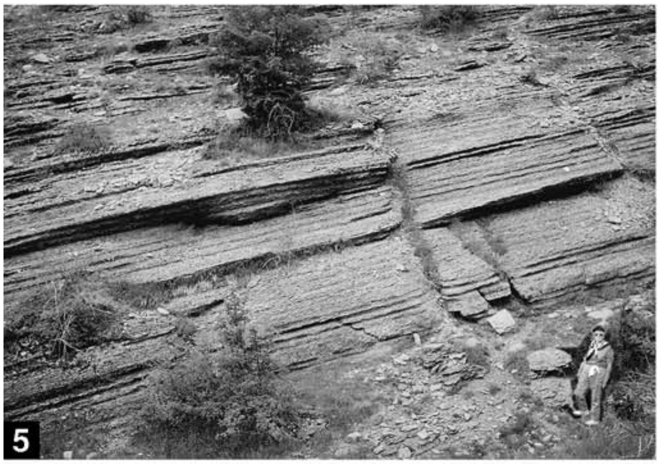
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Геолошки анали Балканскога полуострва Annales Géologiques de la Péninsule Balkanique	67	35–40	Београд, децембар 2006 Belgrade, December 2006

A storm event during the Maastrichtian in the Cauvery basin, south India

Mu. Ramkumar

Abstract. Sedimentary structures in the Kallankurichchi Formation of the Ariyalur Group, South India have been examined with a view of assessing the depositional setting of these rocks. Of the different sedimentary structures, such as cross bedding, cut and fill, etc., hummocky cross stratification is significant as it resulted from a major storm event. This paper deals with the recognized sedimentary structures, their genesis and environmental implications.

Key words: storm event, Maastrichtian, Kallankurichchi Formation, Ariyalur Group, South India.

Апстракт. Седиментне структуре формације Каланкуричи, Аријалур група, јужна Индија, проучаване су ради утврђивања депозиционог простора тих стена. Међу различитим седиментним структурама, као што су укрштена слојевитост, структура спирања итд., брежуљкаста коса слојевитост је значајна као последица деловања снажне олује. У овом раду се говори о утврђеним седиментним структурама, њиховом пореклу и утицајима на депозициону средину.

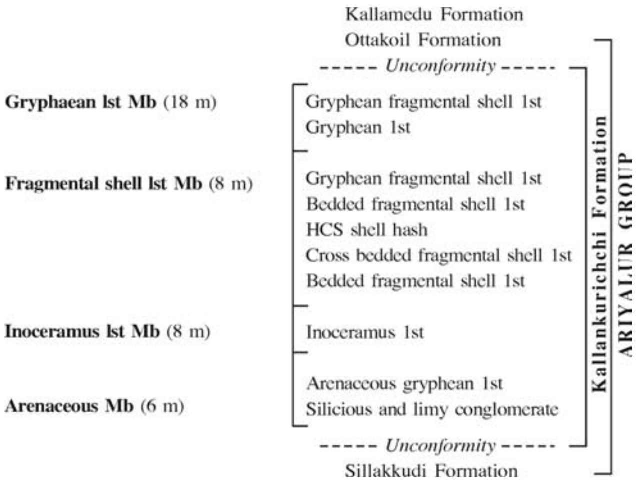
Кључне речи: утицај олуја, мастрихт, Каланкуричи формација, Аријалар група, јужна Индија.

Introduction

Sedimentary structures play a vital role in the interpretation of depositional conditions and hence an attempt was made to understand the depositional environment of the Kallankurichchi Formation of the Ariyalur Group, Tamil Nadu based on its sedimentary structures. The study area is situated east of the town Ariyalur and forms a part of the Kallankurichchi Formation (Fig. 1). The general stratigraphic setup is as follows (after Sastry *et al*., 1968; Chandrasekaran & Ramkumar, 1995).

	Age	Formation	Gross Lithology
Ariyalur Group	Maastrichtian	Kallamedu Formation	Sandstone
		Ottakoil Formation	Sandstone
		Kallankurichchi Formation	Limestone
		----- *Unconformity* ----	
	Campanian	Sillakkudi Formation	Sandstone
		----- *Unconformity* ----	
		Trichinopoly Group	

In the study area, the Kallankurichchi Formation is a prominent carbonate unit and is exposed as isolated outcrops (Guha & Senthilnathan, 1990). The formation is 40 m thick and has N–S extension of 35 kilometers with a width varying from 500–3500 m. Based on the faunal composition, Maastrichtian age has been assigned by Sastry *et al*. (1972) and later refined to Lower Maastrichtian by Ramamoorthy (1991) & Radulović and Ramamoorthy (1992). Hart *et al*. (2000) speculated the commencement of the deposition of this formation during the late Campanian–Earliest Maastrichtian. The generalized lithological succession of this formation was provided by Ramkumar (1999) and is presented herein.



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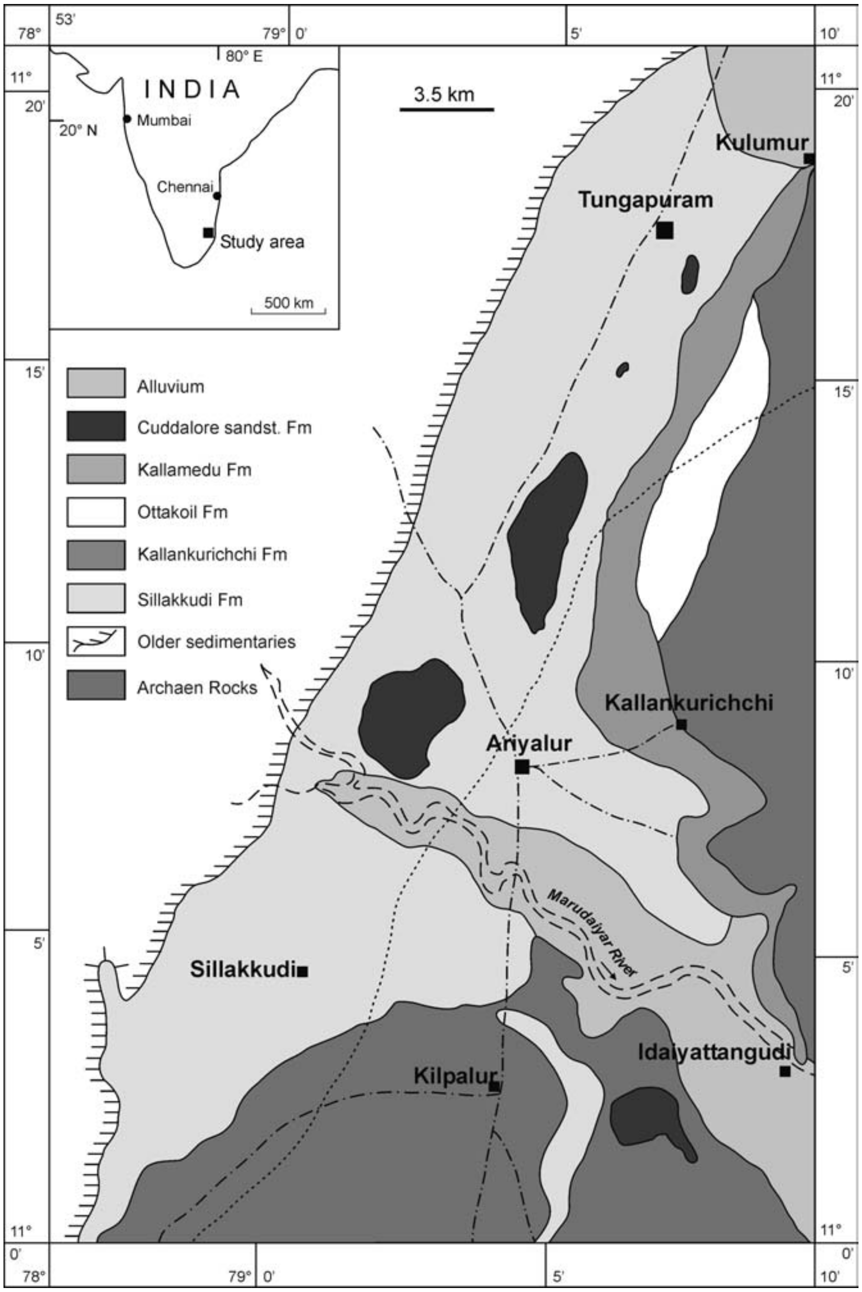

Fig. 1. Location map of the study area

The rocks of this formation consist predominantly of skeletal limestones and fragmental limestones analogous to the bank and bank-derived materials of NELSON *et al*. (1962). They contain whole shells and bioclasts of mollusca, bryozoa, foraminifera, brachiopoda, echinodermata, ostracoda and algae. Minor to significant amounts of peloid, quartz, lithoclasts and intraclasts are also observed. The six standard types of microfacies of WILSON (1975) are recognized from this formation (RAMKUMAR, 1995) and interpreted to have been deposited in a distally steepened carbonate ramp setting (RAMKUMAR, 1999). The depositional history of this formation was elucidated by RAMKUMAR (1995, 1999) and a brief note of it is presented herein.

The Kallankurichchi Formation commenced with a transgression during the Latest Campanian–Early Maastrichtian (HART *et al*., 2000). Towards the top, the conglomeratic deposits show a reduction in proportion and size of siliciclastics which were increasingly replaced by gryphean colonies. In due course, the gryphean bank shifted towards shallower regions and the locations previously occupied by coastal conglomerate become middle shelf, wherein typical inoceramus limestone started developing. The break in the sedimentation of inoceramus limestone was associated with a regression of the sea level, resulting in the erosion of shell banks and middle shelf deposits and their redeposition into biostromal deposits. Again the sea level rose to create a marine flooding surface, as a result of which gryphaean shell banks started developing more widely than before. Towards the top, shell fragments and minor amounts of siliciclastics are observed, indicating the onset of regression and higher energy conditions. The occurrence of a non-depositional surface at the top of this formation and the deposition of shallow marine siliciclastics (Ottakoil Formation) immediately over the carbonates and a conformable offlap of much younger fluvial sand deposits (Kallamedu Formation) are all suggestive of a gradual regression associated with the establishment of a fluvial system during the end of the Cretaceous.

Sedimentary structures

Cross Bedding

Tabular cross bedding is a common in fragmental limestone. The maximum thickness of the cross bedded unit is of the order of 1.8 meters. Due to the presence of shell fragments, the foreset beds do not exhibit well defined layers in the vertical section. However, they do appear as uniform layers on the surface (Fig. 2A). The cross bedding structure is found in a limited region within finely fragmented limestone and can be seen in the southwest wall of mine pit I of the Tancem mine (located west of Kallankurichchi Village and north of the Ariyalur–Kallankurichchi road – Fig. 1). This cross bedded unit can be termed as large scale cross bedding (REINECK & SINGH, 1986). The individual laminae have a more or less uniform thickness varying from 1.5–2.3 cm. The bounding surfaces of the foresets are sharp. The individual laminae can be traced throughout the length. The grains are well sorted irrespective of the nature of the clasts. High roundness is observed in both the bioclasts and peloids. Since this unit is bounded by bank deposits, the cross bedded deposits can be described as large underwater sand dunes developed in the shelf region which might have originated by shoaling waves (Chakraborty, personal communication). Like the carbonate sequence of the Middle Eocene of Peninsular Florida, described by RANDAZZO *et al*. (1990), this cross bedded unit also has abundant burrows.

Cut-and-Fill Structure

Cut-and-fill structures characterized by a shallow concave base and a flat top are common. These are observed in the Tancem mine I along the SW wall of bench I (located west of Kallankurichchi Village and north of the Ariyalur–Kallankurichchi road – Fig. 1). These have a maximum length of 150 cm and a height of 20 cm. These structures occur above the cross bedded strata and form the base of the hummocky cross bedding. The channel fill material does not show any cross bedding and the channels are filled up with fining upwards coarse grained carbonate sand. This carbonate sand consists of minor amounts of intraclasts, ferruginous matrix and fine quartz sand. These channel-fill structures gradually merge into hummocky cross stratification (HCS).

Hummocky Cross Stratification

Hummocky cross bedding is found near the location where cut-and-fill structures predominate. Its characteristics are described herein.

a. The laminae are curved both in hummocks (convex up) and swales (concave up) sectors.
b. The laminations dip at 12°; but the bed sets appear to meet at very low angles in such a way that, at times, they are parallel to the lower bounding surface.
c. Individual laminae have a maximum thickness of 4 cm at swales and 1.8 cm at hummocks, reflective of a thickening (at swales) and a thinning (at hummocks) nature. Maximum wave height is 97 cm and wave length 6 meters.
d. The laminations show no preferred orientation.
e. The rocks showing HCS structures are composed of polished fragmental shells (Fig. 2B, C). These are sandwiched between normal bedded and cross bedded carbonate sand. The upper contact of the hummocky cross stratification unit is also sharp.

Hummocky cross stratification is commonly associated with storm deposits ("Tempstites" of AGER, 1973; KREISA & BAMBACH, 1982; LOOPE & WATKINS, 1989; MENG *et al*., 1997). It is observed on the continental shelf of the northwest Atlantic Ocean in water depths of 10–40 meters. It is also found in tidal flats (MUKHERJEE *et al.,* 1987; WEIDONG *et al.,* 1997). It has been reported from clastic sediments (Bose *et al.,* 1997), as well as from carbonate skeletal deposits (MENG *et al*., 1997; WEIDONG *et al*., 1997). The HCS is interpreted as being due to a combination of storm generated and geostrophic currents (SWIFT *et al*., 1983).

In the present area, reworked autochtonous fauna in the HCS with little lateral variation of texture and structures are found to occur. This suggests that this particular unit of the Kallankurichchi Formation did not receive material from distant sources during the storm. The absence of whole unabraded, well marked layering, edge-polished shell fragments of fossils (Fig. 2B, C), in addition to the occurrence of storm deposits as a single thick unit, etc., suggest that the prevalent major storm might have mobilized already deposited sediments on the bottoms (MENG *et al*., 1997; KROH & NEBELSICK, 2003). According to the descriptions of AIGNER (1982) and AIGNER & REINECK (1982), the exposures of HCS at Tancem mine I SW wall bench I represent a proximal storm bed in view of the following characteristics.

a. This storm depositional unit is a very thick bed.
b. The beds are composite and intermixed with various bedforms and materials.
c. It is composed of bioclasts which are coarse grained (Gravel to coarse sand).

These characteristics of this sequence are spread over short distances and die out towards the east where the size of the bioclasts decreases. Further east, thinly bedded, mud dominated rocks with unabraded fossils are observed, which may represent the distal end of storm beds (TUCKER & WRIGHT, 1990). BOUOUGRI & PORADA (2002) and MENG *et al*. (1997) also observed the deposition of mud and finer grain rich deposits after major storm event in the Neoproterozoic deposits of Morocco. As has been observed in storm associated deposits elsewhere (MENG *et al*., 1997; SAVRDA & NANSON, 2004), due to the reduction of intensity of the storm, the finer grade materials also started to settle and hence, this sequence shows fining upward gradation from gravel to sand. The grains were carried and settled from a suspension cloud. These interpretations are supported by the horizontality of platy shell material with reference to the original sedimentation surfaces. Comparison of these characteristics based on the criteria enlisted by GOFF *et al*. (2004) clearly affirms the storm generated nature to these deposits. This type of typical storm deposit and its distal expression (MARTINI & BANKS, 1989) are interpreted to be of inner and

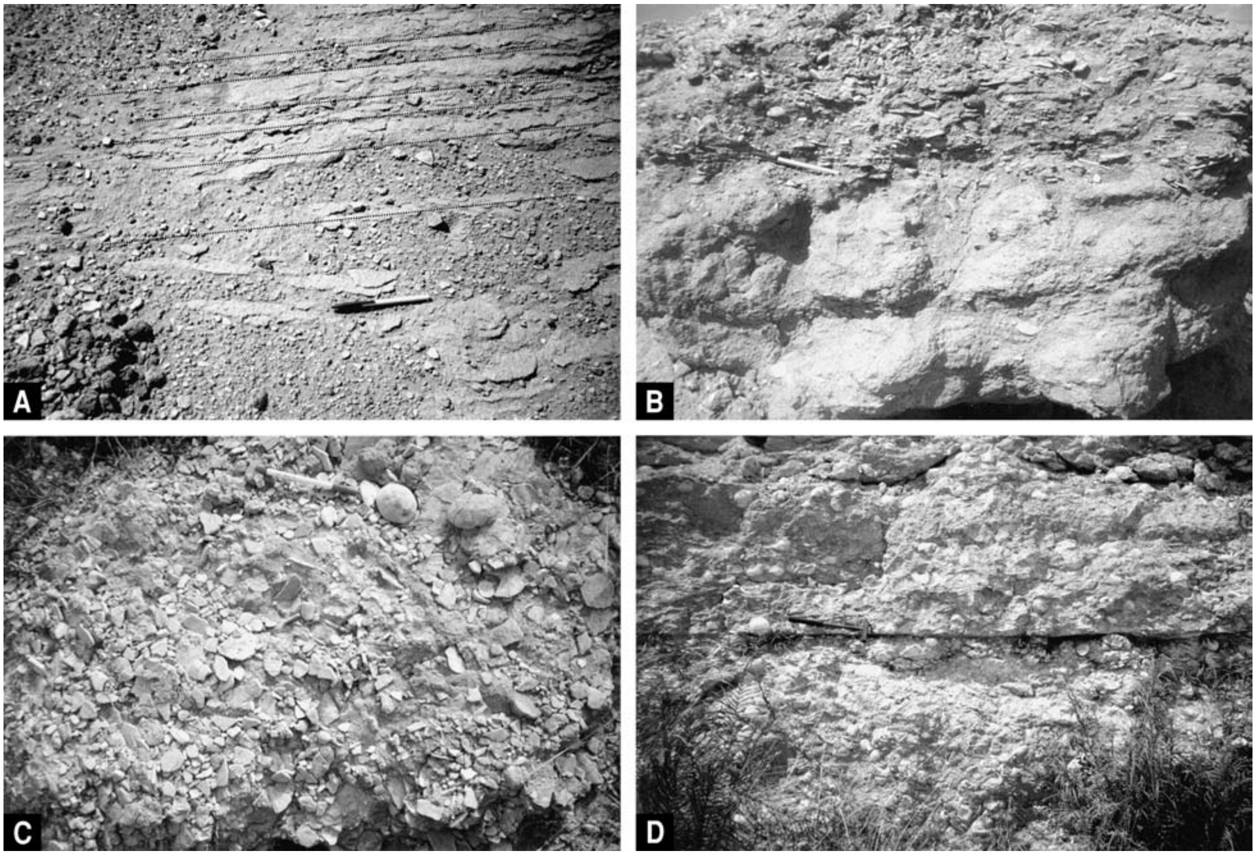

Fig. 2. **A**, Field photograph of planar cross bedded limestone exposed in a mine floor. Due to low dip and planar bedding, the exposure at the mine floor depicts only feeble bedding planes, indicated by dashed lines; **B**, Field photograph showing the nature of the HCS sequence in a mine section. The shoal facies is superposed by HCS beds. Note the sudden change from shoal facies limestone (lower) to edge polished fragmental limestone (HCS bed) as reflected by the sudden change in skeletal composition. Differential compaction of the shoal facies limestone and the HCS limestone has obliterated the sharp erosional bedding plane (indicated by pen in the photograph) between these two units; **C**, Close-up view of the HCS bed showing fragmented and rounded shell material. Also note the presence of Stigmatophygus in life position (indicated by pen in the photograph); **D**, Typical gryphean limestone deposit of the Kallankurichchi Formation. Occurrence of these gryphean colonies over the HCS unit denotes the return of normal depositional conditions after a major storm event.

middle shelf in origin respectively (Droser & Bottjer, 1988; Burchell *et al*., 1990).

Since the storm bed with HCS is found to occur in between normal bedded and cross bedded deposits, the energy of the shoaling waves is presumed to have been short lived. The gradual change of the storm beds to cross bedded, well sorted carbonate sands is indicative of the waning period of the storm. The escape structure in a ‘V‘-shaped burrow at the base of the storm deposit (Heckel, 1972) suggests the sudden appearance of storms. Oyster beds above the bedded and cross bedded carbonate deposits suggest that the colonization of oysters (Fig. 2D) started after the major storm event.

Conclusion

From the nature and sequence of the sedimentary structures, particularly the hummocky cross beds, it can be concluded that during the deposition of the Kallankurichchi Formation, there were storm events, which contributed to the continuous and homogenous deposition of bank deposits and middle shelf deposits. The intensity of the storm event was very high in the deposition of 1.8 meters thick fragmental shell beds. From the change in the nature of sediments within the storm bed, an easterly storm condition has been inferred. The storm deposits of the Kallankurichchi Formation show similarities in broader terms with that of the Cretaceous Mzamba Formation of South Africa, as reported by Liu & Greyling (1996).

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Резиме

Олујна појава за време мастрихта у басену Каувери, јужна Индија

Кречњачке наслаге формације Каланкуричи, Аријалур групе у јужној Индији, садрже карактеристичну асоцијацију седиментних структура ограничених на одређену област у близини села Каланкуричи. Како су седиментне структуре битне за тумачење палеосредине, учињен је покушај да се упознају главни услови таложења, а тиме се дошло и до доказа појава снажне олује током мастрихта у басену Каувери у јужној Индији.

Формација Каланкуручи се састоји претежно од скелетних кречњака и фрагментарних кречњака што одговара обалским и пореклом обалским материјалима. Запажене су мање до знатне количине пелоида, кварца, литокласта и интракласта. Утврђено је шест стандардних типова микрофација ове формације и њихово таложење у средини доњег краја стрме карбонатне рампе.

Табуларна коса слојевитост је честа код слојева поломљених кречњака формације Каланкуручи. Максималне дебљина косо услојених кречњака је реда 1,8 метара. Због присуства одломака шкољака, коси слојеви не показују јасно дефинисане слојеве у вертикалном профилу. Међутим, на површини слојеви изгледају уједначено. Ова косо услојена јединица се може узети да представља косу услојеност већих размера. Како је ова јединица ограничена обалским наслагама, косо услојене наслаге се могу описати као подводне велике пешчане дине највероватније формиране таласима у области оплићалог шелфа.

Структуре спирања (накнадно запуњене вододерине) су честе у проучаваној области и карактеришу се плитком конкавном основом и заравњеним врхом. Те структуре се јављају изнад укрштених слојева и формирају основу брежуљкасте слојевитости. Материјал и испуне вододерина не показују никакву косу слојевитост, већ су вододерине испуњене крупнозрним карбонатним песком који се навише уситњава. Те структуре спирања постепено прелазе у брежуљкасту косу стратификацију (БКС).

На брежуљкасту косу слојевитост се наилази у близини места где преовлађују структуре спирања. Брежуљкаста коса стратификација се обично доводи у везу са олујним наносима. БКС се објашњава да потиче од комбинованог дејства олујне и геострофне струје. Сада су у тој области нађене појаве преталожене аутохтоне фауне у БКС са малом бочном разликом у структури и текстури. Та структура указује да за време олује није стигао материјал са веће удаљености нарочито у ову јединицу формације Каланкуричи. Одсуство читавих, изражено слојевитих, фрагмената фосила шкољки углачаних ивица поред појаве олујних наслага у виду једне дебеле јединице и друго указује да је доминантна, већа олуја можда покренула седименте већ наталожене на дну. На основу тога што је појава олујног слоја са БКС нађена између нормално услојених и косо услојених наслага, претпоставља се да је енергија таласа у оплићалој средини кратко трајала. Постепени прелаз олујних слојева у косо услојене наслаге, добро сортирани карбонатни пескови, указују на период слабљења олује. Структура испирања у бразди у облику слова “V” у бази олујне наслаге указује на наглу појаву олуја. Слојеви острига изнад слојевитих и косо услојених карбонатних наслага указују да је колонизација острига почела пре главне олује.

На основу природе и секвенце седиментних структура, нарочито брежуљкасто-косих слојева, може се закључити да су се за време таложења Каланкуричи формације јављале олује које су допринеле сталном и уједначеном таложењу обаских наслага и наслага средњег шелфа. Олује су биле врло великог интензитета кад су се наталожили слојеви одломака шкољки дебљине 1,8 метара. Из промене природе седимената унутар олујног слоја закључује се да је олују стварао источни ветар. Олујне наслаге формације Каланкуричи показују сличност у ширем смислу са кредним олујним наслагама формације Мзамба у Јужној Африци.

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The Cretaceous/Paleogene (K/Pg) boundary in the Mezdra and Lyutidol syncline, Vratza District (West-Fore Balkan, Bulgaria)

Nikola A. Jolkičev

Abstract. This paper discusses the unjustified assignment (based on calcareous nannofossils) of a large portion of the Maastrichtian strata in the Mezdra and Lyutidol synclines (West Fore Balkan, Bulgaria) to the Paleogene. The co-occurrence of Paleocene nannofossils, reported by some authors, and Maastrichtian macrofossil taxa in these sections indicates diachronism in the appearance of macro- and nannofossils across the K/Pg boundary. Thus, this boundary cannot be precisely localised except if the Maastrichtian fossils are assumed to have been redeposited, but there is no evidence of resedimentation. Maastrichtian macrofossils are found not only within the range of the Paleogene nannofossil zones, but also in sections overlying them in the Kajlâka Formation where new Maastrichtian macrofossil taxa, such as the echinoid *Hemipneustes striatoradiatus* (Leske), appear and some inoceramid and cephalopod taxa range into this unit. These facts shed doubt over the applicability of nannofossils in determining the K/Pg boundary where this has already been firmly documented by macrofauna.

Key words: Cretaceous/Paleogene boundary, Maastrichtian strata, nannofossils, foraminifers, inoceramids, ammonites, echinoids, West Fore Balkan, Bulgaria.

Апстракт. У раду се расправља о неоправданом приписивању (на основу кречњачких нанофосила) великог дела мастрихтског профила у синклиналама Мездре и Љутидола (западни Предбалкан, Бугарска) палеогену. Истовремено појављивање палеоценских нанофосила, о којима пишу неки аутори, са мастрихтским макрофосилним таксонима у овим профилима указује на диахронизам у појави макрофосила и нанофосила на граници K/Pg. Према томе, ова граница се не може прецизно утврдити осим ако се не претпостави да су махстритски фосили били преталожени, али не постоје докази преталоживања. Мастрихтски нанофосили су нађени не само у оквиру вертикалног простирања палеогених нанофосилних зона, већ и у формацији Кајлака где се јављају нови мастрихтски макрофосилни таксони као што је *Hemipneustes striatoradiatus* (Leske) и неки иноцерамски и цефалоподски таксони. Ове чињенице бацају сенку сумње на примењивост нанофосила у одређивању границе K/Pg где је она већ поуздано доказана на основу макрофауне.

Кључне речи: граница креда–палеоген, мастрихтски слојеви, нанофосили, фораминифери, иноцерамуси, амонити, ехиниди, западни Предбалкан, Бугарска.

Inroduction

Upper Cretaceous sediments in the Mezdra and Lyutidol synclines in the Vratza District, West Fore Balkan, Bulgaria are widely distributed and of essential tectonic importance. For a long time, based on erroneously identified fossils, these were assumed to be of Cenomanian age (Zlatarski, 1904, 1905, 1910). This author (Zlatarski, 1905) assigned only a portion of the limestones exposed at the village of Varbeshnitza, northwest of Mezdra and those around the village of Lyuta (now Vladimirovo) to the Senonian. Later, Zlatarski (1910) pointed out that the limestones at Lyuta were certainly of Senonian age but those at Varbeshnitza were of doubtful Senonian age, although he cited some Senonian fossils found earlier by him. He assumed these sediments to be of Cenomanian age but later again referred them to the Senonian (Zlatarski, 1927). The same author (Zlatarski, 1904) assigned a Cenomanian age also to Eocene sandstones resting upon the "Cenomanian" limestones which, as he pointed out, were easily distinguished from the Lower Cretaceous sandstones in the Vratza area. For the first time, Bončev (1932) proved that the Upper Cretaceous sediments in the Fore Balkan to the south of the Iskar River were of Maastrichtian, not Cenomanian, age. Simultaneously, Bončev & Kamenov

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(1932) extended the studies of this stage to the north of the Iskar River – between Mezdra and Roman, and later they (BONČEV & KAMENOV, 1934) continued these to the west – between the rivers of Iskar and Ogosta. Based on inoceramids, cephalopods, echinoids and other macrofossil taxa, they documented in detail the biostratigraphy of the Maastrichtian Stage in the western Fore Balkan. The Maastrichtian age of the Upper Cretaceous sediments in this area was confirmed by all subsequent investigators, based on macrofossil fauna (COHEN, 1946; TZANKOV, 1968; JOLKIČEV, 1982, 1986, 1989, and others).

During recent years, calcareous nannofossils have been assumed to be of extreme importance for the subdivision of Upper Cretaceous and Cenozoic sediments – an importance that, seemingly, cannot be put in question. However, NAIDIN (2002, p. 46) has recently pointed out that "nevertheless we should have some doubts" of the applicability of nannofossils.

Under the influence of nannoplankton euphoria, a number of publications have recently appeared in which the Cretaceous/Paleogene boundary in the study area was traced without taking into account the presence of characteristic macrofauna in the same sections that were subdivided by means of nannoplankton. The Cretaceous/Paleogene boundary as determined by macrofauna was disregarded in these papers.

The macrofaunal data presented below raise questions about the applicability of nannofossils in defining the Cretaceous/Paleogene boundary in the study area.

Facts and discussion

This paper discusses the Maastrichtian strata in the southern limb of the Mezdra syncline and the same deposits in the northern and southern limb of the Lyutidol syncline in the southern parts of the West Fore Balkan (Fig. 1).

The stratigraphic section in these two structures comprises the following lithostratigraphic units in ascending order (JOLKIČEV, 1986): Dârmanci Formation – Lower Maastrichtian; Kunino Formation – Lower Maastrichtian; Mezdra Formation – Lower Maastrichtian and Kajlâka Formation – Upper Maastrichtian (Fig. 2).

The studies of SINNYOVSKY (1991, 1993, 1998, 2001), SINNYOVSKY & CHRISTOVA-SINNYOVSKA (1993) and STOYKOVA *et al*. (2000) all focused on the Dârmanci, Kunino and Mezdra Formations. It is unexplainable why they did not discuss the age of the overlying Kajlâka Formation.

The Mezdra Formation in the two structures comprises three lithological units of variable thickness: the lower unit – microgranular limestones with flint concretions; the middle unit – argillaceous limestones without flint concretions with interbeds or in alternation with marls and the upper unit – microgranular limestones with flint concretions (Figs. 2–4). SINNYOVSKY & CHRISTOVA-SINNYOVSKA (1993, p. 32) referred to the middle unit in the Lyutidol syncline as the "Limestone Formation". In this unit EK. DIMITROVA (Geological Institute, Bulgarian Academy of Sciences (BAS), unpublished data) identified a foraminiferal assemblage (see Fig. 3). From the same strata at the southerly limb of the Mezdra syncline (at the village of Chelopeck), Y. MALIAKOV (Geological Institute, BAS) collected eighteen echinoid tests (now housed at the museum of the Geological Institute, BAS No F.002525 to 002542). Among these, the following taxa have recently been identified (Fig. 2): *Echinocorys conoidea* GOLDFUSS as well as several *Echinocorys* sp. which belong to a group of species morphologically close to *E*. gr. *marginata*/*subglobosa* (of early to late Campanian age; compare ERNST, 1972, 1975; JAGT *et al*., 2004); this may represent a continuation into, or recurrence(?) during the Maastrichtian of such test morphologies (compare JAGT, 2000). In the Maastrichtian type area, these forms occur as well, and are nearly always associated with typical

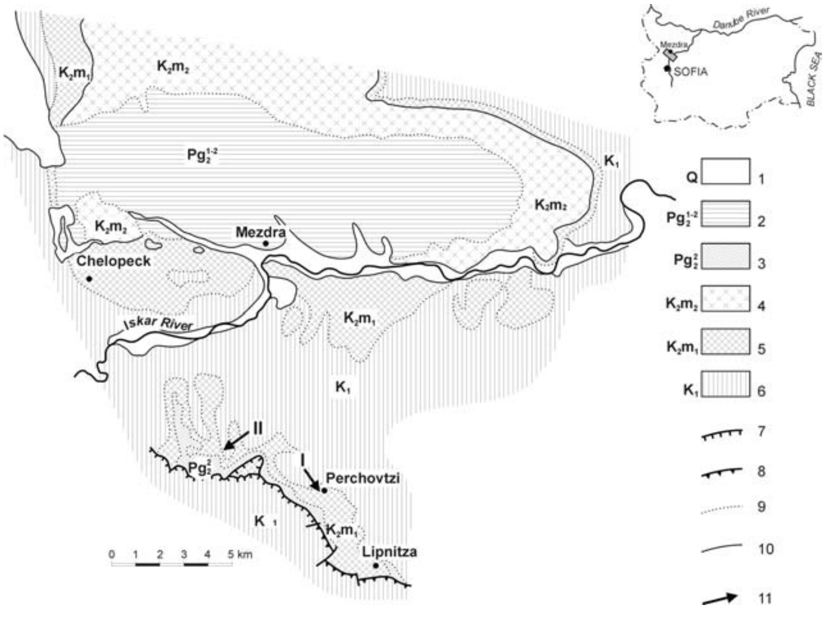

Fig. 1. Sketch map (inset) of Bulgaria with location and geological map of the study area (after TZANKOV *et al*., 1991, modified). **1**, Quaternary; **2**, Lower–Middle Eocene; **3**, Middle Eocene; **4**, Kajlâka Formation – Upper Maastrichtian; **5**, Dârmanci, Kunino, Mezdra formations – Lower Maastrichtian; **6**, Lower Cretaceous; **7**, thrust; **8**, reverse fault; **9**, transgressive boundary; **10**, boundary of Quaternary sediments; **11**, stratigraphic sections (sections I and II, shown in Figs, 3, 4, respectively).

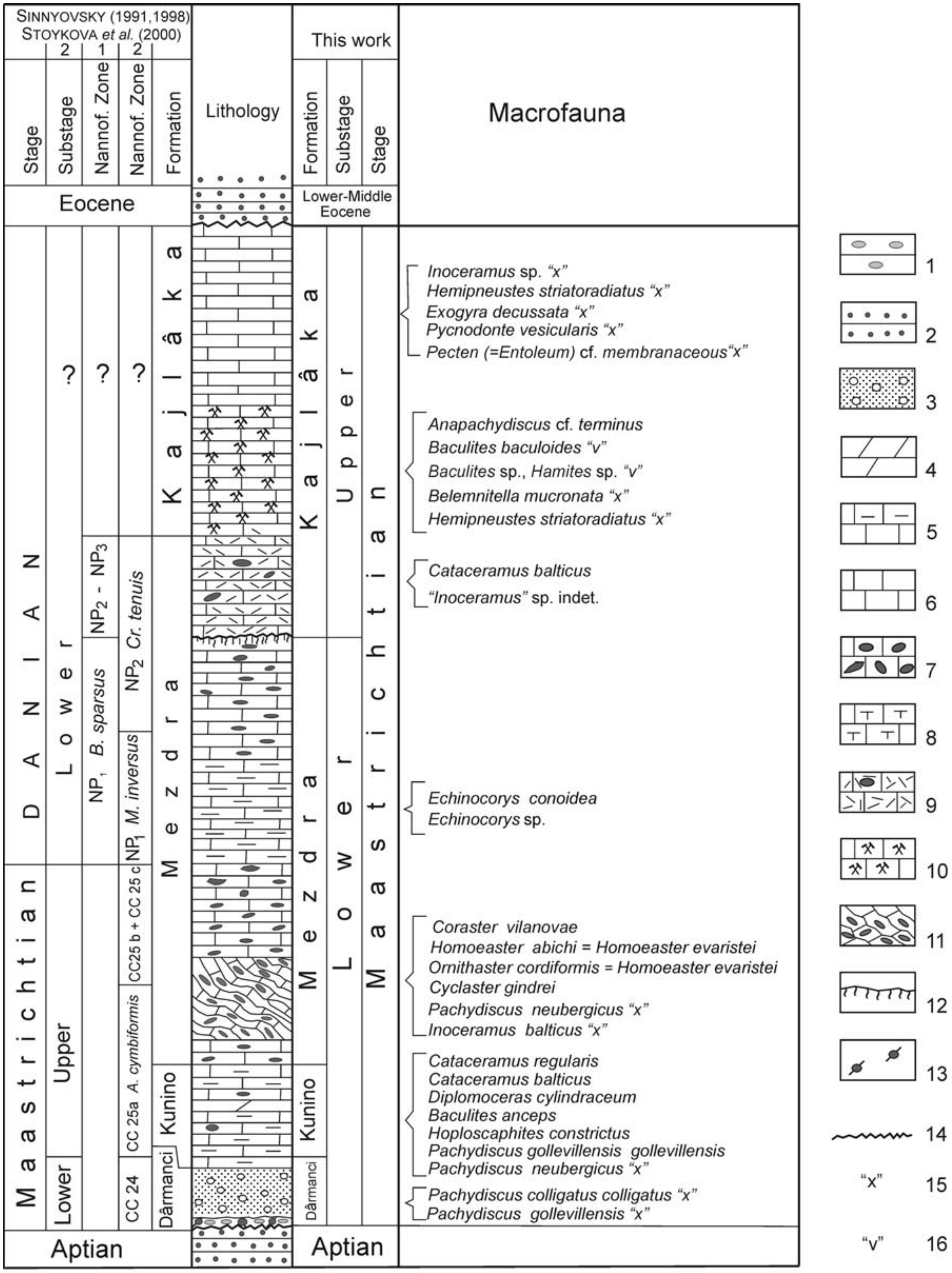

Fig. 2. Generalized stratigraphic section of the Mezdra syncline. **1**, conglomerates; **2**, sandstones; **3**, glauconitic sandstones; **4**, marls; **5**, clayey limestones; **6**, medium- to coarse-grained limestones; **7**, limestones with flints; **8**, chalky limestones; **9**, biomorphic limestones; **10**, "quarry type" limestones; **11**, megaslump; **12**, hardground; **13**, phosphorites; **14**, transgressive boundary; **15**, "x", fossils discovered by Bončev & Kamenov (1934) and probably some of them misidentified; **16**, "v", fossils discovered by Zlatarski (1910) and probably some of them misidentified. The meaning of the question marks "?" in Figs. 2, 3 and 4 is as follows: Fig. 2 – part of the section unstudied by Sinnyovsky (1998) and Stoykova *et al*. (2000); Fig. 3 – the upper part of Mezdra Fm is not shown in the paper of Sinnyovsky & Christova-Sinnyovska (1993, fig. 8); Fig. 4 – the middle part of Mezdra Fm is not shown in the section of Sinnyovsky (2001, fig. 3).

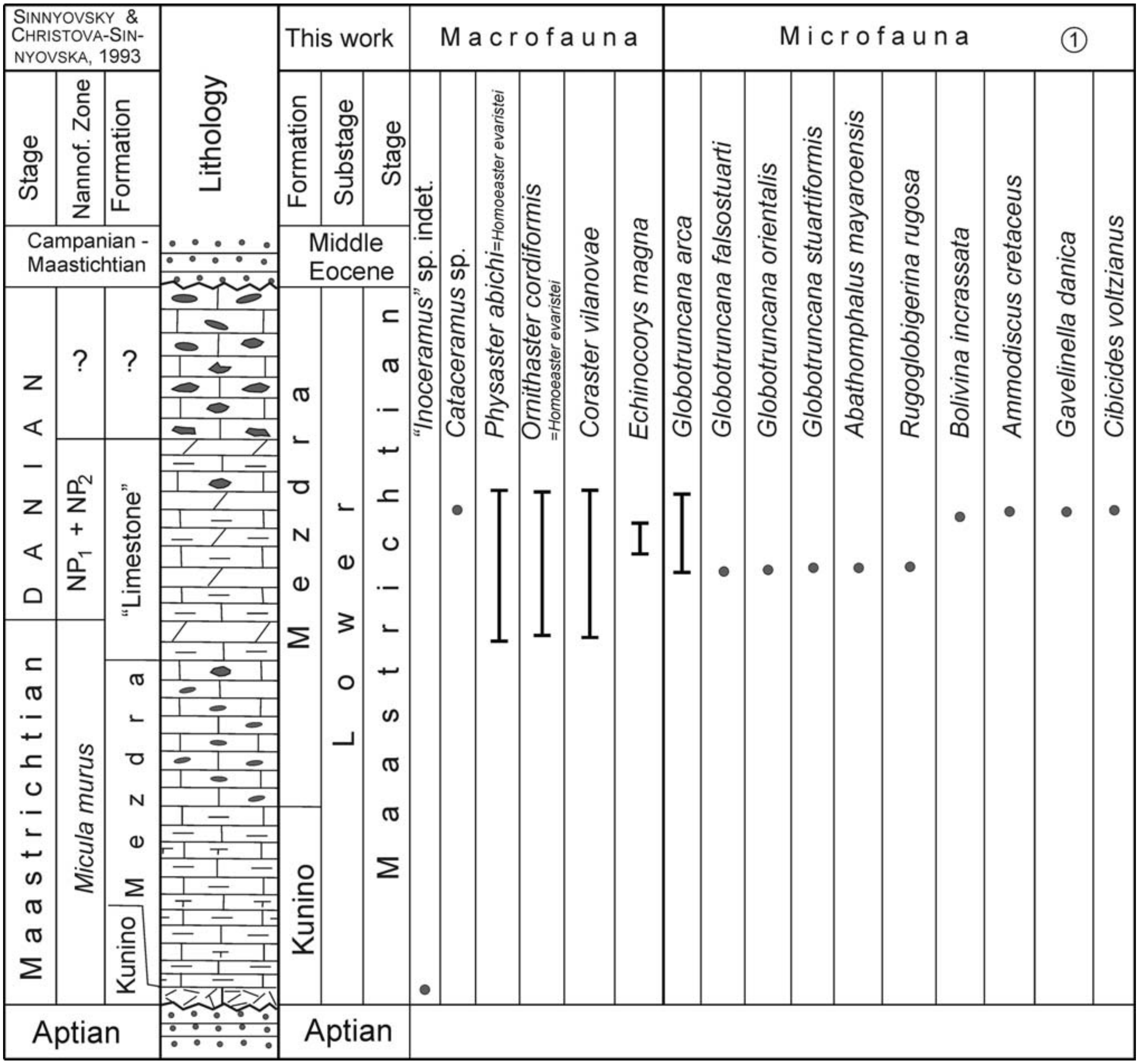

Fig. 3. Schematic stratigraphic section of the Maastrichtian Stage south of the village of Perchovtzi (Section I in Fig. 1). (**1**) - EK. DIMITROVA, unpublished data. For legend see Fig. 2.

Echinocorys gr. *conoidea,* hence a Maastrichtian age is not in doubt (J.W.M. JAGT, pers. comm., April 2006).

In outcrops south of the village of Perchovtzi (Section I in Fig. 1; Fig. 3) and at the Malata reka River (Section II in Fig. 1; Fig. 4), SINNYOVSKY & CHRISTOVA-SINNYOVSKA (1993, p. 38, fig. 8) and SINNYOVSKY (2001, p. 15, fig. 3) did not include the normal and complete magnitude of the Mezdra Formation. Furthermore, in the "Paleocene nannofossil zone" in these sections, single specimens of *Cataceramus* sp. and *Cataceramus* cf. *regularis* (Figs. 5A, B), and a juvenile ammonite undeterminable to the species level cf. Pachydiscidae (Figs. 5C, D) have been found. Ammonites are common in the "limestone formation" along the southern limb of the Lyutidol syncline, SW of the village of Lipnitza (Fig. 1).

In my opinion, the upper portion of the Mezdra Formation in the southern limb of the Mezdra syncline (Fig. 2) above the hardground (SINNYOVSKY, 1991, p. 264, fig. 2; 1998, p. 12, fig. 4; p.14, fig. 6), which comprises bioclastic, medium- to coarse-grained limestones, in fact belong to the base of the Kajlâka Formation. These limestones are analogous to the limestones of Unit 9 in the northern limb of the Mezdra syncline (JOLKIČEV, 1982, pp. 18–19, fig. 7). From these, in the eastern centricline of the Mezdra syncline, a single specimen of *Cataceramus balticus* BÖHM and one of "*Inoceramus"* sp. indet. have been found.

The transitional limestones are followed upwards by whitish (with beige), indistinctly bedded micro- to medium-grained quarry limestones – the so-called "Vratza

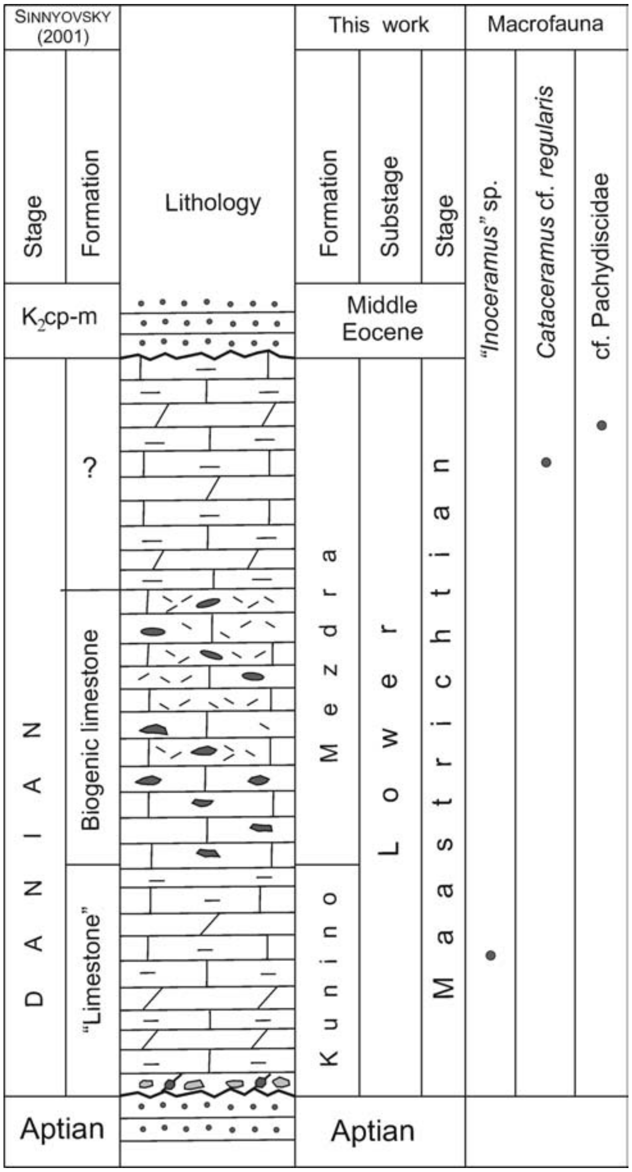

Fig. 4. Schematic stratigraphic section of the Maastrichtian Stage near to the village of Lyutidol – on the left bank of the Malata River (Section II in Fig.1). For dating the "Danian" strata SINNYOVSKY refers to the nannofosil taxa mentioned in the same paper (SINNYOVSKY, 2001, p. 12). For legend see Fig. 2

Stone". From these limestones, ZLATARSKI (1910) mentioned *Baculites baculoides* ROEMER (probably misidentified ammonite), *Baculites* sp., *Hamites* sp.; while BONČEV & KAMENOV (1934, p. 81) reported *Hemipneustes striatoradiatus* (LESKE) and *Belemnitella mucronata* (SCHLOTHEIM) (probably misidentified *Belemnitella*), and JOLKIČEV (1982, p. 18 – packet 9; p. 19, fig. 7) noted a *Pachydiscus gollevillensis gollevillensis* (D'ORBIGNY), = *Anapachydiscus* cf. *terminus* WARD & KENNEDY, 1993 which is figured here (Fig. 5E).

These sediments are overlain by light grey to whitish fine-, medium- to coarse-grained limestones from which BONČEV & KAMENOV (1934, p. 82) collected *Hemipneustes striatoradiatus*, "*Inoceramus*" sp., *Pycnodonte vesicularis* (LAMARCK), *Exogyra decussata* COQUAND and *Pecten* (=*Entolium*) cf. *membranaceus* (NILSSON) (Fig. 2).

In the area of the Lyutidol syncline, the Kajlâka Formation is preserved only in the southern limb of the structure – along the left bank of the Malata River, at the southern end of the village of Lyutidol. There, different horizons of Maastrichtian strata are transgressively overlain by terrigenous Middle Eocene deposits (TZANKOV *et al*., 1991), which SINNYOVSKY & CHRISTOVA-SINNYOVSKA (1993) and SINNYOVSKY (1993, 2001) assumed to be in allochthonous position and of Campanian–Maastrichtian age, as defined by nannofossils (Figs. 3, 4). I subscribe to the transgressive, but not allochthonous, position of the terrigenous sediments upon the Maastrichtian ones. The nannofossil samples have presumably been collected from Upper Cretaceous blocks, included as a common component within Middle Eocene terrigenous sediments.

Disregarding the presence of inoceramids, cephalopods and characteristic Maastrichtian echinoid fauna in the whole section of the Upper Cretaceous series in these structures, SINNYOVSKY & CHRISTOVA-SINNYOVSKA (1993), SINNYOVSKY (1991, 1993, 1998, 2001) and STOYKOVA *et al*. (2000), on the basis of nannofossils, defined the Paleocene age for most of this section (Figs. 2–4). They assumed (pers. comm., 2004) the Maastrichtian inoceramid, cephalopod and echinoid fauna, which occurs in the range of their "nannofossil zones", as well as the macrofauna from the Kajlâka Formation, to have been redeposited. I assert that this does not correspond to the fossil sequences in the section and there is no physical evidence of resedimentation of Maastrichtian macrofossils.

The outcrops of the Mezdra Formation continue into the Fore Balkan and to the west of the Mezdra syncline as far west as the valley of the Ogosta River. There, in a quarry at the village of Lyuta (now Vladimirovo), Vratza District, BONČEV & KAMENOV (1934, p. 80) found *Pachydiscus neubergicus* (VON HAUER) together with numerous echinoids, analogous in specific content to those from the Mezdra Formation in the area of Mezdra (determined also by the present author). SINNYOVSKY (2003, p. 152) analysed the limestones in this quarry for nannoplankton and "proved" that they are of Paleocene and not of Maastrichtian age. SINNYOVSKY is well aware of the presence of Maastrichtian macrofossil taxa at this locality, cited by him in this paper (p. 149), but fails to comment on this fact.

Conclusion

The normal superposition of lithostratigraphic units, which form the limbs of the Mezdra and Lyutidol synclines, as well as their macro- and microfossil content unambiguously confirm their Maastrichtian age.

The co-occurrence of Paleocene nannofossils and Maastrichtian macrofossil taxa in the sections of these

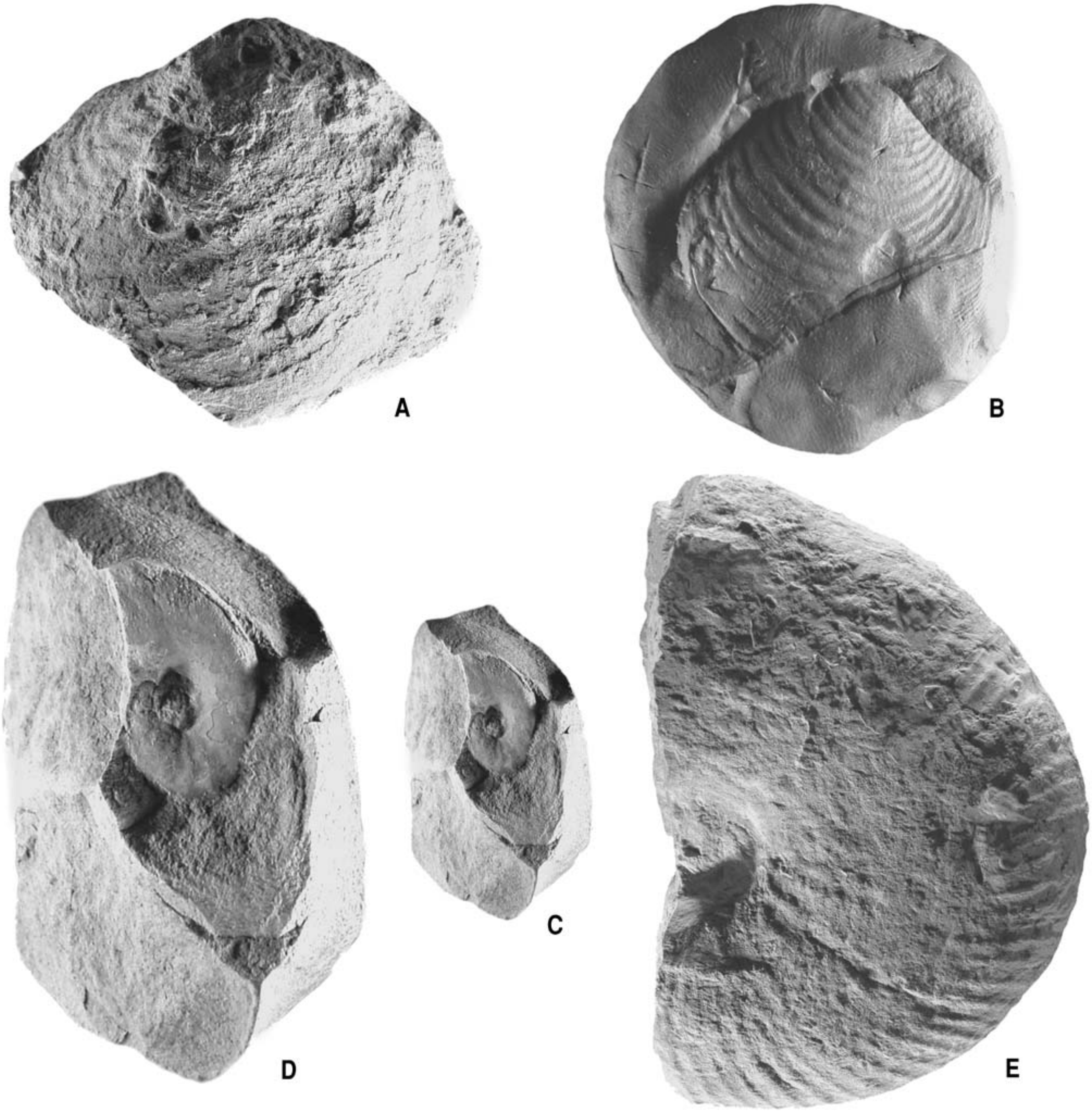

Fig. 5. **A**. *Cataceramus* sp., specimen No SU6030, Mezdra Formation, south of the village of Perchovtzi (section I from Fig. 1; Fig. 3); x 1. **B**. *Cataceramus* cf. *regularis* D'ORBIGNY, specimen No SU6029 (plaster cast), Mezdra Formation, on the left bank of Malata River, near the village of Lyutidol (section II from Fig. 1, Fig. 4); x 1. **C**, **D**. A juvenile ammonite undeterminable to the species level cf. Pachydiscidae, specimen No SU6028, Mezdra Formation, on the left bank of Malata River, near the village of Lyutidol (section II from Fig. 1, Fig. 4); C, × 1; D, × 2. **E**. *Anapachydiscus* cf. *terminus* WARD & KENNEDY, specimen No SU267, found in Kajlâka Formation, the quarry at the village of Varbeshnitza, NW from Mezdra (mentioned in JOLKIČEV, 1982, p. 18, packet 9, p. 19, fig. 7 as *Pachydiscus gollevillensis gollevilensis* (D'ORBIGNY), x 0.5. SU – collection numbers from the Museum of Paleontology at Sofia University "St. Kliment Ohridski".

structures indicates the diachronic appearance of macro- and nannofossils at the Cretaceous/Paleogene boundary. From this viewpoint, the respective boundary cannot be fixed by nannofossils except if it is assume the Maastrichtian macrofossils to have been re-deposited, but this is not the case. Furthermore, the Maastrichtian macrofauna is found not only within the ranges of the "nannofossil zones" but also in the sections overlying them – in the Kajlâka Formation, where a number of new Maastrichtian taxa, such as *Hemipneustes stria-*

toradiatus, appear. Accompaning to this taxon, inoceramids and cephalopods continue to occur (Fig. 2). These facts call into question the applicability of nannofossils for defining the Cretaceous/Paleogene boundary.

This recalls the situation in the type area of the Maastrichtian Stage, where all nannofossil taxa except one (*Biantholithus sparsus*), including the ones held to be indicative of the lower Paleocene, already occur in the underlying Maastricht Formation [(Mai *et al*, 1994; Mai *et al*, 1997a; Mai *et al*, 1997b; Mai, 1999; Mai *et al*, 2003), yet in a different size category], which is well dated by macrofossil taxa as late Maastrichtian (J.W.M. Jagt, pers. comm., 2005).

Diachronism in the occurrences of macro- and nannofossils is observed not only at the boundary Cretaceous/Paleogene, but also at other boundaries, e. g. the Campanian/Maastrichtian boundary in some European outcrops (Jagt & Felder, 2003; Küchler & Wagreich, 1999; Wagreich *et al.,* 2003). Robaszynski *et al.* (1985) also expressed some doubts on the applicability of nannofossils in determining the Campanian–Maastrichtian boundary and pointed out that "the Campanian–Maastrichtian boundary is somewhat difficult to recognize with nannoplankton because of problems in determining the index species and possible diachronism of their appearances and extinctions from the Tethyan to the Boreal realms". Wagreich (1987, p. 85) stated that "no exact correlation of nannoplankton and macrofossil zonation at the Campanian/Maastrichtian boundary for low and high latitudes exists". According to Burnett (1998, p. 137) "stages have been historically defined onshore using macrofossils. In the absence of macrofossil data from oceanic cores, stages boundaries started to be "defined" using microfossil events". Finally, Burnett (1998, p. 137) concluded: "Nannofossils do not define the bases of any Upper Cretaceous stages." This evidence, as well as the data presented above, shows that nannofossils should be used in biostratigraphy with more care in the case of chronostratigraphic boundaries already fixed by macrofauna.

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Резиме

Граница креде и палеогена у синклиналама Мездре и Љутидола у области Враца (западни Предбалкан, Бугарска)

У раду се говори о неоправданом приписивању (на основу кречњачких нанофосила) великог дела мастрихтског профила у синклиналама Мездре и Љутидола (западни Предбалкан, Бугарска) палеогену. Горњокредни седменти у синклиналама Мездре и Љутидола врацке области западног дела Предбалкана у Бугарској широко су распрострањени и великог су тектонског значаја. Bončev

(1932) је први доказао да су горњокредни седименти Предбалкана јужно од реке Искар мастрихтске старости. Мастрихтску старост су потврдили сви каснији истраживачи на основу макрофосилне фауне (COHEN, 1946; TZANKOV, 1968; JOLKIČEV, 1986, 1989; и други).

За кречњачке нанофосиле се последњих година претпоставља да су изузетно важни за расчлањавање горњокредних и кенозојских седимената – важност коју изгледа не можемо да доведемо у питање. Под утицајем нанопланктонске еуфорије недавно се појавило неколико публикованих радова у којима се граница креде и палеогена у проучаваном подручју прати не узимајући у обзир присуство карактеристичне макрофауне у истим слојевима који су расчлањени на основу нанопланктона. Граница креде и палеогена одређена помоћу макрофауне занемарује се у тим радовима.

Макрофаунистички подаци приказани у овом раду покрећу питање применљивости нанофосила за дефинисање границе креда–палеоген у датој области. Занемарујући присуство иноцерамуса, цефалопода и карактеристичне мастрихтске ехинидске фауне у целом профилу горњокредне серије у овим структурама, SYNNOVSKY & CHRISTOVA-SYNNOVSKA (1993), SYNNOVSKY (1991, 1993, 1998, 2001) и STOYKOVA *et al.* (2000) одредили су на основу нанофосила палеоценску старост највећег дела профила (сл. 2–4). Они претпостављају (усмено саопштење, 2004) да су мастрихтска иноцерамска, цефалоподска и ехинидска фауна, која се јавља у границама њихових "нанофосилних зона", као и макрофауна формације Кајлака, преталожене. Ја тврдим да то одговара фосилној секвенци у профилу и да не постоји материјални доказ преталоживања мастрихтских макрофосила.

Нормална суперпозиција литостратиграфских јединица које формирају крила синклинала Мездра и Љутидол, као и њихов макро и микрофосилни садржај недвосмислено потврђују њихову мастрихтску старост. То је приказано у овом раду бројним чињеницама.

Истовремена појава палеоценских нанофосила и мастрихтских макрофосилних таксона у профилима ових структура указује на дијахроничну појаву макро и нанофосила на граници креде и палеогена. Са овог становишта одговарајућа граница се не може утврдити помоћу нанофосила осим ако не предпоставимо да су мастрихтски макрофосили били преталожени, али то овде није био случај. Осим тога, мастрихтска макрофауна је нађена не само у границама "нанофосилних зона" већ и у слојевима изнад њих – у формацији Кајлака, где се јавља више нових мастрихтских таксона као што је *Hemipneustes striatoradiatus*. Поред овог таксона и даље се јављају иноцерамуси и цефалоподи (сл. 2). Ове чињенице доводе у питање применљивост нанофосила за дефинисање границе креда–палеоген.

То потсећа на ситуацију у типској области мастрихтског ката, где се сви нанофосилни таксони осим једног (*Biantholithus sparsus*) укључујући и оне за које се сматра да указују на доњи палеоцен, јављају у подини Мастрихтске формације [(MAI, 1999; MAI *et al.*, 1994, 1997a, 1997b, 2002), мада другачијих димензија], која је поуздано одређена на основу макрофосилних таксона као горњомастрихтска (JAGT, усмено саопштење, 2005).

Дијахронизам у појавама макро и нанофосила запажен је не само на граници креде и палеогена, већ и на другим границама, напр. граници кампана и мастрихта у неким инданцима у Европи (JAGT & FELDER, 2003; KÜCHLER & WAGREICH, 1999; WAGREICH *et al.*, 2003). ROBASZYNSKI *et al.* (1985) такође су изразили сумњу у применљивост нанофосила за одређивање границе кампан–мастрихт и указали да је "границу кампан–мастрихта донекле тешко препознати на основу нанопланктона због проблема утврђивања водеће врсте и могућег дијахронизма њиховог појављивања и изумирања од Тетиса до бореалних области". WAGREICH (1987, стр. 85) констатује да "не постоји тачна корелација нанопланктонског и макрофосилног зонирања на граници кампан–мастрихт за мање и веће географске ширине". Према BURNETT-у (1998, стр. 137), "катови су историјски дефинисани на копну помоћу макрофосила. У недостатку макрофосилних података из океанских језгара, границе катова су почеле да се "одређују" помоћу микрофосила". На крају, BURNETT (1998, стр. 137) закључује: "Нанофосили не одређују базе било којих катова горње креде". Овај доказ као и подаци приказани у овом раду показују да у биостратиграфији нанофосиле треба користити са више пажње у случају хроностратиграфских граница које су већ утврђене помоћу макрофауне.

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Latest Cretaceous mosasaurs and lamniform sharks from Labirinta cave, Vratsa district (northwest Bulgaria): a preliminary note

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Abstract. Preliminary descriptions are given of selected specimens from an assemblage of >65 isolated vertebrate remains, collected in 1985 at the Labirinta cave, situated between the villages of Drashan and Breste, east of Cherven Briag (Vratsa district, northwest Bulgaria), from strata of late Maastrichtian age (Kajlâka Formation). Recorded are a fragmentary lower jaw of a mosasaurine squamate, *Mosasaurus* cf. *hoffmanni* (MANTELL, 1829), with two teeth preserved *in situ*, as well as two isolated teeth of lamniform sharks, assigned to *Squalicorax pristodontus* (AGASSIZ, 1843) and *Anomotodon* sp. Other vertebrate remains in this assemblage include rather poorly preserved fragments of ?skull and appendicular skeleton of mosasaurs, but it cannot be ruled out that other vertebrate groups (?elasmosaurid plesiosaurs) are represented as well. To establish this, the additional material needs to be studied in detail and compared with existing collections; it will be described in full at a later date. A partial phragmocone of a scaphitid ammonite, found associated, is here assigned to *Hoploscaphites constrictus* (J. SOWERBY, 1817) and briefly described as well. This record dates the Labirinta cave sequence as Maastrichtian, as does the echinoid *Hemipneustes striatoradiatus* (LESKE, 1778); tooth morphology of *Squalicorax pristodontus* and a find of the pachydiscid ammonite *Anapachydiscus* (*Menuites*) cf. *terminus* WARD & KENNEDY, 1993 from correlative strata nearby narrow this down to late, or even latest, Maastrichtian. Finally, some remarks on mosasaur and plesiosaur distribution during the Campanian–Maastrichtian across Europe are added.

Key words: Mosasaurs, lamniform sharks, Maastrichtian, Bulgaria, scaphitid ammonites, echinoids, stratigraphy.

Апстракт. Дају се претходни описи примерака одабраних из асоцијације од преко 65 издвојених остатака кичмењака, сакупљених 1985. године у пећини Лабиринта између села Драшан и Бресте источно од Червеног Бриага (Врачански крај, северозападна Бугарска) из слојева горњомастрихтске старости (формација Кајлака). Регистрован је део доње вилице краљуштастог мозазаура, *Mosasaurus* cf. *hoffmanni* (MANTELL, 1829), са два очувана зуба ин ситу, као и два посебна зуба ламниформних ајкула, који се приписују *Squalicorax pristodontus* (AGASSIZ, 1843) и *Anomotodon* sp. Међу осталим остацима кичмењака у овој асоцијацији налазе се доста слабо очувани фрагменти ?лобање и припадајућег скелета мозазаура, али се не искључује присуство и других група кичмењака (?еласмозауридски плезиозаури). Да би се то утврдило, потребно је детаљно проучити допунски материјал и упоредити га са постојећим колекцијама. Потпунији опис ће бити накнадно урађен. Делимични фрагмакон скафитидног амонита, нађен у асоцијацији, приписује се *Hoploscaphites constrictus* (J. SOWERBY, 1817) и укратко се описује. Према овом налазу, као и на основу јежа *Hemipneustes striatoradiatus* (LESKE, 1778); секвенца пећине Лабиринта одређује се као мастрихтска; морфологија зуба *Squalicorax pristodontus* и налазак пахидисцидног амонита *Anapachydiscus*

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(*Menuites*) cf. *terminus* WARD & KENNEDY, 1993 из оближњег изданка формације Кајлака омогућава прецизнију одредбу старости као касни мастрихт, или чак најкаснији мастрихт. Ближа одредба на горњи мастрихт заснива се на морфологији зуба. На крају, дају се и нека запажања у вези распрострањености мозазаура и плезиозаура за време кампан–мастрихта широм Европе.

Кључне речи: Мозозаур, ламниформне ајкуле, мастрихт, Бугарска, скафитидни амонити, јежеви, стратиграфија.

Introduction

In the summer of 1981, a team of speleologists discovered a new, unexplored cave in Upper Cretaceous limestones between the villages of Drashan and Breste, Vratsa district (northwest Bulgaria; Fig. 1). This expedition was organised by the speleoclub ‘Stalacton’, based in the nearby town of Cherven Briag. After a vertical descent of eight metres, the speleologists encountered a labyrinth of several galleries with a total length of about 1 km. The new cave was named Labirinta, Bulgarian for ‘The labyrinth’. On the way back to the entrance, 28 m below the surface, the group came across several fossil bones protruding from the cave wall at two sites (A, B in Fig. 2). During a subsequent expedition to the same area, the speleologist Zdravko Iliev invited two palaeontologists, Drs Stoycho Breskovski and Vassil Popov, who noted that this fossil occurrence was significant and rather unique. Together with Dr Nikolay Spassov they were responsible for a preliminary identification of the bones excavated. They attributed them to the extinct squamate family Mosasauridae and, more specifically, to the genus *Mosasaurus*. In the summer of 1985, a palaeontological excavation was carried out, during which all fossil material accessible was collected, albeit rather chaotically without documenting the exact position of specimens taken from the rock. The material from the two sites was subsequently mixed and transferred to the collections of the National Museum of Natural History Sofia (NMNHS). This excavation has so far been described in a popular paper (GENOV, 1985) only. Part of the material was later sent to the Paleontological Institute of the Russian Academy of Sciences (Moscow) for detailed examination, while the remainder stayed at Sofia. After that, studies came to a halt. A single tooth from this lot was put on exhibit in the Paleontology Hall of the National Museum of Natural History, but the material was never formally published. The current whereabouts of the specimens sent to Moscow is unknown.

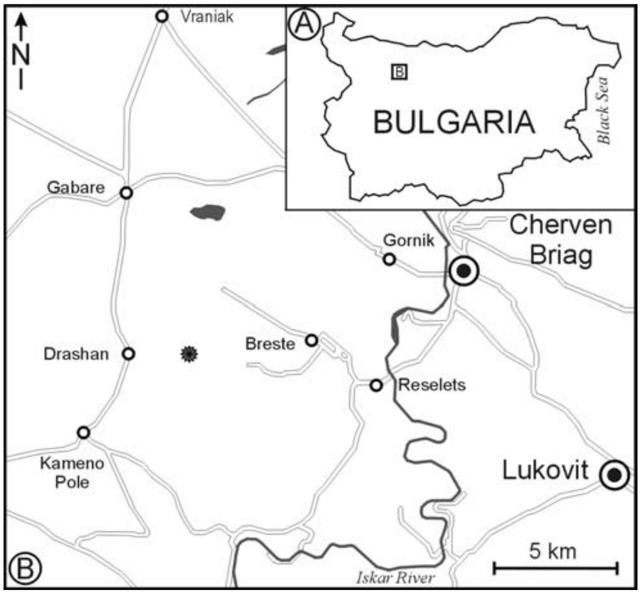

Fig. 1. Locality map of the study area in Vratsa district, northwest Bulgaria; the asterisk denotes the location of the Labirinta cave between the villages of Drashan and Breste.

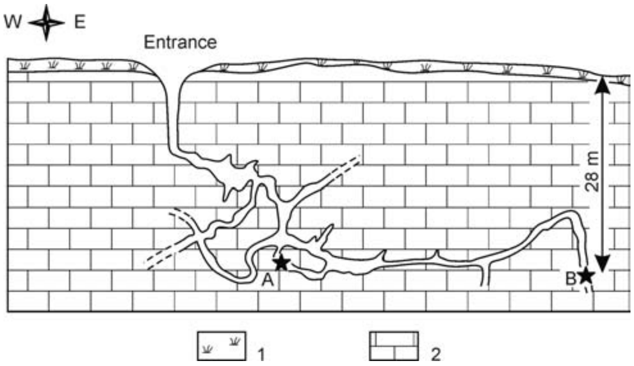

Fig. 2. Schematic vertical section of the Labirinta cave, showing the provenance (A and B) of macrofossil collections described herein; 1, soil; 2, limestones of the Kajlâka Formation.

In total, this lot comprises sixty registered, mostly fragmentary, specimens. Amongst them are two teeth (one of which is described below) associated with part of a jaw bone [NMNHS 11897/1 (*ex* Mos 60) and 11897/2 (*ex* Mos 20)], a radius, single phalanges, ribs, and vertebrae. For the present note, we have selected the following elements for brief discussion: a fragment of a lower jaw (NMNHS 11897/1) of a mosasaurine mosasaur with a tooth preserved *in situ*, and two isolated teeth of lamniform sharks [NMNHS 31362 (*ex* Mos 55) and NMNHS 31363 (*ex* Mos 7)]. Some of the other material is rather poorly preserved, being embedded in an indurated matrix not conducive to mechanical preparation. It may be that the present lot also includes isolated bones of other vertebrate groups, e.g. elasmosaurid plesiosaurs. Our aim is to provide, at a later date, detailed descriptions of this material (held at NMNHS), within the framework of a revision of all mosasaur material from the Upper Cretaceous of Bulgaria known to date, inclusive of the originals of TZANKOV (1939) and NIKOLOV & WESTPHAL (1976).

Of interest is a single scaphitid phragmocone (NMNHS 29929), collected when recovering the vertebrate remains from the Labirinta cave; its stratigraphic value is briefly commented on. Added also are more general notes on mosasaur and plesiosaur distribution in the Campanian–Maastrichtian across Europe, with references to a few recent papers.

Geological and stratigraphical setting

The Labirinta cave is situated within limestones assigned to the Kajlâka Formation (JOLKIČEV, 1986), a unit widely distributed in the Fore-Balkan and Moesian Platform of northern Bulgaria and usually the highest Cretaceous unit in outcrops in this area. Geomorphologically, the entrance to the Labirinta cave is situated in a sinkhole; such karst phenomena are very common in this area and are developed along vertical fractures in the Kajlâka Formation. This formation comprises whitish or beige, medium- to thick-bedded (albeit indistinctly), recrystallised limestones, and varies in total thickness between 10–25 and 200–280 m. In places, the limestones contain organogenic layers composed of shells and detritus mostly of bryozoans, bivalves and echinoids and rarely of gastropods, brachiopods and cephalopods. In the study area, within this limestone unit, there is an interval of 50–60 metres of whitish (with a beige hue), indistinctly bedded, fine- to medium-grained ‘quarry type’ limestones, the so-called Vratsa Stone, famous in Bulgaria and some other European countries for wall tiling (see JOLKIČEV, 2006, fig. 2). We assume that the mosasaurid material described by NIKOLOV & WESTPHAL (1976), and briefly commented upon below, originated from this interval.

The rock which yielded the mosasaur and shark material from the Labirinta cave described herein is a light grey, strongly recrystallised, slightly sandy limestone, containing Mn-oxihydroxide dendrites. This interval of the Kajlâka Formation probably correlates with the highest limestone unit in a section near the village of Varbeshnitsa, northwest of Mezdra. This highest unit overlies the ‘quarry type’ limestones (see description of section in JOLKIČEV, 1982, p. 18, fig. 7; topmost limestone unit 10).

The age assignment of the Kajlâka Formation in the study area has been based mostly on the superposition of this unit on strata of early Maastrichtian age, and on ammonite and echinoid evidence. A single echinoid species, commonly recorded from this formation, *Hemipneustes striatoradiatus* (LESKE, 1778), corroborates a Maastrichtian date [it occurring most commonly in the upper Maastrichtian]. The isolated scaphitid phragmocone (Fig. 3) found associated with the vertebrate remains from Labirinta cave is here assigned to *Hoploscaphites constrictus* (J. SOWERBY, 1817), also confirms a Maastrichtian date. The tooth morphology of the lamniform shark described here as *Squalicorax pristodontus* (AGASSIZ, 1843) allows this to be specified as late Maastrichtian, and the pachydiscid ammonite, *Anapachydiscus (Menuites)* cf. *terminus* WARD & KENNEDY, 1993, recorded from the ‘quarry type’ limestones of the Kajlâka Formation at the nearby village of Varbeshnitsa (see JOLKIČEV, 2006, fig. 5E) narrows the dating more precisely to the latest part of the late Maastrichtian. *Anapachydiscus (Menuites) terminus* is also known from the uppermost Maastrichtian of the Bay of Biscay sections (France, Spain), the southeast Netherlands, northern and eastern Denmark, central Poland, Azerbaijan, Crimea and South Africa (see WARD & KENNEDY, 1993) and the Bjala area of eastern Bulgaria (IVANOV, 1995).

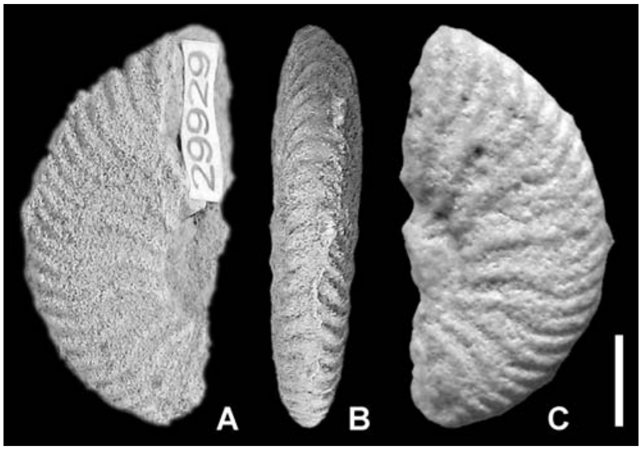

Fig. 3 A–C. *Hoploscaphites constrictus* (J. SOWERBY, 1817), NMNHS 29929, partial phragmocone, in lateral (left and right) and ventral views. Scale bar equals 5 mm.

Preliminary descriptions

To denote the repositories of material described, illustrated and/or referred to, the following abbreviations are used in the text: IRScNB, Institut Royal des Sciences Naturelles de Belgique, Brussels; MNHN, Muséum National d’Histoire Naturelle, Paris; NHMM, Natuurhistorisch Museum Maastricht, Maastricht; NMNHS, National Museum of Natural History, Sofia.

Mosasaurs

From what can be seen, the fragmentary lower jaw with two teeth preserved can be assigned to the mosasaurinae genus *Mosasaurus* CONYBEARE, 1822 (type species: *Mosasaurus hoffmanni* MANTELL, 1829).

***Mosasaurus* cf. *hoffmanni* MANTELL, 1829**
Fig. 4 A–D

Material. A partial lower jaw with two teeth associated (NMNHS 11897); here, we only describe the larger tooth (NMNHS 11897/1). The smaller one, situated more posteriorly, is longitudinally broken and not prepared yet.

Description. Being partially embedded in hard matrix and, in general, of rather poor preservation, the jaw bone cannot be described in detail. From what can be seen (e.g., foramina, overall size and height as compared to tooth size, as well as dental and root structure; Fig. 4A), this fragment is best interpreted as the anterior/mesial portion of a lower jaw (dentary) of an adult individual, pending further preparation. It contains two teeth *in situ*; the larger of these is a sturdy, bicarinate crown (incomplete, tip broken off; Figs. 4 A–D) measuring 43 mm in overall height (as preserved; original height estimated to have been 46 mm), and 19 mm in basal width, in mesodistal direction. It has an elliptical, typically U-shaped cross section (Fig. 4 D), with unequal labial and lingual surfaces; labial face gently convex with no facetting visible, not even proximally; lingual face deeply U-shaped and no facetting seen either. Enamel beading is not well developed; it is seen only in patches proximally. The crown is moderately posteriorly and lingually recurved. Both anterior and posterior carinae are well developed, but partially damaged; carinae minutely serrated over their entire length. The uppermost portion of the root is exposed considerably (Fig. 4A), being reversed conical in shape but lacking a clearly developed 'rim'; its length cannot be determined and the resorption pit is not seen. The extent to which the root is exposed is exceptional in comparison to material of *M. hoffmanni* from the Maastrichtian type area; it could be a pathological feature of the present individual.

Remarks. Tooth morphology (U-shaped cross section, minutely serrated carinae, obliquely positioned carinae, recurvature and enamel beading) allow this material to be compared favourably with *Mosasaurus hoffmanni*, the largest species in the genus *Mosasaurus*. The type material of *M. hoffmanni* is from the upper part of the Maastricht Formation (upper Nekum Member; holotype is MNHN AC 9648) in the type area of the Maastrichtian Stage (St Pietersberg, Maastricht and environs, the Netherlands; see LINGHAM-SOLIAR, 1995; BARDET & JAGT, 1996; KUYPERS *et al*., 1998) and is of late Maastrichtian age (*Belemnitella junior* Zone of authors). In southern Limburg (the Netherlands) and adjacent Belgian territory (provinces of Liège and Limburg), *M. hoffmanni* (or a closely related taxon) first appears, albeit extremely rarely, in the upper Vijlen Member (Gulpen Formation; interval 6), is comparatively rare in the remainder of this formation (Lixhe 1–3 and Lanaye members) and the lower portion of the overlying Maastricht Formation (Valkenburg, Gronsveld and Schiepersberg members), but common in the Emael

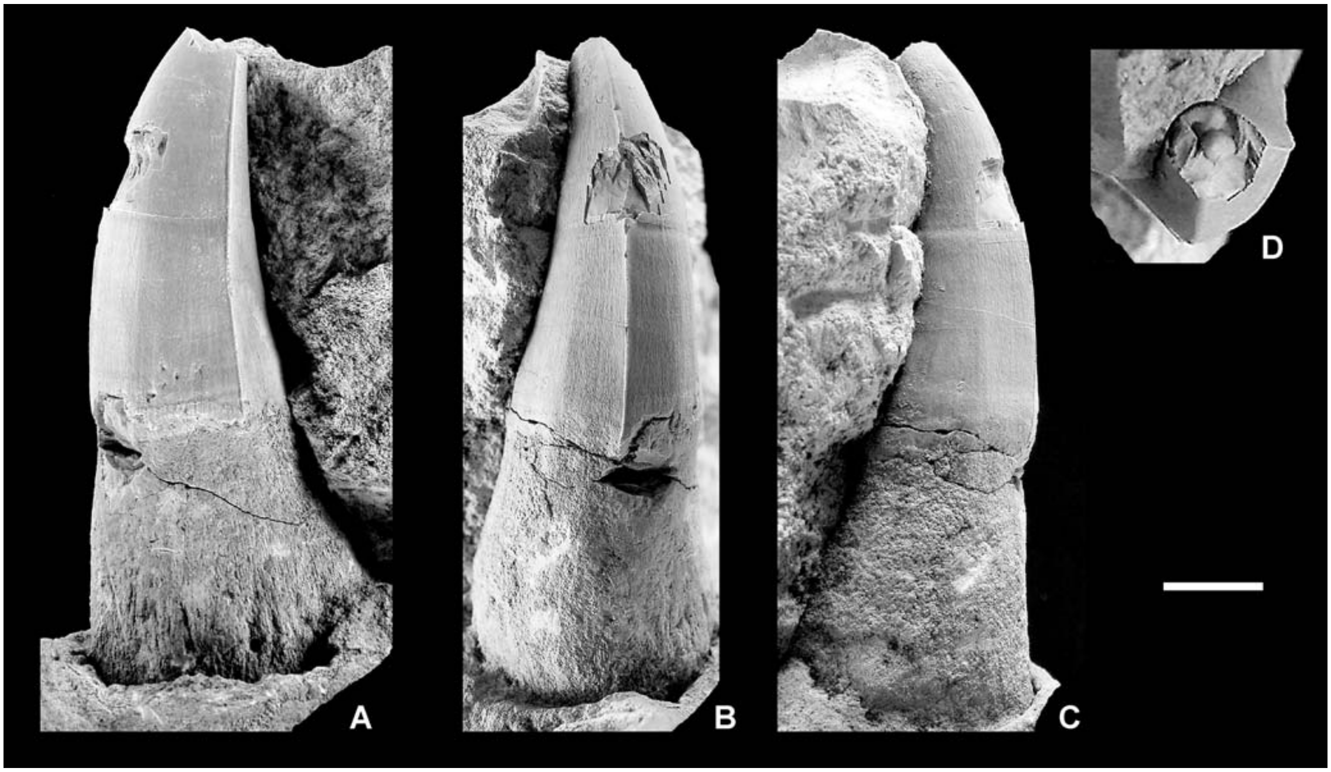

Fig. 4 **A–D**. *Mosasaurus* cf. *hoffmanni* MANTELL, 1829, anterior-mesial fragment of lower jaw (NMNHS 11897/1), with a single tooth preserved *in situ*; tooth shown in labio-mesial (**A**), mesial (**B**), lingual (**C**) and apical (**D**) views, respectively. Scale bar equals 10 mm.

and Nekum members of that formation, in particular. The species ranges to within a metre below the Cretaceous–Paleogene (K/Pg) boundary in the area, and shows a wide range of variation in tooth morphology (facetting, size, enamel structure) which may be ontogenetically controlled (compare MERTENS, 1942); there is currently no evidence of sexual dimorphism in mosasaur taxa. From the Eben Emael area (northwest Belgium) in particular, fairly numerous finds have been made of articulated material in the past decade, inclusive of at least five skulls (JAGT *et al*., 2002). This material is currently being studied, in an attempt to describe the species in detail and better determine its range of morphological and ontogenetic variation. Tooth morphology along the dental and (pre)maxillar rami in particular is assessed to allow isolated tooth crowns and teeth to be assigned with more confidence.

MULDER (1999) relegated the North American (New Jersey) species, *M. maximus* COPE, 1869 into the synonymy of *M. hoffmanni*, and subsequently MULDER *et al*. (2004) expressed doubts over the specific status of *M. lemonnieri* DOLLO, 1889 (holotype is IRScNB R 28 [*ex* 1470]) from the lower Maastrichtian ('Craie phosphatée de Ciply'; *Belemnella obtusa* Zone) at Mesvin, Mons Basin (southern Belgium), noting that it might well be nothing more than a juvenile *M. hoffmanni*. Interestingly, CALDWELL *et al*. (2004) noted that the type lot of *M. lemonnieri* from that unit included a specimen (IRScNB R3211) that would be better assigned to *Moanasaurus*, a genus otherwise known only from the Campanian (Haumurian) of New Zealand (WIFFEN, 1980).

The current stratigraphic and geographic range of *M. hoffmanni*, or of closely related taxa generally referred to in the literature as *M.* cf. *hoffmanni*, corresponds to the late Campanian to latest Maastrichtian in a belt across the Northern Hemisphere (LINDGREN & JAGT, 2005). From west to east this includes Missouri, Alabama, New Jersey (all USA), southern and northeast Belgium, southeast Netherlands, Denmark, northern Germany, central Poland, northwest Bulgaria and Turkey (BARDET & TUNOĞLU, 2002; KIERNAN, 2002; REICH & FRENZEL, 2002; MACHALSKI *et al*., 2003; GALLAGHER *et al*., 2005; TUNOĞLU & BARDET, 2006).

Sharks

The present lot from the Labirinta cave comprises two isolated neoselachian teeth, both assignable to lamniforms, with the larger one representing the anacoracid genus *Squalicorax* WHITLEY, 1949, and the other the ?alopiid genus *Anomotodon* ARAMBOURG, 1952.

Squalicorax pristodontus (AGASSIZ, 1843)
Fig. 5 A, B

Material. A single, incomplete lateral tooth (NMNHS 31362 [*ex* Mos 55]).

Description. This is an element of the lateral file, the crown being well preserved, broad and triangular, with a regularly convex cutting edge, bearing strong serrations. The distal cutting edge is oblique on the whole, with a slight concavity in its upper two-thirds; the labial face is very flat while the lingual one, much more reduced, is slightly convex. The root, not perfectly preserved, is also labio-lingually flattened and shows many small, irregularly spaced foramina on the labial face.

Remarks. Morphologically, this tooth can be identified as *S. pristodontus* beyond any doubt. Considering its general design, a late Maastrichtian age can be assigned to this specimen, by comparison with Maastrichtian material collected from the phosphate series of Benguerir (Morocco), currently under study (H. CAPPETTA, pers. obs.). Thus, this allows the general age assignment of the vertebrate association of Labirinta cave to be narrowed down to the late Maastrichtian.

Occurrence. This species is widely distributed and occurs in the Campanian of Belgium, France and Germany (LERICHE, 1929; ALBERS & WEILER, 1964; VULLO, 2005), but is particularly common in the Maastrichtian of the Netherlands and Belgium (LERICHE, 1929; HERMAN, 1977), Spain (CAPPETTA & CARMELO CORRAL, 1999), northern Bulgaria (TZANKOV & DATCHEV, 1966), Morocco (ARAMBOURG, 1952; NOUBHANI & CAPPETTA, 1997), Angola (ANTUNES & CAPPETTA, 2002), Syria (BARDET *et al*., 2000), New Jersey and Texas (CAPPETTA & CASE, 1975; WELTON & FARISH, 1993) and Brazil (REBOUÇAS & DA SILVA SANTOS, 1956).

Anomotodon sp.
Fig. 5 C–E

Material. A single, incomplete latero-anterior tooth (NMNHS 31363 [*ex* Mos 7]).

Description. This is an element of a latero-anterior file, the crown being fairly high, narrow, and with a sharp apex. The labial face is slightly convex transversely but shows a basal median excavation of triangular outline; the lingual face is transversely convex and completely smooth. On one side, a short oblique heel, slightly and irregularly serrated, can be seen. The root is damaged and only one side is preserved; its lingual face is rather high, and it seems that there was not a long lobe.

Remarks. In lateral teeth of *Anomotodon*, the lingual face of the crown often is devoid of folds. The same is seen in teeth of the genus *Paranomotodon* HERMAN *in* CAPPETTA & CASE, 1975, the crown of which completely lacks folds (SIVERSON, 1992; VULLO, 2005). So, on the basis of the material available, it is quite difficult to give a definite generic assignment. NMNHS 31363 can be compared to *A. toddi* CASE & CAPPETTA, 1997 (p. 142, pl. 5, figs. 1, 2) from the upper Maastrichtian Kemp Clay Formation of Texas, but also to *A. hermani* SIVERSON, 1992 (p. 544, pl. 5, figs. 1, 2) from the lower-upper Campanian of southern Belgium and south-

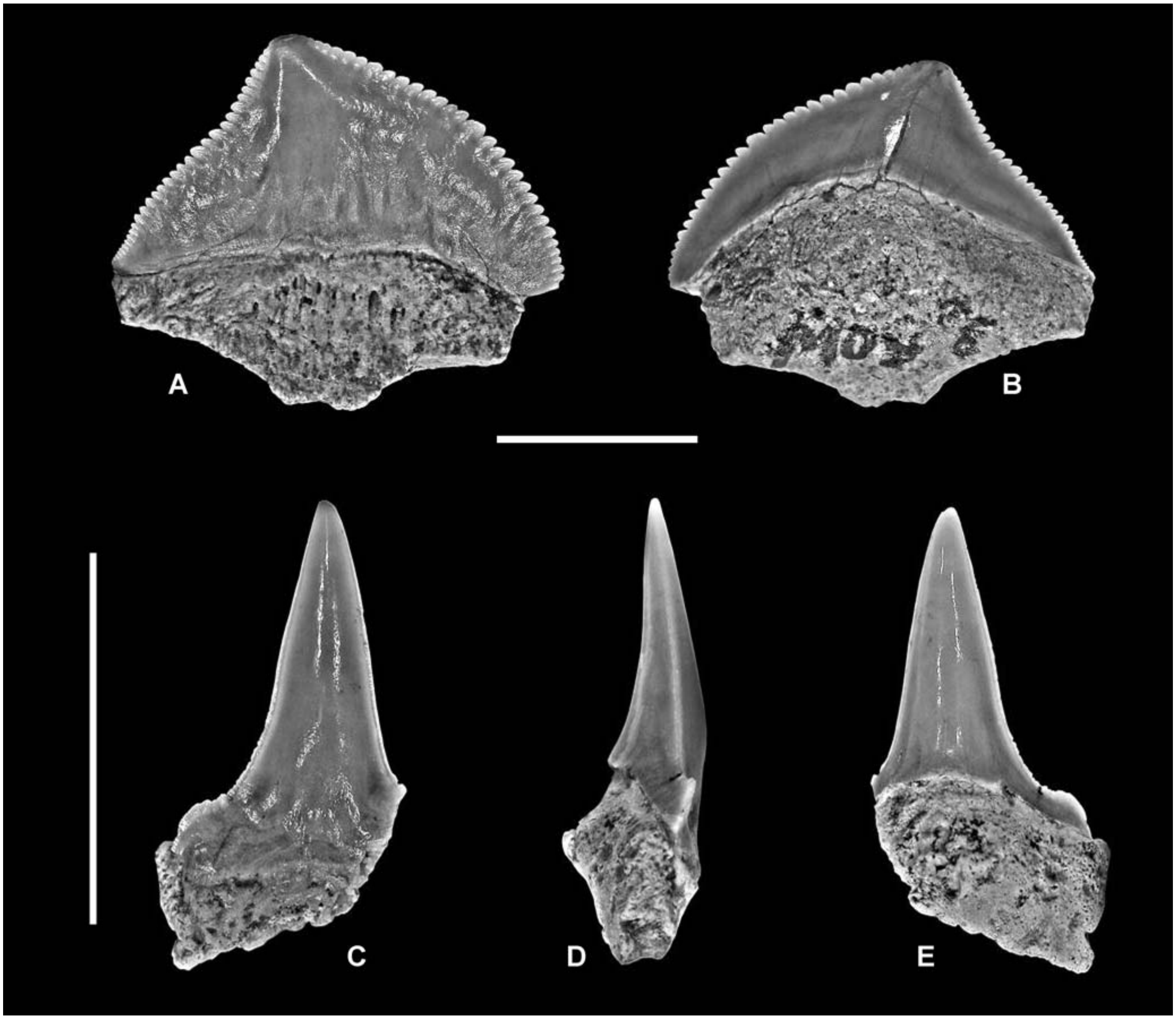

Fig. 5. Lamniform shark teeth. **A, B**. *Squalicorax pristodontus* (L. Agassiz, 1843), lateral tooth (NMNHS 31362), in labial and lingual view, respectively. **C–E.** *Anomotodon* sp., lateral anterior tooth (NMNHS 31363), in labial (**C**), lateral (profile) (**D**) and lingual (**E**) view. Scale bars equal 10 mm.

ern Sweden. Yet, as a definite specific identification is difficult on the basis of a single, imperfect tooth, it appears better to leave it in open nomenclature.

Discussion

Earlier records of mosasaurs from the Upper Cretaceous of northern Bulgaria include dissociated teeth and tooth crowns from near the village of Somovit at the River Danube, listed and illustrated by Tzankov (1939). That author referred these strata to the uppermost Maastrichtian, but his listing [in original nomenclature] of characteristic macrofossil taxa from there shows that a modern revision is called for, as there appears to be a curious mixture of typically Campanian and Maastrichtian elements (e.g., *Echinocorys gibbus*, *Bostrychoceras polyplocum*, *Belemnitella mucronata* and *Parapachydiscus colligatus* being Campanian, and *Hemipneustes striato-radiatus* and *Discoscaphites constrictus* being Maastrichtian). Mosasaur taxa recorded by Tzankov are *Leiodon anceps*, *Mosasaurus giganteus* (= *M. hoffmanni*) and *Globidens fraasi* (= *Carinodens belgicus*). Judging from his description of the matrix and based on the occurrence of the echinoid *Hemipneustes striatoradiatus* (which cannot be misidentified), we assume the material described by Tzankov (1939) to have come from strata assignable to the Kajlâka Formation as well, and thus to be of (late) Maastrichtian age.

Nikolov & Westphal (1976) recorded from an active quarry c. 2 km east of Varbeshnitsa, northwest of

Mezdra, part of the vertebral column (580 mm in length, 14 vertebrae) of a mosasaur, as well as three teeth, possibly of a single individual. The level from which these remains came was described as a unit of white to yellowish, fine-grained limestone quarried for wall tiles, c. 40 m thick, with occasional, scattered flint nodules, yielding the coleoid cephalopod *Belemnitella mucronata* and the holasteroid echinoid *Hemipneustes striatoradiatus*. The quarry exposing the Kajlâka Formation near Varbeshnitsa is in unit 9 of 'quarry-type' limestones (see JOLKIČEV, 1982, p. 18, fig. 7, limestone unit 9), having yielded *Pachydiscus gollevillensis gollevillensis* (*sensu* JOLKIČEV, 1982 = *Anapachydiscus* cf. *terminus*; see JOLKIČEV, 2006). *Anapachydiscus (Menuites) terminus* is characteristic only of the latest part of the late Maastrichtian (see remarks about the species distribution above).

With regard to the macrofossil taxa mentioned in NIKOLOV & WESTPHAL (1976), we wish to observe the following. Amongst belemnite workers, there currently is consensus that *B. mucronata* is a predominantly late Campanian species (CHRISTENSEN, 1997a, b; compare STOYANOVA-VERGILOVA & JOLKIČEV, 1993), which is why the record by NIKOLOV & WESTPHAL (1976) may be taken to refer to another species of *Belemnitella*, probably of the *B. junior* group which in northwest Europe characterises the late Maastrichtian. The presence of *B. junior* in the lower Maastrichtian of southern Limburg (the Netherlands), as noted by KEUTGEN (1996) and CHRISTENSEN *et al.* (2004), could not be substantiated in recent belemnite studies in the area (KEUTGEN *et al.*, in prep.). The holasteroid echinoid *Hemipneustes striatoradiatus* is widely distributed in the Kajlâka Formation (JOLKIČEV, 1989, 2006). This species is a typically Tethyan element, known to range through the whole of the Maastrichtian, but particularly characteristic of the latter part of the stage in the Netherlands, northeast Belgium, French Pyrenees, Navarra and Alicante (Spain), Bulgaria, Georgia, Kazakhstan, northern Caucasus and Tadzhikistan (SMITH & JEFFERY, 2000). In the Maastrichtian type area, *H. striatoradiatus* first appears in the lower Lanaye Member (Gulpen Formation), of late Maastrichtian age (*Belemnitella junior* Zone of authors; equivalents of *tegulatus/junior* Zone *sensu* SCHULZ & SCHMID, 1983) and ranges to the top of the Meerssen Member (Maastricht Formation, IVf-7; *Belemnella (Neobelemnella) kazimiroviensis* Zone) (JAGT, 2000).

Elsewhere in Bulgaria, the age of Kajlâka Formation is late Maastrichtian as based on records of the ammonite *Sphenodiscus binckhorsti* J. BÖHM, 1898 (see TZANKOV, 1982; KENNEDY, 1987; JAGT, 2002, 2005; MACHALSKI, 2005a), as well as on the basis of regular superposition in several outcrops with underlying lower Maastrichtian sediments.

For dating the present material from the Labirinta cave, an associated scaphitid phragmocone (NMNHS 29929) collected during recovery of the vertebrate material, is of importance. This consists of half a whorl (Fig. 3), of c. 22 mm in length (as preserved) and represents a highly compressed, flat-sided phragmocone, with fairly broadly rounded ventrolateral shoulders and a narrow, flattened venter (Fig. 3B). The diameter of the umbilicus is c. 6.5 mm. Ornament consists of up to nine flexuous primary ribs, arising at the umbilical seam (Fig. 3A, C); these are feebly concave and prorsiradiate on the inner flank, convex at mid-flank and concave on the outer flank and ventrolateral shoulder and convex over the venter. Primary ribs divide at mid-flank, and single (or double) intercalatories are inserted on the outer flank, giving a total of c. 22 ribs at this growth stage. No ventrolateral tubercles are seen at the largest diameter, but this may in part be preservation induced. No sutures can be seen.

Comparison with similar-sized material from the Maastrichtian type area (Maastricht Formation, Meerssen Member, subunits IVf-5 and -6; NHMM JJ 8297 and JJ 13341a–c), the upper Maastrichtian of Petites-Pyrénées, France (KENNEDY *et al.*, 1986, pl. 4) and the uppermost Maastrichtian of Turkmenia (JAGT *et al.*, in prep.) shows NMNHS 29929 to be comparatively coarsely ribbed and with a large umbilicus, but this is well within the range of variation documented for corresponding growth stages of *Hoploscaphites constrictus* (J. SOWERBY, 1817). This species is typically Maastrichtian, with records from southern Sweden, Denmark, northern Germany, the Netherlands, southern and northeast Belgium, Cotentin (Manche), Landes and Petites-Pyrénées (all France), Lleida (Spain), Switzerland, Austria, the Czech Republic, Poland, Bulgaria, the Ukraine, Carpathians, Donbass, Transcaspia, Kazakhstan and Kopet Dag, Turkmenia (KENNEDY, 1987; JAGT, 2002; NIEBUHR, 2003; MACHALSKI, 2005a, b). At least at two localities, Stevns Klint (Denmark) and the Maastrichtian type area (the Netherlands), it extends into the lower Paleocene (MACHALSKI & HEINBERG, 2005). On the basis of a detailed study of shell ornament, both of micro- and macroconchs, MACHALSKI (2005a, b) was able to demonstrate the occurrence of several temporal subspecies during the late Maastrichtian. Unfortunately, with but a single, incomplete phragmocone available, it cannot be determined to which of these taxa NMNHS 29929 might be assigned. A more detailed age assignment than Maastrichtian is thus impossible on the basis of this scaphitid record.

Isolated teeth of neoselachian sharks are common in epicontinental Upper Cretaceous strata in Bulgaria. TZANKOV & DATCHEV (1966, pl. 7, figs. 6–16) recorded, for the first time in Bulgaria, the species *Anacorax* (= *Squalicorax*) *pristodontus* (AGASSIZ, 1843) (erroneously spelled *priostodontus*) from Maastrichtian strata near the towns of Pleven and Shumen, and near the village of Somovit on the River Danube, but they did not provide a description of the occurrence of the strata, hence, it is difficult to judge from which exact level of the Maastrichtian their material came. It is presumed that their material also originates from the Kajlâka Formation. The good preservation of the specimen described here allows a comparison with Maastrichtian material from Morocco and, thus, enables a more precise age to be assigned to this specimen.

In summary, the Labirinta cave material is definitely of late (?latest) Maastrichtian date as based on dental morphology of *Squalicorax pristodontus* and on correlation with nearby outcrops, which have been dated on echinoid and ammonoid evidence.

Remarks on mosasaur and plesiosaur distribution across Europe

Mosasaurid occurrences are known from numerous outcrops of Campanian and Maastrichtian strata across Europe. Distinct stratigraphic levels, representing mostly shallow-water, (sub)littoral settings, in three areas in particular have yielded diverse assemblages, and at two of these, more or less completely preserved cranial and post-cranial material has been shown to be relatively common. These are:

1. The lower Maastrichtian 'Craie phosphatée de Mons' (*Belemnella obtusa* Zone) in the Mons Basin, southern Belgium (Lingham-Soliar & Nolf, 1990; Lingham-Soliar, 1992, 1993, 1994, 1999, 2000; Caldwell *et al*., 2004; Mulder *et al*., 2004; Jagt, 2005), which has produced '*Mosasaurus lemonnieri*' (= *M. hoffmanni* juv.?; *Moanasaurus* sp.), *Plioplatecarpus houzeaui*, *Halisaurus ortliebi*, *Prognathodon solvayi* and *Hainosaurus bernardi*. Two additional species, *Prognathodon giganteus* and *Globidens dakotensis*, are of Campanian age (Jagt, 2005);
2. The extended type area of the Maastrichtian Stage (Lingham-Soliar, 1993, 1994, 1995, 1996, 1999; Dortangs *et al*., 2002; Jagt *et al*., 2002; Mulder, 2003a, b; Schulp *et al*., 2004; Jagt, 2005; Schulp, 2006; see also Mulder & Mai, 1999). Here, Maastrichtian strata (Vijlen Member, Gulpen Formation to Meerssen Member, Maastricht Formation) have yielded *Mosasaurus hoffmanni*, '*M. lemonnieri*' (= *M. hoffmanni* juv.?), *Plioplatecarpus marshi*, *Liodon* '*sectorius*', *Carinodens belgicus* and *Prognathodon saturator*. Campanian species include *Prognathodon* '*solvayi*' and *Hainosaurus* sp. (Jagt, 2005);
3. Skåne (southern Sweden, Kristianstad Basin), from where almost no articulated material is known, but which shows a remarkably high diversity in the lower and upper Campanian in particular, having yielded material assigned to *Platecarpus* sp., *Clidastes propython*, *Platecarpus* cf. *somenensis*, *Halisaurus sternbergi*, *Dollosaurus* sp., *Hainosaurus* sp., *Tylosaurus ivoensis*, *Prognathodon* sp. and *Plioplatecarpus* sp. From the lower Maastrichtian, two species are known, namely *Plioplatecarpus primaevus* and *Mosasaurus* aff. *lemonnieri* (Lindgren, 2005a, b; Lindgren & Siverson, 2002, 2004, 2005).

Taken together, these three areas can be assumed to represent all mosasaur taxa distributed across Europe during the Campanian–Maastrichtian, with the possible exception of one or two rare taxa. From other localities in Europe, far less common, and usually highly fragmentary material (with few exceptions), is known. These probably all represent deeper-water settings, far from coastal areas, and include:

4. The Münsterland (northwest Germany; Caldwell & Diedrich, 2005), from where a late Campanian species of *Clidastes* has been recorded;
5. England (Milner, 2002), from where material assigned to *Clidastes* sp. (Santonian–Campanian; Surrey, Sussex), indeterminate mosasaurines (Campanian––Maastrichtian; Norfolk), *Prognathodon* (upper Campanian; Norfolk), indeterminate plioplatecarpines (Santonian–Campanian; Sussex, Hampshire), ?*Tylosaurus* (upper Santonian–lower Campanian; Hampshire, Kent and Yorkshire) and *Leiodon anceps* (= ?*Hainosaurus*, Coniacian–upper Campanian; Sussex, Essex and Norfolk) has been recorded;
6. Southwest Russia and Crimea (Ukraine) (Yarkov, 1993; Storrs *et al*., 2000; Schulp *et al*., in press), with a cranial and post-cranial skeleton of *Dollosaurus lutugini* from the upper Campanian of central Russia and isolated tooth crowns of *Carinodens belgicus* from the upper Maastrichtian of Trudołyubovka, Crimea and Volgogradskaya oblast', Russia;
7. Northern Spain (Bardet *et al*., 1993, 1997b, 1999; Bardet & Pereda Suberbiola, 1996), with isolated teeth and tooth crowns of *Prognathodon solvayi*, *Platecarpus* cf. *ictericus*, *Leiodon anceps*, *Leiodon* sp., *Mosasaurus* sp. and indeterminate mosasaurines;
8. Central Poland (Machalski *et al*., 2003; Jagt *et al*., 2005), with records of *Mosasaurus* cf. *hoffmanni* from the upper Campanian and uppermost Maastrichtian, *M*. cf. *lemonnieri* from the upper Maastrichtian, '*M. (Leiodon)* cfr. *anceps*' [*sensu* Arambourg, 1952] from the same level, *Hainosaurus* sp. 1 and *Prognathodon* sp. from the upper Campanian, and *Hainosaurus* sp. 2 from the upper Maastrichtian, all from sections in the Wisla River valley;
9. France (Bardet, 1990; Bardet *et al*., 1991, 1997a), with records of *Hainosaurus bernardi* and ?*Hainosaurus* sp. from the Santonian–Campanian (Somme, Aude), *Prognathodon giganteus* from the lower Campanian (Champagne) and *Platecarpus* sp. from the lower Campanian of Corbières;
10. Denmark where lower and upper Maastrichtian strata in Jylland and Sjælland have yielded very rare remains, mostly tooth crowns, of *Mosasaurus* cf. *hoffmanni* and *Plioplatecarpus* sp. (Lindgren & Jagt, 2005).

Interesting is also Rügen (northeast Germany) from where historical material of early Maastrichtian age, the current whereabouts of which is unknown, was illustrated by Reich & Frenzel (2002, pl. 2, fig. 4). This isolated tooth from the von Hagenow Collection appears conspecific to material recorded from the upper Maastrichtian of central Poland by JAGT *et al*. (2005), as *Hainosaurus* sp. 2. Representatives of the tylosaurine genus *Hainosaurus* may thus have been more widely distributed in the European Campanian–Maastrichtian than previously assumed. In general, mosasaur remains from deeper-water settings represented by the 'white chalk facies' of northeast Germany and Denmark are comparatively rare; a few isolated tooth crowns of *Mosasaurus* cf. *hoffmanni* are known from Rügen (REICH *et al*., 2005). With the exception of some vertebrae in matrix, none of the vertebrate fossils recorded by LADWIG (1997) from Lägerdorf (northern Germany) can be assigned to mosasaurs; the majority of teeth illustrated are of enchodont teleosts and allies.

In comparison to mosasaurs, plesiosaurs in the Campanian–Maastrichtian are much rarer, and mostly refer to isolated cranial and post-cranial elements, rarely to associated skeletal elements of a single animal. Plesiosaurs appear to have frequented upwelling areas, rich in food; occurrences in shallow-water environments such as the upper Campanian of the Mons Basin (BARDET & GODEFROIT, 1995) and the upper Maastrichtian of the Maastricht area (MULDER *et al*., 2000) are best explained as stemming from floating carcasses, or from animals that only on certain occasions visited the area to feed. Anomalous in this respect is the comparative richness of plesiosaur remains from the Campanian of southern Sweden (PERSSON, 1959, 1962, 1963, 1967); this may be related to preferred feeding and/or breeding grounds in the shallow-water settings of the Kristianstad Basin and other areas there. The find of a partial elasmosaurid skeleton in the uppermost Campanian (*grimmensis*/*granulosus* Zone) of Lägerdorf, northern Germany (MAISCH & SPAETH, 2004) is of special interest, in being one of the very few examples of associated plesiosaur remains known to us.

Conclusions

During the latest Cretaceous (Campanian–Maastrichtian), marine lizard (mosasaurid) species inhabited the epicontinental seas along the northerly margins of the Tethyan ocean. Mosasaurid records from Bulgaria are very rare and refer mostly to fragmentary material, all collected from strata assigned to the Kajlâka Formation. The new material described herein adds to our understanding of mosasaur distribution across Europe during the Campanian–Maastrichtian. For a proper documentation of all of the Bulgarian material, detailed comparisons with identified material contained in museum collections, both in Bulgaria and elsewhere in Europe, are needed. Future fieldwork in northern Bulgaria is also called for, in an attempt to recover more, and stratigraphically well-documented, skeletal remains there.

Rare and randomly distributed cephalopods and other characteristic macrofossils in the epicontinental Upper Cretaceous cause problems in Bulgaria, because there are no zonal subdivisions for the whole of the Upper Cretaceous based on ammonites or belemnites, or any other macrofossil group for that matter. For the Maastrichtian, in particular, only schemes based on microfossils are available. This study shows that shark teeth may be used for dating some strata where other characteristic fossils in the Upper Cretaceous of Bulgaria are missing.

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Резиме

Најмлађи кредни мозазаури и ламниформне ајкуле из пећине Лабиринта, Врачанска област (северозападна Бугарска) – претходна белешка

Из слојева који припадају формацији Кајлака, откривених у пећини Лабиринта у Врачанском крају, северозападна Бугарска, прикупљена је 1985. године збирка остатака кичмењака (укупно 65 регистрованих примерака у колекцијама Народног природњачког музеја у Софији). Од свих ових налаза даје се опис и слике једног дела доње вилице морског гуштера (мозозаура) са два зуба *in situ* и два изолована зуба ајкула (*Squalicorax pristodontus* и *Anomotodon* sp.). На основу величине, јако удубљеног попречног пресека у облику латиничног слова "U", карине и глеђног низа, овај мозозаурски материјал је условно одређен као *Mosasaurus* cf. *hoffmanni*, врсте познате из доњег и горњег мастрихта Мисурија, Алабаме, Њу Џерсија, Данске, северне Немачке, средње Пољске, Холандије, Белгије и Турске. Асоцијација макрофосилних таксона садржи и део фрагмакона скафитида, који припада *Hoploscaphites constrictus* по којем је одређена мастрихтска старост збирке. Морфологија зуба *S. pristodontus* указује на касно мастрихтску старост. Присуство ехинида *Hemipneustes striatoradiatus* такође указује на мастрихтску старост. Налаз пахидисцидног амонита *Anapachydiscus* (*Menuites*) cf. *terminus* из оближњег изданка формације Кајлака омогућава ужу одредбу старости – касни мастрихт. Ускоро ће бити извршена ревизија целокупног регистрованог материјала мозозаура из Бугарске, заједно са наласцима из пећине Лабиринта (који можда садрже и друге групе кичмењака као што су еластосауридски плесиосаури). Осим тога, потребна су нова теренска испитивања у северозападној Бугарској ради прикупљања више, стратиграфски поузданијег материјала. Изгледа да постоји добра корелација Бугарске са другим плитководним, епиконтиненталним срединама у Европи које су се крајем креде налазиле на ободу Тетиског океана. Ретки и случајни наласци цефалопода и других карактеристичних макро фосила у горњој креди стварају проблеме у Бугарској, јер не постоји зонална подела целе горње креде на основу амонита или белемнита, или неке друге макро фосилне групе. За мастрихт, углавном, постоје поделе засноване на микрофосилима. Наша проучавања су показала да се зуби ајкуле могу употребити за одредбу старости тамо где други карактеристични фосили у горњој креди Бугарске изостају.

ГЕОЛОШКИ АНАЛИ БАЛКАНСКОГА ПОЛУОСТРВА ANNALES GÉOLOGIQUES DE LA PÉNINSULE BALKANIQUE	67	65–87	БЕОГРАД, децембар 2006 BELGRADE, December 2006

Trinocladus divnae and *Montiella filipovici* – a new species (Dasycladales, green algae) from the Upper Cretaceous of the Mountain Paštrik (Mirdita Zone)

RAJKA RADOIČIĆ

Abstract. Two new dasycladalean species from the Upper Cretaceous of the Mountain Paštrik, Kukes Cretaceous Unit of the Mirdita Zone are described:

Trinocladus divnae sp. nov. is characterized by variable size of the thallus, relatively narrow main axis, typical *Trinocladus* organization of the laterals and thin calcification limited to the distal part of the thallus which includes a swollen part of secondaries and short tertiaries. Often, the internal portion of the whorls (except sometimes the main stem membrane), tends to dissolve and form dissolution cavities filled with cement.

Montiella filipovici sp. nov. is characterized by a primary skeleton made of a thin individual sheath around a fertile ampullae, often obliterated by recrystallization. Four to six laterals, each giving one secondary and one fertile ampulla located on the upper side of the relatively thick short primary lateral.

Upper Cenomanian limestone with *Cisalveolina fraasi* and *Trinocladus divnae* sp. nov. was deposited immediately before the events that resulted in sea level rising. The middle and upper Cenomanian eustatic-tectonic processes had different effects in the Paštrik shallow water areas, depending on the distance from the basinal part of the Unit. Bathymetric changes in a part of the Paštrik sedimentary area were not significant, even negligible. *Montiella filipovici* is found in the post-*fraasi* shallow water sequence, assigned to the ?uppermost Cenomanian–lowermost Turonian (= *Whiteinella archaeocretacea* Zone *p. p.*; a short stratigraphic gap, in a part of the area, is noted). Shallow water limestone with Turonian taxa, corresponding to the *helvetca* Zone, occurs a few meters upward.

Supplementary note: the species *Cylindroporella parva* RADOIČIĆ is transferred in the genus *Montiella*, the species *Permocalculus elliotti* JOHNSON is transferred in the genus *Trinocladus*, while the species *Trinocladus bellus* YU JING is transferred in the genus *Belzungia*.

Key words: Dasycladales, new species, new combination, Cenomanian, Turonian, Mountain Paštrik, Kukes Cretaceous Unit, Mirdita Zone

Абстракт. Описане су двије нове врсте дазикладалеских алги: *Trinocladus divnae* sp. nov. из ценомана и *Montiella filipovici* sp. nov. из ?највишег ценомана–најнижег турона Паштрика (Кукеска кредна јединица, Мирдита Зона):

Врсту *Trinocladus divnae* sp. nov. карактерише талус веома варијабилних димензија, узана главна оса, организација огранака *Trinocladus* типа и слаба калцификација најчешће ограничена само на дистални дио талуса: око дисталног проширења секундарних и око терцијарних огранака. Унутарња структура, изузев каткада калцифициране мембране главне осе, била је веома подложна дисолуцији. Тако настала празнина обично је бивала испуњена цементом. Векстон-пекстоне са *Trinocladus divnae* sp. nov. Потиче из највишег дијела *Cisalveolina fraasi* зоне

Врсту *Montiella filipovici* sp. nov. карактерише танак примарни карбонатни омотач само око фертилних ампула који је најчешће уништен услед прекристализације. 4–6 релативно кратких масивних примарних огранака носе по један секундарни огранак и једну фертилну ампулу смјештену навише.

