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*Royaume du Maroc*  
*UNIVERSITE MOHAMMED V-AGDAL*  
**Institut Scientifique**  
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N° 26, 2013

## **International Field Symposium “The Devonian and Lower Carboniferous of northern Gondwana”**



**ABSTRACTS BOOK**



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N°26, 2013, Rabat

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International Subcommittee on Devonian Stratigraphy (SDS)  
International Subcommittee on Carboniferous Stratigraphy (SCS)  
IGCP 596 on “Climate Change and Biodiversity patterns in the Mid-Paleozoic”

*in memory of Dr. Volker EBBIGHAUSEN*

22<sup>nd</sup> to 29<sup>th</sup> March 2013

**ABSTRACTS BOOK**

**Ahmed EL HASSANI, R. Thomas BECKER & Abdelfatah TAHIRI (Eds)**

Cover : Devonian succession at Jbel Amelane (panorama shows an overall view of Jbel Amelane, seen from the north, with Emsian to Eifelian section ; box shows Taganic and Kellwasser events top and Famennian at the southern part), Rissani area, Eastern Tafilalt (photo El Hassani)

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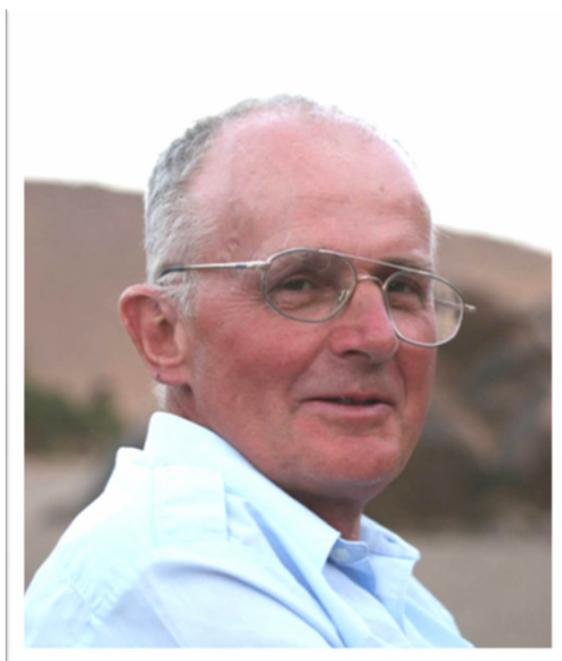
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**Dr. Volker EBBIGHAUSEN**

10.02.1941 – 3.06.2011



*Morocco is world-famous for its beautiful and diverse fossils and has attracted for many years not only the professional community of geologists and palaeontologists, but also a large crowd of amateur collectors. Some of these become so fascinated by landscape, nature and the people that they return to the country over and over and for many years. An even smaller group develops a deeper interest and eventually they turn into specialists that publish their discoveries in a professional manner. Volker EBBIGHAUSEN, mostly in tandem with his close friend Jürgen BOCKWINKEL, collected Moroccan fossils for decades and his/their publications, first in co-operation with experience specialists, but eventually on their own, reached in the last decade the highest scientific level. His work was mostly devoted to Devonian and Carboniferous ammonoids, but he also had a strong interest in brachiopods and he/they would collect the complete fauna at any localities. By this, important material was made available to trilobite, bivalve, gastropod, and coral workers.*

*Volker was born in the Emsland (Lengerich, Tecklenburg County) of northern Germany and attended the gymnasium in Aalen until February 1963. Subsequently, he moved to Munich to study chemistry until 1969. In summer 1972 he received his Ph.D. in chemistry in Munich. Soon after, in 1973, he took a leading position in the analytical laboratories of the famous Bayer AG in Leverkusen. He worked with this large company until his early retirement in 2001. He got married, had two sons, Rodion and Johannes. He started to explore the highly fossiliferous Devonian regions of Bergisch Gladbach and the Eifel region of the Rhenish Massif from his home in Odenthal. Over many years and together with friends he assembled a huge collection of fossils in his home, not only of Devonian age. He was engaged in regional groups of mineral and fossil collectors (Mineralien- und Fossilienfreunde Bayer Leverkusen, Fossilien- und Mineralienbörse Bergisch Gladbach) and became a member of the Paläontologische Gesellschaft. His important collection, including a wealth of Moroccan specimens, has been transferred to the Museum für Naturkunde at Berlin, where he was formally recognized as an independent scientist.*

*I/we had the pleasure of many joint field campaigns with Volker (and Jürgen) since the mid-nineties. This productive cooperation resulted in many joint papers. Our work was in full blossom,*

*when it was abruptly ended by a tragic accident at Volker's home in summer 2011. He was involved with several ongoing and planned publications, many of them also in close cooperation with Dieter KORN and colleagues. He had started to assemble data for a co-authored monograph of the Devonian ammonoids of southern Morocco, which would have met the interest of anybody who bought some of these in the numerous fossil shops of Morocco or elsewhere. Volker (and Jürgen) took an active part in the SDS Field Meetings in the Tafilalt/Maider in 1999 and in the Dra Valley in 2004. During our last field trip and during subsequent visits to Münster he strongly advocated to organize another international field symposium in southern Morocco, which should deal both with the Devonian and Lower Carboniferous. This was agreed by Ahmed EL HASSANI and myself but, in fact, this gathering goes back to Volker's initiative. Therefore, and also in honor of his many and significant scientific achievement concerning the Middle Paleozoic of Morocco, our symposium is devoted to him. He would have appreciated so much to be with us, to show sections, and to discuss aspects of stratigraphy and faunas. His good spirit will be with us.*

*R. Thomas BECKER*

*(photos by J. BOCKWINKEL & S. HARTENFELS)*



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## (P) Conodont biostratigraphy and the timing of facies changes at Ain-as-Seffah (Oued Cherrat Zone, Moroccan Meseta)

ABOUSSALAM, Z. S.<sup>1</sup>, BECKER, R. T.<sup>1</sup>, EICHHOLT, S.<sup>1</sup> & EL HASSANI, A.<sup>2</sup>

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The Moroccan Meseta consists of a complex of structural units with a different history of sedimentation, faunas, and synsedimentary tectonics. The Oued Cherrat Zone is an elongated, ca. N-S running unit S of Rabat. It is characterized by thick, widely quarried Emsian and Givetian reefs and significant Eovariscan erosion and re-sedimentation in the Famennian. The general stratigraphy has been investigated by CHALOUAN (1977) and CHALOUAN & HOLLARD (1979). In the southern Oued Cherrat area, there are two separate reef belts west (Dhar-es-Smene Formation) and east (Cakhrat-ach-Chleh Formation) of the deeply incised Cherrat Valley but both yielded Emsian conodonts at their bases. The Givetian reefs were mostly eroded and reworked during a major Famennian uplift event, for example at Ain Dakhla. An isolated small Givetian reef and adjacent deeper water strata occur at Ain-as-Seffah (= Ain-as-Safah) but are discontinuous laterally. The whole complex is assigned to five different members of the Ain-as-Seffah Formation and surrounded by poorly dated, overlying Famennian clastics, which are separated by a ca. N-S running fault zone from the Eifelian Ain-Kheneg-an-Nmer Formation. The latter is a Flinz-type thick succession of dark-grey to black distal turbiditic limestones, up to 20 cm thick, often laminated and with crinoid debris, rarely with small brachiopods, and poorly fossiliferous dark shales. A sample from its upper part produced *Polygnathus angustipennatus*, *Po. eiflius*, *Po. parawebbi*, *Linguipolygnathus linguiformis* Morphotype 1b, *Tortodus kockelianus*, and *Icriodus anterodepressus*, which indicates a top Eifelian age.

Sharply separated beds below and laterally to the lower part of the reef limestone consist of thin-bedded, often lenticular, fine-grained limestones embedded in calcareous shales (Units A-B), grading upwards into flaserlimestone (Unit C). This Member 1 contains *Icriodus obliquimarginatus* and *I. difficilis* at the top of Unit B, which suggests the base of the *difficilis* Zone near the base of the Middle Givetian. *Po. xylus*, *L. linguiformis*, and *L. klapperi* join *I. difficilis* in Unit C and represent a slightly higher level of the same zone. Rare *Belodella* are typical for a peri-reefal setting. Conodonts from the irregularly bedded base of the massive reef limestone (Member 2, basal Unit D), more than 15 m below the sample from the top of Unit B, also belong to the basal Middle Givetian. However, here *I. difficilis* and *I. obliquimarginatus* are associated with *Po. pseudofoliatus* and *Tortodus sardinia*.

Laterally to a position 7-8 m below the top of Unit B, crinoid-rich reef limestone yielded a typical Upper Givetian conodont fauna with *Po. cristatus ectypus*, *Po. aff. pollocki* (fragmentary), and *I. expansus* (ca. *ectypus* Zone). A sample from just 50 cm above the top of Unit C contains the Upper Givetian *Po. pennatus* together with *Po. xylus*. Both samples prove a disconformable contact between the pre-reefal Member 1 and the big boulders of Member 2. It seems that the mass of reefal limestones has been moved in relation to the lateral matrix. The highest reef blocks belong to the top Upper Givetian, based on an association of *Po. dubius*, *Po. ?webbi*, *Po. pollocki*, *Po. paradecorosus*, *Ctenopolygnathus lanei*, *I. subterminus*, *I. tafilensis*, and *I. symmetricus symmetricus*. The persisting conodont record of Member 2 documents a permanent open marine faunal influx, without an outer barrier.

Laterally, at the southern end of the outcrop and more than 10 m below, the reefal facies changes into massive, crinoid-rich brachiopod limestone (Member 3). Its base yielded *Mesotaxis guanwushanensis* (= *falsiovalis*), *Po. dubius*, *Po. webbi*, and *Ancyrodella rotundiloba pristina*, a typical basal Frasnian

assemblage (MN 1 Zone). Just 1.4 m higher, the same massive limestone includes a rich and very diverse conodont assemblage from near the Lower/Middle Frasnian boundary, with *Ad. africana*, *Ad. pramosica*, *Ad. gigas* Morphotype 1, *Palmatolepis transitans*, *Mes. asymmetrica*, "*Mes.*" *unilabius*, and others. In comparison to the thick reef boulders, the Lower Frasnian is strongly condensed and wedges out on the steep reefal slope (to the north). It suggests a stepwise drowning of the small bioherm/carbonate platform by the transgressive pulses of the Frasnian and Middlesex Events.

Member 4 consists of thin-bedded, partly argillaceous flaserlimestones that overlie Member 3 (in the southern corner) or that lie with an unconformity besides the youngest reef blocks. At the base *I. symmetricus symmetricus* and *Po. paradecorosus* dominate over *Ad. gigas* Morphotype 1. A single *Pa. plana* provides an age close to the Middle/Upper Frasnian boundary (at least MN Zone 10) and evidence for a significant hiatus at the base. Laterally, in a sample from 2.8 m above the marker tree, *I. praealternatus*, *I. symmetricus curvatus*, and *Ad. curvata* early morphotype indicate a similar age. At the top (ca. 2 m higher), just adjacent to the last reef boulder, there is a different, strongly deformed fauna with *Ancyrognathus triangularis*, *Ag. coeni*, *I. alternatus alternatus*, *Ad. curvata* late form, *Pa. muelleri*, and others. These indicate MN Zone 12 and a significant condensation of Member 4.

Member 5 is the ca. 65 cm thick, reddish, condensed *Manticoceras* Limestone that formed the original justification for the Ain-as-Seffah Formation. The goniatites (*Manticoceras cordatum* Group) are rather deformed and this is also true for most of the associated conodonts, which complicates their precise identification. Apart from many orthocones and crinoid ossicles, there are diverse palmatolepids, with platform shapes as in *Pa. hassi*, *Pa. proversa*, *Pa. plana*, *Pa. ljaschenkoae*, *Pa. amana*, and *Pa. bogartensis*. *Ad. curvata* late form, *Ag. triangularis*, *Ag. asymmetricus*, and various polygnathids add to a fauna from near the base of MN Zone 13. Therefore, the *Manticoceras* Limestone is tentatively correlated with the Lower Kellwasser level of Germany.

Above the *Manticoceras* Limestone, there is a sudden change to more than 10 m of unfossiliferous red shales that resemble basinal (deep-water) "Cypridina Shales" from the Famennian of Germany. This unit is only developed laterally to the last reef boulders and is here defined as a local Lower Member of the clastic Al Brijat Formation. Its age is not known; in Germany similar red shales are typical for the upper part of the lower to middle Famennian. The main Al Brijat Formation consists above the reef of ca. 30 m shales and siltstones with irregularly bedded and partly folded, lenticular sandstones, and ca. 20 m fine, mostly olive (apart from a red interval at the base) pelites with some siltstone nodules. Laterally, on the next hill to the north, where the small reefal complex has sharply wedged out, these clastics are thicker. Attempts are under way to date them with the help of palynomorphs.

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## **(P) Carboniferous paleosols and paleolandscapes of East European Craton (southern Moscow Basin, Russia)**

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This work is the detailed and interdisciplinary record of the recently discovered Carboniferous paleosols of the Moscow Basin (~20 profiles of Late Mississippian and Pennsylvanian paleosols) which endeavours landscape reconstruction including types of soils and associated terrestrial sediments, moisture and temperature regime, general features of vegetation and soil biota, and temporal changes of the lowstand-phase paleoclimates. Stacked paleosols in shallow-marine sedimentary successions indicate repeated subaerial exposures caused by complex interplay between sediment aggradation and local and/or global changes of sea level (RETALLACK 2001).

The upper Mississippian (Lower Carboniferous) of the Moscow Basin is one of the brightest examples of epeiric – sea carbonate sedimentary systems with numerous subaerial exposure horizons marked by paleokarsts and paleosols. We present the results of the detailed study of three upper Mississippian unconformities at the Mikhailovian - Venevian (2 profiles) and Venevian-Tarusian boundaries. All three profiles are calcimagnesian systems containing micritic calcite and are characterized by the following common properties. Carbonate (35-78 %) is fresh water calcite with  $\delta^{13}\text{C}$  - 2.3 ‰ to -7.1 ‰ of bulk samples and  $\delta^{13}\text{C}$  - 5.9 ‰ to -10.7 ‰ of micrite. The main clay mineral is authigenic Mg-rich trioctahedral smectite-saponite. The beds are almost free from terrigenous material ; the coarse (>2  $\mu\text{m}$ ) fraction besides fresh water calcite nodules contains traces of quartz and dolomite. Authigenic saponite is developed in alkaline media with large MgO (min. 6 %) and small Si and Al concentrations. Such conditions correspond to shallow marine or salted continental basins of arid/semi-arid climates with prevailing evaporate conditions. We interpret the studied beds as extremely shallow-water lacustrine deposits transformed by repeated subaerial emersion episodes into palustrine pedosediments.

In the soil cover of the territory of the Moscow sedimentary basin during the upper Mississippian hydromorphic soils prevailed: histosols, gleysol. The principal clay mineral in all studied soils is smectite (low charge montmorillonite and saponite). Gleysol and automorphic soils contain pedogenic micrite (calcite). Some top layers contain gypsum. Organic matter (OM) is preserved in the form of coal; aromatic structural components dominate.

While the upper Mississippian paleosols are smectitic, in Pennsylvanian paleosols palygorskite (some times sepiolite) became the principal mineral testifying the aridity of Upper Carboniferous climate on the territory of Moscovian sedimentary basin (KABANOV et al. 2010). The majority of upper Moscovian paleosols surveyed formally conform to lithic (rendollic) haplocalcids within the aridisol class or rendzic calcic leptosols. A detailed study of a representative Podolskian paleo-pedon reveals the development of shallow soil carbonate, low alumina/bases and Ba/Sr ratios, enhanced Mn and Sr content, the presence of soil gypsum and opal, and a characteristic peak in magnetic susceptibility, all suggesting a semiarid to arid pedogenic environment. The palygorskite clay of this paleopedon retains 1.1–1.5% of connate organic matter in the form of covalently bound organomineral complexes, and is fulvatedominated, resembling organic matter from extant dryland soils. Under periodical increasing precipitation palygorskite transformed into beidellite via dissolution-precipitation (vertisol) or via structural reorganization (rendzinas). Development of vertisols, pedogenic carbonate accumulations testify to a seasonal type of

climate. Starting from the (regional) Venevian substage with the gradual aridization of climate, the global role of Mg in mineral formation increased: saponite, dolomite, palygorskite, sepiolite.

Most of the studied paleosols contains inborn OM, detectable by chemical methods and  $^{13}\text{C}$  NMR spectroscopy. The organic carbon content and composition depends on the mechanisms of OM protection against diagenetic transformations, which are determined by the mineralogical composition and interactions between mineral matrix and organic molecules. In the case of smectite, OC preserved in a form of aromatic structures (coal, detritus) or is not preserved (vertisols). The remarkable preservation of fulvic type humus in Pennsylvanian (semi) arid paleosols is related to the formation of organo-mineral derivatives, where organic molecules are chemically bound to palygorskite lattice via grafting reactions between organic molecules and Si-OH groups of the mineral matrix. Found palygorskite-humic substances derivatives seems to be the earliest evidence for Pennsylvanian (semi) arid humification and pedogenesis (ALEKSEEVA et al. 2009).

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## Devonian paleosols of Voronezh region (European Russia): geochemistry, mineralogy, paleoenvironments

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Devonian (Givetian- Frasnian) age complex of 4 paleosols was discovered by our team in 2010 in Voronezh region (European Russia). Devonian deposits of Central Russia on the northern slope of the Voronezh Anticline are known as Central Devonian Field (LEBEDEV et al., 2010). These deposits, especially Yastrebovskaya age suite are famous due to the very rich and unique complex of terrestrial Devonian flora (KRASSILOV et al., 1987). Discovered paleosols (PS) are developed from terrigenous material: argillites of Ardatovskaya age suite in the bottom (PS1) and tuff – sandstones of Yastrebovskaya age suite in its middle part and in the top (PS2-PS4). The whole thickness of the complex is about 6 m. All 4 paleosols are characterized by good profiles preservation; one of them (PS 3) – by the unique preservation due to its fast burying under the thick (1 m) layer of dense argillite (Fig.1). The presence of this layer provided the complete preservation of the PS3 profile including organic A-hor and *in situ* large-scale rooting systems of Progymnosperms (Fig. 2). Paleosol profiles contain organic carbon (1.25-2.25 %), which is presented by coaly large plant fragments of *Callixylon – Archaeopteris*, coaly detritus, well preserved spores. Roots have diameter from 0.5 upto 5 cm, the diameter of the most common roots is 1 – 2 cm. On their cross section roots have the concentric structure, outer surface is covered by clay cutans. Roots channels are mineralized, mostly by goethite, sometimes - by pyrite. All profiles have the similar construction: they are differentiated in granulometry and carbonate content, bottom parts of all 4 paleosol have pink colored horizons (2.5 YR 4/2) rich in hematite. Carbonate (upto 6%) is presented by siderite (FeCO<sub>3</sub>). All profiles have pedofeatures – nodules (2.5.Y 4/4). Their size is within the wide range: from 1-2 cm upto 20 cm. The main mineral component of these pedofeatures is siderite. Besides they can contain goethite and quartz. Stable isotope analysis of <sup>13</sup>C of siderite shows the δ <sup>13</sup>C values between (-10) – (-11) ‰, indicating the biogenic (pedogenic) nature of siderite. The main mineral of all 4 soils is kaolinite. Besides some horizons contain different Fe - minerals: ilmenite, siderite, hematite, goethite, pyrite; sometimes additionally they contain quartz. The joint presence of these Fe- minerals can say about the diverse and complicated paleoecological environments of soils formation. The main mineral component of *clay fractions* from all soils is kaolinite. Some horizons are monomineral - 100% kaolinitic, others additionally can contain goethite. The later is most common in the root containing horizons. The obtained characteristics suggest warm – wet with seasonally arid environments of PSs development. Soils were semi – hydromorphic with stages when oxidizing conditions prevailed.

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Fig. 1. The image of Givetian age pedocomplex (without the top PS4).



Fig.2. PS3 with *in situ* roots of Progymnosperms.

## Viséan corals from the transverse Jebel Begaa to Gara El Itima (eastern Tafilalt, Morocco)

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The macrofauna from the Viséan succession of the Tafilalt is, with exception of ammonoids (e.g. Korn et al. 1999; Klug et al. 2006) and gastropods (Heidelberger et al. 2009), not well known, which is in sharp contrast to the well-known and intensively studied faunas of the surrounding Devonian strata. The record of Viséan corals for the eastern Tafilalt is generally restricted to faunal lists, often on the genus level (e.g. Wendt et al. 2001).

In 2011, more than 300 coral specimens have been collected in the eastern Tafilalt from a series of surface outcrops on a south-north orientated transverse starting at the Jebel Begaa and ending at the Gara El Itima. Corals have mainly been found in the carbonated horizons of the Merdani and Zrigat formations, fewer samples are from more detrital facies of the same formations. The recovered fauna is of high interest since it comes from a transitional palaeogeographical position between the faunas of the Saharan platform (e.g. Aretz 2011) and the faunas of the Moroccan Meseta (e.g. Aretz 2010, Said et al. 2012). Here we present the first results of an ongoing study of these specimens.

The oldest coral assemblage was found in a few meter-thick interval composed of a lower conglomeratic part and higher up bedded to massive limestone west and northwest of the Jebel Begaa. Lateral facies changes are relatively common in this interval. The fauna comprises a series of small solitary rugose corals like "*Pentaphyllum*" sp., *Rylstonia* sp., *Sychnoelasma* ssp. 1, *Zaphrentites* s.l. sp., and *Cravenia* ssp., but also medium-sized solitary rugose corals like *Siphonophyllia* sp., *Caninomorpha* indet., *Cyathoclisia* sp., *Koninckophyllum* ? sp., and *Merlewoodia* sp. The only colonial are tabulates corals belonging to the *Michelinia*. This coral assemblage is considered being of Early Viséan age.

An outcrop of massive limestones further towards the Jebel Begaa contains *Sychnoelasma* sp., for the first time colonial rugose corals (*Siphonodendron scaleberense* Nudds & Somerville, 1987) and large-sized siphonophyllids, provisionally attributed to *Siphonophylliasamsonensis*. This fauna is most likely late Viséan in age; *S. scaleberense* being a common taxa in the Asbian (Poty et al. 2006). The bedded limestones forming the upper part of the Jebel are very poor in corals, but they contain the late Viséan marker, *Dibunophyllum bipartitum* McCoy, 1849.

Further late Viséan corals were found on and along the ridge (Hassi Nebech) east of the Jebel Begaa. There, small and large-sized solitary and colonial rugose corals occur (*Zaphriphyllum* ssp.; *Cravenia* sp.; *Siphonophylliasamsonensis*?; *Siphonophyllia* sp. 1; *Rylstonia sguilmensis* Semenoff-Tian-Chansky, 1974; *Amygdalophyllum* sp.; *Siphonodendron* gr. *kleffense* Schindewolf 1927; *Solenodendron horsfeldi* (Smith & Yu, 1943) and tabulate corals (*Michelinia* ssp.).

Further towards the north, the detrital facies (dominantly sandstones and siltstones) of the Mougui Ayoun Formation is devoid of corals. It is in the lower part of the Zrigat Formation, the carbonate environments of which have been described as mud-mounds by Wendt et al. (2001), when corals reappear. This assemblage contains a diverse Asbian coral fauna. *Rylstonia benecompecta* Hudson & Platt, 1927, *Rylstonia* sp., *Rotiphyllum* sp., *Cyathaxonia cornu* Michelin, 1847, *Cyathaxoniarushiana* Vaughan, 1906, and *Zaphrentites* s.l. sp. have been identified among small solitary rugose corals. Large solitary corals are represented by *Merlewoodia* sp., *Siphonophylliasamsonensis*?, *Siphonophyllia*? sp., *Pseudozaphrentoidesalloyteai* Semenoff-Tian-Chansky, 1974, *Caninomorpha* indet., *Dibunophyllum bipartitum* (McCoy, 1849), *Koninckophyllum* cf. *destitum*, *Clisiophyllum*? sp., *Archnolasma*? sp., *Pareynia*

sp., and *Axophyllum pseudokirsopianum* Semenoff-Tian-Chansky, 1974. Colonial rugose corals are *Siphonodendron irregulare* (Phillips, 1836), *Siphonodendron pauciradiale* (McCoy, 1844), *Siphonodendron martini* (Milne-Edwards & Haime, 1851), *Lithostrotion decipiens* (McCoy, 1849), and *Solenodendron furcatum* (Smith, 1925). There are also abundant tabulate corals (*Michelinia* ssp., *Multithecopora* sp.).

The youngest faunal assemblage comes from the highest carbonate levels of the Zrigat Formation just south and east of the Gara El Itima. It again comprises a diverse range of forms and contains *Zaphriphyllum* sp., *Zaphrentites* sp., *Rylstonia sguilmensis* Semenoff-Tian-Chansky, 1974, *Rylstonia laxocolumnata* Semenoff-Tian-Chansky, 1974, *Caninia* sp., *Siphonophyllia* sp., *Dibunophyllum bipartitum* (McCoy, 1849), *Clisiophyllum garwoodi?*, *Axophyllum pseudokirsopianum* Semenoff-Tian-Chansky, 1974, *Axophyllum?* sp., *Diphyphyllum* sp., *Michelinia* ssp., and *Palaecis* sp. This assemblage lacks typical Brigantian markers, but Brigantian ammonoids have been found in this stratigraphic level (Klug et al. 2006).