Горњоценомански кречњак са *Cisalveolina fraasi* и *Trinocladus divnae* sp. nov. депонован је непосредно прије догађаја који су узроковали пораст морског нивоа.

Средњо–горњоценомански тектонско-еустатички процеси различито су се одражавали на плитководни ареал зависно од удаљености односно близине басенског дијела Кукеске кредне јединице. Батиметријске промјене у дијелу овог плитководног седиментационог простора биле су незнатне. Пост-*fraasi* плитководну секвенцу карактерише исчезавање карактеристичних ценоманских фосила, осиромашење биоте,

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теригени принос (кварц), а мјестимично и разорени слојеви. *Montiella filipovici* нађена је у кречњачком слоју ове секвенце са учесталом *Halimeda elliotti* CONARD & RIOULT испод карбонатне секвенце која је латерални еквиваленат *helvetica* зоне. Овај дио пост-*fraasi* стуба приписан је ?највишем ценоману–најнижем турону (= *Whiteinella archaeocretacea* зона *p. p.*).

У додатној биљешци дати су подаци о новим комбинацијама: врста *Cylindroporella parva* пребачена је у род *Montiella*, *Permocalculus elliotti* у род *Trinocladus*, а *Trinocladus bellus* у род *Belzungia*.

Кључне ријечи: Dasycladales, нове врсте, нове комбинације, ценоман, турон, Паштрик, Кукеска кредна јединица, Мирдита зона.

Introduction

The Cretaceous succession of the Mountain Paštrik is an attached platform (superimposed paleogeography) overlaying the Diabase Chert Formation, serpentinite and Tithonian–Berriasian carbonate clastics (flysch auct.) in the north. From the middle Cenomanian into the Turonian, it was a ramp – a transitional stage from the platform to the basin.

Dasycladales, common in some of shallow water Albian, Cenomanian and Turonian levels of this area, have been mentioned or described by PEJOVIĆ & RADOIČIĆ, (1974), CHERCHI *et al.* (1976), CONRAD *et al.* (1977) and RADOIČIĆ (1978, 1983, 1984, 1994, 1998). The present note is a further contribution to the Dasycladales in Paštrik limestone: two new species – *Trinocladus divnae* and *Montiella filipovici* are described.

Systematics

Order Dasycladales PACHER, 1931
Family *Triploporellaceae* (PIA, 1920) emend. BERGER & KAEVER, 1992
Subtribe *Triploporellinae* (PIA, 1920) emend. ASSOULLET *et al.*, 1979

Genus ***Trinocladus*** RAINERI, 1922

According to ELLIOTT (1972, p. 619), the tubular thallus of *Trinocladus* is composed of "Successive verticils of radial branches, each branch showing outwardly widening primaries giving rise to several secondaries, and these in turn to bunches of tertiaries. Branches of the lower verticils may not show the full detail. Branches usually not alternate in position from verticil to verticil."

Based only on a transversal section, RAINERI (1922) maintained that a trichotomic partitition of the laterals is characteristic of the genus. In fact, the main generic feature is the form of laterals: club-shaped phloiophorous primaries with a more or less large subspheric distal part, similar shaped thinner secondaries, four or more per each primary, and bunches of similar short tertiaries.

It should be mentioned that, in some cases, recrystallized or poorly preserved *Trinocladus* tubes were also ascribed to *Permocalculus* or to *Griphoporella*.

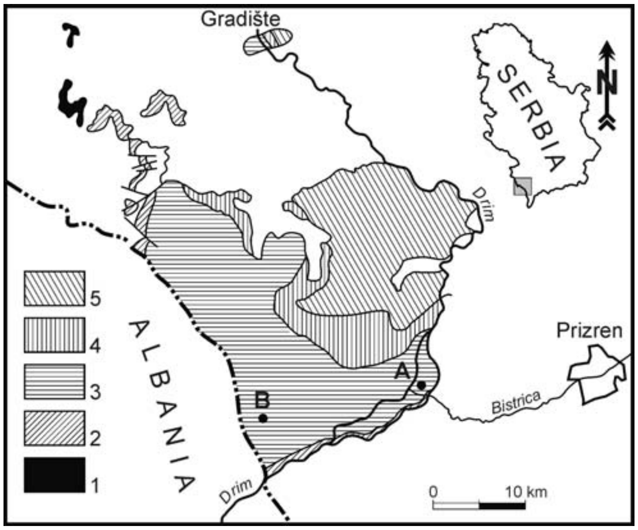

Fig. 1. Geological map of the Paštrik area, according to Geological map 1:000000, sheet Prizren (MENKOVIĆ *et al.*, 1982), simplified (new stratigraphic data are not included). **1**, ?Barremian–Aptian, carbonate breccia and bioclastic limestone; **2**, shallow water Albian sediments, **3**, shallow water lower into upper Cenomanian; **4**, shallow water Turonian carbonates with rudists and hemipelagic limestone; **5**, Senonian, breccia, microbreccia, calcarenites and marly limestone with planktonics. New stratigraphic data are not included. **A**, Bistrica section; **B**, – Vrbnica 1 section.

Specimens of such preservation are often really difficult to distinguish from some *Permocalculus*. Compare: fragments of *T. divnae* in Pl. 3, Figs. 10, 11, with "typical *Permocalculus* debris" illustrated by JOHNSON (1969, pl. 17) and by JOHNSON *in* JOHNSON & KASKA (1965, pl. 14). In the same paper, JOHNSON introduced a new species *Permocalculus elliotti,* which is, in fact, *Trinocladus* (see further text).

Among the species ascribed to the genus *Trinocladus,* there are those which do not have a branching pattern characteristic to *Trinocladus*. Only Paleogene species from China, *Trinocladus bellus* JU YING, 1976, which has a *Belzungia* type of arrangement of the laterals will be mentioned (see farther text).

Trinocladus divnae sp. nov.
Pl. 1, Figs. 1–14; Pl. 2, Figs. 1–8; Pl. 3, Figs. 1–6

Origin of name. The species is dedicated to my friend and colleague Dr. DIVNA JOVANOVIĆ (Belgrade) for her contribution to the study of depositional environments of the Late Paleozoic in northwestern Serbia.

Holotype. Slightly oblique transversal section of the calcareous tube shown in Fig. 2 (= Pl. 1, Fig. 1), thin slide RR2379, sample 013577, author's collection deposited in the Geological Institute, Belgrade.

This section shows the thinly calcified central stem membrane with well preserved insertion points of 4 laterals. The central stem, the irregular space around it and between the laterals were early post mortem filled with matrix. Open pores on the calcareous tube surface (dentate surface) is the evidence of the tertiaries. Primary and secondary laterals are not preserved predominantly due to dissolution in the post-filling phase with matrix. The two primaries are not completely obliterated by recrystalization (arrows). Dissolution cavities were subsequently filled with cement. This space, corresponding to the space of the laterals R1-R3 has an inverted triangular form (Fig. 2, arrows).

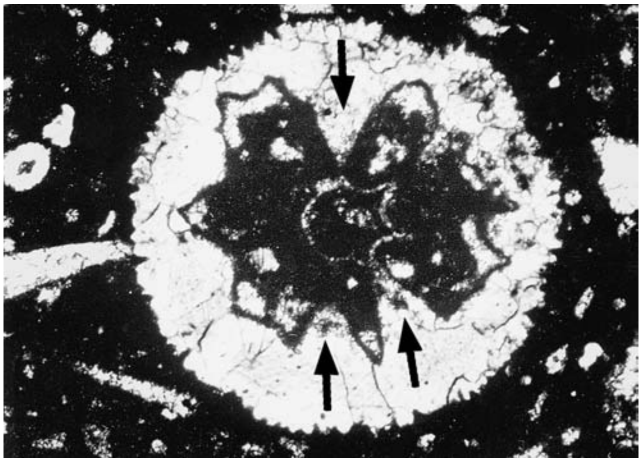

Fig. 2. *Trinocladus divnae* sp. nov. Holotype, slightly oblique section; thin slide RR2379, × 61,5. Notice the insertion points of 4 laterals. Arrows: recrystallized space of inverted triangular forms corresponding to units of laterals (R1-R3), and on the bellow the two pores of primaries which are not completely obliterated by recrystalization.

Isotypes. Specimens in thin slides RR2379 and 2380. Some of them illustrated in Pl 1, Figs. 2, 3, 9, 13; Pl. 2, Figs. 2, 9; Pl. 3, Figs. 2, 3, 4.

Other materials. Specimens in thin slides RR2448-2351 and RR2343-2346a, samples 015126 and 015130 collected in 1973 near Miljaj

Type locality and type level. Southeastern slope of Mountain Paštrik, Graždanik–Bistrica–Dobrušte belt, Geological map SFRJ 1:100 000, Sheet Prizren. The sample 013577 was collected in 1972 from the top part of the *Cisalveolina fraasi* Zone in the Bistrica section north of the Bistrica River, east of the Drim River (Fig. 1).

The Albian–Cenomanian succession of the Bistrica lies on Upper Triassic limestone (= large block sliding, during the Cimmerian event in the Diabase Chert Formation) and on the Diabase Chert Formation (RADOIČIĆ, 1994).

In the Bistrica column, about 220–230 meters thick, *Salpingoporella milovanovici* RADOIĆIĆ (last occurrence) and *Suppiluliumaella schroederi* BARATTOLO (Pl. 7, Figs. 2–4) occur in the level at about 180 meters. Upward from 200 meters, the foraminiferal assemblage is notably impoverished (*Cuneolina*, *Nezzazatinella*, miliolids, a few *Pseudolituonella reicheli* MARIE, *Chrysalidina gradata* D'ORBIGNY and *Pseudocyclammina)*; the new datum is the first occurrence of the "*Bry*2" – bryozoan species known only from the *Cisalveolina fraasi* Zone. *Cisalveolina fraasi* REICHEL was found in the 210th meter. Some beds of the *fraasi* Zone abound in hydrozoans (monospecific assemblage, RADOIČIĆ, 1994, pl. 2, fig. 4). Frequent sponge spicules, *Pieninia oblonga*, bryozoan zoeciums, rare *Heteroporella lepina* PRATURLON, *Montiella parva* (RADOIĆIĆ), microproblematica "*Pr10*", a few corals, gastropods, moluscan fragments, *Neithea* and *Eoradiolites* are also present. Sediments younger than the Cenomanian are outcropping in the western Drim riverside (Western Našec section, RADOIČIĆ, 1994).

Diagnosis. Thallus cylindrical, central stem narrow with whorls consisting of three orders of similarly shaped phloiophorous laterals; 7–8 primary laterals bear probably six relatively long secondaries, giving rise to bunches of fine short tertiaries. Distal widening of the laterals relatively small, maximum of 0,076 mm in the primaries, about 0,050 mm in the secondaries and about 0.025 mm in the tertiaries. Length of the primary lateral is almost equal to both secondary and tertiary laterals. Primary calcification delicate, prevailing around the distal portion of the thallus.

Dimensions (in mm) (extreme in the brackets):

Outer diameter 0.710–1.640 (1.947).

Main stem diameter (given on a few sections only) 0.126–0.177.

Length of primary laterals (= pores) nearly half the wall thickness.

Distance between whorls about 0.100.

Description. *Trinocladus divnae* sp. nov. in the Paštrik material is quite differently preserved. The primary feature of some specimens are completely or prevailing obliterated by recrystallization (Pl. 1, Figs. 4, 7, 10, 11, 14; Pl. 3, Figs. 1–4). Open pores on many of the corpuscule surface (dentate surface) correspond to external moulds of parts of the tertiaries (Pl. 1); their distal unpreserved part form a cortex. In a few sections, some pores, swelling of secondaries or rarely of primaries, are visible (Pl. 1, Figs. 5, 8, 10; Pl. 3, Figs. 5, 6).

Sometimes, the contour of the main stem is also recognizable (Pl. 1, Figs. 5, 9; Pl. 3, Figs. 2, 4), or, even, the calcified stem membrane is preserved as a thin black line (Pl. 1, Fig. 7). The central stem and proximal whorl area, in a few individuals, occur as a cavity with the internal wall surface more or less zig-zag undulated (Pl. 1, Figs. 6, 9, 10).

The internal structure of the specimen shown in Pl. 1, Fig. 9 seems, at a first glance, relatively well preserved. The structure of the whorls, in this case, is obliterated: only contours of the recrystallized whorl areas and the intervening space can be distinguished.

Here, subcircular sections give a wrong impression that they are sections through thick laterals. Most likely, the membrane of the primary and secondary laterals are not at all or only slightly calcified. They were enveloped in a thick mucilage layer giving a cuplike form to every unit of laterals (triangular in the sections), which are, in this case, completely recrystallized.

Individuals with a calcified distal thallus area, around the swelling of secondaries and of dense short tertiaries, are presented as thin-walled cylindrical calcareous tubes, relatively resistant to dissolution and abrasion (Pl. 2, Figs. 1–5). The small sized fragments of this thin fine-porous calcified wall is difficult to recognize as dasycladalean skeleton elements (Pl. 3, Figs. 10, 11).

Comparing these differently preserved specimens, I could not immediately decide: whether they were the same species. Specimens with a thin distal calcification are ascribed to *Trinocladus divnae* because of the somewhat larger size of the thallus and, especially, the difference in the preservation are not of specific maximal values. Smaller individuals generally had an early-altered internal structure (recrystallized) and were more resistant to break. The thin calcareous wall of larger specimens, although resistant to dissolution and abrasion, were more friable.

Relations. This species resembles *Trinocladus tripolitanus* RAINERI the most, which has the same number of primary laterals but is not so variable in the size of the thallus. The main difference lies in the distal part of the thallus: relatively delicate secondaries of *Trinocladus divnae* bear dense bunches of fine tertiaries, forming a resistant thin calcified wall.

Trinocladus sp. from the Maastrichtian of Iraq (RADOIČIĆ, 1979, pl. 2, fig. 4) now is tentatively referred to *Trinocladus divnae*.

It does not exclude that the alga presented as "*Griphoporella* sp." by SCHLAGINTWEIT, 1992, pl. 1, fig. 7 is a thin-walled specimen of *Trinocladus divnae*.

Family *Dasycladaceae* (KUTZING, 1843) emend. BERGER & KAEVER, 1992
Tribe *Dasycladeae* PIA, 1920

Genus ***Montiella*** MORELLET & MORELLET, 1922

The genus *Montiella* is characterized by a simple cylindrical thallus consisting of whorls with two orders of laterals. The primary laterals arranged in quincunxes bear one fertile ampulla and one secondary lateral enlarged distally.

Type species is *Montiella munieri*, from the Montian of Belgium. Other species of the genus, *Montiella macropora*, was contemporary described from the Thanetian of the Paris Basin. Isolated specimens and fragments of the both species were studied by GÉNOT (1978, 1987, *in*: DELOFFRE & GÉNOT, 1982). This author (1982, p. 108) compares the calcareous sleeve of both species with those in some *Neomeris* "chés lesquelles les ramifications primaries et l'extrémité proximale des ramifications secondaires ne sont jamais conservées."

The find of genus *Montiella* in the Cretaceous sediments is of a later datum: when *Cylindroporella elitzae* BAKALOVA and *Cylindroporella benizarensis* FOURCADE *et al*. were transferred to the genus *Montiella* (RADOIČIĆ, 1980, GRANIER, 1990). The difference in the extent of calcification of the Paleocene and Cretaceous *Montiellae* is readily evident. In contrast to the Paleocene species, the known Cretaceous species have a calcareous sheath around the proximal part of the whorl with, in the some specimens, a well preserved morphology of the central stem.

In the calcareous sheath of the known Cretaceous *Montiellae*, as a rule, the pore of the secondary is not differentiated from the pore of the primary lateral (secondary effect). They look like a single pore: a thick and short proximal part with a fertile ampulla followed by somewhat narrower tube, distally enlarged having the protective function of the ampulla.

Besides the fertile ampullae, other whorl elements are often not or only partially preserved. Therefore, it is often difficult to distinguish the calcareous sleeves of *Montiella* from those of *Cylindroporella*. Bearing in mind that the type species of the genus *Cylindroporella* is poorly preserved (some structural elements are obliterated), the question is: what is *Cylindroporella?* The *Cylindroporella* problem is discussed by BARATTOLO and PARENTE (2000).

Montiella filipovici sp. nov.
Pl. 4, Figs. 1–9; Pl. 5, Figs. 1–12; Pl. 6, Figs. 1–4

Origin of name. The species is dedicated to my friend and colleague Dr. Ivan Filipović (Belgrade), for his contribution to the study of on the Paleozoic in Western Serbia.

Holotype. Oblique section of the specimen in Pl. 4, Fig. 1, thin slide RR2328, sample 015117, author's collection deposited in the Geological Institute, Belgrade.

Isotypes. Different sections in thin slides RR2326 to 2336, partly illustrated in Pl. 1, Figs. 2–9; Pl. 3, Figs. 1–12; Pl. 3, Figs. 1–4.

Type level and locality. Sample 015117 was collected in 1973 from the upper part of the section Vrbnica 1, southern slope of the Mountain Paštrik. This section is exposed on the footpath between Vrbnica–Drim and Miljaj–Ninaj, east of the section Vrbnica 2 (CHERCHI *et al*., 1976); Geological map SFRJ 1:100000, Sheet Prizren (Fig. 1).

Lateral equivalents of the sediments with *Cisalveolina fraasi,* in the Vrbnica 1 section, are followed by some ten meters of limestone with a prevailing weak terrigenous influx, changed and impoverished biota. A few meters upward, limestone abounding in halimedacean algae was sampled (15117). This bed, grainstone-packstone dominated by *Halimeda elliotti* Conard & Rioult, *Halimeda* sp. and some other halimedacean algae is the type level of *Montiella filipovici* sp. nov. The association also contains *Montiella parva* (Radoičić), *Terquemella* sp., sparse fragments of *Neomeris* and *Heteroporella lepina* Praturlon, a few foraminifera *Nezzazatinella*, *Cuneolina*, *Reticulinella*, miliolids, rare ostracodes and metazoan fragments (Pl. 6, Figs. 5, 6, 8–11).

Further upwards (the bedding is not well visible), the lower Turonian (equivalent to the *helvetica* Zone) is documented by *Moncharmontia apenninica* (De Castro), *Pseudocyclammina sphaeroidea* Gendrot (sample 015119), and the rudists *Hippurites, Durania*, *Biradiolites* and *Distefanella*. The interval between the *Cisalveolina fraasi* Zone and the lower Turonian is equivalent or partly equivalent to the *Whiteinella archaeocretacea* Zone

In the Gradište succession (the same Cretaceous Unit, Fig. 1), *Halimeda elliotti* occurs abundandly in some beds of the Hemipelagic Sequence (= *Whitinella archaeocretacea* and *Helvetotruncana helvetica* zones). In the Metohija Cretaceous Unit, the abundance of *Halimeda elliotti* associated with *Helvetotruncana helvetica* is known from the Zabel Section (Radoičić, 1993, 1998).

Diagnosis. Thallus with a narrow central stem and whorls commonly consisting of 6 laterals, exceptionally 4 or 5. Primary laterals arranged in quincunxes, short and relatively thick, each bears a fertile ampulla and a secondary lateral. Egg shaped slightly inclined upward fertile ampulla located on the upper side of the primary lateral close to the central stem, a secondary lateral grows from its distal end. Ampulla pedunculus is short, usually not clearly differentiated.

Walls of the fertile ampullae have been individually calcified. This primary calcareous skeleton is altered or partly altered.

Dimensions (in mm) (extreme value in brackets):

External diameter 0.607–0.708 (0.759); the transverse section with 5 laterals in the whorl shown in Pl. 5, Fig. 5 is a specimen with a narrower thallus diameter – 0.430 mm.

Diameter of the main stem (0.075) 0.101 – 0.151 (0.177).

Length of the ampulla with pedunculus 0.180.

Length of the primary laterals up to 0.127.

Diameter of the fertile ampulla up to 0.170.

Diameter of the primary lateral about 0.051.

Description. The surface of the calcareous tubes was more or less eroded, often to half of the fertile ampullae (Pl. 4, Fig. 9). The fertile ampullae primary have been individually calcified as more or less thin carbonate envelopes (about 0.002 mm). A similar primary calcification is not observed in the laterals, they were poorly preserved, probably due to weak or no calcification at all. The form of their distal-cortical part is not known (it seems they were much enlarged). The best example of individual calcified fertile ampullae is the transverse section in Pl. 2, Fig. 5. Some other sections, with a preserved individual sheath around the ampullae, and primary contact between them are illustrated in the same plate. In the same calcareous tube, parts of the skeleton may be differently preserved. An example of the different grade of the obliterated structure in the same whorl is the transversal section in Pl. 2, Fig. 4: well preserved, slight contact between the ampullae observable as a black line, and both, the sheath and the space between the ampullae are almost obliterated by advanced recrystallization in the other part of the section. The mentioned transverse section with 5 laterals is also an example of gradual alteration. The original sheath around the ampullae is preserved in part of this section, and obliterated by recrystallization in the other part.

Relations. *Montiella elitzae* and the very similar *Montiella benizarensis* are species with a larger thallus bearing 6–8 laterals per whorl and more variable dimensions than *M. filipovici* characterized by slightly variable thallus dimensions.

Some specimens of *Cylindroporella elitzae* and *Cylindroporella benizarensis* (are these two species?) are a nice examples of a post-mortem process resulting in axis widening at the whorl level (Radoičić *et al*., 2005, pl 1, fig. 3). The primary calcification around the fertile ampullae in these species is not preserved, except the thin calcification around fertile ampulla of *Montiella elitzae* from Eastern Serbia (Radoičić, 1980, pl. 2, fig. 4), which indicates the same primary calcification as that of *Montiella filipovici*.

A list of algal flora in the Cretaceous of Paštrik

(in alphabetic order):

Acroporella radoicicae Praturlon, Pl. 8, Fig. 3
Bacinella irregularis Radoičić
Charophyta
Coptocampylodon fontis (Patrulius)
Coptocampylodon sp.
Clypeina pastriki Radoičić
Cylindroporella sp. div.
Dissocladella?, Pl. 8, Fig. 5
Halimeda elliotti Conard & Rioult, Pl. 6, Figs. 8–10
Halimeda sp. (spec. nov.?), Pl. 6, Fig. 12
Halimedaceae sp. div.
Heteroporella lepina Praturlon, Pl. 7, Fig. 10
Koskinobulina socialis Cherchi & Schroeder
Lithocodium aggregatum Elliott
Lithocodioidea, different species
Marinella lugeoni Pfender, Pl. 7, Fig. 9
Montiella filipovici sp. nov., Pl. 1, Pl. 2, Pl. 3, Figs. 1–6
Montiella parva (Radoičić), Pl. 4, Pl. 5, Pl. 6, Figs. 1–4
Neomeris (Drimella) drimi Radoičić
Neomeris sp., Pl. 2, Fig. 9
Neomeridae, subgenus?
?*Pseudoclypeina*, Pl. 7, Fig. 6
Pseudolikanella cf. *danilovae* (Radoićić)

Pseudolithotamnium album (PFENDER), Pl. 7, Fig. 7
Salpingoporella hasi CONRAD, RADOIČIĆ & PEYBERNES
Salpingoporella milovanovici RADOIČIĆ
Salpngoporella pygmaea (GÜMBEL)
Salpingoporella turgida (RADOIČIĆ)
Suppiluliumaella schroederi BARATTOLO, Pl. 7, Figs. 2–4
Solenoporaceae
Suppiluliumaella sp., Pl. 8, Fig. 1
Terquemella div. sp.
Trinocladus divnae sp. nov., Pl. 1, Pl. 2, Pl. 3, Figs. 1– 6
Trinocladus aff. *tripolitanus* RAINERI, Pl. 3, Figs. 8, 9
Trinocladus?, Pl. 3, Fig. 7
Triploporella sp.
Digfferent microbial epiliths

Supplementary note

Montiella parva (RADOIČIĆ, 1983) comb. nov.

Cylidroporella parva n. sp., RADOIČIĆ, 1983: pl. 1, figs. 1–5; pl. 2, figs. 1–2; thin slide RR3557, author's collection, Geological Institute of Serbia.

Turonian, Tripolitania, Libya,

The taxon is found in the type level of *Montiella filipovici* and also in association with *Trinocladus divnae* (Pl. 6, Figs. 5–7).The subaxial section, Fig. 7, is the only specimen of this species with preserved pores of primary laterals upward bearing the fertile ampulla (the structure of the genus *Montiella*). The secondary laterals most probably were not calcified.

From the Turonian of Sinai in Egypt, the species was presented by IMAM (1996, 1b, not 1a). A primary calcification of this specimen, the individual sheaths enclosing the fertile ampullae which is the feature of the genus *Montiella,* is well preserved in the part of this recrystallized body.

Trinocladus elliotti (JOHNSON, 1965) comb. nov.

Permocalculus elliotti n. sp., JOHNSON (*in* JOHNSON & KASKA) 1965: pl. 5, figs. 1–5; thin slide 18587, (USNM in Washington, Division of Paleobotany, no 42340), earliest Early Cretaceous, possibly Late Jurassic, Rosario area of Spanish Honduras.

Although the calcareous sheath of the illustrated specimens is diagenetically altered, sections in figures 1–3 give sufficient data on the dasycladalean nature of this species. Branching arrangement – primary, secondary (Figs. 2, 3, on left) and tertiary (Fig. 3) laterals is of the *Trinocladus* pattern. Accordingly, the species is transferred into the genus *Trinocladus* RAINERI, 1922. In order to obtain a diagnosis, a study of the type material is necessary.

Belzungia bella (YU JING, 1976) comb. nov.

Trinocladus bellus. spec. nov., YU JING, 1978: pl. 8, figs. 10 (thin slide 28434) and 11 (thin slide 28435), ?Fig. 9, non Fig. 12, Paleocene–Ypresian of Lungma Region, China

The holotype of *Trinocladus bellus* is a large fragment of the calcareous tube – longitudinal-oblique section through 6 or 7 whorls. The insertion points of the primary laterals in this section are not preserved because the main stem is secondary enlarged. Thick short primaries, thick irregular secondaries and somewhat thinner tertiaries, give rise to further laterals, thin and anarchically arranged. Whorls bearing such arrangement of laterals characterize the genus *Belzungia,* MORELLET, 1908. The new combination refers to two out of four sections illustrated by YU JING (1978, fig. 10), holotype and transversal section in Fig. 12. This species, introduced on insufficient material, is different from other *Belzungia* species by coarser proximal (3 orders) and seemingly somewhat more anarchically arranged distal laterals.

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Резиме

Trinocladus divnae и *Montiella filipovici* – нове врсте (Dasycladales, зелене алге) из горње креде Паштрика (Мирдита зона)

Из горњокредних седимената Паштрика (Кукеска кредна јединица, Мирдита зона) описане су двије нове врсте дазикладалеса:

Triploporella divnae sp. nov.

Холотип. Мало искошен попречан пресјек приказан на табли 1, сл. 1.

Дијагноза. Цилиндричан талус са узаном централном стабљиком која носи, на растојању, пршљенове са три реда огранака сличног облика. 7–8 примарних огранака дају највјероватније по сест секундарених, а ови по сноп кратких терцијарних огранака. Дужина оба, секундарног и терцијарног, приближно је једнака дужини примарног огранка. Калцифициран је често само танак дистални дио талуса (врх секундарних и терцијани огранци), ријетко и мембрана главне стабљике. Секундарне промјене знатне.

Стратиграфски положај и типски локалитет. Горњи (али не најгорњи ценоман), виши дио зоне са *Cisalveolina fraasi* у профилу Бистрице на југоисточним падинама Паштрика.

Montiella filipovici sp. nov.

Холотип. Кос лонгитудинални пресјек приказан на табли 4, сл. 1.

Дијагноза. Талус са узаном централном стабљиком и пршљеновима који обично имају 6, изузетно 4 или 5 огранака. Кратки релативно масивни примарни огранци носе фертилну ампулу и један секундарни дистално проширен огранак. Фертилна ампула је смјештена на гоњојој страни огранка близу централне стабљике, педункулус ампуле веома кратак, нејасно диференциран. Танак кречњачки омотач депонован је појединачно око фертилних ампула.

Примарни, а особито секундарни огранци били су слабо или нијесу уопште били калцифицирани.

Битне примарне црте рода *Montiella* прекристализацијом бивају изгубљене. Стога се такве кречњачке цјевчице могу погрешно приписати роду *Cylindroporella*, премда до данас, с обзиром на лошу очуваност типског материјала, није дефинисано што су битне одлике овог рода.

Montiella filipovici потиче из седимената највишег ценомана–најнижег турона који леже непосредно испод кречњака са туронским микрофосилима и рудистима, а откривени су у профилу Врбница 1 на јужним падинама Паштрика, између Миљаја и Нинаја.

У додатној биљешци дати су укратко подаци о новим комбинацијама за врсте *Cylindroporella parva* RADOICIC, 1983, која је преведена у род *Montiella* MORELLET & MORELLET, 1922, *Permocalculus elliotti* JOHNSON, 1965 која је пребачена у род *Trinocladus* RAINERI, 1922 и *Trinocladus bellus* YU JING, 1978, која је пребачена у pод *Belzungia* MORELLET, 1908.

PLATE 1

Figs. 1–14. *Trinocladus divnae* spec. nov.

1. Holotype, slightly oblique transversal section also shown in text-fig. 2; arrows: the inverted triangular form corresponds to the space of the laterals; thin slide RR2379; × 33.

2, 3, 6, 12. Oblique (2, 6, 12) and transversal (3) section. Specimens of the same preservation; thin slides RR2379, 2379, 2338, 2338; × 33.

5. Transversal section of the damaged specimen; notice poorly preserved primary latera (arrow); thin slide RR2344; × 53.

7. Oblique section (fragment). Pores of the primary and the secondary laterals have been completelly obliterated by recrystallization, thin central stem membrane with insertion points of worls is well preserved; thin slide RR2336; × 33.

8. Oblique section of recrystallized specimen with 3 deformed pores of primaries. Dense pores of tertiaries are visible at the top of the section; thin slide RR2338; × 33.

9. Oblique section of the strangely preserved specimen (see text); thin slide RR2379; × 33.

10. Slightly oblique transversal section with rare primary pores; thin slide RR2346/1; × 33.

11, 14. Recrystallized specimens, transversal sections: thin slide RR2344; × 33.

13. Oblique section of damaged specimen, the preservation similar to those in fig. 9, thin slide RR2378; × 33.


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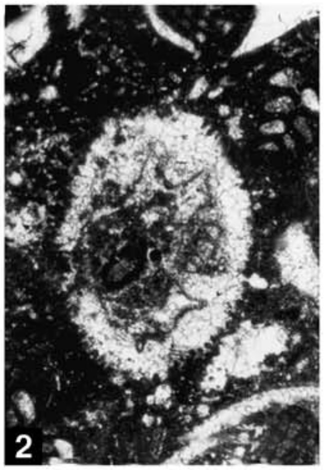
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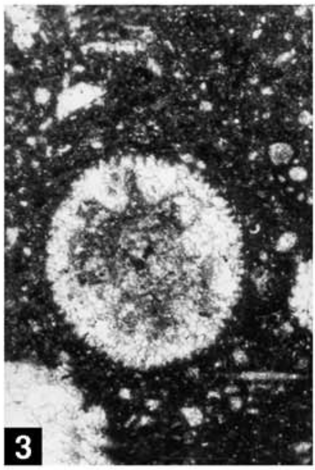
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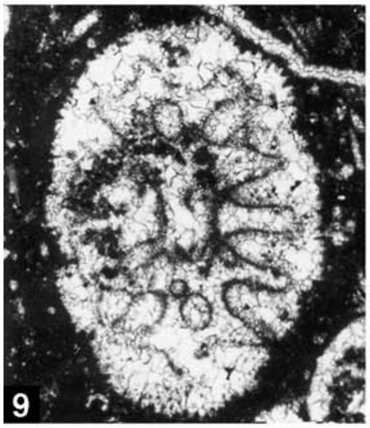
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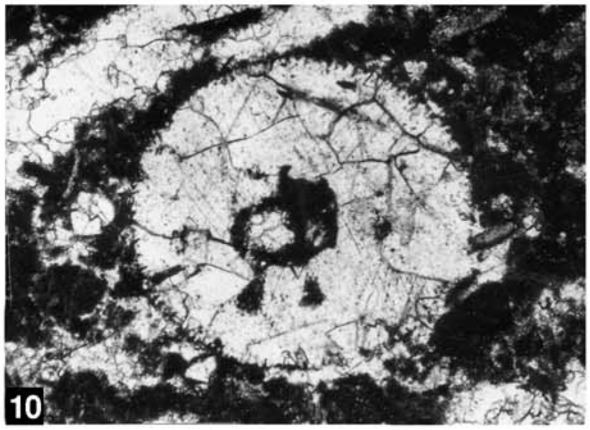
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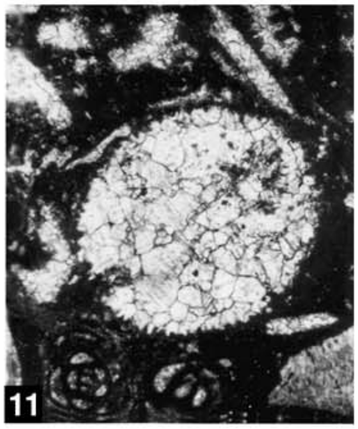
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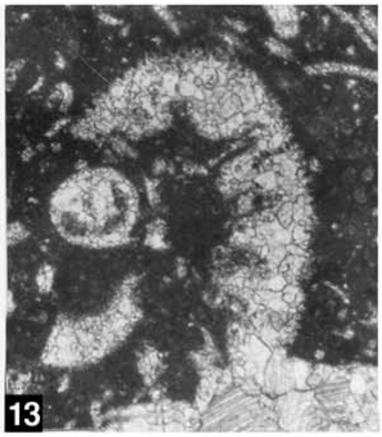
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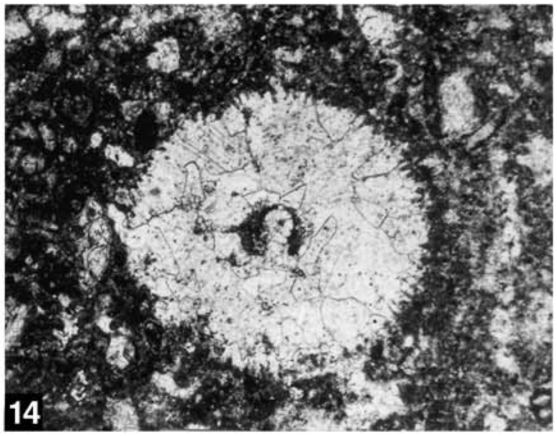
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PLATE 2

Figs. 1–8. *Trinocladus divnae* spec. nov.

1, 4, 5, 7. Oblique and transversal sections of the specimens with calcified more or less thin distral area Fig. 1 is the largest specimen of the collection; in its middle tangential section of the fragment with pores of swollen parts of secondaries; thin slides RR2344, 2346, 2344/1, 2379; × 33.

2, 3. Fragments of longitudinal section; swollen parts of the secondaries are well preserved; thin slides RR2380, 2343/1; × 33.

6. Damaged specimens of partially recrystallized interior, some pores of seconradies recognizable; thin slide RR2343/1; × 33.

8. Fragment of prevailing recrystallized specimen with two primary pores and open pores of tertiaries at the outer surface; thin slide RR2344/1; × 52.

Fig. 9. *Neomeris* sp.. Oblique section, thin slide RR2345; × 20.


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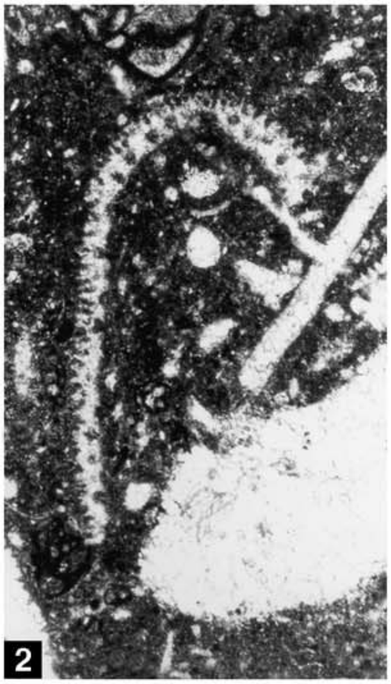
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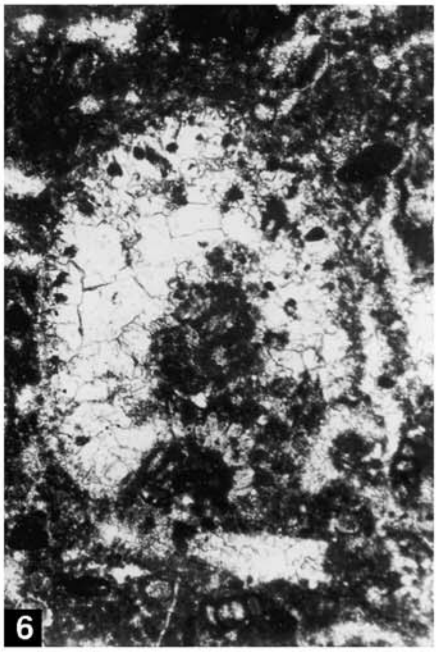
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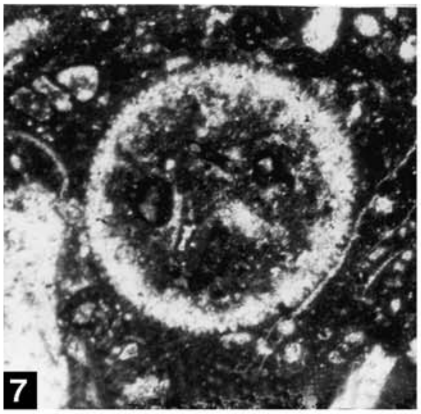
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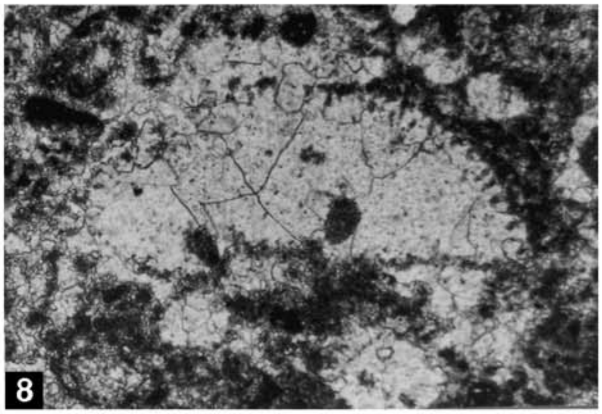
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PLATE 3

Figs. 1–6. *Trinocladus divnae* spec. nov.

1–4. Oblique (1, 2, 4) and transversal (3) section. Specimens obliterated by recrystallization; the contours of central stem, in the specimens 2 and 3, are recognizable, slightly deformed central stem of the specimen 1 is filled by micrite; thin slides RR2345, 2380, 2379, 2379; × 33.

5. Fragment of tangential section with pores of secondary laterals, thin slide RR2343/1; × 33.

6. Slightly oblique transversal section of recrystallized specimen with rare, pores of secondary laterals; thin slide RR2343/1; × 33.

Fig. 7. *Trinocladus* sp. Thin slide RR2380; × 52.

Figs. 8, 9. *Trinocladus* aff. *tripolitanu*s RAINERI. Fragmnts; thin slides RR2344 and RR 2346; × 52

Figs. 10, 11. *Trinocladus divnae* spec. nov. Fragments of calcified thin distal part, thin slides RR2348, 2346/1; × 30.


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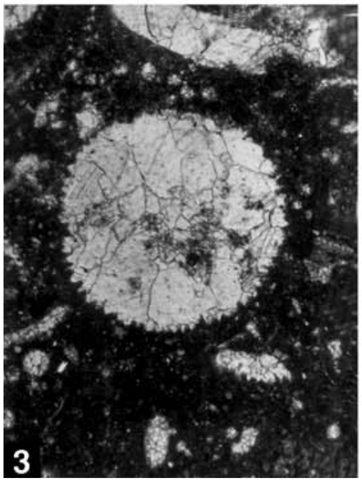
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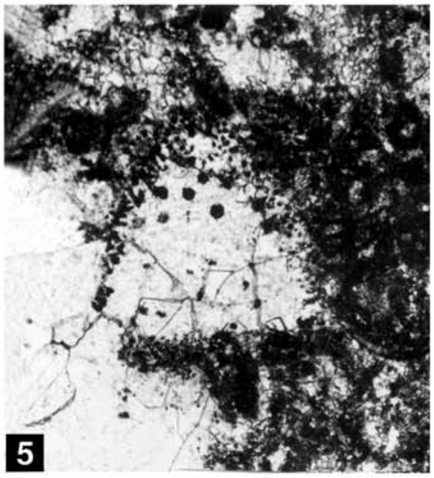
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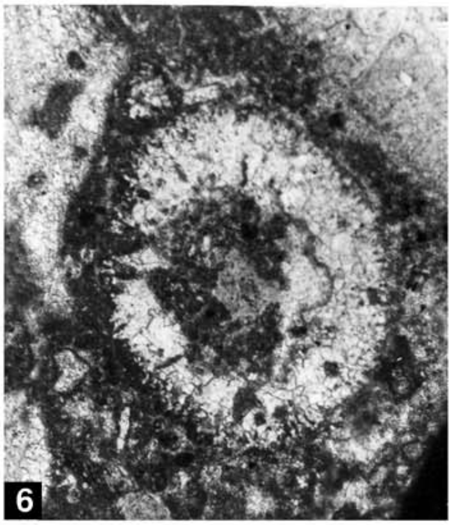
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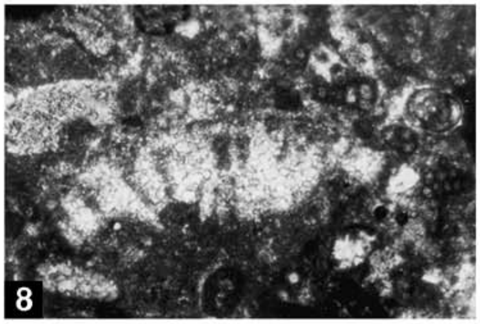
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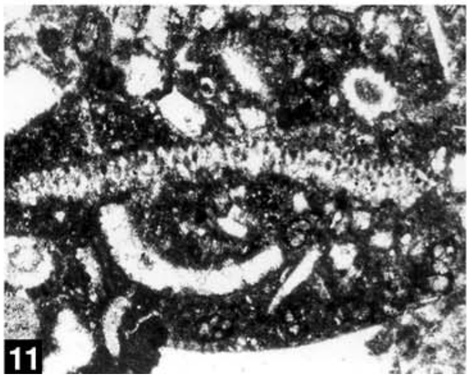
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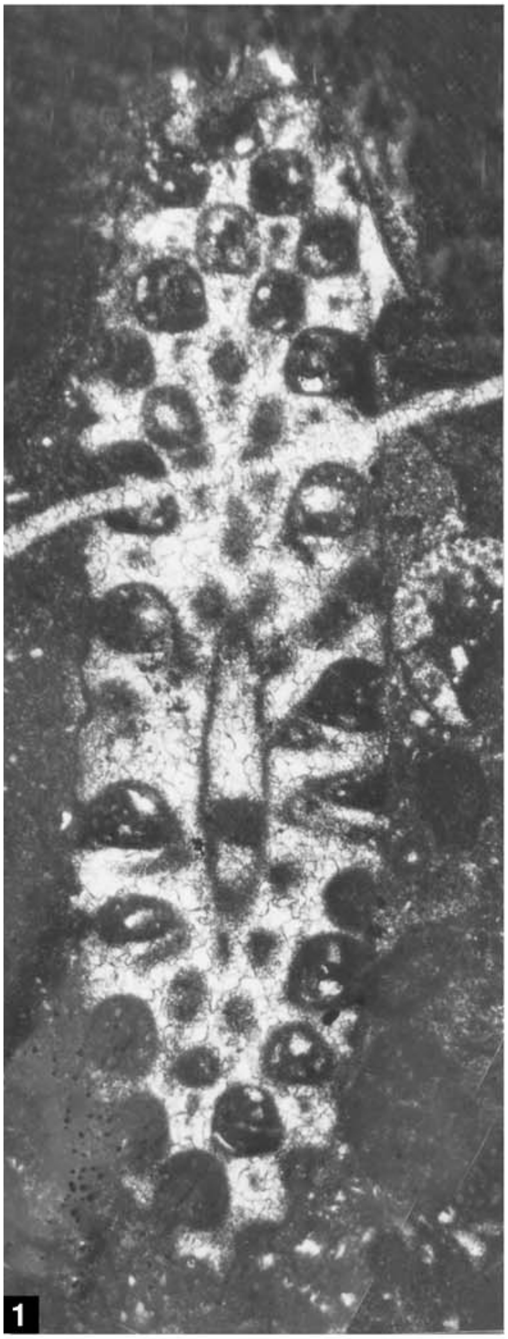
PLATE 4

Figs. 1–9. *Montiella filipovici* spec. nov.

1. Holotype, sub-longitudinal oblique section; thin slide RR2328; × 60.

2–8. different more or less oblique sections; thin slides RR2334; 2335/2, 2330, 2335, 2327, 2326/1, 2326; 2 = × 40; 6, 7, 8 = × 60; 4, 5 = × 63.

9. slightly oblique longitudinal section; thin slide RR2328; × 40.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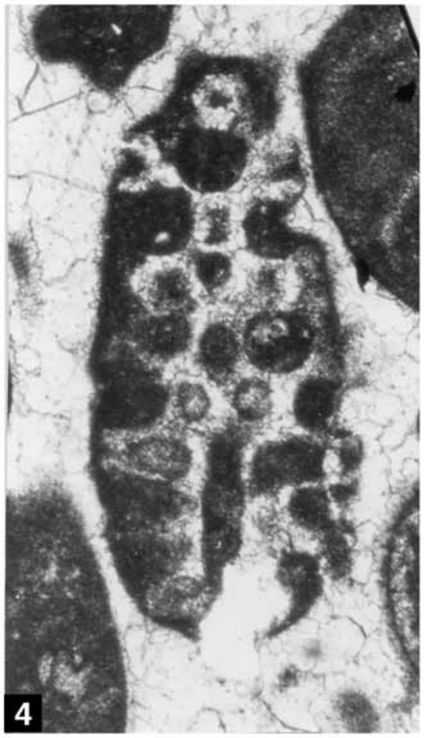
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PLATE 5

Figs. 1–12. *Montiella filipovici* spec. nov.

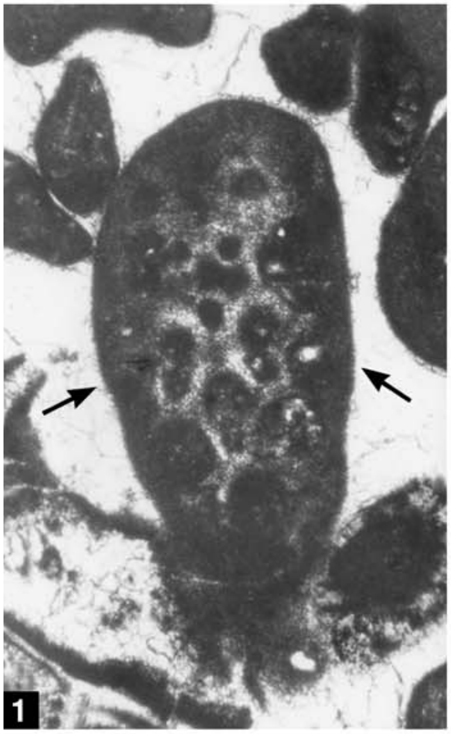
1. Tangential section. Notice characteristic section through primary laterals and ampulla (arrows); thin slide RR2327; × 63.

2, 3. Oblique sections; thin slides RR2335, 2326/1; × 63.

4–10. Transversal sections; thin slides RR2327/1, 2332, 2332, 2326/1, 2327, 2326, 2335/2; × 60; 9 = × 63.

11. Tangential-oblique section, fragment; thin slide RR2331; × 63.

12. Transversal-oblique section of damaged fragmen; thin slide RR2331; × 60.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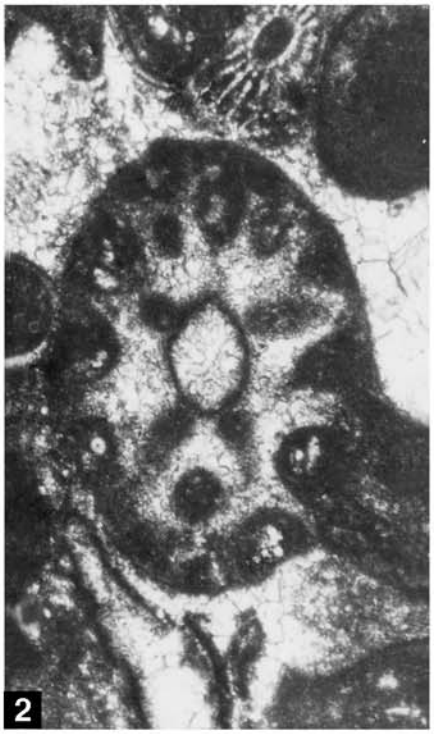
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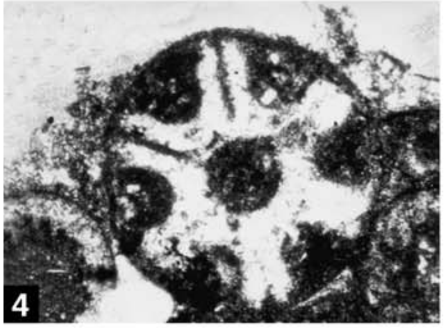
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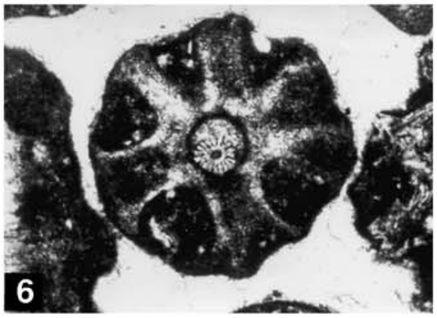
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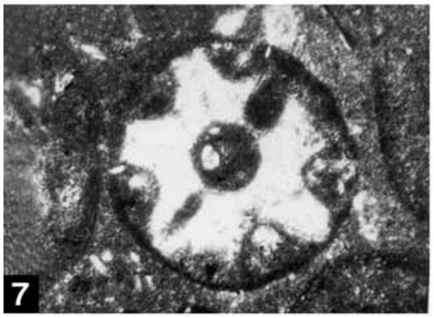
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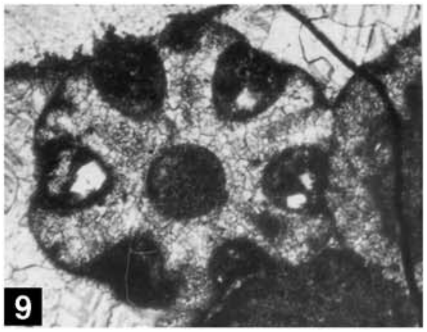
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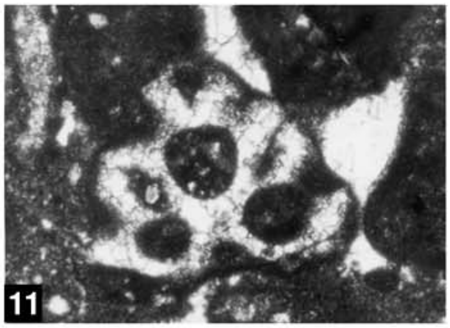
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PLATE 6

Figs. 1–4. *Montiella filipovici* spec. nov. Different oblique sections; thin slides RR2327, 2333, 2333, 2331; × 60; 4 = × 63.

Figs. 5–7. *Montiella parva* (RADOIČIĆ). Longitudinal (5, 6) and oblique section (7); thin slides RR2336, 2327 (type level of *M. filipovici*), 2379 (type level of *Trinocladus divnae*); × 63.

Figs. 8–10. *Halimeda elliotti* CONARD & RIOULT. Longitudinal, oblique and transversal section, thin slides RR2330, 2328, 2236/1; 8 = × 40; 9, 10 = × 45.

Fig. 11. *Neomeris* sp. Fragment; thin slide RR2331; × 60.

Fig. 12. *Halimeda* sp. (spec. nov.?). Thin slide RR2332; × 34.


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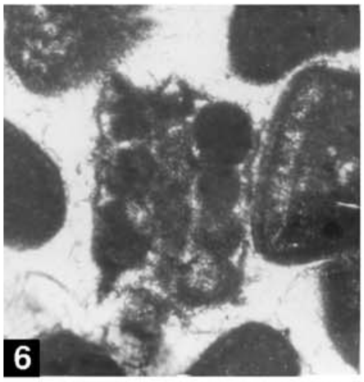
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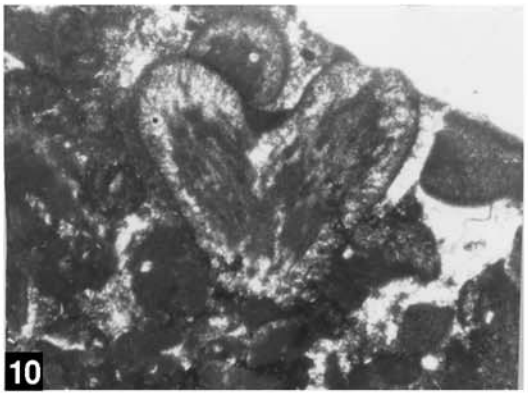
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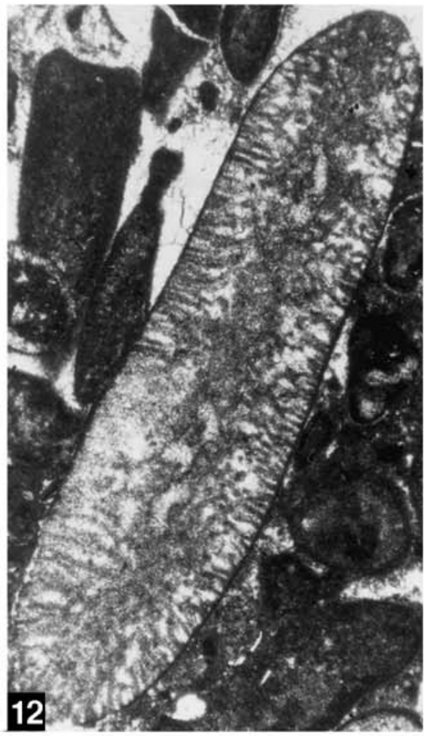
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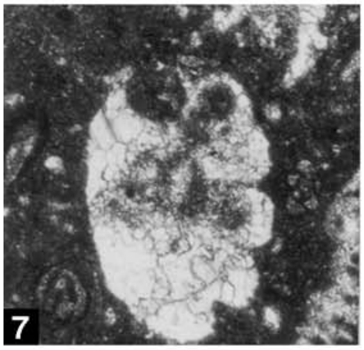
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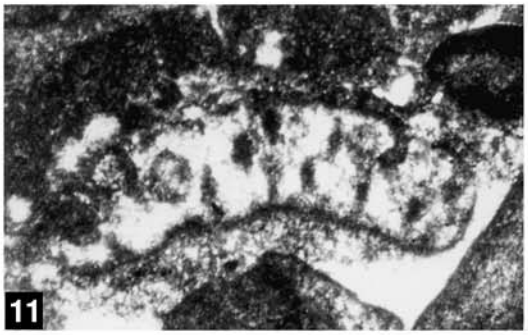
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PLATE 7

Figs. 1. *Salpingoporella hasi* CONRAD, RADOIČIĆ & REY. Microfacies; thin side RR2389; × 37.

Figs. 2–4. *Suppililuimaella schroederi* BARATTOLO (until recently known only from Apennines – BARATTOLO, 1984); thin slides RR2381, 2381/1; × 33.

Fig. 5. Problematic microfossil *"Pr-10"* and fragment of *Neomeris*; thin slide RR2345; × 44.

Fig. 6. *Pseudoclypenia*?. Fragment; thin slide RR6015 (*Valdanchella dercourti* Zone); × 67.

Fig. 7. *Pseudolithothamnium album* (PFENDER) (monospecific assemblage); thin slide RR23471/1; × 10.

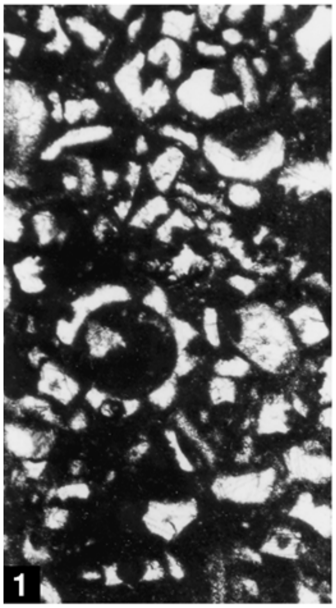
Fig. 8. Sponge spicules, type level of *Trinpocladus divnae*; thin slide RR2379; × 33.

Fig. 9. *Marinella lugeoni* PFENDER. Thin slide RR2340; × 36.

Fig. 10. *Heteroporella lepina* PRATURLON. Fragment; thin section RR3246; × 36.

Figs. 11–12. Ostracods from the type level of *Montiella filipovic*i; thin slide; RR2332; × 60.

Fig. 13. *Jurella*? from the type of *Montella filipovici*; thin slide RR2331; × 60.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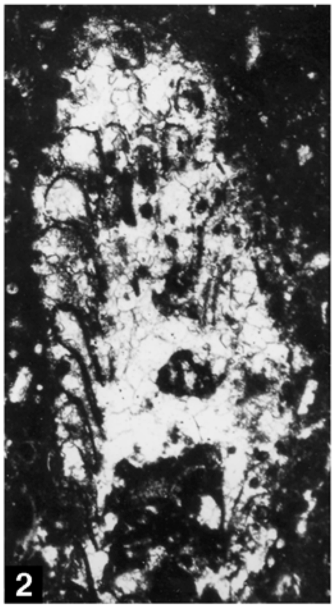
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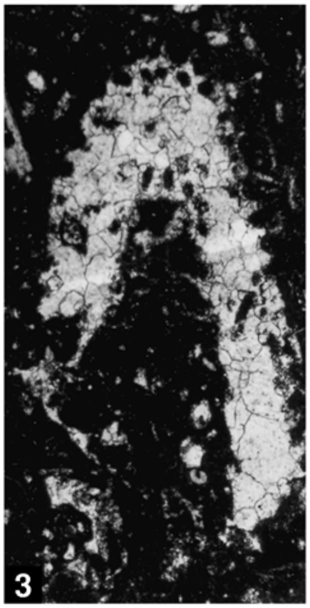
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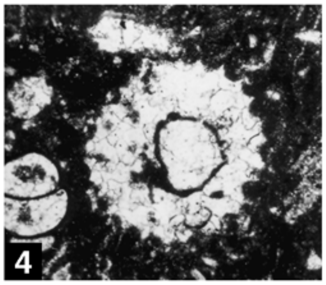
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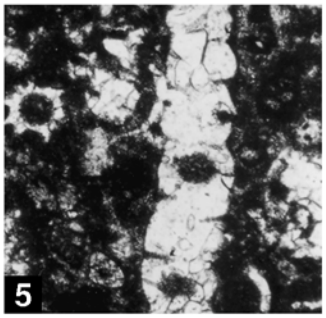
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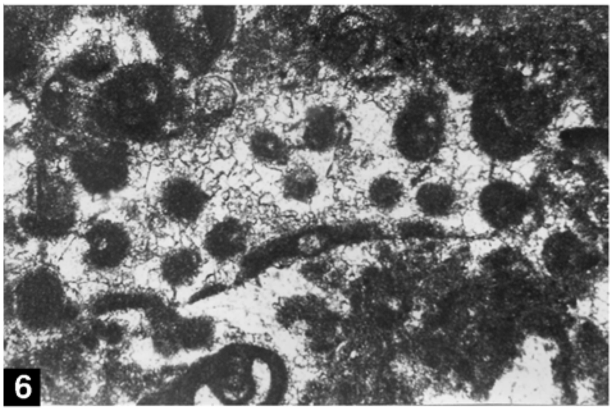
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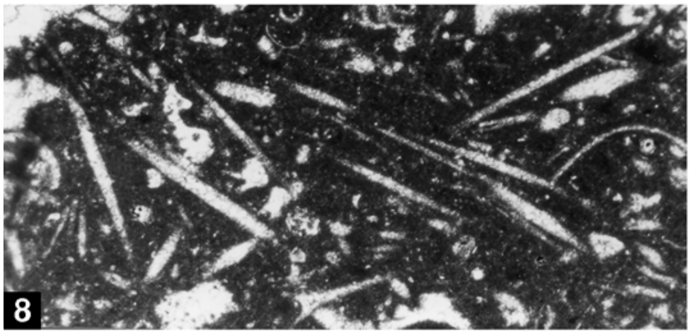
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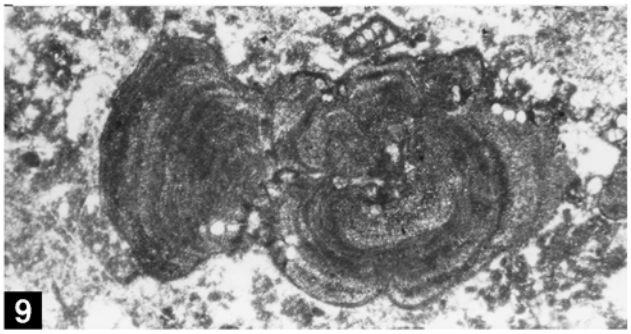
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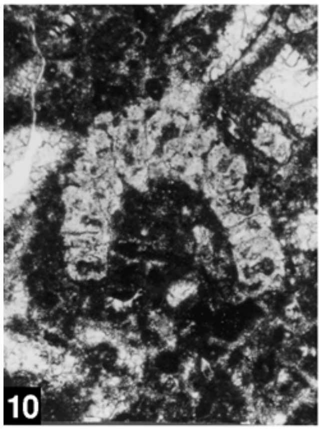
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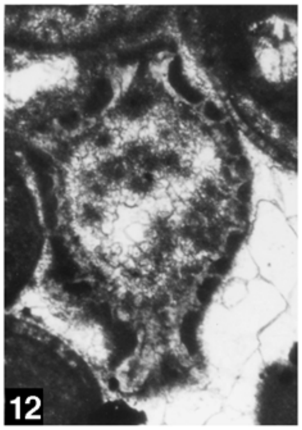
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PLATE 8

Figs. 1–6. Aptian bioclastic sediments of the northern Paštrik:

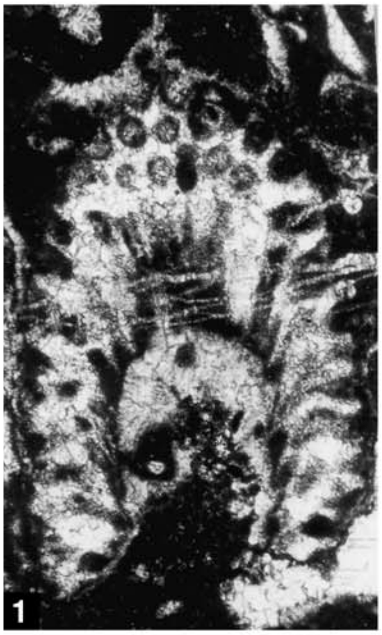
1. *Suppiluliumaella* sp. Thin slide RR2422; × 35,5.
2. *Salpingoporella pygmaea* (GÜMBEL). Thin slide RR2422; × 35,5.

3, 4, 6. *Acroporella radoicicae* PRATURLON (3) algal crusts (4) and *Planomalina* sp. (6); thin slide RR2045; 2045/1; 3 = × 43; 4, 6 = × 30.

5. *Dissocladella*?; thin slide RR2422, × 60.

Figs. 7–10. Problematic microfossil "*Pr 11*" from the Lower Cenomanian of the Vrbnica; Thin slide RR2318/4; 7, 8 = × 85; 9, 10 = × 175.

Figs. 11–13. Problematic microfossil "*Pr 10*" in association with *Trinocladus divnae*; thin slides RR2344, 2344, 2343; × 130.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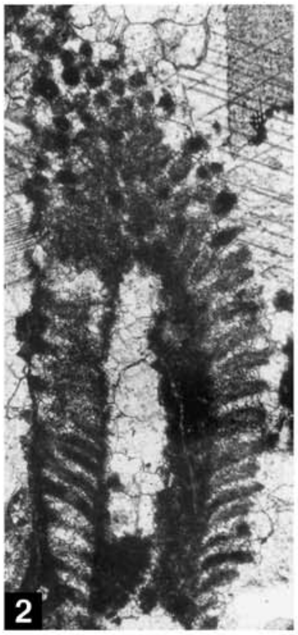
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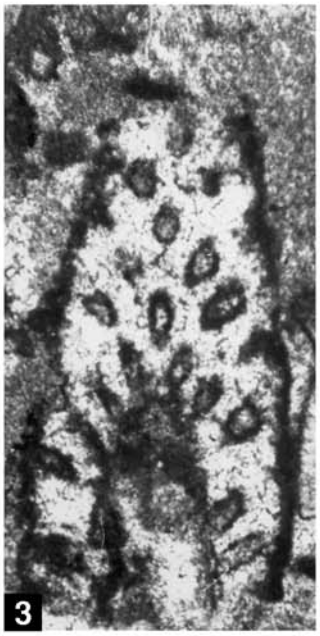
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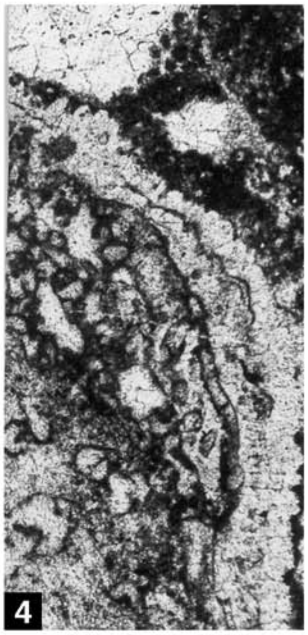
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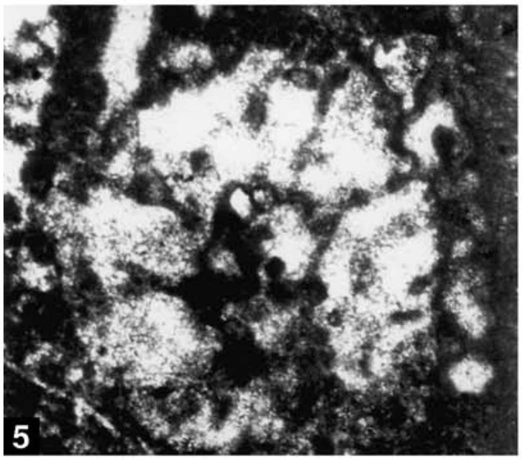
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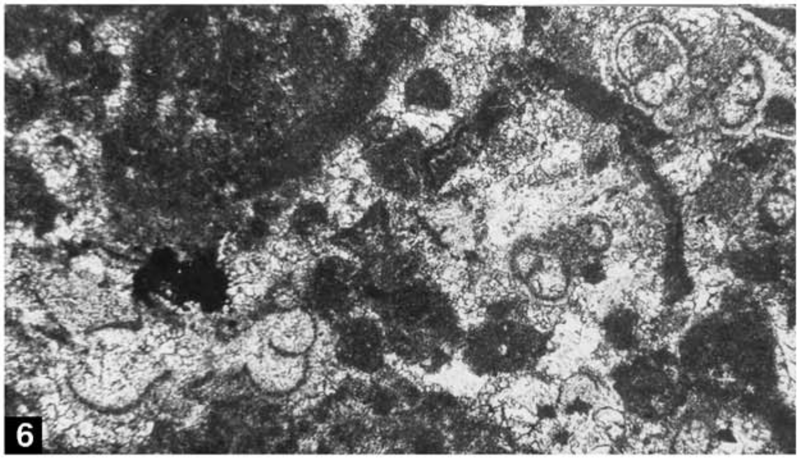
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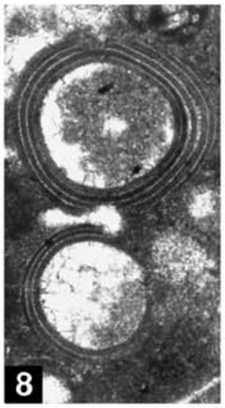
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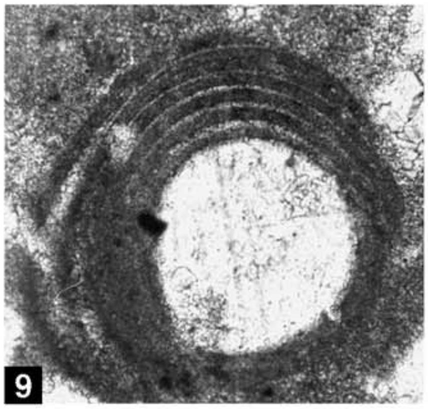
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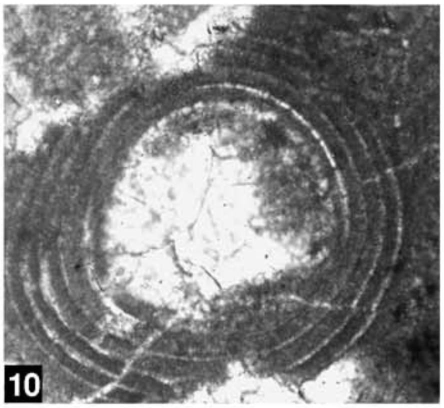
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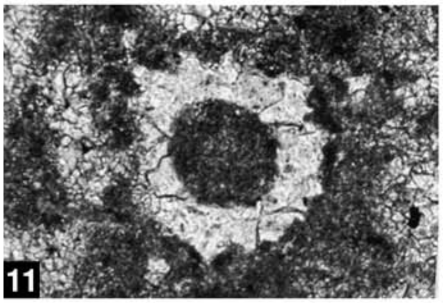
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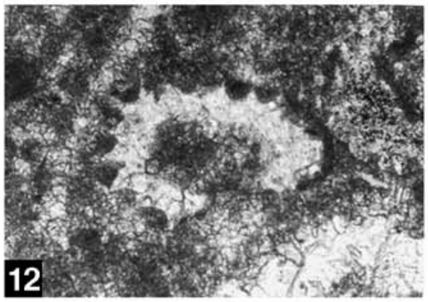
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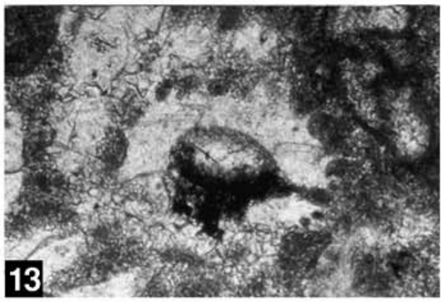
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ГЕОЛОШКИ АНАЛИ БАЛКАНСКОГА ПОЛУОСТРВА ANNALES GÉOLOGIQUES DE LA PÉNINSULE BALKANIQUE	67	89–100	БЕОГРАД, децембар 2006 BELGRADE, December 2006

Late Miocene ostracodes of Serbia: morphologic and palaeoenvironmental considerations

LJUPKO M. RUNDIĆ

Abstract. About 11.5 million years ago, a tectonic uplift of the Eastern and Western Carpathians separated the Pannonian Basin from the rest of the Paratethys. This orogenesis event caused an unconformity between the Sarmatian brackish sediments and the Pannonian lake-sea deposits. More than 6 Ma later, in these parts of the Paratethys, changes in the geographic framework, hydrological conditions and brackish – caspibrackish water chemistry led to the disappearance of restricted marine forms of life. A few euryhaline and marginal marine species survived this environmental change. Among the ostracodes, some originally freshwater taxa, such as *Candoninae,* entered the lake-sea. Many lineages show gradual morphological changes. The older, low diversity ostracode fauna from the Lower Pannonian dispersed to the endemic species and genera during the Upper Pannonian. This interval is assigned as the "bloom time" for many ostracodes, both qualitatively and quantitatively. This time sequence is the last appearances of genera such as *Aurila, Cytheridea*, *Propontoniella,* etc. and simultaneously, the first appearances for many new genera, such as *Zalanyiella, Serbiella, Camptocypria, Sinegubiella* etc. During the Pontian, migration processes were present. Therefore, it can be supposed that many eastern Paratethyan forms have Pannonian origin.

Key words: Late Miocene, ostracodes, morphology, paleoenvironment, Serbia.

Апстракт. На прелазу између средњег и горњег миоцена, пре око 11,5 милиона година, тектонска издизања у Карпатима су довела до одвајања Панонског простора од остатка Паратетиса. То је резултирало дискорданцијом између сарматских, бракичних и панонских, каспибракичних наслага. Скоро 6 милиона година касније, промене географских прилика, хидролошких услова и формирање ослађене водене средине, довеле су до потпуног изумирања морских организама. Само неколико еврихалинских форми као и оних који су настањивали приобалне делове успело је да се прилагоди и преживи. Код остракода, слатководни облици попут кандонина, све више настањују такво велико језеро-море док неки филогенетски низови показују постепене измене. Старије панонска, слабије разноврсна остракодска фауна еволуира у неке ендемичне облике током млађег панона када долази и до правог процвата остракода, како по броју врста тако и по броју јединки. То је период последњег појављивања родова *Aurila, Cytheridea*, *Propontoniella* односно време првог појављивања родова *Zalanyiella, Serbiella, Camptocypria, Sinegubiella*. За време понта, запажени су миграциони процеси на овом простору и сматра се да многи облици који живе у источном Паратетису имају панонско порекло.

Кључне речи: млађи миоцен, остракоде, морфологија, палеоекологија, Србија.

Introduction

As a product of Alpine tectonics in the Late Oligocene and the Early Miocene, a few molasse basins were created along the northern foreland of the uplifting mountain ranges in middle and southeastern Europe (Fig. 1). Although, each basin has its own individual history, their developments display some common features; repeated cycles of isolation from the world oceans, as inferred from the barrenness of the fauna or the presence of endemic organisms, and evolution from fully marine through brackish to caspibrackish and fluvial depo-environments (KOVAČ & MARTON, 1998; PIPÍK, 2000; FORDINAL *et al*., 2006;). Temporary isolation and filling of the western-central Paratethyan basins started from west to east. Firstly, the Alpine molasse

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basin was filled with sediments during the Middle Miocene, then the Pannonian Basin during the Late Miocene–Early Pliocene and, finally, the Dacian Basin during the Pliocene (HORVATH, 1990; MAGYAR *et al.*, 1999). Simultaneously, the eastern Paratethyan basins (Black Sea and Caspian) remained aquatic (SAFAK, 2002; TUNOGLU, 2001, 2002, 2003; WITT, 2003).

During the Late Miocene, most of Serbia belonged to the Pannonian Basin, while a small part of eastern Serbia corresponded to the Dacian area. During this time interval, there was a brief connection between them, but the most important period was during the existence of the so-called "Pannonian–Pontian lake-sea", a deposition area completely isolated and closed from the east by the Carpathian Mountains. Some authors when referring to the Pannonian and Pontian Stages used the name *Lake Pannon* (MAGYAR & GEARY, 1999; MAGYAR *et al.*, 1999; MÜLLER *et al.*, 1999; SZUROMI-KORECZ *et al.*, 2004). The above-mentioned connections between the basins established in the Pontian represented the last phase of the evolution of the western Paratethys. Subsequently, the Pannonian Basin disappeared as a lake-sea and transformed to a marshland environment. On the other side of the Carpathians Mt., the Dacian Basin still continued its existence during the early Pliocene.

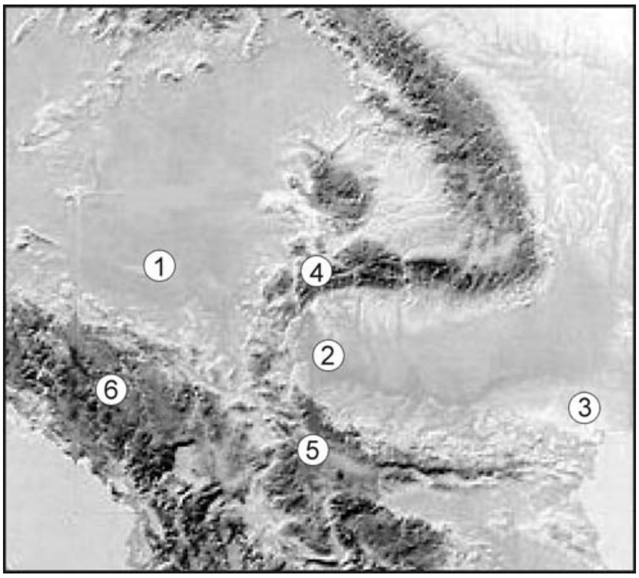


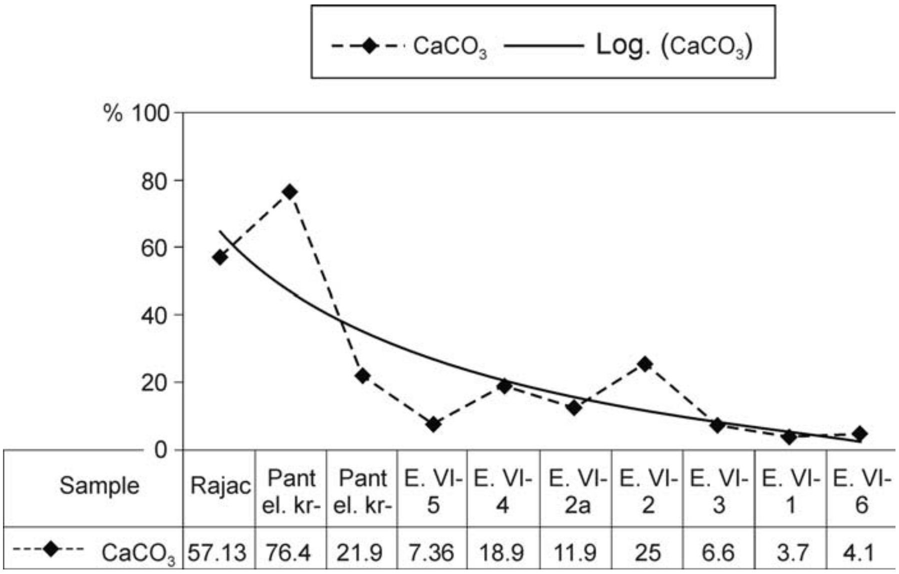
Fig. 1. The Late Miocene Paratethyan province: **1**, the Pannonian Basin; **2**, the Dacian Basin; and **3**, Euxinian Basin; **4**, the Carpathians Mt.; **5**, the Balkans Mt.; and **6**, the Dinarides.

Sedimentological studies have been limited to the investigation of main grain-size classes (rudites, arenites and lutites). Among forty samples, 42% of them are rudites (> 2mm), 38% are sand (0.63–2mm) and 20% are mud (< 0.63 mm). These deposits are slightly dominated by a coarse mean grain-size of gravel and sand (rudites, > 2mm). The distribution of the sedimentary material follows the expected general, fluvial lacustrine model: the coarsest deposits are distributed in the main channels, associated with a more dynamic water regime. On the contrary, the smaller particles occur in zones of low energy (OBRADOVIĆ & JOVANOVIĆ, 1987). Except in the opening of the main channel (the Kolubara Bay), where tidal currents and waves continuously generated high dynamic conditions throughout the year, the mean grain-size pattern does not correspond to a dynamic regime during the major part of the year.

Table 1. CO_2 and $CaCO_3$ content of the well Rgh-107.5, the Kolubara basin, western Serbia.

BOREHOLE Rgh-107.5	SAMPLES		% CO_2	% $CaCO_3$
	No.	Depth (m)		
1	01	05.70–06.80	3.20	7.27
2	16	92.70–93.00	1.20	2.73
3	27	139.60–139.80	2.00	4.55
4	28	144.00–144.20	2.40	5.46
5	29	150.00–150.20	4.40	10.00
6	31	168.70–168.90	2.40	5.46
7	35	192.00–192.20	2.80	6.36
8	41	234.00–234.20	5.60	12.73
9	42	240.30–240.50	14.40	32.73
10	43	247.30–247.50	1.20	2.73
11	44	257.00–257.20	35.60	80.92
12	45	262.40–262.60	29.60	67.28
13	46	267.50–267.70	34.40	78.19

For example, in the well Rgh-107.5 (Kolubara Basin), there are rapid decreases of calcium carbonate from the bottom to the top of the investigated well (Table 1). More consolidated deposits, such as Sarmatian limestones and sandstones, contain up to 80% of $CaCO_3$. On the other hand, semi-consolidated Pannonian and Pontian marls and sands contain a low percent of carbonate, as well as of carbon dioxide. Sample No. 28 (144.00–144.20 m) corresponds to marly sands and marls (OBRADOVIĆ & JOVANOVIĆ, 1987). Sands have a symmetrical grain size distribution and good sorting. Sandy marls shown similar effects. It is concluded that there was multiple alteration in the profile of the fluvial and near shore lake-sea deposits. Fluvial flows dispersed silicoclastites to the near shore parts and temporary deposited them on the land or alluvial environment. Somewhat different cases were observed on the sandy-silts (maximum 21.46% sand) and silts (maximum 98.95% silt) from the Late Miocene sediments of eastern Serbia. These deposits were earlier investigated by KRSTIĆ *et al.* (1992, 1995, 1997) and MIHELČIĆ (1990, 1991). There is a clear trend of decreasing calcium carbonate toward the younger Miocene Stages (Fig. 2). Also, curves of sorting and grain-size values show that the sediments have middle to good sorting and dominantly an asymmetrical grain-size distribution. Some other characteristics, such as pH (7.7–7.9)



Sample	Rajac	Pant el. kr-	Pant el. kr-	E. VI-5	E. VI-4	E. VI-2a	E. VI-2	E. VI-3	E. VI-1	E. VI-6
CaCO$_3$	57.13	76.4	21.9	7.36	18.9	11.9	25	6.6	3.7	4.1

Fig. 2. Decreasing trend of CaCO$_3$ from the Bessarabian to the Pontian (sandy-silts and silts from some localities of the Dacian Basin of eastern Serbia, after DODIKOVIĆ, 2001).

and Eh (+26 to –20), indicate that these deposits are the products of a slightly alkaline and low reduction environment (DODIKOVIĆ, 2001).

Quantitative and qualitative diversity of ostracodes

In the western and central Serbia (an example of the Kolubara Basin), more than 36 genera with 206 ostracod species were identified in the Late Miocene (RUNDIĆ, 1997). The most abundant taxa are representatives of *Candoninae,* including genera such as *Camptocypria*, *Cryptocandona*, *Fabaeformiscandona*, *Hastacandona*, *Lineocypris*, *Pontoniella*, *Propontoniella*, *Serbiella*, *Sinegubiella*, *Thaminocypris* etc. with about 80 species. Among the more abundant genera are *Amplocypris* (25 species), *Cyprideis* (23), *Loxoconcha* (22), *Hemicytheria* (20) and *Amnicythere* (19), which represent together about 90% of the total number of specimens. The findings of *Cypria* (8 species), *Hungarocypris* (4), *Xestoleberis* (3), *Candona* (2), *Ilyocypris*, *Tyrrhenocythere*, *Mediocytherideis* and *Stenocypria* (all with one species) were scarce and they represent only 10% of the mentioned population. Most of them have been recognized as infra/sublittoral. A smaller number of taxa have been found in the intertidal zone. Individual abundance may exceed 400 individuals per sample. Species diversity increases appreciably from the Lower Pannonian and is maximal during the Upper Pannonian (Fig. 3). This basin is the most illustrative example of marginal deposition in the whole of western and central Serbia during the Late Miocene. Other areas have more-less similar characteristics, except the northern province of Serbia (Vojvodina), where there is carbonate development. The best example of this kind is on the Fruška Gora Mt., where the Upper Pannonian is represented by fine grained “cement marls”, which correspond to the deeper part of the Pannonian Basin. The ostracod assemblages contain several genera, such as *Typhlocyprella,*

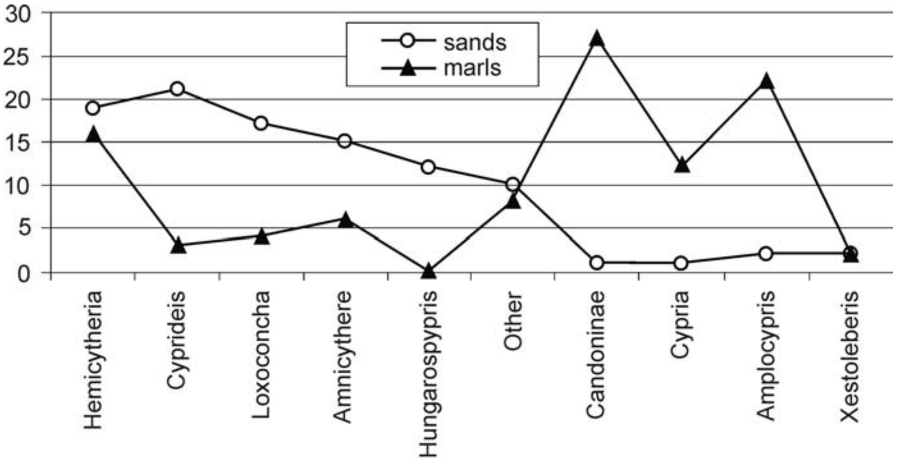


Fig. 3. Species number of main genera in relation to the sediments. The Lower Pannonian of the Kolubara Basin.

Zalanyiella, Camptocypria, *Serbiella*, *Amplocypris*, *Hemicytheria*, *and Cypria*. Apart from *Candoninae*, all other genera are represented by a few species only and their percentage contribution to the total population is small. The most abundant forms are different *Reticulocandona*, *Zalanyiella*, *Serbiella*, *Camptocypria*, etc. It seems that some representatives of *Candoninae* prefer more fine-grained deposits than the others.

In the eastern Serbia (Dacian Basin), there are certain similarities with the Pannonian basin but the ostracodes represent a different paleoenvironment, including a greater numbers and diversity of *Leptocytherinae* (Table 2). Conversely, the *Candoninae* are not as diversified as in the Pannonian Basin. During the Late Miocene (Late Bessarabian – Pontian), the ostracodes had the greatest diversity in the Maeotian and Pontian. The most abundant genera are representatives of *Leptocytherinae* (*Amnicythere*, *Euxinocythere*, *Maeotocythere*) on the one hand, and different forms of *Candoninae* on the other (> 80%). More than others, Dacian and Euxinian species, such as *Amnicythere sinegubi* KRSTIĆ, *A. alizadei* SHEIDAEVA, *A. palimpsesta* (LIVENTAL) *A. subcaspia* (LIVENTAL), *Maeotocythere prebaquana* (LIVENTAL*),* *Euxinocythere immutata* STANCHEVA, *E. suzini* (SCHNEIDER) etc., have been found.

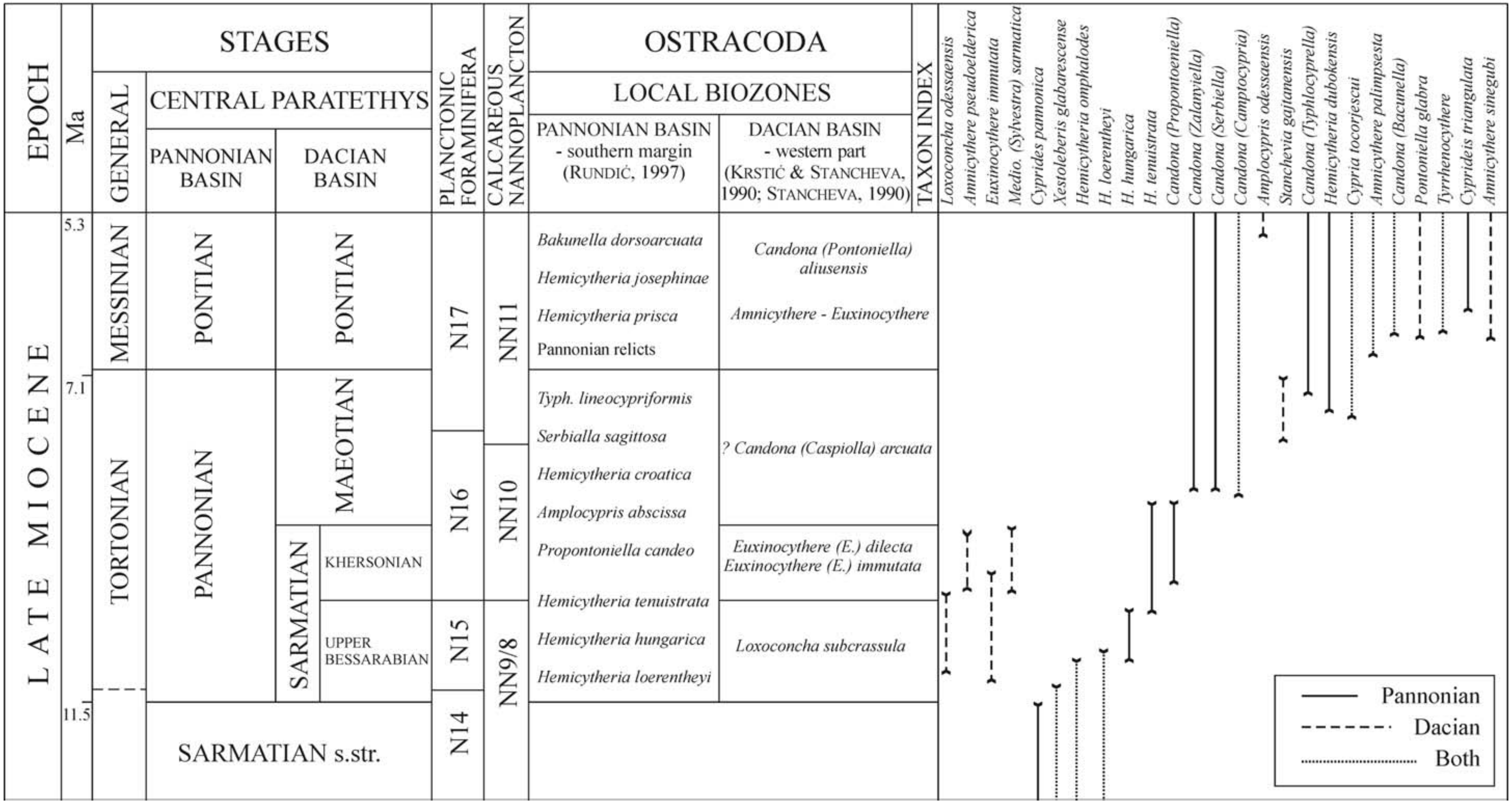
Among the *Candoninae*, *Candona* ex gr. *neglecta* Sars, *C. fagiolata* STANCHEVA, *Camptocypria ossoinensis* KRSTIĆ, *Camptocypria balcanica* ZALÁNYI, *Hastacandona pontica* AGALAROVA, *Bakunella guriana* (LIVENTAL*),* *Reticulocandona orientalis* KRSTIĆ, *Pontoniella paracuminata* KRSTIĆ, have been identified. Genera such as *Loxoconcha, Aurila, and Xestoleberis* are present in Sarmatian with a small number of species and relatively abundant specimens. In the higher stratigraphic levels, in addition to representatives of *Candoninae* and *Leptocytherinae*, the rare *Darwinula*, *Getocytheria, Stanchevia, Mediocytherideis, Tyrrhenocythere* etc. have been found. Most interestingly, representatives of the genus *Cyprideis* have a very small frequency and are limited to a few species, unlike the Pannonian basin. There are also similar results regarding the Pontian ostracodes from the eastern Black Sea region of Turkey (TUNOGLU, 2001, 2002, 2003).

Generally, there is a high diversity in the ostracod assemblages of the Upper Pannonian, Chersonian and also Maeotian. What are the possible reasons for this? First of all, there are some external factors, such as tectonic events and geographical isolation, the consequences of which were relatively stable aquatic conditions and an adequate range of time. If the Sarmatian *s. str.* – Upper Pannonian time period (over 2 Ma) is considered, then it can be said that the time interval was sufficient for taxonomic adaptation. During the older Pannonian, the first 'shock' affects the ostracodes and other fauna. The already freshened and relatively well-adapted Sarmatian brackish forms could not tolerate more freshening. Some of genera are completely absent, for example: *Cytheridea*, *Miocyprideis*, *Cyamocytheridea*, *Aurila,* etc. However, some other representatives, such as *Loxoconcha*, *Leptocythere*, *Hemicytheria,* etc., managed to survive the critical point and later developed special features as result of the adaptation. Mesohaline to freshwater and river-marsh genera, such as different *Candona*, *Cypria*, *Amplocypris*, and *Hungarocypris*, played the most important role. They are the markers for the future new ostracod assemblages, which consequently occur at the beginning of the Upper Pannonian. The second important event in the ostracod development during the Pannonian was the first appearance of many new taxons, such as *Zalanyiella*, *Camptocypria*, *Serbiella*, *Lineocypris*, *and Thaminocypris,* etc. There was also an increase in the ostracod abundance and diversity. An individual abundance may exceed a few hundred individuals per sample. At the same time, there was a high rate of speciation, which was probably the result of some optimal conditions and natural factors, such as geographic isolation, lake-sea adaptation, reproductive mode and dispersal, sexual selection, etc. (MARTENS, 1997). According to these authors, the tempo and manner of speciation were very different in various ancient lake assemblages. An example of a high diversity phenomenon, considered by MAYBURY & WHATLEY (1988), was the Upper Pliocene faunas of Cornwall, England and northwest France. They concluded that "*the high diversity symbolize general result of many factors as favourable preservation, high abundance, competent niche exploitation, the mixing of brackish, marine, cold and warm water species and some degree of allopatric speciation due to partial isolation of faunas because of the incomplete transgression of the Armorican and Cornubian massifs*".

It seems that there are many similarities between these and the Upper Pannonian assemblages. It was the time of the highest development of the faunas of the Pannonian basin. Biostratigraphically, it was the acme zone for many species (an example, the genus *Hemicytheria* represent 75% of all individuals).

After the Pannonian, tectonic subsidence of the Carpathian belt resulted in the reconnection of the Pannonian and Dacian Basins during the Pontian for the last time. At the beginning, thanatocoenoses were slightly different with many transitional types. Later, these differences became more visible. The faunal assemblages had new elements, such as *Bakunella*, *Tyrrhenocythere,* etc. During the Upper Pontian in this widespread environment, there was a trend of decreasing biodiversity among the ostracodes, unlike the previous period. The habitat was more saline and it resulted in species reduction. Analogous ideas were presented by BODERGAT (BONADUCE & SGARRELLA, 1999). The samples have numerous specimens but with relative few species. On the other hand, the faunas show an eastwards migration trend as some genera occupied eastern provinces and finally the Dacian Basin. In my opinion, many of the ostracod genera from the Dacian area have a Pannonian origin. In the Dacian Basin, for example,

Table 2. The correlation chart of the Late Miocene of Central Paratethys of Serbia and local ostracodes biozonation.



EPOCH	Ma	STAGES: GENERAL	STAGES: CENTRAL PARATETHYS: PANNONIAN BASIN	STAGES: CENTRAL PARATETHYS: DACIAN BASIN	PLANCTONIC FORAMINIFERA	CALCAREOUS NANNOPLANCTON	OSTRACODA: LOCAL BIOZONES: PANNONIAN BASIN - southern margin (RUNDIĆ, 1997)	OSTRACODA: LOCAL BIOZONES: DACIAN BASIN - western part (KRSTIĆ & STANCHEVA, 1990; STANCHEVA, 1990)
LATE MIOCENE	5.3	MESSINIAN	PONTIAN	PONTIAN	N17	NN11	*Bakunella dorsoarcuata* *Hemicytheria josephinae* *Hemicytheria prisca* Pannonian relicts	*Candona (Pontoniella) aliusensis* *Amnicythere - Euxinocythere*
	7.1	TORTONIAN	PANNONIAN	MAEOTIAN	N17 N16	NN11 NN10	*Typh. lineocypriformis* *Serbialla sagittosa* *Hemicytheria croatica* *Amplocypris abscissa*	*? Candona (Caspiolla) arcuata*
				SARMATIAN: KHERSONIAN	N16	NN10	*Propontoniella candeo*	*Euxinocythere (E.) dilecta* *Euxinocythere (E.) immutata*
				SARMATIAN: UPPER BESSARABIAN	N15	NN9/8	*Hemicytheria tenuistrata* *Hemicytheria hungarica* *Hemicytheria loerentheyi*	*Loxoconcha subcrassula*
	11.5		SARMATIAN s.str.		N14	NN9/8		

there are very abundant and diversified associations with *Amnicythere, Pontoniella*, *Candona*, *Bakunella* and *Tyrrhenocythere* in the younger levels of Pliocene (OLTEANU, 1998, 2000). There are also similar results regarding some representatives of molluscs (MÜLLER *et al*., 1999).

Morphological and phyllogenetical relationships (an example of the genus *Hemicytheria*, POKORNY)

An attempt has been made to correlate morphological features with some phylogenetical and paleoecological characteristics of the genus *Hemicytheria*. In earlier ostracod references, there are many different considerations concerning the genus *Hemicytheria*. In some, it was considered as an independent genus like in the first Pokorny description (POKORNY, 1955; STANCHEVA, 1971; SOKAČ, 1972; OLTEANU & VEKUA, 1989; OLTEANU, 2001; RUNDIĆ, 2002). Others considered it to be one of the subgenus: *Heterocythereis* (*Hemicytheria*) and *Heterocythereis* (*Tyrrhenocythere*) – MORKHOVEN (1962); *Graptocythere* (*Hemicytheria*) or *Aurila* (*Hemicytheria*) – KRSTIĆ, 1985, 1990. STANCHEVA (1971) divided the genus *Hemicytheria* on the three subgenus: *Getocytheria*, *Hemicytheria s. str*. and *Tyrrhenocythere*. The main parameters for these taxonomic relationships were: carapace morphology, hinge structure, muscle scars pattern and type of marginal pore canals. The anatomy of the soft body of recent representatives of the subfamily *Hemicytherinae* were also explored (MCKENZIE & BONADUCE, 1993).

Based on paleontological standards, some of the characteristics of the fossil representatives of the genus *Hemicytheria* could be shown. Above all, based on the carapace morphology, there are two main groups of *Hemicytheria*: one with a pitted and the other with a reticulated sculpture. It is not possible to clearly determine the phylogenetic evolution of *Hemicytheria*. It can be assumed because of biostratigraphical results in Early Pannonian deposits, that there was a branching off of the phylogenetic lineage in the Lower Pannonian. SOKAČ (1972) believed it probably occurred in the Upper Pannonian. The older Pannonian forms have pitted sculptures [*Hemicytheria ampullata* (MEHES), *H. hungarica* (MEHES)] which presented descendant species for smooth branch (*H. marginata* SOKAČ) and reticulate branch (*H. reticulata* SOKAČ, *H. dubokensis* KRSTIĆ). SOKAČ also considered the smooth forms to be phylogenetically younger because the marginal pore canals can be straight as well as bifurcated, whilst the pitted forms have only straight pore canals. The oldest Pannonian species, *H. omphaloedes* (REUSS) and *H. loerentheyi* (MEHES), also have a reticulated surface. Therefore, it could be said that the branching off occurred during the Lower Pannonian because there are representatives with both types of sculptures (only if the oldest forms of *Hemicytheria* are considered as the independent subgenus – *Graptocythere*). During the Upper Pannonian, there are forms with both types of ornamentation. In the Pontian, only three reticulated species remained whilst the pitted ones had disappeared. The ascendant/descendant problems are impossible to solve at this moment, because of the necessity for a very detailed instars analysis. If the marginal pore canals are considered, SOKAČ (1972) correctly claimed that the younger forms have both straight and bifurcated pore canals while the oldest pitted *Hemicytheria* have only straight ones. Meanwhile, investigations have shown that all the older species have straight pore canals, both the pitted and reticulated morphotypes.

The carapace size can reveal some of the rules of the development of *Hemicytheria*. Diagrams of the mean value of length and the L/H ratio are correlated and they show the trend of ostracod carapace development. In these diagrams (Figs. 4, 5), two trends in shell development can be seen. The first one is the continuous tendency of the value to increase from the Lower Pannonian to the Upper Pannonian, when some of the *Hemicytheria* have a maximal value. The second one, the opposite trend is visible during the Pontian when the *Hemicytheria* species decreased in value and lost some of their features, while some species disappeared completely. Most of large Upper Pannonian hemicytherids died off. During the Lower Pontian, there are transitional Pannonian/Pontian forms but they are not of great significance and disappear afterwards. From the relatively numerous Pannonian hemicytherids, only a few species survived [*H. josephinae* (ZALÁNYI)*, H. portafericae* RUNDIĆ and *H. pejinovicensis* (ZALÁNYI)] in the Upper Pontian. Small forms represented only 5% of total number of species. There is gradual increase in value during the Pannonian. In the Late Pannonian, the ostracodes "bloom" and all of *Hemicytheria* species have maximal dimensions. Forms of 1 mm in length represented about 40% of the total species and they dominated throughout the Upper Pannonian. Some transitional Pannonian/Pontian types have large shells but later the carapace trends to decrease in size. The best examples are Late Pontian species, which have smaller dimensions. The five greatest forms, according to the size of their shells, represented about 25% of the total species and all of them occurred during the Upper Pannonian and Pontian, especially (*H. dubokensis* KRSTIĆ*, H. portafericae* RUNDIĆ – Figs. 3, 4).

It can be concluded that there was a gradual disappearance of hemicytherids during the Late Miocene in the Pannonian Basin, as well as during the Maeotian in the Dacian Basin. This was the result of an increasingly freshening environment, which led to a completely freshwater biotope at the beginning of the Pliocene. The low percentages of dissolved carbonates and low salinity tolerance were not sufficient for the existence of hemicytherids. Meanwhile, later in the Dacian Basin there were similar forms represented by the genus

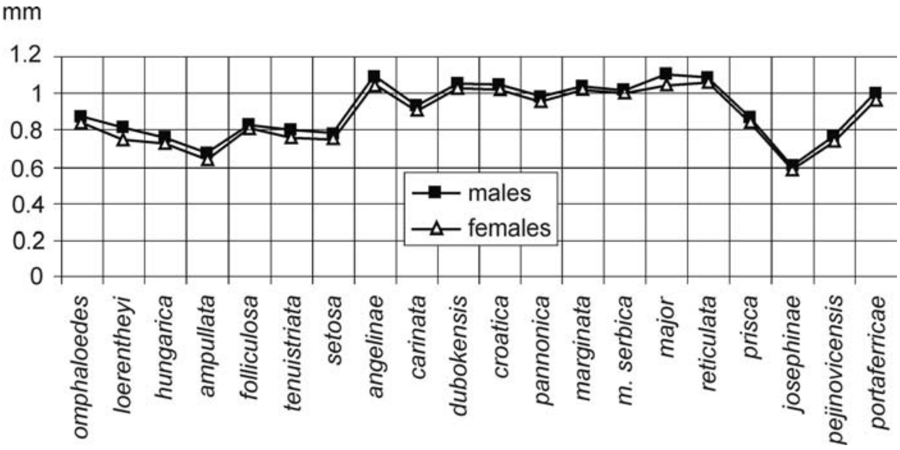
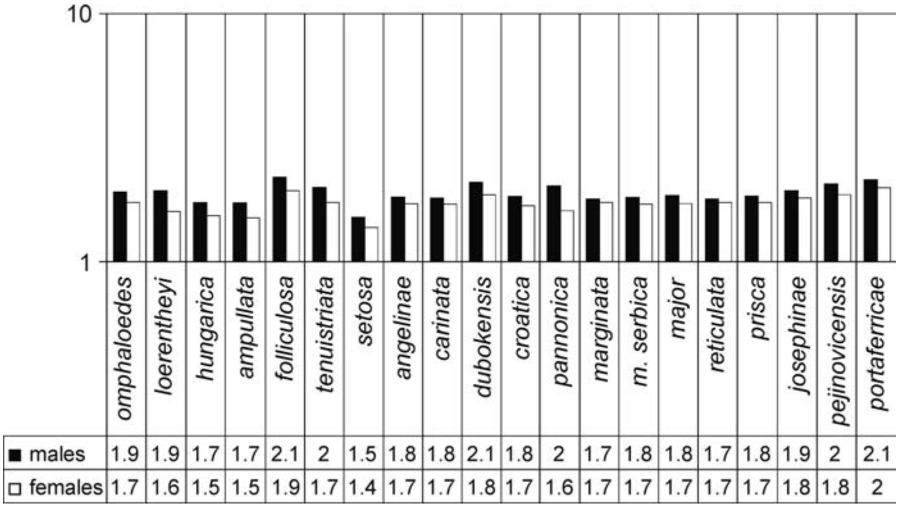


Fig. 4. The mean value of the length (mm) for the all of *Hemicytheria* species during the Pannonian and Pontian.



	omphaloedes	loerentheyi	hungarica	ampullata	folliculosa	tenuistriata	setosa	angelinae	carinata	dubokensis	croatica	pannonica	marginata	m. serbica	major	reticulata	prisca	josephinae	pejinovicensis	portaferricae
■ males	1.9	1.9	1.7	1.7	2.1	2	1.5	1.8	1.8	2.1	1.8	2	1.7	1.8	1.8	1.7	1.8	1.9	2	2.1
□ females	1.7	1.6	1.5	1.5	1.9	1.7	1.4	1.7	1.7	1.8	1.7	1.6	1.7	1.7	1.7	1.7	1.7	1.8	1.8	2

Fig. 5. Length/height ratio for some *Hemicytheria* species.

Tyrrhenocythere (Krstić, 1975; Krstić & Stancheva, 1990; Olteanu, 2000; Olteanu & Vekua, 1989; Stancheva, 1990). The genus *Tyrrhenocythere* has only been found in two wells in the Kolubara Basin during the Pontian (Rundić, 1997), despite the fact that more than 40 wells have been investigated in a diameter of about 30 km. The findings of the Pannonian realm from other locations are very limited. This could be explained by the fact that during the Pontian both basins communicated and migration processes between their inhabitants occurred for the last time. In addition, some other genera found in the Dacian Basin, such as *Bakunella, Pontoniella* and *Mediocytherideis* and probably many molluscs, have a western origin. After the Pontian, the Pannonian Basin finally disappeared.

Paleoenvironmental considerations

The main paleoenvironmental interpretations are based on the quantitative and qualitative analyses of ostracod assemblages and their morphological reflection to the main physico-chemical parameters of the habitat, including temperature, oxygen content, ionic concentration, salinity, etc.

According to Peypouquet (1979) and Rundić (1998), there is a connection between the sculpture and the Mg/Ca ratio of the carbonate of carapaces. Calcite crystals, constituting the hard layer of the carapace, are built perpendicular to the cuticle and their thickness is proportional to the Mg content of the crystal network. Frequently, the most gradated types are found in seawater conditions, the Mg content of which is higher than in freshwater conditions. As stated by Smith & Horne (2002), in nonmarine aquatic systems, the calcite mineral branch point represents an important step in determining the major ion composition of natural waters. In dilute water, below approximately 0.3 g/l, the most common major ions are bicarbonate, calcium and magnesium. With increasing salinity, these ionic concentrations rise, until calcite saturation is attained and calcite precipitates, at a total ionic concentration of approximately 0.3 g/l. This is the calcite branch point, the first mineral branch point in natural waters. Beyond this point, at about 1.4 g/l, the water becomes depleted in calcium and enriched in bicarbonates, or vice versa, resulting in a solute path towards bicarbonate enriched, calcium depleted saline water, or bicarbonate depleted, calcium enriched saline water. Ultimately, other mineral branch points are attained (gypsum, for example) and further changes in the major ion composition occurs [EUGSTER & JONES (1979), *in*: SMITH & HORNES, 2002]. However, some authors consider that this hypothesis is not sufficient. Namely, studies of the New Caledonia area have shown that, despite a high Mg/Ca ratio, the morphotypes are not more gradated than in others sites, and certainly less than the specimens living in seawaters where this ratio has only standard values. Other parameters, such as calcium content of the available food, pH and Redox potential (HOIBIAN *et al*., 2000), may have a contributing effect on carapace calcification. Ornamentation of many ostracod shells is affected by the Mg/Ca ratio of lake water: with Mg/Ca >1, the carapaces are heavily reticulate and the opposite ratio favors smooth ostracod valves. For example, CHIVAS *et al*. (1983) noted the connection between the thickness and ornamentation of an ostracod carapace and the water depth for fossil material obtained from Lake Buchanan in Queensland. They found that ostracodes belonging to the genus *Paracypris* associated with charophytes had a thick shell with a rarely visible reticulation and commonly no spines. Assemblage with charophytes indicates shallow water (up to 2 m), since this is the preferred habitat of these algae. In the same core, specimens of *Reticypris* with reticulated and often spinose shells are thought to indicate deeper water (> 2 m), since remains of shallow-water charophytes are absent (CARBONEL *et al*., 1988).

Essential in paleoenvironmental studies of the Serbian Late Miocene ostracod assemblages are 1) geographical position and 2) time range. It must be realised that there are extensive erosion and unconformity between Sarmatian restricted marine sediments and younger, caspibrackish deposits (RUNDIĆ, 1995, RUNDIĆ & MITROVIĆ,1998). This, by mountains enclosed system, was affected by continental water, which resulted in a brackish lake-sea and the nearly total extermination of marine biota. Only a few ostracod genera were able to survive these environmental changes (*Aurila*, *Loxoconcha*, *and Xestoleberis*). The newly formed habitats and gradually expanding lake-sea created a kind of "ecological gap", and stimulated the rapid evolution of survivors and the immigration of freshwater dwellers from the marshes, ponds and rivers (most of *Candona*, *Darwinula, Ilyocypris,* etc.). As with other fossils and in a still existing *long living lake* (MÜLLER *et al*., 1999), the originally low diversity fauna radiated into a large number of related endemic species and genera in the expanding and ecologically unoccupied lake-sea. The best examples for this are the many genera of Candonids, which had a radiated development during the Pannonian Stage. From the small number of species and genera during the Lower Pannonian (such as *Propontoniella* and *Cryptocandona*), in the Upper Pannonian there was an expansive evolution of these lineages (*Serbiella, Zalanyiella, Camptocypria, Sirmiella, Lineocypris, Typhlocyprella,* etc.) in both species and specimens. They are immigration forms from the marginal rivers and swamps, but most of them are the result of time resolution. Most of them are endemic species and lived only in this realm, such as *Hungarocypris*, which is a typical near shore dweller, preferring a sandier type of stratum. It is scarce in associations from fine-grained sediments. Its large and massive carapace must be the result of a rich ionic concentration, as well as living in warm, oxygenated and clear water (RUNDIĆ, 1991). However, the fossil record includes only three species of *Hungarocypris* with numerous individuals during the Pannonian–Pontian. The genus *Amplocypris* shows similar characteristics. The appearance of corpulent forms, strong carapaces and clear ornamentation give evidence for a shallow-water basin type with mobile flow and an important donation of land material. In normal oxygenated habitats, increasing ionic concentrations led to an increase in the number of sculptured morphotypes. The best examples are representatives of the family Cytherideidae (*Hemicytheria*, *Loxoconcha*, *Leptocythere*, and *Cyprideis*) with mostly ornamented forms. Analyses of the genus *Hemicytheria* in both Pannonian and Pontian deposits show that more than 90% of the species have ornamented shells. Taphonomic analysis of fossil associations, particularly the numerous specimens of *Hemicytheria setosa* RUNDIĆ, shows that during the Late Miocene, salinity was not the principal influence on the occurrence of valves and the type of ornamentation. This ornamented form lived on coastal and mobile parts of the Pannonian lake-sea and its ornamentation is a reaction to the adaptation. On the other hand, the Upper Pontian nodose forms lived in environments with decreased salinity and increased organic matter and silica (*Leptocythere*, *Ilyocypris*). In the Kolubara coal basin, these forms were discovered in sediments with a rich organic content. Based on sedimentological studies, data concerning transport and depositional mechanisms, which appear with intermittent alteration between fluvial and coastal deposition, are obtained. Fluvial flows brought and deposited more silicoclastics in the coastal regions, but in the alluvial part of the land, this occurrence was only periodic (well Rgh-107.5). MÜLLER *et al*. (1999), suggesting that the Pannonian–Pontian lake-sea continuously shrank in the north, due to the prograding deltas. They stated that the more southerly shores were much less affected by progradation. The shoreline remained in more or less the same position for a longer period of time. Along these shores, successions of paralic lignite beds were formed (STEVANOVIĆ *et al*., 1990). In phases of highstand, the lake-sea enlarged in some areas in the south and the shoreline became simpler as large islands became flooded. Coarse-grained clastic rims formed around the islands. From the very beginning of the lake-sea, these coarse clastics were often capped by white calcareous marl or limestone in the southern part of the basin, in locations far from the river mouths.

Modern analogues of Pannonian–Pontian ostracodes from the modern Ponto-Caspian regions have been used to estimate the paleosalinity of the lake-sea. These comparisons suggest that the average salinity of the Pannonian–Pontian lake-sea may have been approximately 10–12 ‰, but that significant local differences existed in shallow parts, such as the southern "shelf" dominated by river deltas. Natural water examination gives data similar to the results of the paleoecological approach. Down from the lacustrine prodelta silt, which is a regional stamp in the basin, inborn waters have salt in an amount approximately, or sometimes even higher, than that of normal marine salinity. Innate waters from above the silt, however, are diluted because of mixing with meteoric water, as shown by the diagenetic carbonates. The salinity of these diluted waters is about 5 ‰, which sets the lower limit for the original lake-sea salinity (Müller *et al.*, 1999).

Conclusions

According to qualitative and quantitative analysis of the ostracodes, the results of a factor analysis applied to the most frequent species to the population structures, it is possible to distinguish several different assemblages based on salinity:

- *Stenocypria, Cyprinotus, Cypridopsis, Ilyocypris, Darwinula* and *Candona* characterize the oligohaline environments (0.5–8 ‰ of NaCl) and indicate a shallow, coastal part of river mouths or in swamps. They occurred in the lowest parts of the Pannonian and in the Upper Pontian.
- *Hungarocypris, Thaminocypris, Propontoniella, Cypria, Serbiella, Camptocypria, Zalanyiella, Typhlocyprella, Pontoleberis,* etc. represent the dominant group of the caspibrackish–mesohaline habitats (8–18 ‰). All of them are related to the entire Pannonian–Pontian and lived in shallow infralittoral environments covered by water plants.
- Assemblages with *Cyprideis, Amplocypris, Leptocythere*, *Hemicytheria, Loxoconcha* and *Xestoleberis* represents more meso- to polyhaline (18–30 ‰) habitats and these associations occupied most of the southern Pannonian estuary (the Kolubara Basin, the Velika Morava Bay, eastern Serbia). Estuaries are semi-enclosed coastal waterbodies in which there is a salinity gradient from fully marine (35 ‰) to fresh water. The salinity varies due to tidal and seasonal influences.
- Ornamented forms of ostracodes lived in the coastal parts and the sculpture is a reaction to adaptation.
- Increases of ionic concentrations resulted in increased diversity (for ex. mostly of Upper Pannonian ostracodes)
- A deficiency of oxygen and a low Mg/Ca ratio in the water can eliminate sculptured morphotypes.
- Nodose ostracodes from the Upper Pontian (*Amnicythere, Ilyocypris*) lived under conditions of low salinity and increased organic matter.

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Резиме

Млађе миоценске остракоде Србије: морфолошка и палеоеколошка разматрања

Пре око 11,5 милиона година, тектонска издизања у Карпатима су довела до одвајања Панонског простора од остатка Паратетиса. То је резултирало дискорданцијом између сарматских, бракичних и панонских, каспибракичних наслага. Почетком плиоцена, промене географских прилика, хидролошких услова и формирање ослађене водене средине, довеле су до потпуног изумирања морских организама. Само неколико еврихалинских форми као и оних који су настањивали приобалне делове успело је да се прилагоди и преживи. Код остракода, слатководни облици попут кандонина, све више настањују такво велико језеро-море док неки филогенетски низови показују постепене измене. Старије панонска, слабије разноврсна остракодска фауна еволуира у неке ендемичне облике током млађег панона. Горњи панон је време процвата остракода, како по броју врста тако и по броју јединки. То је период последњег појављивања родова *Aurila*, *Cytheridea*, *Propontoniella* односно време првог појављивања родова *Zalanyiella, Serbiella, Camptocypria, Sinegubiella*. За време понта, запажени су миграциони процеси на овом простору и сматра се да многи облици који живе у источном Паратетису имају панонско порекло.

Остракоде које су настањивале Панонску провинцију, а посебно њене маргиналне делове, делимично су се разликовале од оних које су живеле у Дакијској провинцији. Основни разлози леже у чињеници да су те две провинције у појединим етапама током старијег горњег миоцена – панона, биле међусобно одвојене. Свака биопровинција је имала своје посебне карактеристике. Панонска је била више изолована и формирала је временом специфичну остракодску заједницу. Старије асоцијације су биле сиромашније јер су егзистовале у једној новој средини у односу на дотадашње услове. То се посебно добро види на примеру из Колубарског басена. Млађе панонске асоцијације су много више разноврсне и имају много више јединки. Основни разлог је довољно дуго време адаптације и касније повољни животни услови. Појављују се потпуно нови родови као одговор на већ консолидоване животне услове. Током горњег панона десиће се прави процват у еволуцији остракода на панонском простору. То се посебно односи на неке таксоне као што су кандониде и хемицитериде које доживљавају експанзију. С друге стране, у Дакијској провинцији која је у исто време више комуницирала са црноморским басеном, присутне су друге групе остракода које указују на ту везу, а посебно лептоцитерине. Такође, има и потпуно другачијих таксона којих нема у Панонском басену (*Getocytheria*, *Stanchevia*). Интересантно је рећи да род *Cyprideis* има врло малу заступљеност за разлику од Панонске провинције где је један од руководећих облика.

Посебно је добро проучен род *Hemicytheria* са свим својим врстама (20) и који показује све поменуте трендове у развоју остракодске фауне који су владали током панона и понта.

Са палеоеколошког аспекта посматрано, током млађег миоцена на простору обе провинције, егзистовало је неколико различитих типова средине у којима је живела остракодска заједница. Прву карактерише присуство родова који насељјавају олигохалинске (< 8 ‰), плитке обалске делове речних ушћа и мочвара: *Stenocypria, Cyprinotus, Cypridopsis, Ilyocypris, Darwinula, Candona*. Присутне су током старијег панона и млађег понта. Друга, доминантна заједница (*Hungarocypris, Thaminocypris, Propontoniella, Cypria, Serbiella, Camptocypria, Zalanyiella, Typhlocyprella, Pontoleberis*) представља каспибракичну, мезохалинску средину (< 18 ‰) односно инфралиторал обрастао воденом вегетацијом. Трећа остракодска асоцијација у којој су представници *Cyprideis, Amplocypris, Leptocythere*, *Hemicytheria, Loxoconcha* и *Xestoleberis* (> 18 ‰) одговара мезо и полихалинском биотопу односно најчешће већини тадашњих панонских и дакијских естуара (Колубарски басен, Великоморавски залив, и други заливи у источној Србији).

У вези са претходним, могуће је донекле успоставити међусобну зависност биотопских услова са изгледом и саставом остракодске љуштуре. Примећено је да, на пример, нодозне форме настањују средине које имају низак ниво растворене соли односно повећан проценат органске материје (родови *Ilyocypris* и *Amnicythere* у Колубарском басену). Сматра се и да смањен ниво кисеоника односно низак однос Mg/Ca у води, може довести до елиминације орнаментисаних форми остракода.

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Spectroscopic study of barite from the Kremikovtsi deposit (Bulgaria) with implication for its origin

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Abstract. Different genetic types (endogene and supergene) of barite from the Kremikovtsi deposit (Bulgaria) were studied by Laser-induced time-resolved luminescence (LITRL), Infrared (IR) and Raman spectroscopy. The IR spectra of the endogene barites are quite similar to those reported in the literature and do not show appreciable differences among them. The IR spectra of the supergene barites are almost identical to those of the endogene ones in respect to the positions of the vibrational modes ν_1, ν_2 and ν_4 of SO_4^{2}, except for a shift of 3 cm^{-1} for the ν_3 band. They displayed a presence of additional bands, which are close to the ν_3 and ν_1 modes of CO_3^{2-} in calcite. The Raman studies support the suggestion that the supergene barite contains traces of calcite.

The modern LITRL technique allowed the detection of several luminescent centers in the endogene barite. Eu^{3+} luminescence was identified for the first time in barite. The different emission spectra at 266 and 532 nm excitations suggest there are at least 2 structural positions for Eu^{3+} in the barite crystal lattice. The luminescent spectra also revealed a rather unusual violet-blue Nd^{3+} emission, which usually occurs in the IR spectral range, as well as emissions of Ce^{3+}, Eu^{2+}, Tb^{3+}, Ag^{+}, Sn^{2+}(?) and UO_2^{2+}.

The oxidation state of cations isomorphically present in the barite crystal lattice suggests the endogene barite in the Kremikovtsi deposit precipitated from reduced fluids supposedly subjected to cooling (conductive/convective) and oxidation (mixing with seawater).

Key words: barite, UV Time-resolved Luminescence, IR spectroscopy, Raman spectroscopy, Kremikovtsi deposit, Bulgaria.

Апстракт. Различити генетски типови барита (ендогени и супергени) из кремиковачког лежишта у Бугарској проучавани су спектроскопским методима LITRL, IR и Раман. Инфрацрвени спектри ендогених барита слични су спектрима из литературе и не показују међусобне разлике. IR спектри супергених барита су скоро идентични спектрима ендогених барита у погледу положаја вибрационих пикова ν_1, ν_2 и ν_4 SO_4^{2-} осим за помак од 3 cm^{-1} код ν_3. Спектри су приказивали присуство додатних трака које су блиске пиковима ν_3 и ν_1 CO_3^{2-} у калциту. Раманова спектроскопска проучавања иду у прилог наговештаја да супергени барит садржи трагове калцита.

Савремена техника LITRL омогућава откривање неколико луминесцентних центара у ендогеним бариту. По први пут је у бариту утврђена луминесценција Eu^{3+}. Различити емисиони спектри при побуђивању од 266 и 532 nm указују да постоје бар два структурна положаја Eu^{3+} у кристалној решетки барита. Луминесцентни спектри су такође открили необичну љубичасто-плаву емисију Nd^{3+}, која се обично јавља у инфрацрвеном опсегу спектра, као и емисије Ce^{3+}, Eu^{2+}, Tb^{3+}, Ag^{+}, Sn^{2+}(?) и UO_2^{2+}.

Степен оксидације катјона изоморфно присутних у кристалној решетки барита указује да је ендогени барит у лежишту Кремиковци наталожен из редукованих флуида за које се претпоставља да су били изложени хлађењу (кондуктивно односно конвективно) или оксидацији (мешањем са морском водом).

Кључне речи: барит, ултравиолентна луминисценција, инфрацрвена спектроскопија, Раман спектроскопија, лежиште Кремиковци, Бугарска.

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Introduction

A large share of the known and mined reserves of various metals (Pb, Zn, Cu, Ag, Hg, Sb, less Fe and Mn) are concentrated in carbonate-hosted polymetallic deposits of the sedimentary exhalative (SEDEX) type. A large Phanerozoic deposit of this type, the Kremikovtsi iron(+Mn)-barite-sulphide deposit in Bulgaria, which has been mined for iron, also contains large reserves of barite ore (29 million tons of barite ore and over 60 million tons of $BaSO_4$). Investigations of this deposit have mainly been focused on the iron ores (Damyanov, 1998, and the references therein). There are just a few studies on the mineralogy and geochemistry of barite (Atanassov & Vassileva, 1987; Zlatev & Mladenova, 1997; Vassileva *et al*., 2001; Dimova, 2006) and only one of them (Atanassov & Vassileva, 1987) presents a brief spectroscopic (IR and thermoluminescence) characterization of the Kremikovtsi barite. All these approaches, however, are based on routine conventional techniques (X-ray diffractometry, emission spectroscopy). In order to gain more insight into the barite structure and cation valence with implications for its origin, a set of modern methods (LITRL, IR and Raman spectroscopy) was used. Here, spectroscopic data for barite from this deposit, with emphasis on the results obtained using the laser-induced time-resolved luminescence technique, are reported.

Geological setting

The Kremikovtsi deposit (Fig. 1) lies in the southernmost part of the Kremikovtsi–Vratsa ore district, located in the eastern part of the Western Balkanides, which belong to the northern branch of the global Alpine–Hymalayan collisional orogenic belt on the Balkan Peninsula. The lithology of the region is presented by Paleo-, Meso- and Neozoic sedimentary rocks (Fig. 1). A major tectonic element is the Kremikovtsi thrust. The deposit is hosted in Middle Triassic dolomitic limestones in the western part of the Kremikovtsi thrust sheet. It consists of stratiform and lenticular iron formations and barite orebodies subparallel to the allochton bedding plane and subvertical (pipe-like) bodies of low-grade sulfide mineralization (Damyanov, 1998).

Two main genetic types of barite were found in the Kremikovtsi deposit – endogene and supergene (Atanassov & Vassileva, 1987), and 3 main morphological types of endogene barite bodies were distinguished: veins and lenses, veinlets and nests, impregnations (Dimova, 2006). Supergene barite occurs as colloform

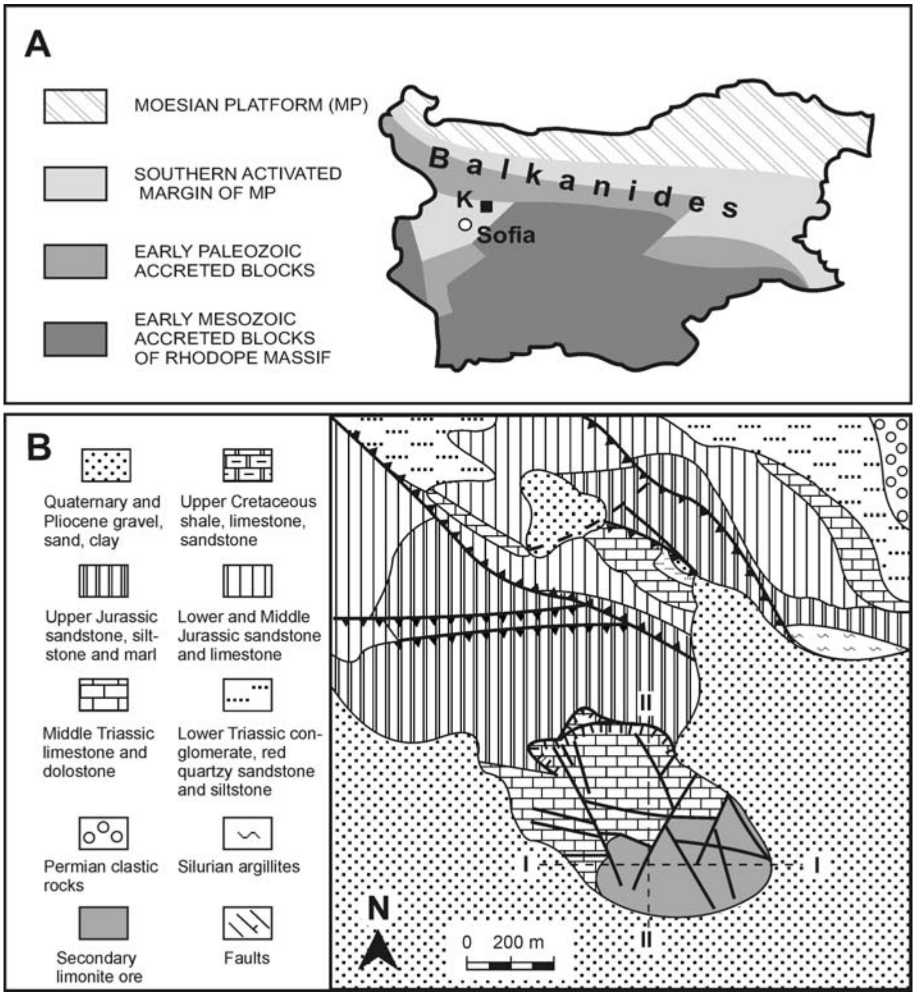

Fig. 1. Location (A) and geological map (B) of the Kremikovtsi ore field (after Damyanov, 1998).

crusts and crystal aggregates (Atanassov & Vassileva, 1987).

Material and methods

Isomorphic substitutions in barite were studied in 38 samples by Infrared Spectroscopy (Perkin Elmer Spectrum One fourier transform infrared spectrophotometer, transmission mode, KBr/sample = 300mg/1mg pellets) and 7 samples by Raman Spectroscopy (SPEX Raman Spectrometer, Ar$^+$ laser excitation at 488.0 nm, P=150 mW, step 1 cm^{-1}, integration time 1s, powder sample in capillary tube).

The luminescence spectra of 7 barite samples were investigated under 2nd, 3rd and 4th YAG harmonics (532, 355 and 266 nm, correspondingly) pulsed lasers excitations. The spectra observed at the geometry of 90° were analyzed by INSTASPEC equipment, enabling time-resolved spectra acquisition with the following facilities: delay times and strobe pulse duration 20 ns – 9 ms, spectral detection range 300–900 nm (1200 channels, spectral resolution 0.5–1 nm, gratings with 300 and 600 lines/mm), detector type – intensified CCD matrix. The luminescence spectra were measured at room temperature (300 K).

X-Ray powder diffraction (Philips PW1710 diffractometer, Cu K$_\alpha$, U=45 kV, I=40 mA, 2–70° 2θ, step 0.02° 2θ, 2 s/step; DRON M3, Co K$_\alpha$, U=40 kV, I=40mA, 2–70° 2θ, step 0.05° 2θ) was used to control the purity of the barite samples.

Results and discussion

Barite has an orthorhombic structure (2/m 2/m 2/m), where the S is situated in tetrahedral coordination with O and the Ba is surrounded by 12 oxygens of 7 SO$_4$ tetrahedra (James & Wood, 1925). The BaO$_{12}$ polyhedra and the SO$_4$ tetrahedra are edge-bound. The BaO$_{12}$ polyhedra are irregular: six of the Ba–O distances are 2.77–2.81 Å, and the other 6 – 2.91–3.32 Å, which suggests a "sheet" structure parallel to {001}.

IR and Raman Spectroscopy

The sulfate group has 4 fundamental vibrational modes: one nondegenerate (ν_1), one doubly degenerate (ν_2), and two triply degenerate (ν_3 and ν_4). The IR spectrum of barite exhibits several significant bands: 2 strong bands corresponding to asymmetric stretching and bending (ν_3 and ν_4), and 2 weak ones – to symmetric stretching and bending (ν_1 and ν_2). The IR spectra of the studied endogene barites from the 3 main morphological types of bodies and from different mineral associations (Fig. 2) are very similar each other and to that reported by Omori (1968; Table 1). Microprobe analyses show that Sr is the main isomorphic substitution for Ba in barite structure (SrO 0–4.5%; Dimova, 2006). Adler & Kerr (1965) found spectral shifts of the stretching modes (ν_3 and ν_1) to lower frequencies with increasing cation mass, such as the case of substitution Ba for Sr. The IR spectra of the studied endogene barites with different Sr contents (up to 4.5 % SrO) do not show any appreciable spectral shifts.

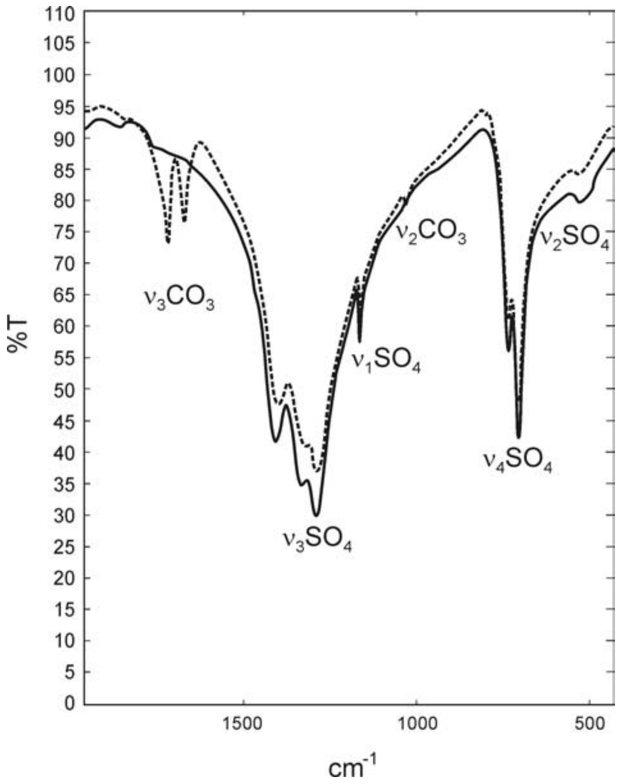

Fig. 2. IR spectra of endogene (bold line) and supergene (dashed line) barite from the Kremikovtsi deposit.

The IR spectra of the supergene barite (Fig. 2) are almost identical to those of the endogene ones in respect to the positions of the vibrational peaks, corresponding to the modes ν_1, ν_2, ν_4, except for a shift of 3 cm^{-1} for ν_3 (Table 1). A significant difference in the IR spectra of the supergene barite in comparison with the endogene one is the presence of additional bands at 1437, 1400 and 874 cm^{-1}, which correspond to the ν_3 and ν_1 modes of CO$_3^{2-}$ in calcite (Fig. 2; Table 1). This suggests that the supergene barite has traces of calcite, undetectable by XRD (<2%).

All the Raman spectra of the studied barite samples are dominated by an intense ν_1 band (symmetric stretching of SO$_4$ tetrahedra) at 988 and 984 cm^{-1} for endogene and supergene barite, respectively (Fig. 3). The other characteristic bands, ν_2, ν_3 and ν_4, reported in the literature, are also present (Fig 3; Table 2). The Raman spectra of the supergene barites show additional bands at 1054, 710 and 270 cm^{-1} (Fig. 3; Table 2). The Raman studies support the suggestion that the supergene barite contains traces of calcite: ν_1(1086 cm^{-1}), which coincides with ν_3 of the SO$_4$ group and ν_4 (710 cm^{-1}) bands.

Table 1. Peak positions (cm^{-1}) of the Infrared modes of barite and calcite.

IR modes of SO_4^{2-}	Endogene barite, Kremikovtski deposit	Supergene barite, Kremikovtski deposit	Barite (OMORI, 1968)	IR modes of CO_3^{2-} in calcite (NAKAMOTO, 1997)
ν_1	982	982	980	
ν_2	467	468	470, 439	879
ν_3	1179, 1118, 1083	1176, 1115, 1182	1180, 1120, 1080	1429-1492
ν_4	610	610	633, 608	706
Additional bands		1437, 1400, 874		

Table 2. Peak positions (cm^{-1}) of the Raman modes of barite and calcite.

Raman modes of SO_4^{2-}	Endogene barite, Kremikovtski deposit	Supergene barite, Kremikovtski deposit	Barite (OMORI, 1968)	Raman modes of CO_3^{2-} in calcite (NAKAMOTO, 1997)
ν_1	988	984	987	1087
ν_2	462, 453	460, 454	460, 451	
ν_3	1167, 1140, 1084	1134, 1086	1167, 1140, 1083	1432
ν_4	647, 618	644, 618	646, 630, 617	714
Additional bands	1104	1054, 710, 270		

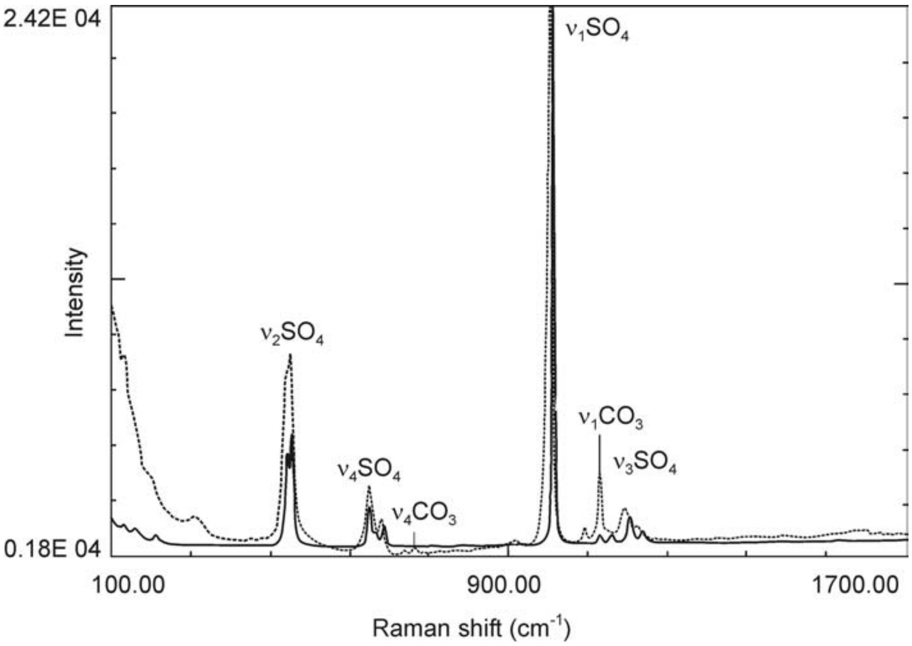

Fig. 3. Raman spectra of endogene (bold line) and supergene (dashed line) barite from the Kremikovtsi deposit.

Laser-induced time-resolved luminescence

Barite is one of the first luminescent materials from which the famous "Bologna stone" was obtained. It has been known for a long time that some barite specimens are fluorescent under UV exposure and emit white, yellow, green or orange light. In order to understand the nature of this phenomenon, the luminescence spectra of barite have been examined via UV, thermal and X-ray excitations. These studies allow the detection of the appearance of different luminescence bands from the UV to the red part of the spectrum.

Laser-induced time-resolved luminescence is a modern technique, which allows discrimination between centers with emission in the same spectral range, but with different decay times. The method involves recording the intensity in a specific time gate at a given delay after the excitation pulse, whereby both the delay and the gate width have to be carefully chosen. Such manipulation allows to be recorded separately the emissions corresponding to different decay times and to avoid the overlapping of the emissions produced from different centers. Laser-induced time-resolved luminescence of barites have detected and identified luminescence centers of Bi^{3+}, Bi^{2+}, Ag^{+}, Cu^{+}, REE^{3+} (Gaft *et* al., 2001a; Gaft *et al.*, 2005) in addition to UO_2^{2+} and Eu^{2+} revealed by Steady-State Photoluminescence (Tarashchan, 1978; Gaft *et al.*, 1985; Gorobets & Rogojine, 2001). Thus the information about the redox state of the different cations in the barite crystal lattice could throw more light on the origin of the barite.

The luminescence spectrum (excitation of 266 nm, without delay, broad gate of 9 ms) of the studied endogene barite contains 2 relatively narrow ultraviolet bands: one peaking at 306 nm and the other at 375 nm (Fig. 4A). The first band has a very short decay time and disappears after D = 50–100 ns. Such a combination of spectrum and decay time parameters is evidence that the luminescence is connected with Ce^{3+}. The emission of Ce^{3+} corresponds to transitions between $5d^1$ and $4f^1$ electronic configurations. The second band has a longer decay time of approximately 1 µs and belongs to Eu^{2+} (Gaft *et al.*, 2005). The Emission spectra of Eu^{2+} result from electronic transitions between $4f^7$ and $4f^65d^1$ electronic configurations. After a delay of several µs, the Eu^{2+} emission becomes much weaker and very weak narrow lines appear, peaking at 488, 544 and 615 nm (Fig. 4B). These lines are connected with trivalent rare-earth elements, which are characterized by relatively long decay times of hundreds of µs: the first 2 lines certainly belong to Tb^{3+} and the last one is principally typical for Eu^{3+} (Gaft *et al.*, 2005). While Eu^{2+} luminescence is common, Eu^{3+} emission is here detected for the first time in barite. Under excitation of 532 nm, a luminescence line at 615 nm dominates the spectrum, accompanied by several lines near 700 nm (Fig. 4C). Such a behavior confirms that Eu^{3+} is responsible for these luminescence lines. The emission of Eu^{3+} corresponds to f–f transitions – from the excited $^5D_0$ level to the ^7F_j (j=0, 1, 2, 3, 4, 5, 6) levels of the $4f^6$ configuration.

Some samples do not show Eu^{3+} luminescence. The relative intensities of Ce^{3+}, Eu^{2+} and Eu^{3+} in different samples are different: Ce^{3+} emission could be stronger than that of Eu^{2+}, while the luminescence of Eu^{3+} is relatively intensive. The different emission spectra at excitations at 266 and 532 nm suggest at least 2 structural positions for Eu^{3+} in the barite crystal lattice (Fig. 4C, D). Under excitation of 355 nm, a very broad orange-red band is detected (Fig. 4E), which was previously ascribed to Ag^+ luminescence (Gaft *et al.*, 2005). One of the samples showed a relatively strong Ag^+ luminescence dominating the spectrum even with an excitation of 266 nm, which is not optimal for this emission (Fig. 4F). The emission bands of Ag^+ result from d^9s–d^{10} transitions.

Under 266 nm excitation, without delay and with a broad gate of 9 ms, the luminescence spectrum contains 2 relatively narrow ultraviolet bands: one peaking at 320 nm and another at 375 nm (Fig. 4G). The latter band belongs to Eu^{2+}, while the origin of the first one is difficult to suggest. A similar emission has been found in synthetic barite artificially activated by Sn (Gaft *et al.*, 2005), but such an interpretation needs further support. The rather unusual narrow lines (at 370 and 391 nm) which appear in the luminescence spectrum with a longer delay of several µs (Fig. 4H) belong to trivalent rare-earth elements. They have been preliminary ascribed to violet-blue emission of Nd^{3+}, but this identification should be further clarified. The weak broad green band (at around 500 nm) may be related to uranyl emission, well known in barite (Fig. 4H).

The LITRL study of supergene barite did not show any luminescence. This could be connected with the lower impurity level in supergene minerals in general. A thermo-luminescence study of barite (Krivovichev, 1971) showed that the intensity of luminescence depends on the Sr concentrations in barite. The available data on the chemical composition of the Kremikovtsi supergene barite (Atanassov & Vassileva, 1987) showed very low Sr contents, which accounts for the absence of luminescence. Another possible explanation is that the luminescence is quenched by components with high-energy phonons, such as water or organic matter (Gaft *et al.*, 2005).

The main prerequisite needed for a mineral to display luminescent properties, in the case that the luminescent centers are minor elements, is a similarity of the ionic radii and charges of the host and isomorphic elements. It is known that the luminescent centers Eu^{2+} (1.24–1.40 Å), Ce^{3+} (0.88–1.02 Å), Nd^{3+} (0.99–1.15 Å), Tb^{3+} (0.89–1.09 Å), Ag^+ (1.13–1.26 Å), Sn^{2+} (0.93 Å) substitute for Ba^{2+} (1.35–1.44 Å) or Sr^{2+} (1.10–1.27 Å), which are in 12-fold coordination in the barite structure (Gorobets, 2002; Gaft *et al.*, 2005; all ionic radii are for the 6-coordination form). A possible accommodation for the established in this study Eu^{3+} (0.97–1.13 Å) is also isomorphic substitution for Ba^{2+} or Sr^{2+}. The presence of uranyl molecules in barite is considered to be a result of chemical adsorption (Gorobets, 2002).

Samples from both morphological types of endogene barite (veins and lenses, and veinlets and nests) fall into 2 groups: (1) barite with Eu^{2+} and Sn^{2+}(?), or Eu^{2+} and Ce^{3+}; (2) barite with Eu^{2+}, Eu^{3+} and Ce^{3+}. Barite samples from the first group contain isomorphic cations in their reduced form (Eu^{2+}, Sn^{2+}, Ce^{3+}), which implies that the barite precipitated from reduced fluids. A possible manner of this precipitation is conductive/convective cooling of the transporting fluid, which reaches the point of barite saturation with no substantial oxidation. The barite from the second group has reduced (Eu^{2+}, Ce^{3+}) as well as oxidized (Eu^{3+}) cations. This suggests that the barite from these samples precipitated from reduced transporting fluid subjected to oxidation. It can be speculated that both morphological types of barite are the result of conductive/convective cooling and mixing of hydrothermal fluid with seawater. This corroborates the most recent submarine hydrothermal model for the formation of the Kremikovtsi deposit (Damyanov, 1996a; Damyanov, 1998): sub- and on-seafloor precipitation upon cooling and mixing of hydrothermal fluid with seawater.

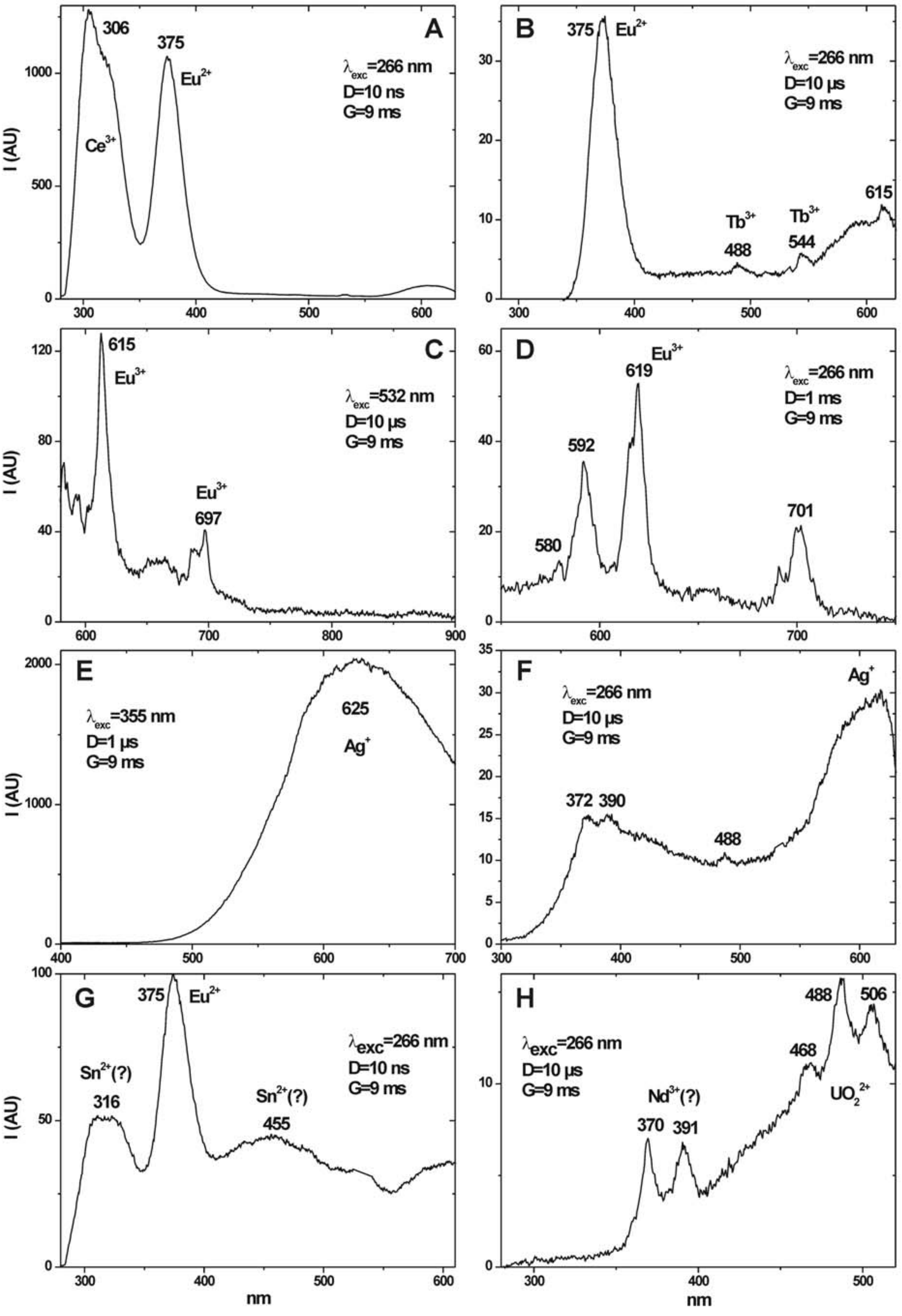

Fig. 4. Time-resolved luminescence spectra of endogenic barite from the Kremikovtsi deposit demonstrating different emission bands: (A) Ce^{3+}, Eu^{2+}; (B) Eu^{2+}, Tb^{3+}; (C) Eu^{3+}; (D) Eu^{3+}; (E) Ag^{+} (F) Ag^{+} (G) Eu^{2+}, Sn^{2+}; (H) Nd^{3+}, UO_2^{2+}.

Conclusions

The data obtained in the present study allow the following conclusions to be drawn:

(1) There is no structural difference depending on the varying Sr content (0–4.5 wt.%) in all the morphologic types of endogene barite, according to the IR and Raman studies. Supergene barite shows IR and Raman spectra identical to those of the endogene one with the only difference being the presence of traces of calcite.

(2) The laser-induced, time-resolved luminescence technique is a suitable tool for the identification of rare-earths (undetectable with conventional methods) in barite and the discrimination of their oxidation states.

(3) Along with the common Eu^{2+} luminescence, the emission of Eu^{3+} was detected for the first time in barite. There are at least 2 structural positions for Eu^{3+} in the barite crystal lattice.

(4) It seems that the endogene barite in the Kremikovtsi deposit was precipitated from reduced fluids subjected to cooling (conductive/convective) and oxidation (mixing with seawater).

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Резиме

Спектроскопско проучавање барита из лежишта Кремиковци (Бугарска) са импликацијама његовог порекла

Лежиште гвожђа(+Mn)-барита-сулфида Кремиковци у Бугарској садржи различите метале и велике резерве руде барита. Пошто се лежиште експлоатисало углавном због гвожђа, претежно су вршена испитивања гвоздене руде. Постоји само неколико минералошких и геохемијских студија барита. Циљ овог проучавања је боље упознавање са структуром и катјонском валентношћу барита са импликацијама његовог порекла. Различити генетски (ендогени и супергени) и морфолошки (жице и сочива, као и жилице и гнезда) типови барита испи-

тивани су применом спектроскопских метода ласерски редуковане луминесценције (LITRL), инфрацрвене анализе (IR) и Рамонове анализе.

Инфрацрвени спектри ендогеног барита сасвим су слични спектрима објављеним у литератури. Не постоји структурна разлика која зависи од променљивог садржаја Sr (0–4,5 теж.%) ни у једном морфолошком типу ендогеног барита. Инфрацрвени спектри супергеног барита скоро су идентични спектрима ендогеног барита у погледу положаја вибрационих пикова ν_1, ν_2 и ν_4 SO_4^{2-}, осим за помаке од 3 cm^{-1} код ν_3. Ти спектри су показивали присуство додатних трака, које су блиске пиковима ν_3 и ν_1 CO_3^{2-} у калциту. Раманова проучавања иду у прилог наговештаја да супергени барит садржи трагове калцита.

Савремена техника LITRL, као погодан начин утврђивања ретких земаља (које се не могу открити класичним методима) и разликовања њихових степена оксидације, омогућила је утврђивање неколико центара луминесценције у ендогеном бариту. Луминесценција Eu^{3+} је по први пут утврђена у бариту. Различити емисиони спектри при побуди од 266 и 532 nm указују на присуство бар два структурна положаја Eu^{3+} у кристалној решетки барита. Луминесцентни спектри су такође открили доста необичну љубичасто-плаву емисију Nd^{3+}, која се обично јавља у опсегу инфрацрвеног спектра, као и емисије Ce^{3+}, Eu^{2+}, Tb^{3+}, Ag^{+}, Sn^{2+}(?) и UO_2^{2+}.

Степен оксидације катјона изоморфно присутних у кристалној решетки барита указује да је ендогени барит у лежишту Кремиковци наталожен из редукованих флуида за које се претпоставља да су били изложени хлађењу (кондуктивно односно конвективно) и оксидацији (мешањем са морском водом).

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Crystallite size distribution of clay minerals from selected Serbian clay deposits

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Abstract. The BWA (Bertaut-Warren-Averbach) technique for the measurement of the mean crystallite thickness and thickness distributions of phyllosilicates was applied to a set of kaolin and bentonite minerals. Six samples of kaolinitic clays, one sample of halloysite, and five bentonite samples from selected Serbian deposits were analyzed. These clays are of sedimentary, volcano-sedimentary (diagenetic), and hydrothermal origin. Two different types of shape of thickness distribution were found – lognormal, typical for bentonite and halloysite, and polymodal, typical for kaolinite. The mean crystallite thickness (T_{BWA}) seams to be influenced by the genetic type of the clay sample.

Key words: kaolinite, bentonite, halloysite, BWA technique, Serbia.

Апстракт. Мерење просечне дебљине кристалита филосиликата и њихове дистрибуције извршено је на узорцима каолинских и бентонитских минерала помоћу BWA (Bertaut-Warren-Averbach) метода. Проучено је шест узорака каолинитских глина, један узорак халојзита и пет узорака бентонита из изабраних лежишта у Србији. Те глине су седиментног, вулканогено-седиментног (дијагенетског) и хидротермалног порекла. Утврђена су два различита облика дистрибуције дебљине кристалита – логнормална типична за бентоните и халојзит, и полимодална типична за каолинитске глине. Просечна дебљина кристалита (T_{BWA}) изгледа да зависи од генетског типа узорака глине.

Кључне речи: каолинитске глине, бентонит, халојзит, BWA метод, Србија.

Introduction

The size distributions of crystallites can be measured by powder X-ray diffraction (XRD) because the widths of the XRD peaks broaden as the crystallite size decreases, if the influence of associated components on the degree of disorder of clay minerals (as presented for kaolinite by Galan *et al*., 1994) is eliminated by adequate sample preparation. The interpretation of distribution and the shapes of crystallite thicknesses, measured by the Bertaut-Warren-Averbach (BWA) method, can then be related to crystal-growth mechanisms according to the theoretical approach of Eberl *et al*. (1998a).

The BWA technique has been applied to the measurement of illite particle thickness (Eberl *et al*., 1998b), to measure the crystallite size distribution of kaolin minerals (Šucha *et al*, 1999), to explore crystal growth mechanisms for illite and smectite (Środoń *et al*., 2000; Mystkowski & Środoń, 2000), to study the diagenetic evolution of the crystallite thickness distribution of illitic material (Kotarba & Środoń, 2000), weathering processes which affected smectite and illite/smectite (Šucha *et al*., 2001), and crystallite-size changes of pyrophyllite during grinding (Uhlik *et al*., 2000). Eberl *et al*. (1998a) studied the growth mechanism of minerals based on the shapes of the crystal size distribution.

Different clay deposits in Serbia have been explored and studied for many decades (Simić, 2001, 2004; Simić & Jović, 1997; Radosavljević *et al*., 1994; Stangačilović, 1970a, 1970b), but the crystallite size of the clay minerals has never been determined.

The main goal of this study was to measure the thickness and thickness distribution of kaolinite and smectite crystallites by the BWA technique and to compare the results with those obtained for similar clays from Slovakia and some other world deposits, and to

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check if the mean crystallite size depends on the origin of the clay.

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Materials and methods

Twelve samples of kaolinites, halloysite, and smectites from selected deposits in Serbia were used for this study. The kaolinites were collected from the Vrbica (sample VRB), Ćirinac (CIR-Z), Lazine (L-1), and Košarno (KOS-5A) deposits (Aranđelovac basin), the Rudovci (RUD-3) deposit (Kolubara basin) and the Jasenovac coal mine, the halloysite was from the Novo Brdo deposit, and the smectites were from the bentonite deposits or occurrences Popovac (POP-1), Mečji Do (MD), Bogovina coal mine (BOG-I), Bivolica (BIV), and Drmno coal mine (D-3). These deposits were selected on the basis of their different clay minerals, genetic types and parent rocks.

Prior to analyses, < 2 mm fractions were separated from the bulk samples by sedimentation. Separated fine fractions were used for X-ray diffraction (XRD) analysis of oriented specimens. The oriented specimens were prepared by sedimentation of the clay suspension (10 mg/cm^2) onto glass slides. All specimens were analysed by XRD using a Philips PW 1710 diffractometer equipped with Cu radiation with a graphite monochromator. The step size was 0.02° 2Θ with a counting time of 5s for the oriented specimens.

The resulting basal reflections of the clay minerals were used for the determination of the mean crystallite thickness (crystallite = X-ray scattering domain) and thickness distribution by means of the BWA techniques (DRITS *et al*., 1998) using the MudMaster program (EBERL *et al*., 1996). The XRD method of crystallite size determination is based on the observation that XRD peaks broaden regularly as a function of decreasing crystallite size. The first basal reflection of all samples was subjected to BWA analysis in the recommended two theta intervals between 6 and 13° for the kaolinite (Fig. 1) and 2.5 to 7.5, for the smectites. All kaolinite samples, except the halloysite sample NB, contain illite in the clay fraction. Therefore, the illite peaks were chopped by the program PkChopr (Fig. 2). A longer XRD exposition time (5 s) was used to obtasin smooth XRD patterns for the analysis.

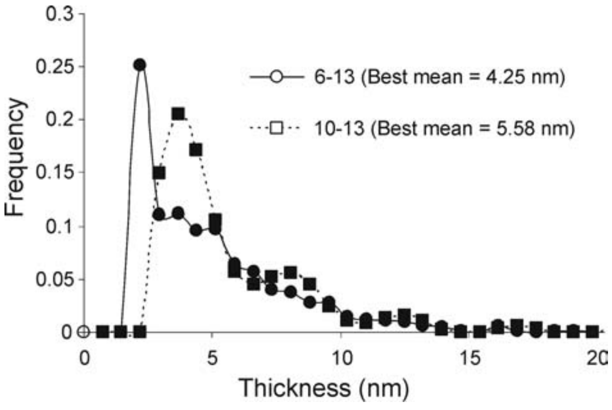

Fig. 1. Changes in the distribution and mean thickness of halloysite (sample NB) after using the incorrect analysed area (10–13° 2θ) in comparison with the recommended area (6–13° 2θ).

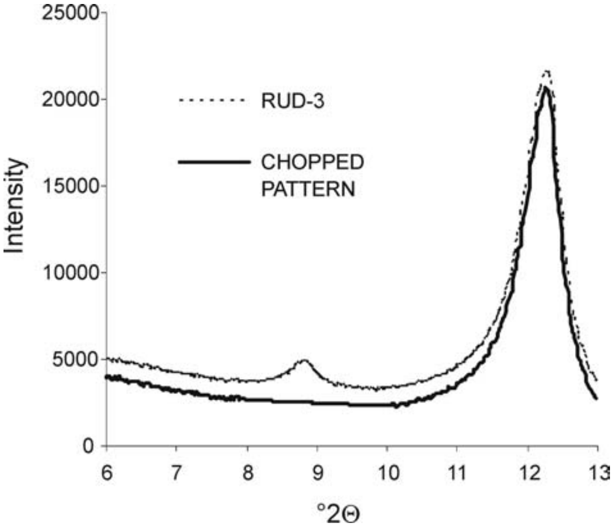

Fig. 2. Example of the modification of an XRD pattern before BWA-analysis the by PkChopr program.

Scanning electron images were taken from fresh rock chips coated with gold using a Jeol JXA 840 scanning electron microscope (SEM).

Geological features of the studied deposits

Samples of 6 kaolinites, 1 halloysite and 5 smectites from deposits in three different geological environments and origin were used for this study. Kaolinites and smectites from sedimentary (originating in a weathering crust and transported into sedimentary basins), volcano-sedimentary (formed by diagenesis of volcanic ash in a subaqual and subarial environment) and of hydrothermal origin were studied. The geological setting of each sample is indicated in Table 1. The analysed samples represent a selected collection of kaolinitic clays and bentonites from the economically most important deposits in Serbia, one sample of halloysite and one sample of Miocene tonstein, recently discovered in eastern Serbia.

Table 1. Basic geological features of the studied clays.

Sample	Genetic type	Age	Clay mineral	Basin	Deposit
Kaolinite					
VRB	Sedimentary	Miocene	Kaolinite	Aranđelovac	Vrbica
CIR-Z	Sedimentary		Kaolinite		Ćirinac
L-1	Sedimentary		Kaolinite		Lazine
KOS-5	Sedimentary		Kaolinite		Košarno
RUD-3	Sedimentary	Pontian	Kaolinite	Kolubara	Rudovci
JAS-5A	Volcano-sedimentary	Miocene	Kaolinite	Krepoljin	Jasenovac
Halloysite					
NB	Hydrothermal	Oligocene-Miocene	Halloysite (7Å)		Novo Brdo
Smectite					
POP-1	Volcano-sedimentary	Lower Miocene	Smectite	Paraćin	Popovac
MD	Volcano-sedimentary	Lower Miocene	Smectite	Zaplanje	Mečji Do
BOG-1	Sedimentary (?)	Miocene	Smectite	Bogovina	Bogovina
BIV	Sedimentary (?)	Lower Miocene	Smectite	Svrljig	Bivolica
D-3	Sedimentary	Pontian	Smectite	Kostolac	Drmno

The sedimentary kaolinitic clays were deposited in different basins, but in similar lacustrine settings. The kaolinitic clays from the Aranđelovac basin (Vrbica, Ćirinac, Lazine, and Košarno deposits) were formed by weathering and redeposition of materials from Bukulja granite (SIMIĆ, 2004). Kaolinite is the dominant clay mineral, with small amounts of illite. The clay from the Košarno deposit also has subordinate smectite. The parent rocks for the clays from the Rudovci deposit (Kolubara basin) are dacitic rocks and their pyroclastics. Kaolinite is also the most abundant mineral, accompanied by small amounts of smectite and traces of illite (SIMIĆ, 2004). The length of transport in both the Aranđelovac and Kolubara basins ranged from several hundred meters to 2–3 km, hence crystal disintegration during transport may have some influence on the crystallite size and thickness. Halloysite from the Novo Brdo deposit is of hydrothermal origin (MAKSIMOVIĆ & NIKOLIĆ, 1978). The kaolinitic clay from the Jasenovac coal mine is a typical tonstein, formed by the diagenetic alteration of volcanic tuff (ŽIVOTIĆ & SIMIĆ, 2003, unpublished report).

The smectite samples are also of different origin. Both Popovac bentonites, interbedded in marlstone in the quarry near the Paraćin Town and the Mečji Do deposit near the Vlasotince Town, are typical volcano-sedimentary rocks formed as a result of "in situ" subaqual alteration of the volcanic tuff. The bentonites from the Bogovina coal mine (East field) and the Bivolica deposit, near the Svrljig Town, are most probably the products of reworking of the weathering crust of andesitic rocks. The bentonite from the Drmno deposit (Kostolac coal basin) is of sedimentary origin (SIMIĆ *et al.*, 1997), but the primary source of clay minerals has not yet been established.

Results and discussion

Typical XRD patterns of each genetic type of clay are shown in Fig. 3, and the results of the BWA measurements of the kaolinite, halloysite and smectite samples in Table 2.

The T_{BWA} value of the sedimentary kaolinites studied varies between 5.55 and 7.91 nm, with an average value of 6.48 nm. The curves of all five samples are polymodal (Fig. 4), indicating that the samples consist of two or more generations of crystals with different thickness. The average T_{BWA} of the Serbian sedimentary kaolinites is slightly higher than the T_{BWA} of Slovakian sedimentary kaolinites, but, at the same time, significantly smaller than the T_{BWA} of selected world kaolinites (Table 3).

The BWA measurements confirmed the previously obtained geological, mineralogical and geochemical data that the weathering conditions during the Upper Oligocene–Lower Miocene did not lead to the origin of a well-developed kaolinitic weathering crust, neither in Serbia (MAKSIMOVIĆ & NIKOLIĆ, 1978; SIMIĆ, 2004), nor in Slovakia (KRAUS, 1989).

The halloysite sample of hydrothermal origin has a rather small mean crystallite thickness of 4.25 nm and a polymodal distribution pattern (Fig. 4). The T_{BWA} values of the Serbian and Slovakian halloysites are very similar, indicating a similar stage of hydrothermal alteration of the primary rocks. The distribution shapes of

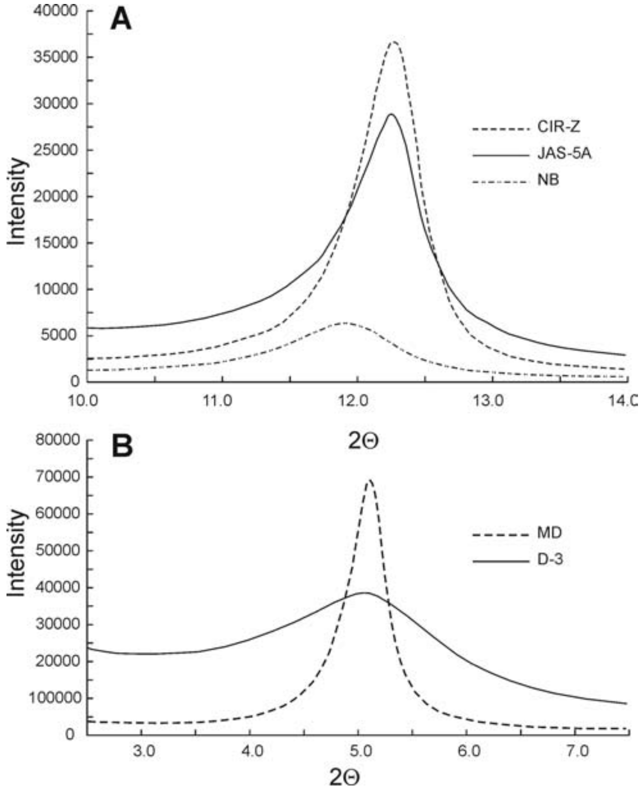

Fig. 3. Typical XRD patterns of the studied samples. A) Sedimentary kaolinite (CIR-Z), volcano-sedimentary kaolinite (JAS-5A), and hydrothermal halloysite (NB). B) Volcano-sedimentary smectite (MD), and sedimentary smectite (D-3).

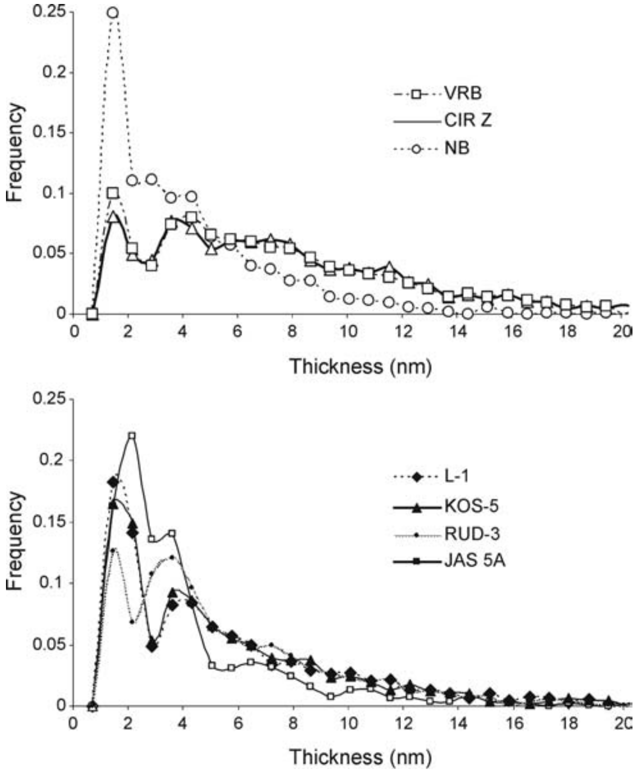

Fig. 4. Crystallite size distribution of kaolinites obtained by the BWA technique.

Table 2. List of clays used for BWA-analysis and the input and output data.

Sample	Genetic type	Position of peak maximum (θ)	d-spacing (Å)	Analysed area (θ)	Best mean (nm; extrapolated; area-weighted)
Kaolinite					
VRB	Sedimentary	12.27	7.21	6–13	7.53
CIR-Z	Sedimentary	12.29	7.20	6–13	7.91
L-1	Sedimentary	12.21	7.24	6–13	5.64
KOS-5	Sedimentary	12.30	7.19	6–13	5.55
RUD-3	Sedimentary	12.25	7.22	6–13	5.75
JAS-5A	Volcano-sedimentary	12.28	7.20	6–13	4.32
Halloysite					
NB	Hydrothermal	11.89	7.44	6–13	4.25
Smectite					
POP-1	Volcano-sedimentary	5.14	17.18	2.5–6	9.00
MD	Volcano-sedimentary	5.11	17.29	2.5–6	10.12
BOG-1	Sedimentary (?)	5.15	17.15	2.5–5.36	5.67
BIV	Sedimentary (?)	5.13	17.20	2.5–5.32	5.79
D-3	Sedimentary	5.07	17.41	2.5–4.88	5.21

Table 3. Average mean crystallite thickness (T_{BWA}) of Serbian, Slovakian and selected world kaolinites and halloysites. Values for the Slovakian and selected world kaolinites are from ŠUCHA *et al*. (1999).

Country	Genetic type	Average T_{BWA} (nm)
Kaolinites		
Serbia	Sedimantary	6.48
Slovakia	Sedimantary	4.81
World (selected)	Sedimantary	13.83
Serbia	Volcano-sedimantary	4.32
Halloysites		
Serbia	Hydrothermal	4.25
Slovakia	Hydrothermal	3.50

these two hydrothermal halloysites are different, as the Slovakian sample has an asymptotic shape and the Novo Brdo halloysite a polymodal one. The polymodal distribution of the Novo Brdo halloysite seems to be a combination of one lognormal and one asymptotic distribution. The asymptotic distribution is typical for samples with small T_{BWA} and could be characteristic for early stages of formation (EBERL *et al*., 1998a).

The tonstein sample of volcano-sedimentary origin also has a small mean crystallite thickness of 4.32 nm with a polymodal distribution pattern, but similar to lognormal (Fig. 4). The tonstein from the Jasenovac mine is generally weakly crystallized according to the XRD (Fig. 5) and at least two generations of kaolinitic minerals can be observed on the SEM image (Fig. 6), confirming the polymodal distribution shape.

The smectites from the volcano-sedimentary type have higher mean crystallite thickness with an average value of 9.56 nm than the smectites from the sedimentary bentonites with an average value of 5.56 nm. The crystallite size distributions for the volcano-sedimentary samples are lognormal (Fig. 7). Their shapes are quite different from the sedimentary types. The volcano-sedimentary smectites have identical distribution shapes with a theoretical lognormal distribution (Fig. 8A). The distribution of smectites from the sedimentary bentonites is different from the theoretical lognormal shape (Fig. 8B).

The mean thickness of smectites from selected world volcano-sedimentary bentonites varies from 6 to 9 nm (MYSTKOWSKI & ŚRODOŃ, 2000; MOL, 2001). The relatively wide range of T_{BWA} values of volcano-sedimentary smectites does not support the idea of using it to distinguish the origin in general. However, the measurement of T_{BWA} has sense for the differentiation of the origin of a bentonite in smaller regions, as was observed for the Serbian bentonites. A similar difference was found for in situ volcano-sedimentary and transposed bentonites from middle Slovakia (both types were characterized by ŠUCHA *et al*., 1996). Smectites originating from the in situ alteration of andesitic volcanoclastics have higher T_{BWA} values (up to 7 nm) in comparison with smectites originating by the redeposition of alteration products (5.5 nm).

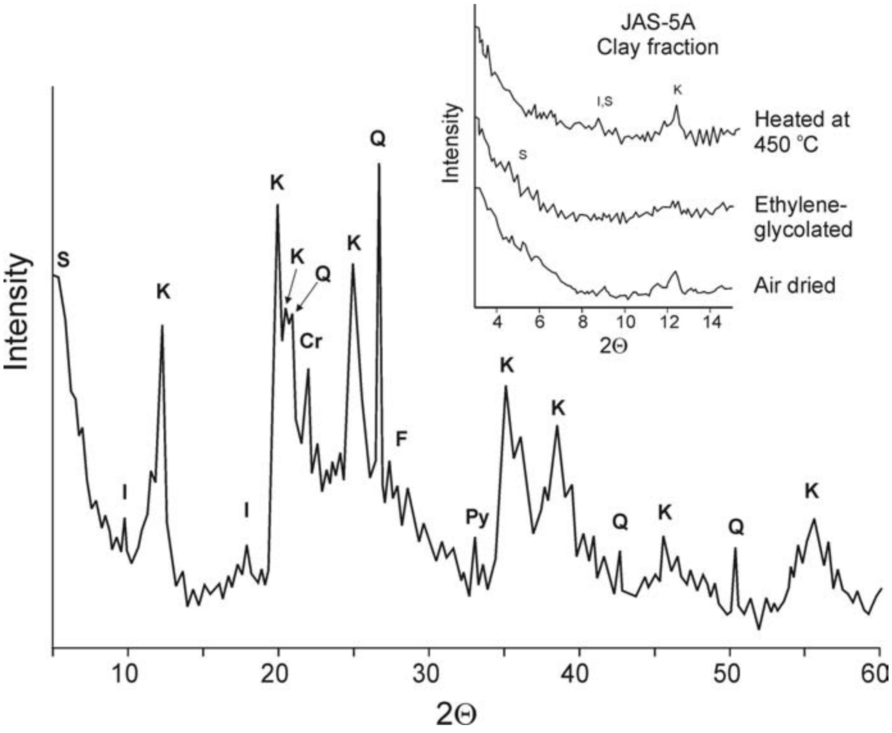

Fig. 5. XRD pattern of the kaolinite sample JAS-5A.
K – kaolinite,
S – smectite,
I – illite,
Q – quartz,
Cr – crystobalite,
F – feldspar,
Py – pyrite.

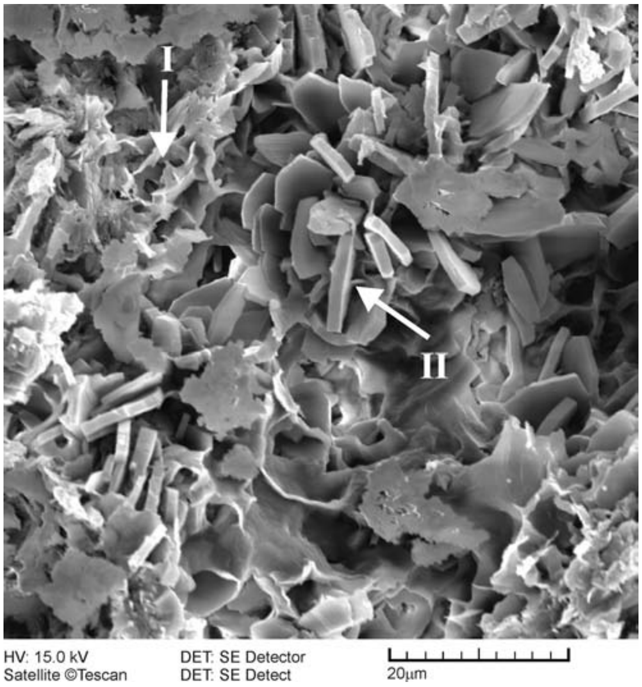

Fig. 6. SEM image of the kaolinite sample JAS-5A, showing two different particle generations (I, II).

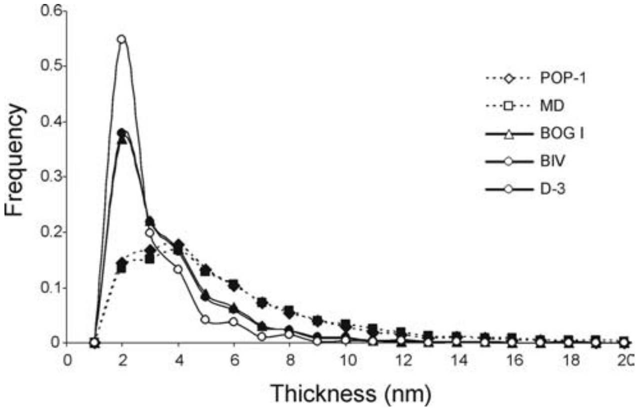

Fig. 7. Crystallite size distribution of the smectites obtained by the BWA technique.

Conclusions

Sedimentary kaolinites from five deposits in Serbia have a low mean crystallite thickness indicating a poorly developed kaolinitic weathering crust from which these clays were redeposited, a situation similar to Slovak kaolin deposits. The role of crystal disintegration during transport may also influence the crystallite size. The shape of the crystal size distribution is polymodal for all samples, most probably as a result of the presence of different kaolinite generations.

Volcano-sedimentary (diagenetic) tonstein from the Jasenovac coal mine has a very low mean crystallite thickness, typical for a weakly crystallized material, and a polymodal distribution shape, due to at least two generations of kaolinite particles.

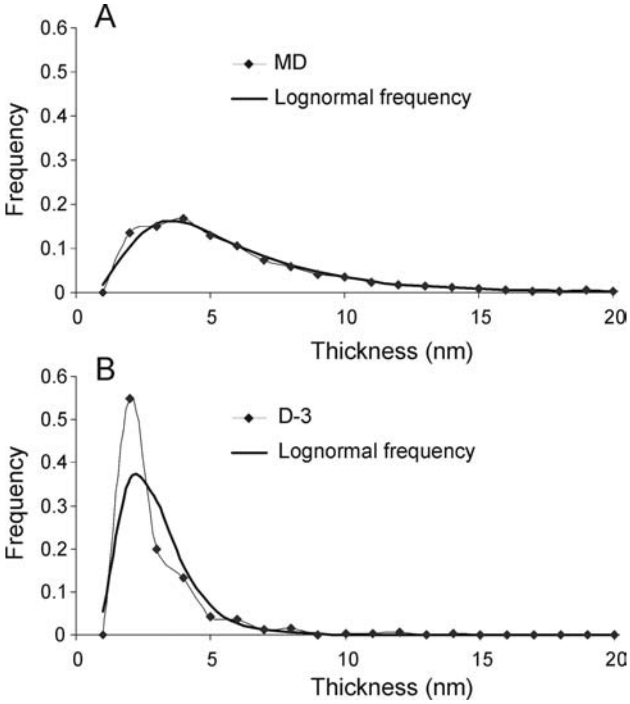

Fig. 8. Comparison of the measured (BWA) and the theoretical lognormal distributions of smectite particles. A) volcano-sedimentary type, B) sedimentary type.

Hydrothermal Novo Brdo halloysite also has a very low mean crystallite thickness and a polymodal distribution.

Two diverse shapes of the theoretically lognormal distributions were observed for the smectites. They correspond to different genetic types of bentonites – sedimentary and volcano-sedimentary. The mean crystallite thickness is also different in the sedimentary and volcano-sedimentary bentonites, with an average T_{BWA} of 5.56 and 9.56 nm, respectively. This means that "in situ" alteration of volcanic ash under subaqual conditions led to the formation of well-crystallised smectite with thicker crystallites.

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Резиме

Дистрибуција дебљине кристалита минерала глина из изабраних лежишта глина Србије

У овом раду измерене су дебљине кристалита каолинита и смектита и њихова дистрибуција методом BWA на узорцима из лежишта у Србији. Након извршених проучавања направљена је компарација са сличним глинама из Словачке и из неких светских лежишта.

Репрезентативни узорци 6 каолинитских глина, једног халојзита и 5 бентонита из лежишта ствараних у различитим геолошким условима: седиментним лежиштима глина (створених у корама распадања а затим транспортована у седиментне басене), вулканогено-седиментним лежиштима (насталим дијагенезом вулканског пепела у подводним или приповршинским условима) и хидротермалним лежиштима. Генетски тип лежишта и основне геолошке информације приказане су у табели 1.

Седиментни каолинити из пет проучаваних лежишта Србије одликују се малом дебљином кристалита, што указује на слабо развијен тип кора распадања из којих су ти минерали глина преталожени, као и на утицај уситњавања честица приликом транспорта. Таква ситуација је веома слична као код лежишта каолина у Словачкој. Облик криве расподеле дебљине кристалита је код свих узорака полимодални, највероватније као последица присуства различитих генерација каолинита.

Вулканогено-седиментни (дијагенетски) тонштајн из лежишта угља Јасеновац показује веома малу дебљину кристалита, типичну за слабо искристалисали материјал, и полимодални облик дистрибуције захваљујући присуству најмање две генерације каолинитских честица.

Хидротермални халојзит из лежишта Ново Брдо такође се одликује веома малом дебљином кристалита и полимодалним обликом расподеле.

Код испитиваних бентонита утврђено је присуство два различита типа теоретски логнормалне расподеле дебљине кристалита, које одговарају различитим генетским типовима бентонита – седиментном и вулканогено-седиментном. Средња дебљина кристалита је такође различита код седиментних и вулканогено-седиментних бентонита, са просечним T_{BWA} од 5,56 и 9,56 nm.

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Belgrade, December 2006

In memoriam

Проф. др Предраг Николић
(1928 – 2005)

Прошле године преминуо је Предраг Николић, доктор геолошких наука, редовни професор у пензији, дугогодишњи Декан Рударско-металуршког факултета у Бору и Декан заједничког Рударско-геолошко-металуршког факултета Београд-Бор Универзитета у Београду.

То су само најосновнији подаци из блиставе каријере примерног човека, врхунског професора и научног радника, који је много тога оставио по чему ће га се сећати генерације студената, његови сарадници, колеге и пријатељи.

Немерљив је допринос Професора Николића за формирање Рударско-металуршког факултета у Бору и његов развој. Мање су познате и велике заслуге за развој Рударско-геолошког факултета у Београду на коме је Професор Николић провео другу половину своје каријере. По мишљењу многих колега био је један од најбољих, а можда и најбољи Председник Државне комисије за оверу минералних резерви. Поред тога Професор Николић нам оставља вредан научни опус од 146 објављених научних радова, 6 уџбеника и 4 монографије. Написано већ довољно говори о изузетној личности професора Николића. Простор нам не дозвољава детаљан приказ животног пута и дела професора Николића па ће он бити скроман, управо онакав какав је био профеоср.

Професор Николић је рођен 31. јануара 1928. године у Горњем Дреновцу код Прокупља, где је завршио гимназију. На геолошко-палеонтолошкој групи Природно-математичког факултета Универзитета у Београду дипломирао је 1952. године, а докторирао 1962. године. Радну каријеру је почео 1953. године као руднички геолог и руководилац геолошко-рударских истраживања у руднику угља “Добра срећа” код Књажевца, затим у рудницима ватросталних глина – Аранђеловац и књажевачким, односно тимочким рудницима угља “Књажевац”. Од 1962. године ради на Рударско-металуршком факултету у Бору као наставник, и то најпре као доцент, затим као ванредни професор и од 1969. године као редовни професор. Од 1980. године је у радном односу на Рударско-геолошком факултету у Београду. Био је 11 година руководилац Рударско-металуршког факултета у Бору, напре као продекан, затим у континуитету као декан. Био је две године заменик декана јединственог Рударско-геолошко-металуршког факултета Београд-Бор и две године декан истог факултета. Остаће упамћен и његов рад као Председника одбора за “Геонауке” у Министарству за науку и технологију Републике Србије. Био је дугогодишњи Главни и Одговорни уредник научног часописа Зборник радова Рударско-металуршког факултета и Института за бакар у Бору. Руководио је и Одбором за високо школство у Заједници усмереног образовања и био члан Председништва исте заједнице. За свој успешан рад стизала су и признања. Прво, 1967. године, Октобарска награда Града Бора, а затим 1977. године Орден рада са златним венцем.

Научни и стручни опус професора Николића је веома обиман и разноврстан па га није лако и једноставно приказати. У основи, био је геолог (стратиграф, тектоничар), али је врхунске резултате остварио и у домену истраживања лежишта минералних сировина, посебно угљева. Слободни смо да кажемо, ако је прва геолошка љубав професора Николића била Тимочка зона, онда је друга љубав угаљ.

Незаобилазан је и велики допринос професора Николића за сагледавање геолошке грађе и тектонског склопа, али и природне потенцијалности наших Карпато-балканида. Посебно се то односи на једну од најинтересантнијих и најсложенијих геолошких јединица у литератури познатој као Тимочка зона. Почев од докторске дисертације у којој је на оригиналан начин приказао стратиграфске односе, посебно горњокредних и палеогених творевина, затим тектонски склоп и магматизам једног дела ове зоне. Потом, следе десетине радова у којима професор Николић шири своја истраживања на целу зону. На крају, све то обједињује у једну врхунску монографију “Тимочка зона Источне Србије – геологија и минералне сировине”. Довољно је само набројати поглавља из ове изузетне монографије па да се сагледа њена свеобухватност и значај. У првом поглављу је детаљан приказ стратиграфије, структурних односа и тектонике, као и критички осврт на досадашња истраживања ове зоне. Друго поглавље обрађује лежишта угља и уљних шкриљаца у јужном делу ове зоне, детаљно по локалитетима – рудницима. Приказ обухвата од геологије до резерви. У трећем поглављу детаљно су обрађена лежишта бакра у северном делу Тимочке зоне. Бројни су радови у којима је обрађивао лежишта угљева у Источној Србији. Довољно је поменути Влашко поље, Добра срећа, Подвис, Лубница, Звездан др. Као врхунски истраживач, сва своја, и истраживања својих сарадника, обједињује у три вредне монографије о угљевима. Професор Николић је један од најзаслужнијих аутора за објављивање дванаестотомене едиције “Геологија Србије” и десетотомног енциклопедијског речника “Геолошка терминологија и номенклатура”.

Посебно издвајамо као непревазиђен монографски рад “Тектоника Карпато-балканида Србије”, који је професор Николић објавио заједно са својим, по нашем мишљењу, најближим колегом и пријатељем, покојним професором Милодрагом Анђелковићем. Реч је о најдетаљнијем приказу стратиграфије и тектонике наших Карпато-балканида. То је незаобилазна литература за наше, али и и за геологе суседних земаља.

Научни опус професора Николића повезује једна лепа и вредна дивљењу научна нит. Основне проблеме које је у почетку своје каријере почео да ради, после вишегодишњих истраживања је обједињавао и сублимирао у монографијама.

Наставни рад за професора Николића био је светиња. Наставник је дужан да обезбеди уџбеничку литературу – био је његов став. Имао је неколико издања универзитетског уџбеника Основи геологије и Опште геологије и први уџбеник на нашем језику из Геотектонике.

Све то говори да је професора Николића одликовала огромна енергија, вредноћа, упорност, креативност и систематичност, па ето одговора за сјајну професорску, научно-истраживачу и људску каријеру.

Нама који смо имали привилегију да са њим пијемо кафу, а понекад и нешто жешће, и причамо не само о геологији, ово написано није потребно, оно је уписано у нама, али ради млађих и оних који долазе, ради Рударско-геолошког факултета и српске геологије име професора Предрага Николића треба да буде уписано крупним – најкрупнијим словима.

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Book Review:

Mittelmiozäne Ostracoden aus dem Wiener Becken (Badenium/Sarmatium, Österreich)

[Middle Miocene Ostracods from the Vienna Basin (Badenian/Sarmatian, Austria)]

by Martin Gross

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This book is a typical paleontological work concerning the Middle Miocene ostracodes from the Vienna Basin and their taxonomy and biostratigraphy. It consists of the following chapters: (1) Introduction, (2) Regional Geology, (3) Material and Sample Preparation, (4) Systematic part, (5) Biostratigraphy and (6) References. Technically, it is a correctly prepared book and illustrated with very high quality figures (6), tables (4) and plates (55).

All the samples originate from the area of Bad Deutsch Altenburg – Hainburg/Donau (eastern margin of the Vienna Basin, 40 km ESE from Vienna). 64 taxa (species) are described and illustrated on 55 plates with 591 SEM-microphotographs. It represents 37 genera and 15 families of the Order Podocopida. Very good quality, high-resolution SEM pictures document the intraspecific variability, juvenile forms and sexual dimorphism of these species as well as the carapace/valve ornamentation.

The terms and morphological features of the ostracodes use here are based on Moore (1961), Morkhoven (1962), Hartmann (1966), Oertli (1985) and Hinz-Schallreuter & Schallreuter (1999). Paleoenvironmental and paleobatimetric interpretations were made on the basis of criteria by Morkhoven (1963), Hartmann (1975), Liebau (1980), Gramann (2000) and Meisch (2000). The systematic part was based on Liebau (1975) and Maddocks & Steiner (1987), as well as Hartmann & Puri (1974) and Meisch (2000).

Based on this study, the author demonstrates that some earlier known ostracodes have a wider stratigraphic range (for example, *Callistocythere postvallata* and *Hemicytheria omphaloedes*). Likewise, *Aurila hispidula*, *Xestoleberis tumida* and *Tenedocythere sulcatopunctata* were discovered for the first time from the Lower Sarmatian.

From the (paleo)ecological point of view, the ostracodes association are grouped into four taphocoenosis and the first three correspond to epineritic, epineritic/phytal taphocoenosis and the last one to epi/mesoneritic taphocoenosis.

In comparison to the foraminifer fauna, most of the ostracod samples (59) belong to the Upper Badenian and only 7 samples are assigned to the Lower Sarmatian. In the chapter of Biostratigraphy (5), one comparative biostratigraphical review of the Middle Miocene (Badenian and Sarmatian) of Central Paratethys is given based on the ostracodes and foraminifers biozonation (after Jiricek & Riha, 1991; Zelenka, 1990). The author distinguished eight ostracod biozones (NO 7 – NO 14), which correspond to the six-foraminifer biozones. This study of ostracodes is complementary to the early-adopted model of ostracod development in the Vienna Basin.

Finally, the very detailed and wide-ranging list of references (341) shows that the author employed both classic papers (Muenster, 1830; Roemer, 1838; Reuss, 1850; Bosquet, 1852, etc.) as well as modern literature.

At the end of this review, a few important remarks can be given:

This book represents an important contribution to study of ostracodes of the Middle Miocene of the Vienna Basin, as well as of the Central (Western) Paratethys area. It is a very high quality, distinct article, both professionally and technically, with numerous SEM-microphotographs which give this book a high professional level. Eventually, some paleo(environmental) interpretations and conclusions are discussed. The absolutely competent approach to the research of fossil ostracodes, as well as the results given here, make this monograph a new upgrades for future taxonomic and biostratigraphic studies of the Middle Miocene of Paratethys.

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Owen, E.F., 1962. The brachiopod genus *Cyclothyris*. *Bulletin of the British Museum (Natural History), Geology*, 7 (2): 2–63.

Rabrenović, D. & Jankičević, J., 1984. Contribution to the study of Albian near Topola. *Geološki anali Balkanskoga poluostrva*, 48: 69–74 (in Serbian, English summary).

Smirnova, T.N., 1960. About a new subfamily of the Lower Cretaceous dallinoid. *Paleontologicheskij Zhurnal*, 2: 116–120 (in Russian).

Sulser, H., 1996. Notes on the taxonomy of Mesozoic Rhynchonellida. *In*: Cooper, P. & Jin, J. (eds.), *Brachiopods*, 265–268. Balkema Press, Rotterdam.

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ГЕОЛОШКИ АНАЛИ БАЛКАНСКОГА ПОЛУОСТРВА
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19. Possible pitfalls in the procedure for paleobiodiversity-dynamics analysis.

20. Bryozoan diversity in Southern Siberia at the Devonian-Carboniferous transition: new data confirm a resistivity to two mass extinctions.

21. How global are the Jurassic-Cretaceous unconformities?

Diversity dynamics of Early–Middle Jurassic brachiopods of Caucasus, and the Pliensbachian–Toarcian mass extinction

DMITRY A. RUBAN

Ruban, D.A. 2004. Diversity dynamics of Early–Middle Jurassic brachiopods of Caucasus, and the Pliensbachian–Toarcian mass extinction. *Acta Palaeontologica Polonica* 49 (2): 275–282.

Taxonomic diversity of NW Caucasus brachiopods changed cyclically in the Early–Middle Jurassic. Diversifications took place in the Late Sinemurian–Early Pliensbachian, Middle–Late Toarcian and Late Aalenian–Early Bajocian, while diversity decreases occured in Late Pliensbachian–Early Toarcian, Early Aalenian and Late Bajocian. Outstanding diversity decline in the Late Pliensbachian–Early Toarcian corresponds to a global mass extinction interval, whose peak has been documented in the Early Toarcian. Similar diversity changes of brachiopods are observed in other Tethyan regions, including the well-studied Bakony Mountains, although in NW Caucasus the recovery after demise have begun earlier. The causes of Pl-To mass extinction in the studied region are enigmatic. Probably, it could be linked to anoxia, but its correspondence to the beginning of transgression is not coincident with the global record, so eustatic causes seem to be doubtful for this region.

Key words: Brachiopoda, taxonomic diversity, mass extinction, Jurassic, Caucasus, Russia.

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Introduction

The Late Pliensbachian–Early Toarcian (Early Jurassic) mass extinction is one of the most significant events in geological history. Although it is often called a "small" mass extinction in relation to such great biotic crises as Permian–Triassic or Cretaceous–Tertiary ones, it has been recognized for many principal groups of fossil organisms: foraminifers, bivalves, brachiopods and ammonoids (Hallam and Wignall 1997). Moreover, it provides an excellent example to study possible causal relations of mass extinctions to anoxia, major sea-level changes and intensive volcanism. The knowledge on the Early Jurassic mass extinction has grown thanks to the studies of Hallam (1961), who analyzed the fossil record, and Jenkyns (1988), who identified oceanic anoxia. In the past decade numerous publications on Liassic biotic changes by Hori (1993), Bassoullet and Baudin (1994), Little and Benton (1995), Aberhan and Fürsich (1997, 2000), Hallam and Wignall (1997, 1999), Harries and Little (1999), Guex et al. (2001), Wignall (2001), Jenkyns et al. (2002), and others have substantially added to previous results. A special study was dedicated to establishing the precise absolute radiometric age of the mass extinction, which is now accepted as 183±2 Ma (Pálfy et al. 2002).

Although one of the main fossil groups used to dicsuss the Late Pliensbachian–Early Toarcian (Pl-To) mass extinction are Bivalvia (Aberhan and Fürsich 1997, 2000; Harries and Little 1999), Brachiopoda seem to be no less important. Their significance in discussion of this decimation has been shown in articles of Hallam (1987), Bassoullet and Baudin (1994), and Vörös (1993, 1995, 2002).

Every mass extinction should be studied both using comparative global and regional data analyses to understand better its appearance in geological time and space. Results of quantitative analysis of taxonomic diversity dynamics of Early–Middle Jurassic brachiopods of Northwestern Caucasus are presented in this paper. The appearance of a biotic crisis in this Tethyan region allows one to extend the area where this mass extinction can be documented.

Geological setting

In the Mesozoic the Caucasus region was located on the northern active periphery of Tethys (Lordkipanidze et al. 1984; Meister and Stampfli 2000). Although brachiopods have been found in Jurassic deposits in this entire region, the most complete data are from NW Caucasus (Fig. 1). In tectonical aspect this territory covers 3 major tectonic zones: Labino-Malkinskaja zone, Pshekish-Tyrnyauzskaja zone and Arkhyz-Guzeripl'skaja zone (Rostovcev et al. 1992).

The Lias to Dogger stratigraphy and lithology of NW Caucasus have been described by Krymgolc (1972), Rostovcev et al. (1992), Granovskij et al. (2001), Ruban (2002), and also in *Stratigrafičeskij slovar'* (1979). The Rhaetian–Hettangian interval corresponds to a major regional hiatus. Smaller hiatuses divide Pliensbachian from Toarcian strata and Aalenian from Bajocian ones. Sinemurian–Pliensbachian deposits about 500 m thick include conglomerates, sandstones, dark shales and rare limestones. Lower–Middle (lower interval) Toarcian deposits (up to 70 m thick) are

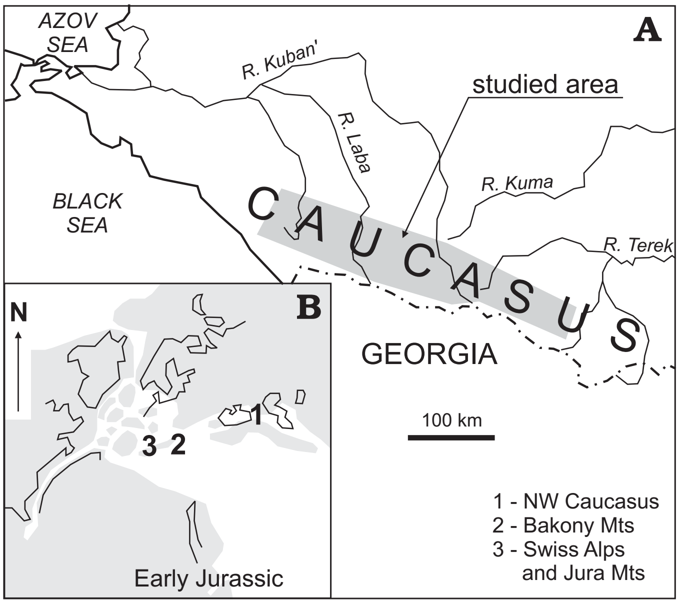

Fig. 1. Geographic location of studied area (**A**) and paleogeographic situation of areas discussed in the text (**B**; studied area indicated by 1). Paleogeographic base map is modified from Owen 1983 and Dommergues et al. 2001.

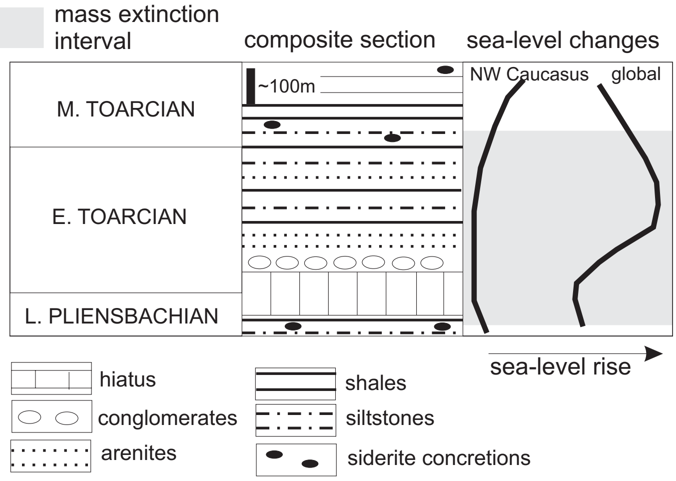

Fig. 2. Liassic composite lithological section of NW Caucasus and comparison of regional and global (simplified from Hallam and Wignall 1999) sea-level changes.

terrigenous (Fig. 2), but in the Middle (upper interval)–Upper Toarcian and Aalenian 1500 m of laminated dark to black shales have been accumulated. Bajocian deposits are represented by dark shales (up to 1200 m), often with interbeds of terrigenous or volcanogenous deposits. Dark shales contain siderite concretions. The highest concetration of them has been observed in the Middle–Upper (lower interval) Toarcian (Ruban 2002). But both rare dark shales horizons and concretions have been found also in the Middle (lower interval) Toarcian deposits.

The Jurassic paleoenvironments of the studied territory have been described in thanks of geochemical studies and numerous paleogeographical reconstructions by Âsamanov (1978). The Sinemurian sea was relatively shallow with normal salinity, the climate was subtropical and humid, and in the Pliensbachian the basin became deeper. Regression took place in Late Pliensbachian–Early Toarcian times followed by a fast transgression that started in the end of Middle Toarcian and in the Late Toarcian–Early Aalenian marine basin was deep. Dark shales of this interval are interpreted as slope deposits (Granovskij et al. 2001). The paleotemperatures of sea-water were about 20–22°C. Regression occured in the Aalenian, and the climate became more moderate. Paleotemperatures decreased to 10°C, while in the Late Aalenian they were once again about 20–23°C. In Bajocian the sea became deep and the climate returned to be subtropical and humid. Paleotemperatures identified were 25–27°C. In general terms, during all the Early–Middle Jurassic the Caucasus basin was not separated from the Tethyan ocean too much.

Materials

Early to Middle Jurassic brachiopods of NW Caucasus are rather well-studied. The first complete revisions have been made by Neumayr and Uhlig (1892), and then by Moiseev (1934) followed by the works by Makridin and Kamyšan (1964), Prosorovskaya (1986, 1989, 1993), and Rostovcev et al. (1992).

The most complete and taxonomically revised data on stratigraphic distribution of brachiopods have been presented by Makridin and Kamyšan (1964), Prosorovskaya (1989, 1993) and Rostovcev et al. (1992). These sources have been used to compile the diversity data, following by some revisions and stratigraphical corrections. In the whole 83 species from the Lower Jurassic and 51 ones from Middle Jurassic were recognized (see Appendix). The stratigraphical ranges of taxa are as detailed as the substages recognized in the ammonite succession (Rostovcev et al. 1992).

Total species quantity, and number of originations and extinctions have been calculated for each substage from Sinemurian to Bajocian to analyze the taxonomic diversity dynamics.

Species diversity dynamics

Species diversity of brachiopods changed dramatically in NW Caucasus (Fig. 3). After the great diversification in the Sinemurian–Early Pliensbachian (up to 54 taxa), a severe diversity decline occured in Late Pliensbachian. This trend continued into the Early Toarcian, when brachiopods disappeared entirely. In the Middle–Late Toarcian a new gradual diversification took place (total diversity reached 14 taxa), although it did not compensate for the previous decline. Once again brachiopods disappeared in the Early Aalenian (probably two taxa are

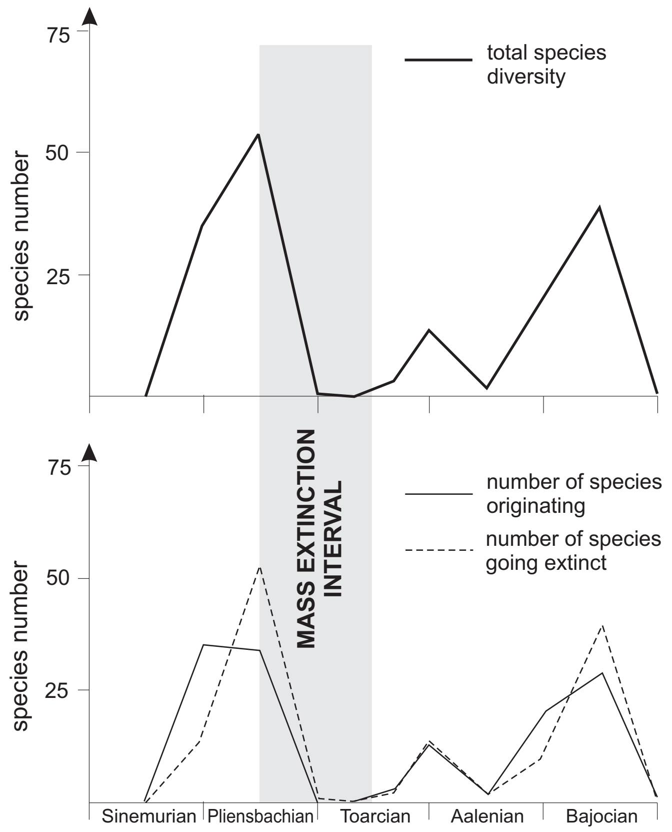

Fig. 3. Total species diversity changes, origination and extinction rates of NW Caucasus brachiopods in Early–Middle Jurassic.

known). Then a new acceleration of diversity is observed during Late Aalenian–Early Bajocian. Later Middle Jurassic brachiopod communities became extinct again. But the most outstanding features of the regional brachiopod history are the Late Sinemurian–Early Pliensbachian acme followed by Late Pliensbachian–Early Toarcian demise.

The origination and extinction rates help us to understand the causes of such turnovers (Fig. 3). Late Sinemurian–Early Pliensbachian diversification has been followed by increase both of origination and extinction. When the disappearance rate became higher, a regional diversity drop occured. Also the Late Pliensbachian–Early Toarcian demise of brachiopods corresponded to collapse of origination rate. The Early Aalenian and Late Bajocian crises also can be explained in the same way.

An analysis of events connected with disappearance of fossils should always take into account so-called Lazarus-taxa (Flessa and Jablonski 1983; Jablonski 1986; Wignall and Benton 1999; Fara 2001). The Lazarus effect causes a significant gap in any taxon's range. Among Early–Middle Jurassic brachiopods of the studied region only three Lazarus-taxa have been found (see Appendix). The interval of their absence covers the Late Pliensbachian–Early Toarcian in the NW Caucasus, and for two of them—also the Middle Toarcian. It is evident, that such a negligible Lazarus effect could not significantly influence the above mentioned results.

Discussion

Regional record of Pl-To crisis.—As shown above, the most intensive drop in brachiopods diversity occured in Late Pliensbachian–Early Toarcian. This time interval approximately corresponded to the mass extinction, whose most intensive phase embraced the Early Toarcian (Hallam and Wignall 1997). As the Caucasus sea was connected with Tethys, there was no barrier to isolate the local biota from the disastrous global factors.

The ecosystem collapse was indeed the most profound in the Early Toarcian, when brachiopods regionally disappeared. However, whilst the decrease of diversity began only in the Late Pliensbachian; the decline of origination rate was documented already in the Early Pliensbachian. In this timespan, the benthos disappearance also intensified, and the initial phase of the crisis took possibly place in the Early Pliensbachian. The recovery of brachiopods has been documented in the Middle Toarcian. But only the intensification of benthic colonization and total species diversity increased (i.e., the repopulation stage) in the Late Toarcian can be considered as a record of the final crisis termination. So, severe extinction have certainly influenced development of the Caucasus brachiopod biota during the Late Pliensbachian–Middle Toarcian.

It is improtant to compare the brachiopods diversity data with those of other shelly benthos like bivalves. The most recent compilative paper devoted to Jurassic pectinoids from the southern regions of the former Soviet Union, including Caucasus, has been published by Romanov and Kasum-Zade (1991). Stratigraphic ranges chart presented by these authors allows the recalculating of the total species quantity changes through the Late Liassic (Fig. 4). After a rather strong diversification of pectinoids in the Pliensbachian a relatively small diversity drop (~30%) is recognizable in the Early Toarcian. The species number was relatively low till the Bajocian when a new rise occured. This diversity changes means: (1) decline of bivalves was not as significant as that of the brachiopods, and (2) it occured after a great diversification event which is similar to those observed for brachiopods.

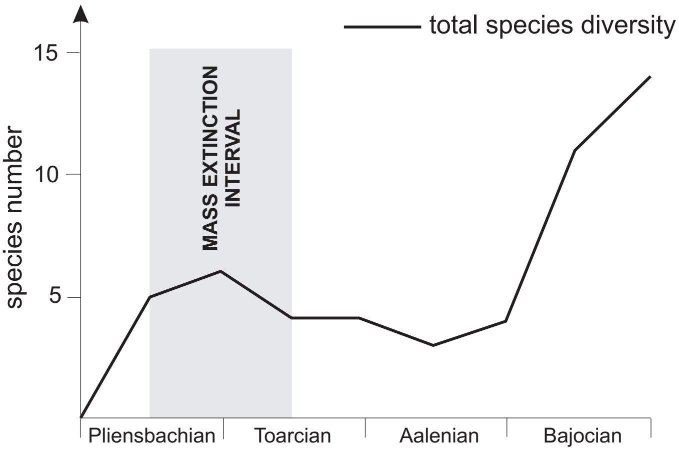

Fig. 4. Diversity changes of Spondylydae, Limidae, Plicatulidae, Anomioidae (Pectinoida) from the South of former Soviet Union (after Romanov and Kasum-Zade 1991).

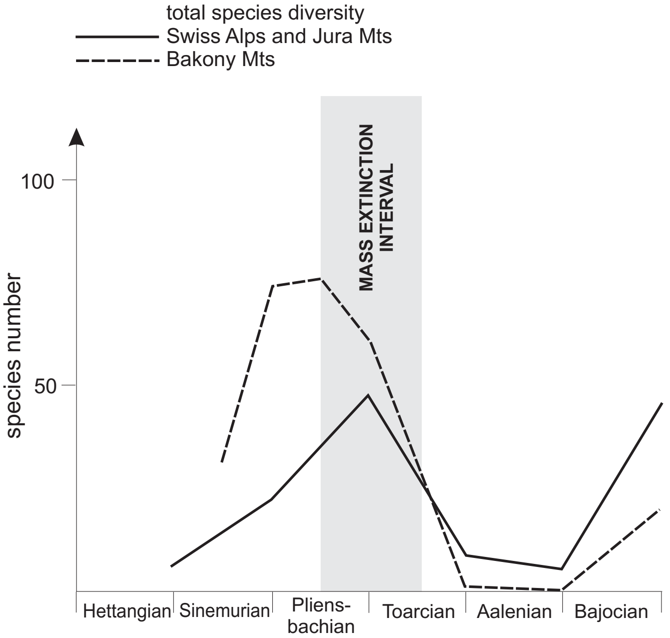

Fig. 5. Brachiopods diversity changes in Lias of Bakony Mts (after Vörös 1993, 1995) and Swiss Alps and Jura Mts. (after Sulser 1999).

Comparison with other regions.—Early Jurassic brachiopod diversity patterns was studied in other Tethyan regions, the best exemplified being in the Bakony Mountains in Hungary (Vörös 1993, 1995). The results show that total species number increased from the Hettangian to the Early Pliensbachian (Fig. 5). Then in the Late Pliensbachian a decrease began, while through the Toarcian–Aalenian interval brachiopods disappeared almost entirely. Origination rate began to decrease in the Carixian, and extinction rate increased the same time but more intensively in the Domerian. A new radiation has been documented in the Bajocian.

Diversity dynamics of Swiss Alps and Jura Mountains brachiopods could be calculated from data of Sulser (1999). Although in this monograph only partial stratigraphical ranges are shown (i.e., where taxa are abundant), the Sinemurian–Pliensbachian diversification phase, and Toarcian–Aalenian diversity fall are evident as well (Fig. 5).

Therefore, temporal diversity patterns of Tethyan brachiopods are very similar to those of NW Caucasus brachiopods although a recovery of diversity in the Late Toarcian shows a shorter crisis duration. But another feature is common for these regions: the mass extinction occurred after significant diversification, which seems to be connected with a radiation phase.

Causes of the shelly benthos collapse in NW Caucasus

The Early Jurassic mass extinction is thought to be causally connected mainly with anoxia and major sea-level changes (see reviews by Hallam and Wignall 1997, 1999; Wignall 2001). As the Pliensbachian–Toarcian boundary transition is marked in Caucasus by a hiatus, it is considered that at this time a local uplift took place; after the accumulation of Early Toarcian inner shelf sediments, the progressive deepening has begun in the Middle Toarcian (Granovskij et al. 2001).

The global sea-level curve (Haq et al. 1987; Hallam 1988, 2001; Hallam and Wignall 1999) shows eustatic rise that occurred distinctly earlier (in the Early Toarcian), and a regressive phase has already begun in the Middle Toarcian. A difference between the NW Caucasus and global sea-level record is evident (Fig. 2), while the results of diversity studies are similar. So, the Pl-To extinction among brachiopods in the studied region possibly should be explained by additional causes. It is necessary to note that previous attempts of explanations of Pl-To biotic crisis by sea-level fluctuations led to conclusion about the absence of direct links between these phenomena (McRoberts and Aberhan 1997).

Anoxic conditions are well established world-wide in Toarcian seas (Jenkyns et al. 2002). Although the broad shallowing trend has been observed, oxygen-depleted conditions can be tentatively inferred due to black colour of shales, common occurrence of siderite concretions and pyrite grains presence (Fig. 2). The dark colour may be interpreted as a result of the organic lamination. In the sections exposed along the Belaja River, the author has found interlayers fully consisted of siderite concretions in terrigenous deposits of the Lower–Middle (lower interval) Toarcian succession (Ruban 2002). Meantime, siderite is not considered as a direct indicator of anoxia.

It is necessary to point out that taking into consideration different regional factors may lead us to the correction of assumptions presented above; e.g., regional subsidence if it was intense could have disguised any local evidence of anoxia. The accounting of the same factor also may enforce us to reestimate the importance of transgressions–regressions. So, this a matter for further studies and discussions.

Nevertheless, the Caucasus record gives another opportunity to analyze the causes of possible anoxia themselves in the light of a recent scenario by Guex et al. (2001). The marine oxygen depletion is connected with the terrestrial plants diversification in the Early–Middle Liassic and linked with extensive land inundation during a transgressive pulse, when this accumulated organic material was able to initiate the anoxia. Macrofloral remains in the Domerian–Early Toarcian deposits are abundantly found in all studied sections (Ruban 2000), which might indicate a major plant diversification just before the beginning of Pl-To mass extinction. The same temporal link is observed in the global record (Philippe et al. 1999).

Conclusions

Three diversification episodes of brachiopod faunas, followed by diversity decline, have been documented in NW Caucasus marine basin, and the benthos diversity changed cyclically in the Early–Middle Jurassic. But the most dramatic change occurred in Late Sinemurian–Early Toarcian when diverse brachiopod communities experienced a severe extinction. Similar diversity patterns have been reported from coeval biota in other Tethyan regions (e.g., Vörös 1993, 1995). Such benthos collapse was obviously related to killing environmental factors operating globally through the Late Pliensbachian to the Early Toarcian (?anoxia), even if the direct causal mechanisms remain still enigmatic in the region studied. Expanding oxygen deficiency is a main candidate, and frequently discussed as a main environmental trigger of Palaeozoic brachiopod demises (e.g., Racki 1998; see also Hallam and Wignall 1997; Harper and Rong 2001; Rong and Shen 2002).

In comparison to the Bakony Mts, Jura Mts., and Swiss Alps, brachiopod recovery began earlier in this northern Tethyan domain. Therefore, the regional diversity pattern of the Early Jurassic biotic crisis could be traced even more clearly, and this is a promising perspective for elaboration of other fossil groups. The Caucasus sequences can serve a "reference" for further high resolution stratigraphical and ecological-geochemical studies of this biotic turning point in the Mesozoic.

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Appendix

List and stratigraphic ranges of Liassic Brachiopoda of Northwestern Caucasus (compiled from Makridin and Kamyshan 1964; Prosorovskaya 1989, 1993; Rostovtsev et al. 1992). S2, Sinemurian; P1, Early Pliensbachian; P2, Late Pliensbachian; T1, Early Toarcian; T2, Middle Toarcian; T3, Late Toarcian.

Species	S2	P1	P2	T1	T2	T3
“Rhynchonella” babala Dumortier						+
“Rhynchonella” latifrons Geyer		+				
“Rhynchonella” reinesi Gemmelaro	+					
“Rhynchonella” ex gr. *obtusiloba* Rollier					+	
“Spiriferina” moeschi Haas	+	+				
Aulacothyris griffini Rollier						+
Aulacothyris resupinata (Douville)		+				
Aulacothyris salgirensis Moissejev		+				
Aulacothyris waterhousi (Davidson)		+				
Bodrakella aff. *bodrakensis* Moissejev		+				
Calcirhynchia plicatissima (Quenstedt)	+	+				
Caucasorhynchia visnovskii Moissejev		+				
Cincta leptonumismalis (Rollier)		+				
Cincta cf. *numismalis* (Valenciennes)		+				
Cirpa borissiaki Moissejev		+				
Cuersithyris radstockensis (Davidson)		+				
Cuneirhynchia dalmasi (Dumortier)		+				
Cuneirhynchia gussmani (Rollier)		+				
Cuneirhynchia persinuata (Rau)		+				
Cuneirhynchia rauei Rollier		+				
Curtirhynchia sp.						+
Digonella subdigona (Oppel)		+				
?*Disculina liasina* Deslongchamps		+				
Flabellirhynchia lycetti (Davidson)		+				+
Furcirhynchia laevigata (Quenstedt)	+					
Gibbirhynchia curviceps (Quenstedt)		+				
Gibbirhynchia gibbosa Buckman	+	+				
Gibbirhynchia heiningensis Rollier						+
Grandirhynchia capitulata (Tate)						+
Homoeorhynchia deffneri Oppel	+	+				
Homoeorhynchia sp. 1		+				
Homoeorhynchia sp. 2						+
?*Linguithyris bimammata* Rothpletz	+					
Liospiriferina alpina (Oppel)		+				+
Liospiriferina obtusa (Oppel)	+	+				
Liospiriferina rostrata (Schlotheim)	+	+				
Liospiriferina cf. *obtusa* (Oppel)		+				
Lobothyris havesfieldensis Rollier						+
Lobothyris punctata (Sowerby)		+	+			
Lobothyris subpunctata (Davidson)		+				
?*Lobothyris ovatissima* Quenstedt		+				
Piarorhynchia juvenis (Quenstedt)	+					
Piarorhynchia rostellata (Quenstedt)		+			+	
Piarorhynchia triplicata (Phillips)		+				

Species	S2	P1	P2	T1	T2	T3
Piarorhynchia variabilis Davidson		+				
Piarorhynchia aff. *albertii* (Oppel)		+				
Praemonticlarella schuleri (Oppel)						+
Prionorhynchia greppini (Oppel)	+					
Prionorhynchia serrata (Sowerby)					+	+
?*Prionorhynchia regia* Rothpletz		+				
Pseudogibbirhynchia jurensis (Quenstedt)						+
Pseudogibbirhynchia moorei (Davidson)						+
Ptyctorhynchia sp.						+
Rimirhynchia rimosa (Buch)		+				
Rudirhynchia belemnitica (Quenstedt)	+					
Rudirhynchia calcicosta (Quenstedt)	+					
Scalpellirhynchia scalpellum (Quenstedt)	+					
Securina partschi (Oppel)	+	+				
Spiriferina haasi Makridin and Kamyshan	+					
Spiriferina haueri (Suess)	+	+				
Spiriferina ilminsteriensis Davidson	+					
Spiriferina walcotti (Sowerby)	+	+				
Spririferina angulata Oppel		+				
Squamirhynchia squamiplex (Quenstedt)		+				
Tetrarhynchia pontica Moissejev	+					
Zeilleria cornuta Quenstedt	+	+				
Zeilleria davidsoni Makridin et Kamyshan	+	+				
Zeilleria engelhardii (Oppel)		+				
Zeilleria indentata (Sowerby)	+	+				
Zeilleria lunaris (Zieten)		+				
Zeilleria mutabilis (Oppel)	+	+				
Zeilleria ovalis Rollier	+					
Zeilleria ovimontana (Boese)	+	+				
Zeilleria perforata (Piette)	+					
Zeilleria rehmanni (Roemer)	+	+				
Zeilleria retusa Martin	+	+				
Zeilleria roemeri (Schloenbach)	+					
Zeilleria scalprata (Quenstedt)	+	+				
Zeilleria stapia (Oppel)	+	+				
Zeilleria subsphaeroidalis Rollier	+					
Zeilleria thurvieseri Boese	+	+				
Zeilleria vicinalis (Schlotheim)	+	+				

ГЕОЛОШКИ АНАЛИ БАЛКАНСКОГА ПОЛУОСТРВА ANNALES GÉOLOGIQUES DE LA PÉNINSULE BALKANIQUE	66 (2004–2005)	9–15	БЕОГРАД, децембар 2005 BELGRADE, December 2005

The Upper Miocene of the Rostov Dome (Eastern Paratethys): Implication of the chronostratigraphy and bivalvia-based biostratigraphy

DMITRY A. RUBAN

Abstract. The Rostov Dome is located in the south of the Russian Platform. In the Late Miocene this area was embraced by the Eastern Paratethys. The implications of a recently developed Neogene chronostratigraphy to the studied area are discussed. The Sarmatian regional stage corresponds to the upper part of the Langhian, the entire Serravalian and the lower part of the Tortonian global stages; the Maeotian regional stage corresponds to the upper part of the Tortonian and the lowermost horizons of the Messinian global stages; the Pontian regional stage corresponds to most of the Messinian and the lowermost Zanclean global stages. A first Bivalvia-based biostratigraphic framework is proposed for the territory of the Rostov Dome. Five biozones were established within the Serravalian–Messinian: *Tapes vitalianus*, *Cerastoderma fittoni–Cerastoderma subfittoni*, *Congeria panticapaea*, *Congeria amygdaloides navicula* and *Monodacna pseudocatillus–Prosodacna schirvanica*.

Key words: regional stages, chronostratigraphy, biozones, bivalves, Upper Miocene, Eastern Paratethys.

Апстракт Ростовска дома налази се на југу руске платформе. За време касног миоцена ова област припадала је Источном Паратетису. Разматран је значај савремене неогене хроностратиграфије за проучавану област. Сарматском регионалном кату одговара део лангиана, цео серавалиан и доњи део тортона; меотском регионалном кату одговара горњи део тортона и најнижи хоризонти месиниана; понтском регионалном кату одговара већи део месиниана и најнижи занклеан. По први пут се даје биостратиграфија горњег миоцена на основу шкољака за област ростовске доме. У оквиру саравалиан–месиниан установљено је пет биозона: *Tapes vitalianus*, *Cerastoderma fittoni–Cerastoderma subfittoni*, *Congeria panticapaea*, *Congeria amygdaloides navicula* и *Monodacna pseudocatillus–Prosodacna schirvanica*.

Кључне речи: регионални катови, хроностратиграфија, биозоне, шкољке, горњи миоцен, источни Паратетис.

Introduction

The study of the Upper Miocene deposits of the Eastern Paratethys began about 150 years ago (ABICH, 1865; ANDRUSOV, 1884; ANDRUSSOW, 1911), but some important questions on their stratigraphy are still unresolved. Correlation between global and regional stages (1) and the development of macrofauna-based biozonation (2) are among them.

The Rostov Dome is a promising area to study the Upper Miocene stratigraphy of the Eastern Paratethys. It is situated in the southern part of the Russian Platform (Fig. 1). The Upper Miocene sedimentary complexes are wide-spread and cover all its territory. The high abundance of fossil bivalves in the Upper Miocene deposits of the Rostov Dome suggests the use of this group for the development of the first regional biostratigraphic framework. To do this, it was first necessary to implement the recently developed chronostratigraphy of the Upper Miocene in order replace the regional scale of the Eastern Paratethys.

Geological setting

The Rostov Dome is situated in the South of the European part of Russia (Fig. 1). In a tectonical sense, it represents a specific structure in the eastern part of the Precambrian Ukrainian Craton, which itself is a great block of the Russian platfrom (LEBEDKO, 1980;

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POGREBNOV *et al.*, 1970). After its uplift in the Cretaceous–Paleogene, the dome has several times been covered by the sea from the south.

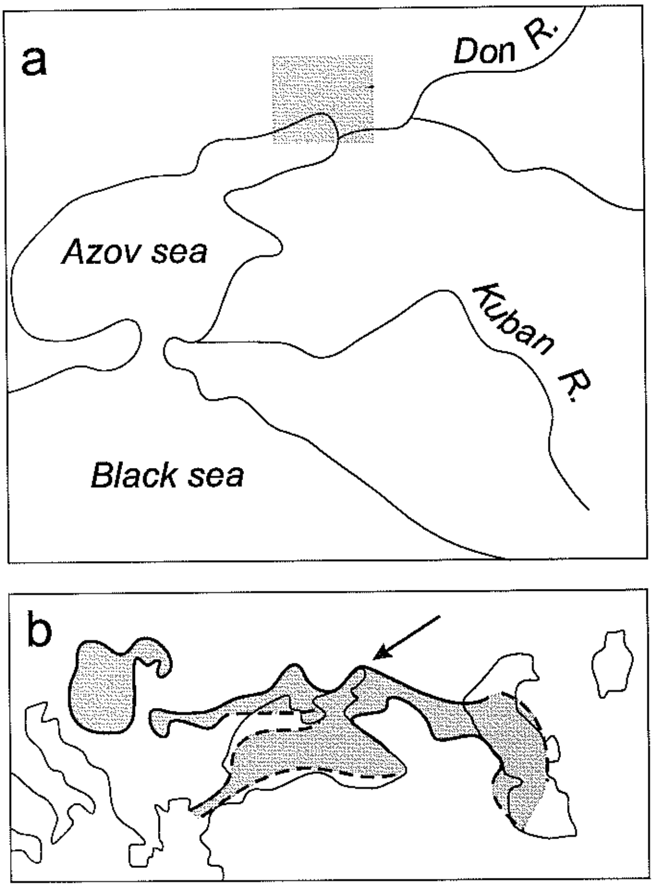

Fig. 1. Geographical (a) and palaeogeographical (b) locations of the studied area (shaded in a and indicated by an arrow in b). Palaeogeographical reconstruction after NEVESSKAJA *et al.* (1984).

In the Neogene, all the south of the European part of Russia was occupied by a large basin. It was a remnant of the previously existing Neotethys, which, after the orogeny in the alpine regions, became divided into 2 parts – the Mediterranean and the so-called Paratethys, consisting of Western (Pannonian), Central and Eastern Basins. The Paratethys originated at the end of the Paleogene, and its isolation strengthened cyclically from the Oligocene until the Pliocene (ILINA *et al.*, 1976; NEVESSKAJA *et al.*, 1984; NOSOVSKIJ, 2001; RÖGL, 1998, 1999; ULANOVSKAYA, 1998). The territory of the Rostov Dome is located at the northernmost periphery of the Eastern Paratethys (Fig. 1). It was embraced by sea during the maximums of cyclically repeating transgressions, when a relatively large and wide Tanaiss palaeobay originated (RUBAN, 2002a). Various sediments accumulated during these times – clays, silts, sands, marls, limes. But the most typical Upper Miocene deposits are skeletal limestones, consisting completely of shells of bivalves, less gastropods and their remains of different size.

Although the Upper Miocene deposits are well-exposed in outcrops within the studied area, they have been investigated only occasionally during the XX century by BOGATCHOV (see RODZJANKO, 1970) and later by RODZJANKO (1970, 1986). Also they have been characterized in few monographs, e.g., PAFFENGOLTS (1959), IVANITSKAJA & POGREBNOV (1962), but these descriptions mostly summarized the results of the above mentioned researchers.

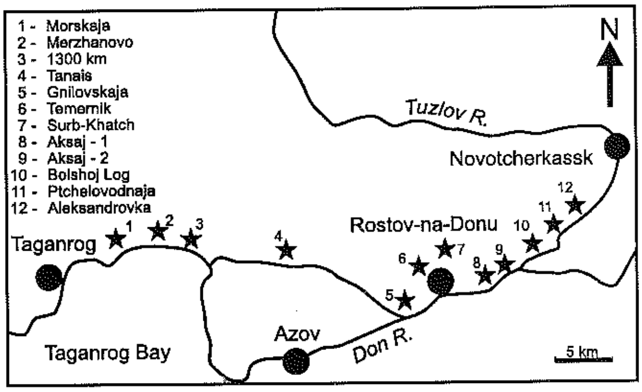

Fig. 2. Location of the studied sections of the Upper Miocene deposits of the Rostov Dome.

The author studied 12 sections of the Upper Miocene strata of the Rostov Dome (Fig. 2) and made a lithostratigraphic framework (Figs. 3, 4). Taganrogskaja, Rostovskaja, Donskaja, Merzhanovskaja and Aleksandrovskaja Formations were formally defined (RUBAN, 2002b). The Janovskaja Formation was established previously by RODZJANKO (1986). Additionally, RUBAN & YANG (2004) proposed a first sequence stratigraphic framework for the Upper Miocene deposits of the Rostov Dome.

Bivalves from the Upper Miocene deposits were studied (RUBAN, 2002b). Identification of the species was made according to general overviews (with taxonomic descriptions and figures) of Neogene bivalves of the Eastern Paratethys presented by ILINA *et al.* (1976) and NEVESSKAJA (1986).

Implication of the chronostratigraphic scale to the Eastern Paratethys

Normalization of the general stratigraphic framework of the Eastern Paratethys, i.e. to correlate global and regional stages, is an important task, because this will enable correlations of biostratigraphic units, which may be defined in the Upper Miocene of the Rostov Dome, to be made to adjacent and even far-located regions.

When in the XIX century differences between the Mediterranean and the Paratethys were established, the general problem of Neogene strata correlation between

Fig. 3. Generalized lithostratigraphy of the Upper Miocene deposits of the Rostov Dome.

these territories appeared. Differences in the stratigraphy between the Western and the Eastern Paratethys arose. In the 1980s and 90s, the Mediterranean stratigraphic scale of the Neogene coupled with world-wide data underwent revision by the International Comission on Stratigraphy (ICS) in order to develop a globally-significant chronostratigraphic scale. This procedure is ongoing, and a new precise chronostratigraphic scale is "under construction". When the development of the recent chronostratigraphic scale began, the difficulties in making a correlation between the global and the regional Eastern Paratethys stratigraphies strengthed again.

The "International stratigraphic Guide" (SALVADOR, 1994) proclaims stages as units with a global sense. Thus, it cannot be defined essentially for a particular region, because the geologic time was not different in the palaeospace. Meanwhile, when stratigraphers begin to define further global units, there is often insuficient data to enable the consideration of globally-recognized horizons in the studied interval. In this way, separate standards of stages appear. Each of them is valid for a single region. The stages in such standards are regional stages.

Nowadays, there are at least two intervals for which regional stages are widely used: Cambrian (PALMER, 1998; ZHURAVLEV, 1995) and Carboniferous (MENNING *et al.*, 2000, 2001; WAGNER & WINKLER PRINS, 1983). Discussions about which regional stages are preferable are ongoing. However, every time, when evidence is obtained which enables larger global units to be defined (as in the case of the Upper Miocene), it is not necessary to use regional stages. It is clear that different chronostratigraphies for particular regions should no

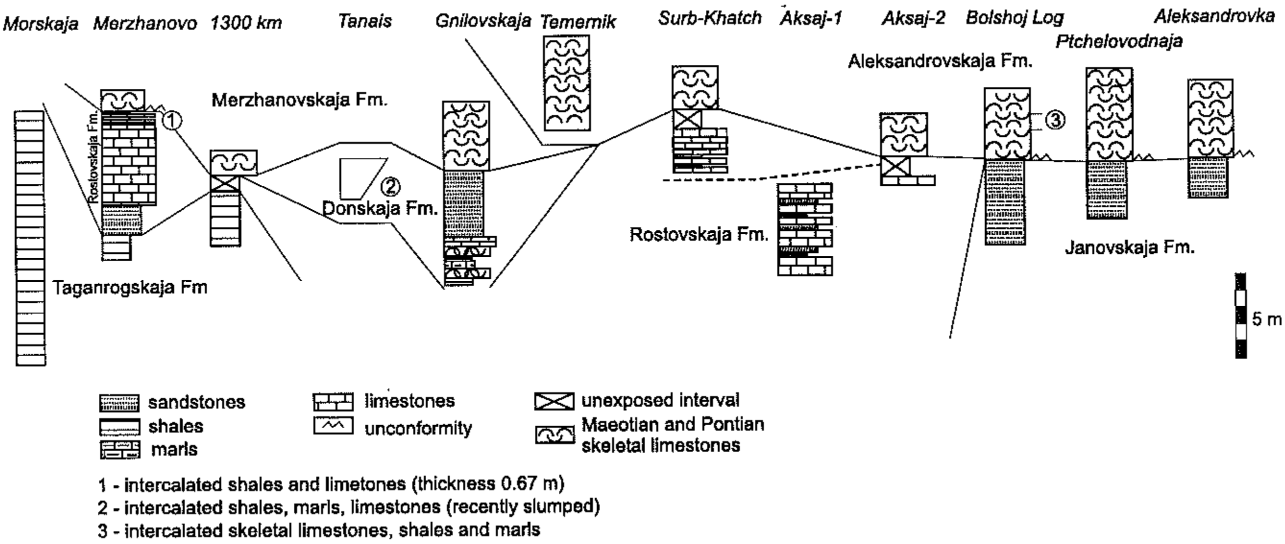

Fig. 4. Correlation of the sections of the Upper Miocene deposits of the Rostov Dome.

longer exist, because the geological time was the same at every point on the Earth's surface. Otherwise, chaotic nomenclature, not representing the true geologic history will result.