The establishment of the original spatial distribution of the different coral morphologies and organizations is hard to make due to often poor to moderate outcrop conditions and coral specimens widely distributed as loose specimens on more or less flat surfaces. It seems reasonable that small and often undissected solitary corals are from the more marly or shaly parts of the succession, whereas the larger and colonial rugose corals are more often found in poorer and more massive carbonate facies.

From a palaeobiogeographical point, the coral assemblages are typical Viséan faunas of the western Palaeotethys. Our first results indicate that the eastern Tafilalt assemblages show not surprisingly strong similarities to the nearby Viséan basins of the Sahara characterizing the northern margin of Gondwana. This is well seen in the number of species, which have often been considered endemic to the Béchar Basin. These species are so far the main difference with the faunas of the mobile Variscan belt towards the north. The typical colonial axophyllid corals characterizing the basal Brigantian in Armorica (Central Europe and Iberia) have not been present in our assemblages. This may be the result of unfavorable regional facies, overall rareness of these forms in Northern Africa, or a bias in our sampling.

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## (P) Conodont characteristics of the Emsian-Eifelian deposits in the Sakaska section (the South Urals)

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The Emsian-Eifelian boundary at the base of *partitus* Zone is established with uncertainty in the Urals. In the western South Urals subregion, the Emsian-Eifelian interval is represented by the Biya and Afonian Horizons. There are quite a few sections, where zonal index species were discovered. The first finds of *Polygnathus costatus partitus* are from the Taltiyian Horizon in the eastern Middle Urals (SAPEL'NIKOV et al. 1981) and from the Biya Horizon in the western South Urals (MASLOV et al. 1983), which correspond to the regional *Zdimir pseudobaschkiricus* - *Megastrophia uralensis* brachiopod Zone.

The Biya Horizon is characterized in typical sections by biohermal limestones with rich benthic fauna: stromatoporoids, gastropods, brachiopods, ostracodes, trilobites, crinoids, and others.

The Sakaska section is located in the south of the West Uralian Folded Zone, at the right bank of the Belaya River. A special feature of this section is the exposed Biya-Afonian boundary, which is rarely found in other sections. Through the whole of the exposed Biya deposits, remains of tentaculites, ichthyofauna and conodonts are found. Grey and light-grey organogenic and organogenic-detrital bedded limestones, 21.5 m thick, compose the sequence. Microorganoclastics components and bed thicknesses vary.

Overlying Afonian deposits (regional *Agoniatites*-*Bornhardtina plana*-*Chaskothyris tchernyschewi* Zone) form an alternation of dark-grey bituminous tentaculitic limestones and black clayey-siliceous shales. Afonian rocks are folded and it is difficult to estimate their thickness (ca. 50-60 m).

The Biya-Afonian boundary interval is 2.0 m thick and marked by Bed 17, terrigenous-carbonate rocks in alternation with black to grey, clayey-siliceous shales and black bituminous tentaculitic limestones (1-5 cm) with scattered iron balls. The rocks are disharmonically folded. The base of this bed is the Biya-Afonian Horizon boundary.

Biya deposits in the Sakaska section are divided into three conodont zones (ABRAMOVA & ARTYUSHKOVA 2004). The main part (Beds 1-14, 21.3 m thick) belongs to the *patulus* Zone. The zonal species was found 3.1 m below the Biya roof but we add the lower interval with *Polygnathus kimi* to this zone, too. The interval 0.27 m below the Biya roof corresponds to the *partitus* Zone (Bed 15, 7 cm thick). Above, the *costatus* Zone begins 0.1-0.2 m below the Biya roof and includes the subsequent Afonian deposits.

The analysis of the conodont distribution through the section shows that the conodont richness is not even in different beds. In the lower part the diversity is poor in taxa. In detrital limestones the conodont content is higher. Cosmopolitan species, such as *Polygnathus serotinus*, *Polygnathus linguiformis bulynecki* and *Polygnathus linguiformis linguiformis*, number up to 20 specimens per 0.8 kg of rock, reaching 66 specimens at end of the *patulus* Zone. Numerous *Pseudooneotodus* sp. and taxa of *Panderodus* appear in Bed 7, passing through Beds 8-14. They outnumber other species at the end of the *patulus* Zone. Many taxa that are widespread in the *patulus* Zone disappear in the *partitus* Zone. Some taxa of *Icriodus* appear in the *partitus* Zone. The beginning of the *costatus* Zone is characterized by a renewal of the conodont complex, which includes the previous taxa of the *costatus* Group. Deposits of the latter do not contain benthic fauna.

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**Plate 1:** Conodonts from the Biya and Afonian deposits (upper Emsian and lower Eifelian) of the Sakaska section (western South Urals). All pictures are in equal magnification.

**Figs. 1, 2, 9, 10, 21** – *Polygnathus serotinus* TELFORD, 1975.

1, 2 – upper views; 10 – lower view, Sample 01009, Bed 5; 9 – lower view, 21 – upper view, Sample 0988, Bed 13.

**Fig. 3** – *Polygnathus kimi* MASHKOVA & APEKINA, 1981.

Upper view, Sample 01007, Bed 4.

**Figs. 4-6, 11, 14, 15, 25** – *Polygnathus linguiformis bultyncki* WEDDIGE, 1977.

4 - upper view, Sample 01029/1, Bed 7; 5, 15, 25 – upper views, Sample 01009, Bed 5; 6 – upper view, Sample 01026, Bed 3; 11 – upper view, Sample 0987, Bed 13; 14 – upper view, Sample 01001, Bed 9.

**Fig. 7, 31, 32** – *Polygnathus costatus sogdianensis* BARDASHEV, 1991.

7 – upper view, Sample 0977, Bed 16; 31 – upper view, Sample 0978, Bed 15 ; 32 – upper view, Sample 01020, Bed 20.

**Fig. 8** – *Polygnathus linguiformis pinguis* WEDDIGE, 1977.

Upper view, Sample 0977, Bed 16.

**Fig. 12** – *Polygnathus* aff. *willii* BARDASHEV, 1991.

Upper view, Sample 01004, Bed 8.

**Fig. 13** – *Polygnathus foliformis* SNIGIREVA, 1975

Upper view, Sample 0980, Bed 14.

**Fig. 16** – *Tortodus* sp.

Lateral view, Sample 0987, Bed 13.

**Fig. 17** – *Polygnathus trigonicus* BISHOFF & ZIEGLER, 1957/

Upper view, Sample 0978, Bed 15.

**Fig. 18, 22** – *Polygnathus costatus patulus* KLAPPER, 1971.

18 – Upper view, Sample 0998, Bed 10; 22 – upper view, Sample 0979, Bed 15.

**Fig. 23** – *Polygnathus* aff. *benderi* WEDDIGE, 1977.

Upper view, Sample 0978, Bed 15.

**Fig. 24** – *Polygnathus costatus oblongus* WEDDIGE, 1977.

Upper view, Sample 01018, Bed 19.

**Fig. 26** – *Icriodus* aff. *norfordi* CHATTERTON, 1978.

Upper view, Sample 0979, Bed 15.

**Fig. 27, 29** – *Polygnathus costatus partitus* KLAPPER, ZIEGLER & MASHKOVA, 1978.

27 – Upper view, Sample 0978, Bed 15; 29 – upper view, Sample 0979, Bed 15.

**Fig. 28** – *Polygnathus linguiformis linguiformis* HINDE, 1879.

Upper view, Sample 01000, Bed 10.

**Fig. 30** – *Pandorinellina expansa* UYENO & MASON, 1979/

Lateral view, Sample 0979, Bed 15.

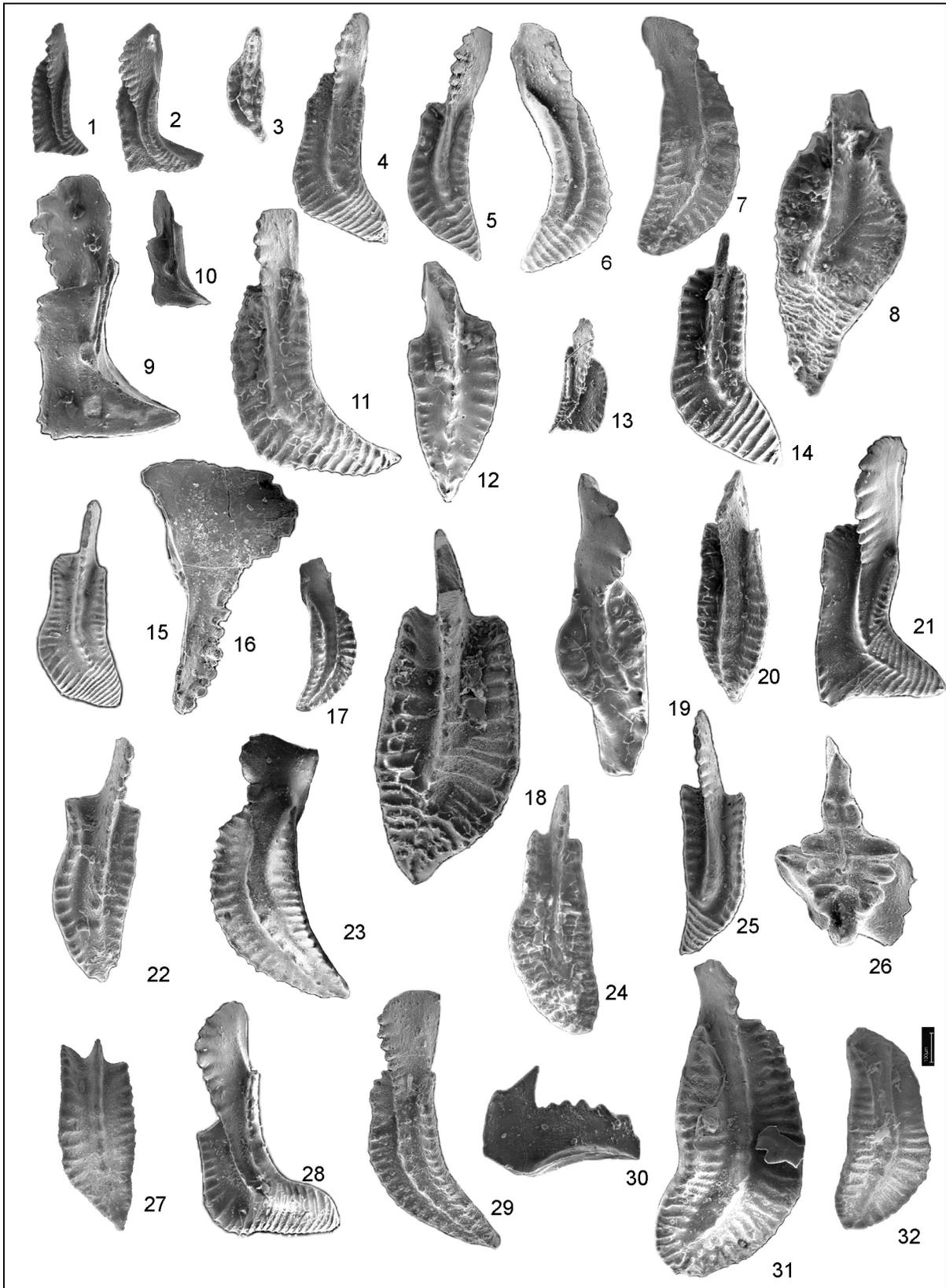


Plate 1. Conodont distribution in the Sakaska section (South Urals).

## **(P) Carboniferous in Central East Iran Microplate. Mississippian/Pennsylvanian boundary interval.**

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Our interest focused on the Mid-Carboniferous Event and its influence on conodonts, which are very sensitive to environmental changes. All global bio-events can be related to environmental changes such as sea-level changes and anoxic events. Concerning the bio-event characteristics of Carboniferous conodont faunas, there are indicators for global changes at the Mississippian/Pennsylvanian boundary. The Sheshangosht and Howz-e-Dorah sections are examples for the replacement of dominant Lower Carboniferous genera *Gnathodus* and *Lochriea* by post-event dominant genera *Declinognathus* and *Idiognathodus* in the Upper Carboniferous. After the bio-event related extinction, there is a drastically increasing diversity of new conodont genera descendent of *Gnathodus*. The Mid-Carboniferous event coincides with a strong regression documented by terrigenous sedimentation, cross-bedded facies, and oolitic sandy limestones. Carboniferous corals are widely distributed in shelf or ramp environments and contribute to the formation of reefs. A crinoidal unit in Sheshangosht section is situated at the Mississippian/Pennsylvanian boundary, indicated by two conodont zones: *Rachistognathus muricatus* Zone and *Declinognathus noduliferous* Zone. The *R. muricatus* Zone is the uppermost conodont zone of the Mississippian (upper Serpukhovian), whereas the *D. noduliferous* Zone is a marker of the Middle Carboniferous boundary between the Mississippian and Pennsylvanian and is the lowermost conodont zone of the Pennsylvanian.

The presence of coral beds in all studied sections stimulated our interest as well as the differentiation of environmental conditions in the sedimentation of the studied sections after the widespread of Lower Carboniferous shale and carbonates, replaced by strongly condensed carbonates. This study provides a detailed sampling across the Crinoidal Limestone in the Darchaleh, Howz-e-Dorah, and Sheshangosht sections in comparison to the previously studied Ramsheh, Asadabad (BONCHEVA et al. 2007), and Howz-e-Dorah (YAZDI 1999) sections. It is aimed to prepare a stratigraphical correlation scheme for all studied sections

Continental shelf deposits of the Upper Devonian have extended further to the Lower Carboniferous in most localities of Iran, after a negligible gap, except at Zagros and Azerbaijan. As a result of tectonic movements, related to the Hercynian orogenic phase, the extensive areas of the Iranian Platform sustained erosion, so that much thickness of lower Paleozoic rocks is missing, for example in the Chah-Riseh area - Northwest Isfahan (GOLAMAILAN 2007, YAZDI et al. 2000), in Zefreh - Northwest Isfahan (BRICE et al. 2006), in Soh-Northwest Isfahan (ADHAMIAN 2003), and in the Dalmeh – Yazd area (HAIRAPETIAN & YAZDI 2003).

Regional data show that on the contrary to the Upper Devonian and Lower Carboniferous, when sea level has been at a maximum level, glaciers covered most parts of Gondwana in Namurian and later in the Sakmarian series. A glaciation event has been correlated with the lowest sea level in the upper Namurian and lower Westphalian. The variations of Carboniferous lithofacies in Central Iran are considerable and it seems that the Lower Carboniferous platform, in Central Iran, covered structural blocks with different sedimentary characteristics. Faults had a main role in their separation.

Carboniferous coral horizon of the Iran Microplate reveal as a richly fossiliferous oolitic limestone conodonts together with brachiopods, gastropods, bryozoans, and ostracodes. The presence of ooides

comes from restricted near-coast marginal-marine parts of the carbonate platform. Fragments of macrofauna show as well a littoral environment, with low salinity waters in shallow shelf platform conditions. A widespread lower Moscovian transgression makes the environment perceptible for the flourishing of cosmopolitan species including conodonts. It gives opportunities for a reliable correlation of basins by marine sedimentation.

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## **(P) New Data on the Stratigraphy of the Lower and Middle Devonian of Salair (West Siberia, Russia)**

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Recent biostratigraphic studies of Salair sequences clarify the understanding of sedimentation and of the development of biota in the Salair-Kuznetsk segment of the Devonian paleobasin. At the same time these raise a number of debatable questions. The following results were obtained:

1. The age of the Kaibalian Suprahorizon (lower Lochkovian) was clarified. In the parastratotype of the Sukhoi Formation near the Gur'evsk town, the Devonian conodonts *Pelekysgnathus* cf. *P. serratus* JENTZSCH and *Pandorinellina exigua philipi* (Klapper) were found. At the top of the Tomsky Zavod Formation and Peetz Formation *eurekaensis* Zone conodonts were identified (IZOKH & CHERNIGOVSKY 2011).

2. A significant improvement of the volume and age interpretation of the Salairka Horizon was achieved. First of all this applies to the Upper Salairka Beds. This stratigraphic unit is now proposed to be allocated in the interval of Beds 15-20 of the type section of the Telengitian Suprahorizon (Section B-819, YOLKIN et al. 1986). The Upper Salairka Beds correspond to the *kitabicus* (its upper half) and the *excavatus* (complete) conodont zones, the brachiopod *Protodouvillina praedistans* Zone, and the ostracod *Miraculum biclivosum* Zone. Earlier, the interval of the Upper Salairka Beds included only Bed 15 of Section B-819, and the interval of Beds 16-20 were regarded as a facies analogue of the Belovo Limestones in a terrigenous type of section. Conodont data from the Belovo Horizon in its type outcrops showed the inaccuracy of such correlations (IZOKH et al. 2012).

3. The Belovo Horizon is abolished. In all the type sections of the Belovo Horizon near Gur'evsk town and along the Baskuskan Creek, *Polygnathus serotinus* TELFORD was recorded. This allows to consider the Belovo (rugose) coral limestones as a facies analogue of the Upper Shanda Beds.

4. A debate concerning the understanding and interpretation of the facies variability within the Shanda Horizon remains. In typical sections of the Shanda Horizon (Akarachkino Quarry) conodonts older than the *serotinus* Zone were not found, whereas traditionally the Shanda Horizon was aligned with the *nothoperbonus* - *serotinus* conodont zones. In Section B-819, the base of the *nothoperbonus* Zone was established at the base of Bed 21, which is accompanied by the first appearance of a number of zonal species of the Shanda Horizon, including the brachiopods *Protodouvillina grandicula* (GRATSIANOVA) and *Leptodontella zmeinigorskiana* (PEETZ in BUBLICHENKO), and the ostracod *Miraculum bisulcatum* BAKHAREV. The stratigraphical volume of the Shanda Horizon in section B-819 has previously been considered to comprise the interval of Beds 21-26, whereas by conodonts this interval corresponds to the *nothoperbonus* Zone. The situation is further complicated at higher stratigraphic levels of Section B-819: In Beds 41 and 42, along with typical Eifelian ostracods, conodonts corresponding to the lower part of the *inversus* Zone were recorded (according N.G. IZOKH). These data require further analysis.

5. The boundary between the Shanda and Mamontovo Horizons of the Salair was placed at the base of red terrigenous beds (change to a regressive stage from a transgressive one), overlying different facies of the Upper Shanda Beds. This boundary is observed at the north-eastern side of the Akarachkino Quarry. In a number of the Shanda Horizon sections conodonts of the *serotinus* Zone and the ammonoid *Mimagoniatites bohemicus* (BARRANDE) were found.

6. The Mamontovo Horizon is now considered to comprise two units: the lower Malaya Salairka Beds and the upper Pesteryovo Beds. A previously included upper third unit, the so-called Akarachkino Beds

(BAKHAREV et al. 2004, 2011), are now regarded as a separate stratigraphic unit. The Malaya Salairka Beds are not characterized by conodonts and ammonoids. The first Eifelian conodonts, *Polygnathus costatus partitus* and *Po. costatus costatus*, were established in the lower part of the Pesteryovo Limestone in the northwestern part of Malaya Salairka Quarry. Ammonoids typical for the lower Eifelian *Pinacites jugleri* Zone were also recovered there. Taking into account this biostratigraphic level, the authors place the boundary between Emsian and Eifelian stages in the top of the Malaya Salairka Beds. In the upper part of the Pesteryovo Limestone of Malaya Salairka Quarry, a conodont association of the *australis* Zone was found.

7. As a regional and global biomarker of the Akarachkino Horizon, an isochronous ammonoid assemblage was discovered in a number of sections. *Agoniatites* cf. *vanuxemi* (HALL), *Agon.* cf. *nodiferus* (HALL), *Fidelites* sp., and *Cabrieroceas salairicum* NIKOLAEVA were found in the upper part of the Akarachkino Horizon type section near the town Gur'evsk. A similar ammonoid assemblage was discovered in the Safonovo Formation sections near Prokopyevsk town and near Zarechnaya village (Bed 11 of Section B-8333), on the right bank of the river Bolshoy Bachat, Kuznetsk Basin (BAKHAREV et al. 2011). This ammonoid assemblage allocates their level to within a narrow biostratigraphic interval of the upper Eifelian *Agoniatites costulatus* Zone. In some sections of the Safonovo Formation (near Prokopyevsk town, Kuznetsk Basin), conodonts of the *kockelianus* Zone, proving an upper Eifelian age, were found. Close to locations with these ammonoids, brachiopods corresponding to the *Indospirifer pseudowilliamsi* Zone, including the zonal species were found. Data mentioned above allowed us originally to consider the Safonovo and respectively Kerlegesh (as stratigraphically older) formations as a facial analogue of the Akarachkino, Pesteryovo and partly Malaya Salairka Beds of the Mamontovo Horizon (BAKHAREV et al. 2011). However, authors are still not in complete agreement concerning this interpretation. First of all, such correlation is not consistent with the data on brachiopods, in particular with the brachiopods association of the Kerlegesh Formation (Horizon). Brachiopod associations from the Kerlegesh Formation include a number of Stringocephalida representatives: *Chascothyris salairica* RZHONSNITSKAYA, *Newberria* (= *Denckmannella*) *damesi* (HOLZAPFEL), *N. circularis* (HOLZAPFEL), and *Bornhardtina* sp. (RZHONSNITSKAYA 1968). There are no records of these brachiopods association below the Givetian Stage in the world literature. Only additional data can help to resolve this complicated and contradictory situation.

8. The Izyly Formation age was re-assigned from the Givetian to the early Frasnian. The brachiopod assemblage belongs to the *Plicathyris sibirica* Zone. The ostracods *Kozlovskiella* sp., *Knoxiella*? sp., *Fabalicypriis holushurmensis* (POLINEVA), *Illativella bicornis* (SCHEVTSOV), *Cryptophyllus* sp., *Marginia sculpta multicostata* POLINEVA, *Bairdia carinata* POLINEVA, and *Uchtovia cyrlinae* POLINEVA prove an early Frasnian age as well. At the base of overlying Vassino Horizon, the conodont *Ancyrodella lobata* BRANSON & MEHL is characteristic for the Middle-Upper Frasnian *hassi-linguiformis* zones. It confirms an approximately early Frasnian age for the Izyly Horizon below.

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## **(P) New biostratigraphic insights from the early Mid-Devonian Choteč Event**

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Samples from rock residues of the Choteč Limestone at its type locality (Na Škrábku quarry at Choteč village) provided masses of three-dimensionally preserved palynomorphs which belong to the prasinophycean algae. In addition, a few mazuelloids and scolecodonts have been observed, whereas acritarchs, chitinozoans and spores are missing in particular parts of the section. Those forms are present at only a few levels.

The Basal Choteč Event in its type area, the Prague Basin, is documented by the occurrence of the Choteč Limestone (Choteč Formation, basal Eifelian) and its equivalents, just above the Lower-Middle Devonian boundary.

In the Na Škrábku section, the respective lithology changes from light-grey bioturbated limestone of the Třebotov Limestone unit to an alternation of dark-grey bioclastic limestones with dark-grey laminated marls and slightly bioturbated middle-grey to dark-grey limestones of the Choteč Limestone unit. The prasinophycean algae occur in a dark bioclastic limestone, 20 cm above the base of the Choteč Limestone. Assemblages of prasinophytes (e.g., *Tasmanites/Leiosphaeridia*) occur in high abundance just above the base of Choteč Limestone and most probably represent a phytoplankton bloom during the Basal Choteč Event. Potential causes for the accumulation of an almost monospecific assemblage of the figured organic-walled phytoplankton (OWM) and possible relations/effects of this event are discussed.

At present, the Choteč Event is mostly regarded as minor, although it is globally documented through distinct changes of facies, biota and sea-level fluctuations. In this context, investigations of time-equivalent successions in different areas were recently started. First results of the on-going studies from several sections of the Appalachian Basin will be shown, based on new faunal (e.g., dacroconarids) and palynological data in comparison with those from the type section. In the Appalachian Basin, focus is on the lower members of the Onondaga Formation (Edgecliff and Nedrow members) in order to get more information on the position and duration of the Choteč Event.