In the meantime, Russian stratigraphers traditionally continue to use the regional stratigraphic scale for the Neogene deposits, which includes regional stages differing from chronostratigraphic (i.e., global) stages (Fig. 5). A widely accepted version of such a regional scale was proposed by NEVESSKAJA *et al.* (1984, 1986) and NEVESSKAJA (1986) . Therefore, there is an urgent need to correlate global and regional stages and to implicate the chronostratigraphy to the Eastern Paratethys in order replace the regional standard and abandon it forever.

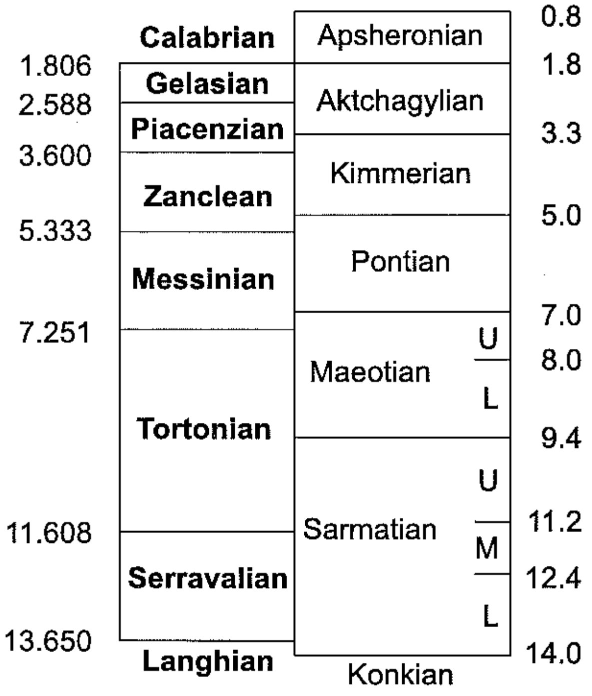

Fig. 5. Correlation between the chronostratigraphic units of the Upper Miocene–Pliocene and the regional stages of the Eastern Paratethys (see text for sources of the absolute ages).

A possible way to correlate the Neogene chronostratigraphic and regional stages is to compare the absolute ages of their boundaries. For the Eastern Parathethys these ages were evaluated precisely by TCHUMAKOV *et al.* (1992), and then discussed several times (TCHUMAKOV, 2000a, b). For the recently employed chronostratigraphic units, the absolute ages are recommended by the ICS (GRADSTEIN *et al.*, 2004) and some of them have been defined in the Global Stratotype Sections and Points (GSSPs) (CASTRADORI *et al.*, 1998; HILGEN *et al.*, 1998, 2000a, b; RIO *et al.*, 1998; VAN COUVERING *et al.*, 2000). For the formal definition of absolute ages of the Messinian, all the Pliocene stages were preferred. ICS recommendations (GRADSTEIN *et al.*, 2004) were used for the Laghian, Serravalian and Tortonian stages.

The results of a correlation by absolute ages (Fig. 5) suggest the Sarmatian regional stage corresponds to the upper part of the Langhian, the entire Serravalian, and the lower part of the Tortonian global stages. The Maeotian regional stage embraces the upper part of the Tortonian and the lowermost horizons of the Messinian global stages. And finally the Pontian regional stage mostly corresponds to the Messinian with only the uppermost part corresponding to the lowermost Zanclean. It is evident, that Miocene/Pliocene boundary, located at the base of the Zanclean in the global scale, has a different position in the Eastern Paratethys, where check meaning the Zanclean is established at the base of the Kimmerian.

Bivalvia-based biostratigraphy of the Rostov Dome

Abundant bivalves remains are the characteristic feature for all the Upper Miocene strata of the Rostov Dome. The analysis of taxa ranges allows the development of the regional biozonation based on this fossil group. Previous studies of the Eastern Paratethys (ILINA *et al.*, 1976; NEVESSKAJA, 1986; NEVESSKAJA *et al.*, 1986) resulted only from malacofaunal support for the regional stages and their substages and from the occasional identification of specific units, called "beds with", which, in fact, are something like acme-zones or assemblage zones. The present study of the Rostov Dome, however, permits the development of a Bivalvia-based biostratigraphy.

The definition of the biostratigraphic units (biozones) was made according to the recommendations of the ICS (SALVADOR, 1994). The difference of terms "first occurence level" (FOL) and "last occurence level" (LOL) from "first appearance datum" (FAD) and "last appearance datum" (LAD) is assumed as the one proposed by PAVIA & MARTIRE (1997). Five distinct biozones have been defined in the Upper Miocene strata of the Rostov Dome (Fig. 6). The correlation established between regional and chronostratigraphic stages helped in the asignment of these biozones to global stages.

***Tapes vitalianus* Interval Zone** corresponds to the interval from the pre-Upper Miocene malacofauna assemblage (not represented in the studied sections) to the LOL of *Tapes vitalianus* ORBIGNY. Further studies are necessary to revise this zone, as its lower boundary is undefined. Age: Langhian–Serravalian; Lower Sarmatian. Reference sections: Morskaja, Merzhanovo.

***Cerastoderma fittoni – Cerastoderma subfittoni* Total Ranges Zone** corresponds to the interval from the FOLs of *Cerastoderma fittoni* (ORBIGNY) and *C. subfittoni* (ANDRUSOV) to the LOLs of these taxa. Age: Serravalian–Tortonian; Middle Sarmatian. Reference section: Merzhanovo.

***Congeria panticapaea* Interval Zone** corresponds to the interval from the FOL of *Congeria panticapaea*

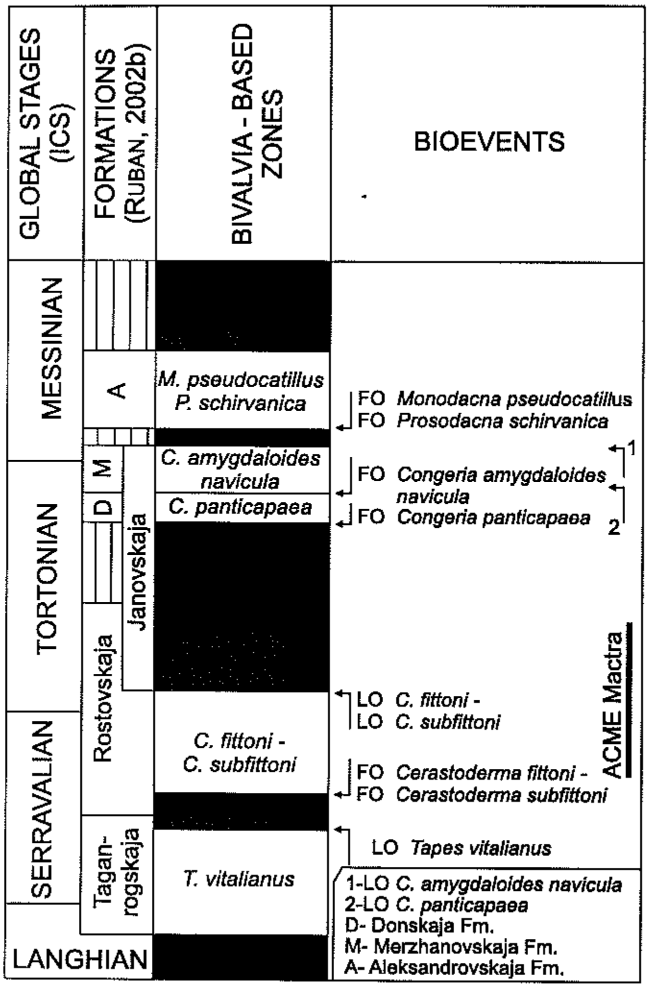

Fig. 6. Proposed Bivalvia-based biostratigraphy of the Upper Miocene of the Rostov Dome.

ANDRUSOV to the FOL of *C. amygdaloides navicula* ANDRUSOV. It is important to note that the LOL of *C. panticapaea* ANDRUSOV is above the upper boundary of this zone. Age: Tortonian; "beds with *C. panticapaea*", lower part of the Upper Maeotian. Reference section: Gnilovskaja.

***Congeria amygdaloides navicula* Total Range Zone** corresponds to the interval from the FOL to the LOL of *Congeria amygdaloides navicula* ANDRUSOV. Age: Tortonian–Lowermost Messinian; "beds with *C. amygdaloides navicula*", upper part of the Upper Maeotian. Reference sections: Merzhanovo, 1300 km.

***Monodacna pseudocatillus – Prosodacna schirvanica* Interval Zone** corresponds to the interval from the FOLs of *Monodacna pseudocatillus* BARBOT and *Prosodacna schirvanica* ANDRUSOV to the upper disconformal boundary of the Upper Miocene sedimentary complex. Age: Messinian; Lower Pontian. Reference sections: Bolshoj Log, Ptchelovodnaja, Aleksandrovka.

Unzoned intervals include hiatuses (at the base of the Donskaja and Aleksndrovskaja Formations) and short intervals where zonality could not be established because of the scarcity of fossils remains (the transition between Taganrogskaja and Rostovskaja Formations, upper part of Rostovskaja Formation, and Janovskaja Formation).

All the above mentioned zones were defined by characteristic taxa bioevents. The last ones are very easy to be determined in the stratigraphic record. All these events seem to be isochronous at least within the area of the Rostov Dome.

Conclusions

The comparison of absolute ages permits a correlation of the global and regional stages for the Eastern Paratethys to be made.

Studies of the Upper Miocene deposits of the Rostov Dome resulted in the definifion of five distinct Bivalvia-based biozones. The implicated chronostratigraphy coupled with the Bivalvia-based biozonation seems to be a real alternative for replacing the previously developed regional stratigraphy, based on the definition of regional stages.

Further research should be aimed at extending the defined biozones to the entire territory embraced by the Eastern Paratethys.

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Appendix

In addition to the formal definition of the Merzhanovskaja Formation (RUBAN, 2002b), a detailed indication of particular beds of skeletal limestones in the Merzhanovo stratotype section is presented below (see this section location and whole composition in Fig. 2, 4). The Beds are numbered from base to top.

LOWER MEMBER (beds 1–10) – 1.26 m
bed 1 – 0.05 m
bed 2 – 0.15 m
bed 3 – 0.04 m
bed 4 – 0.08 m
bed 5 – 0.02 m
bed 6 – 0.11 m
bed 7 – 0.08 m
bed 8 – 0.18 m
bed 9 – 0.40 m
bed 10 – 0.15 m

UPPER MEMBER (beds 11–16) – 0.32 m
bed 11 – 0.03 m
bed 12 – 0.03 m
bed 13 – 0.04 m
bed 14 – 0.06 m
bed 15 – 0.06 m
bed 16 – 0.10 m
Total thickness 1.58 m.

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Резиме

Горњи миоцен Ростовске доме (Источни Паратетис): значај за хроностратиграфију и биостратиграфија на основу шкољака

Ростовска дома налази се на југу руске платформе. За време касног миоцена ова област припадала је Танајском палеозаливу Источног Паратетиса. И ако се миоценски седименти ове области проучавају око 100 година њихова детаљна стратиграфска подела још није приказана. На основу проучавања 12 изданака горњомиоценских успостављена је литостратиграфска (укупно 6 формација) и секвентна стратиграфија.

Регионални катови се углавном употребљавају и имају предност у Источном Паратетису. Њихова корелација са хроностратиграфским катовима предложених од стране Интернационалне комисије за стратиграфију је неопходна, зато што ће то омогућити корелацију стратиграфских јединица, као што се то може дефинисати за касни миоцен ростовске доме, са суседним или чак удаљеним областима. Када су катови, који су глобално препознатљиви, предложени, изгледа да није потребно да се употребљавају регионалнои катови, који су прихватљиви само у оним случајевима када се расправља о хроностратиграфској подели.

Покушана је примена недавно установљене неогене хроностратиграфије на област Ростовске доме. Ово је остварено кроз упоређење апсолутних старости граница две поменуте врсте катова. Сарматском регионалном кату одговара горњи део лангиана, целом серавалиану и доњем делу тортона. Меотском регионалном кату одговара горњи део тортона и најнижи хоризонти месиниана. Понтском регионалном кату одговара већи део месиниана и најнижи занклеан. Установљено је да граница миоцен/плиоцен, која се налази у бази занклеана, има различит положај у Источном Тетису, где је успостављена у бази кимерианског регионалног ката.

У горњомиоценских седимената ростовске доме нађени су многобројни остаци шкољака. Анализа распрострањења таксона дозволила је по први пут успостављење биостратиграфије на основу шкољака за територију Ростовске доме. У оквиру интервала серавалиан-месиниан установљено је пет биозона: *Tapes vitalianus*, *Cerastoderma fittoni–Cerastoderma subfittoni*, *Congeria panticapaea*, *Congeria amygdaloides navicula* и *Monodacna pseudocatillus–Prosodacna schirvanica*. Интервали без зона укључују хијатусе (у бази Донске и Александровске формације) и краћих интервала где зоналност није могла бити успостављена због ретких фосила (прелаз између Таганрогске и Ростовске формације, горњи део ростовске формације и јановске формације).

Наредна проучавања би требала бити усмерена на дефинисању биозона целе територије која је припадала Источном Паратетису.

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Major Paleozoic-Mesozoic unconformities in the Greater Caucasus and their tectonic re-interpretation: a synthesis

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Abstract

Four major unconformities are present within the Paleozoic-Mesozoic strata of the Greater Caucasus, a region that was derived from the Afro-Arabian margin of Gondwana during the Ludlow, then docked at the Laurussian margin near the European Alpine terranes, and moved eastwards during the Late Triassic-earliest Jurassic to occupy its present position. The Ordovician unconformity, which includes part of the Middle Ordovician, formed due to uncertain tectonic activity on the Gondwanan margin. The mid-Permian unconformity, which encompasses the Guadalupian to early Lopingian, might have resulted from the Saalian Orogeny accompanied with strike-slip activity. The Triassic/Jurassic unconformity, which encompasses the mid-Rhaetian to early Sinemurian, developed after the long-distance movement of the Greater Caucasus Terrane along a shear zone or, alternatively, after the global sea-level fall. These three mentioned major unconformities are correlated with similar unconformities in some European, north-central African, and Arabian basins. The mid-Jurassic unconformity, which encompasses the mid-Bathonian to the Early Callovian, might have been caused by arc-arc collision in the Caucasian sector of the northern Neotethys or by the major sinistral transtension from Europe to the Himalayas.

Keywords: Unconformity, Palaeotectonics, Paleozoic, Mesozoic, Greater Caucasus.

Introduction

The Greater Caucasus is a large region, which stretches between the Black Sea and the Caspian Sea, and it embraces some areas of southwestern Russia, northern Georgia, and northern Azerbaijan (Fig. 1). The whole Caucasus is a large Alpine-type region, which is presentely located at the junction of the Eurasian, Anatolian, Arabian plates and Iranian terranes (Bird, 2003). It is composed of two principal domains (terranes), which are the Greater Caucasus and the Lesser Caucasus. They are divided by the Transcaucasian depressions, which are named the Rioni Depression and the Kura Depression (Fig. 1). The geological record from this region provides many new and intriguing information to recognize and explain the tectono-sedimentary events common for the entire Tethys. Unfortunately, the Greater Caucasus is less known than the adjacent European, Asian, and African regions. The use of traditional, but outdated and false concepts like "geosyncline paradigm" and "formational analysis" by the Russian geologists additionally minimizes the significance of information from this region for the international scientific community. The Paleozoic-Mesozoic sedimentary succession of the Greater Caucasus is characterized by the presence of several *major unconformities* (Ruban, 2006a). However, the stratigraphical data on these unconformities are "dispersed" in numerous references, published mostly in Russian.

This article presents a brief synthesis of knowledge on the major Paleozoic-Mesozoic unconformities of the Greater Caucasus, including new data collected by the author and new data from recent studies by Kotlyar et al. (1999, 2004), Davydov and Leven (2003), and Gaetani et al. (2005). In addition, this article presents a new tectonic interpretation of the major unconformities, according to recent models of the regional evolution of the Greater Caucasus (Tawadros et al., 2006; Ruban, in press), and it also demonstrates that they may be traced in some other regions.

Geological setting

The geological history of the Greater Caucasus has been briefly reviewed by Tawadros et al. (2006), who presented a new model of the tectonic

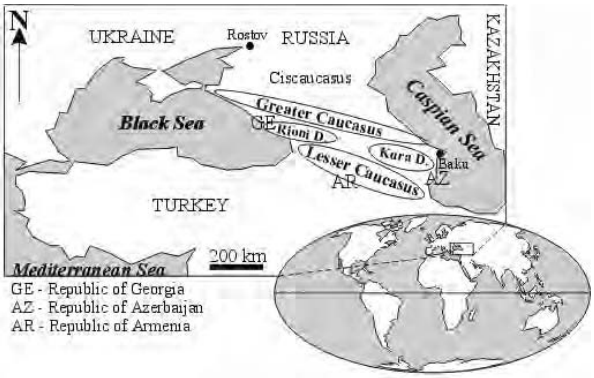

Fig. 1 - Geographic location of the Greater Caucasus.

evolution of this region. To develop this new model, various palaeontological, stratigraphical, and palaeomagnetic data and global plate reconstructions (Dercourt et al., 2000; Cocks and Torsvik, 2002; Stampfli and Borel, 2002; Vai, 1991, 2003; von Raumer et al., 2002, 2003; Vai, 2003; Golonka, 2004; Scotese, 2004; Torsvik and Cocks, 2004) were taken into account. A key point in this model is the activity of planetary-scale strike-slip movements on the northern periphery of the Palaeotethys Ocean (Arthaud and Matte, 1977; Swanson, 1982; Vai, 1991, 2003; Rapalini and Vizán, 1993; Stampfli and Borel, 2002; Ruban and Yoshioka, 2005). To develop this model, it was argued first, that the Greater Caucasus was located somewhere close to the Carnic Alps and Bohemia in the Middle-Late Paleozoic. This was concluded from the similarities of their sedimentary successions and palaeontological records. The next step was to explain how the studied terrane might have reached its present position. This involved the idea of long-distance displacements along the major shear zone. As for the pre-Ludlow time, the position of the Greater Caucasus was evaluated approximately with the global reconstructions of Stampfli and Borel (2002), which permit to understand the palaeogeodynamics of the Hunic terranes in the whole. According to this new model of Tawadros et al. (2006), the following five tectono-depositional phases are recognized within the Paleozoic-Mesozoic history of this region (Tawadros et al., 2006) (Fig. 2): (1) *Gondwanan Phase* (before the Ludlow, the Greater Caucasus was a part of the Afro-Arabian margin of Gondwana open to the Prototethys and then Rheic oceans), (2) *Hunic Phase* (from the Ludlow and until the end-Devonian, the Greater Caucasus and the other European Hunic terranes were shifted to the Laurussian margin and docked somewhere near the Carnic Alps and Bohemia), (3) *Proto-Alpine Phase* (the Greater Caucasus was located at above-mentioned place until the Late Triassic), (4) *Left-Shear Phase* (the clockwise rotation of Africa provoked the left-shear displacements along the Northern Palaeotethyan Shear Zone and, during the earliest Jurassic, the Greater Caucasus reached its present position), and (5) *Arc Phase* (until the end of the Mesozoic, the Greater Caucasus was dominated by the parallel arcs and elongated sea basins on the northern margin of the Neotethys Ocean).

The sedimentary history of the Greater Caucasus has been also reviewed by Tawadros et al. (2006), whereas Ruban (2006a) has presented a composite lithologic section (Fig. 3). Although numerous unconformities are known within the Paleozoic-Mesozoic strata of the Greater Caucasus, only four of them may be called major (Ruban, 2006a). As described below, the major unconformities are present in strata of *Ordovician*, *mid-Permian*, *end-Triassic-early Jurassic*, and *mid-Jurassic* age. The chronostratigraphic framework and numerical ages used in this paper are all taken from Gradstein et al. (2004). The term "unconformity" is used following Bates and Jackson (1987) and Catuneanu (2006).

Major unconformities

Ordovician unconformity

The first Paleozoic unconformity described in this article is present in strata of Ordovician age (Ruban, 2006a). The Ordovician unconformity, which is present between the Upper Formation of the pre-Middle Paleozoic metasedimentary complex and the siliciclastic strata of the ?Ordovician-Silurian Urleshskaja Formation (Fig. 3), has been observed in outcrops in the valley of the Khasaut River (Robinson, 1965). It seems that the Cambrian complexes were altered by metamorphism (Robinson, 1965), and this unconformity is essentially a nonconformity (*sensu* Bates and Jackson (1987) and Catuneanu (2006)). However, the structural framework of the Lower Paleozoic complexes of the Greater Caucasus remains a subject for further discussions.

Below the unconformity, the upper part of the

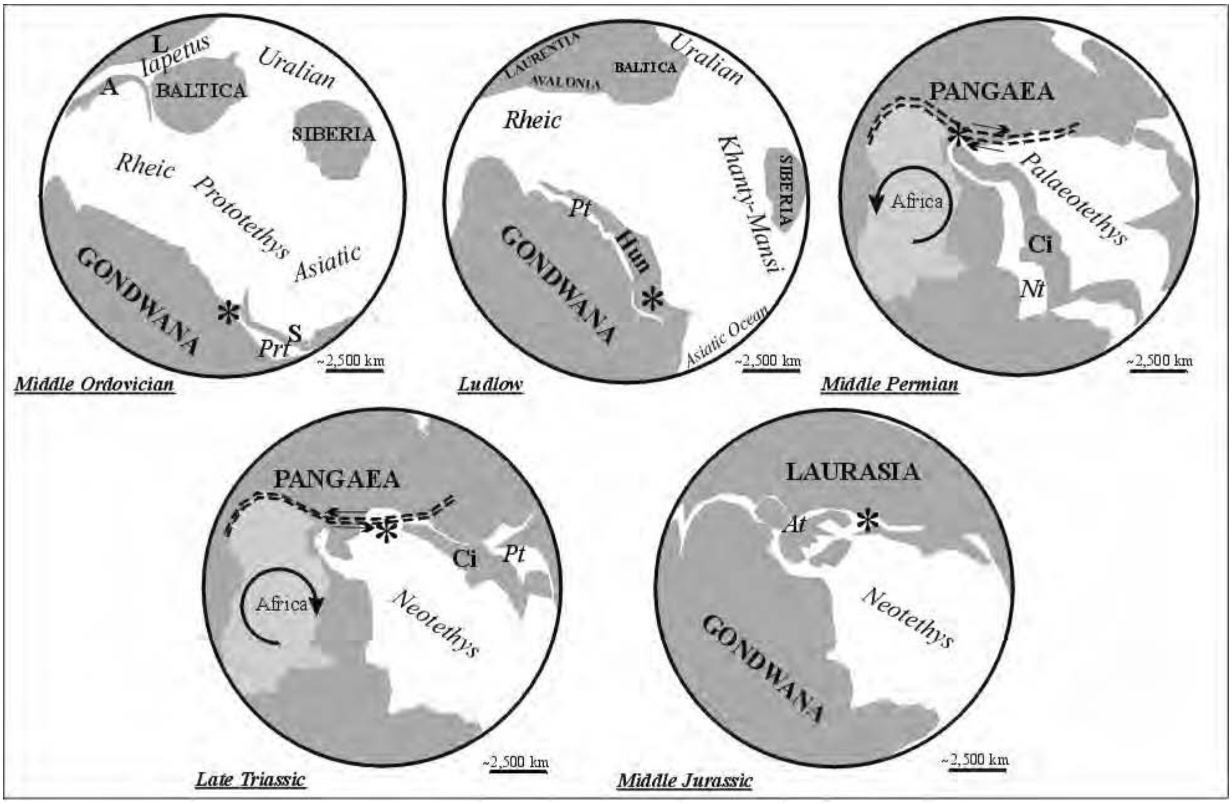

Fig. 2 - The palaeotectonic position of the Greater Caucasus Terrane (marked by asterisk). The palaeotectonic maps are simplified from Stampfli and Borel (2002). Abbreviations: A - Avalonia, S - Serindia, L - Laurentia, Prt - Prototethys, Pt - Palaeotethys, Hun - Hun Superterrane (European+Asiatic Hunic terranes), Ci - Cimmerian Superterrane, Nt - Neotethys, At - Alpine Tethys.

Upper Formation of the pre-Middle Paleozoic metasedimentary complex is composed of phyllites with rare interbeds of sandstones up to 900 m thick. This metasedimentary complex is thought to be of Ordovician age (Paffengol'ts, 1959, 1965), and it overlies the ?middle Cambrian carbonates with *Archaeocyathus* sp. Following these assumptions, the uppermost horizons of the Upper Formation may be upper Cambrian - lowermost Ordovician.

Above the unconformity, the lower Urleshskaja Formation consists of sandstones and conglomerates of unestimated thickness (Obut et al., 1988). The succession starts with the conglomerates, dominated by quartz pebbles, up to 6 m thick (Robinson, 1965). The few microfossils found in these strata (i.e., *Gleocapsamorpha* cf. *prisca* Zall., *Monotrematum* sp., *Trematosphaeridium* sp., *Protolophosphaeridium* sp., and *Turuchanica* sp.) suggest an Early-Middle Ordovician age (Potapenko, 1982; Obut et al., 1988). Within the Urleshskaja Formation, the boundary between the Ordovician and the Silurian deposits corresponds with the appearance of fine-grained siliciclastic rocks (Obut et al., 1988).

Thus, the age of the lower part of the Urleshskaja Formation is unclear, and we cannot conclude, how long was the hiatus, established since the Early Ordovician. If possibly sedimentation restarted already in the Middle Ordovician or even early as suggested from microfossils, this means hiatus was only 10-15 Ma long. In contrast, if this hiatus embraces the most part of the Ordovician, it could have lasted up to 40 Ma.

Mid-Permian unconformity

The second major Paleozoic unconformity in the Greater Caucasus is of Permian age, and it encompasses almost the entire Gaudalupian Series, the Wuchapingian and the lower Changhsingian stages (Fig. 3). Kotlyar et al. (1999, 2004) state that this is an angular unconformity between the Bol'shelabinskaja and Kutanskaja formations

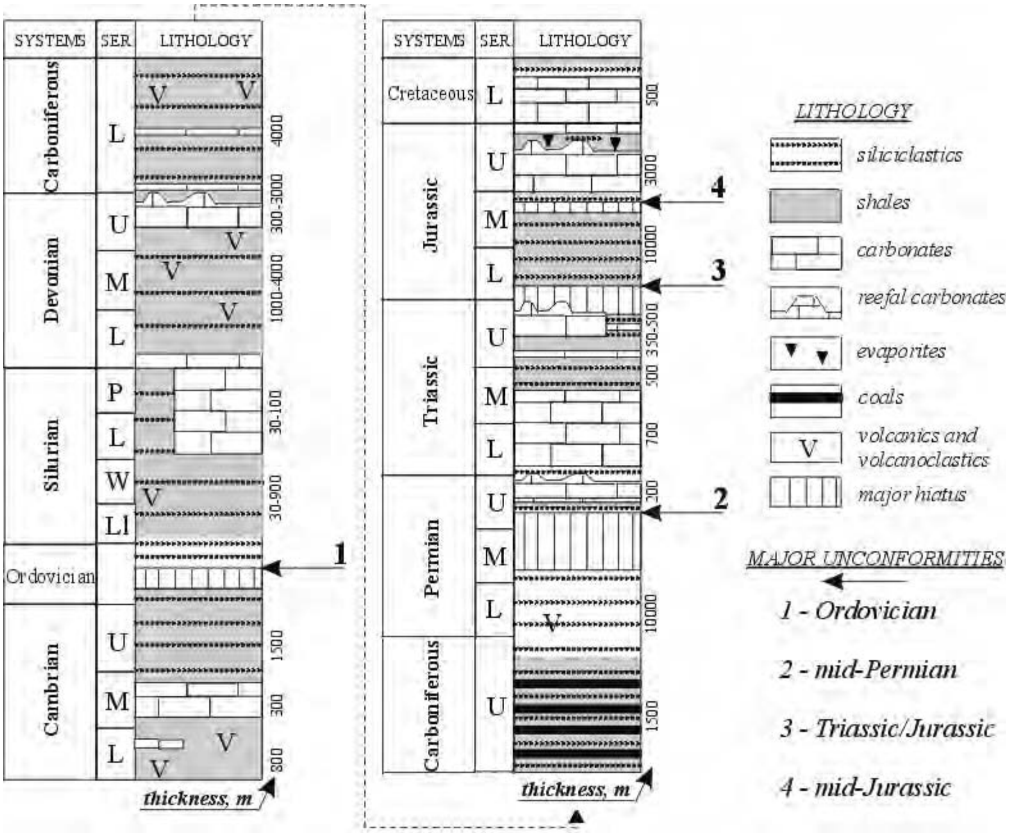

Fig. 3 - The composite Paleozoic-Mesozoic lithologic section of the northern Greater Caucasus (slightly modified from Ruban, 2006a).

in the northern part of the Greater Caucasus, whereas Kotlyar (1977) mentions that the Kutanskaja Formation overlies locally the Bol'shelabinskaja Formation without unconformity. In the southern part of the Greater Caucasus, there is no major unconformity in the Permian strata, which are of marine origin (Miklukho-Maklaj and Miklukho-Maklaj, 1966).

Below the unconformity in the northern Greater Caucasus, the molassic Bol'shelabinskaja Formation consists of siliciclastic red-beds (Miklukho-Maklaj and Miklukho-Maklaj, 1966; Tawadros et al., 2006). The estimations of its thickness vary significantly between 500 m (Miklukho-Maklaj and Miklukho-Maklaj, 1966) and 10,000 m (Kotlyar, 1977). The age of these deposits was established by Davydov and Leven (2003) as the Sakmarian, although the fossil remains from these strata are very poor, and include *Walchia* and *Acanthodes*. The author has also found some unidentified and possibly reworked crinoids. Miklukho-Maklaj and Miklukho-Maklaj (1966) hypothesized that the upper horizons of the Bol'shelabinskaja Formation may belong to the "Upper Permian", which is the Guadalupian Series in the recent chronostratigraphic framework (Gradstein et al., 2004).

Above the unconformity in the northern Greater Caucasus, the Upper Permian Kutanskaja Formation, and overlying Nikitinskaja and Urushtenskaja formations consist of siliciclastics, shales, and carbonates with a total thickness of about 200 m. The age of these overlying strata has been established with numerous palaeontological data (foraminifers, brachiopods, ammonoids, bivalves, gastropods, etc.) as Late Changhsingian (Kotlyar et al., 1999, 2004; Gaetani et al., 2005).

Thus, it seems that the lack of sedimentation lasted during the Guadalupian, Wuchiapingian and early Changhsingian, i.e., its duration was15-20 Ma.

Triassic/Jurassic unconformity

The first major Mesozoic unconformity in the

Greater Caucasus marks the Triassic/Jurassic boundary (Ruban, 2006a) (Fig. 3). This is an angular unconformity or nonconformity at the base of the Lower Jurassic strata (Rostovtsev et al., 1992). Latter are folded less intensively, and they overlie various complexes of the Paleozoic and Triassic.

Below the unconformity in the western Greater Caucasus, the Khodzinskaja Group is composed of limestones (up to 500 m thick), which contain very abundant and diverse marine fauna including ammonoids, foraminifers, brachiopods, bivalves, algae, corals, sponges, echinoids, and bryozoans (Dagis and Robinson, 1973; Jaroshenko, 1978; Rostovtsev et al., 1979; Prozorovskaya, 1979; Gaetani et al., 2005; Ruban, 2006b). The age of this group is established as Norian-Early Rhaetian (Prosorovskaya, 1979; Rostovtsev et al., 1979; Gaetani et al., 2005), although Dagis and Robinson (1973) and then Tozer (1988) suggested the undifferentiated "Norian-Rhaetian sandwich", and Krystyn (1990) showed some doubts about the presence of Rhaetian strata in the Greater Caucasus.

Above the unconformity in the Greater Caucasus, there are several siliciclastic formations that are up to 1,000 m thick (Rostovtsev et al., 1992). A late Sinemurian age for these strata is established with ammonites, foraminifers, and brachiopods (Antonova and Pintchuk, 1991; Rostovtsev et al., 1992; Prosorovskaya, 1993; Ruban, 2004a; Ruban and Tyszka, 2005). The presence of the Hettangian deposits in this region is not confirmed yet, although it is sometimes questionned (Rostovtsev et al., 1992; Ali-Zadeh, 2004).

Thus, the hiatus on the Triassic-Jurassic transition embraced a half of the Rhaetian, the Hettangian and the early Sinemurian. Consequently, It lasted not more than 5-10 Ma.

Mid-Jurassic unconformity

The second major Mesozoic unconformity in the Greater Caucasus is of Middle Jurassic age (Ruban, 2006a) (Fig. 3). This is an angular unconformity. The underlying strata are folded, whereas the overlying strata are only slightly deformed so as to create a monocline. However, in some southern areas of the Greater Caucasus, this mid-Jurassic unconformity is absent (Rostovtsev et al., 1992).

Below the mid-Jurassic unconformity, there are Sinemurian-Bathonian siliciclastic and argillaceous strata up to 10,000 m thick (Rostovtsev et al., 1992; Ruban, 2006a). Above this unconformity, there are siciliclastic and carbonate strata overlain with evaporites up to 3,000 m thick (Rostovtsev et al., 1992; Ruban, 2006a).

The duration of the hiatus, which the mid-Jurassic unconformity corresponds to, varied significantly within the Greater Caucasus (Rostovtsev et al., 1992). In some areas, the lack of sedimentation was established already in the end-Bajocian, while in the other areas, this occurred only in the end-Bathonian. The restart of sedimentation took place also at a diffirent time - it varied between the mid-Bathonian and mid-Callovian (Rostovtsev et al., 1992). Above the mid-Jurassic unconformity in the Laba-Malka Area, the Kamennomostskaja Formation consists of conglomerates, shales, and sandstones with a total thickness of about 7 m (Fig. 4). Very abundant fossil remains, including ammonites, bivalves, brachiopods, belemnites, and plants have been collected from this formation (Lominadze, 1982; Rostovtsev et al., 1992; Pugatchev and Ruban, 2005; Ruban, 2004b, 2005; Gaetani et al., 2005). Its age, which has been established with ammonites and bivalves, is usually assumed as Early-Middle Callovian (Rostovtsev et al., 1992; Ruban, 2004b, 2005) or Middle-Upper Callovian (Lominadze, 1982). However, Gaetani et al. (2005) estimated, that the age of these deposits is the latest Bathonian-earliest Callovian, established with dinoflagellate cysts. In spite of above-mentioned differences and uncertainties, the mid-Jurassic hiatus in the Greater Caucasus lasted not more than 3-4 Ma.

Tectonic re-interpretation

A tectonic interpretation of the *Ordovician unconformity* should be based on the assumption, that the Greater Caucasus was located somewhere on the Afro-Arabian margin of Gondwana during the early Paleozoic (Tawadros et al., 2006) (Fig. 2). This is particularly supported by the presence in the studied region of carbonates with archaeocyaths; these carbonates are also known in the other peri-Gondwanan terranes (Gandin et al., 1987; Cocozza and Gandin, 1990; Courjault-Radeé et al., 1992; Debrenne et al., 1993, 2002; Geyer and Landing, 1995). At this location, two significant events during the Ordovician might have provoked tectonic activity within the Greater Caucasus. The first event involved the con-

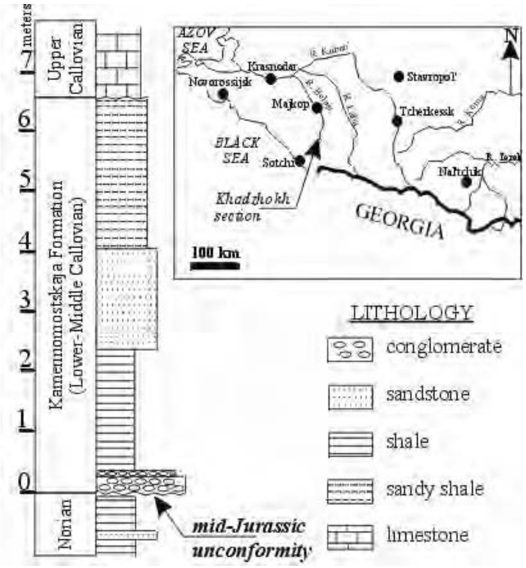

Fig. 4 - The Khadzhokh section (modified from Ruban, 2004b).

vergence along Avalonian-Cadomian arc, which was located off Gondwana. However, if Avalonia merged far northwards with Baltica and Laurentia, then Cadomia might have collided with Gondwana (Stampfli and Borel, 2002; Torsvik and Cocks, 2002; von Raumer et al., 2002, 2003). The second Ordovician event was the amalgamation of the Serindia terranes with Gondwana (Stampfli and Borel, 2002; von Raumer et al., 2002, 2003).

A tectonic interpretation of the *mid-Permian unconformity* should be based on the assumption that the European Hunic terranes collided with the northern margin of the Palaeotethys (Tawadros et al., 2006) (Fig. 2). According to Sampfli and Borel (2002), amalgamation of the Hunic terranes with Eurasia occurred during the Devonian, and resulted the Variscan (Hercynian) Orogeny, which became a remarkable Late Paleozoic tectonic event (Matte, 1986, 1991; Vai and Cocozza, 1986; Franke, 1989; Vai, 1991, 2003; Carmignani and Sassi, 1992; Krainer, 1993a; Bouchot et al., 1997; Dercourt et al., 2000; Franke et al., 2000; Schönlaub and Histon, 2000; Vai and Martini, 2001; Hoepffner et al., 2005, 2006). The Greater Caucasus as a Hunic terrane was a part of the Variscan (Hercynian) Orogen. Regional unconformities known within the Carboniferous sedimentary complexes (see e.g., Paffengol'ts, 1959; Davydov and Levin, 2003), although they are not so major, were, consequently, a result of the this orogeny. Latter was followed by uplift, and then extension and final collapse of the Variscan Cordillera during the Late Paleozoic (Faure, 1995; von Raumer, 1998; Stampfli and Borel, 2002). In the Alpine region of Europe, a mid-Permian unconformity is equated with the Saalian tectonic event (Stille, 1924; Krainer, 1993a). Dextral displacements along the planetary-scale shear zone at the northern margin of the Palaeotethys occurred during the Late Paleozoic (Arthaud and Matte, 1977; Swanson, 1982; Krainer, 1993a; Rapalini and Vizán, 1993; Trümpy, 1998; von Raumer, 1998; Stampfli and Borel, 2002; Vai, 2003; Ruban and Yoshioka, 2005). These displacements perhaps together with the post-orogenic deformations might have resulted the mid-Permian unconformity in the Greater Caucasus.

A tectonic interpretation of the *Triassic/Jurassic unconformity* is based on the assumption that the Greater Caucasus began to move eastwards during the Late Triassic, and reached its present position during the Early Jurassic (Tawadros et al., 2006) (Fig. 2). Possible deformations associated with such a movement and the subsequent collision of the studied terrane with the Russian Platform might have formed this unconformity. However, Ershov et al. (2003) suggested the regional orogeny, which lasted from the Carnian and until the Sinemurian. Gaetani et al. (2005) concluded that a strike-slip regime was present in the Greater Caucasus during the Triassic, and they also expressed some doubts about the appearance of collision during the latest Triassic. Alternatively, a global sea-level fall occurred at the Triassic-Jurassic transition (Embry, 1997; Hallam and Wignall, 1999; Hallam, 2001; Hesselbo et al., 2004, 2007; Haq and Al-Qahtani, 2005; Miller et al., 2005), and it may have resulted in the formation of the Triassic/Jurassic unconformity. The onset of the new subsiding basins after the collision of the Greater Caucasus with the Russian Platform led to an Early Jurassic transgression over those parts of the Greater Caucasus, where Triassic strata did not accumulate. In this case, the tectonic mechanism becomes unnecessary.

During the Jurassic, the Greater Caucasus was located on the nothern active margin of the Neotethys Ocean, where parallel island arcs and basins existed (Tawadros et al., 2006) (Fig. 2). Ershov et al. (2003) suggested that there was an "orogeny" and partial closure of the Greater Caucasian Trough during the Bajocian-Bathonian. Ruban (in press) introduced a different model, in which, during the Middle Jurassic, the

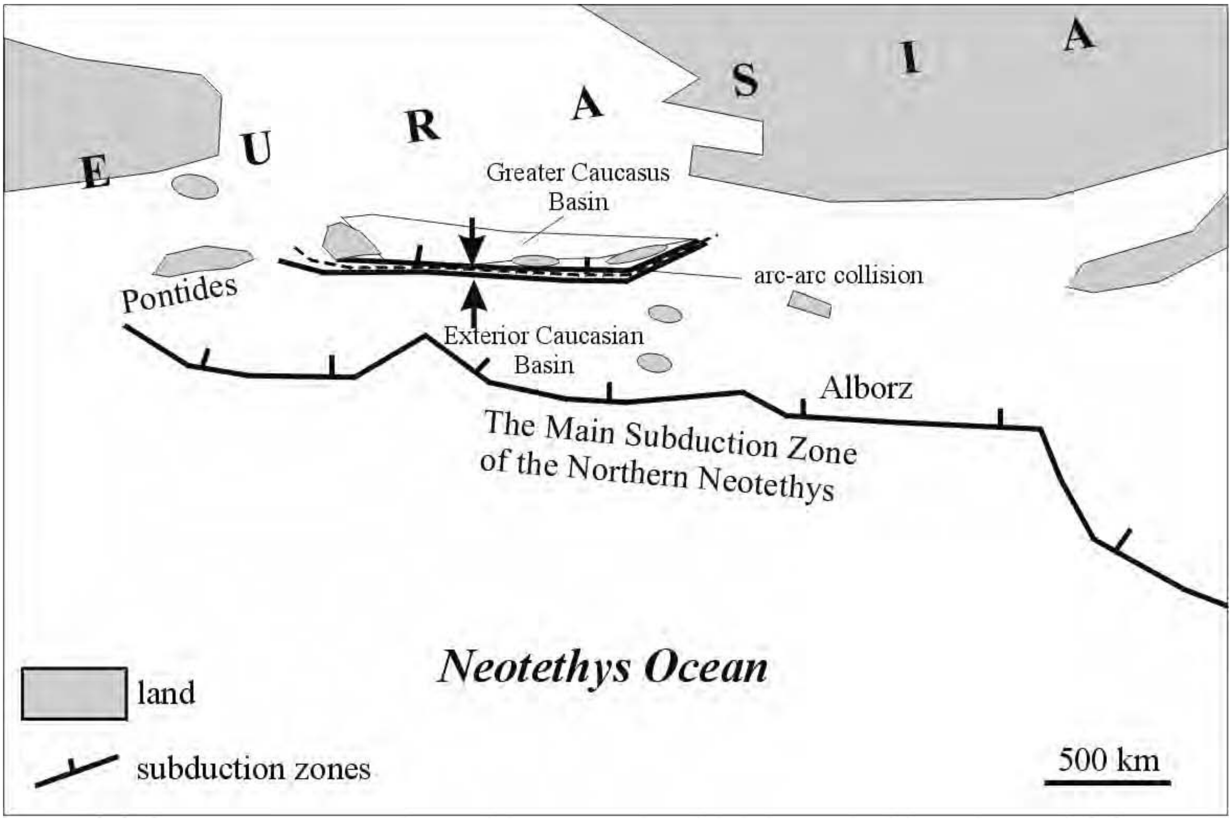

Fig. 5 - The Middle Jurassic tectonic setting in the Caucasian region (from Ruban, in press).

Northern and Southern Transcaucasian island arcs collided and a basin between them was closed (Fig. 5). Such arc-arc collision might have been able to induce the lack of sedimentation, which resulted in the formation of the *mid-Jurassic unconformity*. However, other explanations (e.g., oblique-wrench break-up) are also possible. Moreover, it is conceivable that this unconformity may be associated with a global eustatic fall, which occurred during the Bathonian (Haq et al., 1987; Haq and Al-Qahtani, 2005). However, Hallam (2001) argued that this fall was a regional event, which may be explained only with tectonics.

Discussion

The tracing of four major unconformities in the Paleozoic-Mesozoic strata in the Greater Caucasus (Fig. 3) would help to understand their extent and relation to the planetary-scale tectono-sedimentary events.

The *Ordovician unconformity* of the Greater Caucasus has some analogues in the Pennine Inliers and Midlands Platform of the British Isles and also in the Brabant Massif and Condroz Inlier, where hiatuses are known within the Ordovician strata, while in many other terranes of Avalonian Europe, there are no unconformities similar to that in the studied region (Verniers et al., 2001, 2002). In some basins of Germany, no large Ordovician hiatuses are found (Stratigraphische Tabelle von Deutschland, 2002). However, significant hiatuses are present within the Ordovician strata of Hessen, Lausitz, and Frankenwald (Servais et al., 1998). An unconformity is established at the base of the Ordovician succession within the Moesian terrane (Yanev, 2000). In Turkey, the Late Ordovician hiatus is known, whereas the Early Ordovician hiatus is established within the Istanbul Terrane only (Kozur and Göncüoglu, 1998; Goncuoglu and Kozlu, 2000). However, Göncüoglu and Kozlu (1997) suggested earlier a significant hiatus within the Taurides, which corresponded to the Middle Ordovician. In NW Spain, stratigraphic gaps are present in the Middle Ordovician strata (Gutiérrez-Marco et al., 1999, 2002). In the Montagne Noire, a significant unconformity, similar to that in the Greater Caucasus, is established within the Ordovician (Simien et al., 1999). In Sardinia, the Sardic angular unconformity separates Upper Cambrian-Lower Ordovician strata from Upper Ordovician-Devonian strata (Stille, 1939; Vai and Cocozza, 1986; Leone et al., 1991; Vai and Martini, 2001; Cherchi et al., 2002; Villas et al., 2002). In the Gerrei Tectonic Unit of Sardinia, an unconformity of the Sarrabese Phase is present at the top of the Lower Ordovician (Corradini et al., 2002). Minor, but angular mid-Ordovician unconformities are present in the sedimentary basins of the Northern and Central Africa (Guiraud et al., 2005). In this region, the Cambrian-Ordovician and Ordovician-Silurian transitions are equated with the Sardinian and Taconian orogenic events. One of minor tectonic phases, i.e., the “pre-Caradocian” phase, occurred during the mid-Ordovician (Guiraud et al., 2005). In Morocco, the unconformities are known at the Cambrian-Ordovician and Lower-Middle Ordovician transitions and within the Upper Ordovician, but hiatuses associated with them were not very long, and a broad span of the Ordovician deposits is established there (Geyer and Landing, 1995; Chacrone et al., 2004; Guiraud, 2005). Hoepffner et al. (2006) suggested the “Pre-variscan” tectonic events were very localised in Moroccan domains. On the Arabian Platform, hiatuses are recognized within uppermost Lower Ordovician strata and within Upper Ordovician strata (Sharland et al., 2001). Thus, the Ordovician unconformity of the Greater Caucasus has good analogues in the other Early Paleozoic Gondwanan tectonic blocks. Surprisingly, this unconformity has also close analogues in the eastern North America, where the Middle-Late Ordovician Taconic orogeny is known in the Appalachians (Drake et al., 1989; Hatcher, 1989; Swezey, 2002; Thomas et al., 2002; Hibbard, 2004), and the Early Ordovician Penobscot tectonic event occurred in the Gander Zone (Colman-Sadd et al., 1992; Hibbard and Samson, 1995; Hibbard, 2004; Schoonmaker and Kidd, 2006). Also, the Owl Creek unconformity is traced within the Lower Ordovician strata of the Appalachians (Swezey, 2002).

The *mid-Permian unconformity* of the Greater Caucasus is correlated with Permian unconformities in the Alpine region of Europe, particularly in the Carnic and Southern Alps, the Dolomites, and the peri-Adriatic areas (Krainer, 1990,1993a, b; Venturini, 1990, 2002; Cassinis et al., 1998, 1999; Schönlaub and Histon, 2000; Cassinis and Rocnhi, 2001; Vai and Martini, 2001; Ronchi and Santi, 2003; Vai, 2003). The mid-Permian Saalian unconformities are also present in some areas of the Spanish Pyrenees and Northeastern Iberia (Sopeña et al., 1977, 1988; Arche and López-Gómez, 1996; Vera, 2004). Unconformities are present within the Permian strata of the North Sea (Glennie, 1992, 1997, 2000). In the basins of Germany, an unconformity encompasses the Sakmarian-Roadian to the Capitanian-Wuchapingian (Schneider, 2001; Stratigraphische Tabelle von Deutschland, 2002). Mid-Permian unconformities are present in the Balkan terranes, although not elsewhere (Yanev, 2000). Thus, the mid-Permian unconformity of the Greater Caucasus is evidently traced within Europe, especially within the Rotliegend type area where the Saalian and Altmark unconformities have been identified (Kozur, 1980; Glennie, 1997). However, the sediment preservation above the unconformity began later in the Greater Caucasus than in many other mentioned European regions. In Northern and Central Africa, the lack of sedimentation occurred in many areas during the Permian, and somewhere it was established around the mid-Permian (Guiraud et al., 2005). On the Arabian Platform, the pre-Khuff unconformity (Sharland et al., 2001; Al-Husseini, 2004) may be correlated with the mid-Permian unconformity in the Greater Caucasus. However, in some areas of Oman, the Middle Permian-Early Triassic Khuff Formation overlies the Early-Middle Permian Gharif Formation conformably (Al-Husseini, 2004; Osterloff et al., 2004).

The *Triassic/Jurassic unconformity* of the Greater Caucasus may be correlated with the Early Cimmerian unconformity in European basins. However, this unconformity divides the Norian and Rhaetian strata (Jacquin and de Graciansky, 1998), whereas the Triassic/Jurassic unconformity in the Greater Caucasus was established later, during the mid-Rhaetian (Fig. 3). Embry (1997), Hallam and Wignall (1999), Hallam (2001), Hesselbo et al. (2004, 2007), Haq and Al-Qahtani (2005), and Miller et al. (2005) inferred a major sea-level fall and consequent hiatuses or unconformities at the Triassic-Jurassic transition from both the global and European stratigraphic record. Latest Triassic hiatuses have been identified in Germany (Stratigraphische Tabelle von Deutschland, 2002). In Northern and Central Africa, the Eo-Cimmerian unconformities are present in many basins at the Triassic-Jurassic transition (Guiraud et al., 2005), as well as in Arabia (Sharland et al., 2001).

The *mid-Jurassic unconformity* of the Greater Caucasus is difficult to be traced within Western Europe, because Bathonian and Lower Callovian hiatuses in Europe are quite minor and occur only locally (Jacquin and de Graciansky, 1998; Jacquin et al., 1998). The global evidence for a mid-Jurassic unconformity seems to be unclear (Hallam, 2001). In German basins, the Bathonian-Callovian interval is marked by local unconformities only (Stratigraphische Tabelle von Deutschland, 2002). No major Bathonian-Callovian unconformities, but only minor regressive episodes or local unconformities, are known both in Northern-Central Africa (Guiraud et al., 2005) and in Arabia (Sharland et al., 2001). In the Atlasic Domain, however, significant deformations occurred at the end of the Dogger (Guiraud et al., 2005). In general, sinistral transtension from Alpine Europe toward Himalayan Asia (Vai, 1991, 2003) might have resulted in these minor hiatuses and regressive episodes, and, therefore, a global-scale mechanism to explain the mid-Jurassic unconformity in the Greater Caucasus may be hypothesized.

Conclusions

This article presents a synthesis of knowledge on the major Paleozoic-Mesozoic unconformities in the Greater Caucasus with the following conclusions:

1) four major unconformities are present in the studied region, i.e., the Ordovician, mid-Permian, Triassic/Jurassic, and mid-Jurassic unconformities;

2) a tectonic re-interpretation of these unconformities relates them to the evolution of the Afro-Arabian margin of Gondwana during the Ordovician, post-collisional Variscan events during the Permian, left-shear displacements during the Late Triassic-Jurassic, and the arc-arc collision during the Middle Jurassic;

3) the Ordovician, mid-Permian, and Triassic/Jurassic unconformities are correlated undoubtedly with similar unconformities in the basins of Europe, Africa, and Arabia, whereas the evidences of such correlation for the mid-Jurassic unconformity are less clear.

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Taxonomic diversity structure of brachiopod associations at times of the early Mesozoic crises: evidence from the Northern Caucasus, Russia (northern Neotethys Ocean)

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Abstract. The Northern Caucasus is a large region of southwestern Russia that was located on the northern periphery of the Neotethys Ocean during the Triassic-Jurassic interval. Several crises impacted the evolution of the local fauna of brachiopods during this time as evidenced by changes in the taxonomic diversity of brachiopod associations in the region. The taxonomic diversity structure determines the relative importance of superfamilies to govern the species diversity, and the changes in it are measured with coefficient of rank correlation calculated for any two associations. The available stratigraphic ranges of species, which belong to 113 genera and 22 superfamilies, indicate that a recovery after the Permian/Triassic mass extinction is expressed by a rapid change in the taxonomic diversity structure in the Early Triassic-Anisian. The regional Ladinian crisis, which occurred after an abrupt deepening of the marine basin, did not result in major changes in the structure. The Triassic/Jurassic and Pliensbachian/Toarcian mass extinctions did not produce remarkable turnovers among brachiopods, but those superfamilies which dominated the Early Triassic species diversity of brachiopods rose again after these extinctions. The Pliensbachian/Toarcian event diminished significantly the importance of the Late Triassic superfamilies. The potential Aalenian mass extinction did not affect the taxonomic diversity structure. The changes in this structure recorded in the Northern Caucasus did not correspond to those seen in the Alpine Region.

Key words: brachiopods, diversity, Jurassic, mass extinction, Northern Caucasus, Triassic

Introduction

Mass extinctions are among the most spectacular events in the Earth's history. Interest in them has grown thanks to the studies of Sepkoski (1982, 1993, 2002) and his coworkers (Sepkoski *et al.*, 1981; Raup and Sepkoski, 1982). However, we are still far from understanding the biological and geological mechanisms of these events as suggested from the present syntheses (Taylor, 2004; Over *et al.*, 2005; Erwin, 2006; Roopnarine, 2006; Twitchett, 2006). The assessment of rates of background extinction is also a matter of criticism (Boucot, 2006). One intriguing question concerns the evolutionary consequences of mass extinctions. Jablonski (2004) notes that, in spite of the severity of mass extinctions, these events were not able 'to re-set the evolutionary clock' completely. However, studies of foraminifers suggest that the Permian/Triassic extinction resulted in the reorganization of their assemblages, because their Triassic and Jurassic assemblages were similar to those of the Cambrian and Ordovician (Ruban, 2001). Although we still have only limited knowledge of the major or first-order mass extinctions (the so-called 'Big Five'), information on the minor or second-order mass extinctions is even more limited. It is unclear whether those biotic crises were really less intense than any of the 'Big Five' (Ruban and Tyszka, 2005).

The rich paleontological record of the Northern Caucasus—a large northern Neotethyan region (Figure 1)—allows examination of the early Mesozoic crises among brachiopods, including the influences of the Triassic/Jurassic (T/J) and Pliensbachian/Toarcian (Pl/To) mass extinctions. The former is usually mentioned among the 'Big Five', whereas the latter is considered a minor extinction (Little and Benton, 1995; Hallam and Wignall, 1997; Hallam, 2002; Pálfy *et al.*, 2000, 2002; Taylor, 2004; Hesselbo *et al.*, 2007).

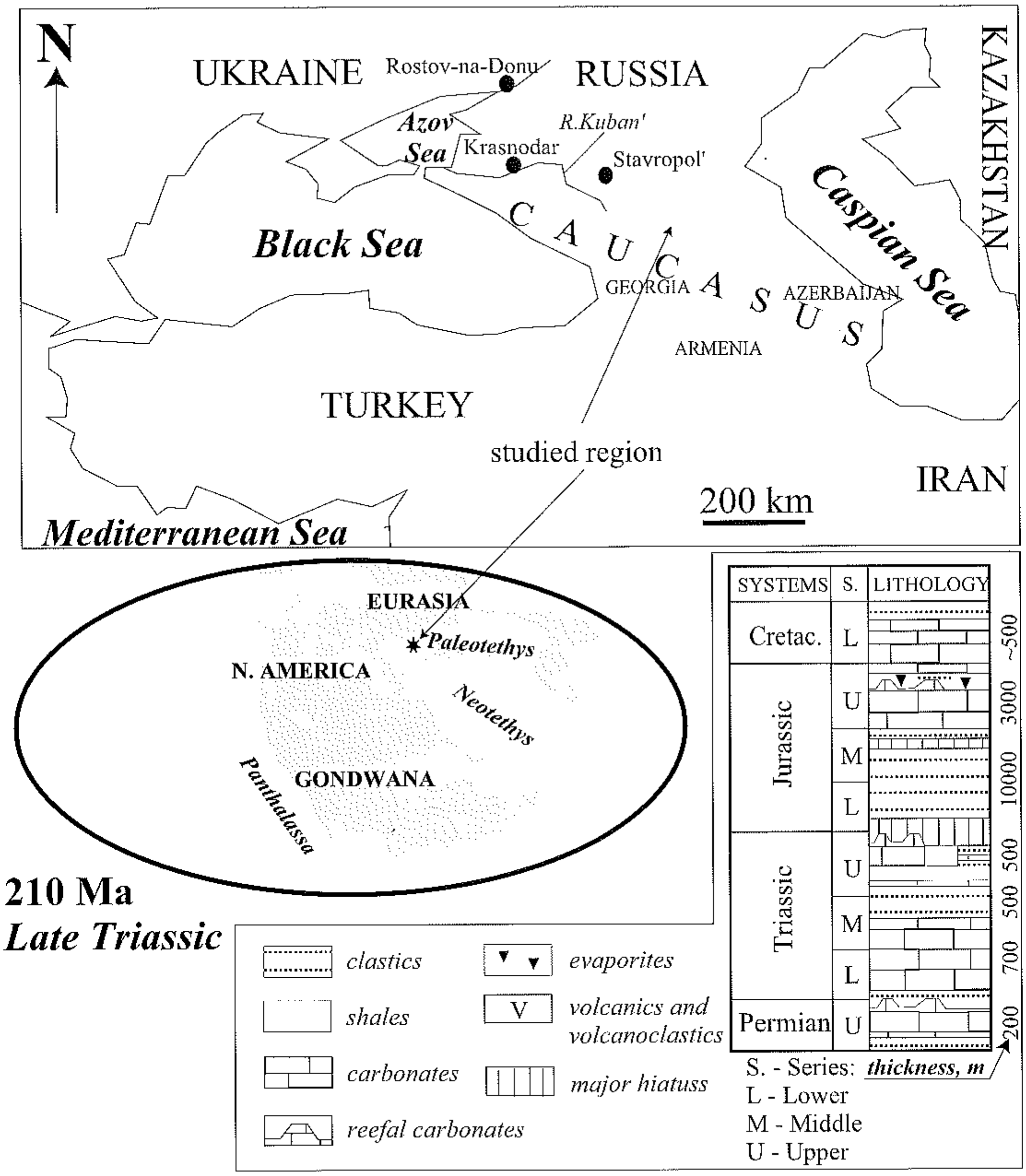

Figure 1. Geographical and paleogeographical (after Scotese, 2004) location of the Northern Caucasus. Generalized composite lithologic section after Ruban (2006c).

Geological setting

The Northern Caucasus is a large region, located in the southwest of Russia (Figure 1). Geographically it corresponds to the northern slope of the Main Caucasus Range, whereas geologically it is a northern part of the Greater Caucasus Terrane. This terrane was identified by Gamkrelidze (1997) and Tawadros *et al.* (2006). It was one of the so-called European Hunic terranes (Stampfli and Borel, 2002; Tawadros *et al.*, 2006), which were derived from the Gondwanan margin in the mid-Paleozoic. Subsequently, the Greater Caucasus Terrane docked somewhere near the Carnic Alps and in Late Triassic-Early Jurassic times, when sinistral strike-slip movements occurred on the Eurasian margin, it was transported eastward to its present position at the south margin of Baltica (Tawadros *et al.*, 2006). Subsequently elongated sea basins divided by arcs evolved in the Jurassic (Ershov *et al.*, 2003; Ruban, 2006a; Tawadros *et al.*, 2006). Thus, during the early Mesozoic, the Northern Caucasus was located in the northern part of the Neotethys Ocean.

The stratigraphy and lithology of the Triassic-Jurassic deposits within the studied region were summarized by Ruban (2004, 2006b,c) and Gaetani *et al.* (2005) (Figure 1). Marine Triassic deposits are known only in the western Northern Caucasus, whereas the Jurassic sedimentary complexes are known in most of the studied region. In general, the 700 m thick Induan-Anisian carbonates are overlain by Ladinian-Carnian turbidites with a thickness of ~500 m. The Norian-lower Rhaetian interval is represented by carbonates including reefal limestones with a total thickness of about 500 m. Locally this sedimentary complex is replaced laterally by a 350 m thickness of intercalated shales and siliciclastics, which are middle Norian in age. The Triassic-Jurassic transition is marked by a major hiatus. The overlying Sinemurian-Bathonian sedimentary complex is up to 10,000 m thick and is dominated by shales and siliciclastics with interbedding of carbonates and coal.

Both Triassic and Jurassic strata contain very rich fossil assemblages, which include brachiopods, bivalves, ammonoids, foraminifers, corals, sponges, and crinoids. As suggested by the newest paleogeodynamic reconstructions (Ruban, 2006a), the study region was a part of the same marine basin, and there is no evidence on the paleobiogeographical differentiation within the Northern Caucasus. The T/J mass extinction was not documented in the Northern Caucasus because of the presence of the noted major hiatus. However, this event may be recorded by the changes in the Early Jurassic assemblages in comparison with those of the Late Triassic as recorded here. The Pl/To mass extinction was documented by Ruban (2004) and Ruban and Tyszka (2005) with data on brachiopods and foraminifers, respectively. However, the influence of this event on bivalves is not clear (Ruban, 2006d). Additionally, the studies of Ruban (2004) and Ruban and Tyszka (2005) indicated an Aalenian mass extinction, which affected brachiopods and foraminifers both in the Northern Caucasus and some other regions of Europe. This may suggest a previously unsuspected Mesozoic mass extinction, although the existing biodiversity curves (see Taylor, 2004) are not detailed enough to document it globally.

Materials and methods

Data on the stratigraphic ranges (per-stage distribution) of the Triassic-Jurassic brachiopods was compiled from a number of available sources and published (Ruban, 2006b,c). Low resolution of the data in some original sources does not permit discussion of turnovers within given stages. In total, 113 genera which belong to 22 superfamilies have been accounted (Table 1). Brachiopod taxa were found in the deposits of most stages in the studied Triassic-Bajocian interval, except the Ladinian and the Hettangian. In the latter case the absence of brachiopods is explained by the presence of a major regional hiatus, whereas the enigmatic Ladinian crisis occurred due to the rapid deepening of the marine basin in the latest Anisian-Ladinian (Ruban, 2006b). Suprageneric taxonomy was verified using the revised edition of the "Treatise on Invertebrate Paleontology" (Williams *et al.*, 1997–2006). Additionally, the taxonomic system employed in the brachiopod database (version 1996) by Doescher (1981, 1990) proved useful.

This paper addresses the taxonomic diversity structure of the brachiopod associations. The latter are considered as entireties of taxa, which existed in each stage of the Triassic and Jurassic. Particular taxa of higher rank may play a different role in the determination of diversity of lower-rank taxa. This forms a taxonomic diversity structure, which evidently changed through geological time (Figure 2). A specific index was proposed to measure these changes (Ruban, 2001; Ruban and Tyszka, 2005). Rst is a coefficient of Spearman's rank correlation (Kendall, 1975) between two associations by presence/absence of higher-rank taxa, where the presence of each of these taxa is indicated by the number of lower-rank taxa, which belong to this higher-rank taxon in the particular association. In our case, the higher-rank taxa are superfamilies, whereas the lower-rank taxa are species (Table 2). The Rate of Transformation of the Taxonomic Diversity Structure (TTDSR) is estimated as 1/Rst. It illustrates the rate of change in the superfamily control of the species diversity, i.e., the changes in significance of each superfamily for the determination of the species diversity. It is always interesting to calculate Rst not only for the successive associations, but for those existing in different times, for example, the Anisian and Pliensbachian associations. This procedure may yield important results (Ruban, 2001).

Changes in taxonomic diversity structure

The taxonomic diversity structure of the brachiopod associations changed significantly during the Triassic-Bajocian in the Northern Caucasus (Figure 3, Table 3). The lower values of Rst in the Early Triassic-Anisian indicates high TTDSR. This was a result of brachiopod recovery and Anisian radiation after the devastating Permian/Triassic mass extinction. The

Table 1. Stratigraphic distribution of the Triassic-Bajocian brachiopod genera in the Northern Caucasus. N.I.—not identified. Stage abbreviations: T1—Induan + Olenekian, AN—Anisian, LA—Ladinian, CA—Carnian, No—Norian, RH—Rhaetian, HE—Hettangian, SI—Sinemurian, PL—Pliensbachian, TO—Toarcian, AA—Aalenian, BJ—Bajocian.

Superfamilies	Genera	T1	AN	LA	CA	NO	RH	HE	SI	PL	TO	AA	BJ
Rhynchonelloidea	*"Rhynchonella"*		1						2	1	2	2	2
Spiriferinoidea	*"Spiriferina"*								1	1			
Rhynchonelloidea	*?Rhynchonelloidea*											1	
Rhynchonelloidea	*Abrekia*	1											
Rhynchonelloidea	*Acanthothyris*											1	3
Dialasmatoidea	*Adygella*				1	1	1						
Dialasmatoidea	*Adygelloides*						1						
Koninckinoidea	*Amphiclina*					1	2						
Koninckinoidea	*Ampliclinodonta*					1							
Dialasmatoidea	*Angustothyris*		1										
Zeillerioidea	*Aulacothyris*									2	1		
Kingenoidea	*Aulacothyropsis*				1		4						
Norelloidea	*Austriella*					1							
Rhynchotetradoidea	*Austrirhynchia*						2						
Pennospiriferinoidea	*Balatonospira*				1								
N.I.	*Bobukella*				1	1							
Wellerelloidea	*Bodrakella*									1			
Wellerelloidea	*Calcirhynchia*								1	1			
Pugnacoidea	*Calvirhynchia*												1
Rhynchonelloidea	*Capillirhynchia*											1	
Wellerelloidea	*Caucasorhynchia*					1	1			1			
Dialasmatoidea	*Caucasothyris*					1							
Loboidothyridoidea	*Cererithyris*												1
Zeillerioidea	*Cincta*									2			
Wellerelloidea	*Cirpa*									1			
Dialasmatoidea	*Coenothyris*		2										
Hemithiridoidea	*Costirhynchia*		1		1								
Pennospiriferinoidea	*Costispiriferina*		1										
Pugnacoidea	*Crurirhynchia*					1							
Ambocoelioidea	*Crurithyris*	1											
Pugnacoidea	*Cryptorhynchia*												1
Dialasmatoidea	*Cubanothyris*					2	2						
Loboidothyridoidea	*Cuersithyris*									1			
Rhynchonelloidea	*Cuneirhynchia*									4			
Rhynchonelloidea	*Curtirhynchia*										1		
Rhynchotetradoidea	*Decurtella*		2										
Zeillerioidea	*Digonella*									1			
Pennospiriferinoidea	*Dinarispira*		2										
Athyridoidea	*Dioristella*		1		1								
Cancellothyridoidea	*Disculina*									1			
Wellerelloidea	*Euxinella*					3	6						
Hemithiridoidea	*Fissirhynchia*					1	1						
Hemithiridoidea	*Flabellirhynchia*									1	1		
Rhynchonelloidea	*Furcirhynchia*								1			1	1
Hemithiridoidea	*Gibbirhynchia*								1	2	1		
Hemithiridoidea	*Grandirhynchia*										1	1	1
Spiriferinoidea	*Guseriplia*						2						
Loboidothyridoidea	*Heimia*												1
Norelloidea	*Holcorhynchella*		1										
Rhynchonelloidea	*Homoeorhynchia*								1	2	1		
Pennospiriferinoidea	*Koeveskallina*		1										
Koninckinoidea	*Koninckina*				1	1							
Suessioidea	*Laballa*					2	3						
Pennospiriferinoidea	*Lepismatina*						1						
Dyscolioidea	*Linguithyris*								1				
Spiriferinoidea	*Liospiriferina*								2	4	1		
Loboidothyridoidea	*Loboidothyris*												1
Loboidothyridoidea	*Lobothyris*					2				3	1	2	2
Athyridoidea	*Majkopella*						3						

Table 1. Continued

Superfamilies	Genera	T1	AN	LA	CA	NO	RH	HE	SI	PL	TO	AA	BJ
Spiriferinoidea	*Mentzelia*		1		2	1	2						
Wellerelloidea	*Moisseievia*					2	1						
Loboidothyridoidea	*Monsardithyris*												1
Loboidothyridoidea	*Morrisithyris*												1
Retzioidea	*Neoretzia*						3						
Wellerelloidea	*Neowelerella*	1											
Norelloidea	*Norella*		1										
Athyridoidea	*Oxycolpella*					3	2						
Hemithyridoidea	*Parvirhynchia*											1	
Athyridoidea	*Pexidella*		1			1							
Norelloidea	*Piarorhynchella*		1										
Rhynchonelloidea	*Piarorhynchia*								1	4	1		
Norelloidea	*Praemonticlarella*										1		
Rhynchotetradoidea	*Prionorhynchia*								1	1	1		
Pennospiriferinoidea	*Pseudocyrtina*					1							
Pugnacoidea	*Pseudogibbirhynchia*										2	2	2
Kingenoidea	*Pseudorugitella*					2	2						
Rhynchonelloidea	*Ptyctorhynchia*										1		
Loboidothyridoidea	*Ptyctothyris*												1
Pennospiriferinoidea	*Punctospirella*		1										
Hemithiridoidea	*Quadratirhynchia*												2
Hemithiridoidea	*Rhactorhynchia*												3
Dialasmatoidea	*Rhaetina*				1	6	4						
Rhynchonelloidea	*Rhimirhynchopsis*					1	1						
Rhynchonelloidea	*Rimirhynchia*									1			
Wellerelloidea	*Robinsonella*						1						
Hemithiridoidea	*Rudirhynchia*								2				
Zeillerioidea	*Rugitela*												1
Norelloidea	*Scalpellirhynchia*								1				
Zeillerioidea	*Securina*								1	1			
Spiriferinoidea	*Sinucosta*		1			1	1						
Loboidothyridoidea	*Sphaeroidothyris*												2
Suessioidea	*Spinolepismatina*					1							
Spiriferinoidea	*Spiriferina*								4	3			
Hemithiridoidea	*Squamirhynchia*									1			
Pugnacoidea	*Stolmorhynchia*											4	2
Rhynchonelloidea	*Striirhynchia*												1
Loboidothyridoidea	*Stroudithyris*											1	
Dialasmatoidea	*Sulcatinella*		2										
Dialasmatoidea	*Sulcatothyris*				1								
Athyridoidea	*Tetractinella*		1										
Hemithyridoidea	*Tetrarhynchia*								1				1
Thecospiroidea	*Thecospira*					1							
Thecospiroidea	*Thecospiropsis*					1							
Loboidothyridoidea	*Triadithyris*					2	1						
Rhynchonelloidea	*Trichorhynchia*											2	1
Rhynchotetradoidea	*Trigonirhynchella*					1	2						
Loboidothyridoidea	*Tubithyris*												2
Wellerelloidea	*Volirhynchia*		2										
Loboidothiridoidea	*Wattonithyris*												1
Dialasmatoidea	*Wittenburgella*					1	1						
Zeillerioidea	*Worobievella*					1	1						
Zeillerioidea	*Zeilleria*					3	6		14	15		3	4
Spondylospiroidea	*Zugmayerella*						1						

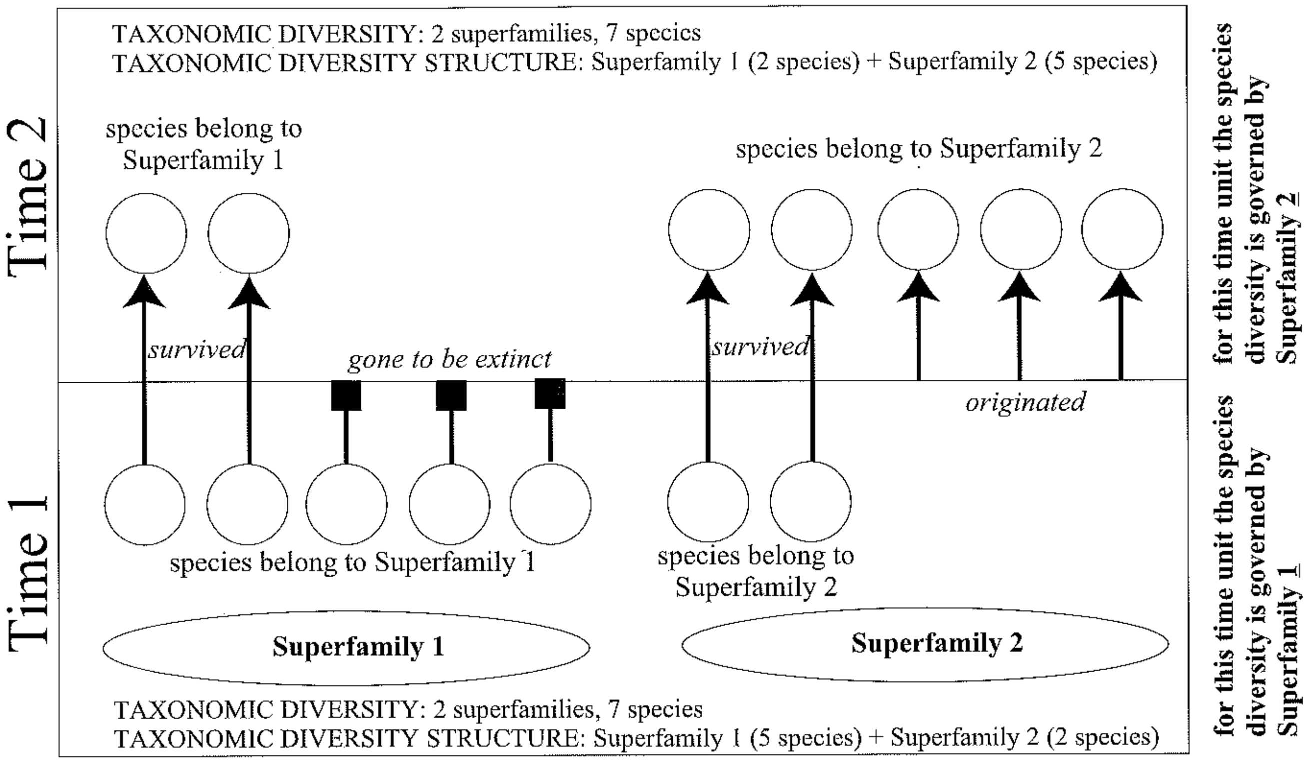

Figure 2. A concept of taxonomic diversity structure and its changes.

Table 2. Stratigraphic distribution of the Triassic-Bajocian brachiopod superfamilies in the Northern Caucasus. Number of species is indicated. Stage abbreviations—see Table 1. N.I.—not identified (genus *Bobukella* is not listed in "Treatise ..."). This matrix was used to calculate Rst (see Table 3).

Superfamilies	T1	AN	LA	CA	NO	RH	HE	SI	PL	TO	AA	BJ
N.I.				1	1							
AMBOCOELIOIDEA	1											
ATHYRIDOIDEA		3		1	4	5						
CANCELLOTHYRIDOIDEA									1			
DIALASMATOIDEA		5		3	11	9						
DYSCOLIOIDEA								1				
HEMITHIRIDOIDEA		1		1	1	1		4	4	3	2	7
KINGENOIDEA				1	2	6						
KONINCKINOIDEA				1	3	2						
LOBOIDOTHYRIDOIDEA					4	1			4	1	3	13
NORELLOIDEA		3			1			1		1		
PENNOSPIRIFERINOIDEA		5		1	1	1						
PUGNACOIDEA					1					2	6	6
RETZIOIDEA						3						
RHYNCHONELLOIDEA	1	1			1	1		5	12	6	8	8
RHYNCHOTETRADOIDEA		2			1	4		1	1	1		
SPIRIFERINOIDEA		2		2	2	5		7	8	1		
SPONDYLOSPIROIDEA						1						
SUESSIOIDEA					3	3						
THECOSPIROIDEA					2							
WELLERELLOIDEA	1	2			6	9		1	4			
ZEILLERIOIDEA					4	7		15	21	1	3	5

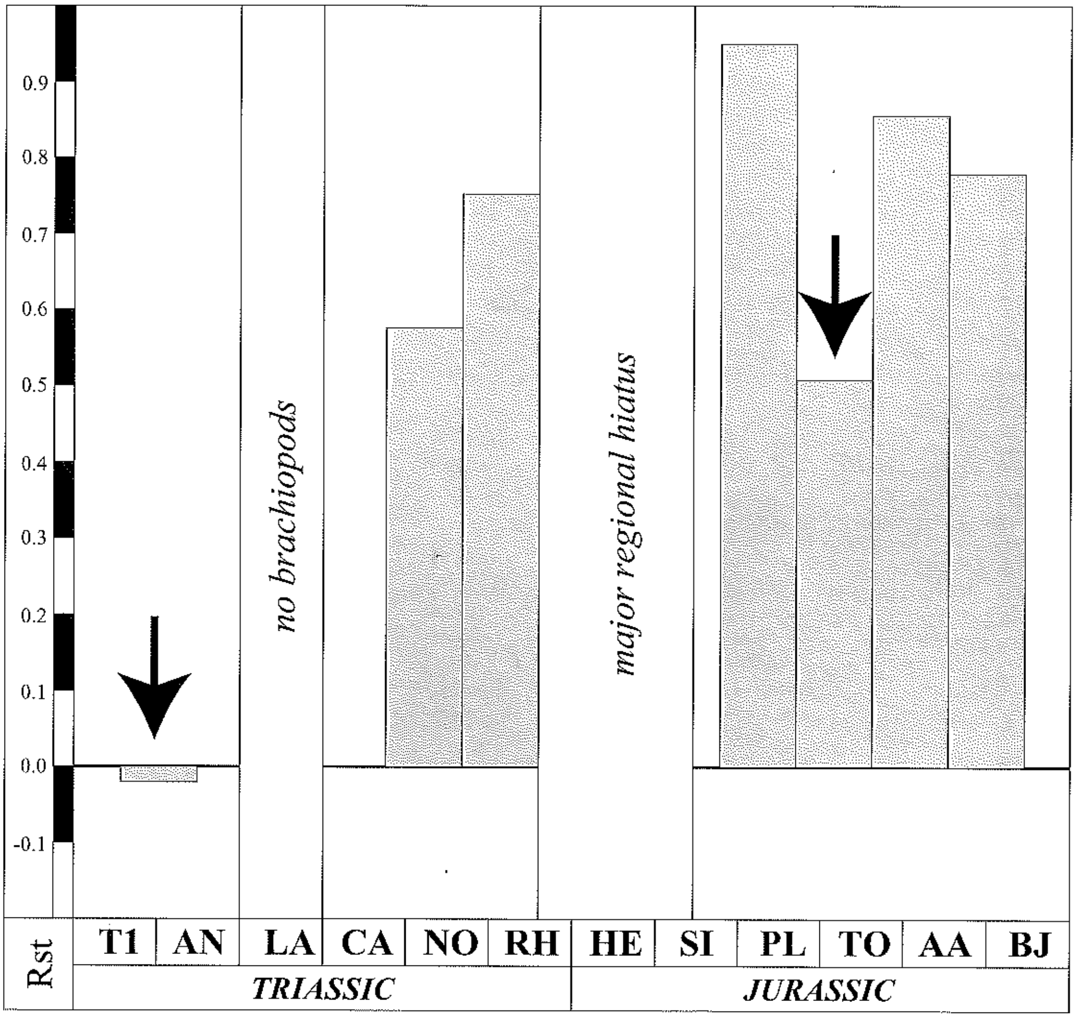

Figure 3. Changes in Rst values through the Early Triassic-Bajocian calculated for the brachiopod associations of the Northern Caucasus. Arrows mark TTDSR accelerations. Stage abbreviations—see Table 1.

Table 3. Rst values for the Triassic-Bajaocian brachiopod associations of the Northern Caucasus. Values higher than +0.2 are highlighted as bold. Significant values are given in italics. Stage abbreviations—see Table 1.

	T1	AN	LA	CA	NO	RH	HE	SI	PL	TO	AA	BJ
T1	1	−0.02	0.00	−0.25	0.02	0.09	0.00	0.05	**0.22**	**0.36**	**0.31**	0.10
AN		1	0.00	***0.58***	***0.49***	**0.39**	0.00	−0.07	−0.10	−0.03	−0.19	−0.22
LA			1	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
CA				1	***0.58***	***0.46***	0.00	0.01	−0.09	−0.13	−0.24	−0.21
NO					1	***0.76***	0.00	0.08	0.13	−0.16	−0.07	0.04
RH						1	0.00	**0.31**	**0.31**	−0.18	−0.18	−0.18
HE							1	0.00	0.00	0.00	0.00	0.00
SI								1	***0.95***	**0.38**	**0.36**	**0.30**
PL									1	***0.51***	***0.52***	***0.47***
TO										1	***0.86***	***0.66***
AA											1	***0.78***
BJ												1

rate decreased significantly in the Late Triassic. During the Jurassic the rate was mostly low, although it accelerated occasionally in the Pliensbachian-Toarcian interval. The taxonomic diversity structure described at superfamily-species organization was stable enough both in the Late Triassic and late Early-early Middle Jurassic.

The species diversity of the Early Triassic association was ruled equally by only three superfamilies (Table 2). In the Anisian, two superfamilies, namely Dialasmatoidea and Pennospiriferinoidea, played the most important role in the taxonomic structure, although they did not dominate in comparison with other taxa. Dialasmatoidea was also the most diverse superfamily both in the Carnian and Norian. In the Norian this superfamily evidently dominated in the composition of the brachiopod species diversity. However, Wellerelloidea also began to diversify in the Norian and by the Rhaetian its representatives were as important as those of Dialasmatoidea. Zeillerioidea also radiated in the Norian-Rhaetian. However, it is difficult to find leaders, which rule the entire species diversity in the end-Triassic. The Sinemurian diversity was dominated by the Zeillerioidea and less by Spiriferinoidea and Rhynchonelloidea. These three superfamilies governed the Pliensbachian species diversity. The Toarcian and Aalenian taxonomic diversity structure was dominated by the Rhynchonelloidea, although Pugnacoidea began to diversify from the Aalenian. In the Bajocian, Loboidothyrioidea became dominant, although some other superfamilies also diversified in this age.

A comparison of nonsuccessive associations reveals some interesting patterns (Table 3). The Ladinian crisis related to the abrupt deepening of the marine basin (Ruban, 2006b) did not lead to significant changes in the taxonomic diversity structure of the regional brachiopod fauna, largely because Rst calculated for the Anisian, on one side, and Carnian, Norian, and Rhaetian associations, on the other, were reasonably high. Even if the T/J mass extinction occurred in the Northern Caucasus (although it is impossible to document it due to a major hiatus), it did not stress brachiopods enough to provoke any remarkable changes in the taxonomic diversity structure. There is a similarity in this structure between the Rhaetian, Sinemurian, and Pliensbachian associations. The Pl-To mass extinction, which depleted the Toarcian assemblages (Ruban, 2004), brought some changes to the taxonomic diversity structure of brachiopods. Relatively speaking TTDSR accelerated at the Pliensbachian-Toarcian transition (Figure 3). Additionally, there is a remarkable difference between the Late Triassic and the Toarcian-Bajocian associations. Rst calculated between the Rhaetian and Toarcian associations is below 0.00, although Rst between the Rhaetian and Pliensbachian associations is as high as 0.31. It is an intriguing question, whether those Late Triassic superfamilies which survived the T/J extinction, but did not survive the Pl/To extinction were 'dead clades walking' (the term of Jablonski, 2004). The Aalenian potential mass extinction did not result in significant changes in the taxonomic diversity structure of brachiopods, although their diversity decreased (Ruban, 2004). In general, all Sinemurian-Bajocian associations were quite similar, and the Rst value for each pair of them is higher than 0.30.

Surprisingly, it is evident that the taxonomic diversity structure of the Pliensbachian, Toarcian, and Aalenian associations was quite similar to that of the Early Triassic association. Rhynchonelloidea and Wellerelloidea were superfamilies which drove the Early Triassic associations and which also played an important role in the Early Jurassic associations. Thus, those superfamilies which were dominants in the Early Triassic became dominants in the Pliensbachian-Aalenian. However, the Early Triassic assemblage was of low diversity, suggesting the need for further studies.

It is possible to hypothesize that paleoenvironmental differences between stages might have reinforced dissimilarities between the relevant brachiopod assemblages. Changes in water depth or character of sedimentation and so on were responsible only for the "ordinary" changes in the brachiopod assemblages, reflected by the Rst values for successive associations. The Rhaetian and Sinemurian paleoenvironments were distinctly different, while the similarity between the respective associations is quite high. In contrast, the conditions in the Toarcian did not differ too much from those in the Pliensbachian, whereas an increase in turnover rate is documented (Table 3). Thus, it is unlikely that those changes in the brachiopods associations at the Triassic-Jurassic and Pliensbachian-Toarcian boundaries were caused by the differences in paleoenvironments.

Discussion

The patterns documented in the Northern Caucasus may be compared with those in the Swiss Alps and the Jura Mountains. The data on those regions have been summarized by Sulser (1999). The suprageneric taxonomy of the Alpine species has been verified in the same way as was done for the Northern Caucasus. It is necessary to note that no Early Triassic, Carnian,

Table 4. Rst values for the Triassic-Bajaocian brachiopod associations of the Alps and Jura (data from Sulser, 1999). There were no brachiopods in the Early Triassic, Carnian, and Norian. Values higher than +0.2 are highlighted as bold. Significant values are given in italics. Stage abbreviations—see Table 1.

	T1	AN	LA	CA	NO	RH	HE	SI	PL	TO	AA	BJ
T1	1	0	0	0	0	0	0	0	0	0	0	0
AN		1	***0.7***	0	0	**0.4**	0	−0.1	−0.1	−0.2	−0.2	−0.2
LA			1	0	0	0.1	0.2	0.1	−0.2	−0.3	−0.3	−0.3
CA				1	0	0	0	0	0	0	0	0
NO					1	0	0	0	0	0	0	0
RH						1	0.1	0.1	0.2	−0.2	0.2	**0.3**
HE							1	***0.6***	**0.3**	−0.2	0	−0.1
SI								1	***0.7***	0	**0.3**	**0.3**
PL									1	0.2	***0.4***	**0.4**
TO										1	**0.3**	**0.4**
AA											1	***0.9***
BJ												1

and Norian brachiopods have been reported from the Alps and the Jura (Sulser, 1999).

The calculation of Rst values for the associations of the Alpine brachiopods (Table 4) demonstrates significant turnover at the Triassic-Jurassic transition, evident from the low Rst values calculated for the Rhaetian and Hettangian and Rhaetian and Sinemurian assemblages. The same values in the Northern Caucasus are higher (Table 3). The next important conclusion is the slight increase in similarity between the Rhaetian and the Pliensbachian-Bajocian associations. This trend was interrupted in the Toarcian. Probably, the total poverty of this association explains the lack of similarity between the Toarcian association and those of other ages. Thus, the patterns documented in the Alps and the Jura contrast to those established in the Northern Caucasus. This might have been related to paleoenvironmental differences between these regions. The Aalenian potential mass extinction provoked a brachiopod diversity drop in the Alps and the Jura (Ruban, 2004), although its influences on the taxonomic diversity structure remain unclear.

Conclusions

This study established the influences of the early Mesozoic crises on the regional evolution of brachiopods. The P/T mass extinction in the Northern Caucasus resulted in a recovery with a consequent increase in the rapid changes in the taxonomic diversity structure of the Early Triassic-Anisian associations. The Ladinian crisis did not provoke any major changes in this structure. The T/J and Pl/To mass extinctions did not lead to significant turnover among brachiopods, but they permitted those superfamilies that dominated in the Early Triassic to rise again. The Pl/To event affected the superfamilies that dominated the species diversity in the Late Triassic. The data from the Alps and the Jura Mountains provide conclusions different from those made in the Northern Caucasus, e.g., a strong turnover at the Triassic-Jurassic transition is reported in the Alpine Region.

Were the early Mesozoic mass extinctions able to reset evolution? Even if our results presented here can be interpreted in the affirmative, such a reset was not complete and, therefore, they do not contradict the observations made by Jablonski (2004).

Acknowledgements

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PALEOZOIC PALAEOGEOGRAPHIC FRAMEWORKS OF THE GREATER CAUCASUS, A LARGE GONDWANA-DERIVED TERRANE: CONSEQUENCES FROM THE NEW TECTONIC MODEL

Dmitry A. Ruban

Abstract - The Greater Caucasus is a large region between the Black and Caspian seas. The new model of its Paleozoic tectonic evolution, coupled with the analysis of distribution of deposits of distinct ages, makes it possible to propose a set of regional palaeogeographic frameworks. Until the mid-Silurian, this region was incorporated into the continental land mass of Gondwana, being subjected either partly or completely to the successive transgressions from the Prototethys Ocean. In the late Silurian-Devonian, terrestrial environments in the Greater Caucasus were related to islands or island chain along the Hun Superterrane, whereas transgressions from oceans located to the north (Rheic, Rhenohercynian, and Khanty-Mansi oceans) and south (Palaeotethys Ocean) established the marine environments. In the Carboniferous-Permian, the Greater Caucasus was linked to the continental landmass of the Laurussian counterpart of Pangaea, and transgressions occurred from the Palaeotethys Ocean. Carbonate platforms and reefs grew at least twice in the Paleozoic history of the study region. In the Famennian, an isolated carbonate platform appeared, whereas a rimmed shelf attached to the Pangaean margin is known from the Changhsingian. In both cases, the equivalents of these events are found in other parts of the "Tethyan Region".

Key words: palaeogeography, terrane, carbonate platform, Paleozoic, Tethys, Greater Caucasus.

Introduction

Present global reconstructions of Paleozoic plate tectonics (Cocks & Torsvik, 2002; Stampfli & Borel, 2002; Von Raumer *et al.*, 2002, 2003; Scotese, 2004; Torsvik & Cocks, 2004) provide a new basis for discussion of the geological history of many regions. This is especially apt in those regions whose evolution has been described in terms of outdated concepts such as the "geosyncline paradigm". An important consequence of the development of new models of regional tectonic evolution is a reconsideration of existing palaeogeographic reconstructions.

The Greater Caucasus is a large region which stretches as a mountain chain between the Black Sea and the Caspian Sea and embraces the territory of southwesternmost Russia, northern Georgia, and northwestern Azerbaijan (Fig. 1). Due to the key position of this region between the Eastern European, Turkish and Iranian domains, reconstruction of its Palaeozoic geological history is significant for the understanding of processes that occurred within the "Tethyan Region". As well, the Paleozoic of this region is of interest because of its economic significance. In some other regions (northeastern Africa, Arabia,

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Fig. 1 - Geographical location of the Greater Caucasus.

Turkey) whose geological history was linked anyhow with the Greater Caucasus, Silurian strata are important source rocks for oil generation (SHARLAND *et al*., 2001; TAWADROS, 2001; TAWADROS *et al*., 2006; VAROL *et al*., 2006). Exploration of the same stratigraphic interval in the Greater Caucasus to evaluate its source-rock potential was suggested by TAWADROS *et al.* (2006). New models have been proposed to describe the geological history of the Greater Caucasus in the Mesozoic-Cenozoic (ERSHOV *et al*., 2003), and, particularly, in the Triassic (GAETANI *et al*., 2005) and Jurassic (RUBAN, in press). The model of NATAL'IN & ŞENGÖR (2005) also has some value in understanding the tectonic evolution of the Greater Caucasus. These replace the traditional "geosyncline" model (see e.g., LAZ'KO, 1975). For the Paleozoic, however, traditional palaeotectonic views have persisted and there has been no appropriate discussion of the regional palaeogeography in the global context. Recently, TAWADROS *et al*. (2006) proposed a comprehensive model of the tectonic evolution of the Greater Caucasus for the entire Phanerozoic (see below for details). In the present paper, the palaeogeographic consequences of this new model are examined, and special attention is given to the determination of the position of the Greater Caucasus with respect to the land masses and oceans existed at those times.

Geological setting

The Greater Caucasus Terrane was identified by GAMKRELIDZE (1997) and later, independently, by TAWADROS *et al*. (2006). At present it is located between the margin of the stable Russian Platform and the Lesser Caucasus Terrane (Fig. 1). It is part of a large orogen deformed and uplifted in the late Cenozoic (ERSHOV *et al*., 2003). The Paleozoic sedimentary complexes, whose total thickness exceeds 20,000 m, are exposed in its central part. The Paleozoic Greater Caucasus Terrane thus forms the core of the present Greater Caucasus. The Paleozoic deposits are dominated by shale and siliciclastic rocks, with the common addition of volcanoclastic and volcanic rocks (RUBAN, 2006) (Fig. 2). However, several episodes of carbonate sedimentation, in the mid-Cambrian, late Silurian-earliest Devonian, Late Devonian and Lopingian, are also known. Coal accumulated in the Pennsylvanian. Major unconformities are recognised within the Ordovician and Permian.

A new model of the tectonic evolution of the Greater Caucasus (TAWADROS *et al*., 2006), which is incorporated with plate tectonic reconstructions by STAMPFLI & BOREL (2002), suggests that this terrane was a part of the Afro-Arabian margin of Gondwana until the mid-Silurian, when margin-parallel breakup occurred (Fig. 3). A new Palaeotethys Ocean was formed, separating a string of terranes, together called the Hun Superterrane, from Gondwana. The Greater Caucasus Terrane was one of these terranes. These migrated from Gondwana towards Laurussia, reaching the latter's margin at some time in the Late Devonian. At the same time, the amalgamation of Pangaea began. Long-distance displacements along a major shear zone, which stretched along the northern Palaeotethyan margin and connected with the Intrapangaean shear zones (ARTHAUD & MATTE, 1977; SWANSON, 1982; RAPALINI & VIZÁN, 1993; STAMPLFI & BOREL, 2002; RUBAN & YOSHIOKA, 2005), resulted in the stacking of terranes westwards, so that the Greater Caucasus Terrane became located close to the Carnic Alps. The palaeontological and lithological records of these two regions are very similar. Some similarity is also observed between the Greater Caucasus and Bohemia. In the Late Triassic, the Greater Caucasus Terrane was repatriated eastwards to its present position along the above-mentioned shear zone. This tectonic model, which has been developed on the basis of palaeontological and lithological interregional comparisons, is also supported by the results of earlier palaeomagnetic studies (SHEVLJAGIN, 1986), which suggest the absence of links between the Greater Caucasus and Russian Platform before the end-Triassic.

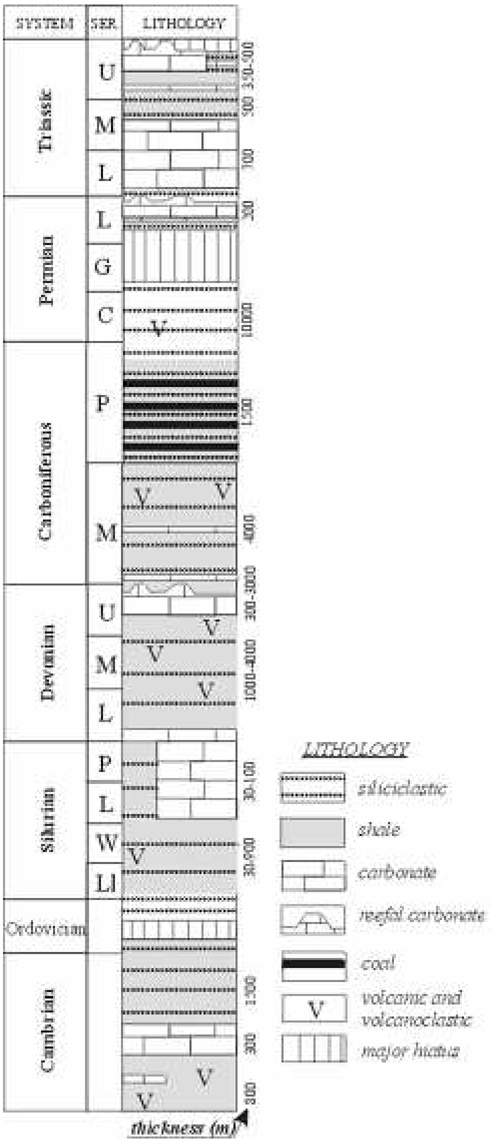

Fig. 2 - Composite section of the Paleozoic-Triassic deposits of the northern Greater Caucasus (modified after RUBAN, 2006). No formal chronostratigraphic units are recognized regionally within the Cambrian-Ordovician.

Materials and methods

The proposed palaeogeographic reconstructions are based on analysis of distribution of the Paleozoic deposits within the Greater Caucasus and interpretation of global reconstructions. The distribution of deposits of distinct ages and their stratigraphic subdivision have been reviewed by PAFFENGOL'TS (1959, 1965), ROBINSON (1965), MIKLUKHO-MAKLAJ & MIKLUKHO-MAKLAJ (1966), ZHAMOJDA (1968), KIZEVAL'TER (1968), KIZEVAL'TER & ROBINSON (1973), ZANINA & LIKHAREV (1975), KOTLYAR (1977), POTAPENKO (1982), OBUT *et al.* (1988), KOVALEVSKIJ & KOTLYAR (1991), BOGOLEPOVA (1997), KOTLYAR *et al.* (1999, 2004) and GAETANI *et al.* (2005). Results from the author's own studies on the Permian Molasse (Cisuralian-Guadalupian), undertaken during past years in the Western Caucasus, were also useful. The following compilation of available data excludes literature citations, in order to avoid multiple repetition of the abovementioned sources.

These data permit the distinction of those parts of the Greater Caucasus that were submarine from those that were exposed as land. This simplest of reconstructions can be incorporated into global plate tectonic reconstructions using the new tectonic model of TAWADROS *et al.* (2006). The global reconstructions of STAMPFLI & BOREL (2002) are preferred because they seem to be the most comprehendsive and well argued. This incorporation allows us to identify the position of the Greater Caucasus relative to the then existing landmasses and oceans, and to relate the regionally documented marine and non-marine environments with them. This creates a set of palaeogeographic frameworks which clearly demonstrate the key features of the regional palaeogeography in the global context. Although further, more detailed studies are necessary, it seems that the drawing of such simple frameworks may form an essential basis for such studies and for further discussion. Moreover, it will be sufficient to correctly plot the Greater Caucasus on global maps constrained for the Paleozoic

Fig. 3 - Palaeogeographical (asterisk) position of the Greater Caucasus. Palaeotectonic maps are simplified from Stampfli & Borel (2002). Oceans are shown as white.

timeslices. Somewhat the same approach was used by Ruban (in press) in his regional Jurassic reconstructions.

The palaeogeographical frameworks have been constrained for nine time intervals, each characterized by a lack of significant palaeogeographical change (Fig. 4). These intervals are as follows: Cambrian, Ordovician, Llandovery-Wenlock (early Silurian), Ludlow-Lochkovian (late Silurian-earliest Devonian), Pragian-Frasnian (Early-Late Devonian), Famennian (Late Devonian), Mississippian (early Carboniferous), Pennsylvanian-Guadalupian (late Carboniferous-Middle Permian), and Lopingian (Late Permian). Stratigraphic nomenclature used in this paper follows the recommendations of the International Commission on Stratigraphy (Gradstein *et al*., 2004; see also official web site: www.stratigraphy.org). Formal stratigraphic names are written with an initial captial letter (e.g., Ordovician, Early Devonian, Changhsingian) to distinguish them from the informal names (e.g., late Paleozoic, early Silurian, mid-Permian). The own series names, which have been adopted formally (e.g., Ludlow, Pennsylvanian, Lopingian), are preferred.

Paleozoic palaeogeographic frameworks

The key elements of world palaeogeography are large continental landmasses, small continents, and large islands or island chains, whose size was comparable with that of small continents (Cocks & Torsvik, 2002; Stampfli & Borel, 2002; Von Raumer *et al*., 2002, 2003; Scotese, 2004; Torsvik & Cocks, 2004). Gondwana - an assembly of plates in the Southern Hemisphere - existed since the Neoproterozoic, although its final amalgamation might have occurred in the beginning of the Paleozoic. Baltica and Laurentia were individual plates until the mid-Paleozoic, at which time they amalgamamted to form a distinct continent, Laurussia. In the late Paleozoic, Gondwana and Laurussia were amalgamated in part, and a superconitnent Pangaea appeared. An example of a small continent is Siberia, but this and other "lesser continental landmasses", which corresponded to relatively small-sized plates or tectonic blocks, were located far from the sector that is meaningful for our study. Large islands or island chains (together called "island masses") were formed via the derivation of ribbonlike terranes. The latter usually moved from

Gondwana towards Laurussia, against the margin of which they docked. These island masses were Avalonia-Cadomia, Hun and Cimmeria. They were detached from Gondwana in the Ordovician, Silurian and Permian respectively (COCKS & TORSVIK, 2002; STAMPFLI & BOREL, 2002). In the Devonian, the Hun Superterrane was divided into the European Hunic terranes and Asiatic Hun terranes. Other island masses were Serindia and Kazakhstan, but both their palaeotectonic and palaeogeographic interpretation still seems to be incomplete. In the Paleozoic, several oceans stretched between Laurentia and Baltica in the north and Gondwana in the south. They may be referred together as the "Tethyan oceans". Although there was a Tethyan palaeobiogeographic realm, there was no unique ocean which might have been called "Tethys Ocean". The best nomenclature of these oceans was proposed by STAMPFLI & BOREL (2002), who identified the Prototethys (early Paleozoic), Rheic (early-middle Paleozoic), Palaeotethys (middle Paleozoic-early Mesozoic) and Neotethys (late Paleozoic-early Cenozoic). These oceans were formed each time new breakup occurred along the Gondwanan margin. The Protetethys Ocean should be distinguished from the Tornquist and Iapetus oceans, which also existed in the early Paleozoic. In the Silurian, the Rheic Ocean was unified geographically with the Prototethys. Additionally, the Rhenohercynian Ocean was opened along the southern margin of Laurussia in the Devonian. COCKS & TORSVIK (2002) and TORSVIK & COCKS (2004) generally support this nomeclature, although their plate tectonic reconstructions differ in somewhat from those of STAMPFLI & BORel (2002) and VON RAUMER *et al.* (2002, 2003).

The palaeogeographical frameworks derived for the Greater Caucasus are presented in Fig. 4 and described below.

Cambrian deposits are known in the central and western parts of the Greater Caucasus Terrane (Fig. 4A). A typical marine fauna, including trilobites, brachiopods and archaeocyaths, has been found there. Although the Cambrian strata outcrop in only few sections, we may hypothesize that the terrane was entirely marine at this time. In the early Paleozoic, the Greater Caucasus was a part the Afro-Arabian margin of Gondwana (TAWADROS *et al.*, 2006). Thus, the sea would have transgressed from the north, i.e., from the Prototethys Ocean.

Although it is generally assumed that no strata of Ordovician age exist in the Greater Caucasus Terrane (Fig. 4B), the presence of Lower and Upper Ordovician deposits is hypothesized here. If the former is true, this region was land at that time. Its location on the Gondwanan periphery allows us to conclude that it was a marginal part of that continental landmass. The Prototethys Ocean was situated somewhere to the north. An intriguing question is whether the study region was embraced by the Hirnantian glaciation, which occurred at the end of the Ordovician (BRENCHLEY *et al*., 1994; SMITH & PICKERING, 2003).

Lower Silurian deposits are confirmed only in the north-central Greater Caucasus Terrane (Fig. 4C). These mark a minor transgressive episode, enforced by the abrupt deepening of the terrane margin as indicated by the dominance of relatively deep-water facies. Therefore, land occupied most of the Greater Caucasus, which was still connected to Gondwana; the marine incursion may be linked to the Prototethys Ocean. These palaeogeographic changes are readily explained by the development of an active margin

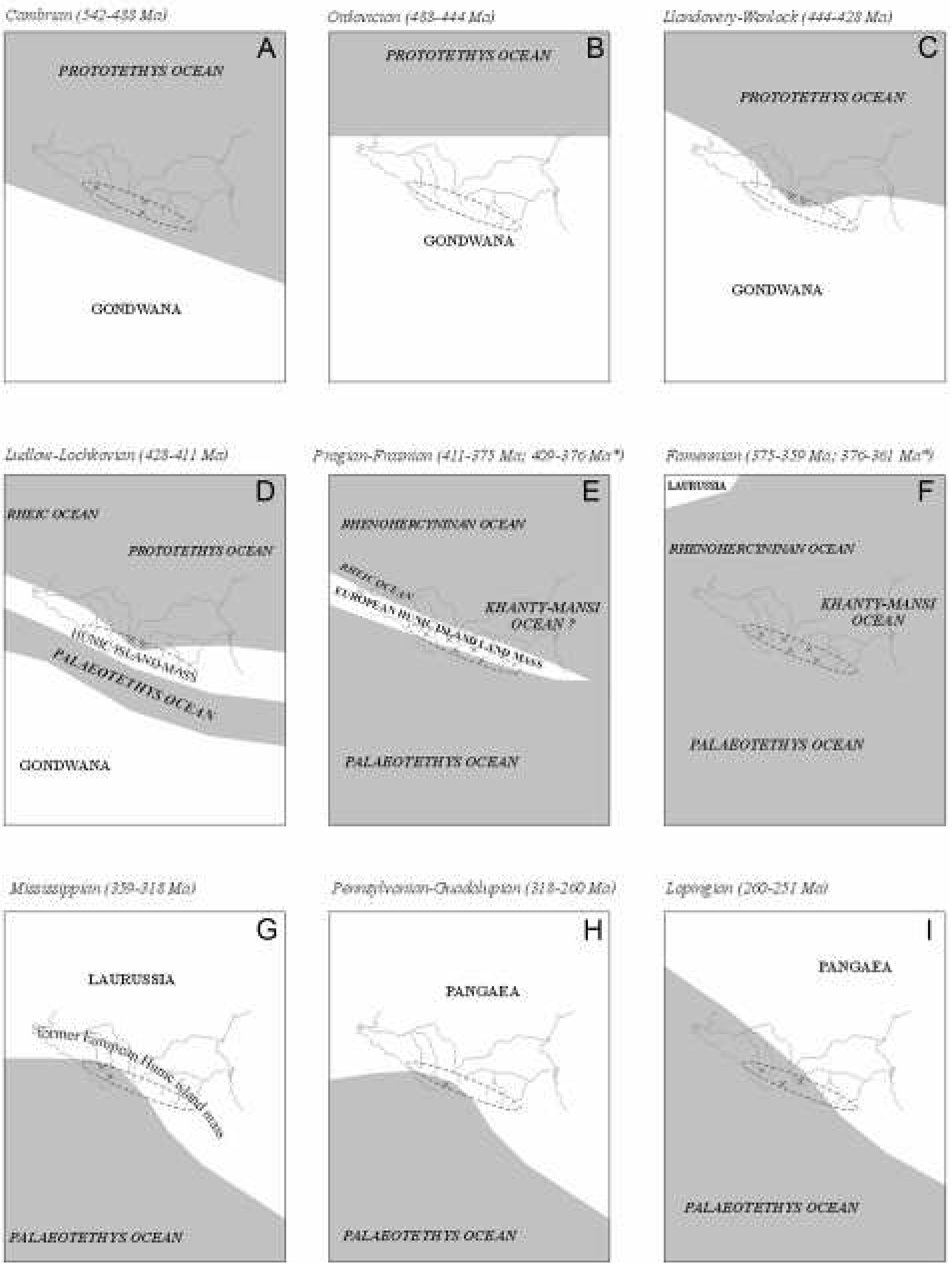

Fig. 4 – Palaeogeographic frameworks of the Greater Caucasus. A) Cambrian, B) Ordovician, C) Llandovery-Wenlock, D) Ludlow-Lochkovian, E) Pragian-Frasnian, F) Famennian, G)Mississippian, H) Pennsylvanian-Guadalupian, I) Lopingian. Present-day geographic outline is shown for general orientation within the deposit distribution for each particular age. Deposit distribution is shown by crosses; oceans are shown as gray; land is shown as white. Position of land and oceans is indicated approximately. The absolute ages are taken from GRADSTEIN *et al.* (2004) and KAUFMANN (2006) (the latter is marked by *).

on the periphery of Gondwana after the partial closure of the Prototethys Ocean (STAMPFLI & BOREL, 2002).

The distribution of upper Silurian deposits was the same as that of the lower Silurian, although succeeding Lochkovian deposits cover a slightly larger area (Fig. 4D). Thus, the general palaeogeographic situation in the northern part of the Greater Caucasus was the same as in the early Silurian. However, dramatic changes occurred to the south. A new ocean opened along the Gondwanan margin; it separated the Hun Superterrane, part of which was the Greater Caucasus Terrane, from Gondwana. Consequently, the land, evidently established in the study territory by the absence of upper Silurian strata in its southern, western and eastern parts, may be referred to the Hunic island mass.

Devonian (Pragian-Frasnian) strata are widely distributed throughout the entire Greater Caucasus (Fig. 4E). However, these mostly comprise thick (up to 4,000 m) volcanic and volcanoclastic deposits. Such a great intensification in regional volcanic activity may be explained in two ways: 1) active contact between tectonic blocks within the assembly of the European Hunic terranes, and 2) location of the Greater Caucasus close to the northern active margin along this superterrane, where the oceans closed (RUBAN & KHOLODKOV, 2006). In the southern part of the Greater Caucasus, marine facies with *Amphipora* and *Favistella* are known. Thus, the volcanic islands of the Greater Caucasus were part of the European Hunic island mass, which separated the closed Rheic Ocean and extended Rhenohercynian Ocean from the large Palaeotethys Ocean. The marine basin periodically transgressed some areas of this island mass, as confirmed by the local presence of rare mid-Devonian carbonates with marine fauna. It is possible to hypothesize that the Greater Caucasus lay on the eastern edge of the European Hunic island mass. If so, the Khanty-Mansi Ocean was located to its north. This ocean moved close to the European Hunic terranes after the partial closure of the Rheic Ocean (STAMPFLI & BOREL, 2002).

Famennian deposits are known throughout the entire Greater Caucasus Terrane (Fig. 4F). In its western and northern parts they are represented by carbonates with abundant marine fauna and reefs. However, relatively deep-water facies are distributed through the southern part of this terrane. Thus, there was no land in the latest Devonian within the study region. The remnants of the Rhenohercynian and Khanty-Mansi oceans united with the Palaeotethys Ocean, which reached its maximum size at that time, to form a single water mass. In the Late Devonian, the European Hunic terranes reached the Laurussian margin. It is possible to say that these terranes abutted Laurussia at their western edge, whereas the Greater Caucasus Terrane was on the eastern edge.

Marine Mississippian deposits are known in the western part of the Greater Caucasus Terrane (Fig. 4G). From the beginning of the Carboniferous, the European Hunic terranes became amalgamated with the Laurussian margin, and the activity on the major shear zone led to their westward stacking (see above). Thus, the sole possibility to explain the marine sedimentation in the western Greater Caucaus is to assume a marine incursion from the Palaeotethys Ocean. Land present in the central and eastern parts of the study region may be assigned to the continental landmass of Laurussia.

Late Carboniferous to mid-Permian (Pennsylvanian-Gaudalupian) strata are known within the western and central Greater Caucasus (Fig. 4H). However, marine facies are known only along its southwestern periphery. Coal-bearing and molassic deposits (the latter dominated by siliciclastic sediments of variegated colour) accumulated in the other areas. Thus the marine basin, again referred to the Palaeotethys Ocean, inundated only a small part of the study terrane, whereas the remainder was land included in the continental landmass of the Laurussian counterpart of the supercontinent of Pangaea.

Late Permian (Lopingian) marine deposits, which were carbonates with the richest fossil assemblages (foraminifers, brachiopods, bivalves, gastropods, trilobites, etc.), are distributed across the western and central parts of the Greater Caucasus Terrane (Fig. 4I). This transgression evidently occurred from the Palaeotethys Ocean, whose narrowing began due to the opening and rapid growth of the Neotethys Ocean in the south. Non-marine environments, if they existed in the eastern part of the Greater Caucasus, are referred to the continental landmass of Pangaea, whose breakup had already begun (Stampfli & Borel, 2002).

Discussion

Two recognizable episodes of carbonate platform and reef growth occurred within the Greater Caucasus in the Famennian and Changhsingian (Ruban, 2005, 2006). For the Famennian, Ruban (2005) hypothesized the presence of a large carbonate platform of rimmed shelf type. However, the present reconstruction (Fig. 4F) does not support such a conclusion. There was no sufficiently large landmass on whose margin such a shelf might have been located. Possibly this was an isolated carbonate platform associated with the drowned land formerly existing along the axis of the Greater Caucasus Terrane. Although Frasnian reefs are much better known, Famennian reefs were also widely distributed globally (Webb, 2002). They are known along the southern margin of Laurussia and in Kazakhstan, South China and Western Australia. It is possible to say that they developed along the entire northern and eastern periphery of the Palaeotethys Ocean, but were absent on its southern periphery. If so, we may confirm that the Famennian was the time when the Greater Caucasus Terrane reached or almost reached the Laurussian margin.

The late Changhsingian reefs, known in the western part of the Greater Caucasus, bounded the carbonate platform attached to the continent (Ruban, 2005). The latter was the Laurussian counterpart of Pangaea. Other carbonate platforms, such as those in the Dinarides (Sremac & Marjanac, 2003; Vlahović *et al.*, 2005), existed on the margins of the Palaeotethys and Neotethys oceans. They were precursors of the large carbonate platforms which appeared in the western part of the "Tethyan Region" in the Mesozoic (Golonka, 2004; Vlahović *et al.,* 2005).

Conclusions

The interaction of global tectonic processes in the "Tethyan Region", i.e., between Laurussia and Gondwana, created a complicated geological history for the Greater Caucasus. The constrained palaeogeographic frameworks (Fig. 4) suggest that the Palaeozoic environments interpreted within the Greater Caucasus were related to several major elements of the global palaeospace, namely the Gondwana and Laurussia continental landmasses, the Hunic island mass (i.e., a unique island or a chain of lesser islands), and the Prototethys, Palaeotethys, Rheic, Rhenohercynian and Khanty-Mansi oceans.

Further studies are necessary to refine our knowledge of the Palaeozoic evolution of the Greater Caucasus. The key topics for these are as follows: 1) justification of the position of the Greater Caucasus in the Cambrian-Ordovician, 2) exploration of the possible regional appearance of the Hirnantian glaciation, and 3) discussion of the mid-Devonian palaeoenvironments, when intense regional volcanism occurred.

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Late Paleozoic Transgressions in the Greater Caucasus (Hun Superterrane, Northern Palaeotethys): Global Eustatic Control

Trangresiones del Paleozoico tardío en el Gran Caucaso (Superterreno Hun, PaleoTesis Norte)

RUBAN, DMITRY A.[1,2]

Abstract

The Greater Caucasus (a mountain chain in southwesternmost Russia, northern Georgia, and northwestern Azerbaijan) is a Gondwana-derived terrane, which was included into the Hun Superterrane (late Silurian-Late Devonian) and then docked at the Laurussian margin of the Palaeotethys Ocean close to the terranes of the present Alps (Late Devonian-Middle Triassic). The Upper Paleozoic sedimentary complex, up to 20,000 m thick, provides a good record to discuss global eustatic changes. Three transgressions are reported in this region, which occurred in the Lochkovian, Frasnian-Famennian, and Changhsingian. The first of them embraced the northern part of the Greater Caucasus, the second was larger and covered this region entirely, whereas the third occurred in its western part only. All of them corresponded evidently to global eustatic rises, and, therefore, their explanation does not require an implication of the regional tectonic activity. These regional transgressive episodes are also known from the Southern and Carnic Alps, Arabia, and Northern Africa. A correspondence between the Late Permian marine sedimentation in the Greater Caucasus and non-marine sedimentation in Spain is established. Thus, they were of planetary extent and the present global eustatic curve is confirmed. The regional transgressions resulted in carbonate deposition, biotic radiations, and reefal growth.

Key words: palaeogeography, Late Paleozoic, sea-level changes, eustasy, Greater Caucasus, Paleotethys, Russia

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INTRODUCTION

The global sea-level fluctuated strongly during the Late Paleozoic. Well-known earlier reconstructions of this eustasy were made by Vail et al. (1977), Johnson et al. (1985), Ross & Ross (1985), Hallam (1992), Ronov (1994), and Hallam & Wignall (1999). Recently, Haq & Al-Qahtani (2005) have proposed a new global eustatic curve. In general, it is concluded that the sea level dropped gradually during the Late Paleozoic, with a pronounced drop in the mid-Permian. Second-order fluctuations were superposed on this trend. Unfortunately, our knowledge of the Paleozoic sea-level changes remains limited (see also Miller et al. 2005). For example, the present curve of Haq & Al-Qahtani (2005) suggests a major regression in the end-Permian, although Hallam & Wignall (1999) argued for a major transgression at this time. By the same token, sea-level changes at the Frasnian-Famennian transition are not clear (Racki 2005). Verification of proposed eustatic curves as well as their details, improvement, and justification is possible only by careful comparison of numerous regional data from across the world. An example from the Jurassic demonstrates that such analysis would significantly contribute both to the identification of the global sea-level changes and their explanation (Hallam 2001). A reconstruction of the global eustatic curve is an enormously difficult task and some doubts are even expressed as to its existence (e.g., McGowran 2005). However, such a curve will be an important key to explain the changes in world palaeogeography, sedimentary environments, and biotic evolution. In the regional record, we document the global sea-level changes by the transgressive and regressive episodes. However, the latter may also reflect (and almost always do anyway!) the regional tectonic influences. Thus, our task is to differentiate between the eustatic- and tectonic-induced regional sea-level changes. Hallam (2001) made an intriguing conclusion, that transgressions are more evident when traced globally, than regressions, which reflect mostly the regional tectonic movements. In the author's opinion, this does not diminish the importance of world correlation of regressive episodes and unconformities, but emphasizes the need for global tracing of the regional transgressions.

The Greater Caucasus Terrane, presently included into the Alpine Mediterranean Belt and located in the Southwest of Eurasia (Fig. 1), provides an exceptional Late Paleozoic record. According to the present tectonic model (Tawadros et al. 2006; Ruban et al., 2007), this region was one of the Hunic terranes identified by Stampfli & Borel (2002), and therefore, its record is meanful for both the Afro-Arabian margin of Gondwana and Variscan Europe. This article is the first, which attempts to give a comprehensive, although brief synthesis of knowledge on the Late Paleozoic transgressions, which occurred in the Greater Caucasus.

GEOLOGICAL SETTING

The Greater Caucasus pertains to a large mountain chain, which is located in the Southwest of Eurasia and connected with the other mountains originated during the Alpine phase of orogenic activity. The Greater Caucasus embraces southwesternmost Russia, northern Georgia, and northwestern Azerbaijan. The Paleozoic sedimentary complexes, whose total thickness is up to 20,000 m, are known in the central part of this territory. They are exposed both in numerous little outcrops and in the continuous sections along the river valleys. Deposits of all three Upper Paleozoic systems are known in the Greater Caucasus (Fig. 1). The Devonian is dominated by volcanics and volcanoclastics, and carbonates are known in the Upper Devonian. The Mississippian is composed of shales, sandstones, and rare carbonates, whereas the Pennsylvanian is represented by non-marine coal-bearing strata. The Cisuralian-

Guadulupian is a typical molasse, whereas the mid-Permian corresponds to a major hiatus. Only in the Changhsingian, did marine sedimentation recommence, when accumulation of sandstones and shales was followed by a remarkable episode of carbonate sedimentation.

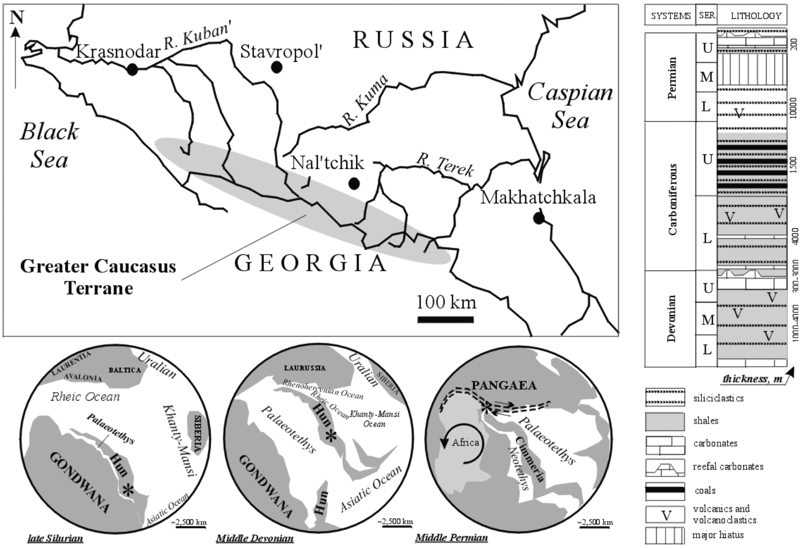

Fig. 1 (Ruban)

Fig. 1. Location of the Greater Caucasus Terrane and a composite lithologic section of the Paleozoic deposits exposed in the northern Greater Caucasus (after Ruban (2006) with additions). The base palaeomaps are simplified from Stampfli & Borel (2002).

The Greater Caucasus Terrane has been identified by Gamkrelidze (1997) and later by Tawadros et al. (2006) and Ruban et al. (2007). The latter authors have also developed a new model, which describes the evolution of the Greater Caucasus (Fig. 1). Until the mid-Silurian, this terrane was a part of the Afro-Arabian margin of Gondwana, i.e., it lay on the southern periphery of the Prototethys and Rheic oceans. In the Ludlow, a breakup occurred along the Gondwanan margin and a ribbon of terranes, called the Hun Superterrane, was formed (Stampfli & Borel 2002; Stampfli et al. 2002; von Raumer et al., 2002, 2003). A new ocean, i.e., the Palaeotethys, was originated in between the Hun Superterrane and Gondwana. This ocean grew in size, whereas the oceans, located between the Hun and Laurussia, were closed. Thus, the Greater Caucasus Terrane together with the other Hunic terranes moved northward. In the Late Devonian, it reached the Laurussian margin. The right-lateral displacements along the major shear zone (Arthaud & Matte 1977;

Swanson 1982; Rapalini & Vizán 1993; Stampfli & Borel 2002; Ruban & Yoshioka 2005) led to the stacking of the Greater Caucasus Terrane somewhere close to the Carnic Alps and Bohemia. In the Late Triassic, the direction of the displacements along the above-mentioned shear zone changed to left-lateral (Swanson 1982; Rapalini & Vizán 1993; Ruban & Yoshioka 2005). The Greater Caucasus Terrane then rapidly reached its present position at the south of Baltica plate.

METHOD: A CONCEPTUAL FRAMEWORK

According to Catuneanu (2006), transgression is defined as a landward migration of the shoreline. It is strongly recommended to make a distinction between transgressions and deepening pulses. Although they are linked in some cases, their true relationships are very complicated (see also Catuneanu 2006). Not in all cases does a transition to facies formed at greater depth, recorded in the sedimentary succession, mark a landward shift of the shoreline. Transgression may have a number of mechanisms, which are distinct in isolated (where are, therefore, not influenced by the global eustasy) and open or half-open basins (Fig. 3).

The Greater Caucasus Terrane was embraced during the Late Paleozoic by marine basins, directly related to the oceans extant at those times. Thus, they were open basins. Consequently, we need to examine two possible explanations of documented transgressions. If they corresponded to the global eustatic events recorded by the curve of Haq & Al-Qahtani (2005), they were eustasy-dominated. When such correspondence is not found, this means tectonic factors were more significant. Alternatively, this also may indicate inaccuracies in the global eustatic curve. The Late Paleozoic transgressions, which occurred in the Greater Caucasus, were recorded thanks to the careful analysis of data on the spatial distribution of deposits of a particular age. These data are contained particularly in the comprehensive overviews by Paffengol'ts (1959), Milukho-Maklaj & Miklukho-Maklaj (1966), Kizeval'ter & Robinson (1973), Obut et al. (1988), Kotlyar et al. (1999, 2004) and Gaetani et al. (2005). The periods, when marine facies became the most wide-spread, were the times of transgressions. Each regional transgressive episode is characterized here in a similar way, i.e., its age, area, sedimentary environments, and possible controls are considered.

A RECORD OF THE REGIONAL TRANSGRESSIONS

Three transgressive episodes may be recorded in the Late Paleozoic history of the Greater Caucasus, namely the Lochkovian, the Frasnian-Famennian, and the Changhsingian episodes.

The Lochkovian regional transgressive episode (D1-RTE) is recorded in the northern part of the Greater Caucasus (Fig. 3). Carbonates with shale interbeds of the upper member of the Manglajskaja Formation, up to 10 m thick, are known there (Obut et al. 1988). In some sections, the lowermost Devonian is represented by shales, siltstones, sandstones with carbonate lenses, whose total thickness reaches 100-150 m. The age of the above-mentioned strata is established precisely with conodonts, and these deposits also contain bivalves trilobites, and tentacultes. The Lochkovian trangsression is evident in the valley of the Malka River, where the Devonian deposits, including basal sandstones with gravels, overlie the lower Silurian strata with an evident disconformity (Obut et al. 1988). Thus, it seems that transgression was directed eastward. At the beginning of the Devonian, the Greater Caucasus Terrane together with the other Hunic terranes moved northward (Stampfli & Borel 2002; Tawadros et al. 2006; Ruban et al., 2007). The absence of the Lochkovian deposits in the southern part of the Greater Caucasus Terrane may be explained by the inclusion of the latter into a large island, which existed along the central axis of the Hun Superterrane. If so, the documented transgression occurred from the Rheic Ocean. This episode was relatively short, because already in the late Early Devonian the sea became restricted in the Greater Caucasus, and volcanoclastic deposition started. D1-RTE undoubtedly corresponded to global eustatic rise (Fig. 3). Even if any regional tectonic activity might have affected the relative sea-level, such influences were minor.

The Frasnian-Famennian regional transgressive episode (D3-RTE) is recorded in the entire Greater Caucasus (Fig. 3). This transgression started in the early Frasnian or even in the end-Givetian, when siliciclastic deposits of the Semirodnikovsksja Formation (its total thickness reaches 1,700 m) were deposited on the Early-Middle Devonian complex composed of volcanics and volcanoclastics. However, the transition between under- and overlying deposits was gradual, and the amount of conglomerates increases upwards (Kizeval'ter & Robinson 1973). These strata are overlain by the carbonates, including reefal limestones, of the Pastukhovskaja Formation, up to 3,000 m thick. The age of these strata is established as Famennian (Kizeval'ter & Robinson 1973). In the southern part of the Greater Caucasus, the upper Frasnian-Famennian Kirarskaja Formation consists of sandstones and shales with limestone interbeds. Stratigraphic relationships between the Devonian formations are not well-justified, and therefore, it becomes difficult to evaluate the direction of this transgression. In the Late Devonian, the Greater Caucasus Terrane was docked at the Laurussian margin, as well as other so-called European Hunic terranes, although a narrow remnant of the Rhenohercynian Ocean remained open between the latter and Laurussia (Stampfli & Borel 2002; Tawadros et al. 2006). It should be further investigated, whether the opening Palaeotethys or closing Rhenohercynian Ocean embraced the studied terrane. This transgression ended at the Devonian/Carboniferous boundary, because it is marked by an erosional surface (Kizeval'ter & Robinson 1973). During D3-RTE, the global eustatic level fluctuated (Fig. 3), and this regional transgression may be related to the pronounced eustatic rise, which occurred in the late Frasnian-early Famennian (Haq & Al-Qahtani 2005). It is necessary to point out the existing misinterpretation of eustatic changes at the Frasnian-Famennian boundary (Hallam & Wignall 1999; Kalvoda 2002; Racki 2005). Earlier eustatic rise in the late Givetian-early Frasnian may have initiated a marine incursion in the studied territory and deposition of conglomerates. Their long accumulation in the Frasnian may be easily explained by the remarkable global regression (Haq & Al-Qahtani 2005), which did not permit a transition to sandstones and shales or carbonates. Thus, as in the previous case, the regional transgressive episode and associated events corresponded well to the global eustasy, and implication of any regional tectonic activity is not necessary.

The Changhsingian regional transgressive episode (P3-RTE) is recorded in the western part of the Greater Caucasus (Fig. 3). This transgression started with the deposition of sandstones, shales, and carbonates of the Kutanskaja and Nikitinskaja with a total thickness exceeding 50 m (Miklukho-Maklaj & Miklukho-Maklaj 1966). But its peak was reached, when the carbonate-dominated Urushtenskaja Formatio with a thickness of more than 100 m, was formed. This formation also includes reefs. All these strata overlie unconformably the Carboniferous-Permian molasse and other older sedimentary complexes. The age of these deposits is now evaluated with brachiopods, foraminifers and other fauna as the late Chaghsingian (Kotlyar et al. 1999, 2004; Gaetani et al. 2005). However, there is some evidence for a conformable contact with the underlying deposits locally (Miklukho-Maklaj & Miklukho-Maklaj 1966). It seems that transgression occurred from the southwest, because the Early-Middle Permian marine environments were established only there. In the Late Permian, the Greater Caucasus Terrane was amalgamated with the Laurussian margin of Pangaea, and it was located somewhere close to the terranes of the present Alps (Tawadros et al. 2006; Ruban et al., 2007). The central and eastern parts of the Greater Caucasus, where the Lopingian deposits are absent, were evidently included into the continental land mass. This transgressive episode was the shortest. It was ended already by the earliest Triassic, because an unconformity is established at the base of the Triassic sedimentary complex (Miklukho-Maklaj & Miklukho-Maklaj 1966; Gaetani et al. 2005). P3-RTE corresponded to a low-amplitude, but still globally-recognizable eustatic rise (Fig. 3). Moreover, recent studies argue that this global transgression strengthened at the Permian/Triassic boundary (Hallam & Wignall 1999; Wignall 2004; Racki & Wignall 2005; Erwin 2006), which is not reflected on the curve of Haq & Al-Qahtani (2005). This well explains marine deposition until the earliest Triassic in the Greater Caucasus. The age of the Abagskaja Formation, which overlies the Urushtenskaja Formation and consists of limestones of about 25 m thickness (Miklukho-Maklaj & Miklukho-Maklaj 1966), is considered as latest Permian-earliest Triassic, because of an extremely impoverished faunistic complex (Miklukho-Maklaj & Miklukho-Maklaj 1966). This biotic crisis is explained by the devastating mass extinction occurring directly at the Permian/Triassic boundary. This gives us the age of the Abag Formation. Thus, the end-Permian regional transgression occurred from the Palaeotethys Ocean thanks to the global eustatic rise. No tectonic forces are necessary to explain this regional episode, although the origin of the Alpine-type structures in the entire Proto-Alpine Region and associated extension (Krainer 1993) might have reinforced transgression.

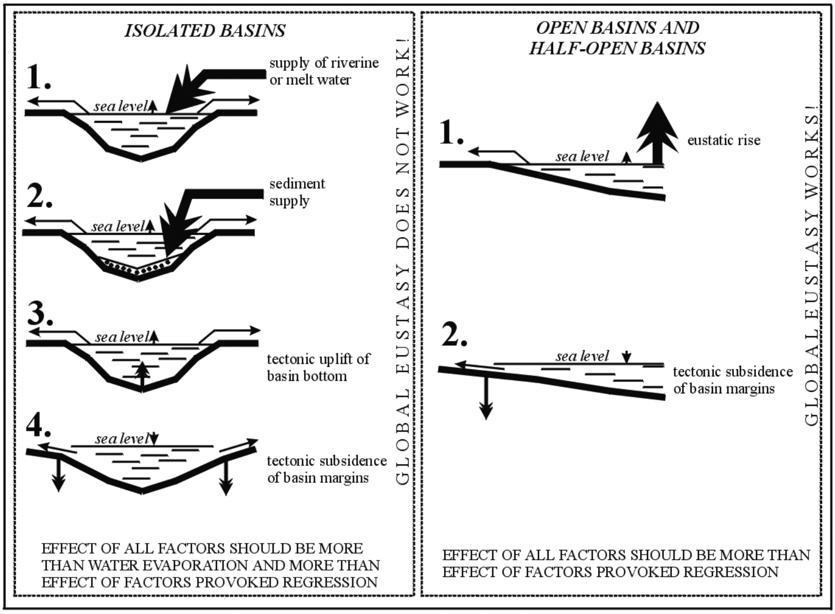

Fig. 2 (Ruban)

Fig. 2. Mechanisms of transgressions in isolated and open and half-open basins. In each particular case, the mechanism of transgression may be complicated and may have included elements from two or more idealized models.

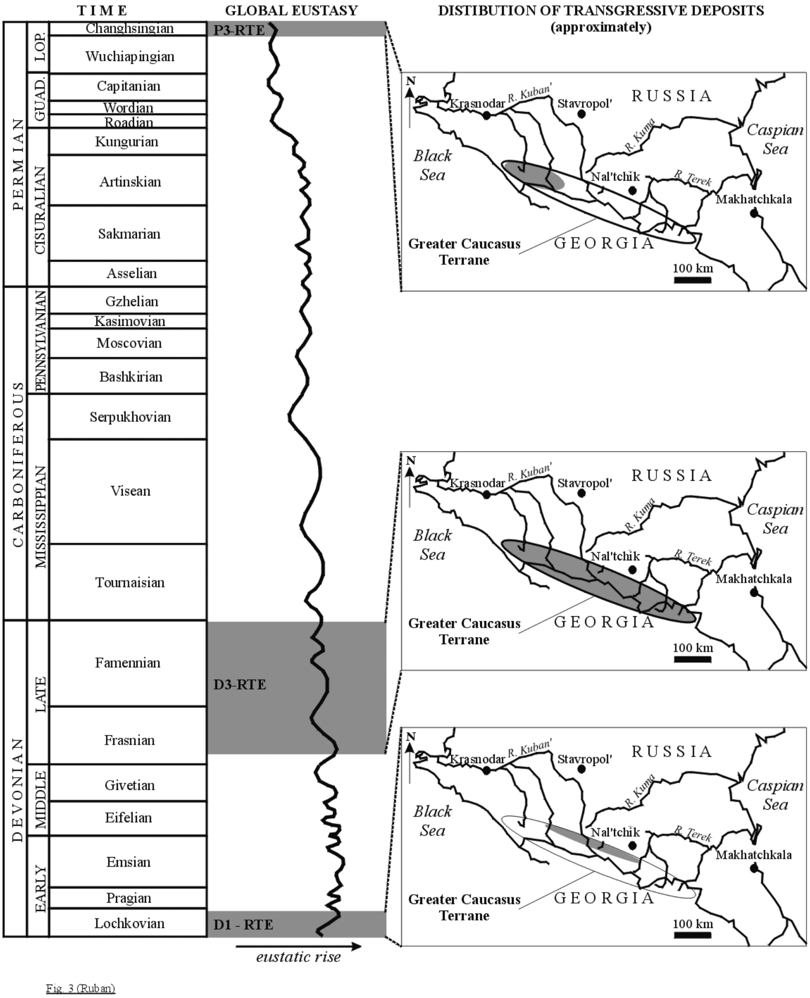

Fig. 3. The Late Paleozoic global eustatic curve (after Haq & Al-Qahtani 2005) and regional transgressive episodes (highlighted by gray). Abbreviations: Guad. - Guadalupian, Lop. - Lopingian. See text for explanation of the regional transgressive episodes.

DISCUSSION

It is important to look for analogues of the Late Paleozoic transgressions, which occurred in the Greater Caucasus. In the Austrian and Italian Southern and Carnic Alps, the Lochkovian and Lopingian transgressions were remarkable events (Krainer 1993; Schönlaub & Histon 1999; Venturini 2002). Although the evidences for Late Devonian shoreline shifts are less clear, a transgressive surface at the base of the lithostratigraphic unit named “calcari a goniatiti e climenia”, whose age is late Frasnian - early Famennian, has been documented in the Carnic Alps (Venturini 2002). In Arabia, transgressions occurred in the end-Silurian, twice in the Famennian, and in the entire Lopingian (Sharland et al. 2001; Haq & Al-Qahtani 2005). In Northern Africa, the Famennian transgression is evident, whereas there was a significant regression in the Lochkovian and at the Permian/Triassic boundary; however, a transgression is known in the Lopingian (Guiraud et al. 2005). Thus, in spite of observed time differences, which may be caused by improperly understood stratigraphic framework and imperfect correlations both in the Greater Caucasus and other regions or by regional tectonic influences, we can suggest that the Late Paleozoic transgressions in the Greater Caucasus were analogous to those in other regions. This indicates their global extent, and also supports the curve of Haq & Al-Qahtani (2005).

It is very intriguing that the latest Permian transgression in the Greater Caucasus coincided with the onset of the Buntsandstein sedimentation in Spain, particularly in the Cordillera Ibérica and the Cordillera Costero-Catalana (Vera 2004). Such a correspondence can be explained by the relation of the Late Permian sedimentation (either marine or non-marine) in the Carnic Alps, Spain, and the Greater Caucasus to the beginning of extension, which embraced at least entire Southern Europe (see e.g., Krainer 1993; Stampfli & Borel 2002).

All three Late Paleozoic transgressions in the Greater Caucasus were expressed by carbonate deposition (see above). Moreover, D3-RTE and P3-RTE occurred at times of reefal growth on the periphery of carbonate platforms (Ruban 2005, 2006). In the Late Devonian, rimmed shelf was attached to the Hun island, whereas in the Lopingian, a carbonate platform of the same type was attached directly to the continental margin of Pangea. Although Khain (1962) argued for tectonic control of the Late Devonian and Late Permian reef distribution in the Greater Caucasus, we may now postulate that the appearance of these reefs might have resulted directly from the eustatically-driven transgressions. By the same token, all reported regional transgressive episodes undoubtedly coincided with biotic radiations. Available data on brachiopods suggest their Early and Late Devonian and Late Permian diversifications (Ruban 2006). The same events are also known in the regional evolution of other fossil groups like bivalves, trilobites, corals, and bryozoans (Paffengol’ts 1959; Miklukho-Maklaj & Miklukho-Maklaj 1966; Nalivkin & Kizel’vater 1973; Obut et al. 1988; Kotlyar et al. 1999, 2004). Transgressions led to the appearance of relatively shallow-marine environments on shelves, which were favorable for marine fauna. When reefal communities grew up, this accelerated biotic radiations as this was previously hypothesized by Ruban (2006).

CONCLUSIONS

Three regional transgressive episodes are known in the Late Paleozoic history of the Greater Caucasus Terrane. They were the Lochkovian, Frasnian-Famennian, and Changhsingian transgressions. Attempted comparison of them with the present global eustatic curve (Haq & Al-Qahtani 2005) and data from other regions (Alpine Europe, Northern Africa, and Arabia) suggests that all these transgressions were eustatically-controlled, and the role of regional tectonic activity to explain them seems to have been insignificant. However, the specifics in the regional tectonic evolution may explain why other transgressions documented by the global curve did not appear in the studied region. Thus, a conclusion that transgressions can be traced globally, analogous to the changes in the global sea level made by Hallam (2001) for the Jurassic, can now also be inferred for the Late Paleozoic. It is also possible to state, that the global eustatic curve of Haq & Al-Qahtani (2005) is confirmed with the data from the Greater Caucasus Terrane.

Another important conclusion from the study is that all three regional transgressions were expressed by carbonate deposition and biotic radiations, and two of them also coincided with reefal growth.

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Natura Nascosta	Numero 35	Anno 2007	pp. 24-35	Figure 4

THE SOUTHWESTERN MARGIN OF BALTICA IN THE PALEOZOIC-EARLY MESOZOIC: ITS GLOBAL CONTEXT AND NORTH AMERICAN ANALOGUE

Dmitry A. Ruban

Abstract - Recent studies have provided evidence for long-termed activity of the shear zones at the southwestern margin of Baltica during the Paleozoic-early Mesozoic. In the late Paleozoic, the Northern Palaeotethyan Shear Zone formed the margin of Baltica and was one of the elements of the global system of megashears. The palaeotectonic analogue of the Paleozoic-early Mesozoic southwestern margin of Baltica may be found in North America, where long-distance displacements occurred at the western and southwestern margin during the Mesozoic-Cenozoic.

Key words: shear zone, palaeotectonic analogue, Baltica, North America, Paleozoic, Mesozoic.

Introduction

The tectonic history of Baltica was reviewed by BOGDANOVA (1993), BOGDANOVA *et al*. (1996), NIKISHIN *et al*. (1996), TORSVIK *et al*. (1996), ARTEMIEVA (2003), and COCKS & TORSVIK (2005). A geologic abruptness of the western Baltica boundary was suggested much earlier (TORNQUIST, 1908). However, many relevant questions remain. The recent studies of the southwestern margin of Baltica (e.g., SAINTOT *et al*., 2003; DADLEZ *et al*., 2005; RUBAN & YOSHIOKA, 2005) suggest that the traditional concepts of the geodynamic evolution of this region need to be re-evaluated.

Geodynamic outline

In the Paleozoic, Baltica, now a part of Eurasia, was a medium-sized continent, which included the present-day territory of European Russia, Ukraine, Belorussia, the Baltic countries, the northeastern part of Poland, and the Scandinavian Peninsula. COCKS & TORSVIK (2005) have described Baltica as a terrane. Two other terms "East European Craton" and "Russian Platform" are often referred to the same territory as "Baltica". To avoid misunderstanding, I propose their definition herein. Baltica is considered only as a plate, which existed since the end-Proterozoic, when a solid continent appeared after the amalgamation of the Precambrian blocks, and already in the Paleozoic included into the larger continent of Laurussia and then Pangaea.

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In contrast, the East European Craton is the present-day tectonic unit, which comprises the second-order structures of former Baltica (Fig. 1). The Russian Platform may be referred to as the assembly of the stable structures between the Carpathians, the Scandinavian Caledonides, the Timan, the Urals, and the Caucasus, i.e., it includes the entire East European Craton, the Scythian Platform (often mentioned as the Scythian Plate), the Peri-Caspian Basin, the Donbass, and some other stable marginal areas. The southwestern margin of Baltica discussed in this paper comprises the present-day structures of the Ukrainian Massif (or the Ukrainian Shield), the Donbass, the Dniepr-Donets Basin, the Pripyat Trough, and the so-called Peri-Tornquist Basin (Fig. 1).

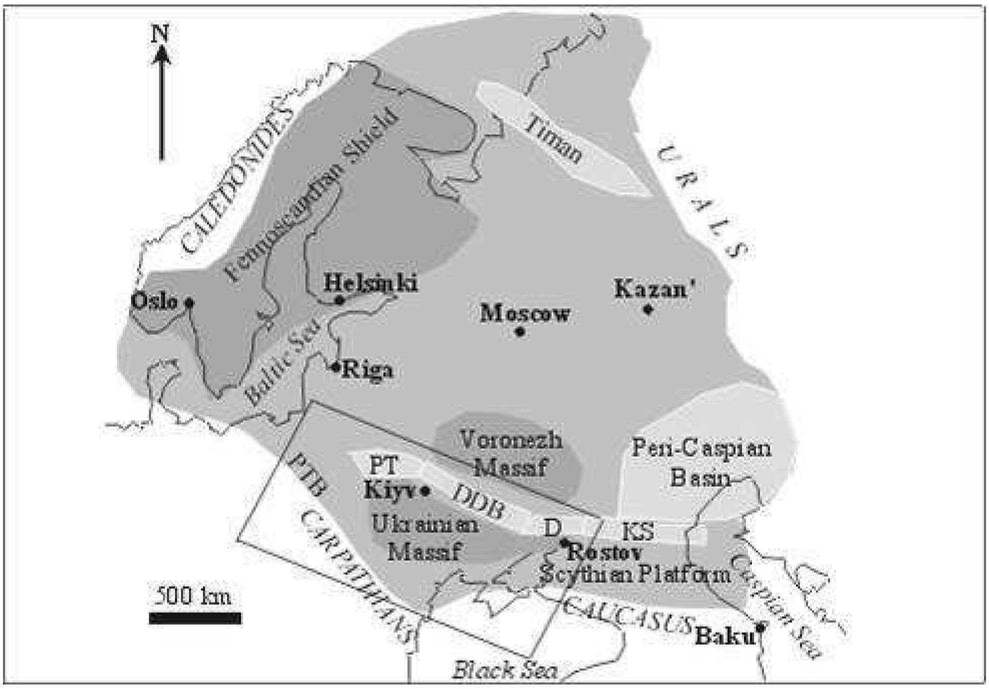

Fig. 1 - The present-day tectonic outline of the East European Craton (significantly modified after NIKISHIN *et al.*, 1996). Abbreviations: PT – Pripyat Trough, DDB – Dniepr-Donets Basin, D – Donbass, KS – Karpinsky Swell, PTB - Peri-Tornquist Basin. Rectangle delineates the southwestern margin of Baltica discussed in this paper.

There was not a unique continent named Baltica until the end of the Proterozoic. The assembly of large terranes in the Palaeoproterozoic did not result in a solid continent as traditionally accepted and described by NIKISHIN *et al*. (1996). Even if an orogeny ended, the boundaries between terranes might have stayed active in the Meso- and Neoproterozoic. BOGDANOVA (1993) and BOGDANOVA *et al*. (1996) argued for the existence of major shear zones with long-distance strike-slip displacements, and rifts between Fennoscandia, Sarmatia, and Volgo-Uralia. Possibly, they were related to previously existing Palaeoproterozoic active structures. In the same way, the Neoproterozoic Pan-African Orogeny (alternatively named the Brasiliano-Pan-African Orogeny) resulted in the amalgamation of many plates and terranes to form the end-Proterozoic-Early Paleozoic Gondwana. This agglutination of continental masses was not an abrupt, but a continuous process, involving multiple and successive collisions (CORDANI *et al*., 2000; HEILBRON & MACHADO, 2003). Major shear zones with strike-slip displacements became the boundaries between amalgamated cratons (CORDANI *et al*., 2000; GUIRAUD *et al*., 2000, 2005; AZZOUNI-SEKKAL *et al*., 2003; CABY & MONIÉ, 2003; KLEIN *et al*., 2005a,b). After the Neoproterozoic, Baltica became an individual ('solid') continent, which generally drifted to the Equator (SCOTESE, 2004). However, ENEROTH & SVENNINGSEN (2004) have argued that Baltica was located close to the Equator by the end of the Proterozoic. Its mid-Paleozoic amalgamation with Laurentia led to the appearance of Laurussia, which in the late Paleozoic was included into the largest supercontinent, namely Pangaea (COCKS & TORSVIK, 2002, 2005; STAMPFLI & BOREL, 2002; SCOTESE, 2004; TORSVIK & COCKS, 2004). After the breakup of Pangaea in the Mesozoic, Baltica, together with Siberia, India, and some large Gondwana-derived terranes, created Eurasia, which has existed until the present day.

Both DADLEZ *et al*. (2005), who analyzed the crustal structure of the Polish Basin, and RUBAN & YOSHIOKA (2005), who discussed the evolution of the Donbass, considered the evolution of the southwestern margin of Baltica in the context of the dynamics along the Paleozoic-Mesozoic major shear zones. The Early Paleozoic counterclockwise rotation of Baltica, documented particularly by TORSVIK *et al*. (1996) and COCKS & TORSVIK (2005), caused dextral strike-slip displacements along the Teisseyre-Tornquist Zone (Fig. 2A). As a result of this movement, a few small terranes were formed, including the Pomeranian Unit and the Kuiavian Unit (DADLEZ *et al*. 2005). It seems that those terranes were not moved far from the Baltic margin, though they were pushed along it. The re-accretion of the detached terranes and the Eastern Avalonian blocks with the southwestern margin of Baltica occurred in the Silurian (DADLEZ *et al*., 2005). Some time later, in the Mississippian or (hypothetically) even earlier in the Middle-Late Devonian, a new and larger break-up of Baltica occurred (RUBAN & YOSHIOKA, 2005). It was caused by activity along the Northern Palaeotethyan Shear Zone. Counterclockwise rotation of Africa stimulated the dextral strike-slip movements along this zone (SWANSON, 1982; RAPALINI & VIZÁN, 1993) and caused the detachment of the Ukrainian Massif from Baltica (RUBAN & YOSHIOKA, 2005). The massif was transported off the margin of Baltica (KOSTJUTCHENKO *et al*., 2004). Rapidly subsiding basins such as the Donbass, the Dniepr-Donets Basin, and the Pirpyat Trough, were opened during the late Paleozoic due to extension between Baltica and the Ukrainian Massif, which became an individual terrane (Fig. 2B). This was somewhat similar to the opening of the Red Sea, which began in the Oligocene and intensified in the Miocene (SHARLAND *et al*., 2001; TAWADROS, 2001; BOSWORTH *et al*., 2005; GUIRAUD *et al*., 2005). Due to this breakup, the Arabian Plate became an individual tectonic block. The stress fields of Arabia and Africa are decoupled and during the next 10 My about 170 km of new crust will form in the Red Sea (BOSWORTH *et al*., 2005).

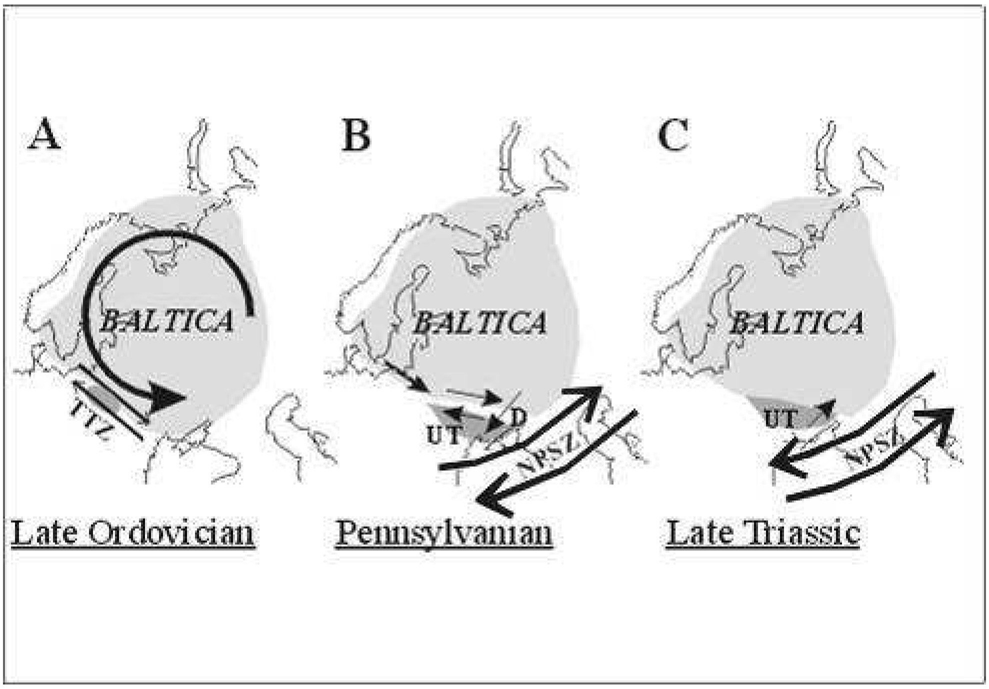

Fig. 2 - The southwestern margin of Baltica in the Late Ordovician (after DADLEZ *et al.*, 2005) (A), the Pennsylvanian (B), and Late Triassic (C). Abbreviations: TTZ –Teisseyre-Tornquist Zone, NPSZ –Northern Palaeotethyan Shear Zone, UT –Ukrainian Terrane, D –Donbass.

The extension hypothesized between the Ukrainian Terrane and Baltica probably took place with a significant oblique component. BELOUS & KOROLEV (1973), ARTHAUD & MATTE (1977), and later DE BOORDER *et al*. (1996) and SAINTOT *et al*. (2003) have shown some dextral displacements along the above-mentioned basins. The Variscan collision along the Teisseyre-Tornquist Zone also occurred with a dextral oblique component (DADLEZ *et al*., 2005). In the Late Triassic, the displacements along the Northern Palaeotethyan Shear Zone changed from dextral to sinistral because of the clockwise rotation of Africa (SWANSON, 1982; RAPALINI & VIZÁN, 1993; RUBAN & YOSHIOKA, 2005). Consequently, the Ukrainian Terrane moved back to the margin of Baltica (Fig. 2C). Compression affected the basin located between them, i.e, the Donbass, as folding and thrusting demonstrate. This Late Triassic phase of contraction was documented by SAINTOT *et al*. (2003).

Thus, during the Paleozoic and at least the early Mesozoic, the palaeotectonic setting of the southwestern margin of Baltica was governed by the activity of the major shear zones. We only have indirect evidences for their continued activity in the Jurassic and the Cretaceous. For example, in the Donbass, Jurassic dykes occur on its southern flank near the Boldyrevka Village in the Rostov Region. Additionally, the Alpine phase of the Donbass deformation appeared in the Late Cretaceous-Tertiary (SAINTOT *et al*., 2003). In the Atlasic domain of Africa, along what was the western branch of the pre-existing Northern Palaeotethyan Shear Zone, shearing continued at least until the Late Jurassic (Löwner *et al*., 2002). Vai (2003) noted that strike-slip displacements might have occurred between the Alps and the Hymalayas in the Jurassic and the same was hypothesized by Ruban (2007) for the Caucasus.

Global context

The Northern Palaeotethyan Shear Zone–was not the only major shear zone in the late Paleozoic. The five shear zones considered most significant are described briefly below and located in Fig. 3.

(1) The Northern Palaeotethyan Shear Zone with right-lateral strike-slip displacements stretched from the Atlasic Domain of Morocco in the west to the Caspian region in the east (Arthaud & Matte, 1977; Stampfli & Borel, 2002; Ruban & Yoshioka, 2005), although Vai (2003) places it closer to the African-Arabian margin of Gondwana than shown in Fig. 3.

(2) The Teisseyre-Tornquist Zone was characterized by dextral shear in Variscan times (Dadlez *et al*., 2005). The model of Arthaud & Matte (1977) relates the activity along some other minor shear zones in Western and Central Europe to the Northern Palaeotethyan Shear Zone.

(3) The Appalachian Shear Zone stretched along the Appalachian Orogen, and it seems to be the westward extension of the Northern Palaeotethyan Shear Zone with

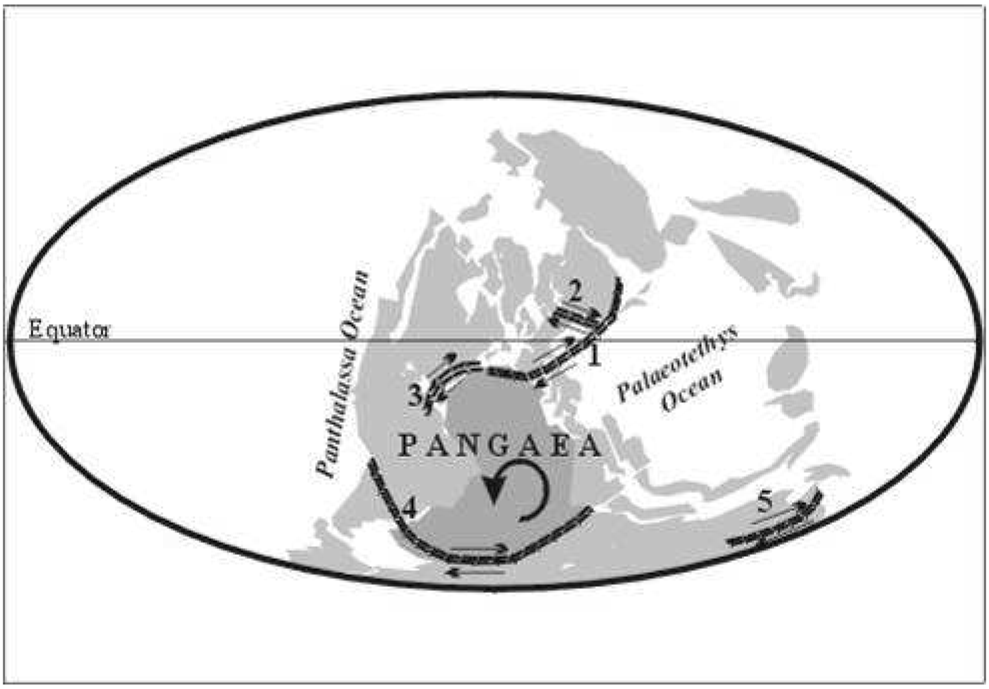

Fig. 3 - Global system of megashears in the late Paleozoic (palaeotectonic base map is simplified from Scotese, 2004). See text for explanation of the numbers of the major shear zones.

the same direction or sense of displacements (ARTHAUD & MATTE, 1977; SWANSON, 1982, 1999A,B, 2006; RAPALINI & VIZÁN, 1993; LAWVER *et al.*, 2002; VAI, 2003; HIBBARD, 2004; HIBBARD *et al.*, 2002).

(4) The Intrapangaean Shear Zone was located within the Pangaea Supercontinent (RAPALINI & VIZÁN, 1993). Megashears are traceable in southern South America (particularly, in the Argentinian basins), and in South and East Africa.

(5) VEEVERS *et al.* (1994) indicated that significant dextral displacements occurred along the East Australian margin of Pangaea. It is unclear whether there was a separate shear zone, or a branch of the Intrapangaean Shear Zone as an alternative to the branch along East Africa proposed by RAPALINI & VIZÁN (1993).

It seems that a global system of major shear zones existed in the late Paleozoic. Its development and activity may be explained in terms of global wrench tectonics (STORETVEDT, 2003), or at least by the rotation of the individual continents amalgamated into Pangaea. The beginnig of the strike-slip activity in the Appalachians and in the Atlasic Domain of Morocco in the Devonian (ARTHAUD & MATTE, 1977; SWANSON, 1999a,b, 2006; HIBBARD *et al*., 2002) corresponded well to the hypothesized Late Devonian onset of the Donbass structures (NIKISHIN *et al*., 1996). If these regions were affected by a unique global system of major shear zones, such a correspondence validates an above-mentioned hypothesis that the Donbass structures were originated with the strike-slip displacements.

Palaeotectonic analogue

A search for palaeotectonic analogues contributes to the understanding of the geological evolution of some poorly-known regions with the information from those better-known. Additionally, it produces a general knowledge on the fundamental mechanisms and regional peculiarities of the tectonic processes. Important examples of such studies were presented by MCCALL (2001), TAIRA (2001), CABY (2003), and RUBAN & YOSHIOKA (2005).

A well studied palaeotectonic analogue of the strike-slip continental margin is found in the west and southwest of North America (Fig. 4). Strike-slip movements, first sinistral and later dextral, occurred in the Sierra Nevada and Cordillera during the Late Jurassic-Early Cretaceous (WYLD & WRIGHT, 2001; DECELLES, 2004). At about 85 Ma, a ribbon of terranes, located along the margin of North, Central and South America, was translating northward (NOKLEBERG *et al*., 2000; JOHNSTON, 2001; SCOTESE *et al*., 2001; LAWVER *et al*., 2002) (Fig. 3A), resulting in the so-called oroclinal orogeny (JOHNSTON, 2001). It is important to note, that the oroclinal orogeny is also interpreted in Eastern Australia, where it was relatred to the activity of the late Paleozoic shear zones (VEEVERS *et al*., 1994). Palaeobiogeographic interpretations of ABERHAN (1998, 1999) allowed him to conclude that Wrangellia and Stikinia, two Canadian terranes, moved northward by 1,000 km or more since the middle of the Early Jurassic. Thus, the major displacement occurred along the western margin of North America in the Mesozoic, and it lasted until the Present (NOKLEBERG *et al*., 2000; SCOTESE *et al*., 2001; LAWVER *et al*., 2002). ANDERSON & SILVER (1979) developed the concept of the Mojave-Sonora megashear with sinistral displacements, which occurred on

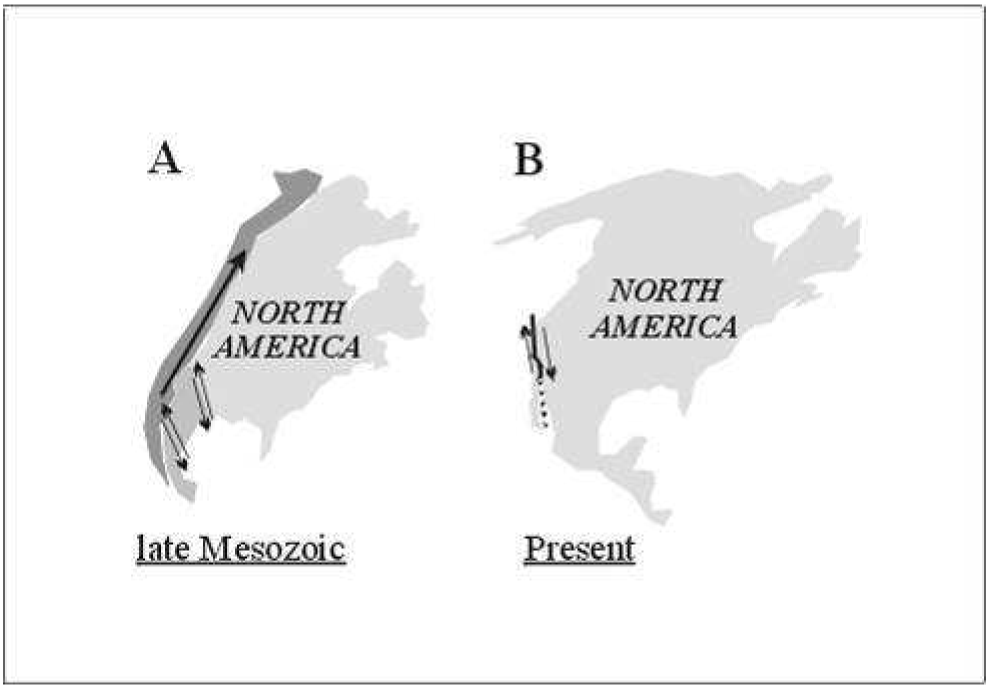

Fig. 4 - The western and southwestern margins of North America in the late Mesozoic (A) and the Present (B).

the southwestern margin of North America in the Late Jurassic. Recently, ANDERSON *et al.* (2005) proposed several subparallel megashears with sinistral strike-slip motions, which crossed the territory of present-day Mexico from southeast to northwest (Fig. 4A). At least some of the faults associated with those megashears might have been active even in the Cenozoic (MCKEE *et al.*, 1984). However, POOLE *et al.* (2005) found no evidence to support the concept of the Mojave-Sonora megashear. From the late Miocene to the Present, dextral displacements (up to 300-500 km) have occurred at the southwestern margin of North America along the San Andreas Fault and associated fault systems and aided in the evolution of the Gulf of California (ATWATER & STOCK, 1998; LEWIS & STOCK, 1998; NAGY, 2000; ARGUS & GORDON, 2001; OSKIN *et al.*, 2001; TSUTSUMI *et al.*, 2001; OSKIN & STOCK, 2003) (Fig. 4B).

The tectonic activity on the western and southwestern margins of North America was characterized by the presence and sometimes dominance of an oblique component during the Mesozoic and Cenozoic. Displacements of small and medium-sized terranes along shear zones were common in the geologic history of this region. Therefore, the western and southwestern margins of North America in the Mesozoic-Cenozoic are proposed as a palaeotectonic analogue of the Paleozoic-early Mesozoic strike-slip southwestern margin of Baltica. Although the forces-causing the activity along the major shear zones in these regions might have been somewhat different, their further comparative study may contribute significantly to the understanding of the geodynamic evolution of these regions.

Conclusions

This brief synthesis allows two conclusions that may aid in the geodynamic reconstruction of Baltica:

1) in the late Paleozoic, the shear zones on the southwestern margin of Baltica were included in the global system of megashears (Fig. 3);

2) a palaeotectonic analogue of the Paleozoic-early Mesozoic southwestern Baltic margin is found in North America - during the Mesozoic-Cenozoic, active megashears played a very important role in the evolution of the western and southwestern margins of that continent (Fig. 4).

In general, this paper should be considered as an invitation to a broad discussion, because many of palaeotectonic ideas presented herein are not finalized and, therefore, need to be verified, modified, or completed.

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Review of Middle East Paleozoic plate tectonics

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ABSTRACT

The Paleozoic Middle East terranes, neighboring the present-day Arabian and Levant plates, are shown by most authors to consist of ten major tectonic units: (1 and 2) the Helmand and Farah terranes of Afghanistan, southwest Pakistan and southeast Turkmenistan; (3 to 6) the Alborz, Central Iran (Lut, Yazd and Tabas) and Sanandaj-Sirjan terranes of Iran, and Northwest Iran (possibly extending into eastern Turkey); (7 and 8) the Pontides and Taurides terranes of Turkey; and (9 and 10) the Greater and Lesser Caucasus terranes between the Caspian and Black seas (Armenia, Azerbaijan, Georgia and southwest Russia). Published plate-tectonic reconstructions indicate that all ten terranes may have broken off from either: (1) the Gondwana Supercontinent in the mid-Silurian as part of the Hun Superterrane; or (2) the Pangea Supercontinent during the mid-Permian - Triassic as part of the Cimmeria Superterrane. To the north of Gondwana and Pangea, three successively younger Tethyan oceans evolved: (1) Proto-Tethys (Cambrian - Devonian); (2) Paleo-Tethys (mid-Silurian - Mesozoic); and (3) Neo-Tethys (mid-Permian - Cenozoic).

Two regional Paleozoic unconformities in the Arabian Plate are generally linked to major regional-scale structural events, and commonly correlated to the Caledonian and Hercynian orogenies. These orogenies took place many thousands of kilometers away from the Arabian Plate and are considered unlikely causes for these unconformities. Instead, the breakaway of the Hun and Cimmeria superterranes are considered as alternative near-field tectonic sources. The older unconformity (middle Paleozoic event), represented by a mid-Silurian to Middle Devonian hiatus in North Arabia (Iraq and Syria), reflects an episode of epeirogenic uplift, which might be related to the mid-Silurian rift of the Hun Superterrane. The younger mid-Carboniferous Arabia-wide angular unconformity involved compressional faulting and epeirogenic uplift, and might be related to the earliest phase of subduction by the Paleo-Tethyan crust beneath Cimmeria (Sanandaj-Sirjan and nearby regions) before it broke off. Based on our review and regional considerations, we assign the Helmand, Farah, Central Iran, Alborz, Sanandaj-Sirjan, Northwest Iran, Lesser Caucasus, Taurides and Pontides to Cimmeria, whereas the Greater Caucasus is considered Hunic.

INTRODUCTION

During the past decade, our general knowledge on the geochronological evolution, paleopositions, paleotectonic compositions and paleogeographic outlines of the Phanerozoic supercontinents has significantly improved (e.g. Dalziel, 1997; Stampfli et al., 2001, 2002; Lindsay, 2002; Cocks and Torsvik, 2002; Lawver et al., 2002; Stampfli and Borel, 2002; Veevers, 2003; Pesonen et al., 2003; von Raumer et al., 2002, 2003; Golonka, 2004; Scotese, 2004; Torsvik and Cocks, 2004). Yet today, many uncertainties persist in reconstructing the geological evolution of the regions adjoining the supercontinents, especially for the Paleozoic Era (Cocks and Torsvik, 2002; Torsvik and Cocks, 2004; Scotese, 2004). These regions are in themselves vast, and consist of numerous plate-tectonic units that are sometimes referred to as microplates, terranes, blocks, structural domains, and sometimes grouped into superterranes. The Middle East is a typical example of a border-region that consists of a complex mosaic of tectonic units (Figures 1 and 2).

We identify the Middle East terranes, bordering the present-day Arabian and Levant plates, in Afghanistan, Iran, western Pakistan, Turkey, southeast Turkmenistan and the Caucasus (Armenia, Azerbaijan, Georgia and southwest Russia) (Figures 1 and 2). Several of these terranes are strongly deformed and stacked within a wide tectonic belt between the Eurasian, Arabian and Indian plates.

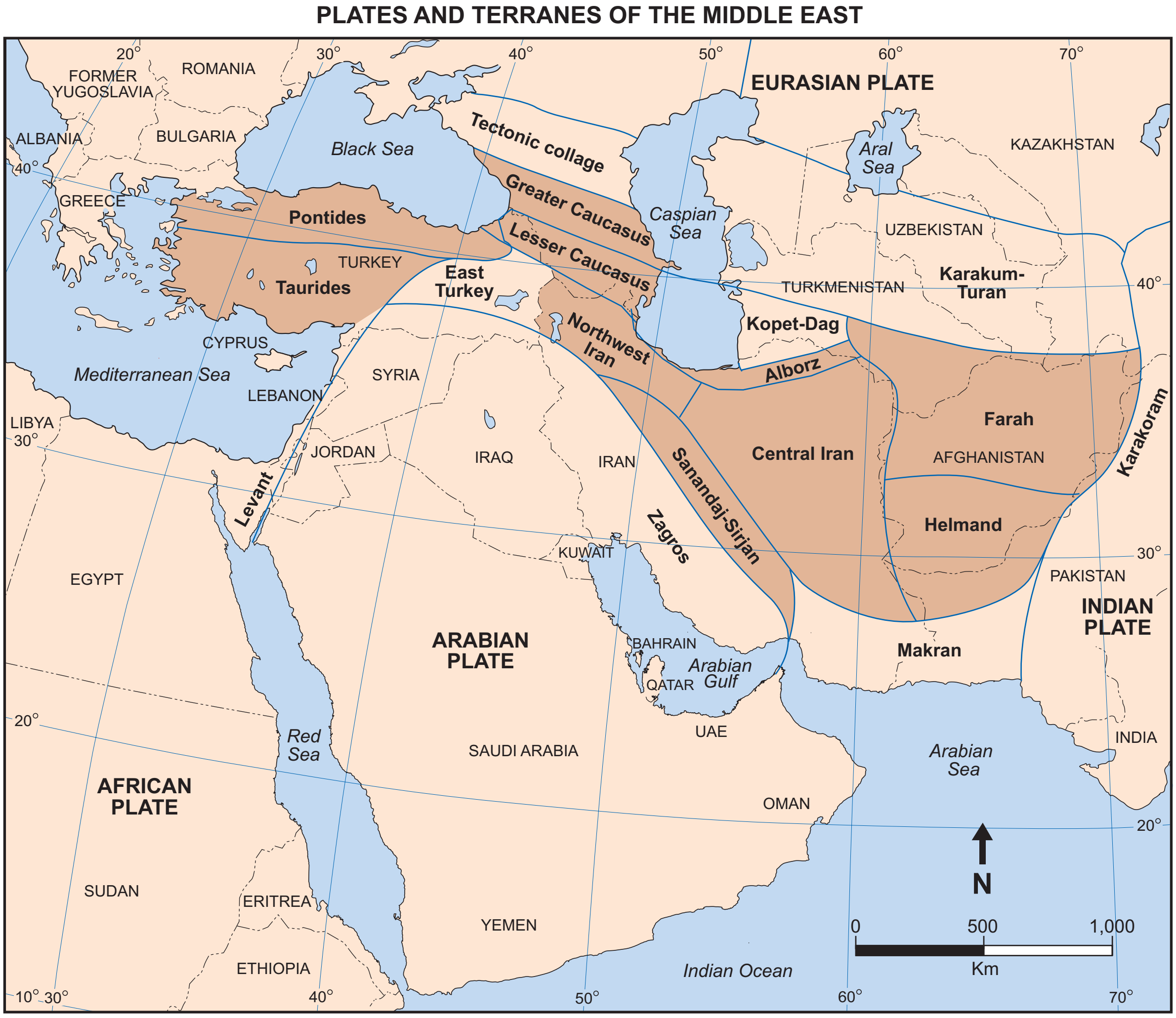

Figure 1: The Middle East region consists of the present-day Arabian and Levant plates and numerous terranes (individual boundaries are shown in blue). During the Paleozoic ten large terranes are variably interpreted to have been adjacent to the Arabian and Levant plates (then attached to Gondwana and later Pangea). The Paleozoic Middle East terranes (colored brown) include Helmand and Farah (Afghanistan, southwest Pakistan and southeast Turkmenistan); Iran's Alborz, Northwest Iran, Sanandaj-Sirjan and Central Iran; Turkey's Pontides and Taurides; and the Greater and Lesser Caucasus between the Caspian and Black seas (Armenia, Azerbaijan, Georgia and southwest Russia). The Makran and East Turkey regions may have a Paleozoic core or could have formed as Mesozoic accretionary terranes.

Although their boundaries are generally traced along well-preserved or/and reactivated Paleozoic fault systems, in some cases the borders remain unclear. Correlation of the sedimentary core complexes, however, suggests that all of these terranes share a common ancestry during some time in the Paleozoic Era.

The Middle East terranes were affected by the evolution of the Paleozoic Tethyan oceans, the Hun (Hunic or Intermediate) and Cimmeria (Cimmerian) superterranes, and the Gondwana and Pangea supercontinents (Figures 3 to 11; e.g. Sengör, 1990; Stampfli, 1996; von Raumer, 1998; Cocks and Torsvik, 2002; von Raumer et al., 2002, 2003; Stampfli and Borel, 2002; Stampfli et al., 2001, 2002; Torsvik and Cocks, 2004; Natal'in and Sengör, 2005; Xypolias et al., 2006). At least three major Paleozoic rift episodes occurred along the margins of Gondwana and Pangea (Figures 4 to 11). The first was in the Early Ordovician when Avalonia broke off from Gondwana. This episode was unlikely to have influenced the Middle East region, which was located about 6,000 km away (Figure 4). The second involved the mid-Silurian breakaway of the Hun Superterrane (Figures 6 and 7), the detailed reconstruction of

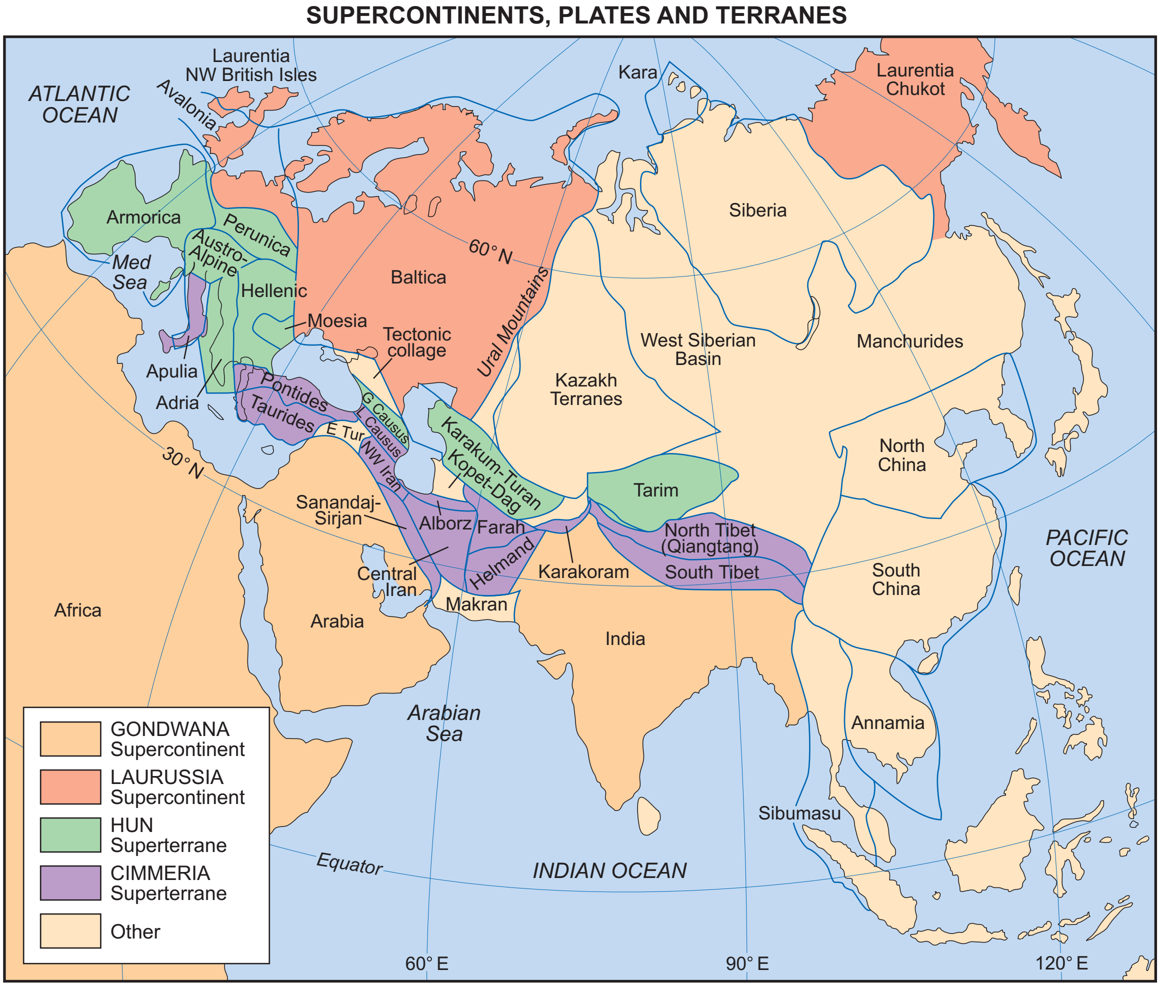

Figure 2: The majority of the plates and terranes discussed in this review are shown after Torsvik and Cocks (2004). Besides the NW British Isles and the Chukot Peninsula in Russia, the Laurentia Supercontinent included North America and Greenland (see Figures 4 and 5). Laurentia later collided with Baltica and Avalonia to form the Laurussia Supercontinent (Figure 7). Besides Arabia, Africa and India, the Gondwana Supercontinent included Antarctica, Australia and Madagascar (see Figure 4). Also shown are terranes that may have been part of the Hun and Cimmeria Superterranes. Note that Karakoram in north Pakistan is Cimmerian and different from the Hunic Karakum-Turan terrane.

which is unresolved due to insufficient paleomagnetic and paleontological data. This episode is relevant to our review because parts of the superterrane may have involved the Middle East terranes. The third episode is the mid-Permian - Triassic breakaway of several Middle East Cimmerian terranes from Gondwana, by then a part of Pangea (Figures 10 and 11).

In most Paleozoic reconstructions, the Middle East region is interpreted as part of the passive margin of Gondwana and Pangea until the mid-Permian - Triassic, when Cimmeria started to rift away, causing the opening of the Neo-Tethys Ocean (e.g. Sharland et al., 2001; Stampfli et al., 2001). In addition, two regional unconformities are recognized. The first corresponds to a mid-Silurian to Middle Devonian hiatus – "middle Paleozoic hiatus" – that is sometimes correlated with the Caledonian Orogeny (e.g. Buday, 1980) (Figures 3, 5 to 7). The second unconformity represents a "mid-Carboniferous hiatus", and is often correlated to the Hercynian Orogeny (e.g. Berberian and King, 1981) (Figures 3 and 9). These correlations do not provide a satisfactory plate-tectonic model for the Paleozoic evolution of the Middle East region because it was located far away from these two orogenies.

Our paper starts with a brief global review of the largest and relatively well-constrained Paleozoic

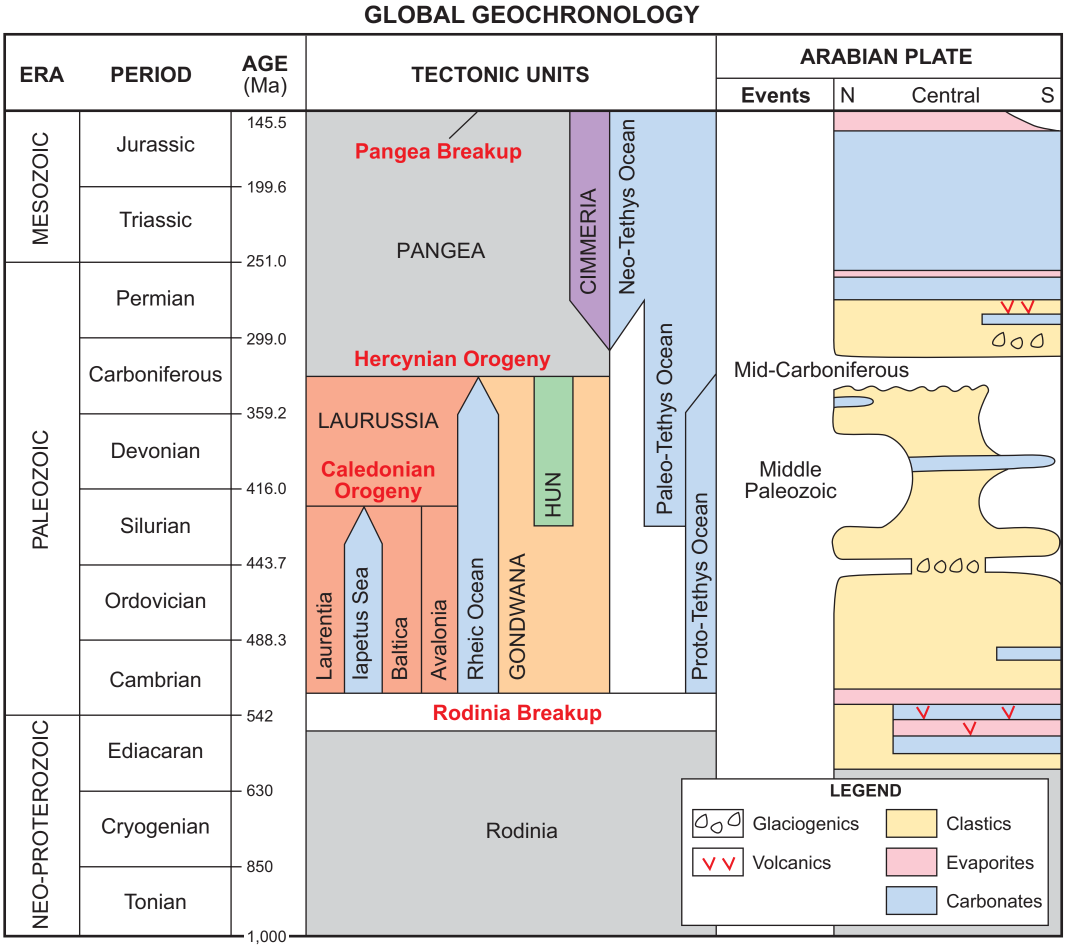

Figure 3: Simplified geochronology of the supercontinents, superterranes, oceans and seas, and generalized Arabian Plate stratigraphic column. The breakup of the Neo-Proterozoic Rodinia Supercontinent is not discussed in this paper. The geological time scale is after the ICS (International Commission on Stratigraphy; Gradstein et al., 2004). Two Arabian unconformities are associated with Gondwana glaciations: latest Ordovician-early Silurian (Figure 5) and late Carboniferous-early Permian (Figure 9). In Arabia the second glaciation followed a mid-Carboniferous structural event that coincided in time with the Hercynian Orogeny (Figure 9). Another North Arabian (and Oman) unconformity is associated with epeirogenic uplifting in the mid-Silurian (Wenlock) to Middle Devonian (middle Paleozoic event) and has been correlated to the Caledonian Orogeny (Figures 5 to 7). The Hercynian and Caledonian collisions were located many 1,000s of kilometers away from Arabia and could not therefore have transmitted significant lateral forces to its crust. Instead two near-field events may have played a more direct role: the breakaway of the Hun Superterrane in mid-Silurian (and possibly Devonian) (Figure 6), and the early compressional evolution of a subduction complex preceeding the breakaway of the Cimmerian Superterrane (Figure 11). Color scheme for tectonic units follows Figure 2.

plate-tectonic units and seaways. We adopt the conventions of the ICS (International Commission on Stratigraphy; Gradstein et al., 2004) by not capitalizing informal qualifiers “late, middle, mid, early, etc.” except where defined (Ordovician and Devonian). After setting the global stage, we discuss the smaller and less constrained units of the Middle East, many of which have unfamiliar names and interpretations. Most of the illustrated global reconstructions follow Cocks and Torsvik (2002) and Torsvik and Cocks (2004), as the involved units are adequately represented. We have also considered the global reconstructions by von Raumer (1998), Stampfli and Borel (2002), von Raumer et al. (2002, 2003), Stampfli et al. (2001, 2002), Scotese (2004), Natal’in and Sengör (2005) and others. Our principal objective is to present the Paleozoic plate-tectonic framework and nomenclature for the Middle East, which can form a basis for further refinements.

GLOBAL PALEOZOIC PLATE-TECTONIC SETTING

Gondwana, Laurussia and Pangea Supercontinents

The global plate-tectonic configuration during the Paleozoic was dominated by three supercontinents: Gondwana, Laurussia and Pangea. Gondwana, the largest supercontinent on Earth from the Late Cambrian to mid-Carboniferous (Figures 2 to 8), consisted of several present-day plates including Arabia, Africa, most of Antarctica and Australia, India, Madagascar and most of South America, with numerous small terranes attached to their margins (Courjault-Radé et al., 1992; Cocks, 2001; Stampfli et al., 2001; Cocks and Torsvik, 2002; Stampfli and Borel, 2002; von Raumer et al., 2002, 2003; Fortey and Cocks, 2003; Scotese, 2004; Avigad et al., 2005).

During the Paleozoic Era, Laurussia was assembled out of three large plates (Avalonia, Baltica and Laurentia, Figures 3 and 5) and several island arcs in a series of orogenic phases (McKerrow et al., 2000; Stampfli and Borel, 2002; Scotese, 2004) (Figures 3 to 7). The collisional assembly of Laurussia caused the Caledonian Orogeny, an event that was redefined by McKerrow et al. (2000) to apply to the closure of the Iapetus Sea (Figures 3, 5 to 7).

- Avalonia consisted of easternmost North America and parts of northwest Europe (Figures 2 and 3; also Avalonian-Cadomian Arc and Orogenic Belt of Linnemann et al., 2000, and Linnemann and Romer, 2002). Avalonia rifted from western Africa (in Gondwana) in the Early Ordovician (Cocks and Torsvik, 2002), and then drifted northwards until it collided with Baltica and Laurentia (Figures 4 to 7).

- Baltica consisted of the Russian Platform and extended to east of the Ural Mountains (Figures 2 and 3, Cocks and Torsvik, 2002).

- Laurentia consisted of most of North America, Chukot Peninsula of eastern Siberia, Greenland, Spitsbergen and the NW British isles (Figures 3 to 5; Cocks and Torsvik, 2002).

Gondwana and Laurussia remained separated by the Rheic Ocean until the mid-Carboniferous (c. 325–310 Ma) when they collided during the Hercynian Orogeny to form the Pangea Supercontinent (Stampfli and Borel, 2002; Scotese, 2004; Figures 3, 8 and 9, Torsvik and Cocks, 2004). In the late Carboniferous (Pennsylvanian) and Permian, Pangea was enlarged with the amalgamation of the Kazakh, Siberia, Kara and other terranes (Figure 10). The combination of Laurussia with these terranes would later in the Mesozoic form the Laurasia Supercontinent, a term that is easily confused with Laurussia.

Three Tethyan Oceans

The names of the Paleozoic oceans that separated the supercontinents are not unique and vary to reflect somewhat different interpretations. The seaway that opened to the north of the Paleozoic Middle East terranes is called the Paleo-Tethys Ocean by some authors (e.g. Sharland et al., 2001; Bykadorov et al., 2003). Others refer to it as the Proto-Tethys (or Asiatic Ocean) and reserve the term "Paleo-Tethys" for the ocean that opened in the mid-Silurian along the trailing edge of the Hun Superterrane (Figures 3, 7 and 8; e.g. Ziegler et al., 2001; Stampfli et al., 2001; Stampfli and Borel, 2002; von Raumer et al., 2002, 2003). Most authors adopt the Neo-Tethys for the ocean that opened in mid-Permian - Triassic along the African-Arabian margin of Gondwana (e.g. Stampfli and Borel 2002; also Meso-Tethys of Metcalfe, 1999; Pindos Ocean of Golonka, 2004; Figure 10). Following Stampfli and Borel (2002), we adopt Proto-Tethys for the early Paleozoic ocean to distinguish it from the late Paleozoic Paleo-Tethys and Neo-Tethys oceans (Figures 3–10).

The interpretation of the lateral extent and evolution of the Tethyan oceans can vary. Hünecke (2006), for example, argued that the Middle-Late Devonian ocean between Gondwana and Laurussia was not as large as depicted by Stampfli and Borel (2002) and Torsvik and Cocks (2004). Stampfli and Borel (2002) and von Raumer et al. (2002, 2003) interpreted that in the Late Ordovician-early Silurian, the eastern branch of the Proto-Tethys Ocean might have closed when Serindia terranes (North China and

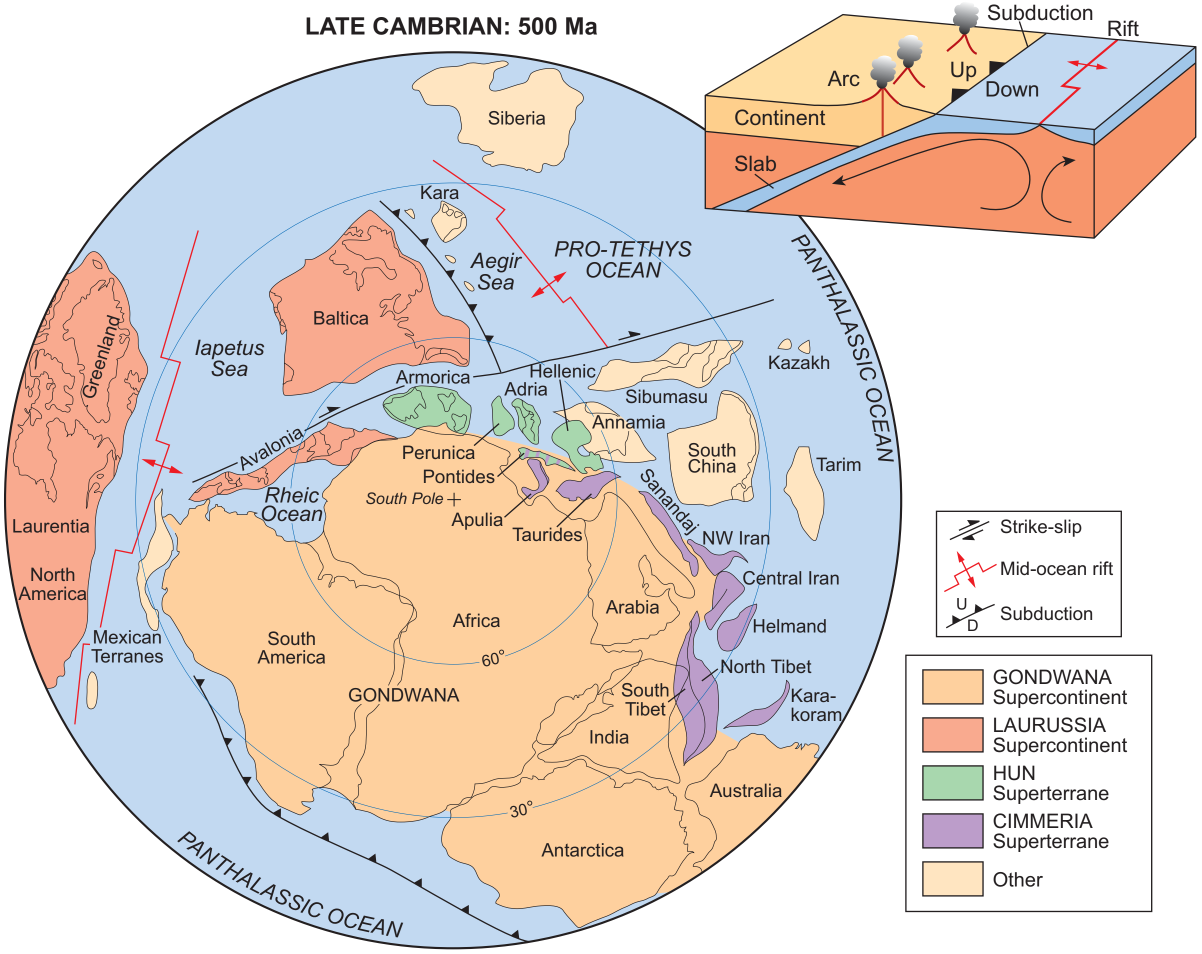

Figure 4: Plate-tectonic reconstruction of the Late Cambrian times (modified after Cocks and Torsvik, 2002). The terranes included in the Hun Superterrane (green) are mostly after Stampfli et al. (2001, see Figure 6). Torsvik and Cocks (2004) show several terranes breaking away from Gondwana in the middle Paleozoic but do not appear to recognize the Hun Superterrane (Figure 7). The Adria terrane (Figures 6 to 9, after Torsvik and Cocks, 2004) was not shown in Cocks and Torsvik (2002) and is located arbitrarily in Figures 4 and 5. Middle East terranes (Figures 1 and 2) not shown are Alborz, Farah and Caucasus. Pontides shown as Hunic by Cocks and Torsvik (2002), but considered Cimmerian in our paper (in stripes). Schematic diagram at top right illustrates oceanic slab subducting below (down or D) the continent (up or U).

Tarim) amalgamated with Gondwana. Several authors interpreted the initial opening of the Neo-Tethys Ocean in early rather than mid-Permian (Vannay, 1993; Garzanti and Sciunnach, 1997; Garzanti et al., 1994, 1996a, b, 1999; Stampfli and Borel, 2002; Angiolini et al., 2003), or to have started north of Australia in the Carboniferous and extended diachronously westwards into the Permian (Stampfli, 2000).

Iapetus Sea and Rheic Ocean

Two more Paleozoic seaways are significant for our review (Figures 3 to 7). The Iapetus Sea, which separated Laurentia, Avalonia and Baltica in the early Paleozoic, closed in the late Silurian when these terranes joined to form Laurussia (Figures 3 to 5). The Rheic Ocean (also Rheic-Mauritania, Rhenohercynian or Hercynian-Rheic) opened in the Cambrian along Avalonia's northerly trailing edge (Figures 3 to 7). During the Devonian-early Carboniferous (c. 420–320 Ma), Gondwana drifted towards Laurussia, closing the Rheic Ocean (Figures 7 and 8, Torsvik and Cocks, 2004). In the mid-Carboniferous, the Hercynian Orogeny occurred along a front between northwest Africa and southeast North America and closed the Rheic Ocean (Figure 9; incorrectly referred to as the Iapetus Sea in Al-Husseini, 2004).

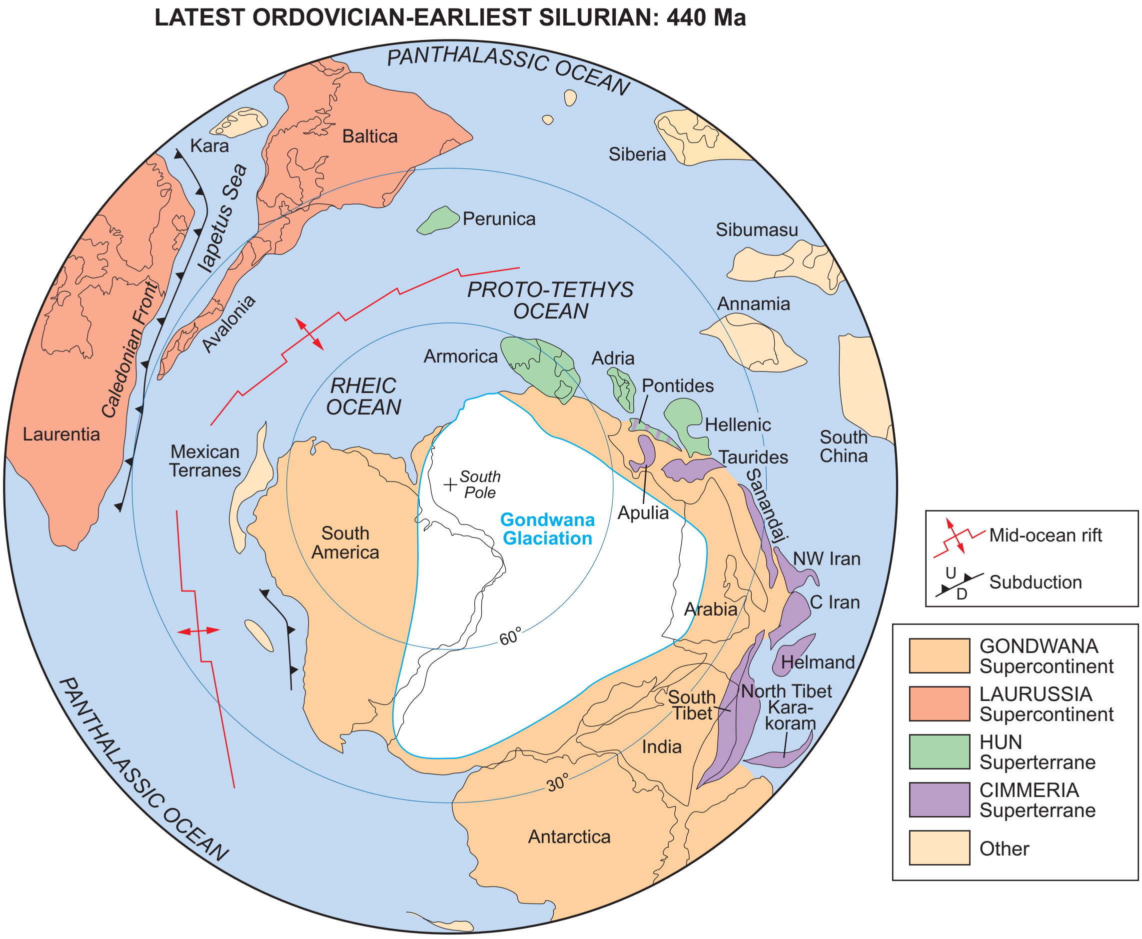

Figure 5: Plate-tectonic reconstruction of the latest Ordovician-earliest Silurian times (modified after Cocks and Torsvik, 2002). During the final Hirnantian Stage of the Late Ordovician, polar glaciers advanced over regions of Gondwana reaching western Saudi Arabia (Vaslet, 1990). Baltica and Avalonia were joined, closing the Iapetus Sea where the Caledonian Orogenic Front was located. Middle East terranes (Figures 1 and 2) not shown are Alborz, Farah and Caucasus. Adria terrane is located arbitrarily (see Figure 4 caption). Pontides shown as Hunic by Torsvik and Cocks (2004), but considered Cimmerian in our paper (in stripes).

Hun Superterrane

Plate reconstruction of the mid-Silurian to mid-Permian northern margin of Gondwana fall into two general models (Figures 6 and 7). Whereas both show the breakaway of various terranes from Gondwana, they differ in detail and lateral extent – particularly near the Middle East region.

In the first model, following Stampfli et al. (2001) and Stampfli and Borel (2002), the ribbon-like Hun Superterrane extended from westernmost Iberia (in Spain) to Qiantang (Figure 6). This superterrane is also referred to as the Hun Composite Superterrane because it is divided into (Figure 6): (1) the northern Hun Cordillera terranes (also European Hunic terranes); and (2) the southern Hun Gondwana terranes (also Asiatic Hunic terranes) (von Raumer, 1998; Stampfli et al., 2001; von Raumer et al., 2002, 2003; Stampfli and Borel, 2002; Schulz et al., 2004). This division reflects the separate evolution of the two sets of terranes after they docked along Laurussia in the late Paleozoic. The Hun Superterrane rifted away from Gondwana in the mid-Silurian (possibly in different phases that lasted into the Devonian) and then drifted towards Laurussia, with which it collided in the Devonian-Carboniferous (Figures 3, 6 to 8). Interpretations of late Silurian paleocurrents indicate that the Panthalassic (north of the Proto-Tethys) waters did not mix with those of the Paleo-Tethys (Johnson et al., 2004), thus suggesting that the Hun Superterrane remained consolidated until at least the Early Devonian.

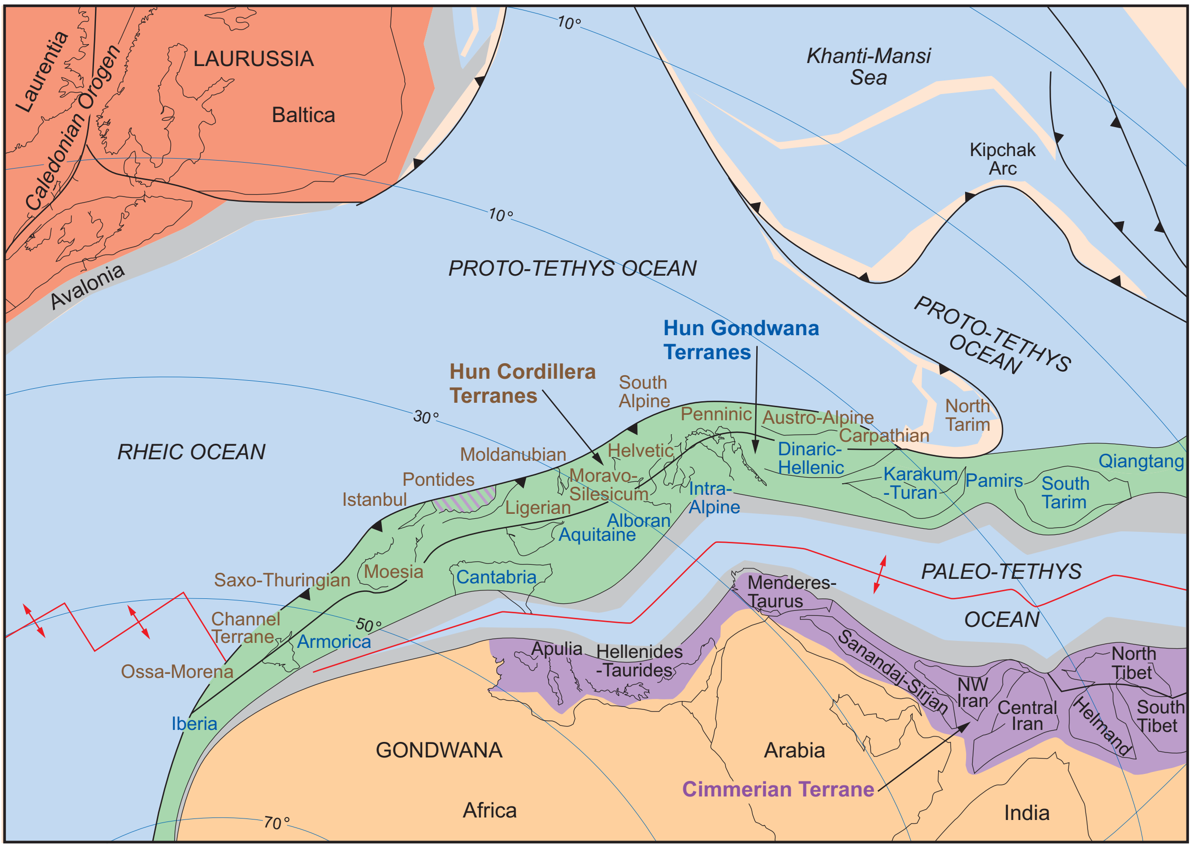

Figure 6: Plate-tectonic reconstruction of the late Silurian (Ludlow) by Stampfli et al. (2001) shows the Hun Superterrane rifting away from the Gondwana Supercontinent. The Paleo-Tethys Ocean formed between Gondwana and the superterrane. The superterrane is divided into the southern Gondwana and northern Cordillera terranes. Stampfli et al. positioned the Istanbul and Pontides terranes (North Turkey) near a subduction zone at the northern boundary of the Hun Cordillera terrane. In contrast, Sengör (1990; see Figure 11) placed the Pontides (in stripes) in the Cimmeria Superterrane and associated it with a middle Carboniferous-Mesozoic subduction-arc that continued into the Sanandaj-Sirjan terrane. Middle East terranes (Figures 1 and 2) not shown are Cimmerian Alborz, Farah and Lesser Caucasus. The Hunic Greater Caucasus may have been located further east beyond Qiangtang.

From west to east the Hun Cordillera terranes included: Ossa-Morena, Channel, Saxo-Thuringian, Moesia, Istanbul, Pontides, Ligerian, Moldanubian, Moravo-Silesicum, Helvetic, South Alpine, Penninic, Austro-Alpine, Carpathian and North Tarim (Figure 6; Stampfli et al., 2001). The Hun Gondwana terranes included: Iberia, Armorica, Cantabria, Aquitaine, Alboran, Intra-Alpine terranes (Adria, Carnic, Austro-Carpathian), Dinaric-Hellenic, Karakum-Turan, Pamirs, South Tarim, Qiangtang, North and South China, and Annamia terranes (Stampfli et al., 2001; Stampfli and Borel, 2002; the latter two easternmost terranes are shown in Figure 1, but not in Figure 6). In the Devonian, the Kazakh terranes may also have been close to the Hunic Cordillera Superterranes (Stampfli and Borel, 2002). The remaining adjoined with Pangea were Apulia, Hellenides-Taurides, Menderes-Taurus, Sanandaj-Sirjan, Northwest and Central Iran, Helmand, North and South Tibet.

In the second model, several terranes broke off and had drifted some distance away from the northwestern margin of Gondwana by the Early Devonian (Figure 7; Torsvik and Cocks, 2004). They also formed a ribbon-like superterrane that vaguely resembles the western part of the Hun Superterrane (compare Figures 6 and 7). The breakaway group included: Rheno-Hercynian, Armorica (includes Iberia), Adria, Pontides, Hellenic and Moesia (Figures 2 and 6). The Rheno-Hercynian and Perunica

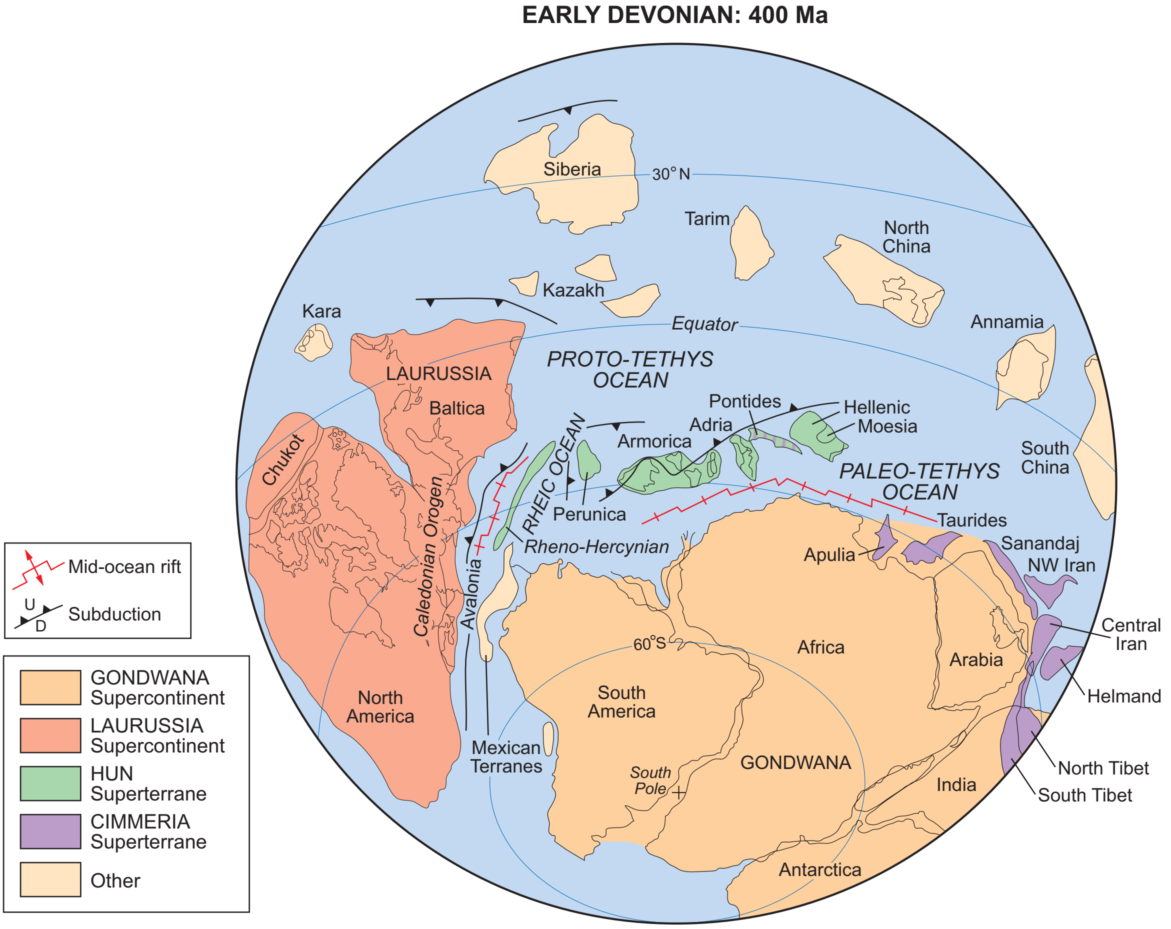

Figure 7: Plate-tectonic reconstruction of the Early Devonian (modified after Torsvik and Cocks, 2004) shows the breakaway from Gondwana of several terranes that together resemble the western Hun Superterrane (Stampfli et al., 2001; Figure 6). Both reconstructions show the Pontides (in stripes) as Hunic whereas we favor a Cimmerian assignment. Middle East terranes (Figures 1 and 2) not shown are Hunic Greater Caucasus and Cimmerian Alborz, Farah and Lesser Caucasus. Note that North Tibet and Qiangtang are synonyms in Torsvik and Cocks, but two different terranes in Stampfli et al.

are considered as separate terrane between the Hunic terranes and Baltica (in Laurussia) (Kríz et al., 2003; Torsvik and Cocks, 2004). Following the breakaway event, terranes adjacent to Gondwana were Apulia, Taurides, Sanandaj-Sirjan, Northwest and Central Iran, Helmand, South Tibet (Lhasa) and North Tibet (Qiangtang) terranes.

Some differences and confusion occur when comparing the two models in detail. The term Qiangtang (also Qangtang) is a synonym for North Tibet in Torsvik and Cocks (2004; Figure 2) and positioned next to South Tibet (Lhasa). In contrast, Stampfli et al. (2001) show Qiangtang as Hunic but North Tibet as Gondwanan (Figure 6). Other confusing terms are Karakum and Karakoram (also spelled as Karakorum). Karakum and neighboring Mangyshlak of Torsvik and Cocks (2004) are equivalent to the Karakum-Turan terrane (Figure 1). Karakum-Turan was not attached to Gondwana in the late Paleozoic (L. Angiolini, written communication, 2006) and probably Hunic (Figure 6). Karakoram is located in northern Pakistan (Gaetani, 1997; Figure 1), which belonged to Cimmeria (L. Angiolini, written communication, 2006). Further studies of the Cambrian-Ordovician rocks in Karakorum, based on the works of Gaetani et al. (1996), Gaetani (1997), Quintavalle et al. (2000) and Rolland et al. (2002), may provide new insights for its early Paleozoic paleoposition. Turan is often mentioned as a plate, but the Russian term ‘plate’ differs in meaning from ‘tectonic plate’, causing some further confusion (Laz’ko, 1975). It remains unclear whether Karakum and Turan formed one or several terranes.

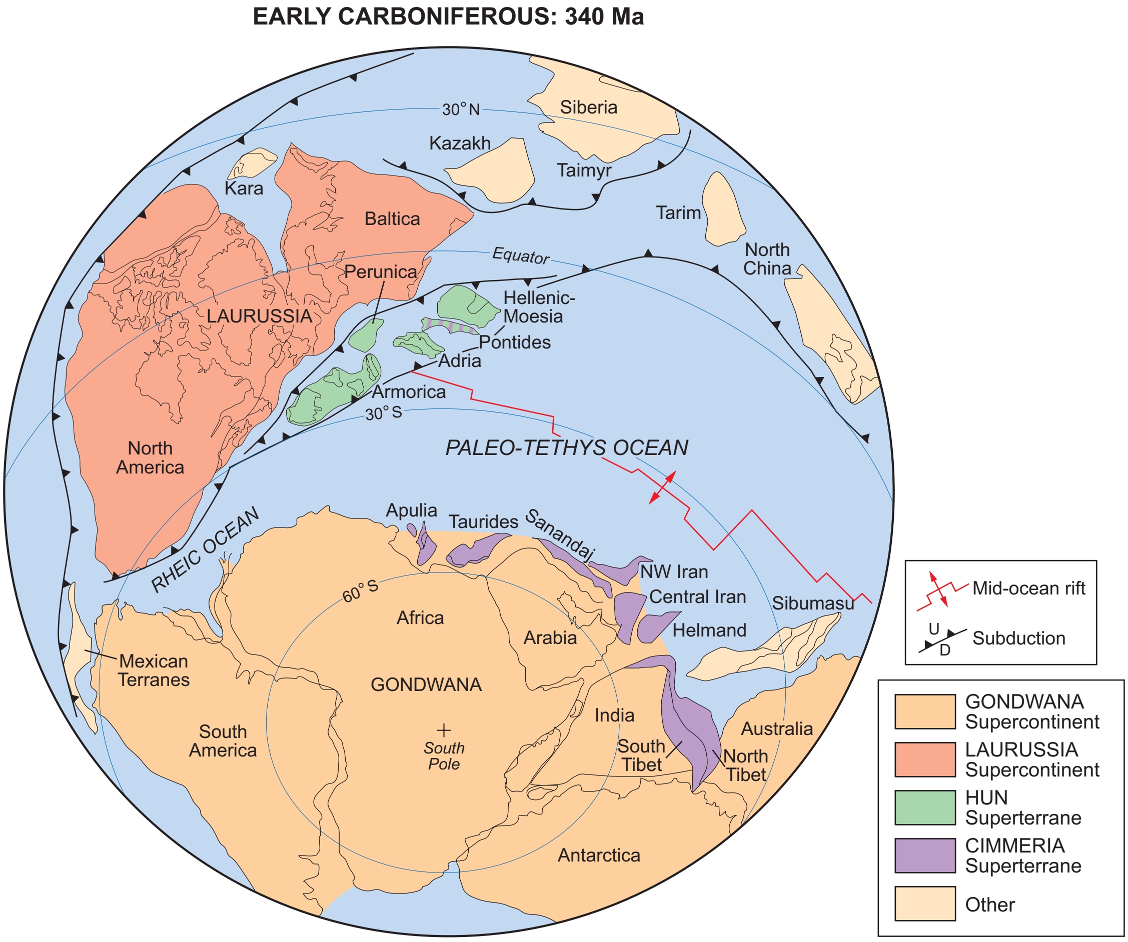

Figure 8: Plate-tectonic reconstruction of the early Carboniferous (Mississippian) (modified after Torsvik and Cocks, 2004). The Hun Superterrane had closed the Proto-Tethys Ocean and the Rheic Ocean was closing as Gondwana and Laurussia drifted towards each other. Middle East terranes (Figures 1 and 2) not shown are Hunic Greater Caucasus and Cimmerian Alborz, Farah and Lesser Caucasus. Pontides shown as Hunic by Torsvik and Cocks (2004), but considered Cimmerian in our paper (in stripes).

Cimmeria Superterrane

In the mid-Permian - Triassic, Cimmeria started rifting away from Pangea and closing the Paleo-Tethys Ocean to the north, while opening the Neo-Tethys Ocean in its wake (Figure 10). Less clear is which terranes were Cimmerian, or Hunic or possibly neither. Torsvik and Cocks (2004, Figure 10) show Cimmeria to consist of Apulia, Taurides, Sanandaj-Sirjan, Northwest and Central Iran, Helmand and North Tibet (Qiantang). They place South Tibet (Lhasa) to the north of India, however noting that it is not constrained by paleomagnetic or faunal content. Stampfli et al. (2001) and Stampfli and Borel (2002) (Figures 1, 2 and 6) included in Cimmeria: Apulia, Hellenides-Taurides, Menderes-Taurus, Sanandaj-Sirjan, Northwest and Central Iran, Helmand, South and North Tibet. A comparison indicates that several Middle East terranes (Northwest and Central Iran, Taurides and Sanandaj-Sirjan) are considered Cimmerian by both groups of authors.

In contrast to the somewhat generalized Cimmeria of some authors (e.g. Sharland et al., 2001; Stampfli et al., 2001; Stampfli and Borel, 2002; Torsvik and Cocks, 2004), Sengör (1990, Figure 11) showed the Cimmeria breakaway event in substantial detail and to consist of three ribbons. He divided North Tibet into East and West Qiangtang and considered the former as the leading Cimmerian ribbon. The "Intermediate" ribbon consisted of East Pontides, Dzirula Massif, Artvin/Karabagh, Sanadaj-Sirjan, Northwest Iran (including Alborz), Central Iran (Yazd, Tabas, and Lut), Farah, central Pamirs (China) and West Qiangtang. The trailing third ribbon included Helmand and South Tibet. The latter two ribbons connected to Australia, and the Neo-Tethys Ocean consisted of several seaways. Significantly for our paper, Sengör's model intepreted a subduction zone along the northeast front of Sanandaj-Sirjan and other northerly terranes, a subject that will be discussed later. The Intermediate ribbon of Natal'in and Sengör (2005) is generally comparable with the Cimmerian Superterrane.

PALEOZOIC OROGENIES AND THE ARABIAN PLATE

In most of the pre-Permian Paleozoic reconstructions (Figures 4 to 9), the Arabian Plate is generally depicted inland from the Tethyan margins of Gondwana, or later Pangea. Until the mid-Permian (Figures 10 and 11), it is shown as bounded by the Middle East terranes and, for the most part, at latitudes of about 30° to 60° south. Two regional hiatuses that were associated with polar glaciations occurred in the Late Ordovician Hirnantian Stage (Figures 3 and 5; Vaslet, 1990; Abed et al., 1993) and in the late Carboniferous - early Permian (Figures 3 and 9; Osterloff et al., 2004).

Two structurally significant unconformities have been recognized in the Arabian Plate (Figure 3). The mid-Silurian (Wenlock) to Middle Devonian hiatus is regionally manifested in Syria and Iraq, and possibly other parts of the Middle East (Brew and Barazangi, 2001; Al-Hadidy, 2007). Because of its age, it was correlated to the Caledonian Orogeny by some authors (e.g. Buday, 1980). In southeast Arabia (Oman), distinct hiatuses occur in the mid-Silurian (Wenlock) to earliest Devonian, and in the Middle Devonian to mid-Carboniferous (Millson et al., 1996; Droste, 1997; Osterloff et al., 2004). It would therefore appear that parts of Arabia could have been uplifted as highlands, most probably sometime between the mid-Silurian and Middle Devonian. These highlands may be related to pre-rift thermal swelling or post-rift isostatic rebound associated with the breakaway of the Hun Superterrane (Figure 6), rather than the Caledonian Orogeny (Figure 5).

The mid-Carboniferous unconformity is sometimes correlated to the Hercynian Orogeny, and the term "Hercynian unconformity" is adopted in regional and local studies by numerous authors (e.g. Stöcklin and Setudehnia 1972; Berberian and King, 1981). In Saudi Arabia, the angular pre-Unayzah unconformity correlates to the mid-Carboniferous hiatus (Figure 3, c. 325–310 Ma, Al-Husseini, 2004; at least Serpukhovian, Bashkirian and early Moscovian, Gradstein et al., 2004). The associated differential structural relief is manifested by broad epeirogenic swells (many 100s of kilometers in lateral extent) and laterally extensive (100s of kilometers) upthrown blocks (several 100s of meters), bounded by transpressional to reverse faults (Wender et al., 1998; Al-Husseini, 2004). The Hercynian Orogeny appears to have been too distant to account for the severity and style of this in deformation in Arabia (Figure 8 and 9).

An alternative to correlating the Hercynian Orogeny to the mid-Carboniferous Arabian unconformity is considered in the interpretation shown in Figure 11 of Sengör (1990). This Early Triassic reconstruction shows a SW-oriented subduction zone of the Paleo-Tethys oceanic crust beneath parts of Cimmeria. Next to the subduction complex, the Podataksasi Arc (a name Sengör derived from the initial letters of Pontides, Dzirula, Adzharia-Trialeti, Artvin-Karabagh and Sanandaj-Sirjan; Figure 11) was mainly a Carboniferous episode of orogenic deformation, metamorphism, and arc-type, calk-alkaline magmatism. This interpretation is based on a detailed study of successions in the involved terranes (see Sengör, 1990). Natal'in and Sengör (2005) included the Podataksasi Arc in the so-called Silk Road Arc, which stretched during the late Paleozoic-early Mesozoic from the Caucasus through north Iran and the Pamirs to China.

Further westwards, Xypolias et al. (2006) extended the subduction-arc model from the Pontides to the Hellenic terrane (External Hellenides) and northeast Greece. U-Pb dating of zircon from a granitic orthogneiss in the Kithira Island (southern Greece) yielded a late Carboniferous age of 324–323 Ma. Taken together with other geochronological data from the Aegean region it provides evidence for a restricted period of plutonism between 325–300 Ma (Xypolias et al., 2006). These authors concluded that northeast Greece (Cycladic and Palegonian basements) and northwest Turkey (Menderes Massif in the Taurides terrane and Sakarya Zone in the Pontides terrane) formed part of Cimmeria.

A reviewer (written communication, 2006) noted that a subduction complex does not transmit compressional horizontal stresses across broad regions and, moreover, a SW-directed compression

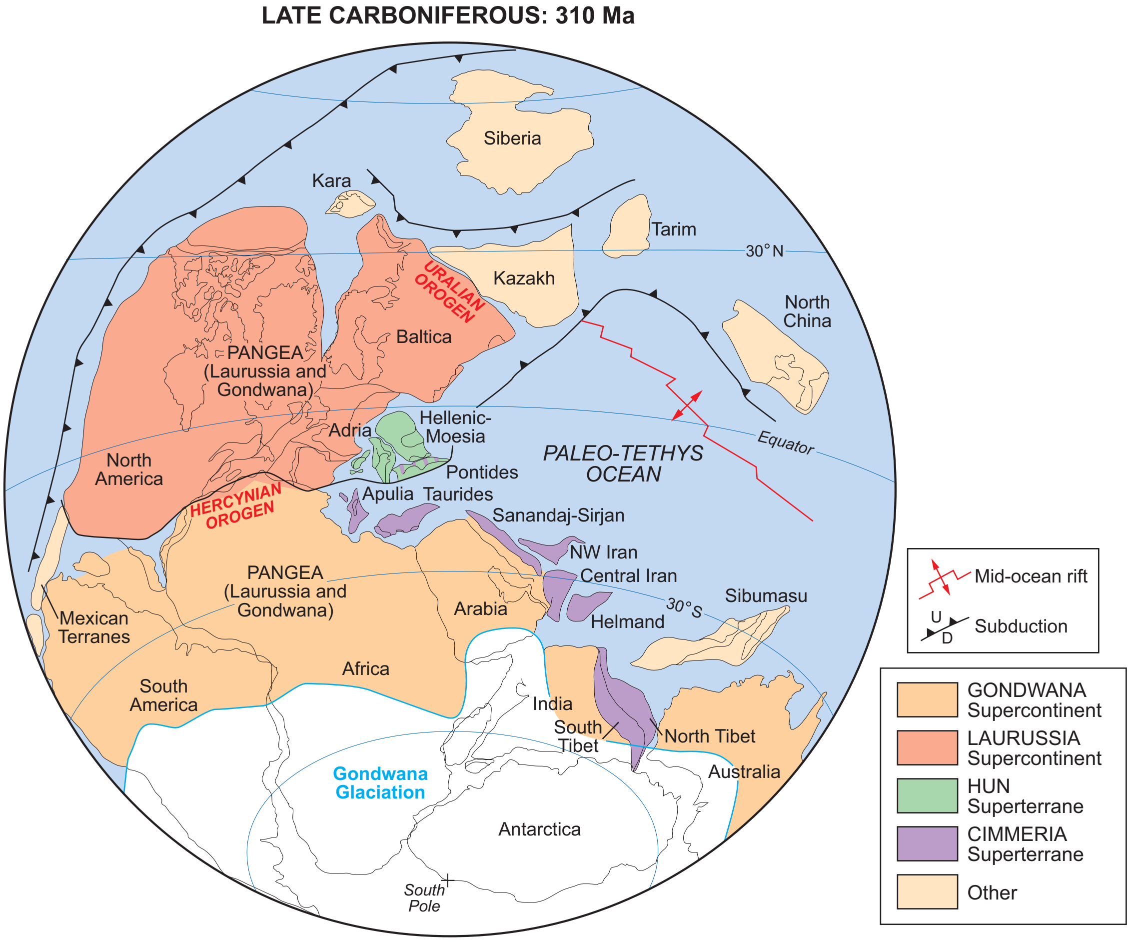

Figure 9: Plate-tectonic reconstruction of the late Carboniferous (Pennsylvanian; modified after Torsvik and Cocks, 2004). In the mid-Carboniferous times, the collision of Gondwana and Laurussia closed the Rheic Ocean causing the Hercynian Orogeny. Also in the late Carboniferous the collision of the Kazakh terranes with Pangea caused the Uralian Orogeny. In the Arabian Plate, an angular mid-Carboniferous unconformity is associated with epeirogenic swells and extensive compressional block faulting and a regional hiatus between about c. 325–310 Ma. The unconformity is widely referred to as "Hercynian" and associated with the Hercynian Orogeny. A more proximal tectonic event that may have caused the deformation of Arabia may have been the initiation of subduction along the Paleo-Tethyan margin of Cimmeria (Figure 11). The late Carboniferous - early Permian Gondwana glaciation extended into southern Arabia. Middle East terranes (Figures 1 and 2) not shown are Hunic Greater Caucasus and Cimmerian Alborz, Farah and Lesser Caucasus. Pontides shown as Hunic by Torsvik and Cocks (2004), but considered Cimmerian in our paper (in stripes).

appears inconsistent with the NS-trending grain of the fault-bounded Arabian structures. He pointed out that the initiation of subduction is commonly associated with a strong pulse of trench suction leading to roll-back, both processes exerting a strong extensional pull on the continental margin overlying the evolving subduction zone. As an analog to the Hercynian Orogeny's influence in Arabia, he suggested the present-day Indian Plate. It is piercing into the anisotropic assembly of Central South Asia, with its effect reaching even the distant east coast of Asia. He concluded that a far-field relationship to Hercynian orogenic forces still cannot be completely excluded in Arabia.

Besides the data supporting a compressional mid-Carboniferous pulse due to a subduction-arc complex (Sengör, 1990, Figure 11; Xypolias et al., 2006), the concerns raised by the reviewer can be addressed. Whereas subduction complexes are indeed driven by slab-pull and associated with back-arc extension in the continental margin, this regime develops after an initial compressional stage. The pre-subduction stage involves first rupturing the entire oceanic crust (10 or more kilometers thick) and initiating

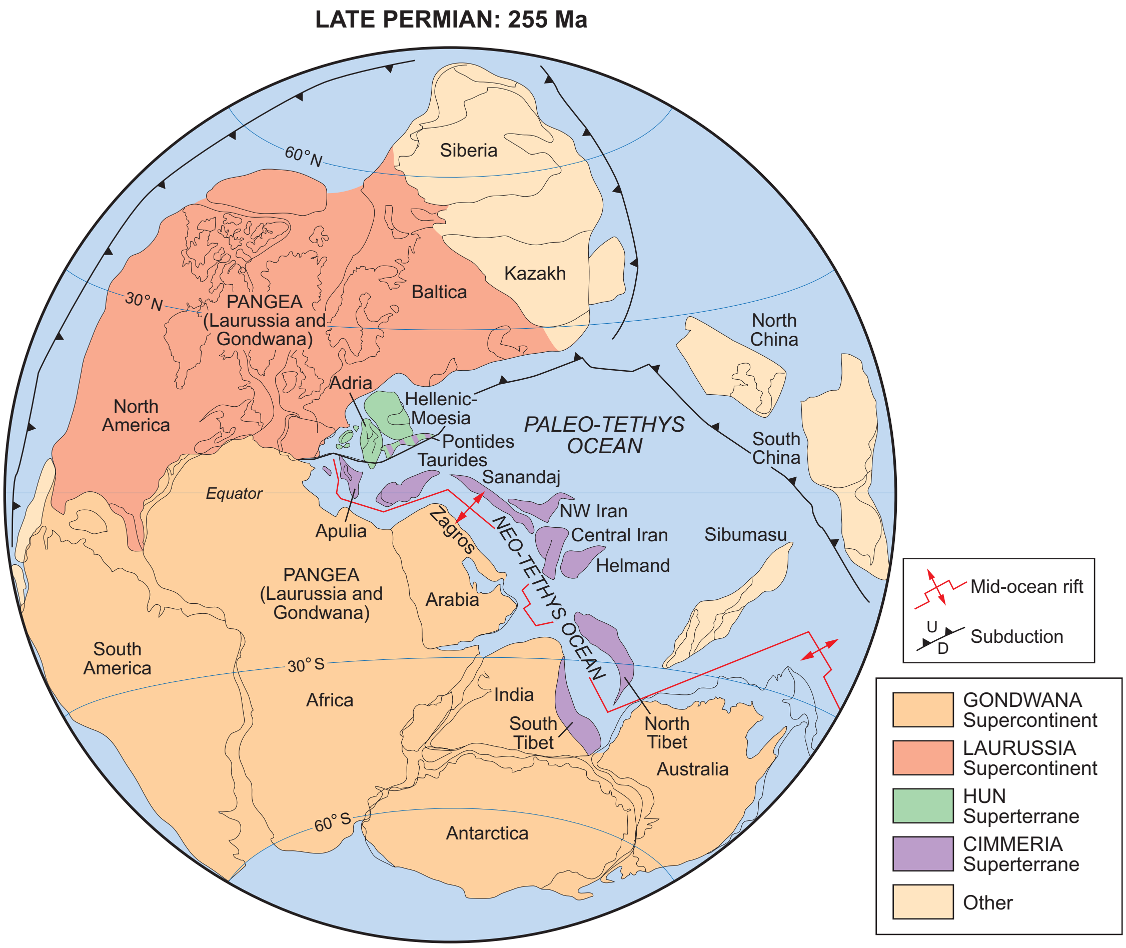
LATE PERMIAN: 255 Ma

Figure 10: Plate-tectonic reconstruction of the late Permian (Lopingian) time (modified after Torsvik and Cocks, 2004). In the mid-Permian - Triassic the Cimmeria Superterrane broke away from Gondwana, now part of the Pangea Supercontinent (Gondwana and Laurussia supercontinents, as well as Kazakh, Siberia and other terranes). Middle East terranes (Figures 1 and 2) not shown are Hunic Greater Caucasus and Cimmerian Alborz, Farah and Lesser Caucasus. Pontides shown as Hunic by Torsvik and Cocks (2004), but considered Cimmerian in our paper (in stripes).

subduction along a new thrust zone. The horizontal forces required to fracture the brittle crust, then bend and push down the oceanic plate are not only compressional but of regional significance. In some cases, the thrust geometry is reversed and the compressional force is great enough to push the oceanic crust above the continental margin resulting in an obduction. It is considered here that the mid-Carboniferous event was an early pre-subduction compressional phase, while the mid-Permian - Triassic was the extensional one.