**(P) Phacopid trilobites across the Zlíchovian/Dalejan, Dalejan/Eifelian and Eifelian/Givetian boundaries in the Prague Basin (Barrandian area, Czech Republic)**

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The occurrence of phacopid trilobites worldwide is typical by extensive radiation shortly after their origin in the uppermost Ordovician, with acme of their diversity and disparity in the Devonian. These trilobites are important component of Silurian and Devonian trilobite associations, up to their abrupt disappearing at the end of Devonian. The occurrence of phacopids has been, however, affected also by several biotic crises. Such distribution pattern in the Zlíchovian up to the Givetian strata of the Prague Basin (Barrandian area, Czech Republic) is discussed herein. For review of previous investigation, see CHLUPÁČ (1977 and 1983). The Zlíchovian phacopids of the Prague Basin are characterised by abundant occurrence but low diversity of *Reedops*. Only one species – *Reedops decorus* occurs since lower part of Zlíchov Formation up to its upper but not uppermost parts, forming typical *Reedops-Odontochile* Assemblage sensu CHLUPÁČ (1983). *Nephranomma modesta* is abundant in the late Zlíchovian. *Boeckops* is less common, being represented by *B. delphinoides* with similar distribution pattern like *R. decorus*. Rare but significant is *Kainops ? chlupaci*. In the late Zlíchovian, *Pedinopariops degener* is characteristic, together with rare *Chotecops glabrens*. The onset of Daleje Event totally rebuilt the phacopid associations. This event presents decline of *Reedops*, *Nephranomma*, *Boeckops* and *Paciphacops* in the Prague Basin. Only one genus of phacopid lineage, which was almost unaffected by this event, was *Pedinopariops*. *P. degener* was only replaced by closely relative *P. superstes superstes* (BARRANDE) which, most probably, represents its adaptation to the changed life environment. Main morphological changes, that enable distinguishing *P. superstes superstes* from *P. degener*, happened in the late holaspid stage (BUDIL et al. in press). *P. superstes superstes* forms the monotypic phacopid community characteristic for the Daleje Shale facies. In the late Dalejan Třebotov Limestone, diversity of phacopids notably increased. *P. superstes superstes* was replaced by *P. superstes superior* and *P. insequens*. *Struveaspis* appeared for the first time in the Prague Basin (*S. marocanica*). There is also sporadic occurrence of *Eocryphops*. In shallower-water Suchomasty Limestone, rich phacopid association with dominant *Pedinopariops major* and *P. regius* and uncommon *Signatops signatus* occur. The Basal Choteč Event presents the decline of *Pedinopariops* lineage but *P. insequens* crosses the Lower-Middle Devonian boundary and extincts in the lower part of the Choteč Formation. *Signatops* disappears meanwhile *Struveaspis* is represented by *S. micrommaeomicromma* with reduced eyes in lower part of the Choteč Limestone and by *S. fugitiva* in its higher part. Eifelian presents the acme of *Chotecops* lineage in the Barrandian area. *Chotecops auspex* is abundant since the lower up to the uppermost part of the Choteč Limestone. *Eocryphops* cf. *termieri* have been also documented in its lower part. In shallower-water Acanthopyge Limestone, rich but low-diversity phacopid association with dominant *Chotecops hoseri*, possibly also questionable *Cordapeltis breviceps* occur. The onset of the Kačák-otomari Event presents abrupt collapse of phacopid associations in the Barrandian area. Last *Chotecops auspex* was gathered two meters below the base of the Kačák Shale (BUDIL 1995). At the base of the lower Givetian Roblín Member, a local and stratigraphically limited occurrence of *Chotecops* aff. *hoseri* have been documented (CHLUPÁČ, 1977). This is the youngest occurrence of phacopids in the Prague Basin.

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## (P) Charcoal, Forests, and Earth's Palaeozoic geochemical oxygen cycle

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Reconstructing secular trends in atmospheric  $\rho\text{O}_2$  is of critical importance for studies of the origin and diversification of tetrapods (WARD et al. 2006, CLACK 2007), arthropod gigantism (GRAHAM et al. 1995) the evolution of flight (DUDLEY 1998), and the role of deep-rooting forested ecosystems in the biotic regulation of the atmosphere (BERNER et al. 2003). A time-series of microscopic charcoal abundance is presented for the latest Early to Late Devonian of Gondwana (Sub-Andean Zone of Bolivia); this is accompanied by a quantitative analysis of reflectance distribution, and the link with atmospheric  $\rho\text{O}_2$  is discussed.

Given that direct sampling is not an option for much of the Phanerozoic, most previous  $\rho\text{O}_2$  curves (e.g., BERNER & CANFIELD 1989, BERNER 2001, BERNER 2006) have been modelled indirectly, based on the geochemical carbon and sulphur cycles, which are the principal controls on atmospheric oxygen (EBELMEN 1845; GARRELS & PERRY 1974). It is becoming clear that these are strongly modified by nutrient cycles (chiefly phosphorus and nitrogen) and external forcing (e.g. volcanic degassing), and regulated by complex feedback mechanisms (LENTON 2001); unfortunately, incorporating these factors into existing models in a realistic way has proven difficult, because many of the underlying assumptions cannot be easily tested. This has led to a lack of agreement between models, and until these discrepancies have been resolved, hypotheses linking biological/evolutionary events to changes in  $\rho\text{O}_2$  are impossible to verify.

The inertinite-group macerals are largely synonymous with fossil charcoal (fusain; SCOTT & GLASSPOOL 2007), and are a common microscopic component both of coals and as dispersed organic matter in other sedimentary rocks. Charcoal is the product of natural wildfires, and there is a well-established positive correlation between atmospheric  $\rho\text{O}_2$  and fire frequency/intensity (BELCHER et al. 2010). By creating and comparing high-resolution inertinite abundance/reflectance time-series for the Silurian – Carboniferous of Euramerica and Gondwana, we intend to explore the hypothesis that the spread of forests significantly increased the production of atmospheric oxygen by increasing organic carbon burial and fundamentally altered biotic regulation of Earth's  $\text{O}_2$  concentration. A particular focus will be the postulated mid-Devonian oxygen low, which is supported not just by models, but also the apparent absence of macroscopic charcoal during this period. Comparing the two datasets will allow us to test this, and also explore possible biases introduced by regional variation of vegetation type, environment, and climate. We present here preliminary results for the latest Emsian – mid-Frasnian of Bolivia, drawn from sampling of the Icla, Huamampampa, Los Monos, and Iquiri Formations.

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## Evolutionary changes in gymnosperm-type trees at the Devonian-Carboniferous boundary: significance of the Gondwanan record of plants

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The Devonian-Carboniferous is a key period in the evolution of land vegetation. It corresponds to the last stage of the terrestrialization process, when plants gradually came to occupy all the surface of the continents and formed the first "modern" ecosystems. A particularly significant part of this process is the appearance and evolution of the first trees and of the first forests. The fossil record indicates that, from the Middle Devonian, the tree habit evolved several times in parallel within groups of plants with different taxonomic affinities and anatomical organizations (e.g., MOSBRUGGER 1990, MEYER-BERTHAUD et al. 2010). Among them, the diversification of lignophyte (progymnosperms and gymnosperms) trees, with their significant biomass, true leaves and deep root systems, is of special importance given its hypothesized impacts on Devonian and Carboniferous ecosystems (e.g., ALGEO & SCHECKLER 1998).

The first lignophytes to evolve the tree habit belonged to the progymnosperm genus *Archaeopteris*. These trees colonized lowlands worldwide during the Middle to Late Devonian and formed the first dense shady forests on Earth (BECK & WIGHT 1988). In Gondwana, *Archaeopteris* is represented by compression of leafy branches and petrified wood from Africa (ANDERSON et al. 1995, MEYER-BERTHAUD et al. 2000, 2004), South America (BERRY et al. 2000), and Australia. *Archaeopteris* became extinct around the Devonian–Carboniferous boundary, possibly in connection with a global crisis in terrestrial plants (STREEL et al. 2000). Following this extinction, very little evidence of lignophyte trees is known before the Late Tournaisian and most data comes from Euramerican localities (GALTIER & MEYER-BERTHAUD 2006, DECOMBEIX et al. 2011a). In this context, our research aims to investigate the patterns of transition between the Devonian and Mississippian forests at a global scale by analyzing the distribution, taxonomic diversity and biology of lignophyte trees during this period.

The first aspect of this work is the search for evidence of lignophyte trees (1) earlier in the Mississippian, i.e., closer to the D/C boundary and the extinction of *Archaeopteris*, and (2) at a wider geographic scale, with a focus on Gondwanan localities.

In Australia, investigations of Middle Tournaisian localities of Queensland in 2005 and 2008 have yielded at least three types of arborescent lignophytes (DECOMBEIX et al. 2011b). One of them, the progymnosperm *Protopitys*, is also known in Europe, which indicate a wide geographical distribution of some of these Mississippian trees. Thanks to these Queensland specimens and to an additional record of *Archaeopteris* wood from the Late Famennian of New South Wales (MEYER-BERTHAUD et al. 2008), a transition comparable to that observed in Europe and North America can now be clearly established for the Australian region.

In North Africa, Moroccan localities of the Anti-Atlas have yielded abundant remains of *Archaeopteris*, which provided key information on the anatomy, biology and architecture of the plant, including its root system (MEYER-BERTHAUD et al. 2000, 2004, in press). For the Mississippian however, only two specimens of lignophyte trees have been reported. The first one, from the latest Namurian-to Westphalian A volcanic sequence of the Tazekka region in Morocco (CHALOT-PRAT & GALTIER 1989), has a wood comparable to some European taxa. The second specimen is a young stem from the Late Tournaisian of the Ahnet region in Algeria (GALTIER & MEYER-BERTHAUD 2006) that represents to date the best evidence of lignophyte trees in North Africa after the extinction of *Archaeopteris*. The detailed anatomical study of this specimen indicates that it represents a new taxon, and it is likely that further investigation of North African deposits would provide novel information on Mississippian lignophyte trees.

To complement these new findings, two types of analyses have been conducted using the data gathered on the anatomy and morphology of lignophyte trees: (1) statistical analyses, to examine the evolution of the vegetative body of these plants between the Devonian and the Mississippian, and (2) phylogenetic analyses, to better assess the progymnosperm (free-sporing plants) or gymnosperm (seed plants) affinities of the Mississippian taxa, for which reproductive structures are mostly unknown. Our results show that the Mississippian taxa had a morphospace overlapping with that occupied by *Archaeopteris* in the Devonian and that most Mississippian taxa appear related to the gymnosperms. Therefore, the hypothesis of a fast replacement at a large geographical scale of *Archaeopteris* by trees that were comparable in their vegetative construction but less constrained in terms of reproduction must be considered.

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**(P) Strunian rugose and tabulate corals from Northwestern Turkey****DENAYER, J.**

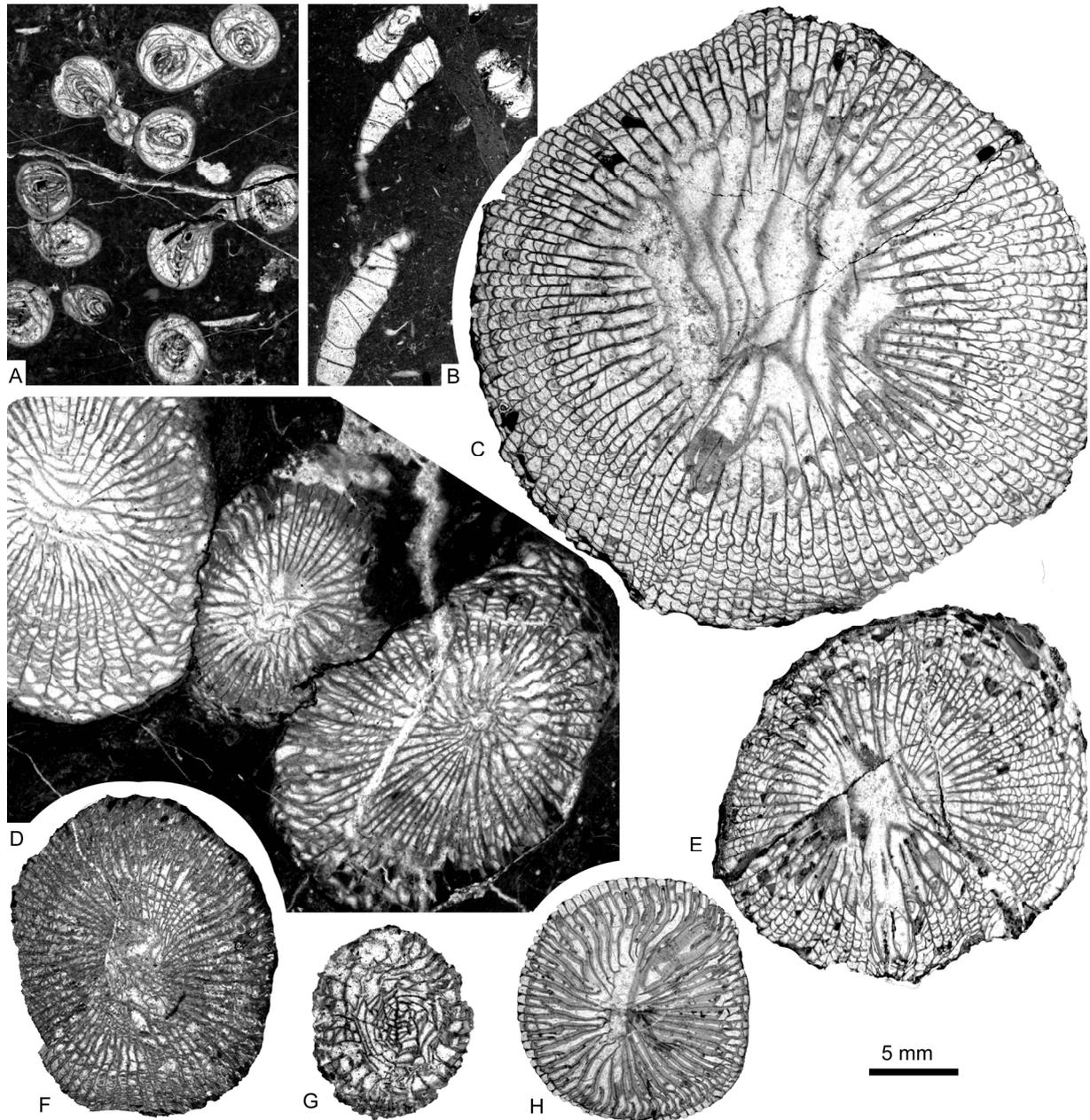
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Strunian (uppermost Famennian) rugose corals are worldwide known but usually form endemic assemblages (POTY 1986). For example, Western European assemblages contains only solitary rugose corals, Siberian assemblages contains rare several colonial corals and coral fauna from South China contain a lot of colonial and solitary genera (CONIL et al. 1982, POTY & XU 1996). In the Istanbul Zone - Zonguldak and Bartın areas, in Northwestern Turkey - the Strunian coral assemblage was discovered in the middle part of the Yılanlı Formation, a thick unit of variegated limestones and dolostones from Middle Devonian to upper Viséan in age. The fossiliferous Strunian corresponds to a c.a. 30 m-thick unit of bedded limestones resting upon a thick unit of massive reddish sandy limestone and dolostone. The facies is mainly bioclastic to peloidal wackestone to packstone grading to grainstone, often slightly dolomitized, typically formed in shallow water platform environment. In the Bartın area, two levels, 40 to 50-cm thick of stromatoporoid boundstone were observed in several localities, forming a marker level through the region. In the Zonguldak area, 60 km westward, these two levels were not recognized but questionably pass to a thick level of bioclastic rudstones with large fragments of stromatoporoids, *Pseudochaetetes* and tabulate corals. The corals were mainly collected in the bioclastic limestone above the stromatoporoid biostrome of the Bartın area. They forms a bottom-level association, mostly dominated by large campophyllids. The rugose corals are dominated by solitary taxa: *Campophyllum* sp. 1 (Fig. 1C, H), C. sp. 2 (Fig. 1E), "*Palaeosmilia*" cf. *aquisgranensis* (Fig. 1F), *Clisiophyllum* aff. *omaliusi* (Fig. 1G), and indeterminate solitary undissepimented rugose corals. Only one colonial coral was collected: cf. *Endophyllum* sp. with an unusual dendroid habitus (Fig. 1D). The Tabulate corals are mainly large syringoporids (Fig. 1A) and an indeterminate Auloporida (fig. 1B). This faunal assemblage has a strong Eastern Europe character, the campophyllids, clisiophyllids and endophyllids being the main component Polish Strunian strata (BERKOWSKI 2002). Chinese taxa - e.g. *Cystophrentis*, *Pseudostelechophyllum*, *Heterostrotion* - or Siberian Strunian taxa such as *Melanophyllidium*, completely lack in Turkey. The Istanbul Zone is supposed to be situated in central Palaeotethys Ocean, along the southern margin of Laurussia during the uppermost Devonian and Mississippian. The rugose corals plaid to some connection with Eastern Europe at this time or mean that both area were under the influence of a common marine current. The Hangenberg event was not recognized in the Turkish localities, except if considering the disappearance of the corals, occurring less than 3 m below the Devonian-Carboniferous boundary based on the foraminifers biostratigraphy (disappearance of *Quasiendothyra*) in the Topluca section, Bartın area. There is no facies change through the boundary but the first carboniferous coral (small *Uralinia*) recover more or less 10 m above the D-C boundary.

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**Fig. 1:** Strunian rugose and tabulate corals from Northwestern Turkey. **A:** *Syringopora* sp. (specimen ET.11.7.a, Topluca section). **B:** indeterminate Auloporida (specimen ET.11.15, Topluca section). **C:** large specimen of *Campophyllum* sp. 1 (specimen ET.11.12.d, Topluca section). **D:** dendroid cf. *Endophyllum* sp. (specimen D.2.1.II.a, Dallica section). **E:** *Campophyllum* sp. 2 with transeptal dissepiment (specimen ET.11.13.II.a, Topluca section). **F:** *Palaeosmia* cf. *aquisgranensis* (specimen G.3.19, Gökgöl section). **G:** *Clisiophyllum* aff. *omalusi* (specimen ET.12a.1.III, Topluca section). **H:** *Campophyllum* sp. 1 (specimen ET.11.12.III.b, Topluca section). Transverse section for all, except B. Scale for all: x3.



## (P) Microfacies and Devonian reef development in the Oued Cherrat Zone (Aïn Khira South and Aïn-as-Seffah), Moroccan Meseta

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The Middle and Upper Devonian (Givetian-Frasnian) is distinguished globally by a maximum development of large reef complexes. They show a highly variable palaeoecology, with different main reef builders (stromatoporoids, tabulate and rugose corals, calcareous algae, Bryozoa) in different regions. They can be developed as large bioherms, consisting of an initial bank facies growing into differentiated reef, slope and platform realms, as smaller-scale atolls sitting on volcanic buildups, as low-relief biostromes, or as mudmounds. Based on their rather southern palaeolatitudinal position on the northwestern shelf of Gondwana, outside the central tropical belt, the poorly investigated reefal limestones of the Moroccan Meseta are of particular importance for Palaeozoic reef studies in a palaeoclimatic context. A facies comparison of reefs of the southern (Hercynian Morocco) and northern (Rhenohercynian Zone) external Variscides, across the at least 3000 km wide western Prototethys, is conducted in the frame of a current DFG project.

In the western Moroccan Meseta, large Givetian carbonate platforms extended from the Atlantic Ocean (Coastal Block) to the Oulmes area (EL HASSANI & BENFRIKA, 2000). A first focus lay on the reefal complex of Ain Jemaa, at the eastern end of this belt (EICHHOLT et al. 2012). Here we present new detailed data on reefal limestones of the Oued Cherrat Zone (Fig. 1 & 2), which extends south of Rabat in the western part of the Hercynian Moroccan Massive (CHALOUAN 1981). Supposed Givetian reefs at Aïn Khira and Aïn Dakhla were previously described by ZAHRAOUI (1991), CATTANEO et al. (1993), and ZAHRAOUI et al. (2000). Intensive quarrying has destroyed many previous outcrops and the Aïn Dakhla Reef proved to be Emsian in age (ABOUSSALAM et al., 2012). Therefore, our work concentrated on a section bordering the new large quarries at Aïn Khira South (Fig. 3) and on a small, isolated reefal outcrop in the southern Oued Cherrat Zone, at Aïn-as-Seffah (Fig. 4).

The Aïn Khira South section is in the northern part of the Oued Cherrat and represented by almost 100 m of limestone beds. Since the Cakhrat-ach-Chleh Formation now falls in the Emsian, a new name, Aïn Khira Formation, is proposed for the N-S trending stretch of Givetian reef limestone east of the Oued Cherrat valley. To give an exact age is difficult, because there are no conodonts at the base or top of the succession. In the middle part (59 m above base) there are limestone beds with *Stringocephalus* (Fig. 5), the brachiopod index fossil for the Givetian. The whole outcrop shows a back reef facies setting with characteristic lagoonal facies types. Fore reef and definite reef core facies types are missing. At the base the limestone consist of a peloidal grainstone with fragments of brachiopods intercalated with a detrital Stromatoporoid-Floatstone with a peloidal grainstone matrix. Such facies types occur near the reef center in shallow, strong agitated water. Associated are *Stachyodes-Thamnopora*-Floatstones and bulbous Stromatoporoid-*Stachyodes*-Boundstones. The latter indicate the proximity of a patch reef or of the reef core. Bioclastic mud-/wackestones with a very poor fossil record (ostracods, gastropods, shell fragments) indicate more restricted and calm realms within the lagoon. Between the bioclastic mud-/wackestones limestone beds with *Stringocephalus* appear; these are known in the Rhenish Massive to occur within the protected wide, shallow platform but were often fragmented during storm events. At the top of the section *Amphipora*-Bafflestones give further evidence for a persisting lagoonal setting, with some shallowing upwards.

The isolated Givetian "reef" of Aïn-as-Seffah is constituted by more than 40 m of massive limestone, here assigned to an extended Aïn-as-Seffah Formation, with the reef limestones forming Member 2. The

"reef" base falls in the lower Middle Givetian and is intercalated with contemporaneous, thin-bedded flaserlimestones and marls (Member 1). The top is covered by Lower Frasnian brachiopod limestones (Member 3), followed by Upper Frasnian flaserlimestone (Member 4) and a distinctive uppermost Frasnian *Manticoceras* Bed (Member 5). The whole unit seems to represent a big, partially slumped mass. Stylolites and solution seams indicate pressure solution caused by tectonic stress during the diagenesis. However, there is no evidence for a large scale internal deformation/distortion of the "reef" blocks; therefore, the internal fabric is still autochthonous. In comparison with other reefal limestones of the Moroccan Meseta, the Aïn-as-Seffah "reef" is only a small bioherm/patch reef that developed on a distal, deeper neritic carbonate platform. There is no distinct zonation into fore reef, reef core and back reef. Detailed studies of micro- and macrofacies enable to establish ten different microfacies types/subtypes. At the base, the limestone consists of a Peloidal-Crinoidal-Grainstone with reef derived detritus (only a few small fragments of stromatoporoids and tabulate corals, e.g. *Alveolites*). Above, coral fragments become larger and more diverse (e.g. *Thamnopora*, *Pachyfavosites*, solitary and colonial rugose corals). Typical facies are *Alveolites*-Stromatoporoid- or *Thamnopora*-*Pachyfavosites*-Rudstones. The overlaying limestones consist of different kinds of coral-stromatoporoid-boundstones (e. g. *Alveolites*-*Thamnopora*-Bind-/Framestones, *Alveolites*-Stromatoporoid-Boundstones, *Thamno-pora*-Boundstones), which constructed a small buildup. In spaces between the corals, the limestone mainly consists of Crinoidal-Grainstones. The consistent presence of neritic conodont faunas proves permanent open marine conditions. The partially (laterally) overlying Brachiopod-Floatstones suggest an extinction of the reef by drowning. The morphological reef top ends with a drowning unconformity, probably caused by current activity.

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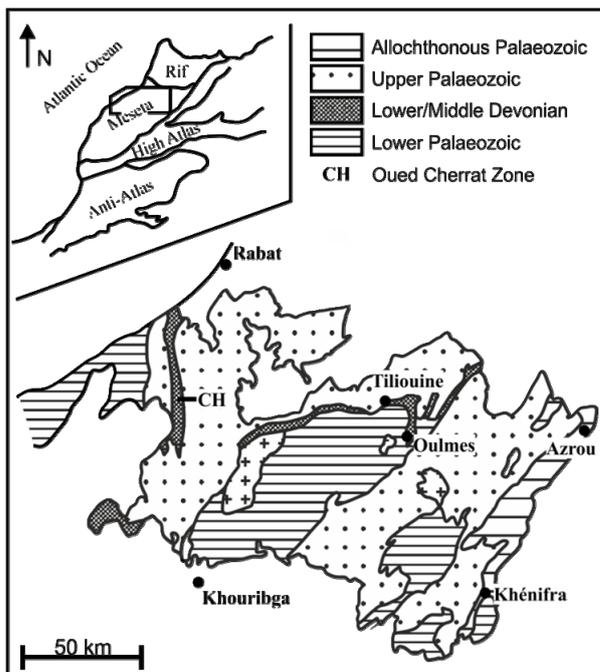


Fig. 1: Geological map of the northwestern Moroccan Meseta (modified after Cattaneo et al. 1993).

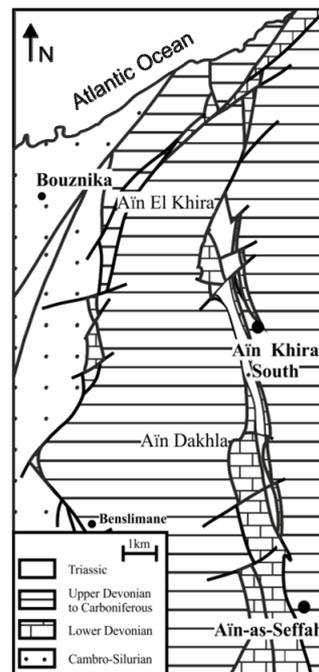


Fig. 2: Location map of the Oued Cherrat sections (modified after Zahraoui, 1991).



Fig. 3: Givetian reef limestones at Ain Khira South.



Fig. 4: bioclastic mud/wackestone with Stringocephalus (black arrow) at Ain Khira South, 59 m. fig. width 17 mm.



Fig. 4: The isolated Givetian "reef" block at Ain-as-Seffah, view from SW. The marker tree indicates the section top.

## (P) Lower Devonian bryozoan faunas of Spain

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Bryozoa are abundant and diverse in the Devonian worldwide (CUFFEY & MCKINNEY 1979). This period was a time of important changes in the structure and global composition of bryozoan faunas (BIGEY 1985, HOROWITZ et al. 1996). However, despite their abundance and importance, Devonian bryozoan faunas in Europe have been scarcely investigated.