The relationship between a SW-directed compression and the NS-trending Arabian uplifted fault blocks was one of mid-Carboniferous transpression along pre-existing NS-oriented fault systems. The Arabian basement manifests a fault system with a NS-, NE- and NW-trending grain that was established in the late Proterozoic and Early Cambrian (Al-Husseini, 2000, 2004). We argue that a SW-directed compressional pulse would have caused the pre-existing Arabian basement-cored structures to be dislocated in a right-lateral transpressional style.

In summary, it seems likely that two regional angular unconformities in the Arabian Plate were related to plate-tectonic events that occurred in the vicinity of the Middle East terranes. The mid-Silurian to Middle Devonian unconformity (middle Paleozoic event instead of Caledonian Orogeny) may have involved the uplift of the northern Arabia margin (and Oman) in Gondwana. The uplift may have occurred along the newly formed Paleo-Tethys Ocean prior to, during or possibly after the breakaway

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Research paper

Evolutionary rates of the Triassic marine macrofauna and sea-level changes: Evidences from the Northwestern Caucasus, Northern Neotethys (Russia)

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Abstract

A diverse Triassic marine macrofauna from the Northwestern Caucasus sheds new light on the biotic evolution after the end-Permian mass extinction. In the early Mesozoic, the study area was located on the northern margin of the Neotethys Ocean. Data on stratigraphic ranges of 130 genera of brachiopods, bivalves, ammonoids, corals, and sponges have been used to calculate the changes in two evolutionary rates, namely faunal transformation rate (FTR) and rate of transformation of the taxonomic diversity structure (TTDSR). The FTR demonstrates the changes in the generic composition of assemblages through geologic time, whereas the TTDSR indicates changes in the generic control of the species diversity. The Triassic marine macrofauna of the Northwestern Caucasus was characterized by very high FTR and TTDSR during the Early Triassic through early Late Triassic. The FTR slowed in the Middle Triassic, and accelerated again in the Carnian–Norian. In contrast, the FTR was abnormally slow in the Norian–Rhaetian. A remarkable turnover among macrofauna occurred at the Carnian–Norian transition. Regional sea-level changes were similar to the global eustatic fluctuations. It is difficult to establish their direct connections with changes in the evolutionary rates, although the turnover at the Carnian–Norian boundary coincided with a prominent regressive episode. In general, high evolutionary rates reported for the Triassic marine macrofauna of the Northwestern Caucasus may be explained as a consequence of the devastating end-Permian mass extinction.

Keywords: Evolutionary rate; Marine macrofauna; Sea level; Triassic; Caucasus

1. Introduction

The greatest Permian/Triassic mass extinction devastated the Earth's marine biota about 251 Ma, and it was followed by a long-term and poorly-known recovery (Sepkoski, 1993; Hallam and Wignall, 1997; Peters and Foote, 2001; Benton and Twitchett, 2003; Bottjer, 2004; Pruss and Bottjer, 2004; Racki and Wignall, 2005; Erwin, 2006). The marine communities recovered only in the Anisian (Erwin, 2006), when the benthic fauna strongly diversified (Komatsu et al., 2004; Ruban, 2006a,b). However, the entire Triassic may be considered a recovery interval, during which biodiversity was less than in the Permian (Peters and Foote, 2001). Significant evolutionary changes, however, occurred during the Triassic; e.g., bivalves began to dominate over brachiopods (Bonuso and Bottjer, 2005). The end of the Triassic was marked by a new mass extinction, which reduced the global marine diversity again (Sepkoski, 1993; Hallam and Wignall, 1997; Hallam, 2002; Pálfy et al., 2002). Thus, the Triassic recovery was unable to compensate for the biodiversity loss after the end-Permian mass extinction. Many factors, including sea-level changes, controlled the evolution of the Triassic marine biota. However, our knowledge of their influences remains limited.

This paper aims at a calculation of the evolutionary rates of the Triassic marine macrofauna of the Northwestern Caucasus (Fig. 1), where 130 genera of ammonoids, brachiopods, bivalves, sponges and corals are known from all Triassic stages (Ruban, 2006a). Previous studies analyzed the diversity dynamics of the Triassic marine biota of the Northwestern Caucasus (Ruban, 2006a,b,c), which was governed in particular by the changes in the basin depth. In this paper, the evolutionary rates are also compared to the sea-level changes, reconstructed for the study

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Fig. 1. Geographical and palaeogeographical location of the Northwestern Caucasus (slightly modified after Ruban, 2006a, 2007b). Palaeotectonics after Scotese (2004).

area. The main goals are to evaluate (1) how rapid was the Triassic evolution of the marine macrofauna, and (2) whether it was controlled by the sea-level changes.

2. Geological setting

The Northwestern Caucasus is a region in the south of Russia between the Kuban River and the Black Sea. During Triassic, the region was located on the northern margin of the Neotethys Ocean (Fig. 1), which was formed together with the northward drift of the Cimmerian terranes (Stampfli and Borel, 2002; Golonka, 2004; Scotese, 2004). The study area was situated much westwards from its present position, i.e., somewhere close to the Eastern Alps (Tawadros et al., 2006; Ruban, 2007a). According to Tawadros et al. (2006) and Ruban (2007a), this region was transported eastward to its present position along the major shear zone stretched along the Eurasian margin during the Late Triassic–earliest Jurassic. The sinistral displacements along this zone and rotation of Africa, which provoked them, have been discussed by Swanson (1982), Rapalini and Vizán (1993), and Ruban and Yoshioka (2005). Gaetani et al. (2005) pointed out the influences of the strike-slip movements in the Triassic evolution of the Northwestern Caucasus. In contrast, the traditional geodynamic model suggests that during Triassic, the study region was located on the southern periphery of the Russian Platform, where orogenic pulses occurred, coupled with extension (Ershov et al., 2003).

The Northwestern Caucasus was located close to the Tethyan region, although the global palaeobiogeographic differentiation was low in the Triassic (Westermann, 2000). This is supported particularly by the Caucasian signature of events documented in the “Boreal” bivalve evolution (Ruban, 2006c).

The Triassic stratigraphy and deposits of the Northwestern Caucasus were reviewed by Dagis and Robinson (1973), Jaroshenko (1978), Prozorovskaja (1979), Rostovtsev et al. (1979), Gaetani et al. (2005), and Ruban (2006a,c). The regional lithostratigraphy is presented in Fig. 2. In general, the Triassic deposits, up to 1700 m thick, are: the Induan–Anisian carbonates, the Anisian–Carnian turbidites, and the Norian–Rhaetian carbonates, including reefal limestones. In some parts of the study area, shales interbedded with clastics and carbonates characterize the Norian. The major regional hiatuses are known in the Early Induan, Late Anisian, at the Carnian–Norian boundary, and in the Late Rhaetian. The age of the lithostratigraphic units was established based on an analysis of macro- and microfauna; ammonoids, brachiopods, and foraminifers were especially important for these purposes (Dagis and Robinson, 1973; Rostovtsev et al., 1979; Prozorovskaja, 1979; Efimova, 1991; Gaetani et al., 2005). Additionally, palynological studies by Jaroshenko (1978) and Gaetani et al. (2005) permitted a precise verification of these stratigraphic constraints.

3. Materials and methods

The data on the stratigraphic ranges of species, which belong to 130 genera of the Triassic marine macrofauna of the Northwestern Caucasus, were compiled from all available sources (Paffengol’tz, 1959; Djakonov et al., 1962; Dagis, 1963, 1974; Shevyrjov, 1968; Dagis and Robinson, 1973; Jaroshenko, 1978; Prozorovskaja, 1979; Rostovtsev et al., 1979; Gaetani et al., 2005). These sources provided the reviews of the regional Triassic palaeontology that contain locality correlations and section-by-section compilations. The original data came from many sections (including those stretched along the valleys of the Belaja River and the Sakhraj River) within the study area. It appears that the compiled information is representative, because every interval of the regional Triassic succession was studied carefully (e.g., Dagis and Robinson, 1973; Rostovtsev et al., 1979).

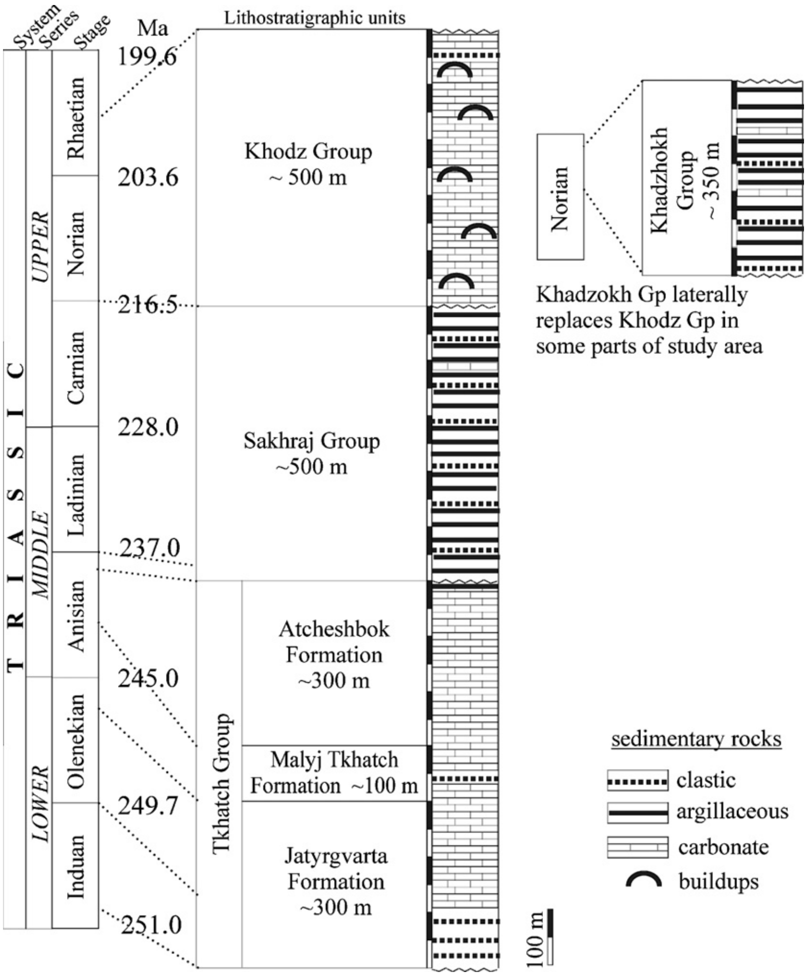

Fig. 2. Triassic lithostratigraphy of the Northwestern Caucasus (after Dagis and Robinson, 1973; Jaroshenko, 1978; Prozorovskaja, 1979; Rostovtsev et al., 1979; Gaetani et al., 2005; Ruban, 2006a,c). Absolute ages of stage boundaries are taken from Gradstein et al. (2004).

These data are essentially the same as those published by Ruban (2006a), although a few taxa identified with "sp." were added. The distribution of fossils in each Triassic stage has been tabulated (Appendix A). It seems impossible to differentiate the data on the Induan and Olenekian, because they often were not differentiated in the available literature sources (e.g., Dagis and Robinson, 1973). The Triassic marine faunas of the Northwestern Caucasus were composed of 130 genera: 42 ammonoid, 61 brachiopod, 17 bivalve, 7 coral, and 3 sponge genera. These genera contain about 250 species, the ranges of which were well documented. When possible, a taxonomic correction was attempted to avoid under- or over-estimation of the evolutionary rates. The conclusions reached in this paper are limited regionally. And it appears to be complete on a regional scale. However, no regional database is complete enough to be used as a sole basis for further global constraints. An analysis of any regional database nevertheless provides the evidence for discussion of the planetary-scale patterns of biotic evolution with regards to their signatures in a given region.

The chronostratigraphy used herein follows that of the International Commission on Stratigraphy (Gradstein et al., 2004; Gradstein and Ogg, 2005, 2006).

Two evolutionary rates have been calculated herein. The methods of their quantitative evaluation were proposed by Ruban (2001a, 2002), and then described by Ruban and Tyszka (2005) and Ruban (2006d, 2007b). The first rate is the faunal transformation rate (FTR), which is estimated as $1/R$. R is a Jaccard's (1901) similarity of two fossil assemblages, which characterize two successive stratigraphic intervals:

$$R = \frac{C}{[(N_1 + N_2) - C]},$$

where C is the number of common taxa for two assemblages, N_1 and N_2 are numbers of taxa in the earlier and later assemblages respectively. The FTR demonstrates how rapidly the changes occurred in the taxonomic composition of assemblages through geologic time. This paper is attempted to evaluate FTR for the generic assemblages of the entire marine macrofauna and particular fossil groups.

The second rate is the rate of transformation of the taxonomic diversity structure (TTDSR), which is estimated as 1/Rst. Rst is a coefficient of Spearman's rank correlation (Kendall, 1970) between two assemblages by presence/absence of genera, where the presence of a genus is indicated by the number of species that belong to this genus in the particular assemblage. In other words, a correlation established takes into account the "weight" of each genus in a given age. This quantitative approach is described in detail by Ruban (2007b). The TTDSR shows how quick the changes in the generic control of the species diversity were, i.e., the change in significance of each genus for the determination of the species diversity. In general, this reflects the changes in the structure of taxonomic diversity and, therefore, in the structure of fossil assemblages. As in the previous case, TTDSR is evaluated for the entire marine macrofauna and particular fossil groups, whose genera and species are examined. It is always interesting to calculate Rst not only for the successive assemblages, but for those which existed in different times, e.g., for the Anisian and Carnian assemblages. This may have very interesting and important results (Ruban, 2001a,b, 2007b).

In this article, two patterns of regional sea-level changes are reconstructed with lithological data, namely transgressions–regressions and deepenings–shallowings. They are independent components of basin dynamics (Catuneanu, 2006; Ruban, 2007c). An appropriate general term to define both patterns is lacking. Tentatively, it is here called "sea-level changes". Based on the analysis of the distribution of the Triassic deposits in the Northwestern Caucasus, with initial information compiled from Rostovtsev et al. (1979) and Gaetani et al. (2005), it is possible to recognize transgressive and regressive episodes. For these purposes, the presence or absence of marine strata in four particular areas of the study region was outlined. These areas were distinguished by Rostovtsev et al. (1979). Then, a cumulative curve for the entire Northwestern Caucasus was evaluated. Evidently, a wider distribution of marine deposits at each given time slice indicates a transgression, and vice versa, i.e., a restricted distribution indicates a regression. A critical analysis of the palaeoenvironmental interpretations of Gaetani et al. (2005) results in recognition of terrestrial, relatively shallow-marine, and slope environments using the same criteria as those by Ruban (2007c). This permits documenting the changes in the maximum basin depth. A somewhat similar approach was adopted by Landing and Johnson (2003). Used for a global reference were global eustatic curve of Haq et al. (1987) slightly modified by Haq and Al-Qahtani (2005) as well as reconstructed global base-level changes by Embry (1997).

4. Evolutionary rates of marine macrofauna

The Triassic marine macrofauna of the Northwestern Caucasus was characterized by a very high FTR during the Early, Middle, and early Late Triassic (Fig. 3). The values of *R* were below 0.1. The FTR slowed in the Middle Triassic and accelerated again in the Ladinian–Carnian. This means very rapid changes in the generic composition of the assemblages until the Norian. In contrast, they were abnormally slow in the Norian–Rhaetian. The *R* value was as high as 0.415, which suggests a half similarity.

The TTDSR of the Triassic marine macrofauna of the Northwestern Caucasus was very high until the late Late Triassic (Fig. 4). Changes in the structure of taxonomic diversity slowed later. However, this evolutionary rate slightly accelerated in the Carnian–Norian. Therefore, this time interval was marked by a great turnover in the structure of fossil assemblages. Thus, the control of species diversity shifted strongly to other genera in the Norian in comparison to the Carnian. In contrast to the Rst index, the value of the *R* index decreased at the Ladinian–Carnian transition. The TTDSR became low in the Norian–Rhaetian. This suggests a high similarity of the taxonomic diversity structure between the Norian and Rhaetian assemblages.

Three fossil groups were responsible for the decrease in both evolutionary rates in the Norian–Rhaetian, namely, ammonoids,

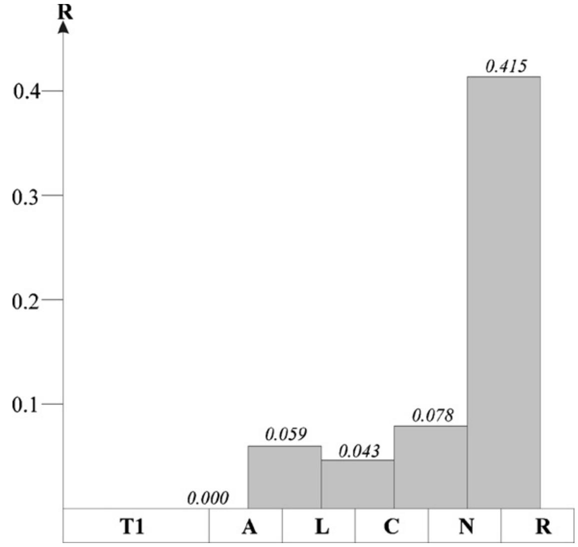

Fig. 3. *R* index of the entire Triassic marine macrofauna of the Northwestern Caucasus. Values of *R* are indicated above the columns. Stage abbreviations: A – Anisian, L – Ladinian, C – Carnian, N – Norian, R – Rhaetian; T1 – Early Triassic.

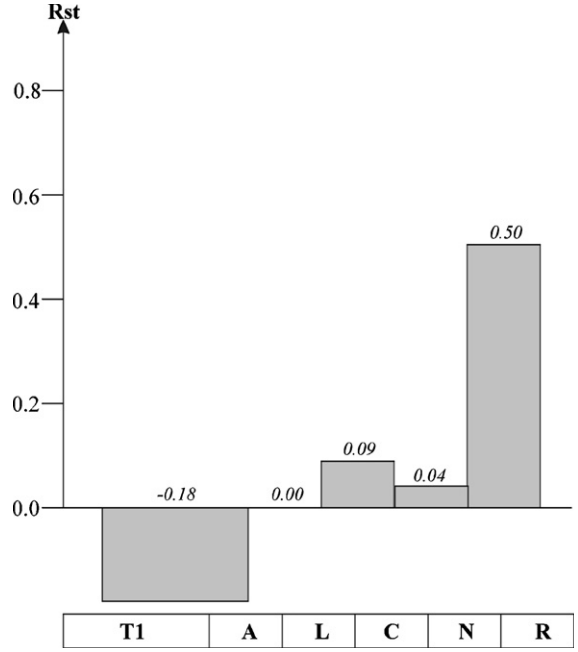

Fig. 4. Rst index of the entire Triassic marine macrofauna of the Northwestern Caucasus. Values of Rst are indicated above the columns. Stage abbreviations – see Fig. 3.

Table 1
R index for the assemblages of particular fossil groups of the Triassic macrofauna from the Northwestern Caucasus

Fossil groups	E. Triassic–Anisian	Anisian–Ladinian	Ladinian–Carnian	Carnian–Norian	Norian–Rhaetian
Ammonoids	0	0.09	0	0	0.56
Bivalves	0	0.11	0.17	0	0
Brachiopods	0	0	0	0.14	0.48
Benthic macrofauna	0	0.04	0.06	0.09	0.39

corals, and brachiopods (Table 1). Many genera of these groups have representatives in both these stages. In contrast, no Rhaetian species of bivalves and sponges are found at all. However, as ammonoids, corals, and brachiopods belong to distinct ecological groups but all demonstrate similarly diminished evolutionary rates across the Norian–Rhaetian transition, it is possible that a real decline in the rate of fauna evolution took place at the Norian–Rhaetian transition. The whole benthic macrofauna, however, demonstrated the higher FTR at the Norian–Rhaetian transition than ammonoids did (Table 1).

A calculation of Rst among all assemblages (Table 2), both successive and non-successive, results in two important conclusions:

(1) The Early Triassic assemblages are more different from the Anisian assemblages than from the Ladinian and Carnian assemblages;
(2) The Norian and Rhaetian assemblages are more different from the Early Triassic, Anisian and Ladinian assemblages in comparison to the Carnian assemblage, although such a dissimilarity of the Rhaetian assemblage is slightly less.

The first conclusion coupled with the relatively high Rst value in the Ladinian–Carnian slightly reduces the importance of biotic changes during the Early Triassic through Carnian. The second conclusion suggests that the turnover at the Carnian–Norian transition was stronger than expected. It was something like a “revolution” in the regional history of the Triassic marine macrofauna. Both conclusions presented above are meaningful for the particular fossil groups (Table 3).

5. Sea-level changes

The documented evolutionary rates of the Triassic marine macrofauna of the Northwestern Caucasus might have been related to the regional sea-level changes. My reconstruction here documents the transgressions–regressions and deepenings–shallowings in the Northwestern Caucasus, which are compared with the global eustatic fluctuations (Haq et al., 1987; Embry, 1997; Haq and Al-Qahtani, 2005) (Fig. 5). Sea-level changes in the beginning of the Triassic are debated. Although Haq et al. (1987), Embry (1997), and Haq and Al-Qahtani (2005) pointed out the low sea level at the Permian–Triassic transition, Hallam and Wignall (1999) argued that a significant transgression occurred. In the Northwestern Caucasus, this transition is marked by the marine facies of the Abag Formation (Prozorovskaja, 1979), but these deposits are capped by an unconformity (Dagis and Robinson, 1973; Prozorovskaja, 1979; Rostovtsev et al., 1979; Gaetani et al., 2005; Ruban et al., in press). Perhaps this earliest Triassic hiatus was a result of local tectonic activity. After the short-term hiatus at the beginning of the Triassic, the sea rapidly transgressed over the study area (Fig. 5). This coincided with the global eustatic rise. However, in spite of global tendencies, a stepwise regression occurred in the middle-late Anisian. A global eustatic fall in the late Anisian corresponded to the regional hiatus. The next transgression, which also was very rapid, occurred in the latest Anisian and reached its maximum in the Ladinian. However, the sea occupied less territory, than during the previous cycle. This transgression corresponded well with the global eustatic rise. There is no evidence for the major late Ladinian regressive episode in the Northwestern Caucasus, although it occurred globally. In contrast, the gradual regional regression in the Carnian occurred at the same time, as the global sea level was rising. The global eustatic fall at the Carnian–Norian transition evidently coincided with the regional hiatus (Fig. 5). In the Norian, a new regional transgression occurred. It was the largest among the other Triassic regionally-documented transgressions, and it had analogues in the global record. Regression occurred abruptly in the Late Rhaetian, which is a well recognized global pattern. Therefore, three transgressive–regressive cycles are established in the Triassic history of the Northwestern Caucasus. Each began with a rapid transgression. Differences between the regional transgressions–regressions and global eustatic fluctuations may be explained in terms of local tectonics. Both eustasy and tectonics are responsible for the regional transgressions and

Table 2
Rst index for the assemblages of the Triassic marine macrofauna from the Northwestern Caucasus

Assemblages	E. Triassic	Anisian	Ladinian	Carnian	Norian	Rhaetian
E. Triassic	–	**−0.18**	−0.07	−0.09	**−0.19**	−0.15
Anisian	–	–	0.00	−0.07	**−0.34**	**−0.28**
Ladinian	–	–	–	0.09	−0.15	−0.05
Carnian	–	–	–	–	0.04	0.06
Norian	–	–	–	–	–	**0.50**

Significant values are highlighted as bold.

Table 3
Rst index for the assemblages of particular fossil groups of the Triassic macrofauna from the Northwestern Caucasus

Assemblages	E. Triassic	Anisian	Ladinian	Carnian	Norian	Rhaetian
Ammonoids						
E. Triassic	–	**−0.40**	−0.16	−0.12	−0.22	−0.21
Anisian	–	–	−0.03	−0.07	−0.23	−0.25
Ladinian	–	–	–	−0.07	−0.13	0.09
Carnian	–	–	–	–	−0.09	−0.09
Norian	–	–	–	–	–	**0.50**
Bivalves						
E. Triassic	–	−0.18	−0.13	−0.09	−0.16	0.00
Anisian	–	–	−0.17	−0.18	−0.46	0.00
Ladinian	–	–	–	0.22	−0.32	0.00
Carnian	–	–	–	–	−0.24	0.00
Norian	–	–	–	–	–	0.00
Brachiopods						
E. Triassic	–	−0.14	0.00	−0.09	−0.17	−0.15
Anisian	–	–	0.00	−0.01	**−0.38**	**−0.36**
Ladinian	–	–	–	0.00	0.00	0.00
Carnian	–	–	–	–	0.22	0.14
Norian	–	–	–	–	–	**0.56**
Benthic macrofauna						
E. Triassic	–	−0.10	−0.04	−0.06	−0.13	−0.10
Anisian	–	–	−0.05	−0.03	**−0.39**	**−0.29**
Ladinian	–	–	–	**0.22**	−0.15	−0.12
Carnian	–	–	–	–	0.00	0.05
Norian	–	–	–	–	–	**0.48**

Significant values are highlighted as bold.

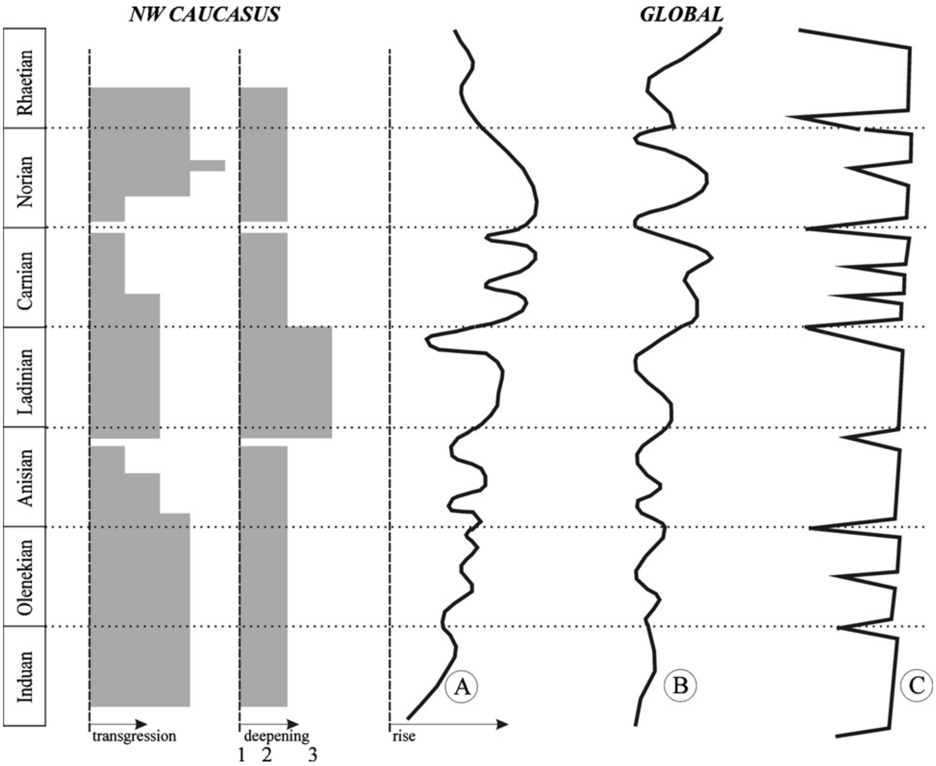

Fig. 5. Regional and global sea-level changes. A – rescaled after Haq et al. (1987), B – rescaled after Haq and Al-Qahtani (2005), C – after Embry (1997). Numbers indicate environments: 1 – terrestrial, 2 – relatively shallow-marine, 3 – slope.

regressions (Catuneanu, 2006), and a balance between them enforces the shoreline shifts.

The Triassic seas in the Northwestern Caucasus were mostly shallow-water (Fig. 5). One long-term and pronounced deepening episode is documented. It occurred in the latest Anisian–Ladinian, when turbidites were accumulated (Gaetani et al., 2005; Ruban, 2006a). It seems that local tectonics enforced significant regional water-depth changes. However, the Ladinian deepening pulse documented in the study area coincided with the peak transgression in Western Europe (Jacquin and de Graciansky, 1998). This pulse may also be related to the eustatic rise documented by Haq et al. (1987), Embry (1997), and Haq and Al-Qahtani (2005). Undoubtedly, the marine basin, which embraced the Northwestern Caucasus in the Triassic, was not stable, but it fluctuated under both eustatic and tectonic controls. Those sharp changes in its extent and depth, which are established herein, indicate that the pulses are meaningful for further comparisons with the available palaeontological data.

6. A comparison of regional evolutionary rates and sea-level changes

The reconstruction of the regional sea-level changes (Fig. 5) allows comparison of them to the documented evolutionary rates (Figs. 3 and 4; Table 1). It seems that they did not directly coincide. However, some interesting relationships may be noted.

An increase of changes in the generic composition of assemblages and a major turnover in the diversity structure at the Carnian–Norian transition corresponds with the regional regressive episode. The latter provoked a significant interruption in the regional evolution of the marine macrofauna. When repopulation occurred in the early Norian, the marine macrofauna differed strongly from the Carnian assemblage because of the absence of transitional genera, which might have appeared in the end-Carnian and survived into the early Norian. It may be asked why there were no such significant changes in the late Anisian when another regression occurred. This can be easily explained by two reasons: (1) my data on the Anisian comprise the entire taxa, which existed both before and after the hiatus, and (2) the late Anisian regression was probably much shorter. Gaetani et al. (2005) suggested that the presence of late Carnian deposits is doubtful, and, consequently, the hiatus at the Carnian–Norian transition might have been even longer than traditionally assumed. Another hypothesis may be formulated. During Late Triassic, the Greater Caucasus Terrane began to move eastward (Tawadros et al., 2006; Ruban, 2007a), and, therefore, in the Norian, the region might be located in a palaeobiogeographically different area from where it was in the Carnian. But such an assumption requires strenuous testing. An easier explanation is a change in the character of sedimentation, which occurred within the Western Caucasus. The Sakhraj Group is dominated by clastic deposits, whereas the overlying Khodz Group is composed of carbonates. However, carbonate beds already appeared in the Carnian (Fig. 2).

Another feature, which may be of interest, is relatively high similarity between the Norian and Rhaetian assemblages, which is suggested by both *R* and Rst values (Figs. 3 and 4; Table 1). The mentioned pair of assemblages corresponds to the transgressive–regressive cycle, i.e., the deposition was not interrupted during the Norian–Rhaetian intervals.

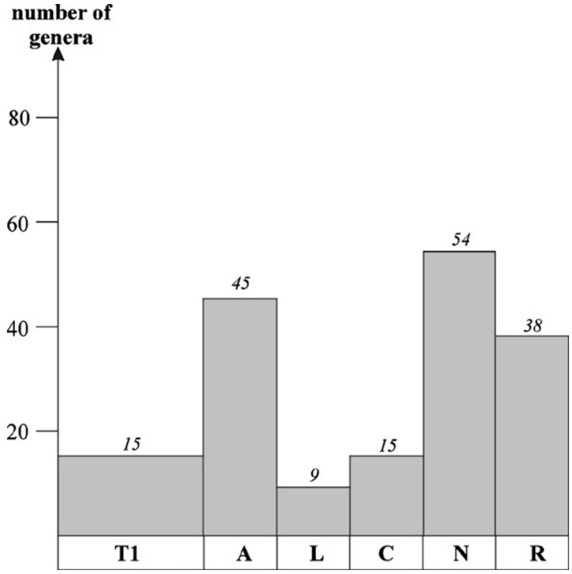

Fig. 6. Diversity dynamics of the Triassic marine macrofauna of the Northwestern Caucasus. Numbers of genera are indicated above the columns. Stage abbreviations – see Fig. 3.

Therefore, sea-level changes were potentially able to control in part the evolutionary rates of the Triassic marine macrofauna in the Northwestern Caucasus.

7. Discussion

Fossil assemblages may be dissimilar not only in their taxonomic composition but also in their total diversity. In such cases, the values of *R* and Rst indices are usually low. To test this explanation for the Triassic marine macrofauna of the Northwestern Caucasus, the total number of genera in each stratigraphic interval has been calculated (Fig. 6). It is evident that the Anisian generic diversity increased about 3 times above that of the Early Triassic. The Ladinian diversity decreased five times. Differences in the number of genera between the Ladinian and Carnian are evident but not great. The Norian diversity was 3.6 times greater than the Carnian diversity, comparable to the Rhaetian diversity. Consequently, differences in diversity were strongly reflected in *R* and Rst values. However, no Early Triassic genera are known in the Anisian (see data in Ruban, 2006a). Therefore, high evolutionary rates in the Early Triassic–Anisian period occurred due to the real turnovers. Differences in the total diversity enforced the *R* and Rst for the Anisian–Ladinian. Three Anisian genera are known in the Ladinian, suggesting that the Ladinian assemblage was formed by the diminishing of the Anisian diversity, but origination of new taxa, and, consequently, turnover were stronger. This is reflected by the *R* and Rst values (Figs. 3 and 4), which are higher than during preceding intervals, although not significantly. It seems that the FTR and the TTDSR at the Ladinian–Carnian interval reflects mostly the turnovers, because of small differences in their total diversity. As for the Carnian–Norian interval, documented acceleration of the FTR and the TTDSR occurred mostly due to the turnovers, but not the differences in the total diversity, because only five Carnian genera are known in the Norian. A similarity of both generic composition and diversity structure between the Norian and Rhaetian assemblages is a result of weak turnover, in which 27 genera are common to these stages. Thus, the real effect of the differences in the total generic diversity on the evaluation of the evolutionary rates was not so great in our case, and the decreases of the FTR and the TTDSR were mostly the result of turnovers.

The structure of global marine communities strongly changed after the Permian–Triassic mass extinction (Sepkoski, 1993; Hallam and Wignall, 1997; Benton and Twitchett, 2003; Bottjer, 2004; Pruss and Bottjer, 2004; Racki and Wignall, 2005; Erwin, 2006). Extremely high regional evolutionary rates, documented in the Northwestern Caucasus, might have been a result of a "restart" of the biotic evolution after the Permian–Triassic mass extinction. By extrapolating the regional knowledge on the evolutionary rates of the marine macrofauna to the global scale, it is possible to conclude that the Permian–Triassic catastrophe was so large as to be followed by a recovery, which lasted until the late Late Triassic. However, data from the other regions are needed to make a complete database in order to verify this conclusion.

8. Conclusions

The Triassic marine macrofauna of the Northwestern Caucasus evolved very rapidly. Rates of change in both the generic composition of assemblages and the structure of their taxonomic diversity were high. The real slowing of the evolution occurred only in the Norian–Rhaetian interval. The diversity structure of the assemblages sharply changed at the Carnian–Norian transition. The Triassic sea-level changes in the Northwestern Caucasus corresponded generally to the global eustatic fluctuations, although the local tectonic movements induced the particular features of the regional sea-level changes. The regional sea-level changes were among the important factors influencing the evolutionary rates of the marine macrofauna.

The calculation of the evolutionary rates of marine faunas may be significant for the chronostratigraphic developments. Using the Caucasian data, Tozer (1988, 1990) recognized the Rhaetian as a provincial substage of the Norian. The FTR and the TTDSR in the Norian–Rhaetian interval were lower than during earlier times (Figs. 3 and 4). But R and Rst values were about 0.5, which is not as large, and, consequently, transformations in macrofaunal assemblages at the Norian–Rhaetian transition were relatively slower. Thus, there is no clear evidence to support combining the Norian and Rhaetian into a unique stage.

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Appendix A

Triassic macrofossil taxa from the Northwestern Caucasus. The number of species in each genus is indicated for stratigraphic intervals. The data are republished from Ruban (2006a) with few additions after a permission from the Editor of "Revue de Paléobiologie".

Genera	Early Triassic	Anisian	Ladinian	Carnian	Norian	Rhaetian
Abrekia	1	0	0	0	0	0
Acrochordiceras	0	3	0	0	0	0
Adygella	0	0	0	1	1	1
Adygelloides	0	0	0	0	0	1
Aegeiceras	0	1	0	0	0	0
Amphiclina	0	0	0	0	1	2
Ampliclinodonta	0	0	0	0	1	0
Angustothyris	0	1	0	0	0	0
Arcestes	0	0	1	0	0	4
Arpadites	0	2	0	0	0	0
Astraeomorpha	0	0	0	0	2	1
Aulacothyropsis	0	0	0	1	0	4
Austriella	0	0	0	0	1	0
Austrirhynchia	0	0	0	0	0	2
Badiotites	0	0	1	0	0	0
Balatonospira	0	0	0	1	0	0
Beyrichites	0	1	0	0	0	0
Bobukella	0	0	0	1	1	0
Cassianella	0	0	0	0	1	0
Caucasites	0	2	0	0	0	0
Caucasorhynchia	0	0	0	0	1	1
Caucasothyris	0	0	0	0	1	0
Cladiscites	0	0	0	0	2	2
Claraia	4	0	0	0	0	0
Coenothyris	0	2	0	0	0	0
Costirhynchia	0	1	0	1	0	0
Costispiriferina	0	1	0	0	0	0
Crurirhynchia	0	0	0	0	1	0
Crurithyris	1	0	0	0	0	0
Cubanothyris	0	0	0	0	2	2
Daonella	0	0	2	0	0	0
Decurtella	0	2	0	0	0	0
Dieneroceras	1	0	0	0	0	0
Dinarispira	0	2	0	0	0	0
Dioristella	0	1	0	1	0	0
Euxinella	0	0	0	0	3	6
Fissirhynchia	0	0	0	0	1	1
Flemingites	1	0	0	0	0	0
Flexoptychites	0	1	0	1	0	0
Guseriplia	0	0	0	0	0	2
Gymnites	0	1	0	0	0	0

Genera	Early Triassic	Anisian	Ladinian	Carnian	Norian	Rhaetian
Halobia	0	0	1	5	0	0
Hodsia	0	0	0	0	1	0
Hoernesia	0	1	0	0	0	0
Holcorhynchella	0	1	0	0	0	0
Hollandites	0	3	0	0	0	0
Indopecten	0	0	0	0	1	0
Japonites	0	1	0	0	0	0
Joannites	0	0	0	1	0	0
Juvavites	0	0	0	0	1	0
Koeveskallina	0	1	0	0	0	0
Koninckina	0	0	0	1	1	0
Laballa	0	0	0	0	2	3
Laboceras	0	3	0	0	0	0
Leda	0	0	1	0	0	0
Lepismatina	0	0	0	0	0	1
Leyophyllites	0	4	0	0	0	0
Limea	0	1	0	0	0	0
Lobites	0	0	0	0	1	0
Lobothyris	0	0	0	0	2	0
Longobardites	0	1	0	0	0	0
Lyssochlamys	0	0	0	2	0	0
Majkopella	0	0	0	0	0	3
Megaphyllites	0	1	0	0	2	2
Mentzelia	0	1	0	3	1	2
Mesocladiscites	0	1	0	0	0	0
Moisseievia	0	0	0	0	2	1
Molengraffia	0	0	0	0	1	0
Monophyllites	0	1	3	0	0	0
Monotis	0	0	0	0	3	0
Montlivaultia	0	0	0	0	1	0
Myophoria	0	0	0	0	1	0
Mytilus	0	1	0	0	0	0
Nannites	1	0	0	0	0	0
Neoretzia	0	0	0	0	0	3
Neowelerella	1	0	0	0	0	0
Norella	0	1	0	0	0	0
Owenites	3	0	0	0	0	0
Oxycolpella	0	0	0	0	3	2
Paleocardita	0	0	0	0	1	0
Paracladiscites	0	0	0	0	0	2
Paradanubites	0	2	0	0	0	0
Paragoceras	1	0	0	0	0	0
Parasageceras	0	1	0	0	0	0
Parussuria	1	0	0	0	0	0
Pexidella	0	1	0	0	1	0
Phyllocladiscites	0	2	0	0	0	0
Piarorhynchella	0	1	0	0	0	0
Pinacoceras	0	0	0	0	1	1
Placites	0	0	0	0	1	1
Posidonia	0	1	1	0	0	0
Proptychites	1	0	0	0	0	0
Pseudocyrtina	0	0	0	0	1	0
Pseudomonotis	0	0	0	0	2	0
Pseudorugitella	0	0	0	0	2	2
Pseudosageceras	1	0	0	0	0	0
Punctospirella	0	1	0	0	0	0
Rhabdophyllia	0	0	0	0	1	0
Rhacophyllites	0	0	0	0	1	2
Rhaetina	0	0	0	2	6	4
Rhimirhynchopsis	0	0	0	0	1	1
Rhynchonella	0	1	0	0	0	0
Robinsonella	0	0	0	0	0	1
Sahraja	0	0	0	0	1	0
Schafhaeutlia	0	1	0	0	0	0
Sinucosta	0	1	0	0	1	1
Smithoceras	0	1	0	0	0	0
Spinolepismatina	0	0	0	0	1	0
Stephanocoenia	0	0	0	0	1	0
Sturia	0	2	1	0	0	0
Stylophyllopsis	0	0	0	0	2	0
Subowenites	1	0	0	0	0	0
Subvishnuites	1	0	0	0	0	0
Sulcatinella	0	2	0	0	0	0
Sulcatothyris	0	0	0	1	0	0
Tetractinella	0	1	0	0	0	0
Thamnastraea	0	0	0	0	2	2
Thecosmilia	0	0	0	0	6	2
Thecospira	0	0	0	0	1	0
Thecospiropsis	0	0	0	0	1	0
Triadithyris	0	0	0	0	2	1
Trigonirhynchella	0	0	0	0	1	2
Velopecten	0	1	0	1	0	0
Volirhynchia	0	2	0	0	0	0
Wittenburgella	0	0	0	0	1	1
Worobievella	0	0	0	0	1	1
Wyomingites	1	0	0	0	0	0
Xenodiscus	0	0	1	0	0	0
Zeilleria	0	0	0	0	3	6
Zugmayerella	0	0	0	0	0	1

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Possible pitfalls in the procedure for paleobiodiversity-dynamics analysis

Pułapki analizy paleo-bioróżnorodności

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Abstract

The changes in the diversity of specific taxa during certain parts of the geological past (paleobiodiversity dynamics) can, in principle, be established by counting the number of the fossil taxa present (worldwide or in a specific study area) in rocks dated for the time interval under study. Numerous obstacles are present, however, for instance in the form of lacking field data, disappeared collections, ambiguous identifications, temporary 'disappearence' of taxa, and dating problems. One major problem is the fact that, particularly in regional studies in some countries, a local, regional or national chronostratigraphic terminology is used rather than the chronostratigraphy recommended by the International Stratigraphic Commision of the International Union of Geological Sciences. This hampers international correlation and makes precise global paleodiversity-dynamics analyses extremely difficult. A reliable insight into the true paleodiversity dynamics requires not only that the various problems are recognized, but also that their consequences are eliminated or, if this is impossible, minimized. This is particularly important if the effects of mass extinctions on fauna and flora are investigated. Each analysis of paleobiodiversity-dynamics analysis of phenomena related to mass extinctions should therefore try to quantify the impact that missing data or inaccuracies of any kind may have on the final results; such an analysis should, in addition, try to find a solution for the major problems, so as to avoid significant inaccuracies of the calculated values. Large electronic databases can help, since about a decade, to diminish possible errors in diversity estimates. Paleobiodiversity should preferably be expressed in the form of values with a certain band with, indicating the inaccuracy, rather than in the form of exact values.

Keywords: biodiversity, taxonomy, biostratigraphy, geochronology, Lazarus taxa

Streszczenie

Zmiany w zróżnicowaniu gatunków w pewnych przedziałach czasu przeszłości geologicznej (dynamika paleo-bioróżnorodności) są z zasady ustalane poprzez zliczanie liczby taksonów skamieniałości (na świecie lub na wybranym obszarze) w skałach datowanych na badany interwał czasowy. Tym niemniej procedura ta napotyka wiele przeszkód, np. w postaci braku danych z jakiegoś obszaru, zagubionych kolekcji, niejednoznacznych identyfikacji, czasowego „zaniku” taksonów czy problemów datowania. Jednym z głównych problemów, zwłaszcza w badaniach regionalnych w niektórych krajach, jest stosowanie lokalnej, regionalnej lub krajowej terminologii chronostratygraficznej, a nie chronostratygrafii rekomendowanej przez Międzynarodową Komisję Stratygraficzną przy Międzynarodowej Unii Nauk Geologicznych. Utrudnia to międzynarodowe korelacje i czyni niezwykle trudnym przeprowadzenie precyzyjnej globalnej analizy dynamiki paleo-bioróżnorodności. Wiarygodny wgląd w prawdziwą dynamikę paleo-bioróżnorodności wymaga nie tylko rozpoznania różnych problemów, ale również wyeliminowania ich konsekwencji, a gdy to niemożliwe, zminimalizowania ich. Jest to szczególnie ważne w przypadku, gdy badane są następstwa masowego wymierania fauny i flory. Dlatego każda analiza dynamiki paleo-bioróżnorodności zjawisk związanych z masowym wymieraniem powinna zawierać próbę ilościowego oszacowania wpływu, jakie brakujące dane lub niedokładności jakiegokolwiek rodzaju mogą wywierać na końcowe wnioski. Taka analiza powinna próbować znaleźć rozwiązanie dla głównych problemów, ażeby uniknąć znaczących niedokładności w obliczonych wartościach. Duże elektroniczne bazy danych, dostępne od około 10 lat, mogą pomóc w zmniejszeniu możliwych błędów przy szacowaniu różnorodności. Najlepiej, gdyby paleo-bioróżnorodność była wyrażana w formie wartości w pewnym zakresie, wskazującym na niedokładność, a nie w formie precyzyjnej wartości.

Słowa kluczowe: bioróżnorodność, taksonomia, biostratygrafia, geochronologia, taksony Łazarza

Introduction

Since the end of the 1970s, hundreds of articles and books have been devoted to changes in global biodiversity, mass extinctions and – more in particular – the changes in biodiversity patterns of numerous fossil groups. Regional patterns were also established for numerous regions worldwide. The studies by Sepkoski (1993, 1997) and his co-workers (Sepkoski *et al.* 1981; Raup & Sepkoski 1982) are of great importance in this context, as they have initiated new developments in historical geology and paleontology. This type of work coincided in time, to mutual benefit, with multidisciplinary studies that have led to the insight that exceptional events in the Earth's history have greatly affected biodiversity in time (e.g., Alvarez *et al.* 1980), but there exists no general agreement yet about extraterrestrial (e.g. impact of a bolide) or Earth-related (e.g. global environmental change) causes that must be held responsible for the various events (e.g., Courtillot 2007).

Most research on paleobiodiversity has concentrated on faunas, commonly marine ones, probably because of their higher preservation potential and because of the commonly lower number and the commonly shorter interruptions (hiatuses) in their record. Diversity changes in terrestrial floras have, in contrast, been studied in much less detail, and paleobotanical evidence for biotic crises is still scarce, which may be ascribed – at least partially – to the much less complete knowledge that is available nowadays about ancient terrestrial floras than marine biota. Attempts to estimate changes in the diversity of fossil plant assemblages have been undertaken by, particularly, Niklas *et al.* (1985), Boulter *et al.* (1988), Nishida (1991), Wing & DiMichele (1995), Tiwari (1996), Boulter (1997), and Philippe *et al.* (1999). The influences of mass extinctions on plants, including their diversity, were discussed by, among others, Tschudy & Tschudy (1986, Retallack (1995), Tiwari (2001), McAllister Rees (2002), and Wing (2004). Data about changes in paleobotanical diversity are still too scarce, however, to be reliable for statistically significant analyses. For instance, the highly interesting hypothesis of Guex *et al.* (2001) and Morard *et al.* (2003), later verified by Ruban (2004) and Ruban & Efendiyeva (2005) – which explains the end-Lias anoxia with the preceding plant growth – could not be supported by quantitatively sufficient data on changes in plant diversity, so that only qualitative and the simplest semi-quantitative data have been used. Most problems regarding the analyses of changes in paleobiodiversity (often referred to as 'paleobiodiversity dynamics') are similar for faunas and floras, although both groups have also their own specific problems because the evolution and preservation of both groups have their own specific aspects.

Some other aspects of paleodiversity-dynamics analysis have been reviewed and discussed by, among others, Benton (1995), Alroy

(2000, 2003), Foote (2000, 2007) and Hammer & Harper (2005). Several other works deal with the influence of differences in preservation potential and of the (in)completeness of the fossil record on diversity measurements and on the interpretation of the findings (e.g., Benton 1995; Peters & Foote 2001; Smith 2001, 2007; Twitchett 2001; Crampton *et al.* 2003; Vermeij & Leighton 2003; Bush & Bambach 2004; Boucot 2006; Peters 2006; Foote 2007). The present contribution is meant to present a brief overview of the more general problems related to the collection and compilation of data.

Collection of data

The first problem, met immediately when starting an analysis of taxonomic diversity dynamics, is how and where initial data (in their most simple form data regarding the stratigraphic range of a particular taxon) have to be collected. This information should, obviously, be as complete as possible; in addition, it should be representative and scientifically correct.

Data from literature can be found in two forms: (1) as dispersed information, spread over numerous publications, each of which is highly incomplete; and (2) as already compiled, fairly complete information on the stratigraphic distribution of the taxa being studied.

It is evident that – if sources of the second type are available – they should be chosen as a starting point. Their disadvantage is, however, that they are commonly outdated. This implies that, even if such compilations exist, additional collection of data from 'dispersed' sources remains essential. Thus, the search for data can be realized in the following three ways:

(1) if there are no sources with compiled data at all, a search for all possible publications and unpublished reports with „dispersed" information should be carried out [as an example: when the diversity studies of the Phanerozoic megaflora from the Northern Caucasus (Ruban 2003) was started, thousands of potential sources were checked, which took about a year];

(2) if there are several sources with compiled, but evidently incomplete or outdated data, they should be chosen as the basic ones, but the data found in them should be complemented with data from other sources with „dispersed" information (in exceptional cases, namely if the sources with compiled data seem neither incomplete nor outdated and if they seem sufficiently representative for the current knowledge, such basic sources may be used without additional search for complementary dispersed information);

(3) if there is only one source with previously compiled, but outdated information, it is necessary to search not only for complementary data in the literature that is devoted to the taxon/taxa under examination, but also for publications that might, as a 'by-product', contain information that might help to avoid shortcomings in the initial data (i.e., missing data, duplicate data).

Apart from carrying out the required literature search with great scrutiny, it is of utmost importance to decide carefully which of the three above approaches should be followed, as the choice of initial data is one of the most important factors that determine whether the inventory of data will result in a complete and reliable set of data that can form an adequate basis for the rest of the procedure.

Compilation of data

The compilation of data, which have commonly to be collected from numerous sources, is one of the most difficult, time-consuming and bothersome activities when preparing a quantitative analysis of paleodiversity dynamics. This holds for both global and regional studies.

Among the problems that are met during data compilation, one of the most important is taxonomical synonymy. Incorrect identifications of taxa, which are sometimes obvious (Sohn 1994; Benton 1995) but which are more often difficult to find out, strongly influence the results of diversity analysis. If the problem of synonymy is ignored, one taxon (e.g. a species, genus or family) may be counted for two or even more, suggesting a too large diversity; on the other hand, fossils that may represent different taxa may have been identified erroneously as identical, thus giving rise to an apparently too small number of taxa, and thus to an unduly low diversity. It is interesting in this context that DNA analysis becomes increasingly applicable to the establishment of taxonomic relationships [not only for fossils from the Pleistocene (see, among others, Rohland *et al*. 2007) but also for fossils that data back from tens of millions of years (see, among others, Wible *et al*. 2007)], thus suggesting that a DNA-based taxonomy may become feasible, as has been predicted already several years ago (Van Loon 1999). This might eventually help solving problems like those of synonyms and homonyms. In some cases (e.g., in the case of planktonic foraminifers), however, genetic explorations demonstrated that the available taxonomic classifications require fundamental re-consideration, which makes it difficult to measure diversity (Kucera 2007). The problems with synonymy should, indeed, be solved precisely, although sometimes (especially when „old" data are used) this cannot be realized due to a low quality of the initial information (e.g. the absence of the original fossil collections, of descriptions used for the taxonomic identification and/or classification, or of figures); it might in many cases even be better not to include such fossils in a paleodynamics-diversity study at all.

An example of synonymy

A characteristic example of a problem raised by synonymy is the confusion about two widespread Late Paleozoic plant genera, namely *Walchia* and *Lebachia*. According to Meyen (1987), *Lebachia* and *Lebachiaceae* are invalid taxa, which should be replaced by *Walchia* and *Walchiacae*; but he also states, remarkably enough, that these names may continue to be used "due to tradition". This 'taxonomic flexibility' is even more remarkable if one realizes that Carboniferous species are commonly indentified as *Lebachia*, whereas Permian remains with the same characteristics are commonly classified as *Walchia*. This 'tradition' is also followed in the Northern Caucasus: *Lebachia* species are found in the Pennsylvanian (Novik 1978), whereas *Walchia* has been described from Early Permian strata (Miklukho-Maklaj & Miklukho-Maklaj 1966). The 'Late Pennsylvanian' (i.e., Kasimovian and Gzhelian stages) assemblage consists of 34 genera, whereas the Permian flora contains only *Walchia* (Ruban 2003). An 'outsider' might conclude that – after the disappearance of the entire 'Late Pennsylvanian' flora – a new genus appeared in the Permian. The actual situation is, however, a sudden degradation of the flora at the end of the Carboniferous, with only one genus surviving into the Permian. This implies that a calculation of the rate of diversity dynamics on the basis of the disappearance of *Walchia* and the appearance of *Lebachia* would yield an incorrect value. Unfortunately, the name *Lebachia* still is found in recent publications, sometimes being even ascribed to the typical Permian *Walchia*, among others by Davydov & Leven (2003) who attempted, although they are not paleobotanists, to present a comprehensive overview for all kinds of stratigraphers/paleontologists.

An example of a problem raised by lack of correct data regarding distribution in time and space

Another significant problem is the frequent absence of clear indications regarding the exact position of taxa in time and space. For example, the presence of a particular taxon may be indicated for a specific study area, without exact data about the precise site or the age of the rocks in which the fossils under study were found; or the occurrence of a particular taxon may be indicated without information about its distribution in zones or even stages. It is, as a rule, highly questionable whether such data could be used, as diversity dynamics should based on "stage-by-stage" or "zone-by-zone" data.

A comparable problem is encountered if regional correlations become almost impossible by the use of different names for the same formation in sites far apart, or if different chronostratigraphic frameworks are used for different regions. A problem that is in many respects similar, but much more severe is posed by the frequent revisions of the geological time scale under the supervision of the International Commission on Stratigraphy (ICS) of the International Union of Geological Sciences (IUGS) (see, as an example, Gradstein *et al*. 2004). In a case of stage boundary re-definition, it may become uncertain whether the stage name has a former or present meaning.

Possible solutions

The above problems for data compilation may be (though sometimes only partly) solved in four ways:

(1) by revising the original data, for instance by re-examining the fossil collections;

(2) by recalculating data „as is", taking into consideration the possible impact of the insufficiently unambiguous data when interpreting the results of the fossil dynamics analysis (but it should be emphasized that such a recalculation almost inevitably reduces the scientific value of the analysis!); this type of recalculation of data was followed by, among others, Ruban (2005) in his discussion of paleontological data gathered in the middle of the 19th century;

(3) by adapting the objectives of the project, i.e. lowering the resolution of the analysis in time and/or space; as an example, the above-mentioned diversity of the Phanerozoic macrofloras from the Northern Caucasus was, as a result of such an adaptation, studied by estimating the number of taxa not per stage (as had been originally envisaged), but per series only (Ruban 2003); other examples have been provided by McGhee (1996);

(4) by using purposely developed numerical equations to predict or to minimize the errors linked to the compilation problems; these equations are similar to those used for evaluation of the preservation bias.

Which of the above approaches should be followed in order to obtain an optimum result, depends on the specifics of the initial data and of the project targets. Sometimes, however, the problems may be so large that the best solution may be to postpone the analysis until more and/or more reliable and unambiguous data become available. It is worthwhile mentioning in this context that the increasing number of electronic paleontological databases provides ever more data for a successful compilation process. The commonly not very clear original source of data from electronic data bases – and this holds even more for data found on internet – can, however, pose a problem; in addition, the reliability of data from not precisely known sources that cannot be trusted on the basis of peer review is dubious. Compilation of data on the basis of not generally recognized electronic sources is therefore not advisable.

Application of the compiled data

Even if data compilation has been successful and a range chart for the various taxa under study has been prepared, the application of the compiled data – for analysis of the fossil diversity dynamics – can meet severe problems. Some of these problems may be exemplified on the basis of a simple hypothetical chart with the stratigraphic distribution of 5 species (belonging to 2 genera) in the Triassic of an imaginary study area (Fig. 1). The data are complete, and, therefore, represent the actual situation. If the diversity dynamics (as defined on the basis of species) within the Early-Middle Triassic is to be analyzed, first the number of species present during each stage must be calculated. This is simple, but the calculating the diversity dynamics is less simple as will be shown here, because some new species appear for the first time, whereas other species become extinct.

To calculate the rate of the origination of new species in the lowermost part of the inves-

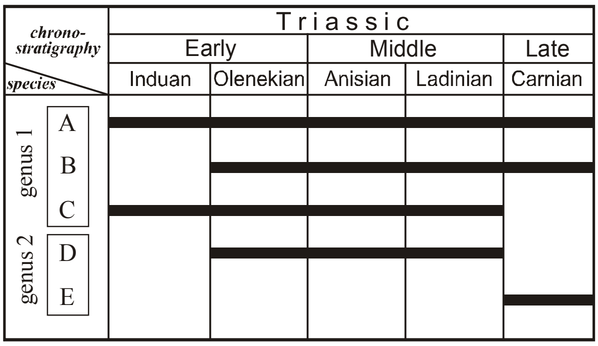

Fig. 1. Example of hypothesized ranges of taxa in the Triassic (see text for explanation)

Fig. 1. *Przykład hipotetycznych zakresów taksonów w triasie (objaśnienia w tekście)*

tigated stratigraphic interval, it is necessary to know how many taxa, absent in the previous stage, are present in this one. Commonly (as in this hypothetical case: Fig. 1) no data on fossils from the underlying strata are available. This implies that it is impossible to calculate the origination rate for the Induan stage, because the origination rate is the ratio between the number of new taxa in a chronostratigraphic unit and the number of new taxa in the immediately older chronostratigraphic unit of the same rank. Moreover, as the rate of change cannot be determined for the Induan, it is not possible to compare this ratio with that of the Olenekian.

In the example of Figure 1, the species A, B, and C belong to genus 1, whereas species D and E belong to genus 2. Genus 2 is represented by species during the Early, Middle and Late Triassic, but in the Carnian the earlier present species (D) was replaced by a new one (E), not present earlier. Data analysis of the Early and Middle Triassic should, obviously, take into account species A, B, C, and D. Species E should, however, also be taken into consideration: a calculation of the generic diversity on the basis of the abovementioned Early-Middle Triassic chart requires also an evaluation of the extinction rate for the last stage, i.e., for the Ladinian. It is therefore necessary to determine whether species that were present in the Early-Middle Triassic survived into the Late Triassic. In addition, it must be checked whether the other species belonging to genera 1 and 2 that did not exist earlier, appeared for the first time in the Carnian. Without such a check, conclusions about the extinction of genus 2 (as in this example) will be incorrect.

This example shows that even well-prepared data may be insufficient for a correct analysis, because lack of data from older and/or younger stages can induce errors. Before starting a quantitative diversity analysis, it is therefore necessary to look for such potential errors and to try to eliminate them.

The Lazarus taxa problem

A specific problem is formed by the so-called Lazarus taxa. The geological record shows frequent interruptions in the chronostratigraphic range of taxa. The re-appearance of a taxon after an interruption is called the Lazarus effect, and taxa showing such interruptions are known as Lazarus taxa. These terms became widely used after the studies by Flessa & Jablonski (1983) and Jablonski (1986). Although the Lazarus effect is linked by some workers to re-appearances after mass extinctions only, it is more logical to consider any re-appearance after interruption in the fossil record as a Lazarus effect (as suggested by Rickards & Wright 2002; and by Ruban & Tyszka 2005), even though the Lazarus effect seems, indeed, most commonly related to mass extinctions (Jablonski 1986, Fara 2001). Several questions related to the Lazarus effect and its influence on the calculation of fossil diversity have been discussed extensively (Flessa & Jablonski 1983; Jablonski 1986; Urbanek 1993, 1998; Senowbari-Daryan & Stanley 1998; Wignall & Benton 1999, 2000; Arz *et al.* 2000; Basov & Kuznetsova 2000; Fara & Benton 2000; Twitchett 2000; Fara 2001; Rickards & Wright 2002; Watkins 2002).

The temporal interruption of the stratigraphic range of a taxon may be due to one of the following causes:

- recurrence, i.e. the appearance of a morphologically similar taxon during evolution (this implies that the interruption is only apparent, as the original taxon is replaced by what is, in fact, another taxon);
- preservation of the taxon in refugia, for instance at times of a global environmental crisis (a widely accepted concept; see, among others, Fara 2001);
- an extreme decrease of the taxon's abundance during a mass extinction, so that the quantity of specimen became too low to trace them in the paleontological record (Wignall & Benton 1999); if the taxon was preserved in refugia, the case is identical to the previous one, but it is also possible that the taxon survived in numerous areas worldwide, but with few individuals;
- temporal disappearance of the taxon from the study area, for instance by migration to the adjacent areas;
- incompleteness of the paleontological record due to hiatuses as a result of non-deposition or erosion, a reduced preserva-

tion of fossils, or errors in sampling (see the review by Fara, 2001);

- taxonomic errors, i.e., incorrect identification of a taxon that appears after the interruption as the same taxon that was present before the interruption (such taxa are also referred to as Elvis taxa: Erwin 2006).

None of the above possibilities should be ignored, and pros and cons for each of these possibilities should be weighted in each particular case. Discussions on how to handle this are still going on (Wignall & Benton 1999, 2000; Rickards & Wright 2000; Twitchett 2000; Fara 2001). Ecological models explaining long survival of rare taxa (Yoshida 2002) support the concept of Wignall & Benton (1999). Meanwhile, refugia seem to play an important role, as suggested by paleoenvironmental studies of both of the geological past and the present (see, among others, Hladil 1994; Hladilova 2000; Riegl & Piller 2003). Possible other explanations for an apparent temporal interruption of the stratigraphic range of a taxon are a reduction in population size (resulting in less individuals that may be found in fossilized form) and a reduction in body size (resulting in less easily found fossil specimens) (Twitchett 2001).

False Lazarus taxa

It is possible to divide the Lazarus taxa into two groups: true and false (Fig. 2). The true Lazarus taxa are those the evolution of which was really interrupted for a particular time interval. In contrast, a false Lazarus effect is obtained if the collected data are incomplete or if taxonomic errors are made while the taxon under study was actually present during all the time span during which its occurrence was only seemingly interrupted. Both survival in refugia and extreme decrease in number are essentially the same in this context: the studied taxa did not really disappear, only no fossil remnants have been found.

A fundamental problem with respect to Lazarus taxa is how to determine whether species or genera before and after the interruption are actually the same, which would imply a true Lazarus effect. If, however, a morphologically similar taxon from before and after the interruption is erroneously considered as the same taxon, this is considered as a 'false Lazarus taxon', and the identification should simply be considered wrong. The fundamental paleontological problem in this context is on what basis species and genera should be distinguished from one another (see also Van Loon 1999; Kucera 2007). This question is closely related to another one: can convergence be so close that it becomes impossible to distinguish between different species (or genera) on the basis of morphology alone? This question is still under much debate, and an answer to this question is badly needed, if only to conclude how the Lazarus effect should be dealt with when determining the fossils' diversity. This is an important key to the systematic paleontology. Fortunately, it seems that application of genetic and other new approaches (such as microarchitectural analyses and investigations at even a molecular scale) may provide solutions for this key problem (see, e.g., Kucera 2007).

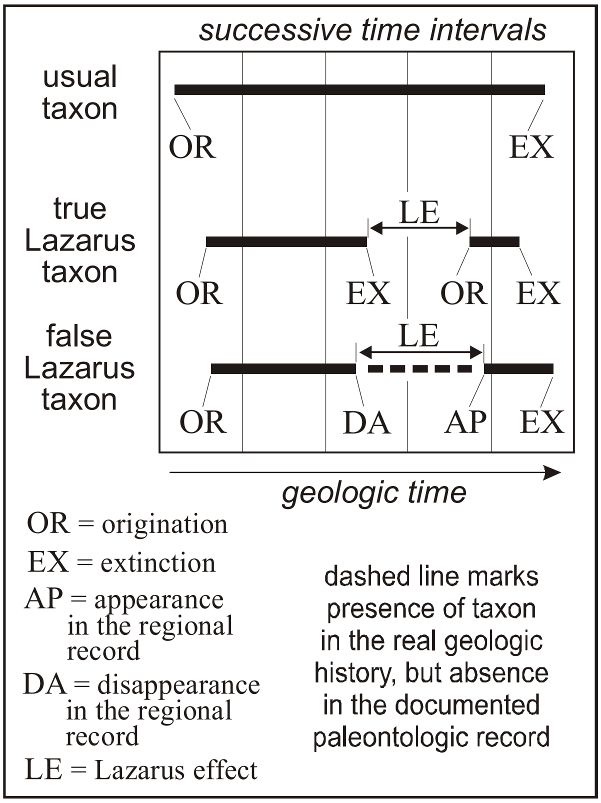

Fig. 2. True and false Lazarus taxa

Fig. 2. *Prawdziwe i fałszywe taksony Łazarza*

Interruption of the stratigraphic range of a taxon increases the (apparent) extinction rate during the interruption interval and thus diminishes the (apparent) total fossil diversity. In contrast, the extinction rate during the next interval will be relatively low, because the ‘extinct’ taxon (re)appears (Fara 2001). A false Lazarus effect thus introduces a difference between the documented and the ‘real’ diversities, and therefore introduces erroneous values for the diversity dynamics. The fossil diversity and the diversity dynamics can therefore be estimated correctly only if the consequences of false Lazarus taxa are taken into account, and if the interruption of true Lazarus taxa is ignored. Dealing with a false Lazarus effect means that it is necessary to analyze the palaebiodiversity during the stratigraphic range as if the occurrence of the pertinent taxon or taxa had not been interrupted.

All Lazarus taxa in the *global* geological record are, following the above-mentioned causes, false ones, except for the case of recurrence, but even in that case one has to consider the possibility that there is only strong morphological resemblance of two different taxa. It seems therefore that the influence of the Lazarus effect cannot be neglected when evaluating global paleobiodiversity. Fara & Benton (2000) and Fara (2001) have indicated how to handle this.

If paleodiversity dynamics are analysed for a relatively small *region*, more complications arise than if a continental or even global analysis is made. The reasons are that (1) the Lazarus effect occurs much more commonly at a small scale than at a large scale (but note that this is true almost exclusively when false Lazarus taxa are involved), and (2) it is more difficult to distinguish between Lazarus taxa that migrated for some time to come back later (for instance as a result of shifting environments due to climate fluctuations), and taxa that are not documented due to an incomplete sedimentary record.

It is, obviously, possible to recalculate the fossil diversity for the possible presence of taxa during the intervals corresponding to their temporal absence in the fossil record. Only the highest probable value (HPV) of the Lazarus effect, which suggests the maximum possible extent of the latter, is thus obtained (Ruban & Tyszka 2005). In other words: the HPV represents the joint effect from both the true and the false Lazarus taxa. When an analysis of regional paleobiodiversity is made, the real diversity must be somewhere between the observed diversity curve and the curve corrected for the HPV (Fig. 3).

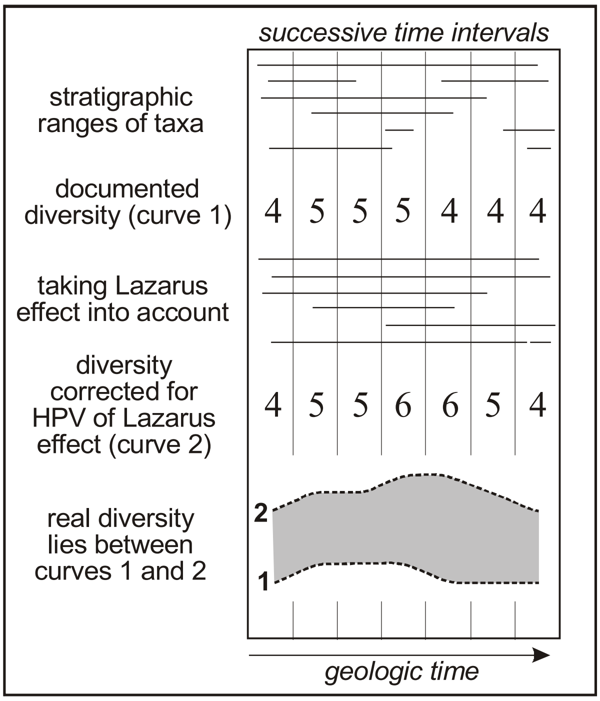

Fig. 3. Highest probable value (HPV) of the Lazarus effect and the estimation of regional fossil diversity.

Fig. 3. *Najbardziej prawdopodobna wartość efektu Łazarza i oszacowanie regionalnego zróżnicowania skamieniałości*

Geochronological problems

A highly important aspect when preparing a quantitative analysis of fossil diversity dynamics is the choice of an appropriate geological time scale. The calculations must be carried out for specific chronostratigraphic units, but the ‘translation’ of lithostratigraphic units into chronostratigraphic units remains a great problem. Recently a great step forwards has been made because the International Commission on Stratigraphy (ICS) has provided an excellent framework for chronostratigraphy (Gradstein *et al.* 2004), so that – in principle – the same geological time table can be used worldwide. This does not imply, however, that is has become easier to ascribe rock units to the correct chronostratigraphic unit. By far most datings of rock units are based on paleontological correlations, but if fossil diversity dynamics are investigated, it would be a vicious circle if the study would be based only on such paleontological data. Just like sedimentary facies shift in space with time, many fossils may show comparable diachronous occurrences. The first or last occurrence of a particular taxon at a certain place, where this occurrence coincides with a chronostratigraphically defined boundary, may therefore have a different age elsewhere in the world. Obviously, areas that are situated far from one another commonly are correlated through a number of intermediate correlations; the resulting inaccuracy of the correlation may be significant (Van Couvering 2000).

An entirely different problem is posed by the fact that the various chronostratigraphic units of one rank (e.g., stages) do not have an equally long duration. A stage that lasts twice as long as the previous one, has, obviously, a great chance to contain more species (and higher-order taxa) than its predecessor. It would be unjustified, however, to deduce that the longer stage is characterized by a higher biodiversity: at any given moment the biodiversity in both stages may be the same (within some band width), but the biodiversity in the longer-lasting stage may, at any given moment, also even be lower than the diversity at any given moment in the shorter stage! Biodiversity and the rate of fossil diversity dynamics should therefore preferably be determined for successive time-spans of approximately equal duration (Van Couvering 2000).

Such an approach may in practice turn out (almost) impossible with our present-day knowledge of correlations. Most areas are still described following a local or regional lithostratigraphic subdivision, and the duration of the time-span during which they were formed can often be estimated only roughly; in many cases attribution to an 'official' chronostratigraphic units is not even certain. This problem has to be solved, however, before reliable diversity dynamics can be determined. This implies (1) that the use of so-called "regional stages", representing a regional time scale, should be avoided whenever possible, and (2) that, more than was done previously, attention should be paid to defining the boundaries between chronostratigraphic units, so that at least a reasonable correlation is established with the recent ICS „International Stratigraphic Chart" (the most recent version is to be found at http://www.stratigraphy.org). Obviously, the decisions and recommendations of the ICS and its subcommissions should be followed. The presentation of data according to the International Stratigraphic Chart is, however, not yet always the case in a few countries (one of them being Russia), which should be regretted deeply as the use of any diverging stratigraphy prevents precise global paleodynamics analyses (or at least makes them more difficult and les reliable).

Geochronology vs. dating

The term 'geochronology' is often considered as a synonym of 'dating' (particularly radiometric dating) or another kind of absolute time estimation, but this is based on misunderstanding (Walsh 2001). It seems, at first sight, that several of the problems sketched above would be solved if all stratigraphic observations were complemented with data about their absolute ages. It is true that some types of fossils evolved so rapidly that they can be considered to represent (geologically) very short time intervals. This is, for instance, the case with Late Carboniferous fusulinids. Only few of such 'time-specific' taxa have been dated precisely, however, by radiometric or other means, but this does not imply that well-dated taxa can always be used as chronostratigraphic markers. Imagine that a marine succession is interrupted by two levels of volcanic ashes that can be dated precisely, and which differ 1,000 years in age. If a species is present in the entire interval between the two ash layers, but neither underneath the lower ash layer, nor above the upper one, this does not imply that the species can be used as a precise time marker: it may well be that the conditions between the deposition of the two ash layers were so different from before and after that the species found a good habitat in the region during the 1,000 'inter-ashes' years, but neither before nor afterwards. Elsewhere, however, the species may have occurred earlier or later. Using the well-dated time range once found for a particular taxon for all subsequent finds, is therefore in most cases without any doubt incorrect.

In addition, radiometric dating is not so absolute as sometimes believed. Much effort has been put by the ICS in establishing absolute ages for the boundaries between chronostratigraphic units (see the 'golden spikes' in the International Stratigraphic Chart), but new absolute ages for the various Devonian stages (Kaufmann 2006) were presented less than two years after the publication of the International Stratigraphic Chart of Gradstein *et al.* (2004). This is not amazing, as dating techniques become ever more refined, as expressed already much earlier by the successive editions of the Elsevier Geological Time Table [compare, for instance, the 4th edition (Haq & Van Eysinga 1987) with the recently published 6th edition (Haq, 2007)], where boundaries were changed sometimes for tens of millions of years. A comparison of the datings for the boundaries within the Mesozoic between the current International Stratigraphic Chart (ICS, 2006) and the 1999 GSA Geologic Time Scale (Palmer & Geissman, 1999) also shows that boundaries shifted in age sometimes more than the time-span of a stage. This means that the inaccuracies in absolute dating are sometimes larger than the duration of the chronological units themselves; it seems that such large changes do not – and will not – occur frequently anymore, but it should be a warning that adaptations of radiometric datings still take place. An example is the boundary (which is most important from a paleodynamics point of view because it is based on the largest mass extinction in the Earth history) between the Permian and the Triassic, which is indicated on the International Stratigraphic Chart (ICS, 2006) as 251.0 (± 0.1) Ma, but which has, shortly after the publication of the 2004 ICS, been found to be 252.6 Ma (± 0.2) (Mundi *et al*. 2004).

Similar conclusions have been drawn for other types of 'absolute datings'. There is, for instance, a gap of several hundreds of years between the varve countings and the C-14 datings for Late Pleistocene and early Holocene glaciolimnic deposits in Scandinavia (e.g. Schove 1977) and for comparable datings elsewhere (e.g. Grayson & Plater 2007). And the necessity to use several types of C-14 dating, is proof in itself that this method has to deal with numerous 'internal' problems (cf. Buck & Bard 2007).

Influence of changing astronomical parameters

Of academic interest only is the fact that paleodiversity dynamics is commonly calculated on the basis of diversity changes that occur in intervals of (usually) millions of years. These 'years' are, however, years with present-day length: ~365 days of 24 hours each. One should realize, however, that these parameters have changed in the course of the geological history. It has been calculated that in the Ediacaran (the end of the Neoproterozoic) a year lasted 444 days and 20.4 hours (Nesterov 1999) with the result that, for instance, the 88 million years that are attributed to the Ediacaran (according to the recent International Stratigraphic Chart it lasted from ~630 to 542 million years ago) lasted only 72 million years according to the astronomical years of the Ediacaran itself. As it is apparent that many organisms reproduce on the basis of yearly cycles, and that the rate of evolution depends (partly) on the velocity of reproduction (one of the reasons why *Drosophila melanogaster* is such a good 'guinea pig' for genetic research), it is obvious that the outcome of paleodiversity-dynamics analyses for Ediacaran (and other old) time intervals will give unduly low values. The change in the duration of a year had been deduced already much earlier, for instance for the Devonian (Wells 1963; Scrutton 1965). More recently, it has been concluded from the frequency modulation analysis of cyclic sedimentary successions that a Jurassic year had a duration intermediate between a Devonian and a recent one (Hinnov & Park 1998), and this confirms that a year on Earth has gradually become shorter. The gradual decrease in the number of days per year was probably compensated – at least in part – by lengthening of the days. This lengthening may, indeed, have had an adverse influence on the reproduction velocity and thus on the value of paleodiversity dynamics, but the net result is still far from clear. It is certain, however, that evolution goes faster with geological time; it even seems likely that more new species are formed per unit of time now than ever before, and probably even at a higher rate than species becoming extinct (Van Loon 2003). In contrast, the paleodiversity dynamics must have been low in the Proterozoic (lack of sexual reproduction will certainly have played a role during this era, but this cannot explain the acceleration of evolution during the Phanerozoic). The reason for the apparently ongoing acceleration of evolution is not well understood, but it means in fact that equal values of paleodiversity dynamics calculated for two time units indicate that the rate of change in the younger unit is *relatively* lower than that in the older unit.

Types of units for which the paleodiversity dynamics can be analysed

Paleodiversity dynamics can be determined for five types of units:

- non-diachronous lithostratigraphic units (results will be enforced by paleoenvironmental specifics);
- stages or epochs (results will be influenced by the uncertainties regarding the current chronostratigraphy);
- biozones (appropriate for particular fossil groups only);
- beds (appropriate only for the analysis of a given section; this is a common procedure particularly for microfossils);
- millions of years (not truly meaningful for fossils; moreover, fossils can rarely be dated within geologically restricted time boundaries).

It seems to us that events that affected biodiversity significantly, such as mass extinctions and sudden faunal explosions, can be used to establish time units that comprise the same number of biodiversity-affecting events (this extends the purpose of event stratigraphy as explained, particularly, by Walliser (1996) and Brett & Baird (1997). A possible alternative ecological approach in this context has been introduced by McGhee *et al.* (2004).

Conclusions

The above-mentioned problems concerning the quantitative analysis of paleodiversity dynamics (for both animals and plants) are presented in Fig. 4. All problems can commonly be solved, though often only in a time-consuming way. It is not uncommon, however, that the problems can be overcome only partly, because of insufficient, ambiguous and/or insufficiently accurate data. In some cases the lack or inaccuracy of information may be even so significant that a detailed paleodiversity-dynamics analysis is not worthwhile; in other cases the problems can be minimized or taken care of in one way or another.

Fortunately, ongoing work in this field, supported by large electronic databases (e.g. NMITA, PaleoTax, The Paleobiology Database,

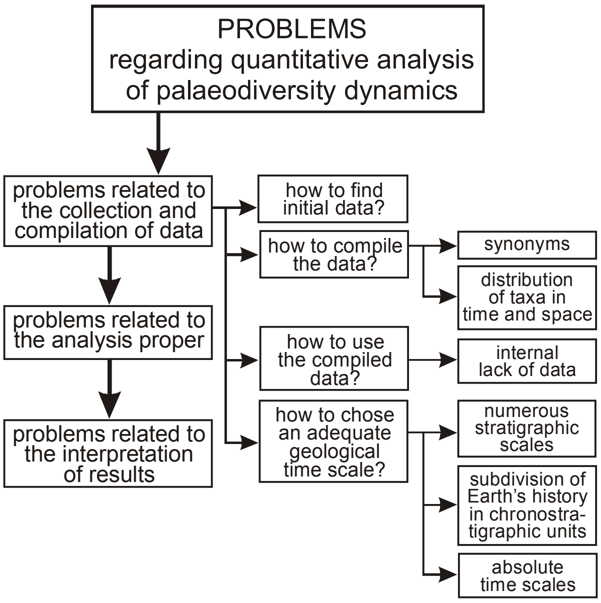

Fig. 4. Schematic overview of the problems related to the quantitative analysis of changes in paleobiodiversity

Fig. 4. *Schematyczne podsumowanie problemów związanych z ilościową analizą zmian w paleo-bioróżnorodności*

MIOMAP, FAUNMAP, Global Pollen Database, NOW, etc.) (Benton 1995; Budd *et al*. 2001; Alroy 2003; Löser 2004; Foote 2007), may help to diminish possible errors in the diversity estimates. Each analysis of paleodiversity dynamics or of phenomena related to the consequences of mass extinctions should, however, (1) try to quantify the impact that missing data or inaccuracies of any kind may have on the final results, and (2) try to find a solution for the major problems, so as to avoid significant inaccuracies of the calculated values.

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Bryozoan diversity in Southern Siberia at the Devonian–Carboniferous transition: New data confirm a resistivity to two mass extinctions

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ABSTRACT

Two spectacular mass extinctions, namely the Kellwasser and the Hangenberg events, affected the Earth's biota during the Late Devonian–Early Carboniferous. Diverse assemblages of bryozoans are known from the Fasnian–Tournaisian deposits of Southern Siberia, which includes such regions as the Kuznetsk Basin, the Kolyvan'–Tom' Zone, the Rudny Altay, and the Gorny Altay. Our new data establish the stratigraphic ranges of 154 species, which belong to 56 genera and 21 families. These data have been analyzed quantitatively to measure the patterns of diversity dynamics and the changes in the taxonomic diversity structure. The species diversity increased at the Frasnian–Famennian transition. After a short-term, but dramatic decrease in the middle Famennian, the total number of species accelerated in the late Famennian and the Tournaisian. A similar pattern occurred for both genera and families, although the radiation at the Frasnian–Famennian transition was not as large, and total family diversity decreased slightly in the Tournaisian. Changes in total species diversity differed between orders of bryozoans. Changes in taxonomic diversity structure were moderately rapid. The Tournaisian species diversity was governed by the other genera comparing to the early time intervals. Thus, our regional data suggest that bryozoans generally survived both the Frasnian/Famennian and the Devonian/Carboniferous mass extinctions, confirming similar conclusions made earlier with global data. However, a mid-Famennian regional crisis (diversity drop) in the evolution of bryozoans is documented within Southern Siberia. Bryozoan radiations coincided with pulses of basin deepening that occurred during both mass extinction intervals, perhaps explaining why bryozoans were resistant to extinction. The mid-Famennian crisis might have been caused by a rapid regression coupled with basin shallowing.

1. Introduction

The marine biota was very diverse during the Devonian (Sepkoski, 1993; Benton, 2001; Newman, 2001; Peters and Foote, 2001). However, the end of this period was marked by two catastrophes that devastated the Earth's life, namely the Frasnian/Famennian (F/F) and the Devonian/Carboniferous (D/C) mass extinctions often referred as the Kellwasser and the Hangenberg events, respectively. Using the present absolute chronology for the Devonian (Gradstein et al., 2004; Kaufmann, 2006; Menning et al., 2006), the D/C mass extinction persisted for approximately 16 Ma after the conclusion of the F/F mass extinction. A number of possible triggers for these extinctions have been proposed. It is not possible, currently, to determine which, if any, is the likely cause of the extinctions (see review by Racki, 2005). These events might have been linked with major eustatic changes, marine anoxia, global cooling, intense volcanism, and probable impacts (Copper, 1986; Buggisch, 1991; Copper, 1994; Algeo et al., 1995; McGhee, 1996; Hallam and Wignall, 1997; Algeo and Scheckler, 1998; Hallam and Wignall, 1999; Caplan and Bustin, 1999; Racki, 1999; McGhee, 2001; Copper, 2002; Gong et al., 2002; House, 2002; Godderis and Joachimski, 2004; Racki, 2005; McGhee, 2005; Girard and Renaud, 2007).

Were any groups of organisms resistant to inferred environmental perturbations at the end of the Devonian? It appears that the answer is yes. Hallam and Wignall (1997) and later Webster et al. (1998) discussed the effect of the F/F and D/C mass extinctions on the planetary biota and found that bryozoans were not stressed severely by these events. Horowitz and Pachut (1993) and then Horowitz et al. (1996) reported a significant crisis among bryozoans in the mid-Devonian, while the F/F event failed to produce dramatic reductions in diversity. This is confirmed by Horowitz and Pachut (2000), whose data also suggest against any major diversity decline after the D/C event. We are still far from a complete understanding of the causes and selectivity of mass extinctions (Jablonski, 2004). However, an

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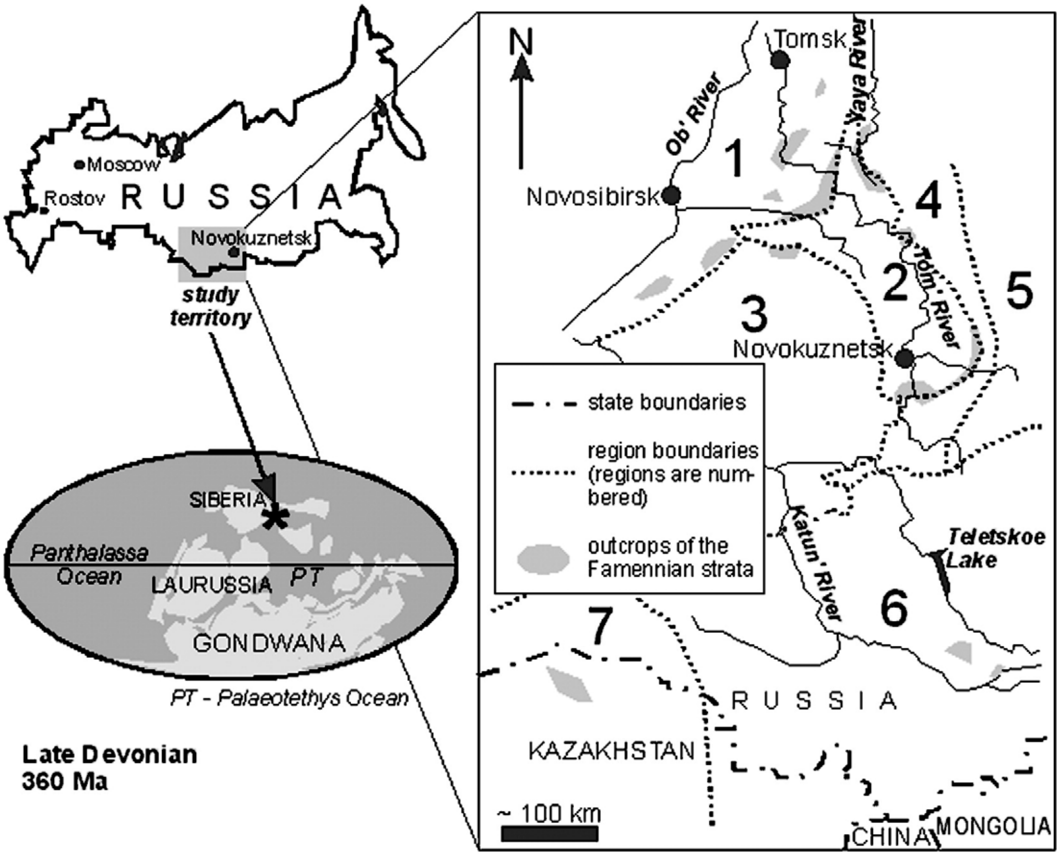

Fig. 1. Geographical and palaeotectonic location of the study territory. Global palaeotectonic map is simplified after Scotese (2004). Numbered regions: 1 — Tom'–Kolyvan' Zone, 2 — Kuznetsk Basin, 3 — Salair, 4 — Kuznetsk Alatau, 5 — Minusa Depression, 6 — Gorny Altay, 7 — Rudny Altay.

evaluation of the evolutionary dynamics of bryozoans from the Late Devonian–Early Carboniferous might be informative. Data are now available from Southern Siberia (Tolokonnikova, 2007) to attempt such a study. This region lies within the core of the present-day Eurasia (Fig. 1). Data include the stratigraphic ranges of 154 species, 56 genera, and 21 families of bryozoans from upper Frasnian, Famennian, and lower Tournaisian strata, making Southern Siberia a very important region to evaluate bryozoan diversity at the Devonian–Carboniferous transition and to discuss the effects of mass extinction on them.

2. Geological setting

The study territory, generally referred as Southern Siberia, includes the Kuznetsk Basin (also called as the Kuzbass), the Kolyvan'–Tom' Zone, the Rudny Altay (the Ore Altay), and the Gorny Altay (the Mountaneous Altay) (Fig. 1). Using present global reconstructions for the Late Devonian and Early Carboniferous (Torsvik and Cocks, 2004; Scotese, 2004) and few regional palaeotectonical constraints (Yakubchuk, 2004; Cocks and Torsvik, 2007), it appears that these regions were attached to the margin of the Siberian Plate or were individual terranes (Fig. 1). The growth of present-day Southern Siberia appears to have been a stepwise process that occurred during the entire Paleozoic and even the Early Mesozoic.

Frasnian, Famennian, and Tournaisian deposits occur in many areas of Southern Siberia (Fig. 1). Their lithostratigraphic framework was established by Krasnov (1982) and was updated by Tolokonnikova (2007). These strata are subdivided into a number of horizons, based on chronostratigraphy and conodont zonation (Gutak and Rodygin, 2004; Tolokonnikova, 2007) (Fig. 2). Because the meaning of the term*horizon*, as it is used in international stratigraphic nomenclature (Salvador, 1994), differs from typical usage, we use the term*unit*. The lowest of the units considered in our study is the Solominski Unit (upper Frasnian), that includes limestones and clastics with a total thickness of up to 800 m. They are overlain by the limestones, clastics, and in places, by volcaniclastics of the Petscherkinski Unit (lower Famennian) with a total thickness up to 800 m. The Podoninski Unit (middle Famennian) consists of red-colored clastic deposits, with occasional limestones and volcaniclastics, with a total thickness exceeding 1600 m (Fig. 2). The Topkinski Unit (upper Famennian) includes clastics and limestones that may reach a maximum total thickness of 500 m. The Krutovski Unit (lower Tournaisian) is dominated by volcanics and volcaniclastics with a total thickness of about 100–1500 m. The uppermost Tajdonski Unit (lower Tournaisian) is represented by limestones and clastics that have a total thickness of approximately 1100 m.

Palaeoenvironmental reconstructions (Bel'skaja, 1960; Gutak and Antonova, 2006a,b; Gutak and Ruban, 2007; Tolokonnikova, 2007) for the Late Devonian–earliest Carboniferous indicate that Southern Siberia was embraced by a large marine basin. The Kolyvan'–Tom' Zone represents the deepest facies deposited in that basin. The Kuznetsk Basin was occupied by extensive, relatively shallow-water, shelf environments. The Gorny Altay lay on a margin of this basin and, together with the Rudny Altay, was a center of regional volcanic activity. The regional Frasnian–Tournaisian transgression–regression (T–R) curve of Southern Siberia was reconstructed from lateral facies analysis, whereas the deepening–shallowing (D–S) curve was developed from analyzing facies stratigraphically. These were derived from Gutak and Ruban (2007). Our T–R curve differs from that proposed earlier by Karaulov and Gretschischnikova (1997) and Yolkin et al. (1997) in the positioning of the D/C boundary and in proposing an alternative interpretation of facies (Gutak and Antonova, 2006a,b; Gutak and Ruban, 2007; Tolokonnikova, 2007). T–R and D–S patterns should always be distinguished (Catuneanu, 2006; Ruban, 2007a).

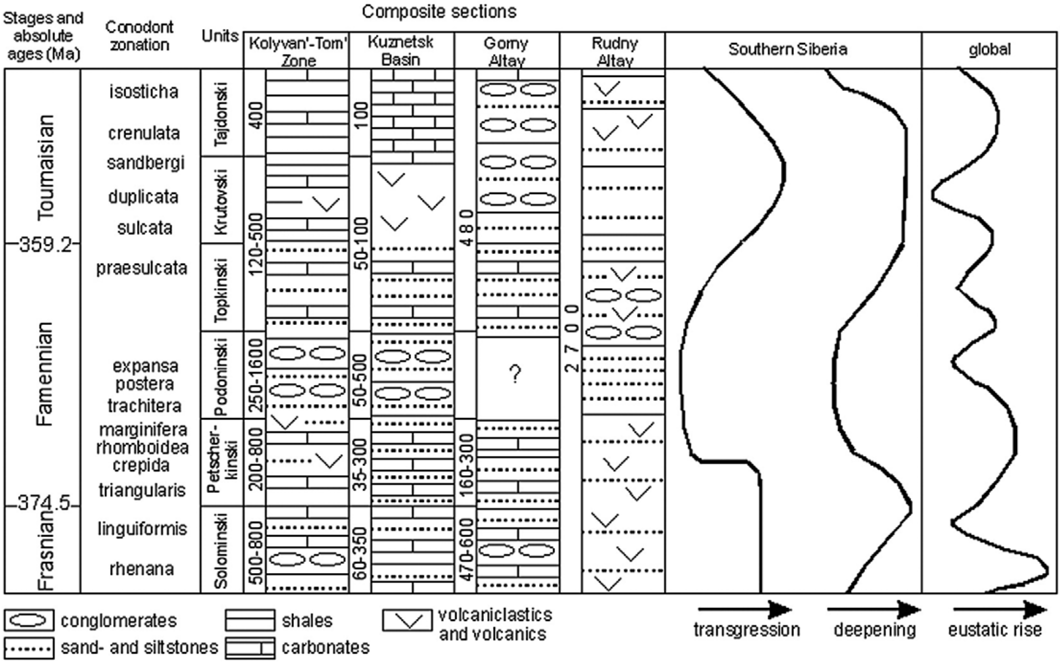

Fig. 2. The Frasnian–Tournaisian composite sections of the studied regions. Chronostratigraphy after Gradstein et al. (2004). Thickness (m) is indicated from the left side of each column. Regional transgressive–regressive and deepening–shallowing patterns are modified after Gutak and Ruban *2007) and the global eustatic changes are given after Haq and Al-Qahtani (2005).

Transgressions and regressions reflect the shifting position of shorelines, whereas deepenings and shallowings mark the changes in the depth of the basin. In Southern Siberia, the shoreline was stable during the Frasnian, and the F/F boundary is marked by a prominent, but short-term, deepening pulse (Fig. 2). A wide and abrupt regression and shallowing occurred in the early–middle Famennian. Marine deposits of this age have a very limited extent, whereas non-marine red-colored clastic deposits are known widely within the study region (Fig. 2). Both transgression and deepening occurred after the late Famennian reaching maximums in the Tournaisian when carbonates, often cherty, dominated regional deposition (Fig. 2). These regional patterns did not correspond well to the global eustatic fluctuations recorded recently by Haq and Al-Qahtani (2005). It appears to be nearly impossible to recognize the action of any planetary-scale events corresponding to their sea-level curve (Fig. 2).

3. Materials and methods

This study is based on data on the Frasnian–Tournaisian bryozoans collected in Southern Siberia by Tolokonnikova (2007). Data from a number of earlier studies (Nekhoroshev, 1926a,b; Krasnopeeva, 1935; Nekhoroshev, 1956; Trizna, 1958; Morozova, 1961; Volkova, 1974; Morozova, 2001) have also been reexamined including the distribution of each taxon (Solominski, Petscherkinski, Podoninski, Topkinski, and Tajdonski units; Supplementary Materials 1). No bryozoans have been found in the Krutovski Unit. The suprageneric taxonomy of the bryozoans examined in this study is presented in Supplementary Materials 2.

For species, genera, and families, we have examined changes in total diversity, number of appearances, and number of disappearances. The absence of bryozoans in the volcanics of the Krutovski Unit is expected (Tolokonnikova, 2007). Thus, to calculate the true number of appearances in the Tajdonski Unit and the number of disappearances for the Topkinski Unit, we have ignored the Krutovski Unit and compared the assemblages directly.

Regional evolutionary changes among bryozoans were measured using the Rst index. This index, and its interpretation, were presented by Ruban and Tyszka (2005), and Ruban (2007b). Rst is a coefficient of the Spearman's rank correlation (Kendall, 1970) between two assemblages, where the presence of a taxon of higher taxonomic rank is indicated by the number of taxa of lower rank, which belong to that higher-ranked taxon in the particular time unit. 1/Rst or 1-Rst shows, how rapid were the changes in the control of the lower-ranked taxa diversity by the higher-ranked taxa. In other words, Rst indicates the changes in the structure of taxonomic diversity. It is intriguing to calculate Rst not only for the successive assemblages, but also for those, existing in different times. In this paper, we have calculated three Rst indices, namely the Genera-species Rst, the Families-genera Rst, and the Families-species Rst. This permits an evaluation of three patterns of regional change in the taxonomic diversity structure of bryozoans and provides a detailed picture of transformations across both the F/F and D/C mass extinctions.

4. Diversity dynamics

In total, 154 species of bryozoans are known from the upper Frasnian, the Famennian, and the lower Tournaisian strata of Southern Siberia. They belong to 56 genera, 21 families, and 6 orders. The most diverse were the orders Trepostomida (8 families, 22 genera, 53 species), Fenestellida (4 families, 16 genera, 54 species), and Rhabdomesida (4 families, 11 genera, 28 species). The Order Cystoporida was characterized by moderate diversity (3 families, 5 genera, 16 species), whereas the diversity of orders Cryptostomida (1 family, 1 genus, 2 species) and Tubuliporida (1 family, 1 genus, 1 species) was very low.

Species diversity nearly doubled at the Frasnian–Famennian (F–F) transition (increased 1.8 times; Fig. 3A) then decreased abruptly. In the late Famennian, bryozoans diversified again and the total number of species continued to rise into the early Tournaisian. Generic diversity rose slightly at the F–F transition (Fig. 3B) followed by an abrupt decline. The total number of genera was high in the late Famennian. A slight increase of 2 genera occurred in the early Tournaisian. It is intriguing that the number of generic appearances dropped dramatically in the early Tournaisian. Two new families appeared at the Frasnian–Famennian transition (Fig. 3C). After an abrupt drop in the mid-Famennian, the total number of families rose again in the late Famennian. A decrease of 2 families occurred at the Famennian–

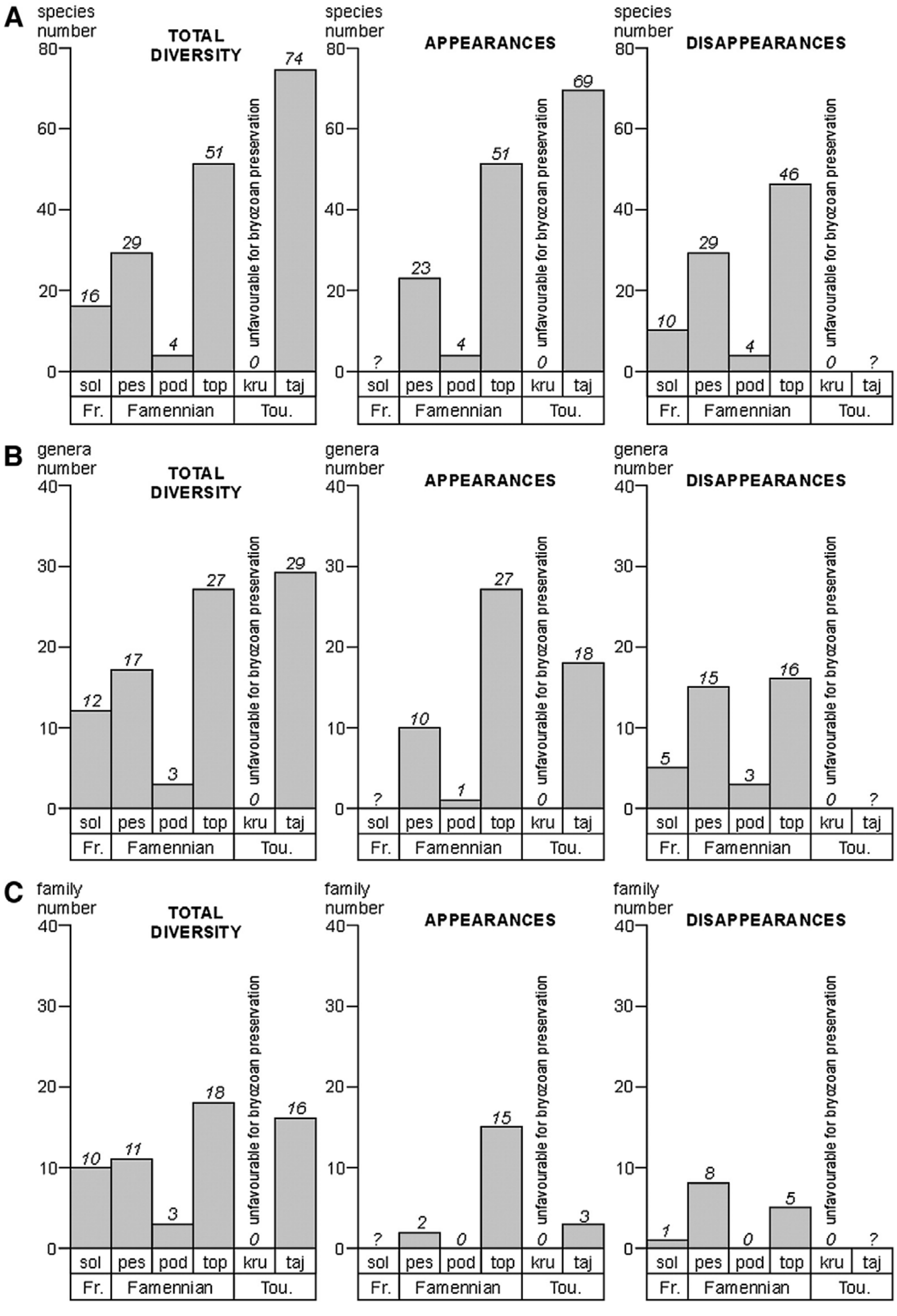

Fig. 3. Species (A), genera (B) and family (C) diversity dynamics of the Frasnian–Tournaisian bryozoans of Southern Siberia. To calculate the numbers of appearances and disappearances the Topkinski Unit and the Tajdonski Unit are compared directly. Abbreviations: Fr. — Frasnian Stage, Tou. — Tournaisian Stage, sol — Solominski Unit, pes — Petscherkinski Unit, pod — Podoninski Unit, top — Topkinski Unit, kru — Krutovski Unit, taj — Tajdonski Unit.

Tournaisian transition (F–T). A comparison of total diversity changes documented at three taxonomic levels suggests differences especially at the F–F and F–T transitions. While species diversity rose significantly at the F–F transition (Fig. 3A), generic and familial diversity increased only slightly (Fig. 3B,C). At the F–T transition, the total numbers of species and genera increased (Fig. 3A,B), whereas the number of families declined slightly (Fig. 3C).

There is no direct evidence supporting either the F–F or the D–C mass extinctions among bryozoans in the regional record of Southern Siberia. There was no major diversity decline at the F–F transition, and only a slight reduction in family diversity at the F–T transition, confirming the conclusions of Horowitz et al. (1996), Hallam and Wignall (1997), and Webster et al. (1998). Updated data on global bryozoan diversity (Horowitz and Pachut, 2000) imply an existence 195 Frasnian species, 153 Famennian species, and 405 Tournaisian species. Thus, the total species diversity declined after the F–F event just in 1.3 times. This is too low in comparison with the Givetian–Frasnian transition, when the total diversity decreased in 3.2 times (Horowitz and Pachut, 2000). As for the D–

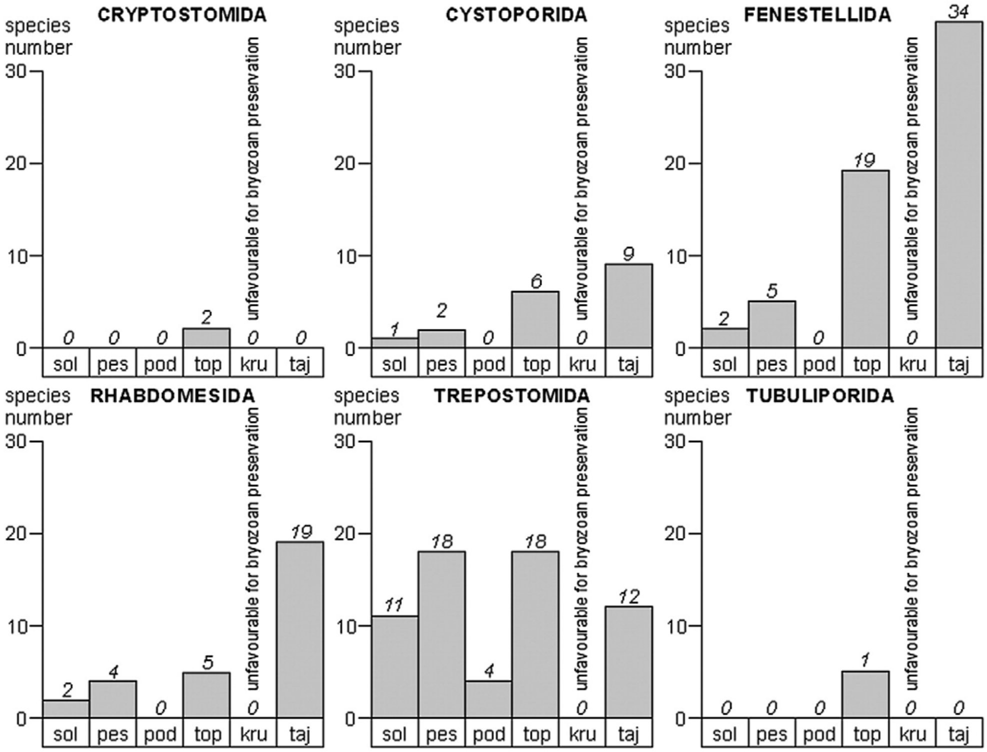

Fig. 4. Changes in the total species diversity among six orders of the Frasnian–Tournaisian bryozoans of Southern Siberia. Abbreviations — see Fig. 3.

C mass extinction, it did not affect either global or regional species diversity at all. On a global scale, the total diversity increased in 2.6 times (Horowitz and Pachut, 2000). The similar acceleration is documented in Southern Siberia.

As potential, but "minor" effect of the F–F and D–C mass extinctions the rate of disappearance of species, genera, and families at the beginning and end of the Famennian increased (Fig. 3A–C). Additionally, the number of appearances of genera and families decreased in the early Tournaisian (Fig. 3B,C). A unique crisis, documented regionally within the studied stratigraphic interval, occurred in the mid-Famennian (the Podoninski Unit) when a short-term but abrupt diversity decline occurred. However, in spite of their low diversity bryozoans are quite abundant in the Podoninski Unit.

The species diversity dynamics of the Cystoporida, Fenestellida, and Rhabdomesida were similar (Fig. 4), with diversity rising throughout the late Frasnian–early Tournaisian. In contrast, the species-level diversity dynamics of the Trepostomida was quite different (Fig. 4). Total species diversity was similar in the early and late Famennian, but declined rapidly in the early Tournaisian. Surprisingly, the Trepostomida was unique in surviving at the time of the mid-Famennian crisis. Both Cryptostomida and Tubuliporida radiated in the late Famennian, but they did not survive into the Carboniferous (Fig. 4). Thus, no orders were removed by the F/F mass extinction, whereas the D/C mass extinction might have been responsible for the disappearance of Cryptostomida and Tubuliporida and the decline of the Trepostomida. A comparison of species diversity (Fig. 3A) with that of orders (Fig. 4) suggests that the decline of Trepostomida did not affect the stepwise increase in bryozoan diversity.

5. Changes in taxonomic diversity structure

Changes in the taxonomic diversity structure of the Frasnian–Tournaisian bryozoans of Southern Siberia are presented in Tables 1–3. The Genera-species Rst suggests relatively slow changes across the F–F transition, whereas the changes that followed were much more rapid (Table 1). This suggests that control of species diversity shifted to new genera in each successive time unit. Rst for the F–T transition also increased. A comparison of the non-successive assemblages suggests 3 important conclusions:

(1) genera responsible for species diversity in the late Frasnian, continued to control species diversity in the late Famennian (Rst for the Solominski and the Topkinski units is as high as 0.24);

(2) no remarkable changes in the taxonomic diversity structure occurred in the mid-Famennian based on an Rst for the Petscherkinski and Topkinski units of 0.48;

(3) changes in species diversity occurred in the Tajdonski Unit at the Devonian–Carboniferous transition that was controlled by genera other than those that were responsible for species diversity at earlier times.

Table 1
Genera-species Rst for the Frasnian–Tournaisian bryozoans of Southern Siberia

	sol	pes	pod	top	taj
sol	1.00	0.54	−0.01	0.24	−0.15
pes		1.00	0.04	0.48	−0.01
pod			1.00	−0.15	−0.12
top				1.00	0.27
taj					1.00

Abbreviations — see Fig. 4

Table 2
Families-genera Rst for the Frasnian–Tournaisian bryozoans of Southern Siberia

	sol	pes	pod	top	taj
sol	1.00	0.81	0.47	0.39	0.22
pes		1.00	0.48	0.37	0.47
pod			1.00	0.15	−0.20
top				1.00	0.65
taj					1.00

Abbreviations — see Fig. 4

Table 3
Families-species Rst for the Frasnian–Tournaisian bryozoans of Southern Siberia

	sol	pes	pod	top	taj
sol	1.00	0.82	0.43	0.13	−0.11
pes		1.00	0.11	0.51	0.29
pod			1.00	−0.13	−0.21
top				1.00	0.68
taj					1.00

Abbreviations — see Fig. 4

The Families-genera Rst indicates some additional peculiarities (Table 2). Changes in diversity structure were minor at the F–F and the F–T transitions but were higher during the middle- and late Famennian. Rst values between the non-successive assemblages were also high, suggesting gradual changes in family control of generic diversity. Similar patterns occur in the Families-species Rst (Table 3). An exception is relatively high Rst calculated for the Petscherkinski and Tajdonski units.

Changes in diversity structure of late-Frasnian–early Tournaisian bryozoans of Southern Siberia suggest moderately rapid evolution. A number of higher-ranked taxa, that controlled the diversity of lower-ranked taxa in the late Frasnian and/or early Famennian, were similarly important in late Famennian and/or early Tournaisian assemblages. No catastrophic patterns are documented, except those established at the Devonian–Carboniferous transition reflected by the magnitude of the Genera-species Rst. Thus, only the effect of the D/C mass extinction can be recognized in the regional record of bryozoans although this effect was weak.

6. Discussion

Transgressions/regressions and deepenings/shallowings might have been important factors, influencing regional diversity changes among bryozoans. A comparison of the reconstructed T–R and D–S curves (Fig. 2) with changes in the total species diversity of bryozoans (Fig. 3A) suggests a better correlation with the regional D–S pattern. Diversity rose as basin depth increased at the F/F boundary. Similarly, late Famennian–middle Tournaisian deepening coincided with rapid bryozoan diversification. In contrast, it appears that bryozoans were able to radiate at times of both a weak regression (early Famennian) and strong transgression (late Famennian–early Tournaisian). Regression and shallowing in the mid-Famennian coincided with a remarkable crisis (diversity drop) among bryozoans.

Racki (2005) suggested that a controversy existed in the interpretation of global sea-level changes across the F–F transition. Hallam and Wignall (1997, 1999) had suggested a eustatic rise, that was later confirmed by Haq and Al-Qahtani (2005). The end-Devonian experienced a global sea-level fall according to Hallam and Wignall (1997, 1999) and Racki (2005), while Haq and Al-Qahtani (2005) suggested the opposite (i.e., an eustatic peak, although preceded by a “minor” fall) (Fig. 2). If bryozoans really diversified as deepening occurred as suggested from Southern Siberia, the global eustatic rises documented by Haq and Al-Qahtani (2005) may explain why this group was little affected by the F/F and D/C mass extinctions on a global scale.

7. Conclusions

The Frasnian–Tournasian bryozoans were diverse in Southern Siberia. Our calculations suggest that both the F/F and D/C mass extinctions did not affect diversity dynamics. Species-, genus-, and family-level diversity did not decline, and no large changes in the taxonomic diversity structure were observed, except for few minor exceptions. Thus, the resistivity of bryozoans to the Late Devonian environmental crises (or the selectivity of the event) is confirmed for a representative regional record from Southern Siberia. Currently, it appears that regional changes in basin depth might have affected bryozoan diversity, causing it to increase as deepening occurred. An enigmatic mid-Famennian crisis might have been caused by rapid regression and basin shallowing. One question awaiting further investigation is whether the mid-Famennian crisis was regional or global in scale. The global data (Horowitz and Pachut, 2000) indicate just a slight general diversity decline in the Famennian. A diversity drop in the middle of this age might contribute to such a decline. To understand why bryozoans survived both the F/F and D/C extinctions on a global scale depends strongly on the causes of the catastrophes. If diversity increased, at least in part, thanks to the eustatic rises, a preference for deeper basins by bryozoans might adequately explain their survival. This is another subject for further investigations.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.palaeo.2008.04.009.

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How global are the Jurassic–Cretaceous unconformities?

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ABSTRACT

The reality of the global-scale sedimentation breaks remains controversial. A compilation of data on the Jurassic–Cretaceous unconformities in a number of regions with different tectonic settings and character of sedimentation, where new or updated stratigraphic frameworks are established, permits their correlation. Unconformities from three large reference regions, including North America, the Gulf of Mexico, and Western Europe, were also considered. The unconformities, which encompass the Jurassic-Cretaceous, the Lower–Upper Cretaceous and the Cretaceous–Palaeogene transitions are of global extent. Other remarkable unconformities traced within many regions at the base of the Jurassic and at the Santonian–Campanian transition are not known from reference regions. A correlation of the Jurassic–Cretaceous global-scale sedimentation breaks and eustatic curves is quite uncertain. Therefore, definition of global sequences will not be possible until eustatic changes are clarified. Activity of mantle plumes is among the likely causes of the documented unconformities.