Lower Devonian deposits of Spain contain rich bryozoan assemblages, which are still insufficiently studied. The focus of the existing research lies on bryozoan faunas from the Lower – Middle Devonian deposits of the Cantabrian Zone, NW Spain. Some few publications deal with the bryozoan faunas from the Moniello Formation (late Emsian – early Eifelian) of Cantabrian Zone (NW Spain): SUAREZ ANDRÉS (1998, 1999a-c), SUÁREZ ANDRÉS & GONZÁLEZ (2000a, b), SUÁREZ ANDRÉS & MCKINNEY 2011.

Recent research expanded the existing knowledge about the taxonomic composition of Lower Devonian bryozoans from Spain. The bryozoan assemblage described from the "La Vid Group" (corresponding partly Esla and Santa Lucia formations, Emsian-Eifelian) in Cantabrian Mountains, NW Spain, comprises 48 species: 11 trepostomes (ERNST 2010), 12 cryptostomes (ERNST 2011), 10 cystoporates (ERNST & BUTTLER 2012), and 15 fenestrates (ERNST 2012). In addition, 14 species were described from the Lower to Middle Devonian (Emsian – Eifelian) Santa Lucía Formation of Abelgas and Paradilla in NW Spain (ERNST et al. 2011a).

A bryozoan fauna containing 16 species was described from a set of small mud mounds occurring in the Lebanza Formation (Lower Devonian, Middle Lochkovian-Lower Pragian) of the Arauz area, NW Spain (ERNST et al. 2012).

Eleven bryozoan species were described from the Lower Devonian (Middle Lochkovian) of Arroyo del Agua, Sierra de Guadarrama, Guadalajara, central Spain (ERNST & MAY 2012).

Still unpublished material includes two rich bryozoan assemblages collected from two localities in the Ossa Morena Zone, Cordoba Province, SE Spain: Peñón Cortado and Guadamez 2. These localities expose most complete reefal sequences of the region, which are estimated being Prahian to Emsian in age (RODRÍGUEZ et al. 2010). These assemblages contain diverse trepostome, cystoporate, cryptostome and fenestrate species. Preliminary results show similarities of these bryozoans to the fauna from the Lebanza Formation (ERNST et al. 2012), and contemporary formations in Bretagne, SW France (unpublished material).

The studied bryozoan faunas refer mainly to the reefal limestones accompanied often by corals, stromatoporoids, brachiopods and echinoderms. Identified bryozoan species display palaeobiogeographic relations to the Prague basin, Northern America and Kazakhstan.

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## The Devonian/Carboniferous boundary and the holotype of *Siphonodella sulcata* (HUDDLE, 1934) in the upper New Albany Shale, Illinois Basin, southern Indiana

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The upper New Albany Shale in southern Indiana consists of the Jacobs Chapel, Henryville, Underwood, and Falling Run beds (descending order) within the upper Clegg Creek Member and the Ellsworth Member (Fig. 1). In Clark and Floyd counties near Henryville and New Albany, Indiana *Bispathodus aculeatus aculeatus* (BRANSON & MEHL, 1934), *Branmehla bohlenana* (HELMS, 1959), *Br. inornata* (BRANSON & MEHL, 1934), *Cryptotaxis culminidirecta* (KLAPPER & PHILIP, 1971), *Palmatolepis glabra* ssp., *Pa. gracilis gracilis* BRANSON & MEHL, 1934, and *Protognathodus* sp. were recovered from the upper Clegg Creek below the Falling Run Bed and indicate the Devonian praesulcata Zone. The holotype of *Siphonodella sulcata* (HUDDLE, 1934) is from Locality 9 of HUDDLE (1934) near the city of Rockford, Jackson County, Indiana. The strata were described as "near the top of the New Albany shale just above gray-green shale containing Devonian Brachiopods ..." This site along US Alt. 31 is now covered. The holotype has been lost, which was first reported by J. HUDDLE in the 1960s (G. KLAPPER, personal communication) and confirmed by C. JOHNSON (personal communication), who indicated that the labeled specimen card is in the Paleontological Museum at Indiana University, but that there is no specimen.

The greenish-gray brachiopod-bearing shale noted by HUDDLE (1934) is the Ellsworth Member, which in Indiana Geological Survey Drill Hole (SDH) 324, eight km to the northwest in Bartholomew County, contains diminutive specimens of *Bispathodus*, *Branmehla*, *Polygnathus communis communis* BRANSON & MEHL, 1934, and *Si. sulcata* indicative of the lowest Carboniferous *sulcata* Zone. The *sulcata* Zone conodonts occur in association with the brachiopods *Subglobosochonetes seymorensis* (HUDDLE, 1933), *Rhipidomella newalbaniensis* HUDDLE, 1933 (emended), *Schuchertella* sp., *Rhynchopora prisca* CARTER, 1988, and *Sphenospira* sp. cf. *S. alta* (HALL, 1867). In the SDH 324 core, brachiopod assemblages in the lower half of the Ellsworth are comprised almost entirely of *S. seymorensis*, with moderately diverse assemblages with *S. seymorensis*, *R. prisca*, and *R. newalbaniensis* in the upper half below the Henryville Bed. None of the brachiopods in the Ellsworth fauna of southern Indiana are known to carryover from older latest Famennian praesulcata Zone strata in the region - the Louisiana Limestone fauna of Illinois and Missouri (WILLIAMS 1943, CARTER 1988). The Ellsworth brachiopod fauna is similar to the earliest Tournaisian brachiopod fauna described from the Glen Park Formation of the western Illinois Basin in Illinois and eastern Missouri by CARTER (1988), which also yields conodonts of the *sulcata* Zone (SCOTT & COLLINSON 1961, COLLINSON in KLAPPER et al. 1971, SANDBERG et al. 1972). A remarkable concentration of conodonts associated with lingulid brachiopods - *Langella?* sp. - was also reported from the black shale of the Henryville Bed in the SDH 324 core. The Henryville contains numerous conodonts concentrated in laminae within the black shale, including *Bispathodus*, *Po. c. communis*, *Pseudopolygnathus* sp., *Si. sulcata*, *Si. duplicata* (BRANSON & MEHL, 1934), *Si. cooperi* HASS, 1959, and *Si. sandbergi* KLAPPER, 1966 indicative of the *sandbergi* Zone.

Based on examination of HUDDLE'S collections from Indiana University and the U.S. Geological Survey the holotype *Siphonodella sulcata* likely came from a conodont-rich horizon in the Henryville Bed. The Devonian-Carboniferous Boundary is placed at the base of the Ellsworth Member, and the base of the

Falling Run Bed in the Clegg Creek Member; a significant disconformity is present between the greenish-gray shales of the Ellsworth and Henryville Bed, and where the Ellsworth is absent, at the top of the upper Clegg Creek black shales at the base of the Falling Run Bed or base of the Henryville Bed.

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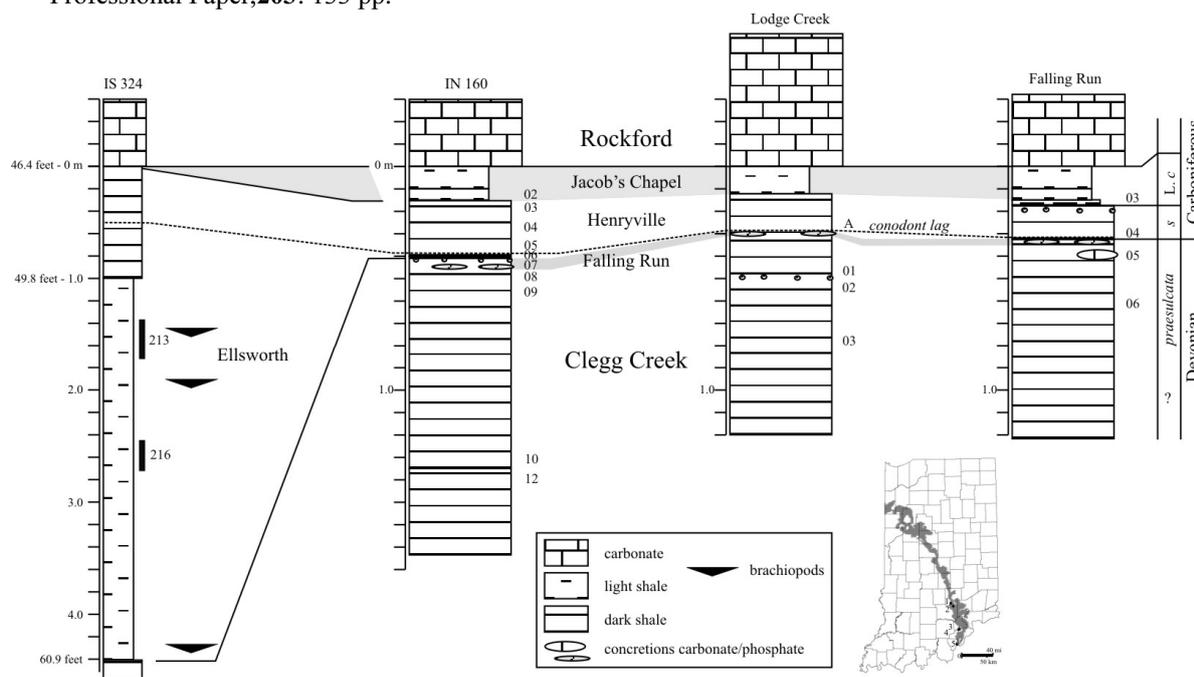
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**Fig. 1.** Stratigraphic nomenclature, measured intervals, and sample distribution from the upper Clegg Creek Member of the New Albany shale in southern Indiana. IS 324 is Indiana Survey Drill Hole 324 – Bartholomew County, IN (1); IN 160 and Lodge Creek – Clark County, IN (3,4); Falling Run – Floyd County, IN (5). Map shows surface distribution of New Albany Shale in Indiana and locality of sample sites; type locality of *Siphonodella sulcata* is (2).

## New data on the Daleje Event from the Barrandian (Bohemia) as a key for progress in Emsian stratigraphy

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The long-term activities of the International Subcommittee on Devonian Stratigraphy have been focused on the division of the Emsian into two new substages. The different stratigraphic levels, close to the Daleje Event, have mostly been discussed as their boundary. The latter event was first recognized in the Barrandian and defined as a bioevent, connected with a pronounced transgression. Its present conception has been inferred from studies of its appearance in deeper environments, with deposition of the calcareous Daleje Shale. Despite its broad impact on environment and biota of Emsian stage no precise biostratigraphic research has been made in carbonate dominated environments of the Barrandian.

Our aim was to make precise qualitative and quantitative biostratigraphical, paleoecological and sedimentological research across the initial phase of Daleje Event in shallower environments with predominantly carbonate sedimentation. Before the present study, there was no section in any carbonate-dominated environment of the Barrandian at the Daleje Event interval that had been studied with a sufficient biostratigraphic resolution. To fill this gap in our knowledge, two sections in the NW part of the Barrandian were studied - the Čeřinka hillside and Čeřinka quarry sections.

Our results clearly provide evidence for the co-occurrence of *N. elegans* and *N. barrandei*; subsequently making the definition of a new subzone possible. This new subzone was also recognized at South Tien Shan and in Spain. Evaluation of the stratigraphic distribution of more than 1250 newly collected dacryoconarid shells from studied sections has resulted in the proposal of a new tentaculite biozonation, providing a much higher stratigraphic resolution than the existing biozonations, which are based on goniatites and conodonts. It is highly unlikely to considerably increase the stratigraphy resolution of goniatite biostratigraphy in the future, simply due to the fact that the goniatites generally occur rather rarely, and it will be impossible to find them in all of the studied beds. However, it has to be noted that the full potential of conodont biostratigraphy has not hitherto been evaluated.

Application of the new high-resolution tentaculite biozonation to the newly gathered sedimentological data has resulted in a more detailed model of the transgression-regression history and faunal changes across the Daleje Event interval in the Barrandian. The research reveals a distinct regression trend in the *Nowakia barrandei* biozone which culminated at the base of the *Nowakia elegans* Biozone (within the *barrandei* - *elegans* Subzone). The beginning of the Daleje Transgression, in the type area for the Daleje Event, occurs just above the lower boundary of the *Nowakia elegans* Biozone, and thus is much older than suggested in previous studies, i.e., at the base of the *Nowakia cancellata* Biozone.

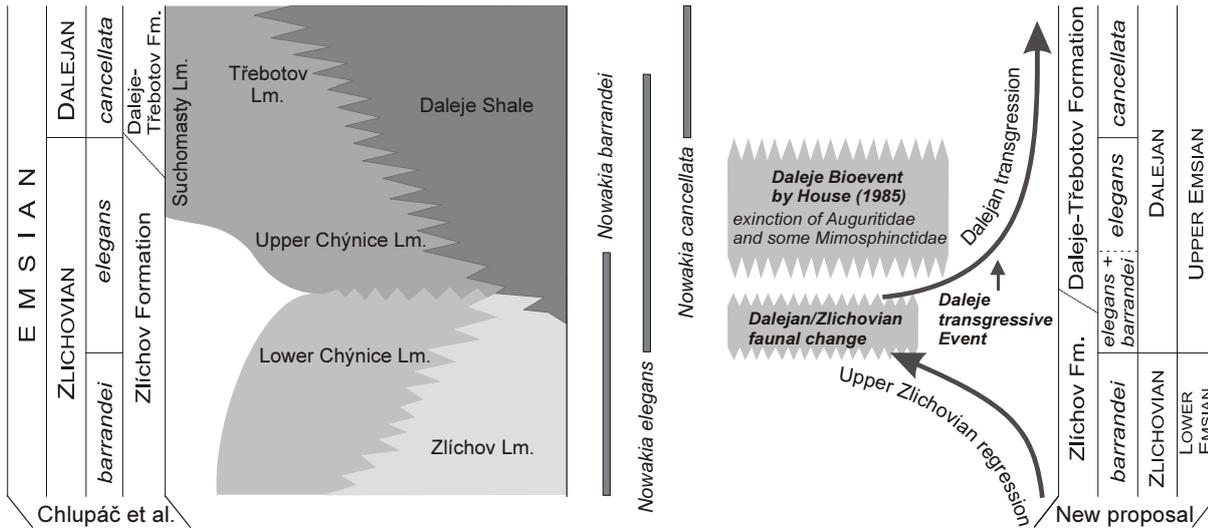
Numerical analysis of more than 2100 specimens, belonging to more than 80 species and coming from the Daleje Event interval revealed a distinct faunal change close to boundary of the *barrandei* and *elegans* Biozones. These results are robust and statistically significant, and they support the earlier observations from deeper water environments.

Taken together, the quantitative analysis of the new paleoecological data suggests a distinct faunal turnover at the base of the *Nowakia elegans* Biozone (within the *barrandei* - *elegans* Subzone). In addition, our study revealed that the Daleje transgression (and thus the Daleje Event) started at the same time and that it was preceded by a distinct regression. These results leads us to redefine the lower Dalejan boundary (FAD of the *N. elegans*) as well as to propose the base of the *Nowakia elegans* Biozone as a

possible level for the definition of the Emsian substage (stage) boundary (i.e., the lower/upper Emsian boundary).

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FERROVÁ, L., FRÝDA, J. & LUKEŠ, P., 2012. High-resolution tentaculite biostratigraphy and facies development across the Early Devonian Daleje Event in carbonate dominated environments of Barrandian (Bohemia). - Bulletin of Geosciences, **87** (3): 587-624. [[http://www.geology.cz/bulletin/fulltext/1336\\_F-F.pdf](http://www.geology.cz/bulletin/fulltext/1336_F-F.pdf)]



Schematic model showing the biostratigraphic position of the lithological units of the Daleje Event interval in the Barrandian as well as a new proposal for the biostratigraphic and lithostratigraphic divisions (ex FERROVÁ et al. 2012).

## (P) Quo vadis, *Platyceras*? – New protoconch data reveals a diphyletic origin of platyceratid gastropods

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Platyceratid gastropods form one of the most common and easily recognizable groups of Palaeozoic gastropods. The oldest members of this group have been recognized in Middle Ordovician strata; the age of the youngest platyceratids is still a matter of debate (BANDEL 1992, 2007, BANDEL & FRÝDA 1999). Typical features of platyceratid gastropods are the high morphological variability of their shells and the close ecological connection at least some of them to Palaeozoic echinoderms (see references in FRÝDA et al. 2008).

A systematic study of Silurian and Devonian platyceratid gastropods provides new data on their early shell ontogeny. Typical cyrtoneuritimorph protoconchs were found in *Praenatica cheloti* (OEHLERT & OEHLERT, 1887) from the Lower Devonian strata of Brittany (W France). The same protoconch type, as in *Praenatica* BARRANDE in PERNER, 1903, was earlier documented in *Orthonychia* HALL, 1843. Both of the latter taxa have often been considered to be subgenera of *Platyceras* CONRAD, 1840. However, some Early Devonian species of *Platyceras* from the Barrandian area develop quite different protoconchs (orthostrophic and tightly coiled), similar to those found earlier in different species of the family Naticopsidae WAAGEN, 1880. Current data clearly evidences that the Palaeozoic platyceratids represent a diphyletic group.

The Palaeozoic platyceratids have been considered to be the stem group for modern Neritimorpha (BANDEL 1992, 2007). On the other hand PONDER & LINDBERG (1997) published an opinion that the Palaeozoic platyceratids are the stem group for the Patellogastropoda. The latter group has no sure Palaeozoic fossil record. This fact is surprising because if they are really the first gastropod offshoot (PONDER & LINDBERG 1997), they had to have split off from the rest of the gastropods (i.e. orthogastropods) before the Early Ordovician (FRÝDA 2012). This controversy has been explained by either their adaptation to intertidal environments that provide a relatively poor fossil record or by the opinion that the Palaeozoic patellogastropods had coiled shells not hitherto recognized among Palaeozoic gastropods. The Palaeozoic platyceratids have been the most frequently discussed as the stem group for the Patellogastropoda. However, the nature of protoconchs (development of a true larval shell) in the both platyceratid lineages testifies against this hypothesis. On the other hand, a derivation of modern neritimorphs from the Palaeozoic platyceratids with tightly coiled protoconchs or from the naticopsids seems to be probable. The strongly convolute neritimorph protoconch (apomorphy of neritimorph crown-group) probably originated after the Permian/Triassic mass extinction event, but before the Late Triassic.

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**(P) El Gara South – new data on Famennian ammonoid and conodont faunas and the *Annulata* Events in the Rheris Basin (northern Tafilalt, Morocco)**

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The Famennian *Annulata* Events are characterized by global overturns of pelagic faunas, especially in ammonoids, and by the wide and sudden distribution of black shale or limestone intercalations. The eustatic rise of the hypoxic Lower *Annulata* Event allowed the spread of *Platyclymenia* faunas, which define the base of UD IV-A and the *Platyclymenia* Stufe of the ammonoid scale. In the international conodont scale, both *Annulata* Events belong to the upper part of the *Ps. granulosus* Zone (= upper part of the Upper *trachytera* Zone, HARTENFELS et al. 2009, HARTENFELS 2011). The hypoxic Upper *Annulata* Event is overlain by regressive and often very fossiliferous cephalopod limestones, the "Wagnerbank" of the Bohlen (Thuringia, e.g. PFEIFFER 1954), and its micritic or marly equivalents.

The Tafilalt in the southeastern Anti-Atlas comprises a shallow pelagic Tafilalt Platform and a deeper pelagic Tafilalt Basin to the east/southeast (WENDT et al. 1984). BECKER (1993) recognized an additional Rheris Basin in the north, which is mostly covered under the Quarternary. Whilst the facies developments and biostratigraphy around the *Annulata* Events of the Tafilalt Platform and Basin have been investigated in detail (e.g., HARTENFELS 2011), there are so far no data for the Rheris Basin. Therefore, the succession at El Gara South (text-fig. 1) was measured bed-by-bed and sampled in two lateral trenches during field work in 2009. Based on a flat landscape, the outcrop conditions are generally poor. The section is situated approximately 12 km northwest of Erfoud, close to the small settlement El Gara (GPS-coordinates: 31°28'23.4"N, 004°21'22.1"N, map sheet 244 Tafilalt-Taouz). In this region, the Famennian was first recognized by WENDT et al. (1984). BECKER (1993) gave a rough cross-section for the exposed Famennian but concentrated on the Nehdenian (UD II). El Gara South is exceptional since it includes taxa that have not been observed anywhere else in the Anti-Atlas.

The two *Annulata* Events are recognizable above a 5.9 m thick succession of middle Famennian marly shales, some yielding small-sized carbonate nodules, with few intercalated red or grey nodular limestones. The UD II-G age of Bed -7 is based on *Maeneceras meridionale meridionale*, whilst *Planitornoceras euryomphalum* marks the UD III-B in Bed -4. Small-sized ammonoids from the overlying Bed -3 are unique for all of the Tafilalt or even for all of North Africa. A poorly preserved *Afrolobites* sp. in association with dominant *Plani. euryomphalum* extends the range of the genus from the Maider to the Tafilalt. Gen. aff. *Protornoceras* n. sp. and a questionable *Varioclymenia* represent new regional records of genera and confirm an assignment to UD III-C<sub>1</sub>. Subsequently, Bed -3b contains many specimens of the marker clymeniid *Sulcoclymenia sulcata* of UD III-C<sub>2</sub>. Associated small-sized goniatites include *Sporadoceras angustisellatum*, *Enkebergoceras varicatum*, and (first record of the genus/species from Morocco) *Protornoceras ornatum*. Rare *Genuclymenia* cf. *angelini* and representatives of the *Protactoclymeniaimplana* Group add to the peculiarity of the fauna.

The two pre-event nodular limestone layers (Beds 1, 2b) yielded relatively poor conodont faunas. Long-ranging taxa, such as *Bispathodus stabilis vulgaris* (= *stabilis* M1), *Branmehla ampla*, *Br. inornata*, *Mehlina strigosa*, *Neopolygnathus communis*, *Palmatolepis gracilis gracilis*, and *Pa. perlobata schindewolfi* dominate. The first Moroccan record of *Pseudopolygnathus granulosus* in Bed 1 enables us to recognize locally a (partial) equivalent of the international *Ps. granulosus* Zone. Due to the rarity of the index species we prefer a local subzone subdivision of the regionally extended (HARTENFELS 2011) *Scaphignathus velifer velifer* Zone. Associated are *Pa. minuta minuta*, *Sc. velifer velifer*, and *Sc. velifer leptus*. The subzonal base should be sought somewhat below the actual record.

The Lower (Bed 3a, 3 cm thick) and Upper *Annulata* Events (Bed 4a, 44,5 cm thick) are developed as brownish to whitish weathered marly shales. They contain abundant small ammonoids, notably *Platyclymenia subnautilina* and *Prionoceras divisum*. None of the UD III ammonoids reach into the Lower *Annulata* Event Bed. Rather, the rarity of ammonoids within the beds slightly below documents a gradual extinction before the onset of the event sedimentation. The blooms of *Platyclymenia* and *Prionoceras* confirm a short-termed peak of eutrophication, as in many other sections. But the food resources remain unclear. Both event layers are separated by a 3 cm thick brownish-red nodular limestone (*Annulata* Intralimestone) dominated again by *Pl. subnautilina* and *Pr. divisum*. Two subsequent, very fossiliferous reddish-green nodular limestones (Beds 5b, 4 cm thick, 6b, 3 cm thick) are regarded as Wagnerbank Equivalents. They yielded *Pl. pseudoflexuosa*, *Pl. subnautilina*, *Pr. divisum*, and *Pr. frechi* and complete the *Annulata* Event Interval. Conodonts from Beds 3b, 4a, and 5b, include, among others, *Alternognathus regularis regularis*, *B. stabilis vulgaris*, *Neo. communis*, and *Pa. perlobata schindewolfi*. They fall in the regionally extended *velifer-stabilis*-Interregnum since the Anti-Atlas sections completely lack all zonal markers, which would enable the recognition of *Polygnathus styriacus*/*Pa. perlobata postera* or *Pa. gracilis manca* Zones.

The overlying cyclic succession (Beds 7a to 26a) consists of green marly shales with intercalated red, green, or grey nodular limestones and reaches a thickness of 3.34 m. This sequence yielded small-sized (especially Beds 25b, 26a) ammonoid faunas with *Pl. annulata*, *Pl. subnautilina*, *Pr. divisum*, *Pr. frechi*, *Trigonoclymenia protacta*, and *Erfoudites* n. sp. There is no record of *Procymaclymenia* or *Protoxyclymenia* in partly rich ammonoid collections. Therefore, it is assumed that the whole interval still falls in the *annulata* Zone (UD IV-A), which considerable thickness is explained by the basinal sedimentary conditions. Within this cyclic sequence, a 12 cm thick, red micritic marker limestone (Bed 19, base of the lateral trench El Gara South II) yielded a sparse conodont fauna that only consists of long ranging taxa. In addition to *Br. ampla* and *Br. inornata*, there are *B. stabilis vulgaris*, *Pa. gracilis gracilis*, as well as *Pa. perlobata schindewolfi*.

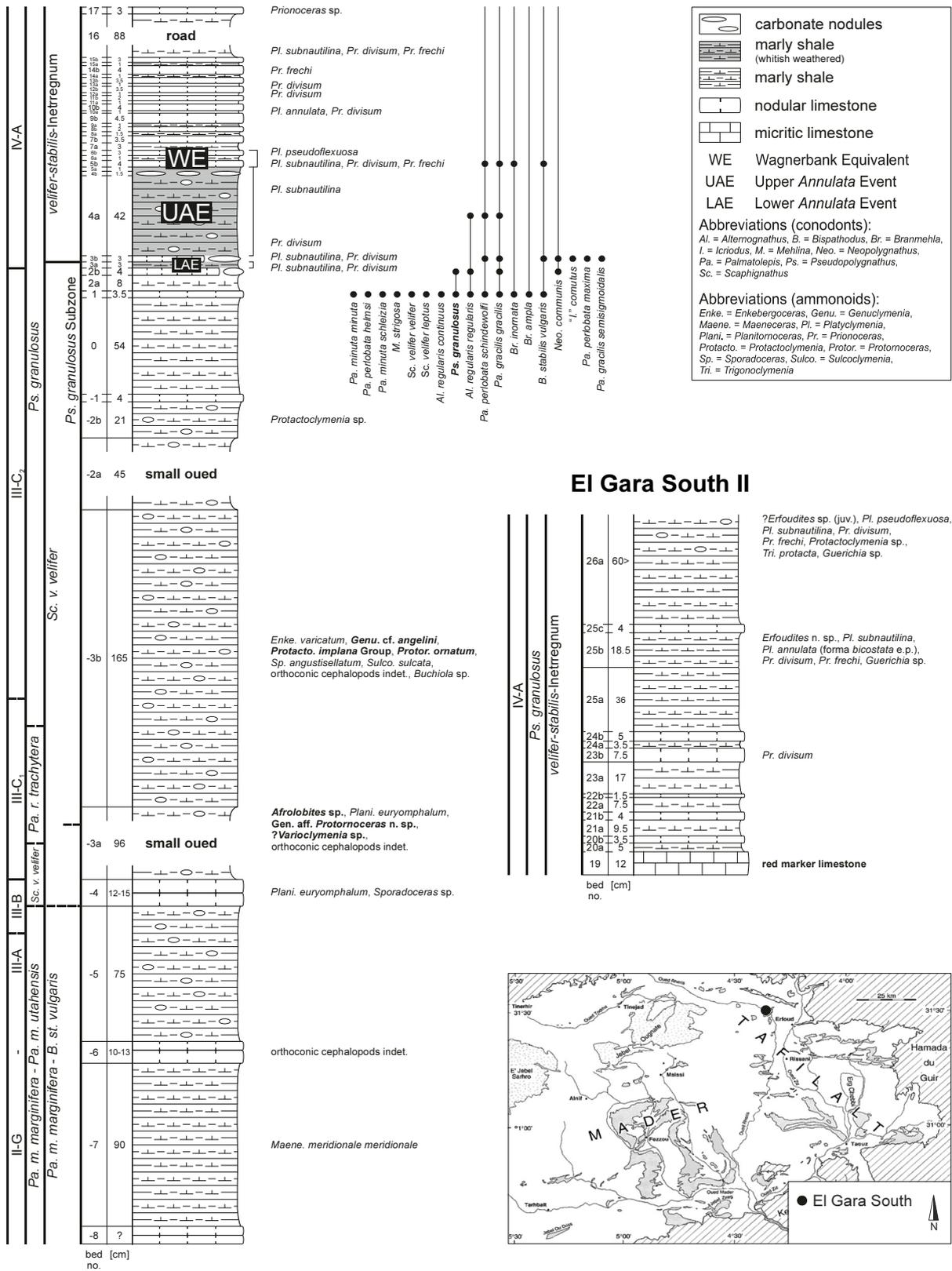
Towards the northern slope of the Tafilalt Platform, at Bine Jebilet (BECKER 1992, HARTENFELS 2011), both *Annulata* Events are similarly developed as *Platyclymenia*-rich greenish marls with small limestone nodules, separated by a nodular limestone. As at El Gara South, the subsequent Wagnerbank Equivalent consists of two beds. In the central Tafilalt at Jebel Erfoud (KORN 1999, HARTENFELS 2011), Bou Tchrafine (BECKER&HOUSE 2000), and Ouidane Chebbi NW (HARTENFELS 2011), there is no dark *Annulata* Shale or Limestone but the Wagnerbank Equivalent can be recognized and consists of a single condensed limestone (Bou Tchrafine, Ouidane Chebbi Northwest) or of two marl-limestone cycles (Jebel Erfoud). In the Tafilalt Basin, at Hassi Nebech (BECKER et al. 2002, HARTENFELS 2011), the succession resembles El Gara South but the Wagnerbank Equivalents are developed as a three-fold limestone. In the Amessoui Syncline (southern Tafilalt Platform, (KORN et al. 2000), a dark grey to black limestone with mass-occurrence of *Platyclymenia* and prionoceratids transgressed and reworked older strata (Takhbtit West, Jebel Ouauoufilal Pass, HARTENFELS 2011).

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### El Gara South I



**(P) Lower Devonian trilobites from a Carboniferous olistostrome near Aït Issa  
(Tinerhir area, southern margin of the Variscan Zone, SE Morocco)**

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The studied locality at Taourirt n'Khellil (= Rhellil), ca. 1.5 km north of Aït Issa (or Ait Aïssa) and in the southern plain below the Jebel Tisdafine, is located just south of the road between the towns of Tinerhir, about 20km to the West, and Tinejdad, about 30km to the East. GPS coordinates are N31° 26' 49,344" W5° 20' 57,948". It represents a thick olistostrome with small to house-sized olistoliths ranging from Ordovician quartzites to Upper Devonian pelagic limestones. They are assigned variably to a separate Taourirt n'Rhelif Formation or to the basal Aït Yalla Formation (DAL PIAZ et al. 2007). The massive, chaotic debris flows are evidence for a first major Variscan uplift, reworking and re-sedimentation event. The source area of the olistoliths probably lay to the south-southeast, towards the stable middle Paleozoic of the Anti-Atlas. The olistoliths include about 25 different lithologies. The distinctive trilobite bearing facies is a light grey, coarse-grained crinoid limestone with a sparitic matrix. Currently a conodont age is lacking. Besides trilobites, crinoids, orthocones, tentaculites, rare bivalves and gastropods (including *Platyceras* and *Orthonychia*) are the main fossil components. The trilobites consist of Phacopida, Proetida, Odontopleurida, and Scutelluida. The latter two are only known from fragmentary pygidia that cannot be identified precisely.

The main component of the Aït Issa trilobite fauna are disarticulated phacopid remains, cephalae and pygidia, which belong to *Reedops maurulus* ALBERTI, 1970. So far, it was only known from cephalae at its type locality in the Tiflet region (NW Meseta), from a calcareous marl bed of the Rechoua Limestone. This unit falls in the upper Pragian with *Caudicriodus celtibericus* (BENFRIKA et al. 2007). Although there are no articulated specimens of *R. maurulus* at Taourirt n'Khellil, all cephalae belong to the same species. Therefore, it is probable that the associated phacopid pygidia also belong to *R. maurulus*. These are much wider (tr.) than long (sag.), with a moderately vaulted axis region strongly tapering posteriorly. The pygidial dorsal furrow is faint and not recognizable within the range of the terminal axial part. Axial-pleural- and interpleural furrows are also faint to not recognizable; only the anterior ones can be distinguished. A pygidial border and border furrow are not developed. There is no tuberculation or granulation. These features clearly resemble other typical *Reedops* species with relatively smooth and inconspicuous pygidia.

Another faunal component at Taourirt n'Khellil is *Cheirurus (Crotalocephalina) gibbusauster* ALBERTI, 1970. Complete and fragmentary cranidia were found, but so far no pygidium. Besides its type location, in the Pragian Megrane Beds near the Tiddas-Oulmes road (northern Meseta, TERMIER et al. 1951), ALBERTI (1970: 89) also reported this species from the Pragian of the famous Hamar Laghdad (Tafilalt) and from the Pragian Tiflet Limestone of the eastern Rabat-Tiflet-Zone (NW Meseta). *C. (Crotalocephalina) gibbusauster* differs from *C. (Crotalocephalina) gibbustifletensis* ALBERTI, 1981, the second known Moroccan subspecies, in a higher vaulted (sag./tr.) glabella (ALBERTI, 1981) and a relative coarse granulation on the posterior part of the glabella and on the median parts of L2 and L3 (compare the *tifletensis* holotype in ALBERTI 1979: pl. 38 figs. 3a-d). HOLLOWAY & NEIL (1982) doubted that *C. (Crotalocephalina)* PRIBYL & VANEK, 1964 and *C. (Crotalocephalus)* SALTER (1853) are independent subgenera, because of intermediate morphological features. In contrast CHATTERTON & WRIGHT (1986) interpreted both as independent and limited the extent of *C. (Crotalocephalina)* to species similar to the type species, *C. (Crotalocephalina) gibbus*, in having short marginal spines on the pygidium and a median lobe or marginal spine. However, LANE (1971) and the above mentioned authors agree that it is hard to

distinguish between the type species of *C. (Crotalocephalina)* and *C. (Crotalocephalus)* only on cephalic morphologies. Therefore, the identification of our cheirurids remains somewhat ambiguous until pygidia become available.

Proetid remains are locally rare: two small cranidia and a single pygidium. One of the cranidia clearly belongs to *Pragoproetustafilaltensis* (ALBERTI, 1964), a Pragian species so far only known from the Lower Hamar Laghdad Limestone (ALBERTI, 1964, 1969). The cranidium is only 3mm long (sag.) and shows the characters given by ALBERTI (1969: fig. 15) for juvenile holaspid specimens. ALBERTI (1969) showed that *P. tafilaltensis* has a high intraspecific variability.

The second incomplete cranidium and the pygidium – only the ventral side is preserved – belong to *Eremiproetus?* RICHTER & RICHTER, 1919. Due to their fragmentary character, a more precise identification is not possible. The cranidium has some typical features of *Eremiproetus*, such as the (relative to the glabella) lower and broad (sag.) occipital ring, and an upwards arched anterior border with subparallel terrace lines. But there are features which clearly differ from typical members of the genus. The outline of the glabella in the anterior part is more or less rounded, whereas a more or less subquadratic outline would be typical (ALBERTI, 1969). This author also stated that the glabella tends towards a violin shape, which is poorly visible in our incomplete specimen.

The Taourirt n'Khellil/Aït Issa trilobite assemblage is clearly of Pragian age. It combines faunal links with the Tafilalt and the northern Meseta, as it can be expected from its intermediate position at the southern margin of the Variscides. Additional material is necessary to clarify open taxonomic questions, such as the affinities of the odontopleurid and scutelluid remains. It is likely that the complete assemblage includes additional taxa. A more complete knowledge may enable a better identification of the source of the allochthonous Devonian of the southernmost Variscides.

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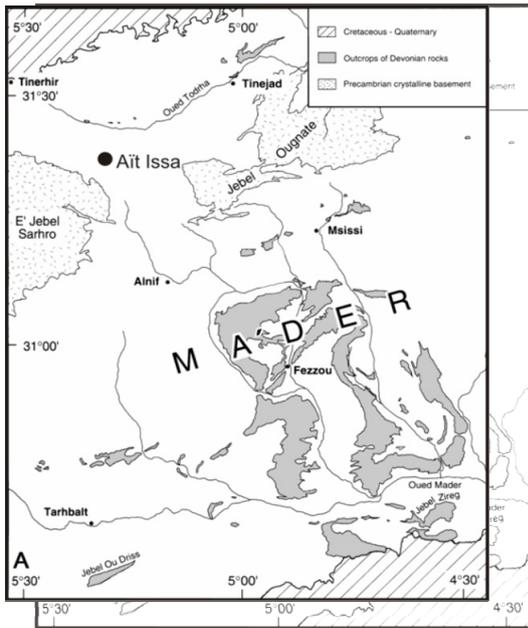


Fig. 1: Geological overview of the Ma'ader Region and the position of Ait Issa at the northwest.



Fig. 2: Overview of the allochthonous Devonian outcrop near Ait Issa.

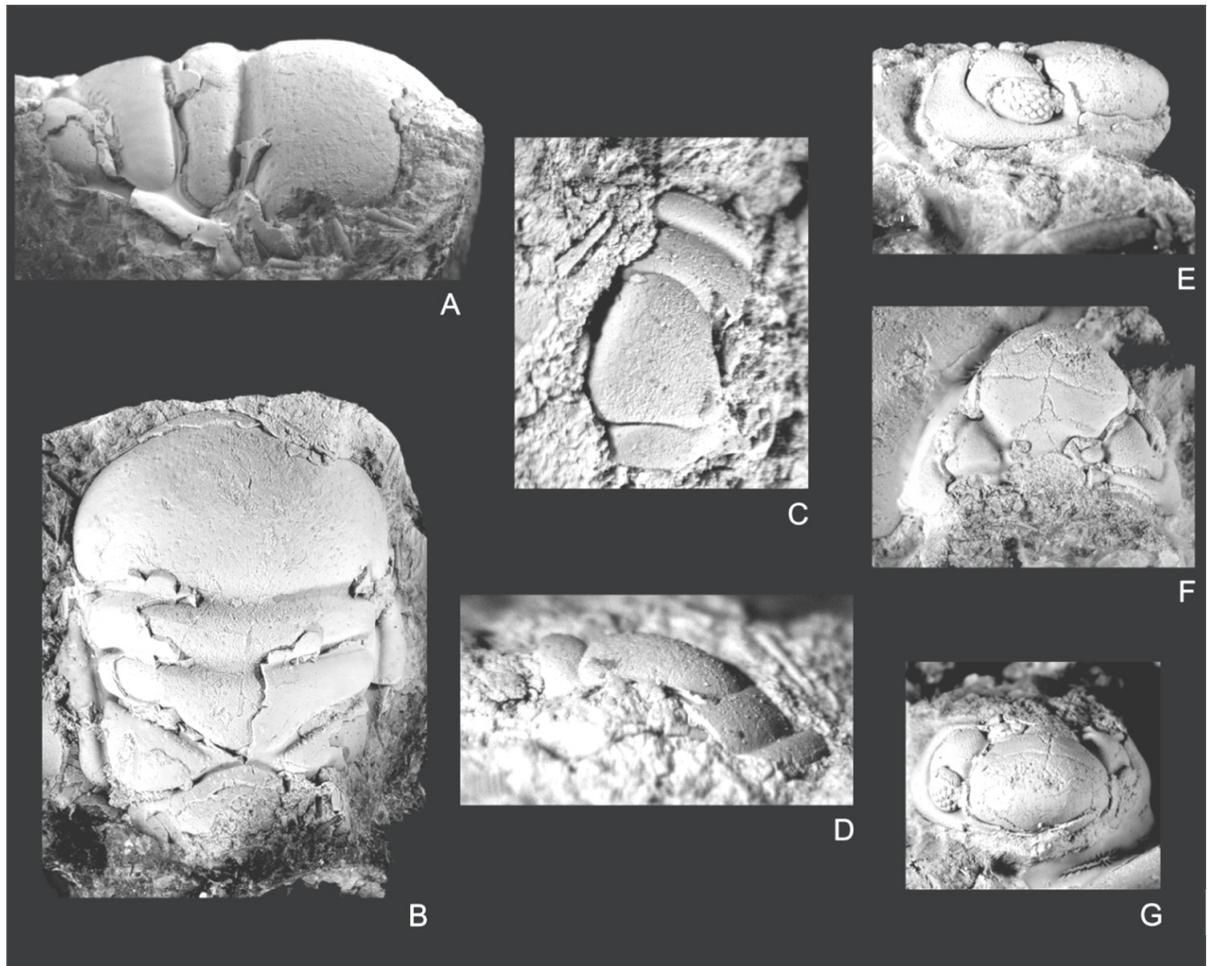


Fig. 3: Pragian trilobites from Ait Issa A-B *Cheirurus (Crotalocephalina) gibbus auster* ALBERTI, 1970 (A) Lateral view x2,5 (B) Dorsal view x2,0; C-D *Pragoproetus tafilaltensis* (ALBERTI, 1964) (C) Dorsal view x10,4 (D) Lateral view x11,0; E-G *Reedops maurulus* ALBERTI, 1970 (E) Lateral view x4,8 (F) Dorsal view x5,1 (G) Frontal view x4,3.

## (P) The Mississippian of the Jerada Massif (NE Morocco)

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### Introduction

Palaeozoic outcrops from the eastern Moroccan Meseta are restricted to a couple of smaller massifs. By areal extent, most important are those of Debdou-Mekkam and Jerada, SW and S of Oujda (Fig. 1, inset). They belong to the internal zone of the Moroccan Variscides, characterized by a poorly dated Late Devonian-Tournaisian “Eovariscan” synmetamorphic folding event (MICHARD et al. 2008, cum lit.). Subsequently deposited Carboniferous strata contain quite monotonous, often poorly dated, and strongly tectonized successions. The Jerada Massif (Fig. 1), a WSW-ENE oriented, maximum 15 km wide, internally folded and faulted synclinorium with shallow dipping northern and steeply dipping southern flank contains a well differentiated marine Mississippian succession. Therefore, it is a key area for reconstruction of basin evolution, geodynamic setting and palaeogeography of the period preceding the main Variscan deformation. Based on detailed mapping and logging, we present results hitherto only partly published in HERBIG et al. (2006), HUCK et al. (2006), ARETZ & HERBIG (2008), ARETZ (2010), and in some additional abstracts and give an overview of current studies.

### Stratigraphy

HERBIG et al (2006) revised the lithostratigraphic subdivision of BERKHLI et al. (1999), which was based on repeated successions in tectonic slices, and first established common formations for the northern and southern flank of the synclinorium (Fig. 2).

Lowermost, the **Oued Defla Formation** (BERKHLI et al. 1999) consists of rhyolitic, rhyodacitic, dacitic and andesitic volcanics, volcanosedimentary rocks and minor intercalations of reddish and greenish shales. Maximum thickness along the northern flank is about 250 m; along the southern flank according to own mapping 1100–1700 m. A late Viséan age (OWODENKO 1976, BERKHLI et al. 1999) had been inferred from fossils from overlying strata (Oued Es-Sassi Fm., see below), resp. from adjoining tectonic slices. ARETZ & HERBIG (2008) discussed a Tournaisian base of the formation, judged from an extended time span necessary for sedimentation of the overlying Çafçaf Fm. (HERBIG et al. 2006).

The **Çafçaf Formation** (HERBIG et al. 2006) is a succession of variegated radiolarian-bearing cherts, siliceous shales and intercalated pyroclastics. A strata-bound manganese mineralization is of importance. Thicknesses along the northern flank vary from 30-150 m, but attain only 2-10 m in the south. A single reworked rugose coral colony from the lower third of the formation, *Siphonodendron irregulare*, gives the oldest reliable biostratigraphic datum in the Jerada synclinorium, indicating a mid to late Viséan age.

The succeeding **Oued El Koriche Formation** (HERBIG et al. 2006) is a succession of interbedded siltstones, calcareous greywackes, calciturbidites, polymictic conglomerates with sedimentary components, and tuffites, 75-180 m thick along the northern flank. In the south, additionally limestone boulders and limestone olistoliths up to 200 m long and 60 m thick occur in at least two horizons within the 15-85 m thick formation. The olistoliths are gravitationally reworked (parts of) microbial-sponge buildups (ARETZ & HERBIG 2008; see also HUCK et al. 2006, who did not yet recognize the allochthonous nature). In the north, foraminifers indicate the early Brigantian (VACHARD & BERKHLI 1992: lower V3c, samples G16-G23, SR3). Rare reworked corals favour an “early and/or late Asbian age” (ARETZ 2010). This is in accordance with some *Goniatites globostriatus* from the overlying Oued Es-Sassi Fm. in the south, which

indicate the (early) late Asbian for the fossil horizon (D. KORN in ARETZ & HERBIG 2008). New conodont and microfacies samples are currently studied.

On top, the **Oued Es-Sassi Formation** (HERBIG et al. 2006) is a quite monotonous series of dark shales with goniatite-bearing sideritic nodules and few intercalations of sandstones, calcareous sandstones and limestones. Thicknesses in the north are 130-250 m, in the south 55-150 m. Based on rugose corals, a late Asbian age is probable (ARETZ 2010). This is consistent with the above mentioned find of *G. globostriatus* and the oldest goniatite assemblage from the Chebket el Hamra outlier at the northeasternmost end of the Jerada synclinorium (KORN & EBBIGHAUSEN 2008; see below).

Strong facies and thickness variations are recorded in the following **Koudiat Es-Senn Formation** (BERKHLI et al. 1999), which is characterized by the entry of carbonate rocks. Two shallowing-upward cycles occur at the northern flank. In the western and middle segment, limestones, dark shales and minor sandstones alternate in the lower cycle parts. In the upper parts, mostly pure limestones become oolitic towards the top. In the upper cycle a dekameter-sized coral patch reef was recorded (HERBIG et al. 2006, ARETZ 2010). The cycle is locally capped by an up to 20 m thick ignimbrite below goniatite-bearing Namurian shales that overly with erosional unconformity. Towards the eastern segment, carbonate sediments in both cycles are completely suppressed by a conspicuous sandy intercalation. East of it, calcareous sandstones are predominant; limestones are completely silicified. At the southern flank, a single shallowing-upward cycle includes autochthonous microbial-sponge buildups, 100-450 m wide and 10-50 m high, and overlying, some meter-sized microbial-metazoan buildups with abundant rugose corals, both of shallow-water origin (ARETZ & HERBIG 2008). Namurian shales follow on top of a cartographic angular unconformity. In the north, the formation is about 175-265 m thick, in the south up to 150 m at the buildups and 50-80 m off-buildup. Calcareous microbiota are late Brigantian on both flanks (upper V3c, VACHARD & BERKHLI 1992, BERKHLI et al. 1999). ARETZ (2010) concluded a "late Asbian and/or early Brigantian age" based on rugose corals and favoured a late Brigantian to basal Serpukhovian hiatus on top of the formation. For an about 10 m thick package of fine-grained limestones with few intercalated siliciclastics at the westernmost end of the syncline ("Oued Agaia Limestone", ARETZ & HERBIG 2008), BERKHLI et al. (1999) claimed a Serpukhovian (E2, Arnsbergian) age. Currently, carbonate microfacies from detailed sections are studied.

West and southwest of Touissit (Fig. 1, inset) the easternmost outliers of the northern flank of the synclinorium crop out. The monotonous Mississippian succession (KORN & EBBIGHAUSEN 2008) begins with 20 m of tuffites and turquoise-coloured cherts overlain by more than 50 m of silty shales with thin sandstone beds. These are equivalents of the Çafçaf and Oued El Koriche formations. Overlying, about 200 m of nodule-bearing dark shales include very thin limestone intercalations. They bear a rich goniatite fauna of Late Asbian to late Brigantian age; the uppermost Brigantian ammonoid zones were not proved. According to age and lithology, these are distal, more basinal equivalents of the Oued Es-Sassi and Koudiat Es-Senn formations. They are named **Chebket El Hamra Formation** (new).

### **Basin evolution, geodynamic setting and palaeogeography**

The Late Tournaisian(?) to latest Viséan (late Brigantian) succession represents a single basin fill sequence (HERBIG et al. 2006). The cyclic development postulated by BERKHLI et al. (1999) has to be rejected (see also KORN & EBBIGHAUSEN 2008). After extrusion of rhyolitic to andesitic volcanics at deep-seated faults, basin plain sedimentation (Çafçaf Fm.), toe-of-slope and lower slope sedimentation (Oued El Koriche Fm.) started to level the volcanic submarine relief. The Oued El Koriche Fm. testifies the collapse of a southern carbonate shelf with marginal microbial-sponge buildups; more distal facies is developed in the north and northeast. Basin fill continued with slope sediments (Oued Es-Sassi Fm.). On top, a shallow, mixed carbonate-siliciclastic platform with different build-up types developed before the latest Viséan emergence of the basin (Koudiat Es-Senn Fm.). Emergence at the southern flank was probably earlier and according to a cartographic angular unconformity more pronounced. The basin was deepening towards the northeast, where the most distal deposits are found (Chebket El-Hamra Fm.).

The geodynamic setting of the basin is not yet clear. Different models are used to explain the structural relations between eastern and western Meseta (see MICHARD et al. 2008, fig. 3.36). HERBIG et al. (2006) and ARETZ & HERBIG (2007) suggested a back-arc basin, resp. a basin dominated by strike-slip faults. HERBIG & ARETZ (2007) postulated the Jerada basin - part of a larger basin comprising at least parts of the eastern Meseta - to be a continental margin type back arc basin strongly dissected by strike-slip faults due to oblique, south-directed subduction occurring further north, resp. northwest. Such a scenario was recently used by MICHARD et al. (2008, cum lit.) for a new geodynamic model of the Moroccan Variscides, strengthening our suggestion (Fig. 3). Current source studies of siliciclastic sediments from the Jerada basin might further elucidate the problem.

A repeatedly suggested eastern prolongation of the eastern Meseta into the internal zones of the Beticorifean arc (Malaguides-Ghomarides) and the Kabylies is not supported by our palaeobiogeographic data. Rugose corals comprise many taxa known from NW Europe, but taxa typifying the Malaguides or the Montagne Noire, i.e. the southern branch of the European Variscides and the Palaeotethys are missing (ARETZ 2010). Also the goniatites are typically Rhenohercynian, with many taxa in common with the South Portuguese Zone and the Rhenish Massif (KORN & EBBIGHAUSEN 2008, KORN et al. 2012).