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Introduction

Phanerozoic unconformity-bounded sequences defined in North America by Sloss (1963) have led to the development of global cycle charts and a possible eustatic curve (Vail *et al.*, 1977; Haq *et al.*, 1987). An updated curve for the entire Phanerozoic has been recently proposed by Haq and Al-Qahtani (2005). However, any eustatic constraints should always be tested to avoid errors and inconsistencies (Miall, 1992; Hallam, 2001; Catuneanu, 2006). Thus, it is sensible to return to the original concept of Sloss (1963) and to attempt broad correlation of the unconformities across the world. Interregional correlations like those previously performed by Soares *et al*. (1978), Petters (1979), Ross and Ross (1985), Embry (1997) and Hallam (2001) suggest the efficacy of such an approach.

During the past decade, new stratigraphic frameworks have been established for the Jurassic and Cretaceous successions of a number of important sedimentary basins of Eurasia, Africa, and America. These provide enough data to substantiate global-scale sedimentation breaks during this time interval.

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Geological setting

To be global, an unconformity should be present not only in many regions in the world, but it should also be traceable in far-located regions with various tectonic settings and sedimentation types. A number of regions worldwide with the newly-established Jurassic–Cretaceous stratigraphic frameworks have been selected in order to correlate the Jurassic–Cretaceous unconformities (Fig. 1). Their tectonic settings differed strongly, which helps to avoid recognition of unconformities that result from specific tectonic events. Most of the above-mentioned basins were dominated by marine sedimentation, except the non-marine basins of Africa. In the Neuquén Basin, continental sedimentation since the Late Cretaceous has been established (Howell *et al.*, 2005). This permits us to outline the nearly-global unconformities traceable within a broad spectrum of depositional settings. Good correspondence in stratigraphic architecture between marine and non-marine strata is not so unusual (Catuneanu, 2006; Ruban *et al.*, in press). In all the studied regions, the lithological composition of the Jurassic–Cretaceous strata is quite diverse and the total thickness exceeds several hundreds of metres.

Available stratigraphic frameworks for the studied regions allow delineation of the hiatuses, which are established within at least the main part of each region. These hiatuses mark regional unconformities. An interregional correlation then becomes possible. We do not omit regional unconformities with probable tectonic origin, because an interregional correlation is itself important for understanding the nature of hiatuses.

Correlation of unconformities

Numerous unconformities are recognized, but not one unconformity is identified within all studied regions (Fig. 2). However, five nearly-global unconformities are common for at least 2/3 of the studied regions. They characterize the base of the Jurassic (T–J), the Tithonian–lower Valangianian interval (J–K), the Albian–Cenomanian (K1–K2), the Santonian–Campanian (S–C), and the Maastrichtian–Danian (K–T) transitions.

A striking feature of all the above-mentioned unconformities is their strong diachroneity. This is especially significant for the T–J and the J–K transitional intervals. They cannot be recognized by any unique surface, but only by a concentration of regional hiatuses. Three Cretaceous unconformities seem to be less diachronous. This diachroneity may have at least two possible causes, namely (1) errors in the dating of the unconformity in given regions, (2) tectonic influence. In the first case, a diachrony can be proclaimed as an artefact of the

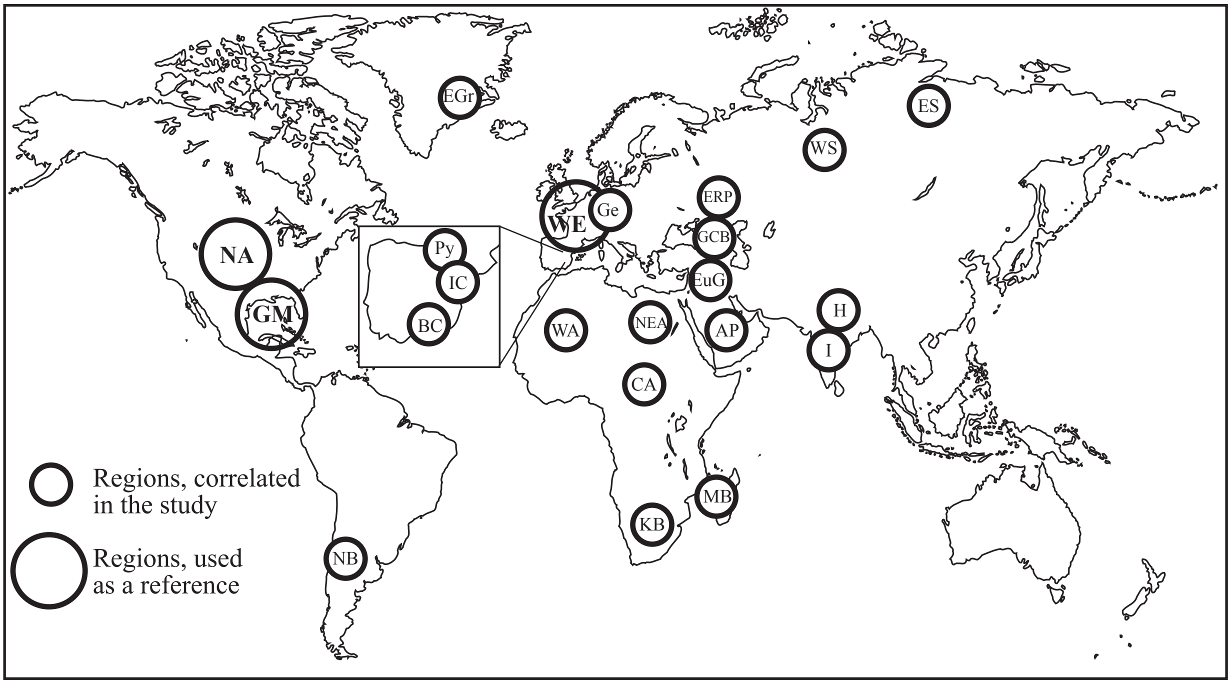

Fig. 1 Location of regions, considered in this study. ERP – eastern Russian Platform, Ge – Germany, Py – Pyrenees-Basque-Cantabrian domain, IC – Iberian Cordillera, BC – Betic Cordillera, EGr – East Greenland, WS – West Siberia, ES – East Siberia, EuG – Euphrates Graben, AP – Arabian Plate, WA – Western Africa, NEA – Northeastern Africa, CA – Central Africa, KB – Karoo basins, MB – Morondava Basin, GCB – Greater Caucasus Basin, NB – Neuquén Basin, H - Himalayas and adjacent blocks, I – India, WE – Western Europe, GM – Gulf of Mexico, NA – North America.

stratigraphic analysis. In contrast, tectonic influences are able to create a true diachroneity. It is possible to hypothesize two kinds of tectonic influences. Any given unconformity may have resulted from the global-scale tectonic pulse. If so, diachroneity of such an activity resulted in diachroneity of the unconformities. Alternatively, an unconformity is resulted from a global eustatic fall. But the age of the unconformity corresponds to the time of such a fall only in those regions that were tectonically stable. Tectonic uplift causes the unconformity to occur earlier, whereas subsidence causes a later unconformity.

It is sensible to correlate the identified unconformities with those established earlier for other regions. We concentrated our attention to three such regions. They are Western Europe, where the principal unconformities have been used to outline the major cycles of sedimentation (Jacquin and de Graciansky, 1998), North America, where some key unconformities were used as major sequence boundaries (Sloss, 1963, 1988) and the Gulf of Mexico, where Salvador (1991) identified a number of extensive unconformities. We observe that the J–K, the K1–K2, and the K–T unconformities are identified in these reference regions (Fig. 3). However, only the oldest is established in all three regions. Note that two unconformities relevant to the latter in the Gulf of Mexico, are not considered by Salvador (1991) among major. Intriguing are the T–J and the S–C unconformities. Their global extent is evident from our correlation (Fig. 2), but it is difficult to identify them in three reference regions (Fig. 3). However, one should take into consideration the absence of pre-Late Jurassic record in the review of data from the Gulf of Mexico by Salvador (1991) and the presence of the Early Cimmerian unconformity at the Norian/Rhaetian boundary.

Sedimentation breaks and eustatic curves

As the five above-mentioned sedimentation breaks are known from many regions and can be labelled as *potentially*-global, they might have been caused by eustatic drops. We use the present Phanerozoic curve of Haq and Al-Qahtani (2005), the curve of Hallam (1988) and Hallam (2001) for the Jurassic, and the curve by Miller *et al.* (2005) for the Late Cretaceous to test this hypothesis (Fig. 3). The first two curves are based on the global compilation of data. However, Hallam (2001) pointed out that the earlier constraints by Haq *et al.* (1987) were based on information from the North Sea and some European sections. The curve reconstructed by Miller *et al.* (2005) is based on data from the New Jersey margin, although compared with those from other regions.

The T–J, the J–K and the K–T sedimentation breaks corresponded to the eustatic lowstands depicted by the curve of Haq and Al-Qahtani (2005). In contrast, the K1–K2 break coincided with a remarkable global sea-level rise. The situation at the S–C transition is unclear, although some eustatic drops are known from there. The other curve (Miller *et al.*, 2005)

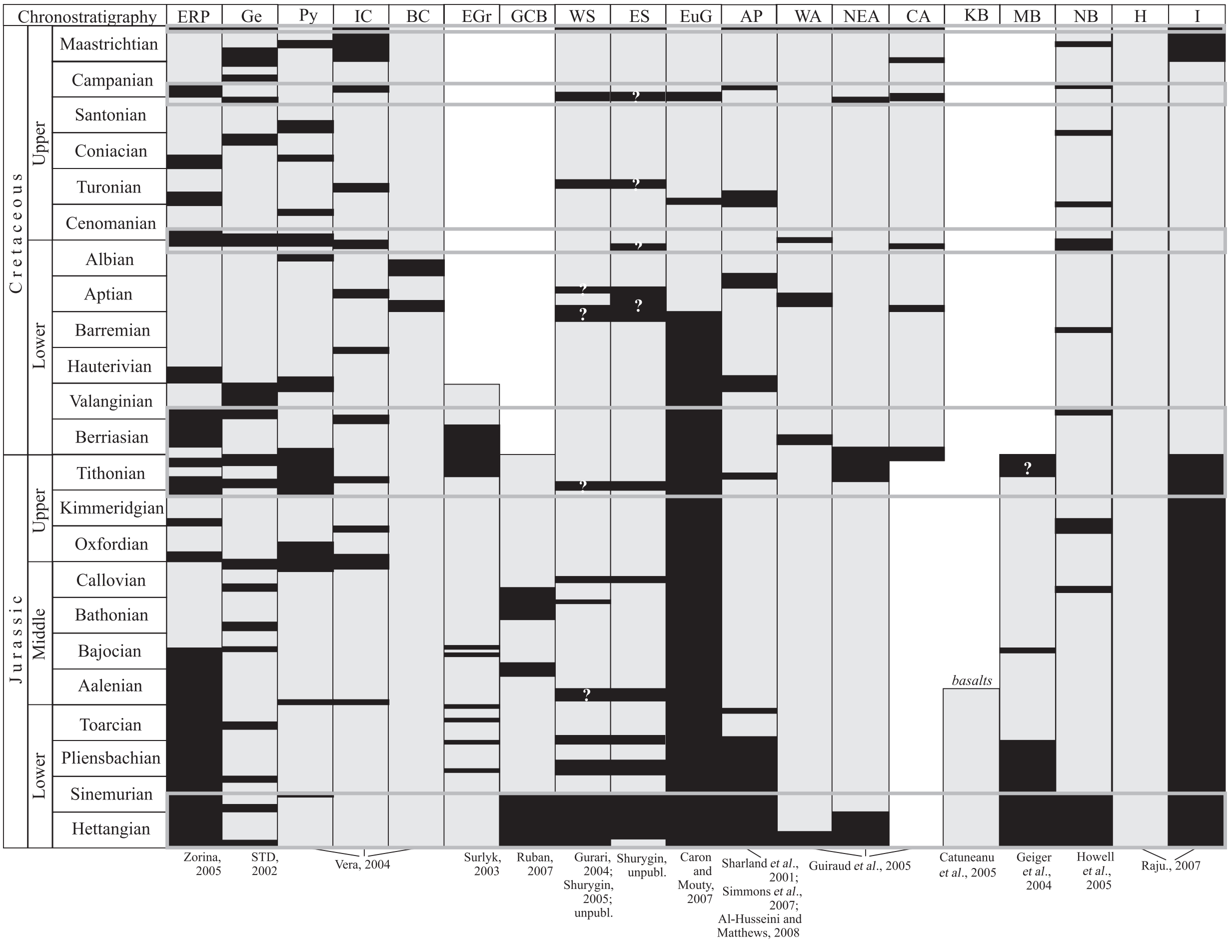

Fig. 2 The Jurassic–Cretaceous hiatuses in the studied regions. The considered intervals are highlighted as grey. No sufficient data are available for the rest intervals of some regions, which are shown by white. Bold dark-grey rectangles outline the common sedimentary breaks. See Fig. 1 for abbreviations.

establishes that both the K1–K2 and the S–C sedimentary breaks corresponded to the prominent global sea-level falls. However, why are there no global sedimentation breaks associated with some other remarkable eustatic drops, such as those in the Bathonian (Haq and Al-Qahtani, 2005) or Turonian (Haq and Al-Qahtani, 2005; Miller *et al.*, 2005)? In some regions, we may easily recognize the unconformities, which correspond to these eustatic falls (Fig. 2). But in some other regions, these unconformities do not exist. This may be explained by the above-mentioned tectonic influences. Rapid tectonic subsidence in some regions did not allow the unconformities to be formed during a time of eustatic fall. Moreover, in deep-marine environments, we do not necessarily see unconformities caused by the sea-level falls. Alternatively, those global falls, not reflected by the presence of wide-spread sedimentation breaks, are artefacts. According to Hallam (2001), regressions were mostly regional events, whereas transgressions are better traced across the world. If so, one should be careful interpreting the eustatic falls. The results of our unconformity correlations agree with Hallam (2001), suggesting the presence of many drops on the global sea-level curves not reflected by wide-spread unconformities, but also question the reliability of proposed eustatic curves.

If potentially global unconformities exist, they may be used to outline the global sequences similar to those proposed by Haq *et al.* (1987). However, it is necessary to remember that a sequence is bounded by both unconformities and correlative conformities (Van Wagoner *et al.*, 1988; Catuneanu, 2006). If so, sequence boundaries may exist at those levels, where unconformities are not so common across the globe. Thus, a more precise reconstruction of global sea-level changes will permit us to outline the global sequence boundaries.

Discussion

Although correspondence between documented sedimentation breaks and eustatic curves remains controversial, one may emphasize that only eustasy could cause them. If so, the next question is what factors would

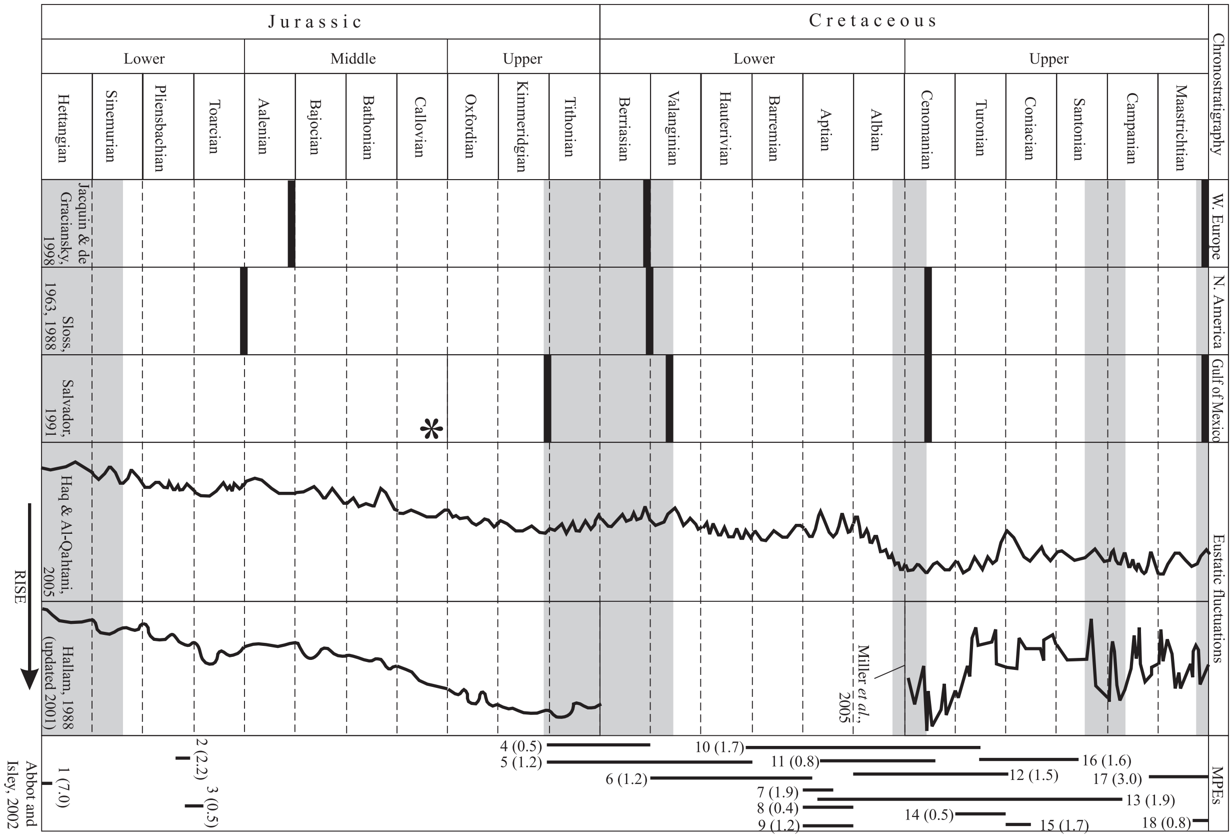

Fig. 3 Unconformities of reference regions, common sedimentary breaks for the studied regions (grey lines), eustatic fluctuations and MPEs. The horizontal scales of the eustatic curves are not identical. Each MPE is shown as a range accounting the dating errors. 1 – Central Atlantic, 2 – Karoo Province, 3 – Ferrar Dolerite, 4 – Magellan Rise, 5 – Shatsky Rise, 6 – Parana–Serra Gelal, 7 – Ontong Java Plateau, 8 – Wallaby Plateau, 9 – Manihiki Palteau, 10 – Alpha Ridge Plateau, 11 – Hess Rise, 12 – Kerguelen Plateau, 13 – Venezuelan–Colombian, 14 – Broken Ridge, 15 – Rio Grande, 16 – Madagascar, 17 – Peary Land, 18 – Deccan. Number in brackets means a size of volvanic province (×10^6 km^2). *Salvador (1994) constrained sedimentary cyclicity since the Late Jurassic.

control the T–J, the J–K, the K1–K2, the S–C and the K–T eustatic drops? Two best candidates are changes in the ice-volume and tectonics. Data from Gondwana establish that the climate was generally warm during the most of the Cretaceous with a noted exception at the Jurassic/Cretaceous boundary, when temperatures dropped by about 10 °C (Scotese, 1998; Anderson *et al.*, 1999; Scotese *et al.*, 1999). A recognizable cooling took place near the end of the Cretaceous (Keller, 2001; Nordt *et al.*, 2003). Miller *et al.* (2005) suggested episodic occurrences of ephemeral glaciations during the Late Cretaceous. Undoubtedly, these cooling phases were able to cause some eustatic drops. However, glaciation episodes are also known from the Pliensbachian (Morard *et al.*, 2003), the Callovian (Dromart *et al.*, 2003), the Early Cretaceous (Alley and Frakes, 2003) and the Turonian (Frakes and Francis, 1988; Frakes and Krassay, 1992; Frakes *et al.*, 1992), but none of them is associated with large unconformities. Several regional hiatuses can be brought into correspondence with these climatic episodes (Fig. 2). But why were other minor cooling phases more important in producing global-scale sedimentation breaks? Moreover, relative to the J–K transition, the present evidence relies on controversial climatic interpretations for this time (Husinec and Read, 2007; Zorina and Ruban, 2007).

Tectonic events such as supercontinent amalgamations and break-ups caused long-term influences on the global sea level (Miller *et al.*, 2005). But to explain relatively short-term sedimentary breaks like those documented by our study within the Jurassic–Cretaceous interval, only abrupt and intense tectonic processes are likely. The clue is given by Hallam (2001) who underlined an importance of the large-scale plume tectonics for sea-level changes at the Triassic–Jurassic transition. The available record of episodes of mantle plume activity (MPE) (Abbott and Isley, 2002) permits us to relate the T–J unconformity with the Central Atlantic MPE, the J–K unconformity with the Magellan Rise MPE and the Shatsky Rise MPE, the K1–K2 unconformity with the Alpha Ridge Plateau MPE, the Hess Rise MPE, the Kerguelen Plateau MPE, and the Venezuelan–Columbian MPE, and the K–T unconformity with the Peary Land MPE and the Deccan MPE

(Fig. 3). No plume activity is known around the Santonian/Campanian boundary (Abbott and Isley, 2002). But taking into consideration the uncertainty in the age of the Venezuelan–Colombian MPE, one may suggest its coincidence with the S–C sedimentation break. Many other MPEs took place at times, when no sedimentation breaks occurred (Abbott and Isley, 2002).

Neither climatic nor tectonic origin of the Jurassic–Cretaceous sedimentation breaks should be excluded, but our knowledge of them remains incomplete. Moreover, local tectonic subsidence could have countered their influences on regional sedimentation. The above-mentioned considerations suggest that MPE is a more likely cause of the potentially global sedimentation breaks.

Conclusions

An attempted correlation of the Jurassic–Cretaceous unconformities established in a number of regions with new or updated stratigraphic frameworks allows recognition of three potentially global sedimentation breaks, which occurred at the Jurassic–Cretaceous, the Lower–Upper Cretaceous and the Cretaceous–Palaeogene transitions. The unconformities established at the base of the Jurassic and at the Santonian–Campanian transition are not recognized in the reference regions of Western Europe, North America and the Gulf of Mexico, but they are common within those regions considered herein. Moreover, there are no unconformities existing within all considered regions. Five unconformities mentioned above are diachronous and their relationships with the eustatic falls are uncertain, because of the differences in the global sea-level constraints. The Jurassic–Cretaceous potentially global unconformities might have been caused by glaciations or MPE, among which the latter appears to be a more likely cause.

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of the Hun Superterrane. The mid-Carboniferous unconformity (instead of Hercynian unconformity) may have resulted from deformation caused by a compressional subduction-initiation phase along the outer margin of Cimmeria before it broke off in the Middle Permian-Triassic.

DISCUSSION OF THE MIDDLE EAST TERRANES

In this section, we discuss the various plate-tectonic interpretations of the individual Middle East terranes (Figures 1 and 2). We highlight conflicting interpretations and suggest preferred interpretations where data and regional considerations allow. The terrane-by-terrane sections follow from present-day northwest to southeast (Figure 1).

The Pontides and Taurides Terranes

Turkey (Figure 1) presently occupies the active collision zone between the Arabian and Eurasian plates (Bird, 2003), and interpretations of its Paleozoic history vary from a single terrane to several. Scotese (2004) positioned Turkey next to the Levant and Egypt (Figures 1 and 2) throughout the Paleozoic and early Mesozoic, and in the Cretaceous showed it drifting northwards until it collided with Eurasia in the middle Cenozoic (c. 30 Ma). Similar single-terrane models involving only Mesozoic rifting were adopted by others (e.g. Grabowski and Norton, 1995; Sharland et al., 2001).

Göncüöglü and Kozlu (2000) separated Turkey into the northern Pontides and southern Taurides by a Paleozoic ocean that closed in the Carboniferous. They correlated post-collisional granitoids and suggested that the Taurides was originally Gondwanan. As discussed earlier, Sengör (1990, Figure 11) considered the Pontides (and Kersehir) and Taurides (Menderes-Taurus) terranes to be parts of two Cimmerian ribbons with the intervening Neo-Tethys (Inner Taurides) Ocean.

Several authors, however, interpreted the Pontides as Hunic and the Taurides as Cimmerian (Figures 6 and 7; Stampfli et al., 2001; Cocks and Torsvik, 2002). Moreover, based on a detailed analysis of foraminiferal paleobiogeography and plate tectonic review, Kalvoda (2002) concluded that Turkey was Laurussian, rather than Gondwanan. An investigation of the Carboniferous (Viséan) foraminiferal and algal paleobiogeography suggests that the Taurides was located along the northwestern border of the Paleo-Tethys (Okuyucu and Vachard, 2006).

Therefore it appears that the Paleozoic paleopositions of the main Pontides and Taurides terranes of Turkey remain unresolved in the literature. As discussed above, we favor the interpretation of these two terranes as Cimmerian and specifically within the regional context of the mid-Carboniferous subduction-arc complex (Figure 11, Sengör, 1990; Xypolias et al., 2006).

The Caucasian Terranes

The Greater Caucasus terrane is presently located south of the Russian Platform (Gamkrelidze, 1997; Tawadros et al., 2006) (Figures 1 and 2, i.e. Baltica), and its Paleozoic sedimentary complexes crop out in the central Greater Caucasus Mountains (Ruban, 2006). Paleontological data from Silurian (Ludlow) carbonates (bivalve and ammonoid assemblages, Bogolepova, 1997), Pennsylvanian paleobotanical data (Anisimova, 1979), and middle and upper Paleozoic paleomagnetic data (Shevljagin, 1986) suggest that the Greater Caucasus was not a part of Baltica, as traditionally proposed (e.g. Laz'ko, 1975; Bykadorov et al., 2003). The faunal and floral assemblages, as well as the lithostratigraphic architecture, are similar to those of Hunic Perunica and Carnic Alps (Central and Alpine Europe, Figures 2 and 7). Moreover, its lower Silurian mainly clastic and volcaniclastic succession resembles that of the Hunic margin of the Paleo-Tethys.

In the Middle-Late Devonian (until Famennian) about 4,500 m of mixed volcaniclastics and volcanic rocks were deposited in the Greater Caucasus (Kizeval'ter and Robinson, 1973). The volcanic activity may have been due to tectonism between the Greater Caucasus and other Hunic terranes. Alternatively, the magmatic activity may have been associated with the closure of the Rheic Ocean (Stampfli and Borel, 2002; Figure 6). We therefore follow Tawadros et al. (2006) in assigning the Greater Caucasus to the Hun (probably Cordillera) Superterrane.

The early Paleozoic location of the Greater Caucasus before the Hunic breakaway is uncertain. Tawadros et al. (2006) positioned it along the African-Arabian margin of Gondwana, but without constraining data. In the mid-Paleozoic, it was located near the easternmost extremity of the Hun Cordillera terranes with westward strike-slip dislocation along the northern Paleo-Tethys Shear Zone in the Carboniferous - Middle Triassic, and eastward dislocation in the Late Triassic - Early Jurassic (not depicted in figures in this paper). Such a late Paleozoic to Mesozoic shear zone may have stretched along the southern margin of Laurussia and connected with an intra-Pangean shear zone (Arthaud and Matte, 1977; Swanson, 1982; Rapalini and Vizán, 1993; Lawver et al., 2002; Stampfli and Borel, 2002; Bykadorov et al., 2003; Vai, 2003; Garfunkel, 2004; Natal'in and Sengör, 2005; Ruban and Yoshioka, 2005; Tawadros et al., 2006).

Stampfli and Borel (2002) positioned Kazakhstan (or parts of it) along the easternmost part of the Hun Superterrane suggesting proximity to the Greater Caucasus in the Devonian. Available paleontological data does not support this suggestion. The trilobite species *Paciphacops* occurs in the upper Silurian to Lower Devonian strata and its distribution encompasses the circum-Pacific (Merriam, 1973; Wright and Haas, 1990; Ramsköld and Werdelin, 1991; Edgecombe and Ramsköld, 1994). Its presence in Kazakhstan (Maksimova, 1968) and absence in Europe suggests the former was located on the margin of the Panthalassic Ocean, i.e. too far to be Hunic. This is also confirmed with other paleontological data (Blodgett et al., 1990; Campbell, 1977; Chlupác, 1975; Kobayashi and Hamada, 1977; Maksimova, 1972; Ormiston, 1972; Perry and Chatterton, 1976; Pedder and Oliver, 1990; Pedder and Murphy, 2004).

The Lesser Caucasus (Transcaucaus) terrane is presently located south of the Greater Caucasus, and north of Turkey and Iran (Figures 1 and 2). Interpretations based chiefly on paleomagnetic and paleontological data (Lordkipanidze et al., 1984; Gamkrelidze, 1986), indicate that it was apparently a separate terrane. It appears to have drifted northwards together with Cimmeria ("Iran-Afghan" microcontinent of Gamkrelidze). In the absence of conflicting evidence, we assign the Lesser Caucasus to Cimmeria. We also conclude that the paleopositions of the Caucasus along the margin of Gondwana or within the two superterranes remain unconstrained.

East Turkey, Northwest Iran and Alborz Terranes

The Eastern Turkey, Northwest Iran and Alborz regions are inconsistently interpreted in published reconstructions. Sengör (1990) interpreted Eastern Turkey as a Neo-Tethyan accretionary prism. Northwest Iran is considered Cimmerian and similarly depicted by several authors (e.g. Sengör, 1990; Sharland et al., 2001), but is sometimes referred to as the Alborz terrane by others (Stampfli et al., 2001; Torsvik and Cocks, 2004). In this review we consider Northwest Iran and Alborz as separate terranes (Figures 1 and 2).

Based on paleobiogeographic studies, Kolvoda (2002) suggested that the Alborz terrane was a part of the late Paleozoic Laurussia Supercontinent. Angiolini and Stephenson (in press), based on a re-examination of early Permian (Asselian-lower Sakmarian) brachiopods of the lower Permian Dorud Formation in the Alborz Mountains and a new study of palynomorphs from the same formation, also concluded that there is little affinity with Gondwana and the peri-Gondwanan region. Brachiopod fauna shows affinities with those of Baltica (Urals and of the Russian Platform), and to a lesser extent to the Trogkofel Limestone (Carnic Alps) in the west. The palynomorph assemblage is completely different from those recorded from the Asselian-Sakmarian *Granulatisporites confluens* Biozone, which is ubiquitous in the Gondwana region. L. Angiolini (2007, written communication) and coworkers, based on their studies and published data, concluded that the Alborz, Northwest and Central Iran remained adjacent to one another throughout most of the Paleozoic. This is reflected by the continuity and common evolution of their Paleozoic sedimentary rocks, and uniform distribution of biota. They attribute the similarity of the fossil record to the Urals to surface currents and the low latitudinal position of the Iranian terranes.

In summary, Eastern Turkey may not have been a Paleozoic terrane. The Alborz, Northwest and Central Iran terranes were apparently adjacent to one another. Their paleobiogeographic signature suggests a Laurussian affinity, but in the absence of more definitive data we follow most authors and assign them to the Cimmerian Superterrane.

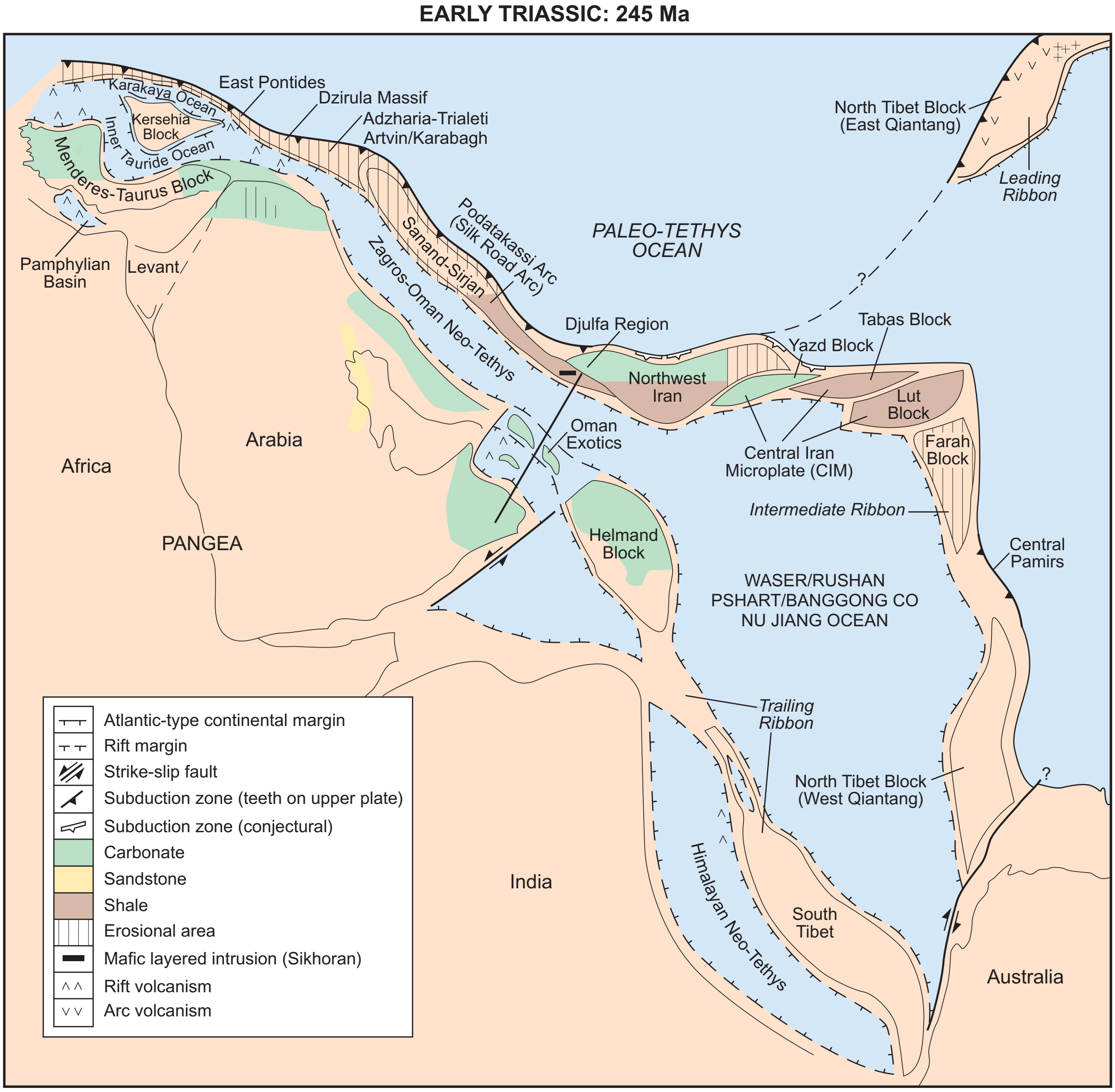

Figure 11: Plate-tectonic reconstruction of the Early Triassic (modified after Sengör, 1990). Sengör interpreted the breakaway of the Cimmeria Superterrane to consist of three ribbons that detached at different times from Pangea. L. Angiolini (2007, written communication) and coworkers believe that this model is probably valid although it is very difficult to prove in detail. For example, their data suggests that the Helmand terrane may have been attached to the middle ribbon. Sengör interpreted the embryonic Neo-Tethys Ocean to consist of several seaways: Inner Taurides, Zagros-Oman, an unnamed sea south of Oman, Himalayan and Waser/Rushan-Nu Jiang. A subduction-arc complex is interpreted along the Paleo-Tethyan margin adjacent to the Sanandaj-Sirjan terrane and possibly extending further to the southeast and west. The initiation of subduction in mid-Carboniferous times may have caused the epeirogenic swells and compressional faulting in the Arabian Plate.

Sanandaj-Sirjan Terrane

The Sanandaj-Sirjan terrane (Figures 1, 2, 4 to 11) was attached to the Zagros Mountains (and the Arabian Plate) until it broke off as part of Cimmeria in the mid-Permian - Triassic (Berberian and King, 1981; Sengör, 1990; Grabowski and Norton, 1995; Stampfli et al., 2001; Sharland et al., 2001; Scotese, 2004). Most authors show the Paleozoic position of Sanandaj-Sirjan adjacent to the Zagros Suture, effectively implying that today it occupies the same approximate position as 250 million years ago.

This seems remarkable as it was involved in the opening and closing of the Neo-Tethys Ocean. During the opening it may have subducted the Paleo-Tethys (Sengör, 1990), and during its closing, the Neo-Tethys (Ghasemi and Talbot, 2005).

Central Iran

Sengör (1990) divided the Central Iran microplate into the Lut, Tabas and Yazd blocks (Figure 11). Other authors consider Central Iran and Lut as synonyms (Stampfli and Borel, 2002; von Raumer et al., 2002, 2003; Torsvik and Cocks, 2004; Scotese, 2004; Golonka, 2004); or two neighboring terranes: Central Iran and Lut, or Yazd and Lut (e.g. Sharland et al., 2001; Stampfli et al., 2001). We adopt Sengör's Central Iran terrane and follow others by considering it as Cimmerian.

Zagros Mountains and Makran Region

The Zagros Mountains region in southwest Iran forms a part of the Miocene-Pliocene collision zone between the Arabian and Eurasian plates (Figure 1). This region was a part of the Arabian Plate from the late Neoproterozoic to the present-day (Berberian and King, 1981; Sepehr and Cosgrove, 2004). During the Permian-Triassic (Figures 10 and 11), the opening of the Neo-Tethys Ocean along the Zagros Suture Zone was accompanied by normal faulting and horsts and graben systems (Sepehr and Cosgrove, 2004).

South of the Zagros Mountains, the Makran region in Iran and Pakistan (Figure 1) consists of the Inner Makran ophiolites and the Cenozoic Makran and Saravan accretionary prisms (McCall, 1997, 2002, 2003). This region is associated with the NE-directed subduction of the Gulf of Oman oceanic crust (a remnant of the Neo-Tethys Ocean) beneath Iran. The Makran core may have amalgamated with Central Iran and Sanandaj-Sirjan during the Triassic (McCall, 2003). Therefore, Makran may have formed a part of Mesozoic Cimmeria.

Helmand and Farah Terranes

Afghanistan, western Pakistan and southeast Turkmenistan are cored by the southern Helmand and northern Farah terranes and considered Cimmerian (Figures 1 and 2; Sengör, 1990; Sharland et al., 2001; Stampfli et al., 2001; Golonka, 2004). Scotese (2004) adopted Sengör's (1990) model showing Helmand and Farah formed parts of two Permian-Triassic ribbons (Figure 11). Together with Karakoram in north Pakistan, the Farah and Helmand terranes are considered Cimmerian.

CONCLUSIONS

Recent publications that interpreted the Paleozoic tectonic units of the Middle East and their paleopositions were reviewed in the global context of supercontinents and exhumed vast oceans, to individual terranes. Adjoining the Arabian and Levant plates, ten Paleozoic Middle East terranes were apparently involved in the evolution of the Gondwana and Pangea margins and the Hun and Cimmeria superterranes. The Cimmerian terranes that broke off from Gondwana in mid-Permian - Triassic appear to have been: (1 and 2) Turkey's northern Pontides and southern Taurides; (3 to 6) Alborz, Central Iran (Lut, Tabas and Yazd), Sanandaj-Sirjan and Northwest Iran; (7 and 8) Helmand and Farah of Afghanistan, western Pakistan and southeast Turkmenistan; and (9) the Lesser Caucasus. The Greater Caucasus may have been Hunic.

The Caledonian and Hercynian orogenies occured far away from Arabia. Correlation between these two orogenies and deformations in Arabia can be misleading. They imply that far-field stresses were transmitted many thousands of kilometers from the orogenic fronts to Arabia's crust. The terms Caledonian and Hercynian should not be applied to the tectonic evolution of Arabia. Instead two significant and more proximal tectonic events were identified as possible near-field sources of regional deformation. The mid-Silurian breakaway of the Hun Superterrane is identified as a candidate that may be related to the mid-Silurian to Middle Devonian (middle Paleozoic) uplift in North Arabia and possibly Oman. The initiation of subduction, which could have preceeded the mid-Permian - Triassic breakaway of Cimmeria, is considered a possible force for the regional mid-Carboniferous faulting and epeirogenic deformation in Arabia.

Middle East plate-tectonic models require much more data and investigations if they are to be firmly constrained. The first step is to adopt common boundaries and names for the terranes, not only for the Middle East, but also of those in Asia and Europe (Figures 1 and 2). The second step requires constructing a regional tectono-stratigraphic framework that crosses from the interior of the Arabian Plate and its outer margins (Oman, Zagros, North Iraq, Syria and Southeast Turkey) to the ten and possibly more Middle East terranes. The framework requires correlating stratigraphic rock units that are much better constrained by age (biostratigraphy), paleontology and tectonics. Additionally, paleomagnetic and age data, together with the descriptions and interpretations of volcanic rocks could better clarify many aspects of the tectonic events.

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Mesozoic marine fossil diversity and mass extinctions: an experience with the middle XIX century paleontological data

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Abstract

The recalculation of the paleontological data of A. d'ORBIGNY on the marine fossils distribution within the Mesozoic stages was attempted. About 18.000 species and 1200 genera were accounted. It was surprisingly found that fossils diversity changes constrained with the 150 years old data seem to be enough similar to ones calculated with the recent data. Most of the Mesozoic mass extinctions might be identified in the middle of the XIX century. That means the leading role in the growth of paleontology belongs not only to the data collecting, but to the analytical approach developments.

Key words

Diversity, mass extinctions, catastrophism, fossil record, completeness, Mesozoic.

1. INTRODUCTION

Thanks to the excellent compilations of SEPKOSKI (1982) and his co-workers (e.g., SEPKOSKI *et al.*, 1981) we obtained a real basis to discuss the fossil diversity changes and particulary such outstanding events in geological history as mass extinctions. Later biodiversity trends were several times updated (BENTON, 1995; PETERS & FOOTE, 2001; SEPKOSKI, 1993; SIGNOR, 1985). Meantime, the data are still incomplete and they will be much enlarged and improved later; e.g., recently only ~10-15% of Paleozoic megafossils were described (BOUCOT, pers. comm.). Therefore, a question arises: what may be changed in our imaginations on the diversity changes and the mass extinctions due to further paleontological studies? One of the possible way to answer it is to compare the estimations of fossil diversity changes based on recent and past data compilations. This also allows to clarify what could do the paleontologists with their data more than a hundred years ago. Such retrospective analyses were already attempted few times with other data (e.g., PADIAN & CLEMENS 1985; SMITH, 2001).

In the middle of the XIX century an "explosion" in the paleontological studies began. The first really comprehensive data compilations were appeared. One of the best attempt was made by Alcide d'ORBIGNY, whose contributions to studies of all principal fossil groups are enormous. His "Prodrome..." (ORBIGNY, 1850a, b, 1852) was a complete reference of the fossil distribution for the 1850s.

The main goal of this article is to analyze the Mesozoic marine fossil total diversity recalculating the data of A. d'ORBIGNY and then to compare results with the present knowledge on the marine biodiversity changes in the Earth's history.

2. MATERIALS AND METHODS

The d'ORBIGNY's "Prodrome..." contains information on about ~18.000 species and ~1200 genera of bivalves, brachiopods, bryozoans, echinoids, foraminifers, etc. Although not all fossil groups were accounted, it is possible to assume these data characterized enough representatively the whole marine biodiversity. The main difficulty to analyze them is connected with the misunderstanding of the time intervals used. Finally it was decided to concentrate the attention only on the Mesozoic (and Danian), which stages may be more evidently attributed to the present subdivisions.

The "database" of d'ORBIGNY allows to analyze total species and genera diversity per stages (TSD and TGD respectively).

3. MARINE FOSSIL DIVERSITY IF ANALYZED IN THE XIX CENTURY

TSD as it might be analyzed by d'ORBIGNY quickly accelerated in the Triassic (Fig. 1). In the average it

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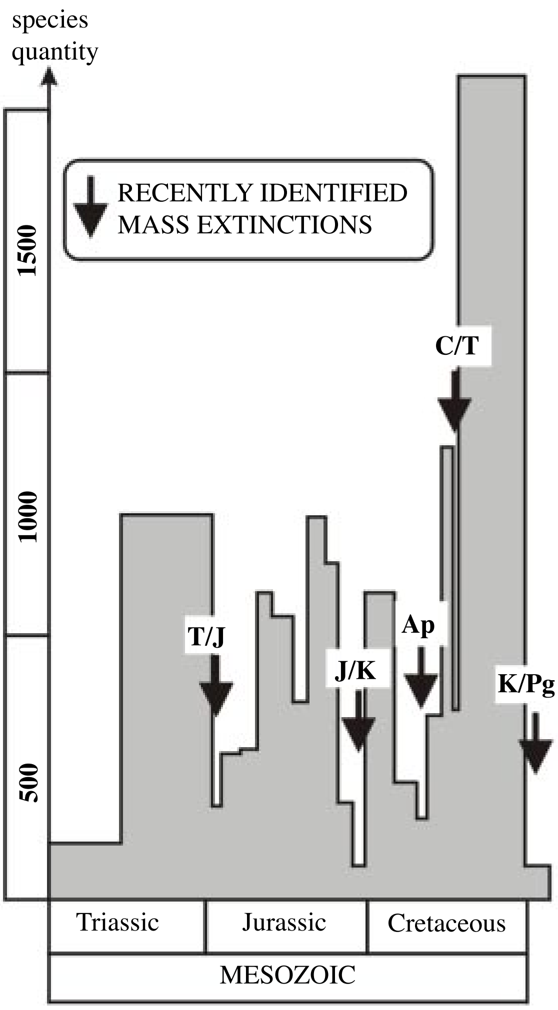

Fig. 1: The total species diversity changes in the Mesozoic calculated with the data of ORBIGNY (1850a, b, 1852).

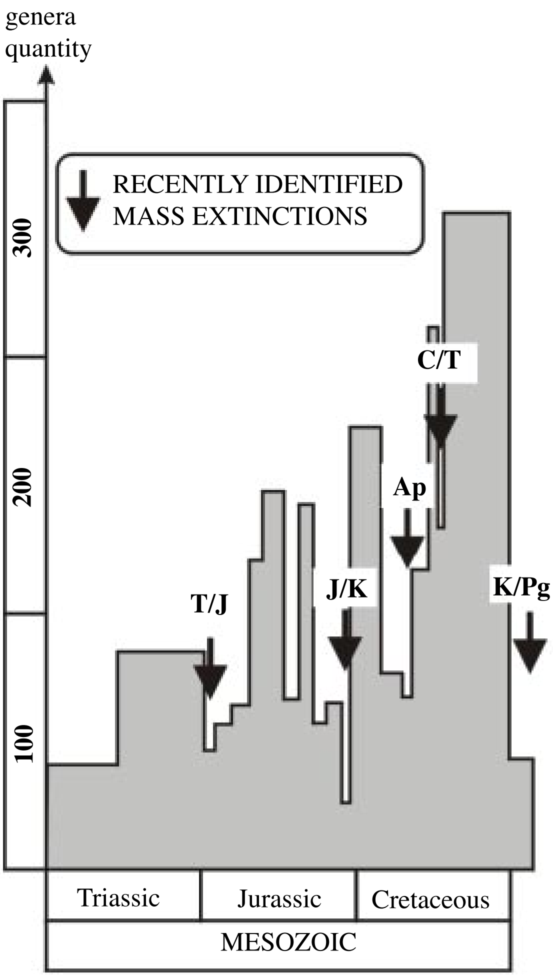

Fig. 2: The total genera diversity changes in the Mesozoic calculated with the data of ORBIGNY (1850a, b, 1852).

stayed at the same level until the mid-Cretaceous, when a new significant rise began. An abrupt decrease of the species number occured in the Danian. But the taxa quantity significantly fluctuated through the Mesozoic. The most intense TSD drops are documented for the beginning of the Jurassic, in the Callovian, in the end of Late Jurassic, in the Aptian and Turonian. Some long intervals corresponded to the several stages of the recent chronostratigraphical scale were defined by d'ORBIGNY as unique stages (Conchylian and Saliferian in the Triassic, Neocomian and Senonian in the Cretaceous). Therefore, it is impossible to discuss the TSD changes in detailes at them. Moreover, the diversity peaks observed for some of them should be reduced after their differentiation into the several distinct stages.

TGD changes are comparable with above mentionned patterns of TSD (Fig. 2). Only minor differencies may be documented. All the diversity drops established for species have also been recorded for genera.

4. DISCUSSION

Let's imagine that Alcide d'ORBIGNY analyzed marine fossils diversity in the middle of the XIX century. Results he received were not surprising for him and his contemporaries. They well coincided with the concept of "catastrophism" wide spread that times. Abrupt changes may easily be attributed to the catastrophic events in the Earth's history, which led to the disappearance of fossils. Just note, in reality the understanding of the Earth's history of d'ORBIGNY fluctuated between CUVIER and LAMARCK (LAURENT, 2002), i.e. between "catastrophism" and "evolutionism". But finding intervals of fossils disappearance, he had to chose the first concept without any doubts. This might turn all the further development of paleontology and geology. It is difficult to say was it good or bad, but none of these events mentionned above occurred in real history.

And are results of d'ORBIGNY's data recalculation surprising for us ? Both negative and positive answers are possible.

As we know enough much on the global fossils diversity (PETERS & FOOTE, 2001 ; SEPKOSKI, 1993) and on the mass extinctions (COURTILLOT, 1999 ; HALLAM & WIGNALL, 1997 ; RAUP, 1993 ; RAUP & SEPKOSKI, 1982) the results of d'ORBIGNY 's data recalculation seem not to be surprising for us. The diversity increase in the Triassic may be interpreted as a recovery after the mass extinction at the Permian/Triassic boundary. The diversity drops in the beginning and the end of Jurassic, in the Aptian and Turonian are corresponding well to the Triassic/Jurassic, Jurassic/Cretaceous, Early Aptian and Cenomanian/Turonian mass extinctions. Finally the significant drop of TSD and TGD in the Danian seems to be a result of famous K/T event. The unique unusual event is a diversity drop in the Callovian, which can be explained by the incompleteness of d'ORBIGNY 's data.

On the other hand, the results of our recalculation are very surprising : all Mesozoic mass extinctions (except one during the Early Toarcian) could be documented 150 years ago – even such short-termed events as recorded in the Early Aptian and at the Cenomanian/Turonian boundary. In fact, the intensity of mass extinctions recorded by the "past" data is uncomparable with estimated at now ; also the relation of strength between distinct extinction events differs from one established today (PETERS & FOOTE, 2001). This is caused by the high degree of d'ORBIGNY's data incompleteness (absence of many taxa found later, taxonomical errors, restriction to the Western Europe, stratigraphical uncertainties, etc.). But even that fact, many of presently stated events might be recorded in the middle of the XIX century, is too significant itself.

It was pointed out above d'ORBIGNY accounted ~1200 genera, while SEPKOSKI - ~36.000 ones. This means the quantity of paleontological data increased in 30 times during the century and a half, although the results of recalculations are somewhat similar. Anyway this seems to be possible, because d'ORBIGNY evidently collected data on the most common taxa. Further studies led to the relatively uniform increase of the paleontological information for each interval.

It is interesting, SMITH (2001), comparing results of PHILLIPS (obtained in 1860s) to ones of SEPKOSKI (1993) and BENTON (1995), also stated enough high similarity between their curves.

Now we can try to answer the question outlined in the beginning (see Introduction). It is possible, that even very significant enlargement and improvement of paleontological data in the future may not lead to the great changes in our principal imaginations on the marine fossil diversity changes and the mass extinctions, although many new detailes will become clear. The development of a science needs something more than a collecting of data, and evidently in the future paleontologists could do much more interesting analyses using even our present data. Perspectives of the fossil record exploration lie mostly in the area of development of the analytical techniques (see also JABLONSKI, 1999).

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Diversity dynamics and mass extinctions of the Early–Middle Jurassic foraminifers: A record from the Northwestern Caucasus

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Abstract

The Early–Middle Jurassic foraminiferal assemblages of the Northwestern Caucasus, including a total of 315 species and 68 genera, were analysed to establish the principal diversity patterns at substage level of resolution. An overall conclusion is that the number of species varied significantly in contrast to the number of genera. The most diversified were Late Sinemurian–Pliensbachian, Late Toarcian–Early Aalenian, and Late Bajocian assemblages. Significant diversity drops have been documented in the Early Toarcian and in the Middle Aalenian. The first of them corresponds to the well-known mass extinction, which regionally was initiated in the Pliensbachian and terminated when rapid Late Toarcian diversification began. The second event could be considered as a regional evidence of a new mass extinction that appeared in the Middle Aalenian and was most likely associated with the Aalenian/Bajocian regional anoxic event also recorded in the Carpathian part of the Western Tethys. The most intense foraminiferal turnovers took place in the Toarcian, and in the Middle Aalenian–Early Bajocian. The Lazarus-effect in spite of its high value has not much influenced the estimation of principal diversity trends.

Keywords: Foraminifers; Diversity; Origination; Mass extinction; Early–Middle Jurassic; The Caucasus; Tethys

1. Introduction

Evolution of the Early–Middle Jurassic marine biota in response to palaeoenvironments is a significant subject to be studied with the help of microfauna. Foraminifera are especially useful microfossils in such studies due to their relatively high diversity, widespread occurrence, and a high preservation potential. Early–Middle Jurassic

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assemblages dominated by benthic foraminifera of the suborder Lagenina were far from actualistic assemblages known from Late Cretaceous, Tertiary or Recent oceans. At the same time, several new foraminiferal lineages appeared, including planktonic foraminifera, that dominated post-Jurassic assemblages. Our recent knowledge on Jurassic foraminifera from the Western Tethyan and and Boreal realms is already considerable. Nevertheless, data of comparable quality and resolution are still not available that is especially evident in the taxonomic data from Western and Eastern Europe. First attempts summarizing global dynamics of Jurassic foraminifera based on genera and higher taxonomic levels were presented by Tappan and Loeblich (1988), and later by Basov and Kuznetsova (2000). The problem appears when we try to compile data at species level at the species level from different palaeogeographic areas. In this case, the best method is to focus on selected areas and look at dynamics at regional scales. The next step would be to compare different faunal dynamics from all available palaeogeographic areas.

The Caucasus seems to be a promising region for analyzing dynamics of Early–Middle Jurassic foraminiferal assemblages (Figs. 1 and 2). Their record, including over 300 species, in north-west Caucasus is substantial (Antonova, 1958, 1962, 1975; Antonova and Pintchuk, 1991; Alieva et al., 1991; Antonova et al., 2000; Mamontova, 1956, 1957; Rostovtsev et al., 1992). Currently used foraminiferal regional biozonation (Alieva et al., 1991; Antonova and Pintchuk, 1991; Rostovtsev et al., 1992) is characterized by resolution comparable with that of brachiopods or even ammonites (Fig. 2; see Rostovtsev et al., 1992). Moreover, due to the lack of uniform macrofaunal record, it becomes especially significant for both stratigraphical and palaeoenvironmental studies (Ruban, 2002a).

This paper presents an attempt of quantitative analysis of the Early–Middle Jurassic foraminifers from the North-western Caucasus with a view to

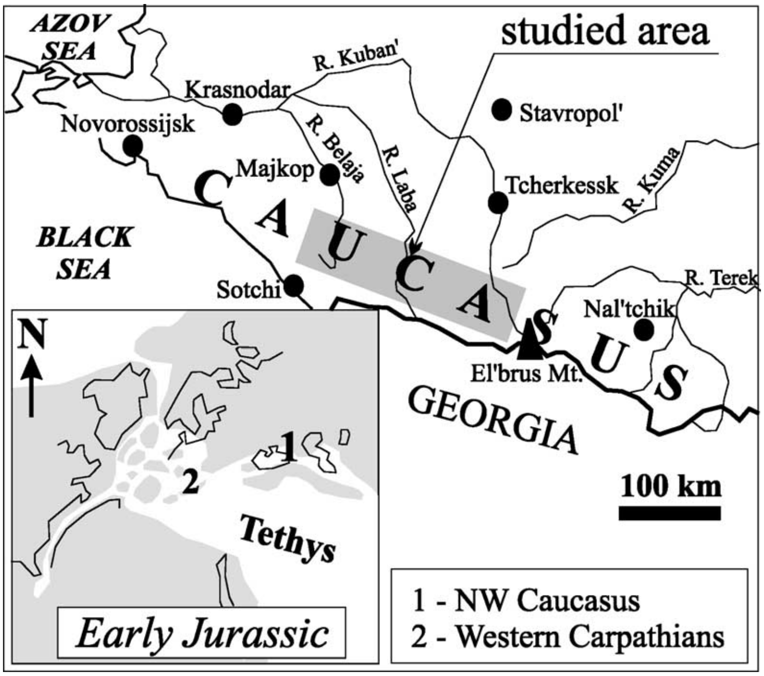

Fig. 1. Geographical location of the NW Caucasus (paleogeographic base map is modified from Owen, 1983; Dommergues et al., 2001). Land areas shaded.

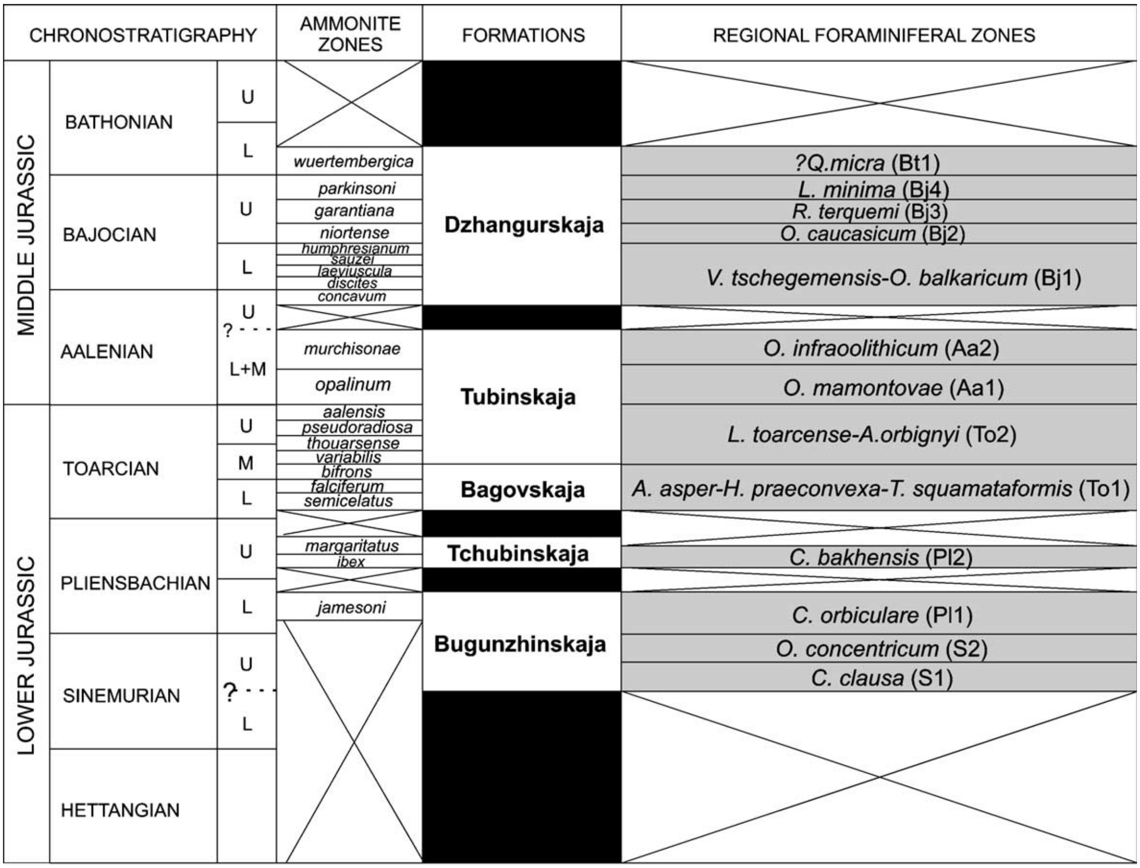

Fig. 2. Stratigraphy of the Lower–Middle Jurassic deposits of the NW Caucasus (zones after Alieva et al., 1991; Antonova and Pintchuk, 1991; Rostovtsev et al., 1992, with slight modifications).

obtain regional trends in taxonomic diversity dynamics and to evaluate possible influences of the global mass extinctions.

2. Geological setting

2.1. Lithostratigraphy

The studied area (Fig. 1) represents the western part of the Labino–Malkinskaja tectonic zone, which was incorporated into the active structures of the Northwestern Caucasus (Lordkipanidze et al., 1984; Meister and Stampfli, 2000).

The deposits of the Early–Middle Jurassic interval have already been investigated (Krymholz, 1972; Prosorovskaja, 1979; Rostovtsev et al., 1992; Granovskij et al., 2001; Ruban, 2002a). They are subdivided into 5 formations (Figs. 2 and 3) (Prosorovskaja, 1979; Rostovtsev et al., 1992). The Bugunzhinskaja Formation (Upper Sinemurian–Lower Pliensbachian) consists of sandstones and dark-grey claystones up to 150 m thick, with calcareous interbeds. A hiatus embraces the middle part of the Pliensbachian. It is overlain by the terrigenous Tchubinskaja Formation (Upper Pliensbachian), 300–350 m thick. A hiatus on the top corresponds to the uppermost Pliensbachian–lowermost Toarcian. Above the hiatus, the Bagovskaja Formation (Lower–Middle Toarcian) starts with conglomerates (mixed grains of different size) and succeeding sandstones with intercalations of claystones, altogether about 700–800 m thick. There follows the Tubinskaja Formation (Middle Toar-

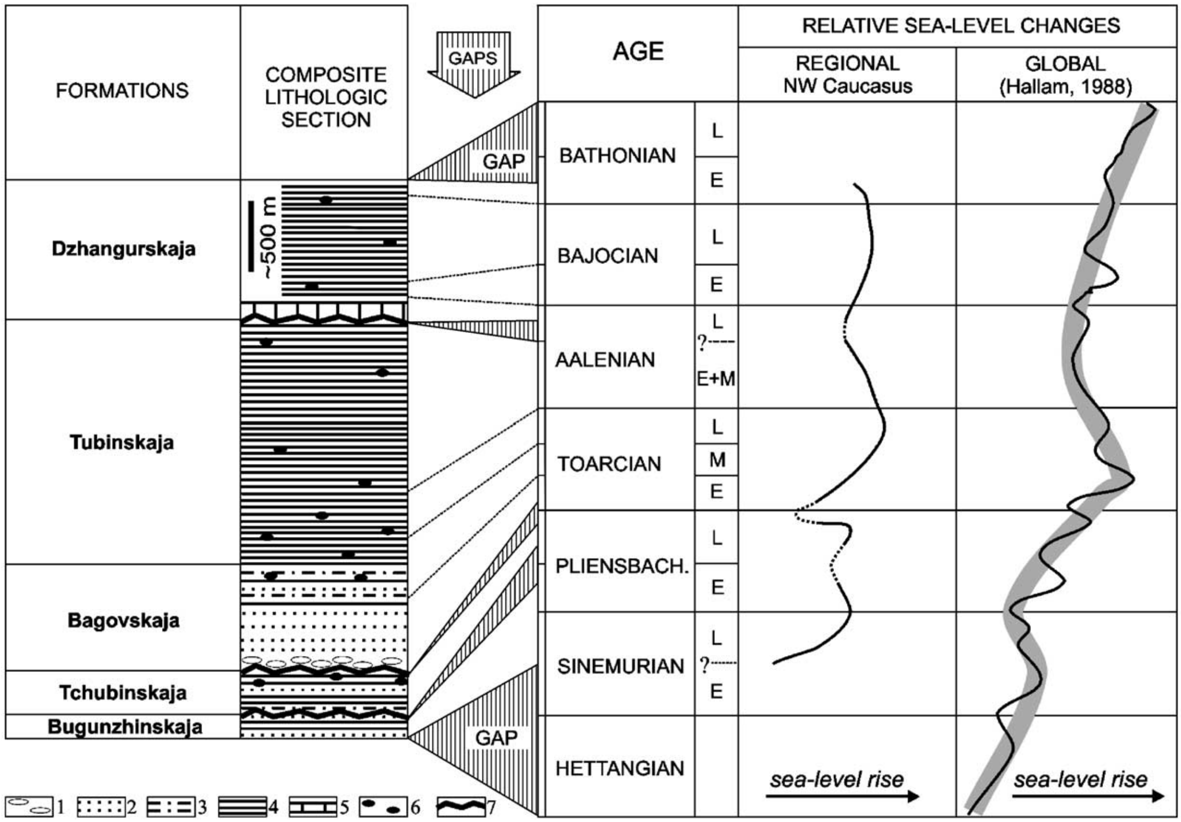

Fig. 3. Composite lithologic section of the Lower–Middle Jurassic and relative sea-level changes in the NW Caucasus compared to the global eustatic curve (after Hallam, 1988). 1—conglomerates, 2—sandstones, 3—siltstones, 4—claystones, 5—limestones, 6—siderite concretions.

cian–Middle Aalenian) consisting of dark grey to black shaly claystones, so-called "black shales", up to 1800 m thick. A hiatus, probably corresponding to a middle part of Aalenian, separates the deposits of the Tubinskaja Formation from the deposits of the Dzhangurskaja Formation (uppermost Aalenian–lowermost Lower Bathonian). This formation is represented by a 1 m thick pinkish detrital crinoidal limestone, followed by dark claystones and siltstones (altogether up to 1000 m thick). These dark grey to black claystones and siltstones of all levels contain siderite concretions and dispersed pyrite grains (Granovskij et al., 2001; Rostovtsev et al., 1992; Ruban, 2002a). An interval, corresponding to a major regional hiatus, covers a considerable part of the Bathonian. The terrigenous Callovian deposits overly older deposits with the angular unconformity.

2.2. Biostratigraphy

An ammonite zonation was developed for the Caucasus by Rostovtsev et al. (1992). Regional zones are cited below to facilitate the interregional correlation of the units used in this paper. None zones were established in the Sinemurian. This ammonite zonation is quite limited within the Pliensbachian to: *Uptonia jamesoni* (upper part of the Bugunzhinskaja Formation), *Tragophylloceras ibex* and *Amaltheus margaritatus* zones (Tchubinskaja Formation, although foraminifers suggest that relation of its lower part to the *T. ibex* Zone is doubtful) were defined in successive order, but sedimentary gaps exist between them. Further 7 regional zones were defined in the Toarcian: *Dactylioceras semicelatum*, *Harpoceras falciferum*, *Hildoceras bifrons* (Bagovskaja Formation), *Haugia* *variabilis*, *Grammoceras thouarsense*, *Dumortieria pseudoradiosa*, *Pleydellia aalensis* (lower part of Tubinskaja Fm). Aalenian zones include *Leioceras opalinum*, *Ludwigia murchisonae* (upper part of the Tubinskaja Formation) and *Graphoceras concavum* (not well established in NW Caucasus). The following zones were defined in the Bajocian: *Hyperlioceras discites*, *Witchellia laeviuscula*, *Otoites sauzei*, *Stephanoceras humphresianum*, *Strenoceras niortense*, *Garantiana garantiana*, *Parkinsonia parkinsoni*. A single unit represented by the *Parkinsonia würtembergica* Zone was established in the base of Bathonian. Both Bajocian and Bathonian zones are attributed to the Dzhangurskaja Formation. A detailed correlation of ammonites-based zones of the Caucasus and reference regions of Western Europe is in progress now, but a preliminary comparison to Spanish (Sandoval, pers. comm. 2004) and French sections indicate their utility for the chronostratigraphical correlation.

Regional biozonation has been developed based on the studies of foraminiferal assemblages from the Lower and Middle Jurassic of the NW Caucasus (Alieva et al., 1991; Antonova and Pintchuk, 1991; Rostovtsev et al., 1992). All units identified were proposed as “zones” or “beds with”. In fact, they are abundance (acme)-assemblage zones (sensu Salvador, 1994). This zonation is linked to the chronostratigraphy of the Early–Middle Jurassic (Fig. 2). It was finally done by Antonova and Pintchuk (1991), who established correlations between the regional zonation and other, mostly European, regions. This zonation is directly attributed to the formations (Rostovtsev et al., 1992). In this paper, we generally accept proposed stratigraphical framework except for the Aalenian–Bajocian transition. Brachiopods suggest the lowest horizons of the Dzhangurskaja Fm may be related to the Upper Aalenian (Ruban, 2003). Therefore, we extended the lower boundary of the Bj1 foraminiferal zone a little downwards (see Fig. 2).

Foraminifers have a larger potential in the NW Caucasus for the development of the regional Lower–Middle Jurassic biostratigraphy than ammonites (Ruban, 2002a). Additionally their record is more uniform, while ammonites often are restricted to the relatively short intervals within the stratigraphic succession. We would like to stress that the presented regional foraminiferal zonation (Fig. 2; see Rostovtsev et al., 1992) has just local or regional value limited to the NW Caucasus. Foraminifers have been used to correlate a large number of small outcrops presenting several meters thick sections. Comparison of the NW Caucasus to different Tethyan and Boreal areas would be possible after verification of all described foraminiferal taxa.

2.3. Palaeogeography and palaeoenvironment

The studied basin (Fig. 1) was located at the northern Tethyan periphery in the Early–Middle Jurassic (Lordkipanidze et al., 1984; Meister and Stampfli, 2000). The deposits mentioned above accumulated in a marine sedimentary basin, which was relatively warm during the whole Early and Middle Jurassic interval except for the Early Aalenian when palaeotemperatures probably dropped to ~10 °C (Jasamanov, 1978). Marine transgressions occurred during the Sinemurian–Early Pliensbachian, Early–Middle Toarcian and Early Bajocian intervals, while regressions took place at the end of Pliensbachian, in Middle Aalenian, and in Middle–Late Bajocian (Fig. 3). The “black shales” possibly have been accumulated on the shelf margin (Granovskij et al., 2001). Their sedimentologic characteristics, such as coloration, abundant siderite concretions, and synsedimentary and/or early diagenetic pyrite mineralization suggest deposition under dysoxic to anoxic conditions.

We suppose that the silled basin model best describes conditions and sluggish circulation in such a stratified dysoxic basin (see Bernoulli and Jenkyns, 1974; Wignall, 1994) recorded in the Lower–Middle Jurassic of North-west Caucasus. Large terrestrial areas under a wet or at least seasonally wet climate surrounded the basin in Early Jurassic (e.g., Hallam, 1994). Sedimentation of the described above thick deposits took place in a tectonically active basin with a relatively high subsidence. These conditions are probably responsible for a relatively high average sedimentation rate calculated at 10–15 cm/ka.

A general palaeogeographical position of NW Caucasus is still under debate due to the lack of high-quality data on all faunal groups. Dommergues (1987) locates this area within the Euro–Boreal domain, while Westermann (2000) considers it as a part of the Tethyan Realm. It seems, however, that this region shows some transitional features between both realms.

3. Materials and methods

Several reviews of the Early–Middle Jurassic foraminiferal data from the NW Caucasus were presented (Alieva et al., 1991; Antonova and Pintchuk, 1991; Rostovtsev et al., 1992; Antonova et al., 2000). They include both biozonation development, interregional correlations and taxonomic descriptions. This paper is based on the comprehensive information on biostratigraphic ranges of all 315 recorded species within this region, refereeing to the established foraminiferal zones. These data, presented in sources mentioned above, were compiled by Antonova, Zozulja and Pintchuk, and provided with permission by T.N. Pintchuk (see Appendix). The data were obtained from hundreds relatively small sections and available boreholes. Unfortunately, the NW Caucasus lacks extensive continuous sections, thus, the generalized data are used for analysing diversity trends instead of studying them in selected type sections. Most explanatory data on sampled horizons were already presented by Alieva et al. (1991), as well as, Antonova and Pintchuk (1991).

Among different measures of diversity usually chosen to be analysed (see Buzas, 1979), the most significant are: total diversity (total number of species or genera), appearance, disappearance, origination and extinction rates, number of singletons (sensu Foote, 2000), as well as two indices used to estimate rates of association changes (Ruban, 2001, 2002a).

Making a distinction between the terms “appearance” and “origination”, and “disappearance” and “extinction” is necessary because of a high amount of the so-called Lazarus-taxa (Flessa and Jablonski, 1983; Jablonski, 1986; Wignall and Benton, 1999; Fara, 2001). If a break in a taxon range is recognized, this taxon may appear or disappear several times, while only once it can be originated or go extinct forever. To avoid the Lazarus-effect, it is also necessary to recalculate data accounting for possible presence of taxa at intervals corresponding to their temporal absence in the fossil record. This way we document only the highest probable value (HPV) of this effect. If the Lazarus-effect were connected with the real temporal disappearance (regional or global) or migration, and not with the incompleteness or defects of fossil record, the break in a taxon regional range would be real. In this case we would consider disappearances as extinctions and appearances as originations. In fact, we can hardly conclude unequivocally, what is the nature of the Lazarus-effect (especially when analysing regional data). Considering the Early–Middle Jurassic foraminifers from the NW Caucasus, their Lazarus-effect seems to be an effect of incompleteness of the fossil record, so the HPV possibly is not so different from the real effect value.

Evaluation of the number of appearances/originations in the first (lowermost) zone within the succession is not possible because there is no data on the preceding zone to compare. The disappearances/originations in the terminal zone might be an effect of sedimentary hiatus in the Upper Bathonian, when foraminifers might have still existed.

Two indices of associations changes rate proposed by Ruban (2001, 2002b) have been used. The first of them is R, which reflects the Jaccard similarity (Jaccard, 1901; Shi, 1993) of two fossil assemblages, characterized stratigraphic intervals:

$$R = C/[(N_1 + N_2) - C],$$

where C is the number of common taxa for two intervals, and N_1 and N_2 are the taxa quantities in the lower and upper intervals respectively.

Faunal transformation rate (FTR) is estimated as $1/R$. This rate shows how quickly changes in taxonomical composition of assemblages took place through geological time. A similar method, but with Simpson’s coefficient, has been used by Hallam (1983), as well as, Smith and Tipper (1986) to evaluate macrofaunal evolutionary turnovers. Another index is R_{st}, which reflects a similarity of taxonomical structure of diversity in assemblages. It is determined as a simple coefficient of Spearman rank correlation (Kendall, 1975) between two assemblages by presence/absence of the genera, accounting species number in each of genera. The rate of transformation of taxonomic diversity structure (TTDSR) could be estimated as $1/R_{st}$. It shows changes in the generic control of species diversity, i.e., changes in significance of each genus for the determination of species diversity.

4. Species and genera diversity dynamics

The species diversity dynamics of foraminifers in the Early–Late Jurassic of the NW Caucasus is not characterized by any kind of stability (Fig. 4). After a slight diversification in the Late Sinemurian–Pliensbachian (S2–Pl2 zones), a sharp decrease of the total species number occurred in the beginning of Toarcian (To1 zone). It was caused primarily by a high value of extinction rate in the Pliensbachian and less by the decline of origination rate. But just after this Toarcian event a fast recovery took place (To2 zone). The diversity maximum was reached during the Late Toarcian–Early Aalenian (To2, Aa1 zones). Then a new diversity drop occurred (Aa2 zone), which was connected with a high rate of extinction documented in the Aa1 zone. A gradual recovery that began at the end of Aalenian (Bj1, Bj2 zones) did not compensate the previous decrease. A relatively high rate of originations was preserved during the first half of Bajocian (Bj1, Bj2 zones). Then, until the end of the Bajocian the species number did not change much, while the Bathonian foraminiferal assemblages became impoverished due to a major regional regression episode.

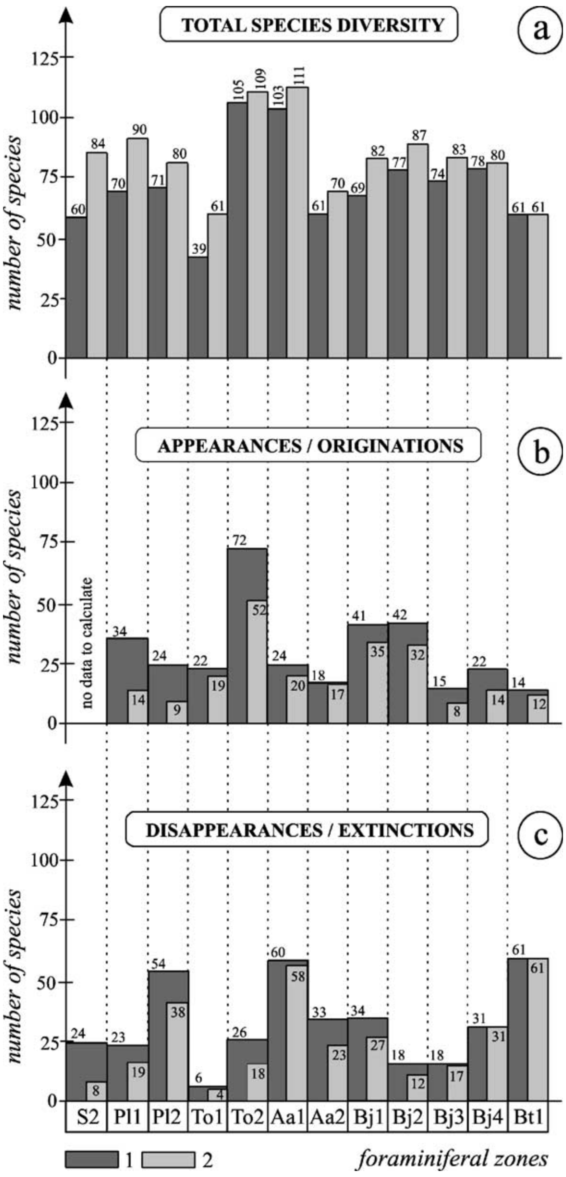

Fig. 4. Species diversity of foraminifers (per zones): total diversity (a), appearances/originations (b), disappearances/extinctions (c). Columns: 1—“original”, 2—accounting for the Lazarus-effect. See Fig. 2 for foraminiferal zones abbreviations.

In contrast, the genera diversity dynamics seems to be rather stable in the Early–Middle Jurassic (Fig. 5). The number of genera did not vary so much as the species number. The most significant event was a decline of diversity in the To1 zone followed by a recompensed diversification documented already in the To2 zone. The diversity drop in the Aa2 zone was not very great. It is evident that overall diversity changes of foraminifers in the NW Caucasus resulted mostly in species (but not generic) turnovers. That means strong variations of average species number within the relatively stable number of genera.

Calculations of R and R_{st} indices changes suggest that the most intense transformations of assemblages took place within the intervals of Pl2–To2 zones and Aa1–Bj2 zones, when both the specific and generic compositions have been renewed (Figs. 6 and 7). It is evident that principal transformations at species level took place earlier than at the level of genera. That could be possibly explained by the above-mentioned higher stability at the generic level. In fact, this shifted turnover at this level resulted from low changes in number of extinctions in contrast to relatively high appearances/originations of new genera just after both extinctions levels, i.e., in To2 and Bj1–2 zones (Figs. 4–6).

The singletons number (i.e., quantity of species appeared, existed and disappeared within the same single interval) changed cyclically during the Early–

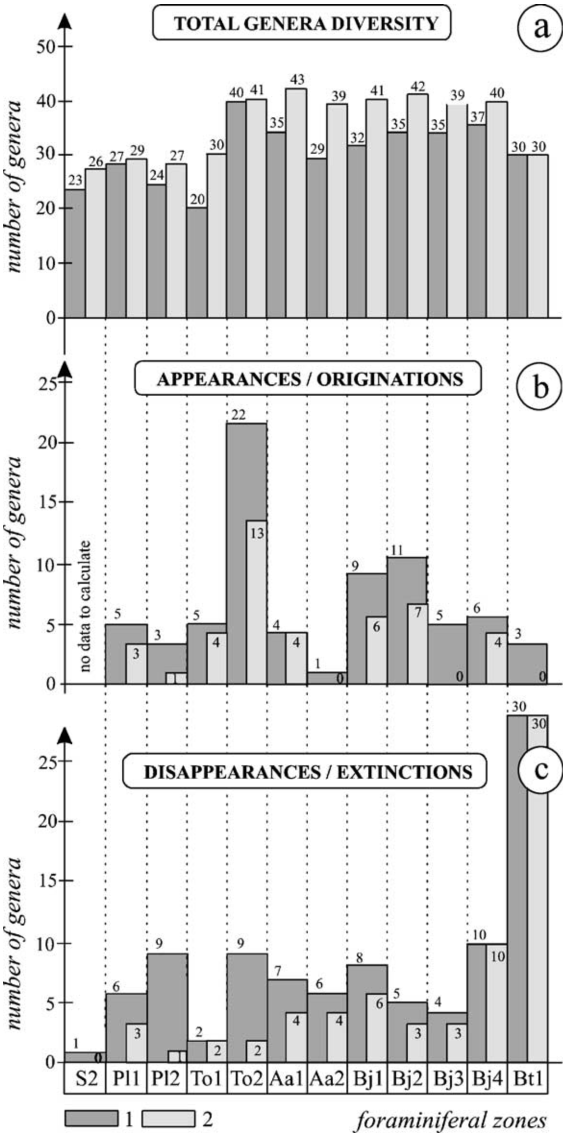

Fig. 5. Generic diversity of foraminifers (per zones): total diversity (a), appearances/occurrences (b), disappearances/ number of species going extinct (c). Columns: 1—“original”, 2—accounting for the Lazarus-effect. See Fig. 2 for foraminiferal zones abbreviations.

Middle Jurassic (Fig. 8). The highest values coincided mostly with intervals of significant diversifications. It is necessary to point out that the acceleration of singletons quantity in Bt1 zone is simply explained by the regional termination of development of all taxa, including just-originated ones, due to regression of the sea.

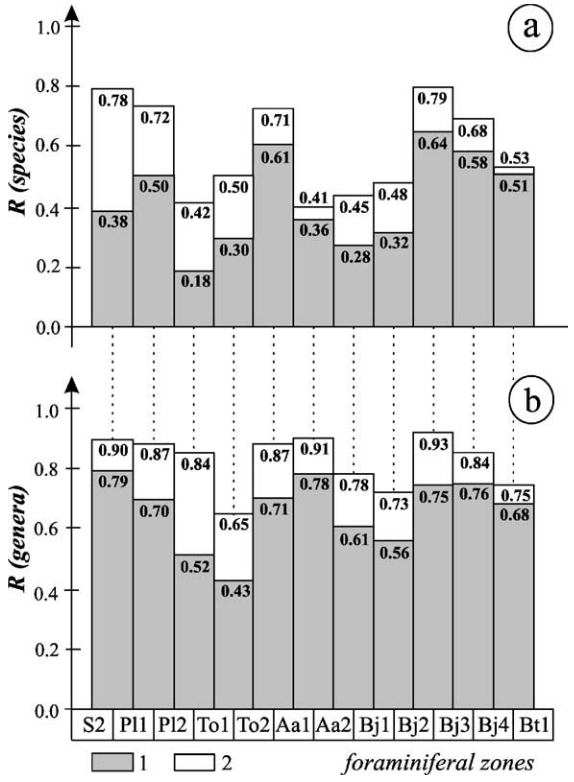

Fig. 6. R index changes (per zones): species (a), genera (b). Columns: 1—“original”, 2—accounting for the Lazarus-effect. See Fig. 2 for foraminiferal zones abbreviations.

Analysing all components of foraminiferal diversity, comparisons between curves calculated “as original”, and as accounting for the Lazarus-effect were made.

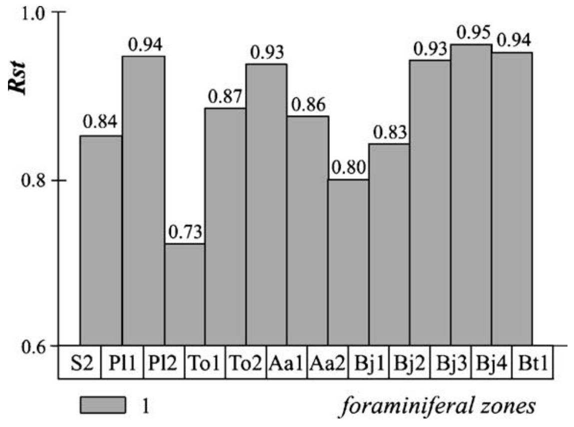

Fig. 7. R_{st} index changes (per zones). Column: 1—“original”. See Fig. 2 for foraminiferal zones abbreviations.

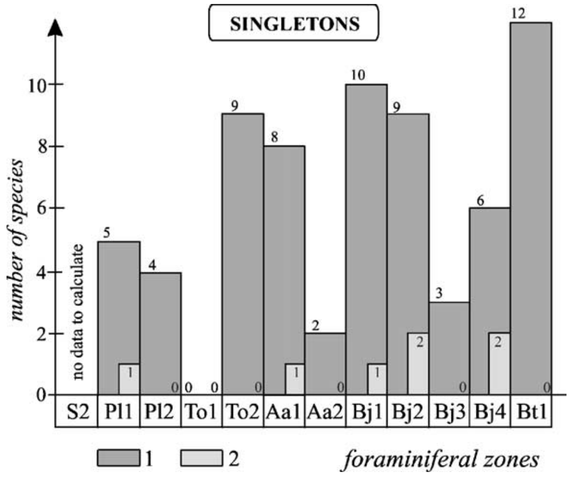

Fig. 8. Number of singletons accounting for the Lazarus-effect (per zones). Columns: 1—species, 2—genera. See Fig. 2 for foraminiferal zones abbreviations.

The result is quite surprising. Although the value of this effect is relatively high, that does not distinctly influence the estimation of diversity trends. The indices of the total diversity and *R* become higher, while, in contrast, rates of appearance/origination and disappearance/extinction decrease. Nevertheless, relative values follow the same trend in the succession of events. Therefore, the Lazarus-effect has not significantly influenced the results of foraminiferal diversity calculations from the Early–Middle Jurassic deposits of the NW Caucasus.

5. Interpretation and discussion

In the Early Jurassic marine biota were subject to a significant mass extinction coincided to major sea-level changes and appearance of anoxia, which led to the diversity drop in the Late Pliensbachian–Early Toarcian (Hallam, 1961, 1987; Jenkyns, 1988; Hori, 1993; Little and Benton, 1995; Aberhan and Fürsich, 1997, 2000; Hallam and Wignall, 1997, 1999; Guex et al., 2001; Harries and Little, 1999; Hylton et al., 2000; Jenkyns et al., 2002; Pálfy et al., 2002; Vőrős, 2002; Wignall, 2001; Ruban, 2004). Unfortunately, the impact of this event on foraminifers, although evidently established (Bassoullet and Baudin, 1994; Bassoullet, 1997), has been less extensively discussed.

The most comprehensive global data also suggest that diversity drop occurred in the Toarcian (Tappan and Loeblich, 1988). Tappan and Loeblich (1988) have reported a slight generic diversification in the Pliensbachian, followed by the relatively low-diversity Toarcian–Aalenian interval. A significant diversification occurred in the Bajocian, while in the Bathonian the number of genera slightly declined. Regional Early–Middle Jurassic trends of the foraminifers diversity dynamics calculated coincided with the global tendencies, except for the Bajocian diversification which has not been evidently documented in the NW Caucasus. Ruget and Nicollin (1997), based on the Western European data suggested diversification in the Sinemurian, which perhaps also took place in the NW Caucasus (S2 zone). They also pointed out two significant renewals of assemblages, which took place in the Early Pliensbachian and Early Toarcian. Data from La Rambla del Salto Section in Spain by Herrero (2001) also allow pointing a significant turnover at the Pliensbachian–Toarcian transition.

The first renewal is enigmatic in the NW Caucasus because in this particular case the Lazarus-effect brings misunderstanding into our results. But another strong intensification of the assemblages transformation (i.e., high values of the faunal transformation rate 'FTR' and the rate of transformation of taxonomic diversity structure 'TTDSR') is truly documented in the studied area. It was also noted by Ruget and Nicollin (1997) this transformation was taking place gradually. In the NW Caucasus, duration of this event covers a quite long interval represented by the Pl2−To2 zones.

Our results from the NW Caucasus support the idea of the significant mass extinction during the Early Jurassic that strongly influenced foraminiferal assemblages. The total species diversity of foraminifers in the To1 zone decreased to 55% of the level recorded in the Pl2 zone (Fig. 4). Thus, a strong extinction took place after the time representing the Pl2 zone. Taking into account the Lazarus-effect, it is suggested that this extinction was already initiated after the time representing Pl1 zone, but it was not intense (11%). We should nevertheless be aware that exclusion of the Lazarus taxa gives a very stable total species number in both Pl1 and Pl2 zones.

This described Pl2/To1 crisis was not that evident at generic level (disappeared 16.6% of recorded genera). Actually, this bioevent began earlier in the Pliensbachian when at the Pl1/Pl2 disappeared 11% of genera (Fig. 5). A rapid recovery of species and genera took place at the time of To2 zone and it completely compensated the previous demise. As a result, the To2 assemblage was even more diversified than pre-event assemblages at specific and generic levels. Thus, we can document duration of the main mass extinction of foraminifera in the NW Caucasus as the latest Late Pliensbachian to the earliest Middle Toarcian. A slight gradual (or stepwise) decrease has actually been recorded from the uppermost Lower Pliensbachian upwards, considering relations of foraminiferal zones to chronostratigraphic units (Fig. 2). In fact, it is quite likely that the above mentioned major extinction event took place around the *falciferum* ammonite zone, correlated with the middle part of the To1 zone. The pre-*falciferum* foraminiferal assemblage of the To1 zone was probably impoverished due to adverse environmental conditions associated with deposition of coarse-grained transgressive sediments, including conglomerates and sandstones (Fig. 3).

Bassoullet and Baudin (1994) have shown that a crisis affected foraminifers just at the Pliensbachian/Toarcian boundary and lasted during the time interval represented by the *tenuicostatum* ammonite zone; then, a rapid diversification began. Global data on all marine biota suggest the beginning of their decimation at the end of Pliensbachian (Pálfy et al., 2002). In both mentioned cases, the Middle Toarcian diversity acceleration is shown as a rapid event, but not as "overcompensated" as it was documented in the NW Caucasus. Should these differences in appearance of mass extinction be explained by any region-specific palaeoenvironmental changes? The answer depends much on what causes of the Pliensbachian-Middle Toarcian event we chose as principal ones. The most promising causes are major sea-level changes and anoxia (see Hallam and Wignall, 1997, 1999). The first seem to be more doubtful because during the interval of mass extinction sea-level changed cyclically, but crisis continued both during regression and during transgression. On the other hand, the peak of mass extinction in the global record corresponds to the maximum of transgression (Hallam and Wignall, 1999), while the Early Toarcian in the NW Caucasus was a time of beginning of the transgression, which reached its maximum in the Late Toarcian (Fig. 3).

Detailed estimation of influence of anoxia on extinction of foraminifera is not possible in the NW Caucasus due to lack of detailed sedimentological, geochemical and palaeoecological studies. Abundance of siderite concretions, "dispersed" pyrite grains, and dark to black color of shales (Granovskij et al., 2001; Rostovtsev et al., 1992; Ruban, 2002a, 2004) suggest that dysoxic to suboxic conditions could be documented in the whole Sinemurian–Bathonian interval. This might have been resulted from isolation of this marine silled basin from other adjacent Tethyan basins connected via relatively shallow seaways (Lordkipanidze et al., 1984).

The question is whether anoxia was the strongest in the Early Toarcian. We suppose that transgressive coarse-grained deposition took place under better oxygenation conditions associated with higher hydrodynamic conditions. Dysoxic conditions probably developed later, i.e., during successive deepening of the basin in the Middle–Late Toarcian. Surprisingly, these dysoxic conditions did not stop a rapid diversification of foraminifers. Most likely, these latest Middle–Late Toarcian well-diversified assemblages disprove an existence of anoxic, or suboxic bottom water conditions. Thus the factors controlling mass extinction of foraminifera in the regional NW Caucasus aspect is still enigmatic.

The Pliensbachian–Toarcian transition in the NW Caucasus is embraced by a short, but regionally recognized hiatus, which may partly or completely explain decline of the taxa number. We can hypothesize that the latest Pliensbachian regression caused a regional emergence of the basin and inevitable disappearance of all taxa. The successive Early Toarcian transgression brought about colonization of a new assemblage from adjacent basins. This new foraminiferal assemblage recorded in the To1 zone was less diversified (45% drop in total species number) and consisted of 17 "old species" and 22 newcomers. A total of 54 species did not come back, including 38 of those which did extinct (Fig. 4). We can therefore presume the extra-basinal faunal impoverishment took place either during existence the regional emersion level or during deposition of transgressive gravel and sandy facies (Figs. 2 and 3). The latter case is more likely because coarse-grained arenaceous facies were deposited during the time represented by the Early Toarcian *semicelatus* and *falciferum* zones. This is the time iterval, corresponding to *tenuicostatum* and *falciferum* zones, when the global extinction event associated with the rapid sea level rise and development of widespread anoxic conditions occurred (see Hallam, 1987; Little and Benton, 1995; Hallam and Wignall, 1999).

It is worthwhile to look at the recovery and recolonization patterns—diversification coinciding with late stages of the relative sea level rise and even its maximum (Figs. 3 and 4). It seems that a high sea level improved connections (via sills) with adjacent Tethyan and Boreal basins and enabled rapid immigration of new taxa and re-immigration of the Lazarus-taxa into the basin. Furthermore, during that time, adaptive radiation followed the global extinction event (Hallam, 1987; Hallam and Wignall, 1997). Overall deepening of the NW Caucasus basin and a shift from sandy to silt and clay facies during Middle and Late Toarcian also facilitated further colonization. It seems to be clear that moderately dysoxic conditions did not disturb this regional trend.

Another significant bioevent is recorded in the Aalenian. The Aa1 zone records a high total species diversity inherited from the To2 assemblage. A drastic species diversity drop (in about 2 times) is recognized in the Aa2 zone. (Fig. 4). Then a gradual recovery began correlated with the Bj1 and Bj2 zones. The causes of this Aalenian crisis are still enigmatic. Dysoxic conditions hypothesized by “black shales” with siderite concretions in the Early Aalenian, existed also in the Bajocian, but this did not stop the recovery of foraminifers. Major sea-water cooling documented regionally in the Aalenian (Jasamanov, 1978) should also be taken into account as a potential cause of extinction. It is also necessary to consider that the drop of species number in this case might also be explained by the incompleteness of palaeontological record.

Global analysis of the foraminiferal diversity suggests that a crisis took place around the Toarcian/Aalenian boundary, and that the recovery was finished in the Early Bajocian (Basov and Kuznetsova, 2000). Tappan and Loeblich (1988) showed that the total generic diversity decrease covered Toarcian and Aalenian. In the Aalenian a weak acceleration took place. Unfortunately, this low-resolution analysis limited to genera only does not allow making further conclusions. In several Spanish sections some patterns of the Aalenian crisis, including diminishing of the diversity, was also documented (Canales and Herrero, 2000; Canales, 2001).

The present results from the NW Caucasus suggest that the Aalenian–Early Bajocian event is not less pronounced than the preceding one observed in the Pliensbachian–Toarcian. This provides a matter of debate on reality of an additional mass extinction. There are no doubts that the results on foraminifers should be further supported by records from other regions and on different fossil groups. The global data on marine biota (Sepkoski, 1993; Markov, 2001) does not provide clear suggestions on particular events in the Middle Jurassic interval. On the other hand, an idea of new mass extinction has recently been supported by Tethyan brachiopods, which record strongly reduced Aalenian assemblages (Vőrős, 1993, 1995; Sulser, 1999; Ruban, 2003, 2004).

We can compare the NW Caucasus record with that of the Western Carpathians. The Pieniny Klippen Belt basin left a sedimentary record of dysoxic to suboxic facies dated from the Pliensbachian to the Late Bajocian (see Birkenmajer, 1977; Tyszka, 1994a,b). Unfortunately, the Early Toarcian Anoxic Event is not well recognized there due to tectonic gaps and limited availability of outcrops. This event is most likely preserved within the Toarcian–Lower Aalenian succession of dark grey to black flysch deposits with interbeding mudstones and sandstones (Birkenmajer, 1977; Birkenmajer and Tyszka, 1996). In contrast, the Aalenian/Bajocian event is documented much better in the Polish, Slovakian, and Ukrainian sectors of the Pieniny Klippen Belt (see Birkenmajer, 1977). It is represented by a relatively thin and widespread horizon of “black shales” with spherosideritic and phosphoritic concretions. This event has already been described as a “regional anoxic event” (Tyszka, 1994a, 1995). The event is associated with impoverishment of microfauna, including extinctions of several foraminiferal taxa, such as *Lenticulina d’orbignyi* (Roemer) (with *Astacolus*-like morphology), *Falsopalmula tenuistriata* Franke, and *Marginulinopsis dictyodes dictyodes* (Deecke). On the other hand, the same event was associated with first appearances (?originations) of some new taxa of agglutinated foraminifers, such as: *Trochammina globoconica*

Tyszka and Kaminski (Tyszka and Kaminski, 1995), *Verneuilinella pieninica* Tyszka and Kaminski, and *Miliammina gerochi* Tyszka (1997) as the first documented appearance of the genus *Miliammina* and the superfamily Rzehakinacea. The upper (Lower Bajocian) part of the "black shales" shows much higher diversity as a result of improvement of conditions, recovery of populations, and colonization by immigrants. It was connected with the sea-level rise, tectonic reorganization of the basin and, in consequence, deepening of sills and widening of seaways (Tyszka, 1994b, 1995).

The history of the Pieniny Klippen Basin seems to be closely linked to the Aalenian extinction event in the NW Caucasus. It also suggests that it is necessary to conduct detailed palaeoenvironmental studies in this Caucasus basin in comparison to other Tethyan and Boreal areas.

Both events documented in the Early–Middle Jurassic of the NW Caucasus are comparable in their intensity with the greatest crisis which occurred at the K/T boundary. Keller (2001) reported that 2/3 of planktic species were extinct at the K/T boundary. In our case 50–75% species went extinct at the Pl2/To1 zones boundary, and 50–60% at the Aa1/Aa2 zones boundary, i.e., around the *opalinum*/*murchisonae* ammonite zones boundary (?Early/Middle Aalenian). If we assume that less than 10% of the total benthic foraminiferal population was influenced at the K/T boundary (Keller, 2001), the Early–Middle Jurassic events could be regarded as significant in comparison to other mass extinction. However, it should be stressed that direct comparison of Jurassic and K/T events is irrelevant because almost a complete turnover of planktonic foraminifers at the K/T boundary is incomparable (see MacLeod, 1996; Keller, 2001).

6. Conclusions

Quantitative analyses of the Early–Middle Jurassic foraminifers taxonomic diversity performed for the NW Caucasus territory (Fig. 1) allow us to make several main conclusions:

(1) there is no stasis in the diversity dynamics of the Early–Middle Jurassic foraminifera recorded (Figs. 4–8);
(2) the diversity changed strongly at species level while at generic one it was relatively stable (Figs. 4 and 5) that is in agreement with other records of the Early Toarcian extinction event (see Hallam and Wignall, 1999; Hylton et al., 2000);
(3) the most intensive foraminiferal turnovers took place at the Pliensbachian/Toarcian boundary followed by the Toarcian, as well as, from the Middle Aalenian to the Early Bajocian (Fig. 6);
(4) significant diversity drops are documented in the Early Toarcian and in the Middle Aalenian. The first event corresponds to the well-known mass extinction. The second event probably represents a new mass extinction that appeared in the middle of Aalenian, possibly around the opalinum/murchisonae ammonite zones (Figs. 2 and 4);
(5) both extinction events were followed by extensive diversification intervals in the middle–late Toarcian and the latest Aalenian–Early Bajocian (Fig. 4);
(6) in spite of high value of the Lazarus-effect, it has not much influenced the estimation of main diversity components (Figs. 4–6).

Further micropalaeontological, sedimentological, and geochemical studies are necessary to explain the diversity dynamics observed, and to compare global and regional evolutionary and palaeoenvironmental changes.

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Appendix A. Supplementay data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.palaeo.2005.03.021.

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Late Paleozoic – Early Mesozoic Tectonic Activity within the Donbass (Russian Platform)

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Abstract: The Donbass is a large structural domain in the south of the Russian Platform, whose origin and tectonic evolution is still debated. Our new model suggests the Late Paleozoic and Early Mesozoic tectonic activity within the Donbass was controlled by the dynamics of the Northern Palaeotethyan Shear Zone. The Late Paleozoic dextral movements along the latter led to the detachment of the Ukrainian Massif and extension of the Donbass. The change to sinistral strike-slip in the Late Triassic resulted in regional contraction and folding.

Key words: Shear zone. Strike-slip. Contraction. Paleozoic. Mesozoic. Palaeotethys. Donbass.

Resumen: El Donbass es un gran dominio estructural de la Plataforma Rusa cuyo origen y evolución tectónica es objeto de debate. Se propone un nuevo modelo según el cual la actividad tectónica durante el Paleozoico tardío y el Mesozoico temprano fue controlada por la dinámica de la Zona de Cizalla Septentrional del Paleotetis. Los movimientos dextrales de esta Zona en el Paleozoico tardío condujeron al despegue del Macizo Ucraniano y a una extensión en el Donbass. El cambio a un movimiento en dirección sinistral durante el Triásico tardío dio lugar a una contracción y plegamiento.

Key words: Zona de cizalla. Strike-slip. Contracción. Paleozoico. Mesozoico. Paleotetis. Donbass.

Recent advances in global palaeotectonic reconstructions (e.g., Stampfli and Borel, 2002; Scotese, 2004; Torsvik and Cocks, 2004) ask for reconsideration of the traditional models of the tectonic evolution of particular regions. The Donbass (alternatively named Donets Basin or Donbass Fold Belt) is a large structural domain in the southern part of the Russian Platform (Fig. 1). It is located between two stable Precambrian blocks: the Ukrainian and Voronezh massifs, and their marginal structures. The Donbass is the eastern and the most "mature" domain of the elongated structure, which also includes the Dniepr-Donets Basin and the Pripyat Trough.The Donbass contains Carboniferous and Permian coal-bearing deposits several kilometres thick, which are strongly folded and disrupted by numerous faults. Overviews of the tectonic activity of the Donbass are presented by Laz'ko (1975), Stephenson et al., (1996), Maystrenko et al., (2003), Saintot et al., (2003), Stovba et al., (2003), Vai (2003), and Kostjutchenko et al., (2004).

Two significant questions arise about the Donbass: (1) why did intracratonic rifting occur in the Late Devonian-Carboniferous, and (2) what forces led to end-Triassic contraction? We present a new model to answer them.

Northern Palaeotethyan Shear Zone

A major shear zone was active on the northern margin of the Palaeotethys in the Late Paleozoic and Early

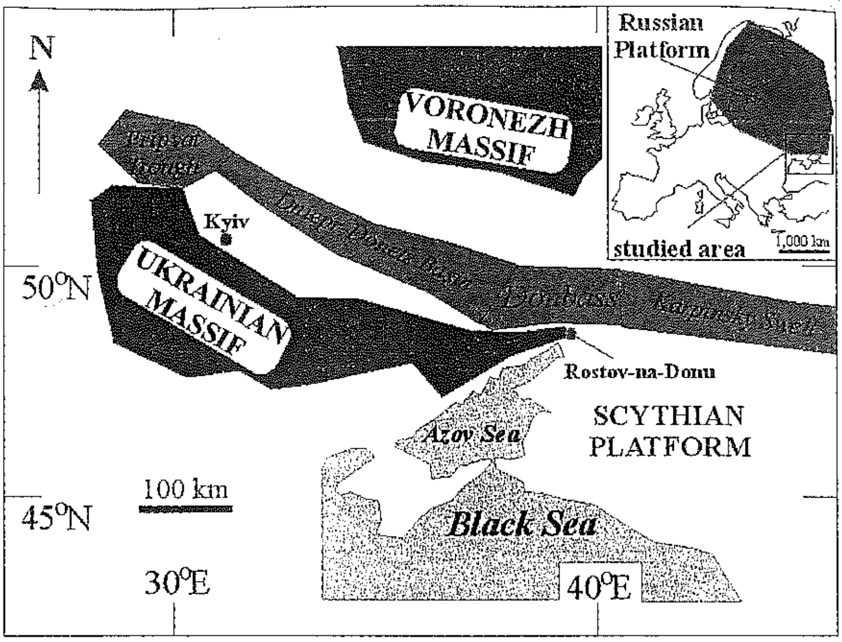
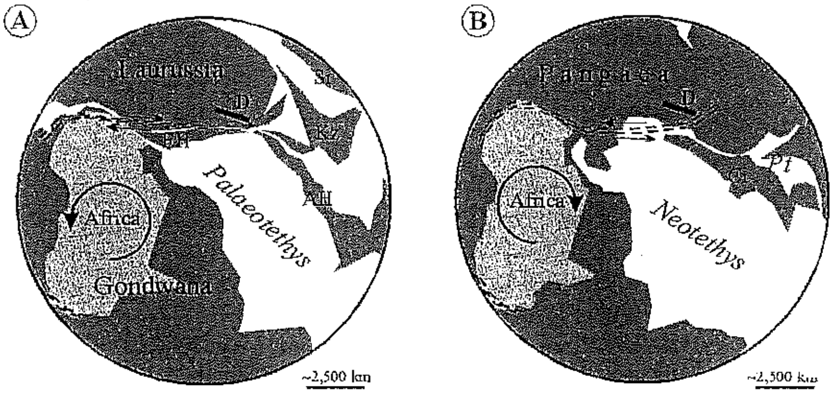


Figure 1. Main tectonic structures in the SW of the Russian Platform (simplified from Maystrenko et al., 2003). Areas of the Precambrian rock exposure are highlighted as black. The Late Paleozoic basins are highlighted as grey. Other areas (white) are covered by the Mesozoic-Quaternary deposits.



Bashkirian (320 Ma)

Norian (220 Ma)

Figure 2. Global palaeotectonic reconstructions for the Bashkirian (Pennsylvanian) (A) and the Norian (Late Triassic) (B); base palaeotectonic maps are strongly simplified from Stampfli and Borel (2002). Double dashed lines trace the shear zones related to the Africa rotation. Abbreviations: D – Donbass, Dniepr-Donets Basin and Pripyat Trough, EH – European Hunic Terranes, AH – Asiatic Hunic Terranes, Kz – Kazakhstan, Si – Siberia, Ci – Cimmerian Terranes, Pt – Palaeotethys.

Mesozoic (Arthaud and Matte, 1977; Swanson, 1982; Rapalini and Vizán, 1993; Lawver et al., 2002; Stampfli and Borel, 2002; Vai 2003; Garfunkel, 2004). Strike-slip movements along this zone were dextral since the Devonian until the Late Triassic (Fig. 2A), and since then it became sinistral (Fig. 2B) (Swanson, 1982; Rapalini and Vizán, 1993; Vai, 2003). These movements stopped in the Middle Jurassic (Rapalini and Vizán, 1993). These displacements were caused by the rotation of Africa (Rapalini and Vizán, 1993). The latter may be explained with the 'global wrench tectonics' (Storetvedt, 2003).

An intriguing question for further studies is connected with the hypothetical relation of motions along the Northern Palaeotethyan Shear Zone in the Late Paleozoic to the strike-slip deformations along the other major shear zone, which was located between the Laurentia and Baltica in the Middle Paleozoic (Lawver et al., 2002).

The global-scale dynamics of the Northern Palaeotethyan Shear Zone could enforce the tectonic evolution of the Donbass.

A tectonic model

Dextral strike-slip movements along the Northern Palaeotethyan Shear Zone occurred since at least the Visean (Lawver et al., 2002; Stampfli and Borel, 2002), but probably began a little earlier, in the Late Devonian,

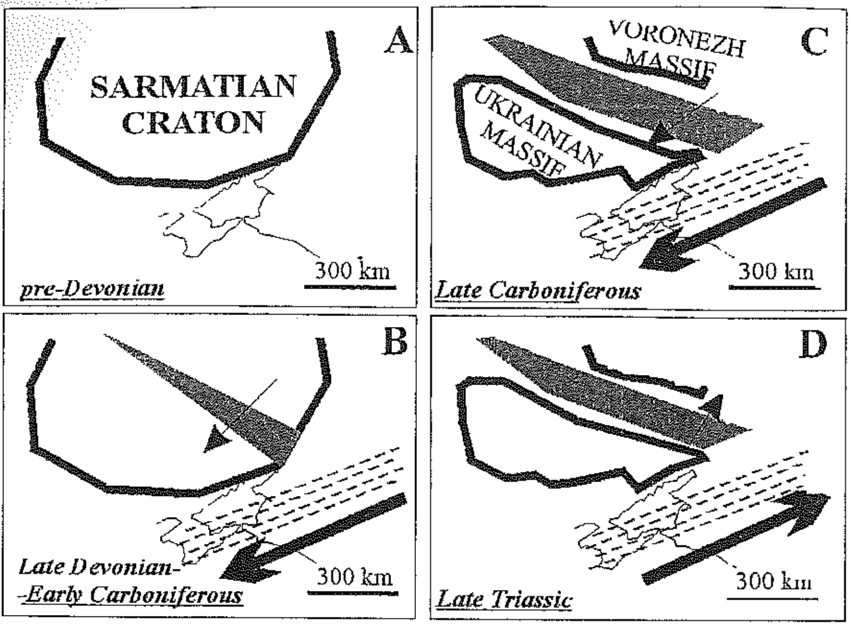


Figure 3. A model of the development of the Donbass from the pre-Devonian (A) to the Late Triassic (D). Dashed lines mark the shear zone. Thick arrows indicate the movements of the land masses in relation to the Russian Platform. Thin arrows indicate direction of the Ukrainian Massif movement. The Pripyat Trough, the Dniepr-Donets Basin and the Donbass are highlighted as grey.

when the European Hunic Terranes just docked to the Laurussia after the partial closure of the Rhenohercynian Ocean (Stampfli and Borel, 2002). These movements influenced intense fault activation and reactivation in European structures (Arthaud and Matte, 1977; Vai, 2003). It seems that dextral strike-slip movements were able to break up the previously stable Sarmatian Craton (Fig. 3A), detaching the Ukrainian Massif (Fig. 3B). This break-up occurred along the tectonically weak zone, which crossed the Sarmatian craton since the Precambrian (Stephenson et al., 1996). Kostjutchenko et al. (2004) estimated that the southeastern part of the Ukraininan Massif was transported 100-150 km off the craton margin. A number of basins (i.e., Pripyat Trough, Dniepr-Donets Basin and Donbass) were formed in this way (Fig. 3C). The regional extension, i.e. between the Ukrainian and Voronezh massifs, in the Carboniferous due to the continued detachment of the Ukrainian Massif seems to be a factor that controlled the deepening of the Donbass and, therefore, provoked deposition of thick sedimentary complexes. The rates of this regional extension, which was also marked by the magmatic activity, are discussed in Stephenson et al., (1996).

The change of movement direction along the shear zone in the Late Triassic caused contraction and folding within the Donbass. Sinistral strike-slip caused the Ukrainian Massif to move back towards the northeast, i.e. towards the Voronezh Massif. Such change of the motion of the Ukrainian Massif in the Triassic was also hypothesized by Kostjutchenko et al., (2004). As the 'mature' structure of the Donbass existed between the massifs, it was contracted (Fig. 3D), with folding of the sedimentary complexes. The main phase of shortening occurred in the Late Triassic (Maystrenko et al., 2003; Saintot et al., 2003; Stovba et al., 2003; Vai, 2003).

Thus, the Donbass evolution was influenced by the African geodynamics because the Northern Palaeotethyan Shear Zone evolution was controlled by the rotations of this continent (Rapalini and Vizán, 1993).

Our model explains the differences between the Pripyat Trough, the Dniepr-Donets Basin and the Donbass. The first is a relatively 'shallow' structure, while the second is deeper, but neither is as deep or deformed as the Donbass (Maystrenko et al., 2003). According to the model presented here, extension at the margin of the Sarmatian Craton increased towards the southeast, i.e., towards the most peripheral part of craton. The latter was especially strongly affected by the strike-slip movements, because shear zone was located along it.

The Donbass tectonic developments has analogues. Similar strike-slip movement along another great shear zone were documented in East Asia during the Mesozoic, particulary in Japan and Korea (Otoh and Yamakita, 1995; Otoh and Yanai, 1996; Yamakita and Otoh, 2000; Sasaki, 2001). The initial dextral movement changed to sinistral in the Early Cretaceous, which influenced the evolution of nearby regions.

Conclusion

Dextral strike-slip movements along the Northern Palaeotethyan Shear Zone detached the Ukrainian Massif from the previousely existed Sarmatian Craton, and the thick Late Paleozoic sedimentary complex was accumulated in the opened basin, i.e. in the Donbass. When these strike-slip movements became sinistral since the Late Triassic, the compression of the Donbass occurred, which was resulted into the strong deformation of the mentionned sedimentary complex.

It is important to note, that the Late Devonian-Triassic movements along the Northern Palaeotethyan Shear Zone also controlled the development of many basins in the North and South America, and Africa (Rapalini and Vizán, 1993).

Further studies are necessary to test deeply the model proposed in this paper.

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Diversity changes of the Brachiopods in the Northern Caucasus: a brief overview

Dmitry A. Ruban

The diversity of the brachiopods in the Northern Caucasus significantly fluctuated throughout the Paleozoic-Mesozoic. Weak diversifications occurred in the Middle Cambrian, Late Silurian – Early Devonian, and Late Devonian – Early Carboniferous. Since the Late Permian brachiopod assemblages became quite diverse. The maximum number of species was reached in the Rhaetian. The Permian/Triassic mass extinction and enigmatic Ladinian crisis, on the other hand, led to regional brachiopod demises. In the Jurassic – Early Cretaceous interval the diversity of brachiopods generally decreased. The strongest drops of species numbers occurred in the Toarcian and Berriasian following the Pliensbachian–Toarcian and end-Jurassic global mass extinctions, and in the Kimmeridgian due to the regional salinity crisis. It is evident that some of the regional brachiopod diversifications coincided with the development of rimmed shelves.

Key words: brachiopods, diversity, mass extinctions, reefs, Paleozoic, Mesozoic, Caucasus

Introduction

Global and regional diversity changes of the brachiopods have been discussed in numerous publications (Hallam 1987; Vörös 1995; Sulser 1996; Racki 1998; Alvarez and Modzalevskaya 2001; Harper and Rong 2001; Rong and Shen 2002). This fossil group has developed in high diversity during the Phanerozoic overall and seems to be one of the most important in the paleontological record.

This article is devoted to the brachiopods known from the Paleozoic and Mesozoic deposits of the Northern Caucasus (Fig. 1). Their diversity changes

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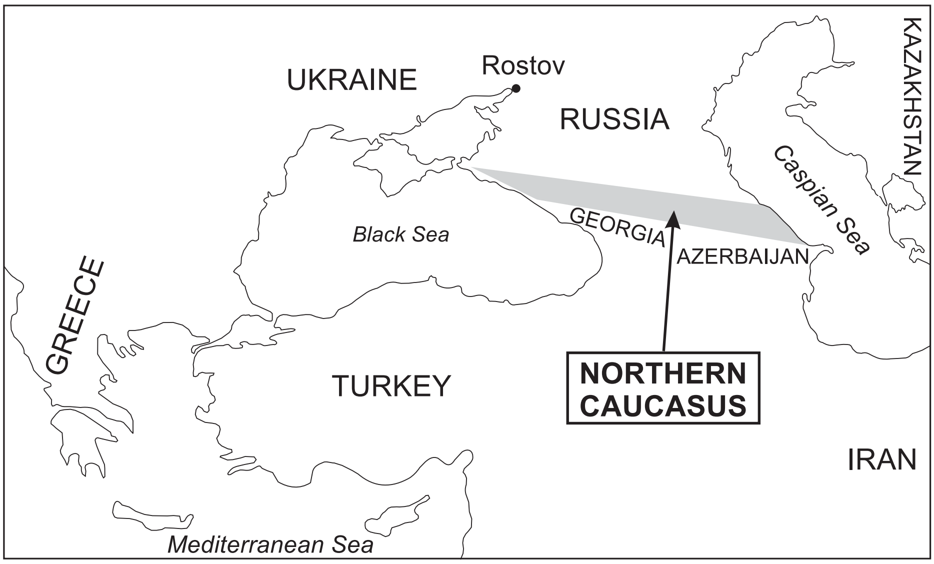

Fig. 1
Geographical location of the Northern Caucasus

have never been discussed before, except for the Jurassic as reviewed by Ruban (2003, 2004a). A comparison between the global and local Caucasian events shows some interesting results.