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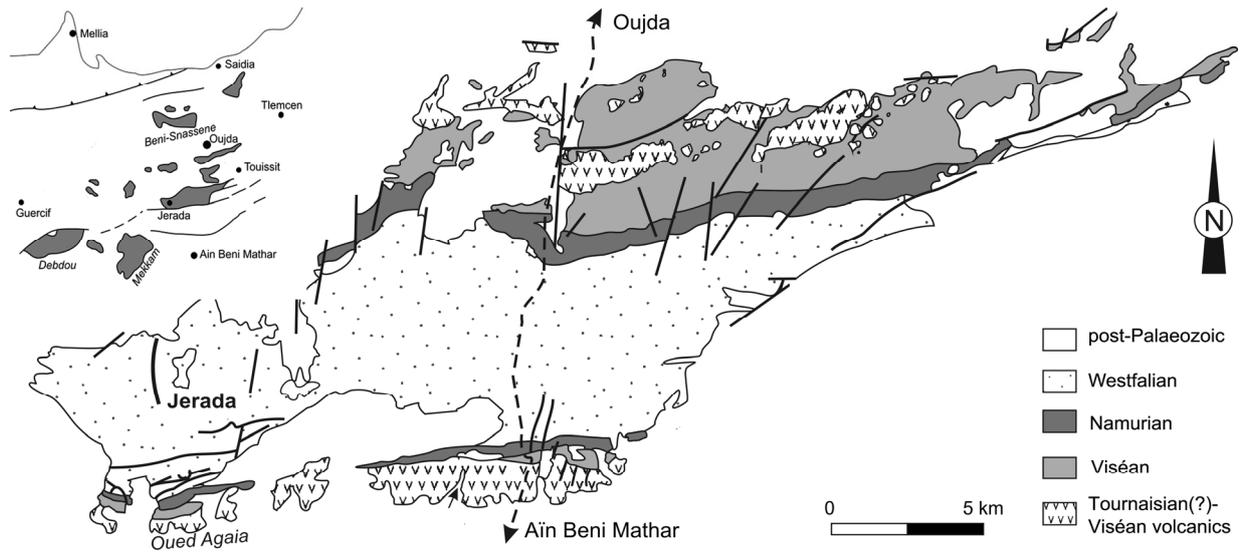


Fig. 1: Geological map of the Jerada synclinorium (modified from OWODENKO 1976 and author's mapping). Inset: Palaeozoic massifs of the Eastern Meseta.

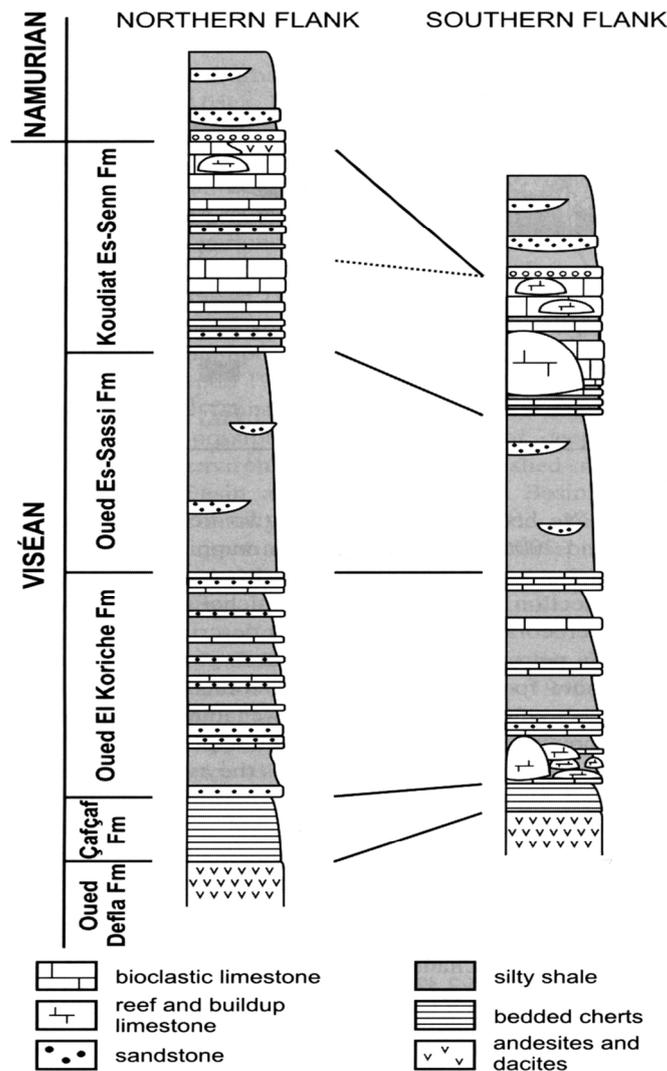


Fig. 2: Schematic lithostratigraphy of northern and southern flank of the Jerada synclinorium, not to scale (from ARETZ 2010).

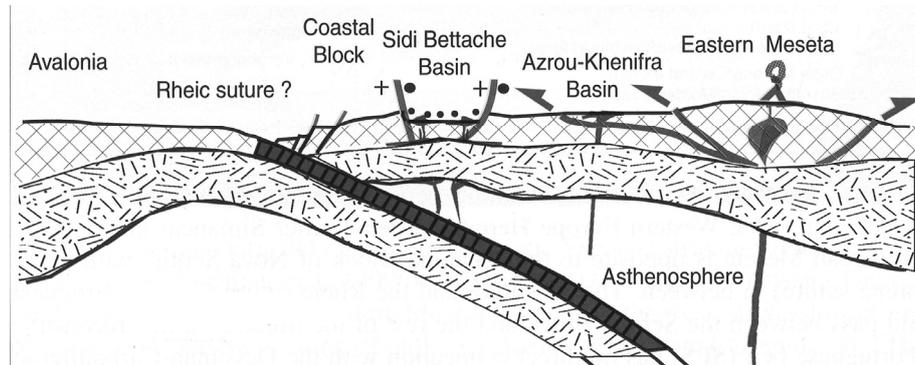


Fig. 3: Geodynamic model of MICHARD et al. (2008) with southeastern dipping subduction and back-arc setting of the Eastern Moroccan Meseta.

**(P) The nature, range, and utility of stratigraphic distributions of morphotypes of *Polygnathus linguiformis linguiformis* (HINDE), Middle Devonian, northern Appalachian Basin**

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The Hamilton Group of New York in the northern Appalachian Basin of eastern North America is a package of marine shales and limestones deposited during the Middle Devonian noted as a time of evolutionary stasis for ~5 million years. The Cherry Valley Limestone, Centerfield Limestone, Windom Shale, and North Evans Limestone in New York State, and the *Hadrophyllum* Bed of the Delaware Limestone in Ohio were analyzed to determine if there were detectable changes in *Polygnathus linguiformis linguiformis* (HINDE) through this interval. Morphometric analysis supports the ranges and differentiation between WALLISER & BULTYNCK (2011)  $\gamma$ 1a and  $\gamma$ 1b morphotypes, and proposes two new morphologies in New York and Ohio, the Cherry Valley morph, and the Centerfield morph (Fig. 1). These four morphotypes have stratigraphic utility within the Hamilton Group of western New York State.

Morphometric analyses were conducted using the software program "MorphoJ" (KLINGENBERG, 2011). "MorphoJ" uses landmarks and covariant analysis to distinguish morphometric variation between the specimens. Five landmark points were chosen for the element shape analysis: the posterior and anterior tips of both the inner and outer platform margins, and the central node. Another physical characteristic chosen to study were the transverse ridges, or ribs, that run across the tongue of the posterior platform. The tongue region begins when the carina comes to an end on the posterior platform, approximately two-thirds down the length of the element. The rib count of each specimen was recorded and an ANOVA test was executed using "Minitab" (MINITAB 16, 2010). A minimum of ten specimens were analyzed from each stratigraphic unit.

All four morphotypes of *P. l. linguiformis* are present during the *eiflius* Zone, within the Cherry Valley Limestone and the Delaware Limestone. Although these units are equivalent, the relative abundances of the morphs differ slightly between the localities. The Cherry Valley morph and the  $\gamma$ 1b morph make up almost the entire population of the Cherry Valley Limestone, while there is a stronger representation of the  $\gamma$ 1a morph and Centerfield morph in the Delaware Limestone. The Centerfield Limestone is the only unit which contained abundant Centerfield morphotypes. Within the Centerfield Limestone there was a reduction of  $\gamma$ 1b morphs, absence of the Cherry Valley morph, and an increasing presence of the  $\gamma$ 1a morphotype. *Polygnathus linguiformis linguiformis* recovered from the Windom Shale (*latifossatus* Zone) are only of the  $\gamma$ 1a morphotype. This population is seen again in the North Evans Limestone and is most likely, the same fauna, in part, reworked into younger strata. The  $\gamma$ 1a morphotype appears to increase in abundance throughout the Givetian and it eventually dominates the entire population. Each morphotype has a distinct range within the Middle Devonian and prove to be stratigraphically significant (Fig. 2).

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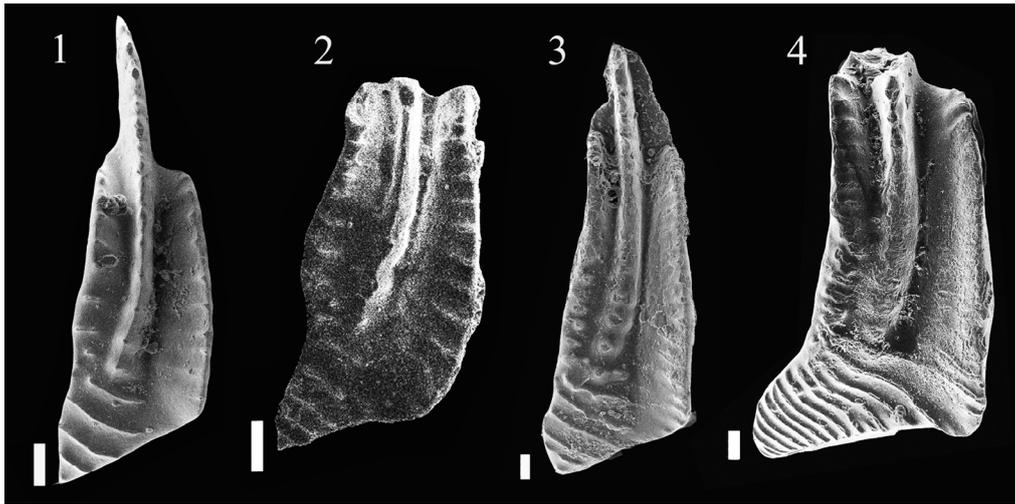


Fig. 1. Four morphotypes of *Polygnathus linguiformis linguiformis* that occur in the Hamilton Group of western New York. 1, Cherry Valley morphotype. 2,  $\gamma$ 1b morphotype. 3, Centerfield morphotype. 4,  $\gamma$ 1a morphotype. Scale bar = 0.1 mm.

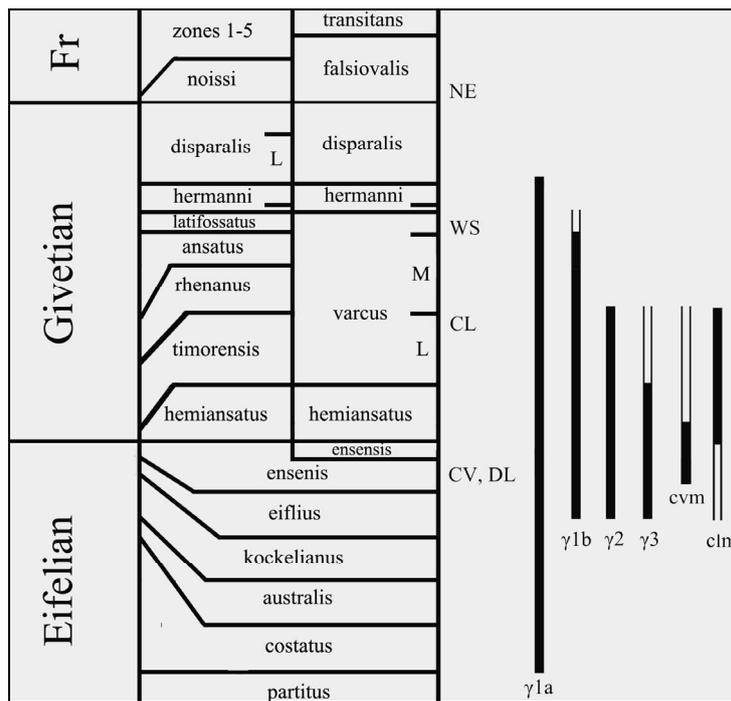


Fig. 2. Ranges of all morphotypes of *Polygnathus linguiformis linguiformis*, including the  $\gamma$ 2 and  $\gamma$ 3 morphs from WALLISER & BULTYNCK (2011) relative to Eifelian-Givetian conodont zonation; cvm, Cherry Valley morphotype; clm, Centerfield morphotype; hollow portions indicate rare occurrence. NE, North Evans Limestone; WS, Mid-Windom Conodont Bed, Windom Shale; CL, Moonshine Falls Bed, Centerfield Limestone; CV, Cherry Valley Limestone; DL, Hadrophyllum Bed of the Delaware Limestone; Fr, Frasnian.

## **Facies and sediment architecture of Upper Devonian limestones at Gara de Mrirt, Eastern Moroccan Central Massif: Resedimentation in response to block faulting**

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The Upper Devonian at Gara de Mrirt consists of calcareous bioclastic contourites and pelagic limestones. Drastic differences in sediment thickness and abrupt lateral facies changes within the pelagic sequences indicate structural control over deposition. Although the "highs" and the "basins" in pelagic carbonate systems have long been recognized in ancient rifted continental margins, relatively little detailed information has been available on facies transitions between the two and on the geometries of contourite and pelagic sediment bodies. Outcrops at the Gara de Mrirt provide new constraints for Devonian palaeogeographic reconstructions, allowing for the development of facies models for these depositional systems.

The investigated area is located within the nappe of Ziar-Mrirt at the eastern margin of the Moroccan Central Massif. The nappe of Ziar-Mrirt is well exposed in the tectonic unit of Mrirt and comprises a series of Ordovician to Lower Carboniferous strata. During the Carboniferous, the nappe pile was deformed by Variscan compressive events and overthrust on the autochthonous unit of Tanadra-Bou-Tazart (e.g., BOULIN et al. 1988, BOUABDELLI 1989, and references therein), which also consists of Ordovician to Lower Carboniferous strata.

Devonian hemipelagic and pelagic sediments of the study region record continuous deposition on continental rises and terraces (passive continental margin sequence). While the presence of Lochkovian sediments has not been confirmed previously, Pragian to Givetian strata mainly consist of hemipelagic shales and siltstones (< 80 m) with rare intercalations of calcareous shale, limestone and more rarely sandy shale (WALLISER et al. 2000). Platy limestones with thin shale intercalations of Frasnian and early Famennian age represent a condensed sedimentary record (~ 8 m) (LAZREQ 1992, BECKER & HOUSE 2000). The fauna is of pelagic origin and includes cephalopods, thin-shelled bivalves, small brachiopods, styliolinids, solitary corals, conodonts and ostracods. The Kellwasser Limestone Horizon signifies the Frasnian-Famennian boundary. The platy limestones are overlain by nodular limestones and marly shales of middle-late Famennian age that include thick limestone conglomerate beds and limestone breccias (10–30 m) in the south-eastern part of the Mrirt unit (LAZREQ 1992, WALLISER et al. 2000). These limestones are of similar pelagic composition. Latest Famennian greenish shales with intercalations of siltstone and siliciclastic conglomerate layers form the transition to similar Lower Carboniferous dark-grey shales.

The Devonian sediments of the Ziar-Mrirt nappe were deposited somewhere along the disintegrated northern continental margin of Gondwana, in southern parts of the narrow Protoatlantic–Prototethys ocean (e.g., see STAMPFLI et al. 2011). The primary plate tectonic relation to the West African craton is difficult to reconstruct due to dextral strike-slip faulting during and subsequent to the Variscan collision (e.g., MATTE 2001).

Integrated field, sedimentological, biostratigraphical and palaeontological studies of the Upper Devonian pelagic carbonate system in the Mrirt unit (LAZREQ 1992, HÜNEKE 2001, BECKER & HOUSE 2000, WALLISER et al. 2000) resulted in the recognition of distinctive patterns of vertical and lateral facies changes that are indicative of pelagic carbonate platforms as defined by SANTANTONIO (1993, 1994). Subsidence and sedimentation are controlled by a pelagic horst-and-halfgraben system that arose from tectonic disintegration and block tilting during middle-late Famennian (Fig. 2). It was initiated by uplift of

the Midelt Zone and/or onset of rifting of the Azrou-Khenifra Basin in the eastern Central Massif (see BEAUCHAMP & IZART 1987, WALLISER et al. 2000).

Limestones of Frasnian and early Famennian age form a succession of virtually uniform thickness (~ 8 m) that overlie the Pragian-Givetian succession of hemipelagic shales and siltstones along an erosional unconformity of regional character. The Frasnian limestones include (in particularity within its lower part) skeletal calcarenites and calcisiltites of pelagic composition and are interpreted as bottom-current deposits (HÜNEKE 2006, 2007).

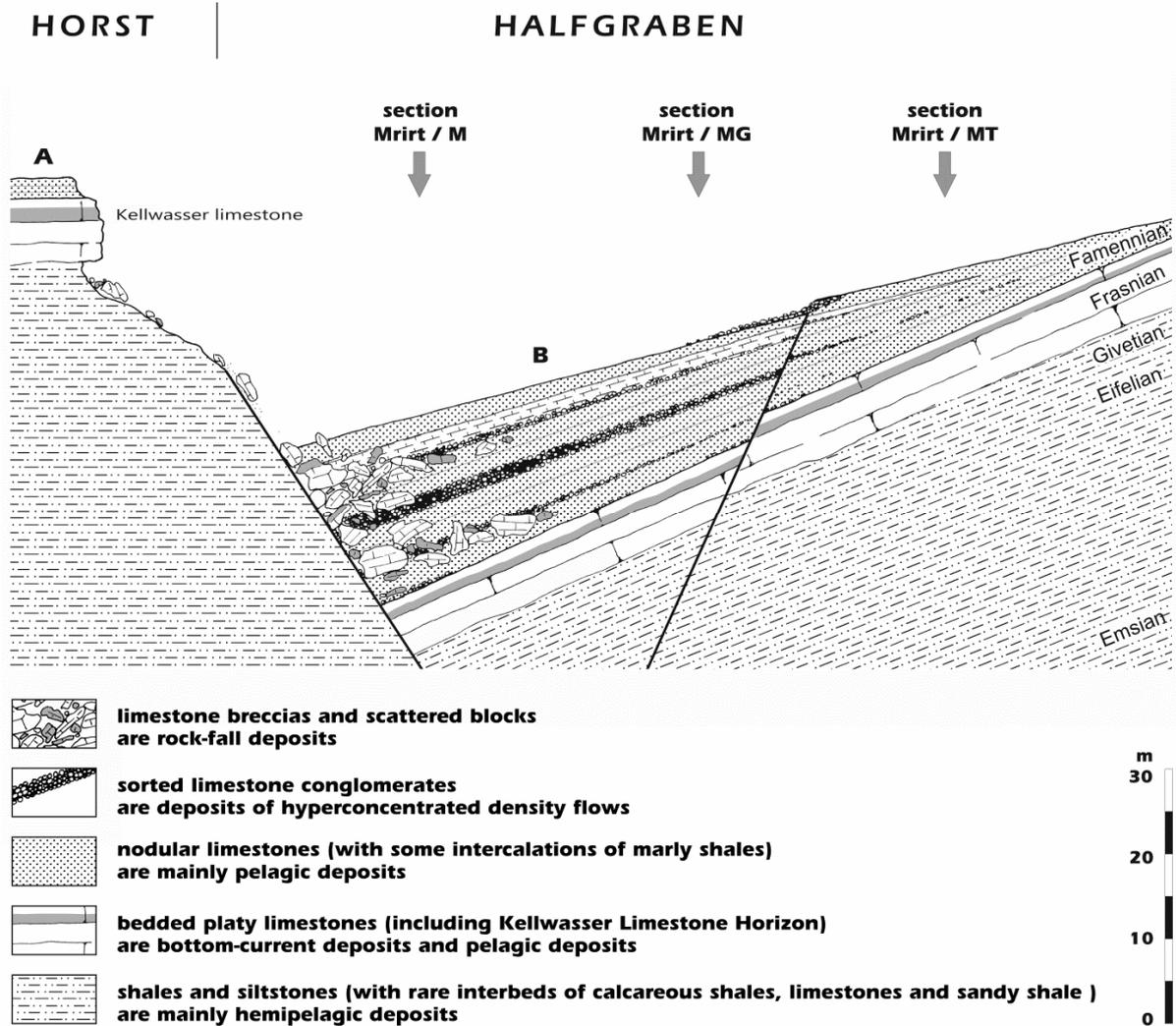
Limestones of middle-late Famennian age are differentiated on the basis of sediment thickness, which varies considerably (10–30 m), and facies association: (A) A condensed pelagic facies association includes deposits characterized by concentrations of macrofossils (mainly cephalopod) due to episodic winnowing of fines, the occurrence of sedimentary hiatuses, and hardgrounds. Most diagnostically, they generally lack gravity-flow deposits. (B) A normal and resedimented pelagic facies association can be recognized as interbedded background pelagic/hemipelagic deposits and resedimented deposits (Fig. 1). The latter include deposits mainly produced by both rock fall and hyperconcentrated density flows. Clasts of these resedimented deposits are composed of pelagic limestones of mainly Frasnian and Famennian age. More rarely, phosphate and chert occur. The rock-fall deposits are breccia beds and scattered isolated large blocks (olistoliths), indicating rock falls from steep submarine cliffs that are a product of synsedimentary faulting. The deposits of hyperconcentrated density flows are sorted limestones conglomerates that fill the erosional relief of channels or form a lens-like geometry. Sediment architecture, pattern of thickness and facies changes indicate deposition on a pelagic carbonate ramp such as produced by a half graben.

The two documented facies associations are reliable markers of the local patterns of faulting and submarine topography and their related sedimentary processes (compare SANTANTONIO 1993, 1994). The condensed pelagic facies association is interpreted to be deposited on horsts, whereas the normal and resedimented pelagic facies associations was accumulated in halfgrabens and along dip slopes of tilted blocks.

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**Fig. 1.** Interpretative facies diagram of the Upper Devonian pelagic carbonate platform at Mrirt illustrating the depositional processes and distribution of facies association A (condensed pelagic), and B (normal and resedimented pelagic) in a horst-and-halfgraben structural setting during the middle to late Famennian (HÜNEKE 2001).

## (P) Uppermost Famennian Conodonts from Kuznetsk basin (South of West Siberia)

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Upper Famennian sequences crop out in the western part of the Altai-Sayan Folded Area (ASFA), within marginal structures of the Kuznetsk Basin. They are represented by the red-colored sequence of the Podonino Formation, which almost lack fossils. First Upper Famennian brachiopods were collected A.V. TYZHNOV (1938) from red rocks in Section Nevsky, who described the section.

Section "Nevsky" is situated 50 km to the north-west of Kemerovo town, on the left side of the Yaya River, downstream from the confluence of its tributary Barzass River. It is exposed near Nevsky village by the Anzhero-Sudzhensk-Kemerovo railway excavations. First data on microfossils were published by GUTAK et al. (2001). Recently this section was re-investigated by collaborators from IPGG SB RAS in order to clarify the biostratigraphic characteristics of ASFA Late Famennian strata.

Section "Nevsky" is represented by red terrigenous rocks with rare carbonate intercalations, often composed of brachiopod shell-rock. Twelve carbonate samples were dissolved for microfossils. They revealed the presence of conodonts, gastropods, ostracods, bivalves, and fish remains. The most diverse fauna was obtained from the lower part of the section. Conodonts occur through all the section, among them: *Siphonodella*, *Icriodus*, *Pseudopolygnathus*, *Polygnathus*, and *Pelekysgnathus* (Pb elements). The conodonts are grey-brown, semitransparent, shiny, well preserved, and have a sharp-angled shape of their elements. They are irregularly distributed along the section. Most abundant assemblages were collected from the base of the section, including *Siphonodella*, *Icriodus*, *Polygnathus* and *Pseudopolygnathus*. The number of conodonts is reduced towards the upper part, where mainly *Pelekysgnathus* (Pb elements) were collected. The conodont association contains: *Icriodus costatus* THOMAS, *Pelekysgnathus* sp. (Pb elements), *Polygnathus delicatulus* ULRICH & BASSLER, *Polygnathus lenticularis* GAGIEV, *Polygnathus parapetus* DRUCE, *Polygnathus symmetricus* BRANSON, *Pseudopolygnathus postinodosus* RHODES, and *Siphonodella praesulcata* SANDBERG.

The distribution of conodonts in the "Nevsky" section allows us to align it with the *praesulcata* Zone of the Uppermost Famennian.

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## **(P) Early Devonian brachiopod faunas, facies and bioevents of the Rhenish Massif (Germany)**

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The Lower Devonian of the Rhenish Massif documents the history of an extensive tropical shelf along the southern border of the Old Red Continent. Changing sea-level and varying rates of subsidence and sedimentation governed the development of palaeogeography and palaeoenvironments. Regarding stratigraphic intervals on formation level, lateral variations of the "rhenotypic" facies and brachiopod assemblages reflect different marine biotopes from the deep shelf to intertidal settings, mostly under the influence of suspended siliciclastic material derived from the continent. Vertical changes, on the other hand, document the successive appearance and disappearance of characteristic brachiopod faunas reflecting ecological-evolutionary subunits of possible supraregional significance. The Pridolian to earliest Eifelian succession of openly marine faunas is characterized, in stratigraphical order, by the spiriferids *Quadrifarius dumontianus*, *Howellella mercurii*, *Multispirifer solitarius*, *Hysterolites hystericus*, *Arduspiriferlatestriatus*, *Euryspiriferparadoxus* and *Paraspirifercultrijugatus*. Major faunal turnovers occur at events separating these faunas, whereas faunal change was relatively modest between the events. Although Hermeskeil, Seifen, Kürrenberg, Saxler, Bornich, Spitznack, Stadtfeld, Berl , Laubach and Kondel events were recognized as the most significant ones in the Rhenish Massif (MITTMEYER 2008), the significance of each of these is poorly known in terms of faunal turnover. The Saxler Event, for example, appears to correlate with the onset of transgressive unit 1b sensu JOHNSON et al. (1985) representing a "basal Emsian Event", and the Berl  Event corresponds to the Daleje Event. The early Eifelian OCA (*orbignyanus-cultrijugatus-alatiformis*) Event (STRUVE 1982) marks the disappearance of the Early Devonian type of fauna and could be a neritic equivalent of the Chotec Event. The faunal changes are chiefly due to regional extinction of taxa and subsequent immigration of new taxa close to an event.

As regards Early Devonian palaeobiogeography, faunal relationships suggest the presence of a separate "Maghrebo-European Subrealm" within the Old World Realm (or even a separate realm). The "Rhenish Province" represents the northern part of this subrealm, opposed to the "North Gondwanan Province" in the south (including, at least, the faunas of Armorica, Celtiberia, Cantabria, Anti-Atlas/Ougarta Chains and Turkey). Neither the presence of an extensive Rheic Ocean nor Palaeotethys Ocean is supported by the faunal data.

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## Intraspecific variability of Devonian ammonoids

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Heritable phenotypic variation is usually thought to be the raw material for evolution and natural selection. Knowledge of the intraspecific variability of any taxon is therefore essential for systematics and diversity studies. Heritable variation is, however, hard to separate from morphological variation resulting from a plastic response to the environment, especially in extinct groups. For instance, a large part of the intraspecific variability in shelled molluscs could be caused by differences in growth rates, which could also explain certain recurrent patterns in intraspecific variability in the shells of coiled mollusc shells. Probably because of a low amount of specimens, poor preservation of study materials and/ or the lack of time and/ or motivation, intraspecific morphological variability has been poorly studied in ammonoids (as it is the case in most other groups of fossil and extant organisms). In ammonoids, this might lead to biases and problems such as, e.g., overlooking the influence of coiling on the intraspecific variability, oversplitting of taxa, and insufficient knowledge of taxa.

We have finished first studies on variability in relation to coiling with loosely coiled ammonoids of the genera *Anetoceras* and *Erbenoceras* from the early Emsian (Early Devonian) of Morocco and could demonstrate high degrees of intraspecific variability in both genera (DE BAETS et al. 2012). We compared the morphometric data of the Moroccan material with published specimens of the same age worldwide. This led to the conclusion that the current amount of valid ammonoid species of the studied interval (> 20) is not supported by differences in the classical dimensional shell characters. Since the suture lines of these ammonoids are extremely simple, they are insufficient to base species exclusively on their suture line courses. This led to the conclusion that the existing wealth of species with loosely coiled shells has originated from oversplitting and has no profound scientific base. Estimates of ammonoid diversity are thus probably too high for the early Emsian.

In the past decade, we have collected and measured many Palaeozoic ammonoids, predominantly of Early and Middle Devonian age, but also of Late Devonian and Carboniferous age. These included species that have the advantages that several of which originated more or less close to the origin of the respective clade and that several show traces of ancestral characters especially in the early parts of ontogeny. An important plesiomorphic trait is the loose coiling, which is seen in several early Emsian ammonoid species. Our material of these very early forms led to the hypotheses that (1) intraspecific variability (at least of some characters) is intimately linked with coiling and therefore, (2) earlier forms are more variable than more derived forms with more tightly coiled shells.

To test the relation of the degree of intraspecific variability of conch parameters with coiling, we measured additional datasets of ammonoids from the late Emsian, the Eifelian, and the Carboniferous; the representatives of which have more or less tightly coiled shells. We used the coefficient of variation as a measure of the range of variability. Preliminary tests appear to support the two hypotheses, but further tests are needed before drawing a conclusion.

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## Shallow water facies setting around the Kačák Event – a multidisciplinary approach

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West of the river Rhine, the Eifel area (Eifel synclines) is the dominating structural unit interpreted as N-S trending axial depression of the Rheinisches Schiefergebirge (RSG). Siliciclastic material was delivered from the north during the Early Devonian and early Middle Devonian (Eifelian) but diminished during Givetian times when shallow subtropical carbonates were established over much of the region. STRUVE (1963) established a depositional model of the Eifel area with a N-S trending basin surrounded by landmasses, which he considered as the so-called "Eifel Sea Street". In contrast to this model, WINTER (in MEYER et al. 1977) defined three facies belts in the Eifel synclines and later FABER (1980) modified this model, based on detailed microfacies studies which gave evidence of rhythmic development of a carbonate platform during the early Eifelian and of a flat shelf lagoon during the late Eifelian and early Givetian, affecting the eastern part of the Eifel synclines. PAPROTH & STRUVE (1982) distinguished between N-, W-, and S-Eifel biofacies based on faunal differences. During the Givetian this facies differentiation broke down to some degree and mainly stromatoporoid/coral biostromes extended over the entire area.

The studied section lies within the Blankenheim syncline, between the villages of Blankenheim and Blankenheimerdorf and comprises shallow shelf mixed carbonate and siliciclastic facies of Middle Devonian age (Eifelian) accumulated on the southern margin of the former Avalonia microcontinent. A recent paper published by ERNST et al. (2011) is focussing on microfacies and bryozoan diversity. Based on the huge diversity of fossils found in deposits around the Kačák Event (e.g., HOUSE 1986) we will present results of a multidisciplinary approach in order to verify the hypothesis of a fundamental sedimentological gap ("Great Gap" sensu STRUVE 1982) in the late Eifelian (Junkerberg / Freilingen Fms).

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## **Multi-proxy stratigraphic analysis of the Devonian-Carboniferous boundary sections in the Central, Western and Southern Europe: a pathway to the better interregional correlations.**

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The end-Famennian Hangenberg Event was a series of the environmental perturbations recorded in marine and terrestrial realms on global scale (CAPLAN & BUSTIN 1999). Proposed as the "natural" Devonian-Carboniferous (D-C) system boundary (WALLISER 1986) the Hangenberg event provides an alternative solution of the D-C boundary. The D-C boundary GSSP is currently subject to extensive criticism due to the taxonomical, palaeoecological and taphonomical problems connected with the *Siphonodella praesulcata* – *Siphonodella sulcata* conodont lineage (PAPROTH et al. 1989, KAISER 2009, KAISER & CORRADINI 2011). Chemostratigraphic correlations based on the stable carbon isotopes were recently carried out between the D-C sections of the Carnic Alps, Graz Palaeozoic, Rhenish Slate Mountains, Montagne Noire, Colorado and Utah (BUGGISCH & JOACHIMSKI 2006, KAISER et al. 2006, KAISER et al. 2008, MYROW et al. 2011). The most distinct correlative feature is the positive peak or peaks of  $\delta^{13}\text{C}$ , which were documented from the *costatus-kockeli* Interregnum (CKI) and *kockeli* Zone in these regions. This shift is regarded as synchronous with the Hangenberg Black Shales, representing a horizon of enhanced organic carbon burial connected with the mass extinction event (CAPLAN & BUSTIN 1999, KAISER et al. 2006).

In this paper, we present a high-resolution, multi-proxy approach to the D-C boundary solution. Methods of the stable carbon isotope stratigraphy, field gamma-ray spectrometry (GRS) and mass specific magnetic susceptibility (MS) have been applied on the D-C sections in the Moravian Karst (Czech Republic), Carnic Alps (Austria), Dinant Synclinorium (Belgium, Northern France), Montagne Noire and Pyrenees (Southern France).

The variations in bulk carbonate  $\delta^{13}\text{C}$  were measured in the Lesní lom, Křtiny (Moravian Karst; KUMPAN et al. in press), Gendron-Celles and Avesnes (Dinant Synclinorium) sections. The  $\delta^{13}\text{C}$  excursion to values around 4 ‰ in the Lesní lom occurs in the CKI and is regarded as synchronous with the above mentioned peaks from (KAISER et al. 2006, 2008). In the Dinant sections, no positive  $\delta^{13}\text{C}$  peaks were documented in the Famennian, possibly due to the stratigraphic gap inferred from the foraminifer biostratigraphy. A sharp positive  $\delta^{13}\text{C}$  excursion at the boundary between the Comblain-au-Pont Fm (or Avesnelles Fm in Avesnes) and Hastière Fm is presumably controlled by facies shift at the base of the Tournaisian succession overlaying the wide stratigraphic gap.

A good correlation tool is the field gamma-ray spectrometry. Covariance between the dose rate (nGy.kg<sup>-1</sup>) and K, Th and U concentrations demonstrates that the gamma-ray signal is driven by Th and K in most of the studied sections. Concentrations of these terrigenous elements are presumably driven by the dilution effect of CaCO<sub>3</sub> (function of carbonate production and detrital input). Clay gamma-ray (CGR) values, used as improved clay volume indicators, were calculated from the spectral values. The redox sensitive Th/U ratio (ADAMS & WEAVER 1958) systematically decreases in the *praesulcata* Zone toward the Hangenberg black shales equivalents or Hangenberg positive  $\delta^{13}\text{C}$  excursion in all of the studied sections. The decrease of Th/U reflects a gradual lowering of the bottom oxygenation with minimum in the CKI coinciding with the black shales of the deep-water Kronhofgraben (Carnic Alps) and Puech de la

Suque (Montagne Noire) sections. These facies coincide with the sharp increase of CGR values. The gamma-ray and geochemical trends at the deep water Saubette section (Pyreneés) are presumably affected by dolomitization, but a sharp peak occurs in the shales of the CKI. In the deep water sections where no black shales occur (Lesní lom, Moravian Karts; Grüne Schneid, Carnic Alps) a sharp increase in CGR takes place above the  $\delta^{13}\text{C}$  peak, just below (upper part of CKI or the *kockeli* Zone) or at the D-C boundary. This is interpreted as slow-down of carbonate production after the hypoxic Hangenberg event phase and/or as accelerated supply of the fine-grained terrigenous material during a sea-level fall (equivalent of the Hangenberg Sandstones). The gamma-ray trends at shallower water succession at the D-C boundary GSSP La Serre are similar to the above-mentioned trends. Higher values are associated with the shales at the base of the CKI and with argillaceous limestones in the upper part of the CKI and *kockeli* Zone. The carbonate ramp successions of the Dinant Synclinorium (Gendron-Celles, Rivage and Avesnes sections) are also characterised by decreasing Th/U ratios from the ?Upper *expansa/praesulcata* Zone (BOUCKAERT & GROESSENS 1976) toward the D-C boundary. Other trends are missing due to the stratigraphic gap.

MS signal decreases in most of studied sections (except Kronhofgraben, Mokra and Saubette) below the Hangenberg Black Shales. Growing MS signal in the CKI to *sulcata* Zone underlines a global regressive character of sedimentation near the D-C boundary.

Our results support the views of WALLISER (1984) who regarded the Hangenberg Event as worldwide synchronous and a natural D-C boundary. Good interregional correlation of the petrophysical and geochemical proxies suggests that they are driven by global oscillations. At the same time, their temporal resolution is higher than that of standard biostratigraphy. The proxies therefore represent a useful tool for chronostratigraphic correlations, which can contribute to the calibration of the conodont zones. Consequently, the Hangenberg event should be addressed in the new biostratigraphic definition of the D-C boundary which is currently under discussion.

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**(P) Conodont and foraminiferal biostratigraphy of the Late Famennian and Early Tournaisian in the Moravian Karst (Moravo–Silesian Zone, Czech Republic)**

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A standard definition of the Devonian-Carboniferous (D-C) system boundary is based on the FAD of conodont *Siphonodella sulcata* in the lineage from *Si. praesulcata*, applicable especially for deeper water facies. The recent studies of conodonts at the D–C boundary recognized serious problems in this lineage both in the GSSP at La Serre and other sections (KAISER 2009, TRAGELEHN 2010). Variety of problems shows also an alternative index to the siphonodellids, the protognathodids (CORRADINI et al. 2011).

Sections in the Moravian Karst (Lesní lom, Mokrá and Křtiny quarries) represent a halfgraben sedimentation where the D-C boundary can be identified in different facies of the Líšeň Fm, ranging from the nodular hemipelagic limestones (Křtiny Lms) of the upper slope of the basin to calciturbidite lower slope facies (Hády-Říčka Lms). The importance of the joint presence of foraminifers and conodonts in calciturbidites is underlined by the manifestation of the Hangenberg event horizon with typical positive  $\delta^{13}\text{C}$  excursion (KUMPAN et al. in press) which makes from the sections in the southern part of the Moravian Karst one of the key areas for the comprehensive biostratigraphic analysis of the D-C boundary beds.

The conodont succession from the Moravian Karst reveals good correlation with other areas (zonation sensu JI 1986 and KAISER et al. 2009 have been applied). All of the seven groups of early siphonodellids (genus *Eosiphonodella* sensu JI, 1986), designed by KAISER AND CORRADINI (2011) have been easily determined from the Moravian Karst. Distribution of this morphologic groups shows an important stratigraphic pattern here, and is slightly different from the chart published by KAISER & CORRADINI (2011). Group 1 ("unusual" *Si. sulcata*) enters before the appearance of Groups 2 and 3 ("typical" *Si. praesulcata* morphotypes). It resembles Uppermost Famennian conodont successions from the Franconia, where, however, *Si. sulcata* and *Si. praesulcata* Morphotypes enters simultaneously (TRAGELEHN 2010) and marking the base of the *praesulcata* Zone. Groups 4, 5 and 7 (the later is the "typical" *Si. sulcata* morphotype) enter and mark the base of the Tournaisian together with the upper protognathodid fauna. The protognathodid fauna is relatively abundant in Lesní lom and partially in the Mokrá quarries, where the FAD of *Protognathodus* species is in accordance with their known global stratigraphic range which is globally rare feature (CORRADINI et al. 2011). In the Křtiny quarry, on the other hand, protognathodid fauna is missing, most probably due to the stratigraphic gap.

The calciturbidite successions with more shallow water allochems yielded so far richest foraminiferal associations at the D-C boundary where the evolution of foraminiferal fauna is calibrated by conodont zonation. The evolution and extinction of the foraminiferal genus *Quasiendothyra* across the D-C boundary represents an important guide which application is, however, problematic because of different interpretations of species ranges. While in Belgium (POTY et al. 2006) and China (HANCE et al. 2011) *Quasiendothyra* is regarded as typical late Famennian guide, in Moravia (Czech Republic) it was reported to range up to the *bransoni* Zone (former Lower *duplicata* Zone; KALVODA & KUKAL 1987) and lower Tournaisian occurrences have been suggested also in the Urals (KOCHEKOVÁ et al. 1987, NEMIROVSKAYA et al. 1993, PAZUKHIN et al. 2009). During recent research at the Lesní lom quarry was *Quasiendothyra* founded even in the *hassi* Zone (former Upper *duplicata* Zone).

The D-C boundary interval is characterized by the FAD of *Tournayellina pseudobeata* close below the D-C boundary and by a sequence of bioevents where beside LADs of quasiendothyrs FADs of

*Neoseptaglomospiranella* and chernyshinellids (see Fig. 1) play an important role similarly as in the Eastern Europe (KULAGINA et al. 2003, PAZUKHIN et al. 2009). The correlation of the recognized bioevents elsewhere is often hindered by glaciostatically driven unconformities and widespread occurrences of unfavourable facies for plurilocular foraminifers (Malevka Beds, *Bisphaera* Beds).

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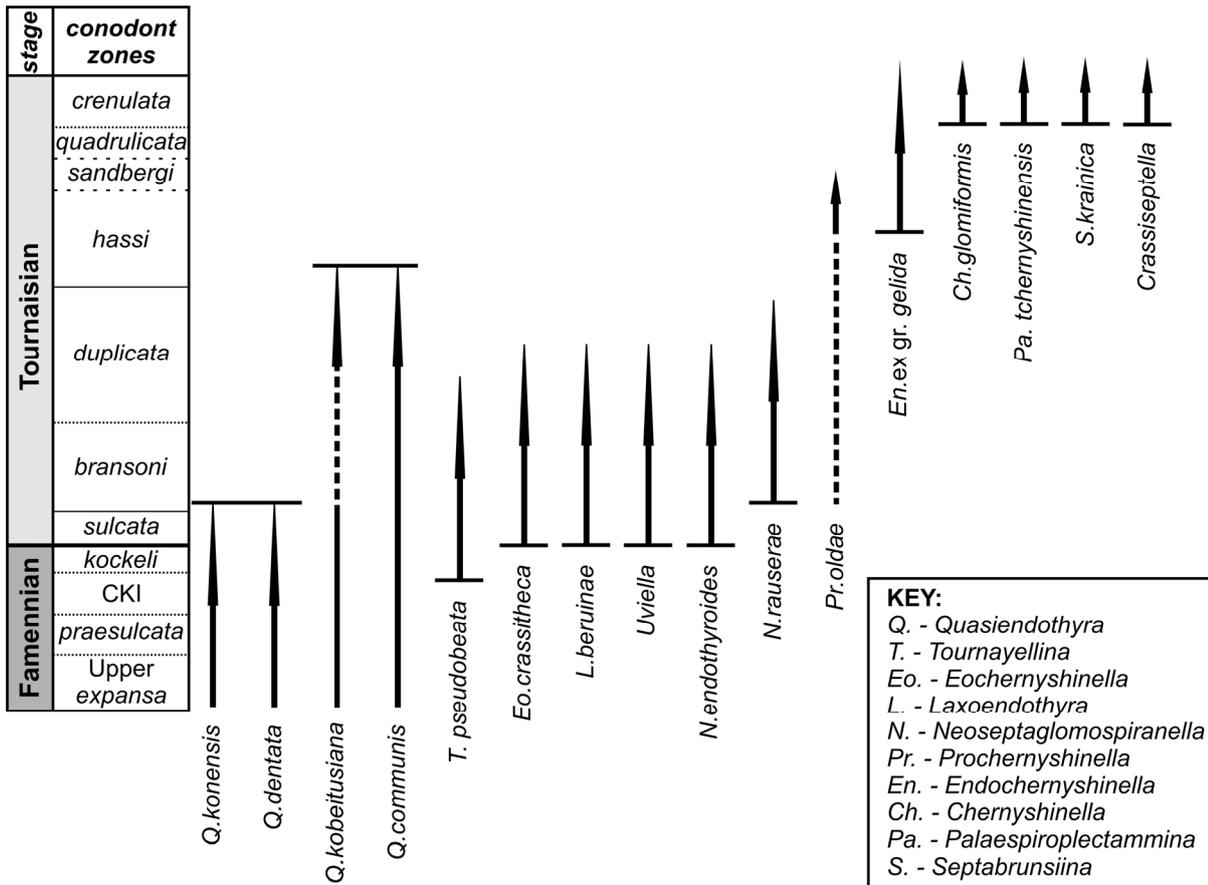


Fig. 1. Stratigraphic range of foraminifers in the Moravian Karst (conodont zones sensu Ji 1986 and KAISER et al. 2009).

**(P) Crinoids of the Silurian - Devonian boundary beds of Transbaikal****KURILENKO, A.V.**

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For many years it has been assumed that the lower boundary of the Devonian in Transbaikal and the Far East is not paleontologically proven. The direct superposition of Lower Devonian sediments over Silurian strata has been discovered in the east of Transbaikal – the Upper Amur region. According to MODZALEVSKAYA (1958), some gradual transition can be observed between them. G.R. SHISHKINA (TURBIN 1994) thought that there was a stratigraphic break in sedimentation, which was explained by the lack of Lower Lochkovian fauna in the area. According to the Amur regional stratigraphic scheme, the boundary of the Omutnaya and Bolshoi Never Suites corresponds to the boundary of the Silurian and Devonian (TURBIN 1994). As a result of the author's work some new data on the boundary sediments within the eastern Transbaikal region (the basin of the Amazar River, the Uteni River) was obtained (KURILENKO et al. 2002, KURILENKO & KULKOV 2008).

The upper part of the Omutnaya Suite in this zone has accumulations of big Rhipidomellidae – *Platyorthis? mugurensis* VLADIMIRSKAYA and can be dated as Pridolian (KURILENKO et al. 2002).

The lower part of the Bolshoi Never Suite is characterized by stem fragments of the crinoid *Scyphocrinites mariannae* YAKOVLEV, which has a short vertical range and which occurs at the base of the Ainasu Horizon in Kazakhstan (STUKALINA 1991). Outside the Transbaikal region, the genus *Scyphocrinites* is well known from the boundary layers of the Skala and Borschov (Taina Beds) horizons in Podolia (the East European platform), which are dated by the occurrence of the graptolite *Monograptus uniformis agustidens* PŘIBYL. *Scyphocrinites* is identified from the Silurian - Devonian boundary deposits of the Urals. They are ubiquitous in the Silurian - Devonian boundary beds of the USA, Morocco, Malaysia, but they are most abundant in the upper part of the Pridolian in the Czech Republic (Bohemia), its analogues in Poland, Bulgaria, Germany (Rhineland, Kellerwald, Thuringia), France (Pyrenees, Cabrieres, Artois), Spain (Pyrenees), Great Britain (Cornwall area), Algeria, and Tunisia (STUKALINA 1977, 1991, WITZKE et al. 1979). In G.A. STUKALINA'S opinion (STUKALINA 1991), the boundary strata of the Silurian and the Devonian contains some plankton representatives of *Scyphocrinites*; they can be considered as a part of the Silurian, as in many regions (Podolia, Kazakhstan and others) they are accompanied by benthic crinoids of the Silurian.

Some higher section parts of the Bolshoi Never Suite in the zone of the Uteni River have crinoids: *Costatocrinus bicostatus* (STUKALINA), *Asperocrinus echinatus* (YELTYSCHEWA), *Anthinocrinus radialis* STUKALINA, *Gurjevskocrinus impalpabilis* DUBATOLOVA, *Gregariocrinus forus* (STUKALINA), *Facetocrinus stellatus* (YELTYSCHEWA & SISOVA), *Tastjicrinus paucicostatus* (YELTYSCHEWA), and the brachiopods: *Dalejina austera* HAVLIČEK, *Plectodontamimica* (BARRANDE), and *Lissatrypa* sp.

Almost all the crinoid species found in Transbaikal (except *Gurjevskocrinus impalpabilis*) are known from the upper part of the Ainasu and Kokbaital horizons of Central and Eastern Kazakhstan (STUKALINA 1991). The occurrence of *Costatocrinus bicostatus* correlates with the Kunzhak and Shishkat horizons of the Zeravshano-Gissarskaya mountainous region of the Southern Tian-Shan. *Anthinocrinus radialis* spreads in the Mitkov Beds of the Borschov Horizon of the East European platform. *Gurjevskocrinus impalpabilis* and *Costatocrinus bicostatus* occur in the Tom'chumysh Horizon of the Salair (STUKALINA 1986, 1991, DUBATOLOVA 1968, 1971). The age of this part of the section can be defined by crinoids as Early - Middle Lochkovian. In N.P. KULKOV'S opinion the brachiopods studied from this location most likely speak in favor of the Lochkovian (KURILENKO et al. 2002, KURILENKO & KULKOV 2008).

Another section with analogous crinoids (of the lower part of the Bolshoi Never Suite) has been found within the eastern Transbaikal region (KURILENKO et al. 2001). In the Makarov Unit (on the right bank of the upper Onon River stream) they are determined at two stratigraphic levels: the lower one contains abundant imprints of discrete remains belonging to *Scyphocrinites mariannae* YAKOVLEV and *Mediocrinus* aff. *medius* YELTYSHEWA; the upper level has *Scyphocrinites mariannae* YAKOVLEV, *Mediocrinus medius* YELTYSHEWA, *Costatocrinus bicostatus* (STUKALINA), *Anthinocrinus radialis* STUKALINA, *Tastjicrinus paucicostatus* (YELTYSHEWA), and *Tolenicrinus lenticularis* (STUKALINA).

In addition to the above mentioned species, *Mediocrinus medius* is known from the Tom'chumysh Horizon of the Salair, Ainasu and Kokbaital horizons of Central and Eastern Kazakhstan. *Tolenicrinus lenticularis* (STUKALINA) is known from the upper part of the Tokrausky, Ainasu and Kokbaital horizons of Kazakhstan. The beds distinguished in Transbaikal and the Sarajnaya and Sauma horizons of the Urals are also considered coeval by the occurrence of the genera *Costatocrinus* and *Mediocrinus*. This also demonstrates an Early – Middle Lochkovian age.

Thus, Silurian and Devonian transitional sediments and faunal complexes have been discovered on the territory of the Transbaikal region by the author. This fact made it possible to record the transitional sediments in the lower part of the Bolshoi Never Horizon (the lower strata of the Bolshoi Never Suite in the Upper Amur region and the Makarov Unit on the right bank of the Onon River).