Stratigraphic setting

According to the paleotectonic reconstructions by Stampfli and Borel (2002) and Scotese (2004), the Northern Caucasus was located on the active northern Paleotethyan and then Neotethyan margin. A thick sedimentary succession was accumulated there during the Paleozoic–Mesozoic (Fig. 2). The Paleozoic deposits are exposed in the central part of the studied region, while the Triassic strata crop out in its western part. The Jurassic–Cretaceous sediments are distributed widely within the entire Northern Caucasus.

During the Early–Middle Paleozoic, deposition of clastics and shale occurred. A major regional hiatus embraced the Ordovician. Episodes of carbonate sedimentation took place in the Middle Cambrian, late Silurian and Late Devonian. Volcanics and volcanoclastics are common in the sedimentary succession, especially in the Devonian. Non-marine deposition dominated in the Pennsylvanian–Middle Permian. Carbonates were accumulated in the end-Permian, and an episode of reef growth is known from the Late Changhsingian.

During the Early Triassic–Anisian interval, carbonate deposition dominated, while accumulation of clastics and shale prevailed in the Ladinian–Carnian. Carbonate deposition and reef growth characterized the Norian–Early Rhaetian.

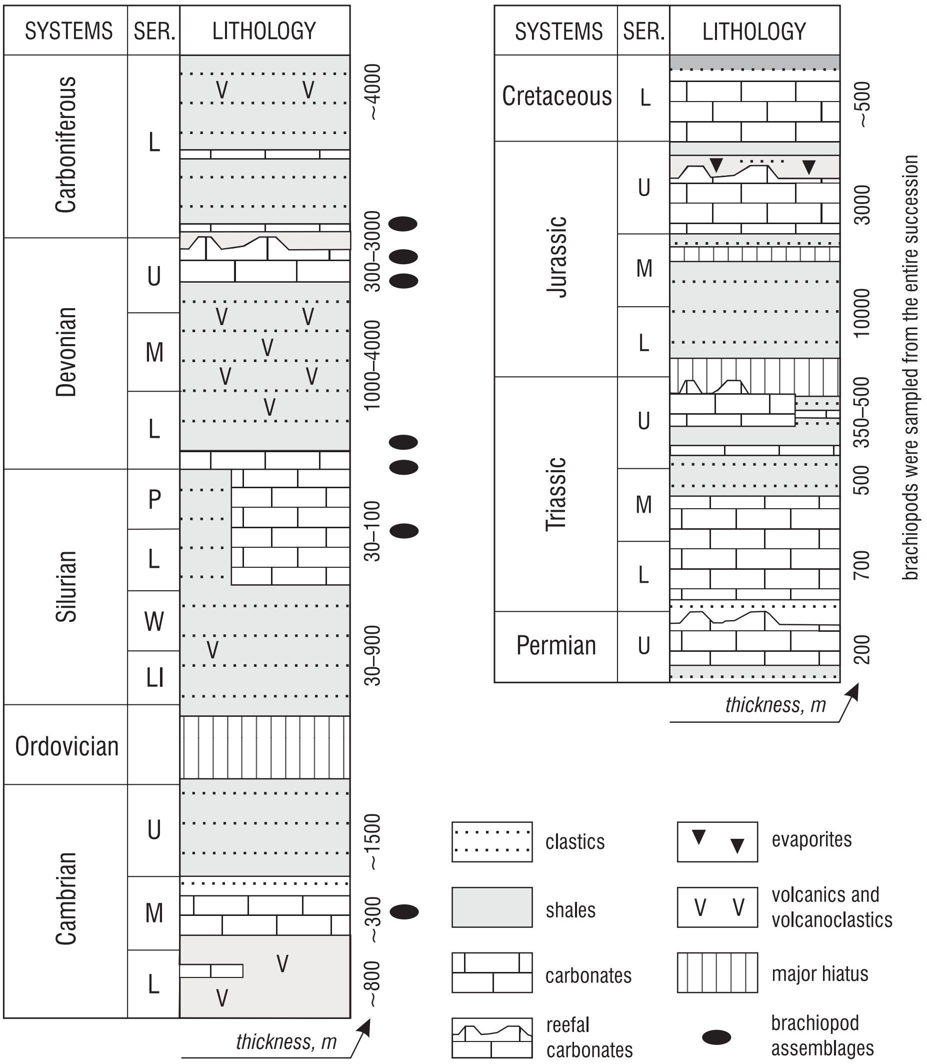

Fig. 2
A generalized lithologic section of the Paleozoic–Mesozoic of the Northern Caucasus

A major regional hiatus spans the latest Triassic–earliest Jurassic. Until the Bathonian, deposition of clastics and shale was common in the Northern Caucasus, while after the Bathonian hiatus, ie., in the Callovian, a large carbonate platform appeared (Kuznetsov 1993; Ruban 2005). The regional salinity crisis, introduced by Ruban (2006), led to the deposition of evaporites together with "color-beds" and red-beds in the Late Jurassic. Carbonate sedimentation was restored in the latest Jurassic and dominated during the Early Cretaceous. In the Aptian-Albian clastics and shale were deposited.

Materials and methods

To obtain as much information on stratigraphic distribution of the brachiopods as possible a data compilation from different sources was attempted. These sources are cited below.

Cambrian: Paffengol'ts (1959, 1965), Tchernysheva (1968); Silurian: Obut et al. (1988); Devonian: Rzhosnitskaja (1968), Kizeval'ter and Robinson (1973), Zanina and Likharev (1975), Obut et al. (1988); Carboniferous: Kotlyar (1977); Permian: A.D. and K.V. Miklukho-Maklaj (1966), Likharev (1968), Kotlyar (1977), Kotlyar et al. (1999, 2004); Triassic: Dagis (1963, 1974), Dagis and Robinson (1973), Rostovtsev et al. (1979); Jurassic: Makridin and Kamyshan (1964), Rostovtsev et al. (1992), Prosorovskaya (1993a, b), Ruban (2004a); Early Cretaceous: Smirnova (1972).

Brachiopods were sampled in numerous sections and small outcrops located within the Northern Caucasus.

Finally, data of two distinct kinds have been collected. The brachiopod assemblages of the Paleozoic are poorly studied and they show little diversity (except for the Permian). The data on them are incomplete, i.e. further sampling is necessary. In some sources only the presence of the brachiopods is stated without indication of the specific taxa. Thus only a general appreciation of their diversity changes in this interval is possible, without quantitative estimates of diversity. However, to take these data into consideration seems to be useful, particularly to simplify further comparisons between the Northern Caucasus and other regions. As for the Permian, numerous brachiopod taxa were collected in the Northern Caucasus, but published sources do not contain their full lists and should be further normalized. Therefore the available data is not sufficient to estimate the Permian diversity quantitatively.

On the other hand the compiled data on the Mesozoic interval is more complete and detailed (see Appendix). Those assemblages have been well studied. This allows quantitative estimations of the diversity changes per stage to be made.

Paleozoic brachiopods

The Cambrian–Carboniferous diversity of brachiopods species in the Northern Caucasus is characterized by very low numbers. Most known brachiopod finds are restricted to the carbonate facies (Fig. 2).

Three weak diversifications have been documented in the Cambrian–Carboniferous history of brachiopods in the Northern Caucasus (Fig. 3). The first one took place in the Middle Cambrian, when *Acrotreta gerassimovi* Lermontova appeared. The global "explosion" of biodiversity occurred at the same time interval (Geyer 1998).

For the second time, brachiopods appeared in the Northern Caucasus in the Ludlow–Přidoli. Poor assemblages existed in the Lochkovian (*Cingulodermus* ex gr. *superstes* (Barrande), *Clorinda pseudolinguifera* (Kozlowski) and *Janius* ex gr.

irbitensis (Tschernyshova)) and Pragian (*Ivdelina* (*Procerulina*) ex gr. *procerula* (Barrande)) before brachiopods disappeared.

The Frasnian assemblage was also very poor. It included *Atrypa posturalica* Markowskii, *Gypidula comis* Owen, *Hypothyridina cuboides* Sowerby and *Spinatrypa* ex gr. *bifidaeformis* Tschernyshova. A radiation strengthened in the Famennian, when 11 species appeared. They are *Cyrtospirifer verneuili* (Murchison), *Cyrtospirifer* cf. *calcaratus* Sowerby, *Cyrtospirifer* cf. *archiaci* Murchison, *Cyrtospirifer* cf. *postarchiaci* Nalivkin, *Isopoma brachyptycta* (Schnur), *Productus* sp., *Productella* ex gr. *subaculeata* (Murchison), *Productella calva* var. *multispinosa* Sokolskaya, *Productella calva* var. *koscharica* Nalivkin, *Pugnax janischevskii* Rozman and *Rhipidiorhynchus* ex gr. *livonicus* Buch. The Frasnian/Famennian transition coincided with a significant change in the taxonomic composition of assemblages. In contrast to the Frasnian, assemblages with *Cyrtospirifer* and *Productella* dominated in the Famennian. In some other regions, e.g. in South China, the Famennian assemblages were also characterized by diverse cyrtospiriferids (Ma et al. 2001). In the East European Platform area, however, cyrtospiriferid-dominated communities appeared earlier – in the Late Frasnian (Sokiran 2001). One of the biggest mass extinctions, the Frasnian–Famennian event, strongly stressed brachiopods (McGhee 1996; Hallam and Wignall 1997; Racki 1998; House 2002); but in the Northern Caucasus we cannot document the patterns of this event due to the low-resolution data. In the Early Carboniferous, the poorest assemblage with a unique *Spirifer* cf. *distans* Sowerby existed. Probably the poverty of those faunas may be explained by the mass extinction that occurred at the Devonian/Carboniferous boundary (Hallam and Wignall 1997). Such a hypothetical conclusion, however, should be further verified.

After a long period of non-marine deposition in the Late Carboniferous–Middle Permian, the sea transgressed into the Northern Caucasus in the Late Permian, which reached its maximum in the Changhsingian. At the same time a strong diversification of brachiopods began. Dozens of species appeared (Kotlyar et al. 1999, 2004). However, at the Permian–Triassic transition, a significant crisis occurred. It led to the short-term but total disappearance of the brachiopods. They are absent in the deposits of the Abagskaja Formation, the age of which is established as the latermost Permian–earliermost Triassic. This event is evidently connected with the global mass extinction, which took place at this time (Hallam and Wignall 1997; Harper and Rong 2001; Erwin et al. 2002; Shen and Shi 2002).

Mesozoic diversity changes

In the Early Triassic brachiopod assemblages were of extremely low diversity. The presence of ?*Crurithyris extima* Grant 1970 (Rostovtsev et al. 1979), which is a characteristic taxon of the lowermost Triassic (Grant 1970), suggests that the repopulation of this group was initiated just after the mass extinction. A major diversification occurred in the Anisian, when a recovery after the Permian/

Triassic crisis was completed (Fig. 3). The Anisian is an important stage in the evolution of shelly benthos, when its diversity rose extraordinarily (e.g. Komatsu et al. 2004).

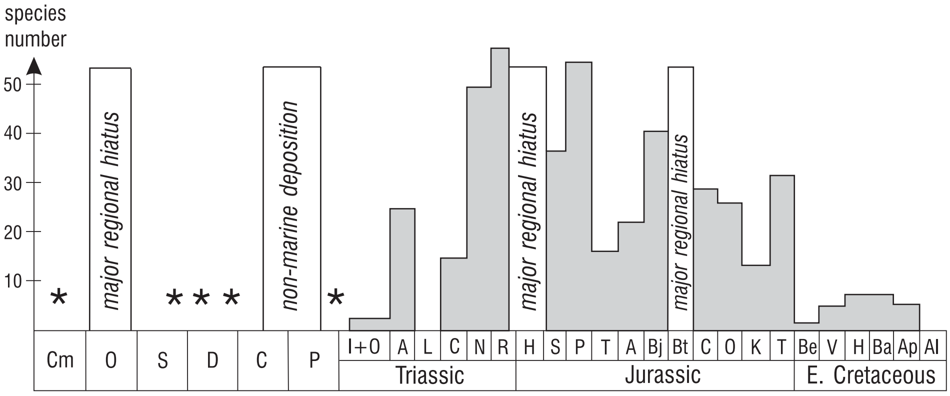

Fig. 3
Total species diversity of the Paleozoic–Mesozoic brachiopods of the Northern Caucasus. Asterisks mark quantitatively unestimated diversifications

However, brachiopods abruptly disappeared already in the Ladinian (Fig. 3). All pre-Ladinian taxa became extinct. As for other fossil groups, i.e. bivalves, ammonoids, foraminifers, they are well represented in strata of this age (Rostovtsev et al. 1979). Moreover, there are no strong differences between the Ladinian and Carnian intervals in the sedimentary succession, although the brachiopod diversity in the Carnian was high enough (Fig. 3). This suggests that differences in the diversity between the mentioned stages cannot be explained by the differences in preservation potential. The existing paleoenvironmental studies (Gaetani et al. 2005) also do not provide clear evidence about the possible causes of the enigmatic Ladinian crisis among brachiopods. In the other regions, particularly in the Alps and Jura Mountains, the Ladinian brachiopods were also not diverse, but their species number is comparable with that in the Anisian (Sulser 1999).

In the Carnian to Rhaetian, a new radiation of brachiopods took place, and by the end of this interval the diversity in the region reached its Phanerozoic high point. The most significant changes took place in the Norian, when the species quantity increased more than threefold.

After a depositional hiatus, which embraced Late Rhaetian, Hettangian and Early Sinemurian, the repopulation of brachiopods began once again. The maximum of diversity was rapidly reached in the Pliensbachian (Fig. 3). But a significant extinction occurred in the Toarcian, which corresponded to the a global event (Hallam 1986, 1987; Vörös 1995). This event was already documented in the studied region (Ruban 2004a, 2006; Ruban and Tyszka 2005).

A new increase of the species quantity continued during the Aalenian–Bajocian. The Bathonian sediments are missing due to a major regional hiatus. In the Callovian, the diversity of brachiopods was relatively high. It decreased slowly in the Oxfordian, but in the Kimmeridgian it dropped abruptly. This is possibly explained by unfavorable conditions due to increasing aridity, shalloving, and a salinity crisis in the Caucasus (Jasamanov 1978; Ruban 2006). However, brachiopods survived this crisis and rapidly recovered after it, in contrast to bivalves (Ruban 2003, 2006). Prosorovskaya (1996) stated the ability of brachiopods to radiate rapidly even during short time intervals with favorable environments.

The last significant diversification of brachiopods in the Northern Caucasus is documented in the Tithonian. A major drop in species number after this age seems to correspond with another global mass extinction, although the last one is sometimes considered to be doubtful (Hallam 1986; Hallam and Wignall 1997; Wignall 2001; Ruban 2004b).

The Early Cretaceous assemblages were very poor. A slight diversity increase is observed during the Berriasian–Hauterivian. Thereafter a similarly weak decreasing trend was established. In the Albian the brachiopod record terminates.

Discussion

The diversity changes of brachiopods in the Northern Caucasus was controlled by several factors.

The regional sea-level fluctuations have been reconstructed only for the Jurassic (Ruban 2006). A comparison between the regional sea-level changes and the brachiopod diversity during this time interval suggests that there are no direct links between them (Fig. 4). The diversity rises in the Pliensbachian and Bajocian corresponded to transgressions, while the maximum of species number in the Tithonian coincided with a regression. The Toarcian and Kimmeridgian diversity drops occurred during transgressions. Undoubtedly sea-level changes were able to influence the brachiopod diversity, while other regional factors (such as salinity crises or anoxia) were superposed on them.

It is very interesting to note that diversity maxima in the Famennian, Changhsingian, Norian-Rhaetian, and Callovian–Oxfordian corresponded to the regional development of rimmed shelves (term after Ginsburg and James, 1974; Read 1985), i.e. carbonate platforms bounded by reefs (Khain 1962; Dagis and Robinson 1973; Kuznetsov 1993; Ruban 2005). Thus, the absence of reefs in the Frasnian may explain why brachiopods were not so diverse in the Northern Caucasus as in the other regions, where Frasnian reefs were abundant (Copper 1994, 2002; Webb 1996; Droser et al. 2000; Baliński et al. 2002). Just after the appearance of reefs in the Famennian in the central part of the studied region (Fig. 2), brachiopod diversity accelerated significantly. The recent observation

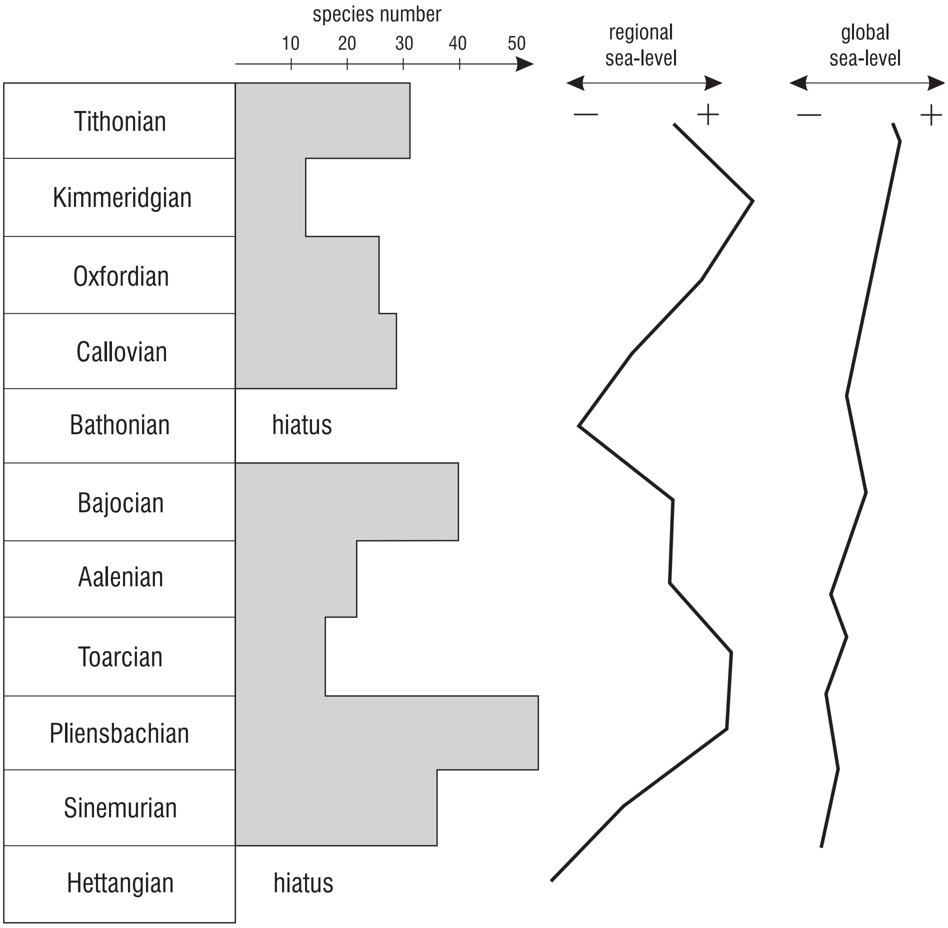

Fig. 4
Jurassic sea-level changes (after Ruban 2006) and brachiopod diversity. The global sea-level fluctuations, given for general reference, are simplified from Haq et al. (1987)

that the Famennian reefs grew spectacularly (Webb 2001, 2002; Shen and Webb 2004) may lead to significant changes in how we visualize the Late Devonian biotic crises.

Two interesting kinds of paleoenvironmental events strongly influenced the regional diversity of the brachiopods. The first is the above-mentioned regional salinity crisis, which occurred in the Kimmeridgian–Tithonian (Jasamanov 1978; Ruban 2006) and diminished the species quantity. Another phenomenon is that of oxygen-depleted conditions. Zakharov et al. (1999) explained the regional faunal crisis at the Permian–Triassic transition by anoxia. Efendiyeva and Ruban (2005) reconstructed the regional chronology of the Jurassic dysoxic and anoxic events. Dysoxic environments dominated during the entire Late Pliensbachian–Middle Aalenian, as well as the Bajocian. When oxygen depletion strengthened in the Toarcian (when anoxia occurred) brachiopod diversity dropped. Ruban (2004a, 2006) and Ruban and Tyszka (2005) also commented on the influences of the Toarcian anoxia on the Caucasian marine biota.

The regional preservation potential might have heightened our estimations of the brachiopod diversity. This is especially significant for the Paleozoic because brachiopods of that age are restricted to the carbonate facies (Fig. 4). As for the Mesozoic, it is difficult to speculate about this because of distinct facies in the each stratigraphic interval. However, high diversity, documented in the Anisian, Norian–Rhaetian and Oxfordian, corresponds to the intervals of the carbonate sedimentation. It is also very interesting to note a high diversity in the Callovian, although those deposits are strongly condensed (Ruban 2004c).

Conclusions

Summarizing the material presented above it is possible to conclude that the diversity of brachiopods in the Northern Caucasus fluctuated strongly through Cambrian–Early Cretaceous times (Fig. 3). The key points in the regional brachiopod diversity changes are located in the Famennian (first significant diversification), Late Permian (the beginning of rapid species number acceleration, followed by mass extinction), Rhaetian (the highest diversity), and in the end-Liassic (the initiation of gradual extinction).

The data should be further updated, and other diversity patterns (e.g. origination, extinction rates, etc.) need to be taken into consideration. It will be necessary to discuss the problem of the relationship of the brachiopod diversity dynamics to regional paleoenvironmental changes (not yet well established, and not dicussed completely in this paper), as well as rates of faunal immigrations/emigrations during the Paleozoic and Mesozoic. Special attention should be paid to the explanation of the Ladinian crisis among this fossil group.

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Appendix

Stratigraphic distribution of the Mesozoic brachiopods in the Northern Caucasus (see text for literature sources). Number of species of each genera in the Triassic–Lower Cretaceous stages is indicated

TRIASSIC

GENERA	Induan+ Olenekian	Anisian	Ladinian	Carnian	Norian	Rhaetian
"Rhynchonella"		1				
Abrekia	1					
Adygella				1	1	1
Adygelloides						1
Amphiclina					1	2
Ampliclinodonta					1	
Angustothyris		1				
Aulacothyropsis				1		4
Austriella					1	
Austrirhynchia						2
Balatonospira				1		
Bobukella				1	1	
Caucasorhynchia					1	1
Caucasothyris					1	
Coenothyris		2				
Costirhynchia		1		1		
Costispiriferina		1				
Crurirhynchia					1	
Crurithyris	1					
Cubanothyris					2	2
Decurtella		2				
Dinarispira		2				
Dioristella		1		1		
Euxinella					3	6
Fissirhynchia					1	1
Guseriplia						2
Holcorhynchella		1				
Koeveskallina		1				
Koninckina				1	1	
Laballa					2	3
Lepismatina						1
Lobothyris					2	
Majkopella						3
Mentzelia		1		2	1	2
Moisseievia					3	
Neoretzia						3
Neowelerella	1					
Norella		1				
Oxycolpella					3	2
Pexidella		1			1	
Piarorhynchella		1				
Pseudocyrtina					1	
Pseudorugitella					2	2
Punctospirella		1				
Rhaetina				1	6	4
Rhimirhynchopsis					1	1
Robinsonella						1
Sinucosta		1			1	1
Spinolepismatina					1	
Sulcatinella		2				
Sulcatothyris				1		
Tetractinella		1				
Thecospira					1	
Thecospiropsis					1	
Triadithyris					2	1
Trigonirhynchella					1	2
Volirhynchia		2				
Wittenburgella					1	1
Worobievella					1	1
Zeilleria					3	6
Zugmayerella						1

JURASSIC

GENERA	Si	Pl	To	Aa	Bj	Bt	Cl	Ox	Km	Tt
"Rhynchonella"	2	1	2	2	2					
"Spiriferina"	1	1								
"Terebratula"							1	1		2
Acanthorhynchia								1		
Acanthothyris				1	3					
Aromasithyris							1	2	1	
Aulacothyris		2	1				2			
Bodrakella		1								
Calcirhynchia	1	1								
Calvirhynchia					1					
Capillirhynchia				1			1			
Cardinirhynchia							1			
Caucasella							1			
Caucasorhynchia		1								
Cererithyris					1					
Cheirothyris										1
Cincta		2								
Cirpa		1								
Colosia								2	1	
Cryptorhynchia					1					
Cuersithyris		1								
Cuneirhynchia		4								
Curtirhynchia			1							
Digonella		1						1		
Disculina		1								
Dorsoplicathyris							2			
Ferrythyris							1			
Flabellirhynchia		1	1							
Furcirhynchia	1			1	1					
Gibbirhynchia	1	2	1							
Goniothyris							1		1	1
Grandirhynchia			1	1	1					
Gusarella							1			
Heimia					1					
Homoeorhynchia	1	2	1							
Ismenia										2
Ivanoviella							3	1		
Juralina								5	1	2
Lacunosella							1	3	2	1
Linguithyris	1				1					
Liospiriferina	2	4	1							
Loboidothyris					1			1		
Lobothyris		3	1	2	2					
Lophothyris									1	
Monsardithyris					1					
Monticlarella									1	2
Morrisithyris					1		1			
Nucleata								2	1	1
Paruirhynchia				1						
Piarorhynchia	1	4	1							
Postepithyris									1	
Praemonticlarella			1							
Prionorhynchia	1	1	1							
Pseudogibbirhynchia			2	2	2					
Ptyctorhynchia			1							
Ptyctothyris					1		3			
Quadratirhynchia					2					
Rhactorhynchia					3					
Rhynchonelloidea				1						
Rhynchonelloidella							1			
Rimirhynchia		1								
Rudirhynchia	2									
Rugitela					1					
Scalpellirhynchia	1									
Securina	1	1								
Sellithyris								1	1	
Septaliphoria							2	2		6
Somalirhynchia								1		
Sphaeroidothyris					2					
Spiriferina	4	3								
Squamirhynchia		1								
Stolmorhynchia				4	2		1			
Striirhynchia					1					
Stroudithyris				1						
Tchegemithyris							2			
Terebratula								1		

LOWER CRETACEOUS

GENERA	Be	Va	Ha	Ba	Ap	Al
Belbekella	1		1	4	1	
Belothyris			1	2	2	
Orbirhynchia				1	1	
Peregrinella			1			
Praelongithyris					1	
Psilothyris		2				
Sellithyris		1	3			
Sulcirhynchia	1					
Symphythyris			1			
Terebrataliopsis		1				
Weberithyris		1				

stage abbreviations: Be – Berriasian, Va – Valangianian, Ha – Hauterivian, Ba – Barremian, Ap – Aptian, Al – Albian

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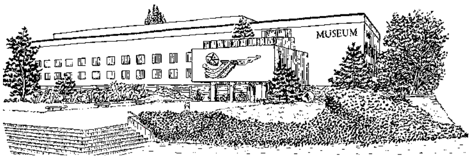

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Diversity dynamics of the Triassic marine biota in the Western Caucasus (Russia): A quantitative estimation and a comparison with the global patterns

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Abstract
Diversity dynamics of discovered Triassic marine biota in the Western Caucasus (Russia) have been studied quantitatively. The stratigraphic distribution of 422 species was taken into account. Foraminifers and brachiopods, followed by ammonoids, were the most diverse groups in comparison with bivalves, corals, algae and sponges. Total diversity, relatively low in the Early Triassic, strongly increased in the Anisian. Diversity decreased significantly in the Ladinian. Repopulation began in the Carnian, and the Norian is characterized by a new «explosion» of diversity. Weak diversification continued in the Early Rhaetian. The faunal diversity documented in the Western Caucasus generally reflected both the influences of the regional palaeoenvironmental changes and the global evolutionary patterns of the Triassic marine biota.

Keywords
Diversity, marine biota, ammonoids, bivalves, brachiopods, foraminifers, Triassic, Western Caucasus.

1. INTRODUCTION

The present knowledge on the Phanerozoic diversity dynamics of the marine biota on a global scale is summarized in the well-known articles of Benton (1995), Peters & Foote (2001) and Sepkoski (1993). But it is always important to evaluate the fossil diversity for the particular territory to test the regional appearance of the global patterns.

All Triassic stages (Induan to Rhaetian) are represented within the Western Caucasus – a region located on the northern periphery of the Neotethys Ocean (Golonka, 2004; Gaetani *et al.*, 2005; Stampfli & Borel, 2002). Its palaeontologic record is characterized by numerous taxa of ammonoids, brachiopods, bivalves, foraminifers, corals, algae and sponges. The study of the diversity dynamics (i.e., changes of the total number of taxa) of Triassic marine biota in the Western Caucasus and its comparison with the global record is attempted in this article.

2. GEOLOGIC SETTING

In the Triassic, the Western Caucasus was situated on the northern periphery of the Neotethys Ocean (Gaetani *et al.*, 2005; Golonka, 2004; Stampfli & Borel, 2002) (Fig. 1). Triassic strata and fossils of this region have been described by Dagis (1963, 1974), Dagis & Robinson (1973), Diakonov *et al.* (1962), Efimova (1991), Gaetani *et al.* (2005), Jaroshenko (1978), Paffengoltz (1959), Popov (1962), Prozorovskaja (1979), Rostovtsev *et al.* (1979), Shevyrjov (1990) and Vuks (2000).

The Triassic deposits of the Western Caucasus unconformably overlie the Paleozoic strata and are subdivided into four groups (Fig. 2). The most ancient is the Tkhatchskaja Group (~700 m) that consists of dominated carbonate deposits. The Sakhrajskaja Group (~500 m) consists of shales with clastic beds; its lower and upper contacts are marked by unconformities. The upper groups were deposited in the same time and they are named as the Khodzinskaja Group, composed of carbonates including reefs (the total thickness is ~500 m), and the Khadzhokhskaja Group, embracing shales with clastic and carbonate beds (total thickness is ~350 m). The Upper Rhaetian-Lower Liassic interval corresponds to a major regional hiatus.

3. MATERIALS AND METHODS

The compilation of all available palaeontologic information provided a database of the Western Caucasus marine biota (see Appendix). The principal sources are monographs and articles of Dagis (1963, 1974), Dagis & Robinson (1973), Diakonov *et al.* (1962), Efimova (1991), Gaetani *et al.* (2005), Jaroshenko (1978), Paffengoltz (1959), Prozorovskaja (1979), Rostovtsev *et al.* (1979), Shevyrjov (1968) and Vuks (2000).

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Fig. 1 : Geographical and palaeogeographical location of the studied region (global palaeogeography at 210 Ma after Scotese, 2004, position of the Western Caucasus is marked by black circle).

In its final version the database contains the distribution per stage of 422 taxa. When possible, taxonomic revisions of taxa were made to avoid under- or overestimation of the diversity due to the synonymy errors. Species identified has "sp." only were excluded, because it is difficult to understand their relation to the completely identified species mentioned in other papers. Species defined with "cf." and "aff." were considered as concrete species to avoid underestimation of the diversity. Similarly suggestions were made by Alba *et al.* (2001).

The analysis of the diversity dynamics of the Triassic marine biota of the Western Caucasus attempted herein is based on the simple procedure of calculation the total number of species in each stage.

4. DIVERSITY DYNAMICS OF THE MARINE BIOTA

The total number of analyzed species of the Triassic marine biota in the Western Caucasus is 422. In decreasing order, foraminifers (175 species), brachiopods (122 species), ammonoids (69 species), bivalves (31 species), corals (17 species), algae (5 species) and sponges (3 species) were registered. The evaluated faunal diversity dynamics is presented in Fig. 3. Diversity changes within the particular fossil groups partly differ from that observed within the total biotic dynamics (Fig. 4).

The Early Triassic diversity remained relatively low. However, the palaeoenvironments, characterized by the dominated low energy carbonate sedimentation below wave base (Gaetani *et al.,* 2005), were favourable enough for the rapid recovery of the marine biota after the crisis, which occurred at the Permian/Triassic boundary. Such assumption is supported by the presence of 35 Early Triassic species, known in the Western Caucasus, which number is not so little. During the Anisian, a strong diversification, recognized in ammonoids, brachiopods, foraminifers and to a lesser scale in bivalves, reached a total species number exceeding 100. The palaeoenvironments were the same as in the Early Triassic (Gaetani *et al.*, 2005). This diversification is especially an interesting event, because the distribution of the Anisian deposits became restricted in comparison with the preceding time intervals (Rostovtsev *et al.*, 1979).

In the Ladinian, diversity decreased significantly in most groups, although less pronounced in bivalves and foraminifers. Species numbers declined strongly in ammonoids, while brachiopods disappeared entirely. The total absence of the last ones is enigmatic. At least, it cannot be explained by the sampling errors, because palaeontological data were collected randomly from the entire Triassic succession (see above mentioned data sources). A repopulation began in the Carnian, with renewed radiation of the brachiopods, bivalves and foraminifers, while the diversity of ammonoids decreased. The Ladinian and Carnian palaeoenvironments were the same. Shales with clastic beds are interpreted by Gaetani *et al.* (2005) as turbidites, which were accumulated in the deep basin. Such conclusion is supported by my own field observations. Thus, abrupt deepening occurred in the Early Ladinian. It was resulted in the dominance of the palaeoenvironments, unfavourable for the pre-existed shallow-water Anisian fauna. In the Carnian, brachiopods, bivalves and foraminifers might have adopted to such environments, in contrast to ammonoids. Alternatively, the Carnian basin might have been shallower, which is documented by the appearance of carbonate rocks in the upper part of the Sakhrajskaja Group (Gaetani *et al.*, 2005). In this case, palaeoenvironments became more favourable to initiate the faunal diversification.

The Norian "explosion" of total marine biota resides in the diversification of brachiopods, though less ammonoids. The number of the Norian bivalves species was the same as in the Carnian. Corals, sponges and algae also appeared in the Norian, while a decline in the foraminiferal diversity is observed. This Late Triassic "explosion" may be directly linked to the dominance of the favourable palaeoenvironments of the carbonate platform and appearance of the diverse reefal communities (Gaetani *et al.*, 2005 ; Ruban, 2005).

In the Early Rhaetian, slight diversification of brachiopods

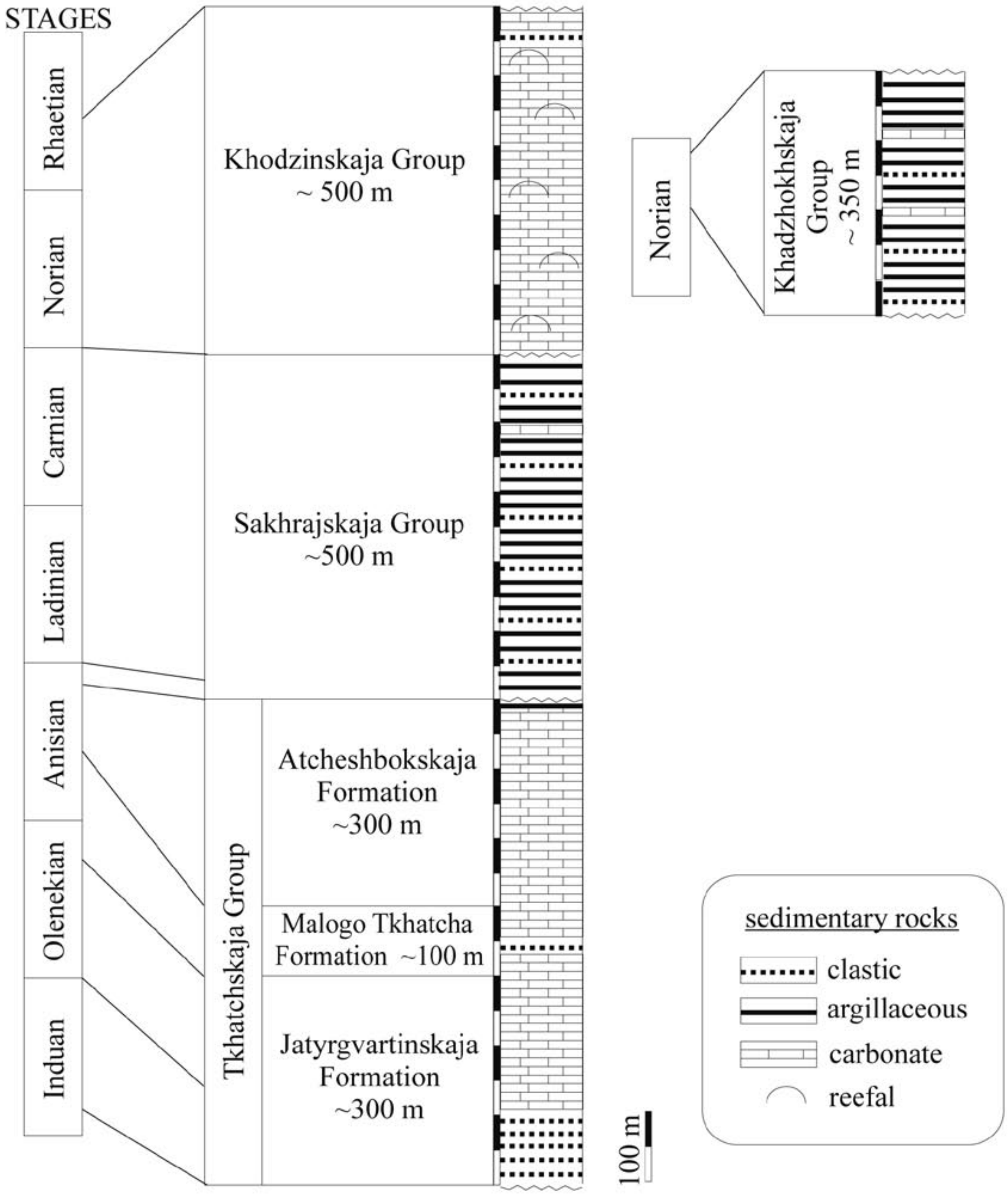

Fig. 2: The Triassic lithostratigraphy of the Western Caucasus (after Dagis & Robinson, 1973; Gaetani *et al.*, 2005; Prozorovskaja, 1979; Rostovtsev *et al.*, 1979 with corrections).

and ammonoids went on and foraminifers recovered, while bivalves disappeared entirely. The Late Rhaetian-Early Sinemurian phase of tectonic activity resulted a major hiatus, which embraced the mentioned time interval.

5. DISCUSSION

These regionally documented diversity patterns (Fig. 3) are somewhat similar to the global trends (Benton, 1995; Peters & Foote, 2001; Sepkoski, 1993). The globally-known recovery after the Permian/Triassic mass extinction is evidently observed in the regional record. The diversity acceleration is documented globally in the Middle Triassic. The present studies (Komatsu *et al.*, 2004) bring the evidences for the similar diversity rise in the Anisian. Bonuso & Bottjer (2005) suggested the improvement of the environmental conditions after the Permian/Triassic mass extinction in the Middle Triassic, which may be considered as the main cause, which provoked diversification at this time. However,

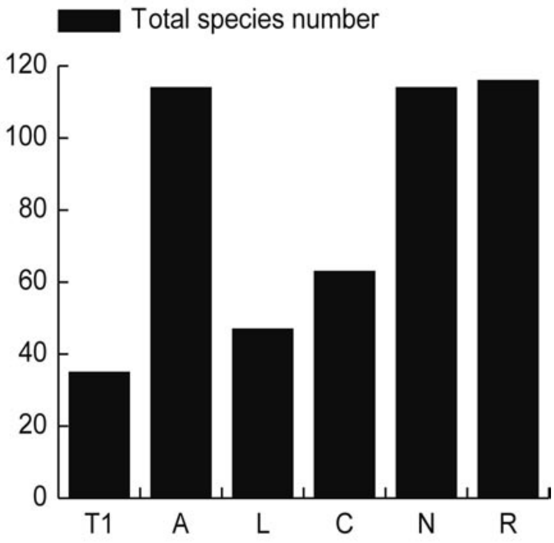

Fig. 3: The total diversity dynamics of the Triassic marine biota in the Western Caucasus. Time intervals: T1 - Lower Triassic (Induan+Olenekian), A - Anisian, L - Ladinian, C - Carnian, N - Norian, R - Rhaetian.

the Ladinian diminishing of the taxa quantity is not well documented by the data of the above mentionned global compilations. This may be explained by their low resolution

The Norian diversity «explosion» corresponded with the global patterns (Benton, 1995; Peters & Foote, 2001; Sepkoski, 1993). Although the decrease in the bivalve diversity is evident (Fig. 4), the globally known end-Triassic interval of the mass extinction (Hallam, 2002) cannot be observed in the Western Caucasus due to the major Late Rhaetian-Early Sinemurian hiatus.

6. CONCLUSIONS

The analysis of the diversity dynamics of the Triassic marine biota in the Western Caucasus permits to make some main conclusions:

1. Diversity increase in the Early Triassic strengthened in the Anisian; it was followed by the decline in the Ladinian. Diversity began to rise in the Carnian. The Norian is characterized by a new diversity "explosion",

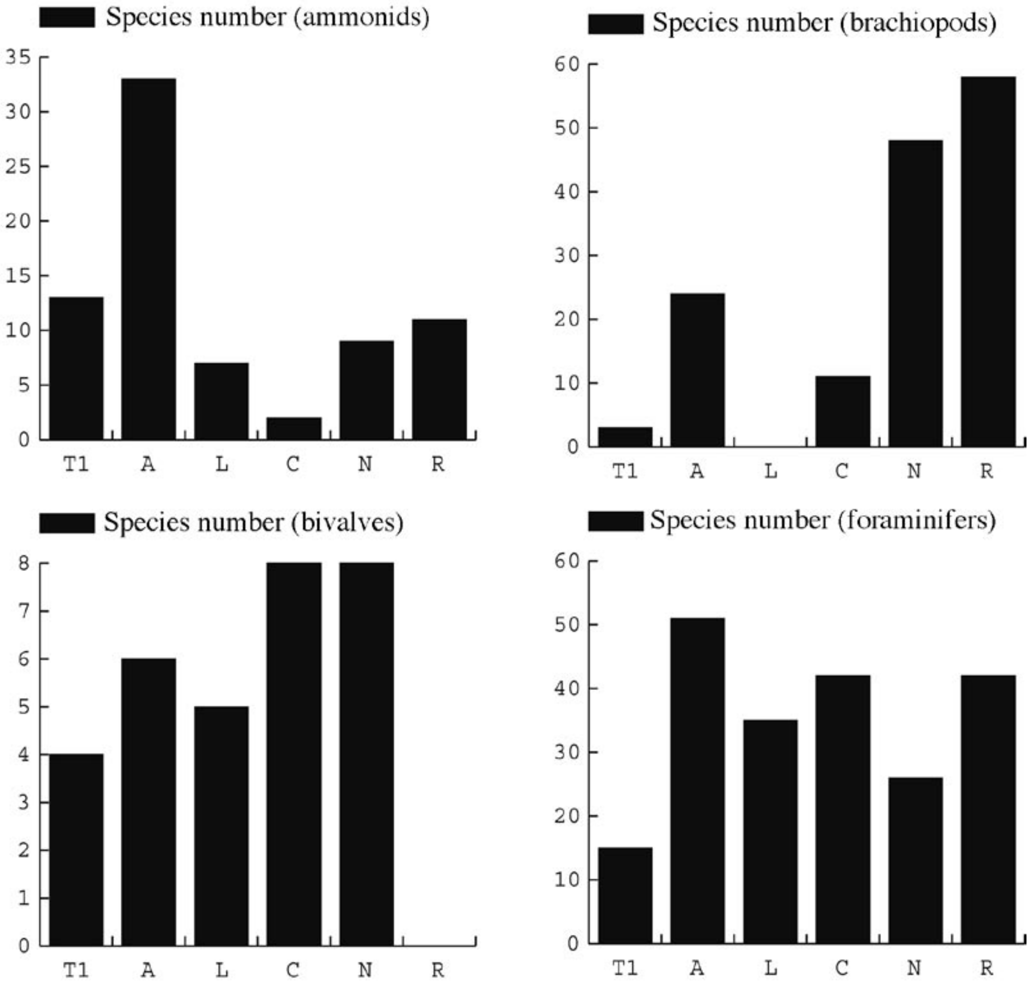

Fig. 4: Changes of the total species number of the Triassic ammonoids, brachiopods, bivalves and foraminifers of the Western Caucasus. For abbreviation of time intervals see Fig. 3.

and the Early Rhaetian is characterized by the very diverse marine biota.

2. Diversity dynamics within the principal fossil groups (ammonoids, brachiopods, bivalves, foraminifers) differ rather strongly.
3. Documented diversity changes may be explained by the changes of the regional palaeoenvironments.
4. The regional diversity dynamics reflected some global tendencies.

It seems that sharp diversity changes of the marine biota may be effectively used to correlate the Triassic strata of far-located regions.

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APPENDIX. Compiled list of genera of the Triassic marine biota of the Western Caucasus. Number of species in each genus in the Triassic stages is indicated.

Genera	Induan + Olenekian	Anisian	Ladinian	Carnian	Norian	Rhaetian
AMMONOIDS						
Acrochordiceras		3				
Aegeiceras		1				
Arcestes			1			3
Arpadites		1				
Badiotites			1			
Beyrichites		1				
Caucasites		2				
Cladiscites					2	2
Dieneroceras	1					
Flemingites	1					
Flexoptychites		1		1		
Gymnites		1				
Hollandites		3				
Japonites		1				
Joannites				1		
Juvavites					1	
Laboceras		3				
Leyophyllites		4				
Lobites					1	
Longobardites		1				
Megaphyllites		1			2	1
Mesocladiscites		1				
Monophyllites		1	3			
Nannites	1					
Owenites	3					
Paracladiscites						2
Paradanubites		2				
Paragoceras	1					
Parasageceras		1				
Parussuria	1					
Phyllocladiscites		2				
Pinacoceras					1	1
Placites					1	1
Proptychites	1					
Pseudosageceras	1					
Rhacophyllites					1	1
Smithoceras		1				

Genera	Induan + Olenekian	Anisian	Ladinian	Carnian	Norian	Rhaetian
Sturia		2	1			
Subowenites	1					
Subvishnuites	1					
Wyomingites	1					
Xenodiscus			1			
ALGAE						
Lithotamnidium					1	
Spongiomorpha					4	
BRACHIOPODS						
«Rhynchonella»		1				
Abrekia	1					
Adygella				1	1	1
Adygelloides						1
Amphiclina					1	2
Ampliclinodonta					1	
Angustothyris		1				
Aulacothyropsis				1		4
Austriella					1	
Austrirhynchia						2
Balatonospira				1		
Bobukella				1	1	
Caucasorhynchia					1	1
Caucasothyris					1	
Coenothyris		2				
Costirhynchia		1		1		
Costispiriferina		1				
Crurirhynchia					1	
Crurithyris	1					
Cubanothyris					2	2
Decurtella		2				
Dinarispira		2				
Dioristella		1		1		
Euxinella					3	6
Fissirhynchia					1	1
Guseriplia						2
Holcorhynchella		1				
Koeveskallina		1				
Koninckina				1	1	
Laballa					2	3
Lepismatina						1
Lobothyris					2	

Genera	Induan + Olenekian	Anisian	Ladinian	Carnian	Norian	Rhaetian
Majkopella						3
Mentzelia		1		2	1	2
Moisseievia					2	1
Neoretzia						3
Neowelerella	1					
Norella		1				
Oxycolpella					3	2
Pexidella		1			1	
Piarorhynchella		1				
Pseudocyrtina					1	
Pseudorugitella					2	2
Punctospirella		1				
Rhaetina				1	6	4
Rhimirhynchopsis					1	1
Robinsonella						1
Sinucosta		1			1	1
Spinolepismatina					1	
Sulcatinella		2				
Sulcatothyris				1		
Tetractinella		1				
Thecospira					1	
Thecospiropsis					1	
Triadithyris					2	1
Trigonirhynchella					1	2
Volirhynchia		2				
Wittenburgella					1	1
Worobievella					1	1
Zeilleria					3	6
Zugmayerella						1
BIVALVES						
Cassianella					1	
Claraia	4					
Daonella			2			
Halobia			1	5		
Hoernesia		1				
Indopecten					1	
Leda			1			
Limea		1				
Lyssochlamys				2		
Monotis					3	
Myophoria					1	

Genera	Induan + Olenekian	Anisian	Ladinian	Carnian	Norian	Rhaetian
Mytilus		1				
Paleocardita					1	
Posidonia		1	1			
Pseudomonotis					1	
Schafhaeutlia		1				
Velopecten		1		1		
CORALS						
Astraeomorpha					2	1
Montlivaultia					1	
Rhabdophyllia					1	
Stephanocoenia					1	
Stylophyllopsis					2	
Thamnastraea					2	2
Thecosmilia					6	2
FORAMINIFERS						
«Frondicularia»						2
«Orthovertella»	1					
«Protonodosaria»	1					
«Tetrataxis»					1	1
Agathaminina					1	1
Ammobaculites		2		1	1	
Ammodiscus	1		2	1	1	
Angulodiscus					1	1
Arenovidalina		4				
Astacolus		2	1	3		
Auloconus					1	1
Aulotortus					3	3
Calcitornella		1	1			
Cornuloculina		1		1	1	1
Cornuspira				1		
Coronipora						1
Dentalina	3	5	6	4		
Diplotremmina		1		1		
Duostomina		1	1			
Duotaxis						1
Earlandia		1				
Galeanella					1	2
Gandinella	1					
Gaudryina				2		1
Gaudryinella						1
Glomospira		2				

Genera	Induan + Olenekian	Anisian	Ladinian	Carnian	Norian	Rhaetian
Glomospirella		3				1
Hoyenella	1	1				2
Hyperammina		2		1		
Ichtyolaria				2		
Involutina					2	2
Labalina				1		
Lagena		1				
Lenticulina			4	3		2
Lingulina		2	4	1		
Marginulinopsis				1		
Meandrospira	2	3				
Miliolipora					1	1
Nodosaria	5	2		6		2
Nodosinella		1				
Ophthalmidium			1	2	3	4
Pachyphloides			1	2		
Pilammina		3				
Planiinvoluta					1	4
Pseudonodosaria		7	9	5		
Quinqueloculina			1			
Reophax			2			
Saccammina		1				
Semiinvoluta					1	1
Spiroplectammina		1		1		
Tetrataxis					1	
Tolypammina		3				
Triasina					1	
Trochammina		1		2	2	3
Trocholina				1	3	4
Turrispirillina						1
Vaginulina			1			
Vaginulinopsis			1			
SPONGES						
Hodsia					1	
Molengraffia					1	
Sahraja					1	

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Taxonomic diversity dynamics of the Jurassic bivalves in the Caucasus: Regional trends and recognition of global patterns

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Abstract

Jurassic bivalves were quite diverse in the Caucasus—a region located on the northern margin of the Tethys Ocean. Stratigraphic ranges of 513 species were compiled. Total species number, already low, declined during the Early Jurassic, reaching its minimum in the Toarcian (27 species). Although this coincided with the Pliensbachian–Toarcian global mass extinction, the latter is not evident for the Caucasian bivalves. Diversity rises occurred in the Aalenian and then in the Bajocian, when 77 species existed. In the Bathonian species number did not change. A diversity explosion took place in the Callovian, when favourable palaeoenvironmental conditions were established. 163 species are known from deposits of this age. This event is recognized globally as generic diversity significantly rose at this time all over the world. When a rimmed shelf was developed in the Late Callovian–Oxfordian, total species number reached its maximum (166 species). But a regional salinity crisis in the Kimmeridgian–Tithonian led to a final diversity drop, when 86 species existed in the Tithonian. Bivalves could not recover at the end-Jurassic, along with other representatives of shelly benthos. Comparison with the other well-known salinity crisis, which occurred in the Messinian in the Mediterranean, suggests that the Late Jurassic event in the Caucasus was much stronger and stressed bivalve assemblages for a longer time. The influence of the end-Jurassic mass extinction has not been fully documented. Throughout the Jurassic benthic assemblages within the Caucasus changed rapidly. This regional diversity dynamics corresponds quite well to the global trends. Both sea-level and marine palaeotemperature changes may have had an indirect influence on regional bivalve diversity, as they fluctuated strongly during the Jurassic.

Keywords: Bivalves; Diversity; Mass extinction; Jurassic; Caucasus

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1. Introduction

Bivalves were an important group among the Jurassic marine biota. They have been characterized by a great number of species and genera. Changes in their diversity were analyzed by Hallam (1972, 1986, 1987), Miller and Sepkoski (1988), Johnson (1990), Aberhan (1994), Little and Benton (1995), Aberhan and Fürsich (1997), McRoberts and Aberhan (1997), and Benton (2001). Most of these studies examined particular patterns of diversity (e.g., extinction rate) or quite short intervals of mass extinctions (i.e., mostly the Pliensbachian–Toarcian event). Therefore, the study of taxonomic diversity dynamics among Jurassic bivalves remains relevant.

Useful evidence of diversity changes among Jurassic bivalves may be obtained from the Caucasus — a large region embracing the territory of Southwestern Russia and the entirety of Georgia, Armenia and Azerbaijan (Fig. 1). During the Jurassic it was located on the northern Neotethyan margin in the "key" position

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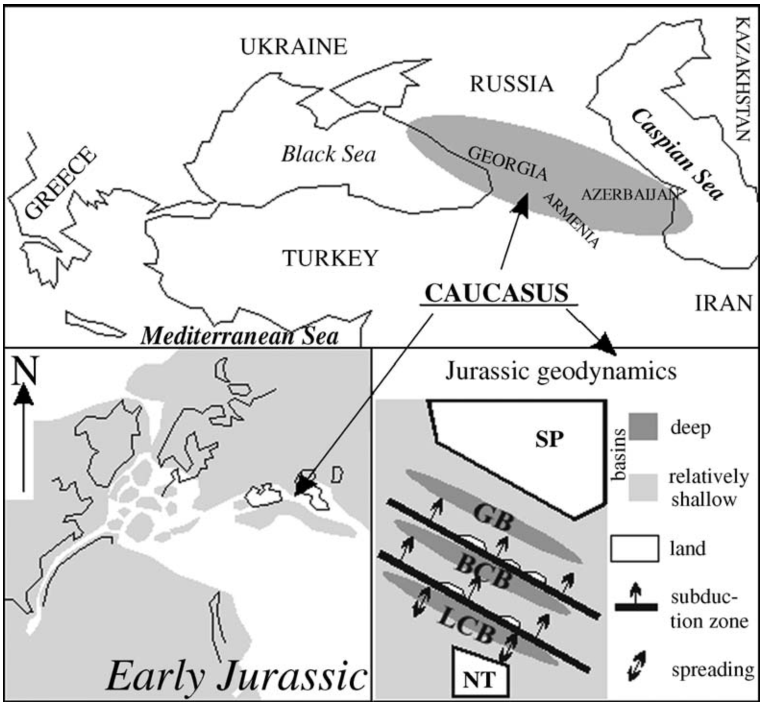

Fig. 1. Geographical and palaeogeographical locations of the Caucasus (Early Jurassic base map simplified after Owen, 1983; geodynamic situation is modified after Lorkipanidze et al., 1984). Abbreviations: SP—Scythian Platform, GB—Greater Caucasus basin, BCB—hypothetic Black Sea–Caspian Sea Basin, LCB—Lesser Caucasus Basin, NT—Nakhitchevan' Terrane.

between European and Asiatic areas. In this region Jurassic deposits are widespread and bivalves are common (Rostovtsev et al., 1992). However, the regional diversity of bivalves has yet to be analyzed, except for a preliminary study devoted to the Late Jurassic (Ruban, 2003a), and an intriguing remark by Hallam (1986) on the Tithonian mass extinction, which stressed bivalves in the Caucasus.

A first compilation of all available data on the stratigraphic distribution of Jurassic bivalve species of the Caucasus is presented herein. Several hundred taxa were evaluated. This study provides an excellent basis upon which to evaluate diversity dynamics of this group through the Jurassic.

2. Geological setting

In the Jurassic the territory of the Caucasus was occupied by marine basins (Fig. 1). They were well connected to one another as well as to the Neotethys ocean (Owen, 1983; Meister and Stampfli, 2000; Stampfli and Borel, 2002; Brunet and Cloetingh, 2003; Golonka, 2004). These basins had distinct tectonic settings; several spreading and subduction zones were concentrated in a relatively narrow belt (Lordkipanidze et al., 1984; Ershov et al., 2003), where several microplates interacted.

Various sediments accumulated in the Caucasian basins during the Jurassic (Rostovtsev et al., 1992). Most of the typical facies, from continental to deep marine, are represented within the study area. In the Greater Caucasus Basin shales and clastic deposits up to 10 000 m thick accumulated during the Early–Middle Jurassic, while carbonates and evaporites (up to 3000 m) were deposited in the Late Jurassic (Tsejsler, 1977; Rostovtsev et al., 1992). The sedimentary succession of the Lesser Caucasus Basin is somewhat similar, although the thickness of the deposits is less; but the main distinctive feature is a great abundance of volcanogenic deposits (Rostovtsev et al., 1985, 1992). During the Jurassic several transgressions occurred within the Caucasus. Their maxima were reached in the Pliensbachian–Toarcian and later in the Oxfordian–Kimmeridgian.

In the Early–Middle Jurassic the Caucasus lay in the subtropical or moderate humid zones. Toarcian palaeotemperatures are estimated as 15–20 °C; in the Early Aalenian they dropped to 5–15 °C, but already in the Late Aalenian temperatures increased to 20–25 °C and stayed at the same level until the end of Jurassic

(Jasamanov, 1978). After the beginning of the Callovian the climate became subtropical to tropical and semi-humid. In the Late Kimmeridgian–Tithonian significant aridization occurred, at which time evaporites accumulated (Jasamanov, 1978; Rostovtsev et al., 1992).

The general palaeogeographical position of the Caucasus is the subject of some discussion. Dommergues (1987) places it in the Euro-Boreal domain for the Early Jurassic. Westermann (2000) mentions it being included in the Tethyan Realm. Perhaps the best interpretation is that region was transitional. This is supported by the taxonomic composition of brachiopod assemblages (Ruban, 2003b). Another significant feature of the palaeogeographical changes in the Caucasus is the drift of this region northward. In the Early Jurassic it was located around 25°N latitude, while at the end of the Jurassic it had moved to 30°N latitude (Lordkipanidze et al., 1984).

3. Material and methods

Bivalves were one of the most diverse and abundant groups among the Jurassic marine biota of the Caucasus. They are found in strata of all stages of this system, except the Hettangian, which corresponds to a major regional hiatus (Rostovtsev et al., 1992). Although they have been described in several quite comprehensive papers (Ptchelintsev, 1931; Khimshiashvili, 1957, 1967, 1974; Azarjan, 1983, Romanov, 1985; Rostovtsev et al., 1985, 1992; Ratiani, 1987; Janin, 1989; Romanov and Ksum-Zade, 1991; Ruban, 2005a), a total compilation of Jurassic bivalve data has not yet been made. Therefore, the first stage of the present study was to compile all available information on the stratigraphic distribution of bivalve taxa for the Caucasus. Taxonomic revisions also were conducted to avoid duplication or missing of information on a single taxon. Finally a database of valid species was constructed, where presence and absence of taxa in each stage were tabulated. Unfortunately, the initial resolution of data and various stratigraphic problems do not permit the data to be assessed below stage level. In total, the ranges of 513 valid species were obtained.

A difficulty was encountered with the common reports of underinterpreted taxa, i.e. species defined with “cf.”, “aff.” and “sp.” To avoid overestimation of species numbers it was decided to exclude species identified to “sp.” only, because it is difficult to understand their relation to completely identified species mentioned in other papers, especially where no figures and/or detailed descriptions were available. As for species defined with “cf.” and “aff.”, they were counted as concrete species, so as to avoid underestimating the diversity. The same assumptions were made in a study of Alba et al. (2001).

A quantitative approach was used to analyze the compiled data, which includes the calculation of *total species number (TSN)*, *appearing species number (ASN)*, *disappearing species number (DSN)*, *number of singletons (SN)*, *R-index* for stages (Sinemurian–Tithonian). Brief explanations of these indices are presented here. *TSN* is the quantity of bivalve species in each stage. All bivalves distributed regionally in the deposits of a particular stage are defined as a single *assemblage*, i.e., each assemblage directly corresponds to the whole stage. *ASN* and *DSN* are indices which show the quantity of species that have appeared or disappeared regionally within a particular stage. It is necessary to distinguish them from number of *originated* or *extinct species* because of the so-called Lazarus-effect (Wignall and Benton, 1999; Fara, 2001). *Appearing species* might have existed earlier and then temporarily have disappeared in the regional fossil record, so that some are not new taxa. *Disappearing species* may appear again in the regional record some time later, i.e., not all of them are extinct. *Singletons* are species that appeared and then disappeared within a single stage (sensu Foote, 2000).

R-index establishes Jaccard’s (1901) similarity between successive assemblages (Ruban and Tyszka, 2005):

$$R = C/[(N1 + N2 - C)],$$

where C is the number of common taxa for two assemblages, and $N1$ and $N2$ are the taxa numbers in each assemblage. *Assemblage transformation rate (ATR)* essentially is $1/R$. It shows how rapidly the changes in the taxonomic composition of assemblages are realized through geological time.

It is important to note that lack of data on the Berriasian does not permit a calculation of DSN and SN values for the Tithonian.

4. Regional diversity analysis

Diversity dynamics of Jurassic bivalves in the Caucasus is characterized by significant changes of principal patterns. TSN stayed at a relatively low level during the Sinemurian–Aalenian, not exceeding 50 species (Fig. 2). A weak declining trend is observed in the Early Jurassic with a minimum in the Toarcian, when only 27 species existed. A first significant diversification occurred in the Bajocian, when TSN reached 77. In

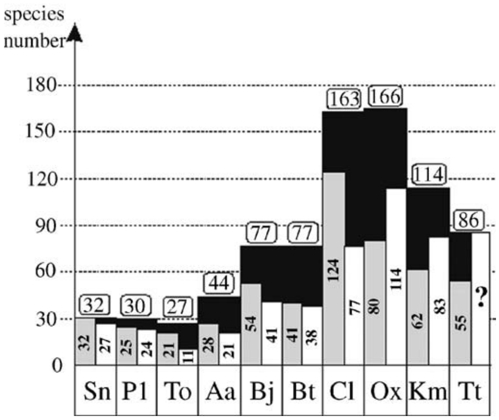

Fig. 2. Principal patterns of taxonomic diversity dynamics of the Jurassic Caucasian bivalves. Columns: black—TSN, gray—ASN, white—DSN. TSN values are indicated above general columns in rectangles, ASN and DSN values are presented in the columns. Stage abbreviations: Sn—Sinemurian, Pl—Pliensbachian, To—Toarcian, Aa—Aalenian, Bj—Bajocian, Bt—Bathonian, Cl—Callovian, Ox—Oxfordian, Km—Kimmeridgian, Tt—Tithonian.

the Bathonian the TSN value did not change. An explosion of diversity occurred in the Callovian, when TSN value doubled to 163 taxa. Only a minor increase is recorded in the Oxfordian (166 species), but later diversity declined significantly until the end of Jurassic. In the Tithonian only 86 species existed. But the diversity of bivalves did not decrease in the Late Jurassic so much as to reach the Toarcian minimum.

ASN and DSN changed in the same way as the TSN (Fig. 2). However, the number of appearances dropped in the Bathonian and Oxfordian. The number of disappearances, on the contrary, accelerated in the Oxfordian and then stayed at a relatively high level in the Kimmeridgian. A comparison between values of ASN for assemblage of a given stage and DSN of the assemblage of the preceding stage (DSNp) may be used to explain the mechanism of TSN changes. During the Early Jurassic the ASN was lower than DSNp, which led to the TSN decrease. But TSN decline should be explained mostly by a general ASN decrease. The Aalenian and Bajocian TSN accelerations occurred thanks to an increase in ASN relative to DSNp. The Callovian diversity explosion took place, when ASN increased to about 3 times greater than DSNp. The Late Jurassic diversity decline was realized because ASN was less than DSNp. Nonetheless appearances rates were quite stable and total diversity decreased realized thanks to the strength of disappearance rate.

SN values reached maxima in the Sinemurian, Bajocian, and especially in the Callovian–Kimmeridgian (Fig. 3). Singletons were the most diverse in the Sinemurian and Pliensbachian, when such taxa composed more than 70% of assemblages. During other intervals of the Jurassic the quantity of singletons in assemblages varied between 25% and 50%.

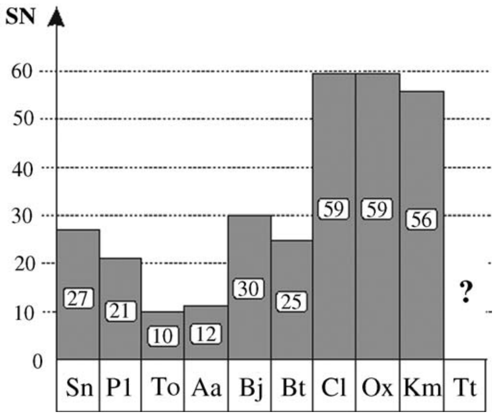

Fig. 3. Singleton quantity among the Jurassic bivalves of the Caucasus. Stage abbreviations—see Fig. 2.

R-index values fluctuated during the Jurassic (Fig. 4). Maxima in similarity among assemblages were reached for the Toarcian–Aalenian, Bajocian–Bathonian, and Callovian–Oxfordian. The turnover of assemblages was quite rapid overall as *R*-index did not reach even 0.40 at maximum, and was usually below 0.30. In the Early Jurassic the transformation occurred because of turnovers of the assemblages taxonomic composition as TSN is comparable for stages. But in the Middle–Late Jurassic assemblage transformations were realized mostly because of rises or falls of the TSN. At this interval when TSN values for successive assemblages are comparable, the ATR decreased and transformation occurred mostly by turnovers (Bajocian–Bathonian,

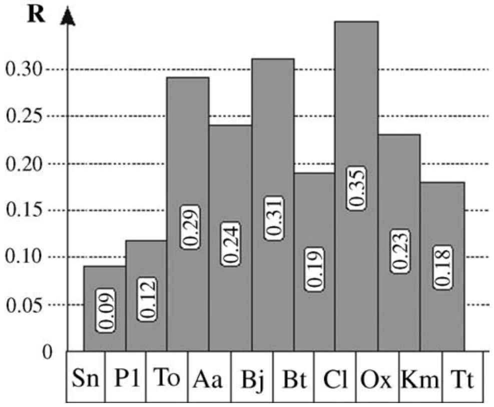

Fig. 4. *R*-index for assemblages of the Jurassic bivalves of the Caucasus. Stage abbreviations—see Fig. 2.

Callovian–Oxfordian). The most intense assemblage changes took place at the Sinemurian–Pliensbachian, Pliensbachian–Toarcian, Bathonian–Callovian, and Kimmeridgian–Tithonian boundaries.

The most diverse genera were *Cardinia* and *Plagiostoma* (Sinemurian), *Pseudopecten* (Pliensbachian), *Retroceramus* (Toarcian and Aalenian), *Mytiloceramus* (Aalenian), *Pleuromya* (Bajocian), *Camptonectes* (Bathonian, Oxfordian and Kimmeridgian), *Chlamys* and *Plagiostoma* (Bathonian–Tithonian), *Pholadomya* and *Pleuromya* (Callovian), *Astarte* (Callovian and Oxfordian), *Spondylopecten* (Oxfordian), and *Diceras* (Tithonian). These genera are characterized by the highest species numbers in the indicated assemblages. It is necessary to point out they were not always the most abundant taxa. For example, *Radulopecten* and *Entolium* dominated in the Early–Middle Callovian (Rostovtsev et al., 1992; Ruban, 2003a), but they were not the most diverse genera within those assemblages.

5. Recognition of global patterns

To explain the observed diversity dynamics of the Caucasian Jurassic bivalves is a significant task. The first step in this direction is to compare the regional versus global records.

An evaluation of the global total diversity changes was attempted by Miller and Sepkoski (1988) and then reconsidered by Benton (2001). Although these studies were aimed at generic diversity (Fig. 5), it seems to be comparable with the species diversity for the general comparison. The total number of bivalve genera was relatively minimal in the beginning of the Jurassic. A weak radiation occurred in the Early Jurassic, which was followed by a short interval when diversity stayed at the same level. In the second part of the Middle Jurassic the number of genera quickly grew. This may be regarded as a "diversity explosion". Late Jurassic assemblages were the most diverse, but there were no significant variations in the number of taxa. The Jurassic/Cretaceous transition is marked by an abrupt decline of bivalves on the global scale.

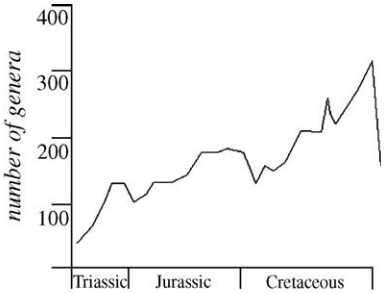

Fig. 5. The global Mesozoic bivalve generic diversity (modified from Miller and Sepkoski, 1988).

A comparison between regional (Fig. 2) and global (Fig. 5) diversity dynamics of Jurassic bivalves shows they were quite similar. Two differences are different diversity changes in the Early Jurassic and an early decline of the Caucasian bivalves in the Late Jurassic. The first difference may be ignored, because both regional and global trends are too weak to be directly comparable. But more detailed data on the Early Jurassic TSN changes of European (Hallam, 1976, 1986) and Andean (Aberhan and Fürsich, 1997) bivalves also suggest a gradual radiation of the Liassic bivalves. The quantity of Sinemurian bivalves known in North America is also a little lower than that of Pliensbachian bivalves (Smith et al., 1994). Thus a mechanism-driven evolution of Caucasian bivalves was distinct in the Early Jurassic. One of the possible explanations lies in the timing of dysoxic/anoxic events. In the Caucasus dysoxic palaeoenvironments were established beginning in the Pliensbachian, when black shales began to be deposited. This may have negatively influenced the bivalve assemblages because of deteriorating environments. As for the Late Jurassic, palaeoenvironments in the Caucasus were extraordinary at this time, which may have affected the bivalve evolution (see Discussion).

It is important to compare the diversity changes of the Jurassic bivalves between the Caucasus (Fig. 2) and Western Europe (Hallam, 1976). Rapid radiation in the Early Jurassic and decline in the Toarcian, which have been documented in the European record, do not appear in the Caucasus (see also above). However, the diversity acceleration in the Aalenian–Bathonian took place in both regions. A significant difference is the Callovian diversity drop in Western Europe, whereas in the Caucasus the bivalve diversity greatly increased during this time. This may be explained by the onset of the carbonate platform development in the Caucasus in the Callovian (Ruban, 2005b), which resulted in a diversity explosion (see below for details). In Western Europe bivalve diversity rose in the Oxfordian, and in the Caucasus species numbers were high during this age. In the Late Jurassic the diversity trends were similar in both regions, where a decline occurred.

Thus quite strong correspondence is observed between the whole regional and global diversity dynamics of the Jurassic bivalves. This is a significant conclusion, because Caucasian basins were relatively isolated. Such correspondence may be explained by the high degree of similarity between bivalves of the Caucasus and other regions, i.e. low degree of endemism — at least at the generic level. Many of the Jurassic bivalves found in the Caucasus are known in the other regions of the world, which is suggested by data on their global distribution (Hallam, 1977). For example, more than twenty Caucasian bivalve genera (*Acesta*, *Camptonectes*, *Entolium*, *Meleagrinella*, *Neocrassina*, *Opis*, *Oxytoma*, *Palaeonucula*, *Pholadomya*, *Pleuromya*, *Trigonia*, etc.) are known even in the far-flung Australasian regions, characterized recently by Grant-Mackie et al. (2000). Hallam (1977) has shown a high proportion of cosmopolitan genera among the Jurassic bivalves. Sha (2003) argued that many Mesozoic bivalves were global colonizers, being characterized by cosmopolitan, anti-tropical or pantropical geographical distributions. Some taxa considered by Sha (2003) (e.g., *Camptonectes*, *Chlamys*, *Gryphaea*) are typical for the Caucasus. Generally, the palaeobiogeographic differentiation in the Jurassic was not so great as later in the Cretaceous and Cenozoic. True superrealms appeared only in the end-Jurassic, and true realms in the Middle Jurassic (Westermann, 2000). Such conclusion is evidently supported by Hallam's (1977) data, which considered only bivalves.

Two globally recognized and remarkable events in the evolution of Jurassic bivalves should be discussed with respect to the data on the Caucasus. They are the Pliensbachian–Toarcian mass extinction and the Callovian diversity explosion.

The Pliensbachian–Toarcian interval is known for a global mass extinction, which particularly stressed bivalves (Hallam, 1972, 1986, 1987; Aberhan, 1994; Little and Benton, 1995; Aberhan and Fürsich, 1997; McRoberts and Aberhan, 1997; Hallam and Wignall, 1997; Harries and Little, 1999; Pálfy et al., 2002). This event was previously recognized in the Northern Caucasus for brachiopods; establishment of anoxic environments seems to be its likely cause (Ruban, 2004a). Our data suggest the TSN of bivalves in the Toarcian was only 3 species lower than in the Pliensbachian (Fig. 2). Although 24 Pliensbachian species disappeared, 21 new species appeared in the Toarcian, which may be explained by both speciation and taxa immigration. Moreover, DSN did not rise in the Pliensbachian. Therefore, although we document a "negative event", it is not so significant as in the other regions such Northwest Europe or South America (Hallam, 1986, 1987; Little and Benton, 1995; Aberhan and Fürsich, 1997). One possible explanation lies in the area of data resolution. In the Caucasus recovery after the Pliensbachian–Toarcian mass extinction began already in the Middle Toarcian (Ruban, 2004a). Hence, at the stage level the short-term effect of the shelly benthos collapse during the Toarcian seems to be minimized by the high values of overall diversity in the Middle–Late Toarcian. Thus to recognize a peak mass extinction regionally (if it existed at all) for bivalves should be a task for further detailed studies. But a distinctive feature of the regional record is the relatively low diversity of the Caucasian bivalves in the Pliensbachian, when other marine fossils, e.g., brachiopods, were strongly diversified. In conclusion, the global generic diversity curve does not reflect the influence of the Pliensbachian–Toarcian mass extinction on the bivalves (Miller and Sepkoski, 1988).

The Callovian diversity explosion of bivalves in the Caucasus (Fig. 2) took place along with a radiation in the entire shelly benthos (Ruban, 2003a, 2005a,b). This occurred as favourable environmental conditions were established from the Early Callovian: climate became warmer, and wide shelves were occupied by sands or calcareous silts (Jasamanov, 1978). Moreover, in the studied sections abundant remains of *Chlamys* (*Chlamys*) *viminea* (SOWERBY, 1826), which was one the Callovian dominants, were found as low as in the Lower Callovian basal conglomerates (Ruban, 2005a), i.e. the bivalve radiation was abrupt. In the Late Callovian a great carbonate platform emerged (Fig. 6). It existed until the end of Jurassic (Rostovtsev et al., 1992; Kuznetsov, 1993). This platform was essentially a rimmed shelf (after Ginsburg and James, 1974; Read, 1985), bounded by barrier reefs constructed by corals, which diversified at this time across the entire Tethys (Martin-Garin et al., 2002). In the Oxfordian conditions evidently remained favourable for bivalves. But it is important to note that bivalves were restricted more to shelf, than to reefal facies (Rostovtsev et al., 1992). Generally, the bivalve diversity explosion established in the Caucasus was not a specific regional feature. The similar generic diversity acceleration occurred on the global scale (Miller and Sepkoski, 1988; Benton, 2001). Moreover, the new diversity curves of the whole planetary marine biota (Benton, 2001; Peters and Foote, 2001) indicate an increase of taxa number in the middle of the Jurassic. This suggests that a diversity explosion documented in the Caucasus may be at least partly explained by the influences of the evolutionary mechanism of bivalves overall or even by global factors, enforced in all marine biotas.

A decrease in the total diversity of bivalves in the end-Jurassic occurred globally (Hallam, 1976, 1986; Benton, 2001), which supports the idea of mass extinction. But in the Caucasus its greatest effect is

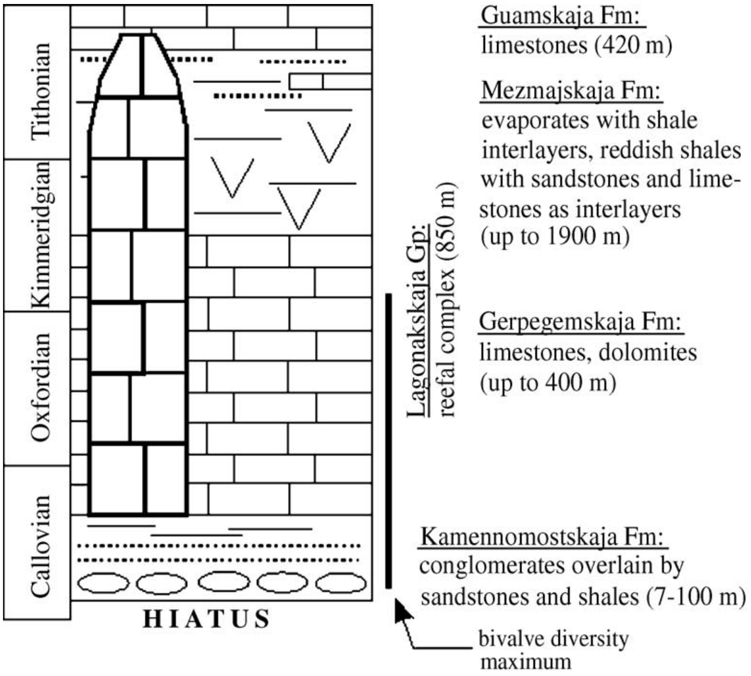

Fig. 6. Stratigraphic extent of the Late Jurassic rimmed shelf in the Western Caucasus (formations and thicknesses after Rostovtsev et al., 1992).

recorded in the Berriasian (Ruban, 2004b), when foraminifers and brachiopods strongly declined. Moreover, in the global record the diversity of bivalves also dropped in the beginning of the Cretaceous (Miller and Sepkoski, 1988). Therefore, it is difficult to discuss how strongly this global event stressed end-Jurassic bivalves regionally. But it is not possible to exclude that the TSN decrease observed in the Tithonian might be explained partially by mass extinction.

6. Sea-level changes as a driving force of diversity dynamics

Sea-level fluctuations are often considered as one of leading factors of diversity changes in bivalves and other marine biota (Newell, 1967; Hallam and Wignall, 1997, 1999; McRoberts and Aberhan, 1997; Smith, 2001). Available data on the Jurassic bivalves of the Caucasus allow a test of this assumption. The first step was an estimation of the regional sea-level changes.

The territory of the Caucasus is subdivided into several dozen areas (traditionally called “zones”, although this is not well appropriated term), distinguished by facies composition of the Jurassic successions (Rostovtsev et al., 1992). A total of 36 “zones” delineate the Hettangian–Bathonian interval, and 26 for the Callovian–Tithonian interval. In the Caucasus the Callovian stage traditionally is attributed to the Upper Jurassic, which is wrong (see scale of Gradstein et al., 2004). A facies interpretation for each of the hundreds of formations was made herein, based on the information of Rostovtsev et al. (1992) and personal field observations. Three main types of palaeoenvironments were defined: continental, shallow-marine and deep-marine. Then for each stage a total number of “zones” with a particular type of these palaeoenvironments was derived. Stratigraphic patterns between distinct facies types mark transgression/regression and shallowing/deepening episodes (Fig. 7). A somewhat similar approach was used by Hallam and Wignall (1999), who calculated the number of shallow formations to evaluate sea-level highstands and lowstands. Smith (2001) measured the outcrop area of the principal types of sedimentary rocks, and Crampton et al. (2003) calculated the same. Peters and Foote (2001) accounted for the number of marine formations and global rock area.

To reconstruct the sea-level curve a specific index (ISL) was evaluated:

$$\mathrm{ISL} = (s + d)/c,$$

where s, d, c are the number of “zones” with shallow-marine, deep-marine, and continental palaeoenvironments respectively. The lower values of this index mark regressions, while higher values indicate transgressions. Shallowing is recognized by an increase in s-number relative to d-number.

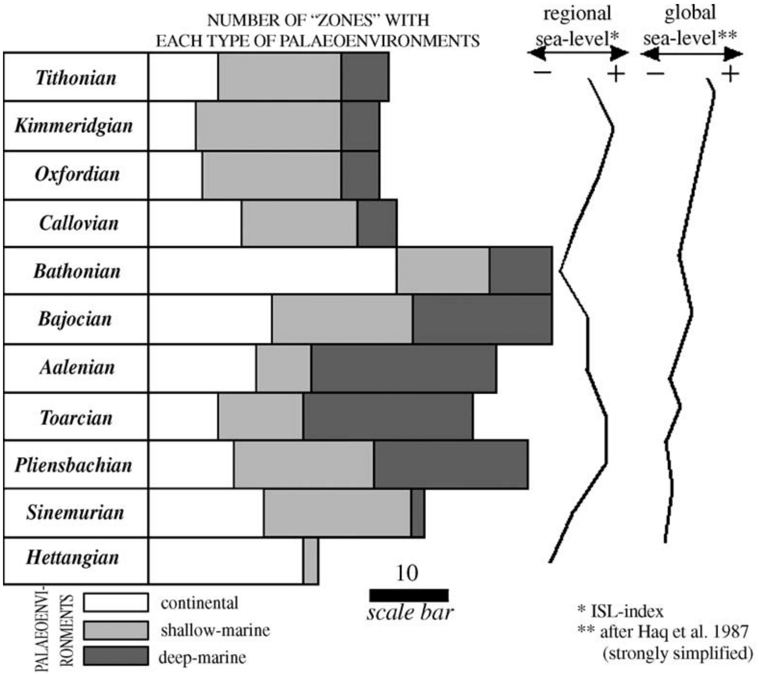

Fig. 7. Changes in the total number of "zones" with interpreted continental, shallow-marine and deep-marine types of palaeoenvironments, regional and global sea-level changes (see text for explanation).

The Hettangian and Bathonian were the time intervals of strong regressions. By contrast, seas transgressed during the Sinemurian–Pliensbachian and later in the Callovian–Kimmerdgian. Deepening reached its maximum in the Toarcian–Aalenian. Therefore, the Early Jurassic transgression is indicated by an increase in depth of the Caucasian basins, while in the Late Jurassic seas were shallow and extensive shelf areas developed.

Comparison between global sea-level changes (Haq et al., 1987; see also Hallam, 2001) and those observed in the Caucasus suggests a regional difference (Fig. 7). This was caused by the regional tectonics. Peaks of transgressions and regressions were not synchronous. The general trends of sea level are somewhat similar, but only in the most generalized view. In the global record a tendency toward transgression is documented for the overall Jurassic, and in the Caucasus, sea-level rise is evident, but it is weak. Globally four transgressive–regressive cycles appeared in the Jurassic, while only two larger ones are recorded within the Caucasus. Sea-level changes recorded in Europe (Jacquin and de Graciansky, 1998; Smith, 2001; Stratigraphische Tabelle von Deutschland, 2002) are a little more similar to those in the Caucasus, but a large difference still exists.

An interesting conclusion may be made. As shown above, the Jurassic bivalves diversity changed globally and in the Caucasus in a similar manner. Therefore, the differences in sea-level changes imply that they may not be considered as a leading factor of these diversity changes. The same conclusion was reached for Early Jurassic brachiopods in the Caucasus and worldwide (Ruban, 2004a).

The quantitative expression of the regional sea-level changes allows a calculation of the simple index of rank correlation (RS) (Kendall, 1970) between ISL, *s*-, *d*-, *c*-number and the total species number dynamics for the entire Jurassic period. A similar analysis was attempted by McRoberts and Aberhan (1997) for Early Jurassic bivalves of Northwest Europe, and also by Crampton et al. (2003) for Cenozoic molluscs.

Results obtained for the Caucasus are surprising. RS for TSN and ISL is only 0.24. RS values for TSN and *s*-, *d*-, and *c*-number are 0.50, −0.31, and −0.31 respectively. In no case are the correlation values significant. A stronger correlation was observed between number of "zones" with shallow-marine palaeoenvironments and sea-level fluctuations. But even in this case the RS values are low. Therefore, it may be concluded that sea-level changes should not be considered as a driving force of diversity dynamics for Jurassic bivalves of the Caucasus.

The absence of a relationship between diversity of the Early Jurassic bivalve species from the Northwest Europe and sea-level changes was reported by

McRoberts and Aberhan (1997). In another well-known example Hallam and Wignall (1999) showed that transgressions/regressions distinctly enforced mass extinctions among the Phanerozoic marine biota. Undoubtedly, sea-level fluctuations should not be excluded from the list of diversity driving factors, but whether they were direct or indirect factors must be considered in every case. Sea shallowing partially controlled the diversity rises among bivalves of the Caucasus, as inferred from the Sinemurian, Bajocian–Bathonian and Late Jurassic, whereas in the other cases (as in the Aalenian or Kimmeridgian) no direct connections are observed. Regardless, further studies based on global and other regional data are awaited to test this conclusion.

7. Discussion

Regional diversity dynamics of Jurassic bivalves from the Caucasus are discussed with respect to the influence of sea-water palaeotemperature changes and the impact of the Late Jurassic salinity crisis.

Sea-water palaeotemperature changes might have controlled the diversity dynamics of Jurassic bivalves within the Caucasus because amplitude changes were significant (Fig. 8). Regional palaeotemperatures of sea-water for each Jurassic stage were established by Jasamanov (1978), who analyzed oxygen isotopic compositions derived from belemnites and other benthic macrofossils. Although such isotopic measurements might lead to wrong interpretations (Longinelli, 1996; Longinelli et al., 2002, 2003; Longinelli, personal communication), results obtained for the Caucasus were compared to palaeoecological studies (Jasamanov, 1978).

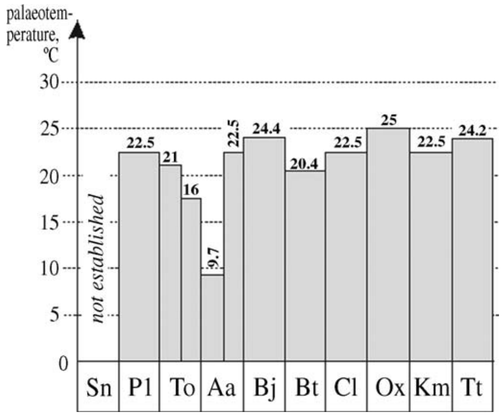

Fig. 8. Jurassic sea-water palaeotemperature changes in the Caucasus (data from Jasamanov, 1978, average values are shown). Stage abbreviations — see Fig. 2.

A comparison between regional bivalve diversity (Fig. 2) and sea-water palaeotemperatures (Fig. 8) suggests there was no direct covariation between them. Even the remarkable Late Toarcian–Early Aalenian cooling did not stress bivalves. Meantime, all Jasamanov's data and, therefore, our conclusions should be further verified to become trustable. It is important to note, that Late Aalenian–Bajocian and Callovian–Oxfordian episodes of sea-water warming corresponded to the evident radiation of bivalves, i.e. diversity rises. Therefore, there was a partial temperature control of the bivalve diversity changes. On the other hand, changes of sea-water palaeotemperatures might have provoked major changes in the structure of the bivalve assemblages, which were not recorded as diversity events. This is an important matter for further studies.

The Kimmeridgian–Tithonian is characterized by the remarkable decline of bivalves (Fig. 2). The causes of this event are readily apparent in the stratigraphic record. During this time interval in many parts of the Caucasus the deposition of evaporites took place indicating an increase in salinity of sea-water (Jasamanov, 1978; Rostovtsev et al., 1992; Kuznetsov, 1993) (Fig. 6). Also it appears that at least parts of some basins were isolated as lagoons. But complete restriction of the Caucasian basins did not occur as suggested from our facies analysis (Fig. 7). Nonetheless, such "hypersaline" environments were unfavourable for bivalves, which showed a significant diversity drop. In comparison to brachiopods, bivalves had not recovered by the Tithonian (Ruban, 2003a). This can be explained by the ability of brachiopods to radiate more rapidly when favourable conditions appear even for a very short time intervals (Prosorovskaya, 1996). Evidently, the diversification of bivalves needs much more time to accelerate.

In discussing the influence of the regional salinity crisis on bivalves in the Jurassic of the Caucasus, a comparison was made with respect to the effect of the other well-known salinity crisis, which occurred in the Messinian of the Mediterranean Sea (Krijgsman et al., 1999). Comprehensive data are available, including stratigraphic ranges of molluscan species known from the Neogene basins of southeast Spain (Demarcq, 1990). This information allows an evaluation of total numbers of mollusc species (Fig. 9). It is evident that the salinity crisis of the Late Messinian did not lead to the temporal diversity decrease of pectinides. Even if their communities were stressed in the Late Messinian, any recovery seems to have been very rapid and completely compensated for any earlier losses. Data on other molluscs (non-pectinide bivalves and gastropods) support this conclusion (Barrier and Montenat, 1990).

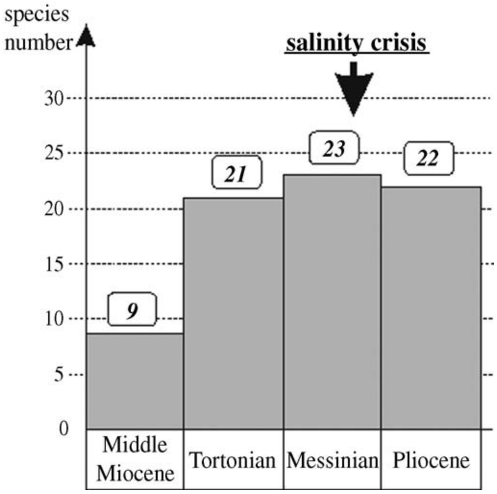

Fig. 9. Total species number of the middle Miocene–Pliocene pectinids in the basins of SE Spain. Initial data extracted from Demarcq (1990).

Therefore, the Late Jurassic salinity crisis in the Caucasus and the much more intense event of the same kind in the late Miocene in the basins of southeastern Spain had different effects on marine faunas. To explain this is a matter for further study.

8. Conclusions

A total of 513 species of bivalves are known from the Jurassic deposits of the Caucasus. In the Early Jurassic diversity among bivalves was low and weakly declining. Later in the Aalenian and especially in the Bajocian the total species number significantly increased by 2.5 times. The next radiation took place in the Callovian. It was so intense that it could be called an “explosion”. The Late Jurassic is characterized by a diversity decrease caused by the regional salinity crisis. The most intense assemblage changes took place at the Sinemurian–Pliensbachian, Pliensbachian–Toarcian, Bathonian–Callovian, and Kimmeridgian–Tithonian boundaries. But only in the first two cases were they caused by true turnovers of taxonomic composition. In later two cases they were connected with the significant diversity changes. This study found that regional and global bivalve diversity changes were somewhat similar. A significant conclusion is that the Callovian diversity explosion is a globally recognized event. But Pliensbachian–Toarcian and end-Jurassic mass extinctions evidently have not been recognized within the Caucasus.

Detailed analysis of sea-level changes (both transgressions/regressions and shallowing/deepening) within the Caucasus, realized by facies interpretation, and marine palaeotemperature changes suggests these factors may not be considered as common direct driving forces of bivalve diversity dynamics. Thus further studies should be aimed at exploring global influences and comparisons with the other regions. This is always important for the estimation of the biodiversity (Alroy, 2003).

An attempted evaluation of principal patterns of the bivalve diversity dynamics and establishment of their correspondence to the global trends of the entire Jurassic make the Caucasus an important region for worldwide comparisons.

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Evolution of NE Africa and the Greater Caucasus: Common Patterns and Petroleum Potential

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Abstract

Comparative geologic studies are important both to find common patterns in the evolution of distinct regions and to reconsider their petroleum potential. We have attempted a comparison of the Phanerozoic evolution between the Northeastern African basins (NEA) and the Greater Caucasus (GC) (Fig. 1). The latter is nowadays a large elongated structure located within the Alpine Belt to the south of the Russian Platform.

Several tectonic/depositional phases have been distinguished in the evolution of NEA (Table 1): 1) Magmatic Arc Phase (Precambrian-Early Cambrian), 2) Graben Phase (Middle-Late Cambrian), 3) Glacial Phase (Ordovician-Silurian), 4) Tectonic Arches Phase (Devonian-Permian), 5) Mediterranean Phase (Triassic-Early Jurassic), 6) Atlantic Phase (Late Jurassic-Early Cretaceous), 7) Alpine Phase (Late Cretaceous-Eocene), 8) Gulf of Suez Phase (Oligocene-Miocene), and 9) Nile Phase (Pliocene-Recent) (Tawadros, 2001, Tawadros, 2003).

The Magmatic Arc Phase was dominated by the Pan-African Orogeny which led to the formation of Gondwana. Gneisses, granites, migmatites, volcanics, and metasedimentary rocks dominate the Precambrian-Early Cambrian succession.

In the Graben Phase extensional tectonics, accompanied by volcanic activities, formed a series of horsts and graben which were filled by shallow marine sandstones and red beds. The Cambrian fractured quartzites form reservoirs at the Attahaddy Field in Libya and the Hassi Massaoud Field in Algeria.

The Glacial Phase was accompanied by a drop in sea level in the Caradocian that led to the formation of incised valleys. In the Caradocian/Ashgillian, peri-glacial sediments filled these valleys. The rise of sea level in the early Silurian Ruddanian stage and the occurrence of an oceanic anoxic event led to the deposition of “Hot Shales” in West Libya, Tunisia, and Algeria. Peri-glacial reservoirs and Hot Shale source rocks form the hydrocarbon system in a number of oil fields in these three countries.

The Tectonic Arches Phase was dominated by the Hercynian Orogeny and the collision between Africa and Laurussia. Formation of tectonic arches, such as the Sirte, Gargaf, and Helez arches and cratonic sag basins, such as the Kufra, Murzuq, and Ghadames Basins, took place during that time. Sandstones, shales, and carbonates were deposited during a gradual sea level fall. The Devonian rocks form reservoirs in West Libya and the Sirte Basin, and Carboniferous rocks form good reservoirs in West Libya, the Sirte Basin, and the Gulf of Suez. The Frasnian Hot Shales are major source rocks in West Libya and Algeria. Thick Permian sediments were deposited in northwestern Libya, the Permian Basin in Tunisia, and the northern Western Desert of Egypt.

The Mediterranean Phase was a period of extension and rifting in Northeast Africa, related to the opening of the Neotethys. Most of the tectonic arches collapsed at that time. Fluvial, deltaic, and shallow marine sandstones and carbonates were deposited. Thick evaporates dominate the succession in northwest Libya and Tunisia. These sediments constitute major hydrocarbon reservoirs in the eastern Sirte Basin and the El Borma Field in Tunisia.

The Atlantic Phase was dominated by the opening of the Central, Southern, and Northern Atlantic oceans. Shallow marine and continental sandstones dominated the succession. The so-called Lower Cretaceous Nubian Sandstones of the Sarir Formation and the fractured quartzites of the Wadi Formation in the Sirte Basin, the Aptian Alamein Dolomite and the Upper Jurassic coaly sandstones of the Khatatba in the Western Desert of Egypt, form lucrative reservoirs. Source rocks are provided by the coaly sediments of the Upper Jurassic Khatatba Formation and the Neocomian Matruh Shale in the Western Desert and by Aptian shales and marls in the eastern Sirte Basin. Economically exploited coal deposits occur at Gebel Maghara in Sinai.

The Alpine Phase started with the collapse of some of the remaining arches. Plate collision and basin inversion, especially during the Santonian and Oligocene, led to the formation of the Syrian Arc System. Carbonates, shales, sandstones, and evaporites were deposited and form effective reservoirs and cap rocks in North Africa. Pinnacle and shelf edge reefs are common especially in Libya. The OAE during the Cenomanian-Turonian, Coniacian-Santonian, Campanian, and Paleocene led to the formation of significant source rocks in NEA. The Campanian source rocks are particularly important in the Sirte Basin (Sirte Shale) and the Gulf of Suez (Brown Limestone). Sandstones of the Waha, Bahi, Baharia, and Abu Roash formations are among the Cretaceous reservoirs. Reefal limestones of the Zelten (Nasser) and Intisar fields are prolific Paleocene reservoirs. Early Eocene nummulitic shoals and bars form prolific reservoirs in the offshore of Tunisia and Libya. The Middle Eocene carbonates are the main reservoir in the giant Gialo Field, with 4 Bbbls of recoverable oil reserves. Economically exploited iron ores (Cretaceous and Eocene) and phosphates (Campanian) in Egypt belong to that phase.

The Gulf of Suez Phase is characterized by the rifting and the opening of the Red Sea, Gulf of Suez, and the Gulf of Aqaba. Reefal carbonates and dolomites, shales (Globigerina marls), sandstones, and evaporites form a complete hydrocarbon system in the Gulf of Suez. Tilted fault blocks, stratigraphic, and combination traps are common. The closure of the Mediterranean Sea and the isolation of the Gulf of Suez led to the deposition of extensive evaporate deposits during the Messinian Salinity Crisis. Incised Valleys (Sahabi, Nile, and Abu Madi channels) were formed inland during the Messinian period. The collision between the African and Eurasian plates led to the formation of a thrust belt and nappes of the Atlas Mountains in Morocco and northern Tunisia and a foredeep basin in the latter. The Serravallian sandstones form the reservoir in the Birsa Field, offshore Tunisia.

The Nile Phase in NEA witnessed the development of the Nile River and its Delta in Egypt. The Messinian incised valleys were filled with shallow-marine sandstones and shales. Fluvial, deltaic, and turbidite deposits were deposited in the offshore of the Nile Delta and are subjected to extensive exploration activities in that area.

Traditional views of the geologic evolution of the GC should be reconsidered, taking into account the new data and new global paleotectonic models. A key point was the identification of the GC as a Gondwana-derived terrane. The close paleoposition of the GC terrane to the Carnic Alps and Bohemian Massif in the Late Paleozoic is argued by faunal and floral similarity, paleomagnetic constraints, and similarities of the sedimentary successions. Such paleoposition of the GC terrane indicates that it was a part of the Hun Superterrane, which was detached from the Gondwanan margin in the Middle Silurian (Stampfli & Borel, 2002).

Tectonic/depositional phases in the evolution of the GC have been distinguished taking into account this new model (Table 1). The history of the GC included 1) Gondwanan Phase (pre-Ludlow), 2) Hunic Phase (Ludlow-Devonian), 3) Proto-Alpine Phase (Carboniferous-Middle Triassic), 4) Left-Shear Phase (Late Triassic-Earliest Jurassic), 5) Arc Phase (Jurassic-Eocene), 6) Paratethyan Phase (Oligocene-Miocene), and 7) Transcaucasus Phase (Pliocene-Recent).

During the Gondwanan Phase (pre-Ludlow), the GC was a part of the Afro-Arabian margin of Gondwana. Numerous evidences have been obtained for the beginning of the Phanerozoic, while special studies of older rocks are still needed. The Cambrian is represented by quartzites, schists and carbonates. The most of the Ordovician is embraced by a major regional hiatus. The Uppermost Ordovician is composed of clastics, which may be related to a periglacial facies. In the Llandovery-Wenlock, schists, clastics, and volcanics dominate.

During the Hunic Phase (Ludlow-Devonian), the GC was a part of the Hun Superterrane, and it was one of the so-called European Hunic Terranes. The Hun Superterrane was detached from Gondwana and drifted northwards to Laurussia. The Paleotethys Ocean was opened at that time. In the Ludlow-Lochkovian interval, carbonates with “Bohemian-type” fauna were accumulated. Volcanics, schists, and sandstones are common in the Pragian-Frasnian succession. A rimmed carbonate shelf existed in the Famennian. At the end of the Devonian, the GC reached the Laurussian margin.

The Proto-Alpine Phase is characterized by strike-slip activity along the Northern Paleotethyan Shear Zone, which was extended westwards as the Intra-Pangaean Shear Zone. Dynamics of these shear zones was caused by the rotations of Africa. In the Carboniferous-Middle Triassic, anticlockwise rotation of Africa caused the right-shear deformations (Swanson, 1982; Rapalini & Vizán, 1993; Ruban & Yoshioka, 2005). At the same time, Hercynian and then Early Cimmerian orogenic events took place in the Proto-Alpine Region, which included the GC. The Mississippian deposits of the GC are schists, clastics, volcanics and carbonates. The Pennsylvanian deposits were continental coal-bearing strata. The Lower-?Middle Permian is a typical red-bed Molasse up to 25,000 m in thickness. A remarkable transgression and a short-lived rimmed carbonate shelf characterized the Late Permian. In the Early-Middle Triassic, carbonates, shales, and sandstones were accumulated.

During the Left-Shear Phase (Late Triassic-Earliest Jurassic) Africa was rotated clockwise, which caused the change of shear-motion direction. The GC moved along the shear zone to its present position to the south of the Russian Platform. In the Norian-Rhaetian, a rimmed carbonate shelf evolved in the GC, while in some areas shales were deposited. The Upper Rhaetian-Lower Sinemurian interval is embraced by a major regional hiatus.

The Arc Phase (Jurassic-Eocene) of the GC evolution comprises the time when island arcs existed at the active margin of the Neotethys (Lordkipanidze et al., 1984). Shales and clastics were accumulated in the Early-Middle Jurassic, but a very large rimmed carbonate shelf existed in the Late Jurassic. In this epoch a regional salinity crisis occurred, and salt was deposited at the same time when reefs existed. Flysch basins evolved during the Cretaceous-Paleogene, and their depth reached its maximum in the Maastrichtian-Danian.

The Paratethyan Phase (Oligocene-Miocene) is characterized by the growth of the Caucasus Orogen. Orogenic chains separated the relatively shallow basin of the Paratethys Sea from the Mediterranean. The salinity crisis, which affected the latter in the Messinian, did not appear in the GC.

During the Transcaucasus Phase (Pliocene-Recent) the principal feature of the GC became the subsiding Transcaucasus Depressions, such as the Rioni Depression and Kura Depression, which are tectonically connected with the Black Sea Depression and the Caspian Depression (Ruban, 2003).

As a result of the comparison of the Phanerozoic evolution between NEA and the GC common patterns for these regions have been established: a) in the Cambrian-Ludlow both NEA and the GC were included in the Gondwanan margin; b) in the Carboniferous-Early Jurassic both studied regions were affected by the major shear zone; c) in the Jurassic-Eocene NEA and the GC were dominated by the development of the Neotethyan structures (Fig. 2).

In NEA, source rocks range in age from the Silurian to the Pliocene and are related to global oceanic anoxic events, except for the Jurassic coals and Miocene marls (Tawadros, 2001). Proven oil reserves of Egypt, Libya, and Algeria are approximately 54 billion barrels.

The proven petroleum reserves of the Azerbaijanian Hydrocarbon Province of the GC are 7-13 billion barrels of oil. They were generated principally by the Majkopian (Oligocene-Early Miocene) source rocks, which were accumulated in locally dysoxic and anoxic environments in the subsiding basins. The Majkopian Group comprises organic-rich shales with interbeds of sandstones (Ali-Zadeh, 1945; Efendiyeva, 2004). Oil was generated from them and stored in the Pliocene Productive Group (traditionally called “Productive Series”), which is the principal exploited reservoir. Depositional environments in the Majkopian Basin were similar to those which existed in the Miocene Gulf of Suez, which generated proven reserves of 1.1 billion barrels of oil. All possible, including minor, petroleum reservoirs in the Azerbaijanian Hydrocarbon Province are shown in Table 2.

The common patterns in the geologic evolution of NEA and the GC suggest that special attention should be paid in the latter to those sedimentary complexes, which are considered as source rocks in NEA (in particular the Silurian, Cenomanian, and Campanian sediments). This suggests that new perspectives in the petroleum exploitation and reserves growth in this region, where the petroleum industry started more than a century ago, still exist.

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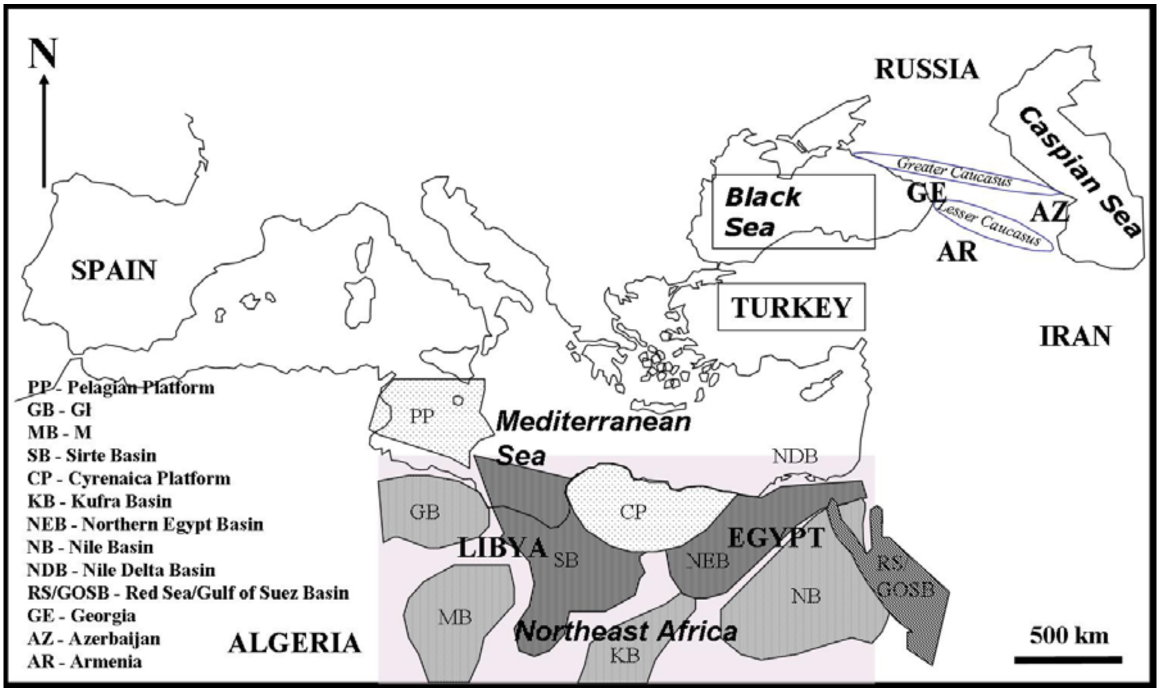
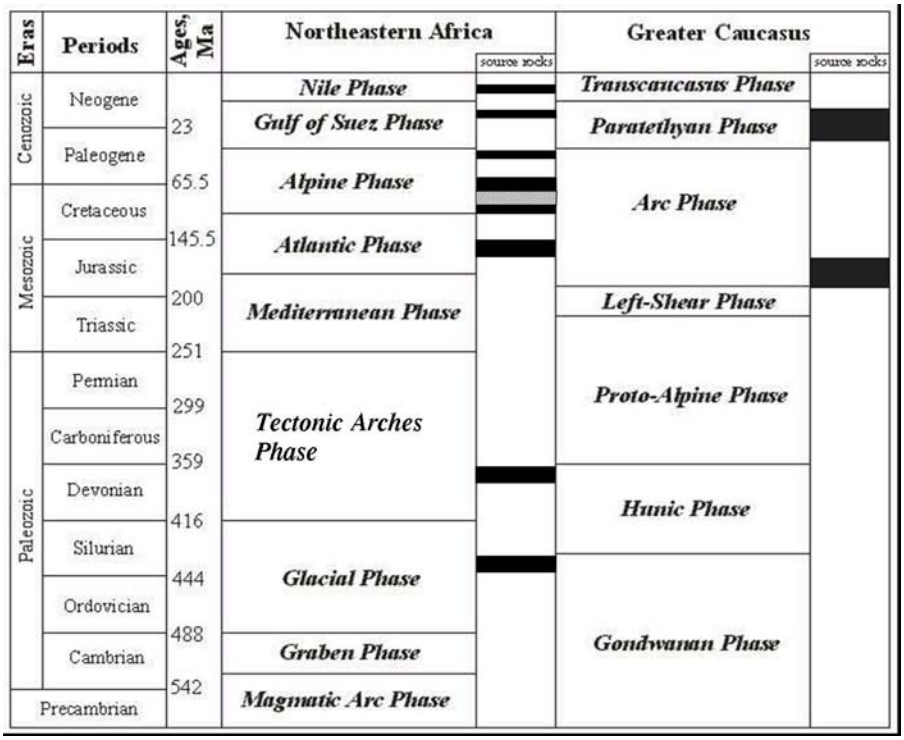


Figure 1. Geographical location of the studied regions (chronostratigraphy and absolute ages after Gradstein et al., 2004).

Table 1. A comparison of the tectonic/depositional phases between NEA and the GC.



Eras	Periods	Ages, Ma	Northeastern Africa	source rocks	Greater Caucasus	source rocks
Cenozoic	Neogene	23	*Nile Phase* *Gulf of Suez Phase*	■ ■	*Transcaucasus Phase* *Paratethyan Phase*	■
	Paleogene	65.5	*Alpine Phase*	■ ■	*Arc Phase*	
Mesozoic	Cretaceous	145.5	*Atlantic Phase*	▨ ■		
	Jurassic	200	*Mediterranean Phase*	■		■
	Triassic	251			*Left-Shear Phase* *Proto-Alpine Phase*	
Paleozoic	Permian	299	*Tectonic Arches Phase*			
	Carboniferous	359				
	Devonian	416		■	*Hunic Phase*	
	Silurian	444	*Glacial Phase*	■	*Gondwanan Phase*	
	Ordovician	488				
	Cambrian	542	*Graben Phase*			
Precambrian			*Magmatic Arc Phase*			

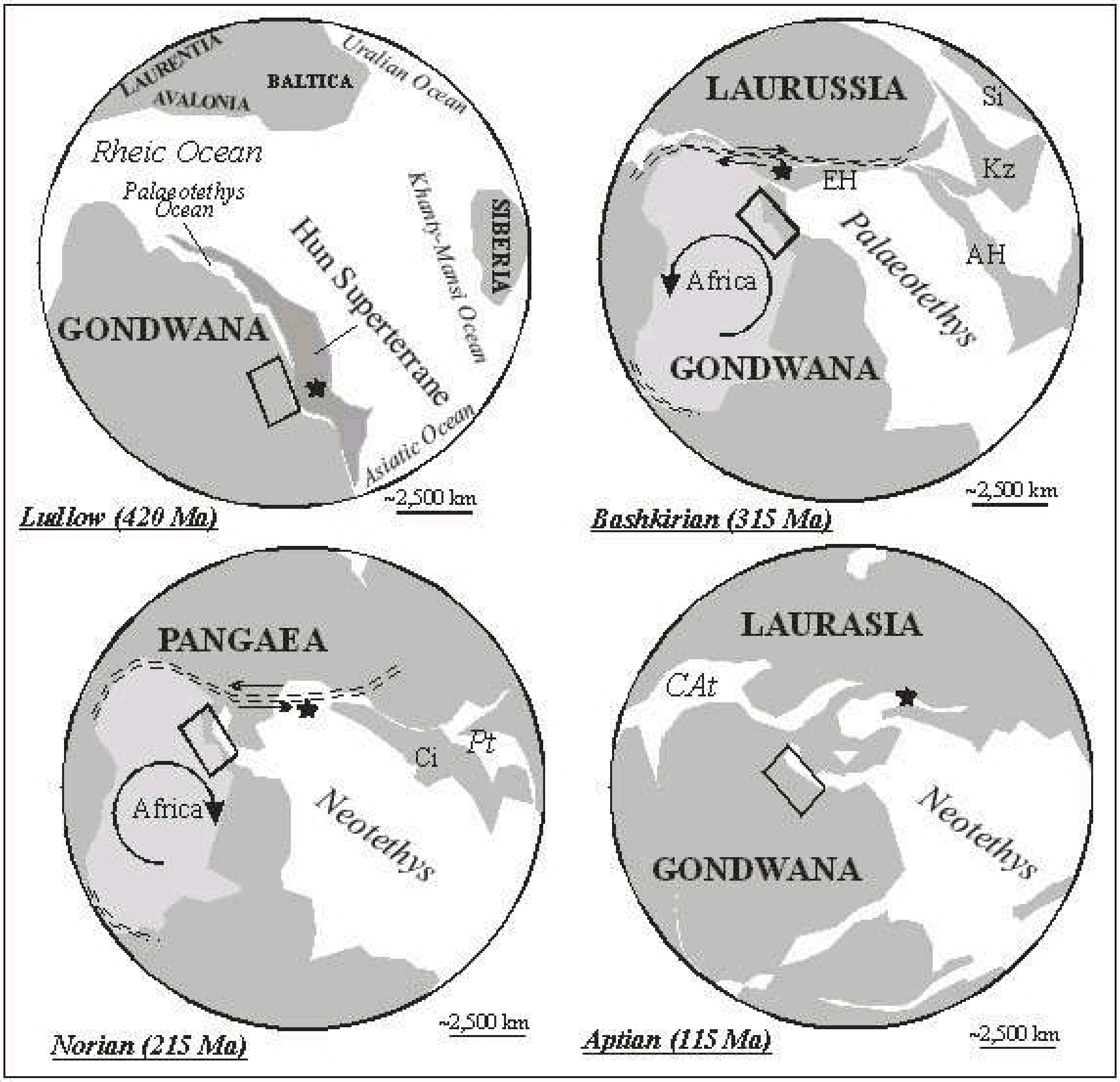


Figure 2. Paleotectonic sketch-maps (modified from Stampfli & Borel, 2002). Rectangle marks NEA, asterisk marks the GC. Abbreviations: EH – European Hunic Terranes, AH – Asiatic Hunic Terranes, Ci – Cimmerian Terranes, Si – Siberia, KZ – Kazakhstan, Pt – Palaeotethys, CAn – Central Atlantics.

Table 2. Possible petroleum reservoirs in the Azerbaijanian Hydrocarbon Province.

Stratigraphy	Region	Outcrop/well Area
Middle Jurassic (Aalenian, Bajocian)	Pri-Caspian-Guba	
Lower Cretaceous (Valanginian, Hauterivian, Albian)	Pri-Caspian-Guba	Sovetobad, Gyadysu, Kurkachidag
Upper Cretaceous-Lower Paleogene (Maastrichian-Danian)	Pri-Caspian-Guba Schamakha-Gobustan, Gyanja	Astrachanka
Paleocene	Pri-Caspian-Guba	Siazan monocline
Eocene	West Azerbaijan	Tarsdallyar, Gyrzundag
Majkop	entire province	Umbaki, Naftalan, Kalamaddyn, Shorbulag, etc.
Chokrak (Middle Miocene)	Gobustan, Apsheron	Umbaki, Duvanny
Diatom Beds (Miocene)	West Apsheron, Shamakha-Gobustan	Binagady, Shabandag, Garadag
Productive Group	Apsheron Peninsula, Apsheron and Baku Archipelagos	40 oil/gas fields