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## **(P) Strunian Glaciation and the Devonian-Carboniferous boundary in the high palaeolatitude record from Bolivia**

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The Late Devonian to Early Carboniferous was a time of major climatic changes; marking the transition from a Mid-Devonian Greenhouse world, to a cooler global-climate punctuated by glaciations. The transition was marked by the Strunian Glaciation, a short and sharp glaciation in the latest Famennian. This was broadly coincident with eustatic sea-level changes and End Devonian Mass Extinction (EDME). Strunian diamictites can be found throughout central South America and even in the mid-palaeolatitude record of the USA (CAPUTO 2008, ISAACSON et al. 2008, BREZINSKI et al. 2008). These diamictites have been dated palynologically within the *Retispora lepidophyta-Indotriradites explanatus* (LE) and *R. lepidophyta-Verrucosisporites nitidus* (LN) palynozones of the latest Famennian (CAPUTO 2008).

Although coincident with significant changes, the Strunian glaciation has received little previous attention. Few detailed sedimentary logs have been published for the high palaeolatitude, and these have limited associated biostratigraphic data. The present study aimed to rectify this situation by using a combined sedimentological and biostratigraphic approach to document, in detail, the Late Devonian and Early Carboniferous in the high palaeolatitude. The Altiplano of Bolivia was selected as a study area based on excellent Late Devonian exposures.

Three sections were measured, Villa Molino, Chaguaya and Hinchaka, and represent deposition within a shallow, wave-dominated marine shelf. The diamictites are thin and are not associated with significant down-cutting. At Villa Molino and Chaguaya, the glacial-diamictites are interpreted to have been deposited in a shallow glacio-marine environment with occasional ice-grounding. The diamictites at Hinchaka are interpreted as relatively distal marine claystones with exotic dropstones. A 120 m thick marine claystone succession rests above the termination of the diamictites at Villa Molino and Chaguaya. Initial biostratigraphic analysis of the basal 15 m of this claystone has shown the presence of *R. lepidophyta* and *V. nitidus*, indicating an LN age and, therefore, a pre D-C boundary termination to the glaciation.

Future work will be to compare the near-field glacial history to that of far-field record of extinctions and eustatic sea-level. In addition, this study will also be used as a comparison to the terrestrial climate record from the palaeo-southern arid zone NE Greenland.

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## (P) The Givetian Conodont subdivision in the Spanish Central Pyrenees and its global correlation

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The intensive worldwide study of Givetian rocks and fossils for more than 30 years led to the establishment of a stable conodont zonation that can be globally applied. In fact, this detailed zonation formed the base for the Givetian three-fold subdivision proposal that was formally voted by the International Subcommittee on Devonian Stratigraphy in 2007. Detailed studies on the Givetian of the Pyrenees started in 2001 when the first comprehensive taxonomic work on the Renanué section was presented (LIAO et al. 2001). Since then, several tons of rocks and many stratigraphical levels have been sampled searching for conodonts. This study has demonstrated that the conodont record in the Pyrenean sections is rich, varied and of global relevance, and therefore, the Pyrenees is an important region in the context of Givetian conodont studies on a zonal and even smaller subdivisions. The purposes of this work are to present the current knowledge on the Givetian conodont sequence from the Spanish Central Pyrenees and to correlate this sequence with the standard one.

The Pyrenean data presented herein come from five selected sections that belong to three different Pyrenean Subfacies areas (VALENZUELA-RÍOS & LIAO 2006): The Renanué section from the Subfacies Renanué; the Ampriú section from the Subfacies Sierra Negra and the Compte, Villech and La Guardia d'Ares sections from the Subfacies Compte.

The Eifelian/Givetian boundary is defined by the entry of the index conodont *Polygnathushemiansatus* at the base of the Bed 123 in the section Jebel Merch Irdane in Tafilalt, Morocco. The boundary is identified by the sequential entry of the taxa in sections Renanué and La Guardia d'Ares. In the section Villech, the index taxa has not been found, but the boundary has to lie in an undefined interval of about 2 ms between conodonts of the *kockelianus-ensensis* Zone below and the Lower *varcus* Zone above.

The *timorensis* Zone is identified in Renanué with the entry of the name-given taxa *P. timorensis*. In other sections (Compte, La Guardia d'Ares, Villech) there is a set of strata below the overlain *rhenanus/varcus* Zone that can be attributed to the *timorensis* Zone.

The *rhenanus/varcus* Zone is identified in the four afore-mentioned sections by the entry of the indexes *P. rhenanus* and/or *P. varcus*.

The *ansatus* Zone is recognized in the four sections by the entry of the nominal taxon *P. ansatus*.

The *semialternans/latifossatus* Zone is recognized in Villech by the lowest record of "*Oz.*" *semialternans*, in Compte by the joint entry of both indexes and in Renanué and La Guardia d'Ares by the conodont record above and below the supposed interval.

The *hermanni* Zone is recognized in all sections. The sequential entry of *P. cristatus cristatus* or *P. c. ectypus* permits identification of lower and upper *hermanni* subzones in Renanué, Compte and La Guardia d'Ares. In Ampriú and Villech only the *hermanni* Zone can be recognized.

The *disparilis* Zone is identified by the lowest occurrence of the nominal taxa in Compte, La Guardia d'Ares and Villech. Its further subdivision into lower and upper, which is based on the entry of *P. dengleri* is also identified in Compte, La Guardia d'Ares and Villech.

The uppermost Givetian *norrissi* Zone is defined by the lowest entry of *Skeletognathus norrisi* and has been identified in Renanué, Compte, La Guardia d'Ares and Villech.

In brief, the set of Pyrenean sections allows identification and recognition of all global standard conodont zones improving the knowledge of this global succession and testing its application in the

Pyrenean region. This, in turn, supports the intended subdivision of the Givetian into three substages based on conodont occurrences.

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## **(P) Late Devonian sequences and biostratigraphy of the South Marhouma region, Beni Abbès, SW Algeria - Preliminary results.**

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The Late Devonian of the Saoura region around Beni Abbès, SW Algeria, is famous since the detailed work on the regionally rich cephalopod faunas by G. PETTER (1959, 1960). These allowed a detailed biostratigraphic subdivision of the Famennian, whereas the Frasnian, which is comparatively much poorer in macrofossils, remained poorly studied. The goal of current research in the frame of a PhD Thesis by A. MAHBOUBI is to establish the conodont biozonation in comparison with sedimentological features and with the bathymetric evolution of the depositional environment. First investigations focussed on the Frasnian of the S Marhouma section, situated some 30 km SE of Beni Abbès.

After the recognition of conodont and goniatite bearing strata of the basal Late Devonian (GÖDDERTZ 1987), Frasnian deposits are analysed for the first time in fine detail for the conodont biozonation, goniatite and trilobite faunas, microfacies, and sequence stratigraphy. The completely exposed 57 m thick succession is organized into 15 to 16 median sequences (probably fourth order sequences). The lower part of the succession is composed of black shales with limestone nodules alternating with several prominent limestone beds with goniatites and phacopid trilobites. Above, the main succession is composed of stylioline-rich, thin-bedded, often laminated pelitic marlstones and rare siltstones yielding numerous levels with beige-pink irregular nodular limestones. Their homogenous micritic texture without clastic input, the scarcity of benthic biotas, bioclasts and the absence of bioturbation and current features indicate rather calm depositional conditions on a distal, subsiding outer platform. In the upper part of the section, episodic basinal conditions prevail, when black shales and silts are intercalated, the topmost of which is the biostratigraphically recognized equivalent of the Upper Kellwasser Horizon. The conodont content is rather poor: only half of the so far 84 dissolved samples yielded a mean amount of 16 conodonts per kg. Six biozones are currently recognized: Upper *falsiovalis*, *transitans*, *punctata*, upper *hassi*, *jamieae* and Upper *rhenana*. The equivalent of the Kellwasser horizon is a 1.2 metre thick black shale deposit. It is sandwiched between grey-reddish, silty marlstones, yielding goniatites, small chonetid brachiopods, bivalves and *Homoctenus tenuicinctus*, and black siltstones yielding limestone septarians with *Palmatolepis triangularis*. It is expected that ongoing work will reveal that the Frasnian and the Famennian conodont biozonations are complete, allowing Marhouma S to constitute a Late Devonian reference section for the entire Saoura region.

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## Terrestrial Climate and Ecosystem Change from the Devonian-Carboniferous boundary to the earliest Viséan interval in East Greenland

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There is significant change in the terrestrial biota at the Devonian-Carboniferous boundary. This occurs in the fish, the early tetrapods, and the land plants as revealed by their spores. This terrestrial record is particularly informative through linking the changes in the terrestrial biota with the climate system, as revealed by the climatically sensitive terrestrial sediments. It is now clear that Devonian-Carboniferous (D-C) boundary marks a profound realignment of the evolutionary trajectory of some groups. This is exemplified by the early tetrapods. The tetrapods reach the D-C boundary as essentially fish with fingers, and with these having variable digits on different taxa. The tetrapods then disappear from the fossil records for some 5 my, an interval known as Romer's Gap. When the tetrapods reappear in the Viséan, they have become both terrestrialized and with a uniform five digits on both manus and pes. The terrestrial sediments of East Greenland contain an integrated record of the palynological and palaeoclimatic record. Importantly these sediments continue into the Early Carboniferous (VIGRAN et al. 1999), through Romer's Gap, and provide a key to understanding this crucial interval. This account is based on new fieldwork from the 2012 season that focused on the Early Carboniferous interval with sections studied from Celsius Bjerg (Gauss Halvø), Rebild Bakker (Traill Ø) and Backlund Ridge (Geographical Society Ø). These are sections that contain the terrestrial D-C boundary and three of these have substantial sections of preserved Early Carboniferous rocks.

The integrated spore and palaeoclimate record from East Greenland shows that the plants survived the cool arid time of the Strunian glaciation to then provide, initially high diversity palynofloras as the climate changed to warm and wet conditions. But as the sedimentary system flooded to form a megalake, with coincident margins, the plants suffered significant habitat loss, ultimately becoming extinct and never returning to the system as the lake dried out. This is quite surprising as the Strunian palynofloras contain a number of very persistent and abundant elements (as exemplified by *Retispora lepidophyta*) which achieved a global distribution and were clearly highly successful and able to survive in many different climate zones.

The fossil record of trees across the D-C boundary is not well understood but it is becoming clear (DECOMBEIX et al. 2011) that major groups such as *Archaeopteris* did not survive to cross the boundary. This collapse of the forest environment is shown here by a quantitative count of fossil plant stems across the D-C boundary from Celsius Bjerg. Substantial sized plant stems are abundant up to the boundary and then immediately disappear at the boundary. This disappearance is coincident with that of fossil fish such as the holoptychiids. Both plants and fish are then absent through the earliest Carboniferous. Large trees only reappear, and then as a different plant group- the lycopsids, with the appearance of warm wet conditions in the Viséan. The spore floras show associated changes and particularly the demise of spores that were produced by the Rhacophyton dominated understorey vegetation to these Famennian forests.

The style of sedimentation also changes immediately above the D-C boundary. A system that was river channels with significant overbank deposits then becomes replaced by a stack of river channels with little preserved overbank. These channels are significantly thicker, faster flowing with higher discharges and contain substantial bedload although without any visible plant debris. The discharge can be high enough to form small Gilbert-type deltas and ephemeral lakes. However, there is an arid season as shown by the ubiquitous presence of calcrete clasts. This change is regarded as representing a time of ecosystem collapse

with the removal of both the forest trees and understorey vegetation with the fluvial system reverting to a more arid and ephemeral character typical of the early Devonian. It is ecosystem collapse that is seen as driving extinctions and environmental change with ROMER'S Gap.

The initial palynofloras above the Gap are striking in being very impoverished and containing only a few common taxa that are generally simple spores. Higher up section the palynological floras recover in this wetter early Tournaisian interval and includes the appearance of lycopod megaspores. However, the character of the sediment then changes to a sustained interval of aridity. This interval contains multiple evidences for arid conditions including the presence of thick vertisol palaeosols that represent a low seasonal contrast within an arid climate. There are no palynomorphs preserved through this interval. The peak of this aridity is tentatively correlated with the Tournaisian glaciation as represented in South Africa by the Miller Diamictite with the post-glacial sea-level rise being equivalent to the Soutkloof Shale and in Western Europe the mid Tournaisian Alaunschiefer.

Following this arid interval the climate becomes wetter with the appearance of immature sandstones followed by thin coals and black shales. These sediments contain abundant *Lycospora* and are accompanied by tree-sized lycopod trunks and are Viséan in age. The interval is now more typically of Coal Measures character and on Backlund Ridge continues (VIGRAN et al. 1999) into the Namurian.

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## (P) Lochkovian conodont biostratigraphy in the South Urals

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Carbonate successions of the early Lower Devonian (Lochkovian–lower Pragian) from the Mindigulovo section (West Zilair Zone, western slope of the South Urals) can be globally correlated using diverse conodont faunas. The number of the documented widespread or cosmopolitan taxa of the Lochkovian and lower Pragian enables the definition of globally recognized conodont successions and the establishment of a refined regional conodont biozonation. The Lochkovian in the South Urals is represented by two Horizons with different facies development: 1. The Siyak Horizon of the lower Lochkovian is characterized by the dominance of shallow-water organogenic limestones of the *hesperius-optima* and *optima-omoalpha* Zones, and 2. The Sherlubai Horizon of the middle and upper Lochkovian is characterized by the dominance of carbonate-terrigenous successions. The highest diversity of conodont faunas can be seen in deeper-water strata of the middle Lochkovian; it provides the best correlation on the basis of the *Lanea* and *Ancyrodelloides* lineages. The presence of many cosmopolitan taxa and recorded single occurrences of representatives of widespread lineages (e.g., *Pelekysgnathus* and *Pedavis*) are promising for a future detailed study and prospective refinement of the regional biozonation. The abundance of conodonts in shallow-water reefal carbonate rocks is very low but the correlation of the Lochkovian–Pragian boundary interval is relatively precise. The present three-fold global subdivision of the Lochkovian Stage can be applied as well to the studied area. The middle and upper Lochkovian is subdivided into five regional biozones (using a binominal system): *omoalpha-oeleanorae*, *oeleanorae-eleanorae*, *eleanorae-trigonicus*, *trigonicus-pandora* beta and *pandora* beta–*sulcatus* eta/*steinachensis* beta. The Lochkovian–Pragian boundary and the lower Pragian time are determined by the presence of eognathodontid and icriodontid stocks. The established regional conodont zonal scale in the South Urals provides a solid basis for a refined correlation with the Lochkovian–lower Pragian carbonate rocks of peri-Gondwana and Euramerica.

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## (P) „Living fossils“ in the Devonian: micromorphic organophosphatic brachiopods

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Organophosphatic brachiopods are rare invertebrates in recent seas. However, they were typical members of the Cambrian Evolutionary Fauna. In the Cambrian, different distinct clades did exist: linguloids, acrotheloids, acrotretoids, siphonotretoids and paterinoids and brachiopod affinity of some bizarre organophosphatic fossils was widely discussed in last years (HOLMER et al. 2008).

Having the acme in the Tremadocian and Floian, a slow decline of organophosphatic brachiopods proceeded in the Ordovician as a response to growing abundance and diversity of bivalves and expansion of carbonatic deposits. Only discinoids clade appeared later in the Ordovician and became firstly important in the Silurian and Devonian. The Ordovician glaciation terminated or almost terminated some of organophosphatic clades: paterinoids extincted and acrotretides and siphonotretids almost disappeared. In the Silurian, Devonian and Carboniferous, linguloids and discinoids became abundant in siliciclastic shallow shelves and this environment occupy to the Recent.

The response to environmental changes in the beginning of Silurian, acrotretoids and siphonotretoids followed two trends – the miniaturization of shell size and occupation of the hypoxic environment. Some but not all discinoids follow the similar trends.

In the Devonian, only *Opsiconidion* and *Havlicekion* are cosmopolitan acrotretoid taxa. They have very small shell size, thin and highly conical shell. This micromorphic acrotretoids disappeared in the Middle Devonian. Siphonotretoids, originally the moderately sized taxa, were assumed extinct in end of the Ordovician. However, a micromorphic *Orbaspina* is present in the offshore limestones in up to latest Lower Devonian in Bohemian and Australia (MERGL 2001, VALENTINE&BROCK 2003). *Paterula*, the genus of Paterulidae (Linguloidea), is unique. Its first occurrence is of the early Ordovician date, it crossed the Ordovician glaciation having the latest occurrence in the lowest Middle Devonian. *Paterula* show a weak morphological change over 80 Ma It was always an epibenthic taxon in a deeper offshore, associated with black shale lithofacies shoreward to graptolite shales. These genera represent long existing “living fossils”, which lived next several tenths of Ma after extinction of other taxa of their clade. Micromorphic adaptation of other organophosphatic clades is known also in the Middle Devonian; an obolid *Microbolus* and a trematid *Opatrilkiella* are good examples.

Miniature size and long existence of these clades can be explained by their adaptation to hypoxic marine environment. Small size and thin shell is a good adaptation to hypoxic condition. Hypoxic condition was in a deeper environment but also inside a sediment of shallow water. Small cavities inside coarse sediment, e.g. crinoidal sands, are good sites for extremely small micromorphic brachiopods. Filtering a nanoplankton just above bottom-water interface in deeper waters and inside the sedimentary cavities is a good strategy how withstand the bivalve competition.

Run away to hostile deeper environment is common evolutionary strategy in many invertebrate groups. This trend is known even in the Recent: *Pelagodiscus* having thin and small shell is a typical deep-ocean discinid.

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## **Contribution of the Famennian plants from Anti-Atlas (Morocco) to the analysis of plant-climate interactions during the Devonian**

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Terrestrial landscapes and the global environment changed substantially as land plants diversified, increased in size and progressively colonized all habitats in lowlands during the Devonian. In a large number of geochemical models, the interplay between vegetation, global climate and carbon cycle is mediated by the roots (BERNER 2004). Root systems are expected to have been deeper as the plants gained height in the Devonian. This evolutionary process accelerated the continental weathering, contributing to the drawdown of the atmospheric CO<sub>2</sub> and ultimately changing the global climate. In the GEOCLIM-Slave Jr model presented by LE HIR et al. (2011), weathering remains a key process in climate-carbon feedbacks. In this model, however, climate-vegetation feedbacks are driven by the geographical distribution of the major plant types which impacted the albedo, thermal properties, roughness and potential evaporation of the continental surfaces. For this type of model, a good understanding of the morphology and ecological requirements of the major types of Devonian plants, together with a correct assessment of the structure of past plant communities was necessary to find the best modern analogs allowing to parameterize the model.

Several Famennian marine deposits from Anti-Atlas have yielded well preserved plant remains. Most are assignable to the Cladoxylopsida, a class of fern-like plants, and to the archaeopteridalean progymnosperms, the sister-group of the seed plants (MEYER-BERTHAUD et al. 1997). Both groups evolved the tree habit in the Middle Devonian and are therefore expected to have impacted the global climate and the carbon cycle during the Devonian. During the past 12 years, the permineralized remains of trunks, branches and roots collected in deposits from Mader and Tafilalt have substantially improved our understanding of the morphologies and growth patterns in these two groups of plants. The cladoxylopid root system consists of small roots that are exclusively adventitious (i.e. borne on stems), branch rarely and form a globose mass surrounding the stem base (SORIA & MEYER-BERTHAUD 2004). In contrast, the archaeopterid roots which measure up to 90 mm in diameter have a similar structure to that of extant gymnosperms. They have abundant wood and display complex branching patterns. They show an unprecedented potential for underground exploration and exploitation of water and nutrients in the Devonian (MEYER-BERTHAUD et al. in press). In addition, whereas cladoxylopsids are made of a large volume of living cells in all phases of their life, archaeopterids depends on the production of wood, a tissue made largely of dead elements, to grow and enlarge (SORIA & MEYER-BERTHAUD 2005, MEYER-BERTHAUD et al. 2010). This information allowed predicting the climate requirements (temperature and water availability of soils) of these two groups of plants. These data were used in the GEOCLIM-Slave Jr model to study the plant/ climate /carbon interactions during the Devonian.

Simulations obtained with that model show how the vegetation contributed to the decrease in the atmospheric CO<sub>2</sub> by increasing the runoff and accelerating the continental weathering. However, these simulations also predict that the reduction of the continental albedo due to vegetational changes resulted in a decoupling between CO<sub>2</sub> and the surface temperatures which may have remained unchanged through most of the Devonian (LE HIR et al. 2011).

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**(P) Upper Devonian and Lower Carboniferous of Western Transbaikalia****MININA, O.R.<sup>1</sup>, NEBERIKUTINA, L.N.<sup>2</sup> & ARISTOV, V.A.<sup>3</sup>**

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The Baikal folded area (BFA) is considered as a part of the Central Asian Belt. The Baikal-Vitim folded system has been outlined by Ruzhentsev et al. (2012). It includes the Vitimkan-Tsipka, Uda-Vitim, and Turkish-Kurba structural-formational zones. Lower and Middle Devonian sediments are carbonatic, reefogenic and terrigenous-carbonate rocks.

Flyshoid greywacke sediments of the Tocherskaya, Muhtunnaya, Amatskanskaya, and Syryhsкая Formations, of the Ognenskaya «Series» (of the Vitimkan-Tsipinska Zone), the Panofskaya and Zumburuskaya Formations (of the Turka-Kurbinskaya Zone), the Himgildinskaya Formation and the Ulzutuyskaya «series» (of the Uda-Vitim Zone) belong to the Upper Devonian – Lower Carboniferous.

The age of the sediments is determined by miospore assemblages (Aristov et al. 2005, Minina et al. 2011, 2012). Conodonts and miospores allow to delimit the accumulation time of these sediments as Famennian to Tournaisian. Conodonts are only found in the Tocherskaya and Panofskaya Formations, miospores are found in all stratigraphic units.

The Tocherskaya Formation is composed of sandstones and polymict sandstones, siltstones, tuff-siltstones interbedded with conglomerates, grits, and silty limestones (lower subsuite). There are the conodonts *Palmatolepis* cf. *triangularis* Sannemann., *Polygnathus* sp., "*Ozarkodina*" sp., "*Ligonodina*" sp. (D3fm), *Pa. perlobata schindewolfi* Müller, *Pa.* cf. *marginifera* Helms, and *Polygnathus glaber* Ulrich & Bassler in limestone. (D1-2 fm), and the miospores *Geminospora basilaris* (Naum.) Pashk., *G. rugosa* (Naum.) Obukh., *Auroraspora varia* (Naum.) Ahmet, *Arreticulisporeta retiformis* (Naum.) Obukh., *Kedoesporites imperfectus* (Naum.) Obukh. etc. (D3-C1), also from limestone. The miospores *Tumulisporeta rarituberculata* (Luber) Pot., *Grandisporeta famenensis* (Naum.) Streel, *Trachytriletes solidus* Naum. (D3fm-C1.t) etc. are found in the middle subformation (sandstones interbedded with siltstones, phyllites, and bituminous limestone).

The conodonts *Pseudopolygnathus triangulus* Voges (C1t) and *Neopolygnathus communis* Branson & Mehl (D3 fm2-C1), and the miospores *Leiotriletes ornatus* Isch., *Dictyotriletes similis* Kedo, *Cyclogranisporites punctulatus* (Waltz) Luber, *Cymbosporites acutus* (Kedo) Byv., *Spelaeotriletes microgranulatus* Byv. var. *minor* Byv., *Euryzonotriletes tersus* (Waltz) Isch., *Hymenozonotriletes ugulatus* Jusch. etc. (C1.) are found in the upper subformation of the Tocherskaya Formation (interbedded with sandstones, siltstones, aleurolites, sandy and clayey limestone, clay and black carbonaceous shale).

The Panofskaya Formation (lower subformation) is composed of limestones interbedded with aleurolites. It includes the conodonts *Ancyrodella binodosa* Uyeno, *Mesotaxis* cf. *falsiovalis* Sandberg, Ziegler & Bultynck, *Icriodus* sp., *Polygnathus* sp., *Ancyrodella* ex gr. *nodosa* Ulrich & Bassler, etc. (D2zv-D3f). The conodonts *Palmatolepis* sp., *Pa. superlobata* Branson and *Icriodus* sp. (D3fm1), and the miospores *Brochotriletes faveolatus* Naum., *Verrucosporites grumosus* (Naum.) Sull., *Archaeozonotriletes formosus* Naum., *Kedomonoletes glaber* (Kedo) Oshurk. etc. (D3fm) are found in the upper subformation (phyllites interbedded with detrital limestone)

The Zumburuskaya Formation (siltstone, clay-siliceous shales with lenses of limestone) includes the miospores *Lycosporeta pusilla* (Ibr.) Sw & B., *Tetraporina prima* Naum., *Euryzonotriletes tersus* (Waltz) Isch., *Dictyotriletes similis* Kedo, *Hymenozonotriletes ugulatus* Jusch., *Spelaeotriletes microgranulatus* Byv. var. *minor* Byv., etc. (C1t-v).

The Muhtunnaya Formation (interbedded sandstones, sandstones, siltstones, detrital limestone) includes the miospores *Lophozonotriletes bellus*KEDO, *Lophotriletes inflatus* (LUBER) NAUM. *Dictyotriletes rotundatus*NAUM., *Verrucosisporites mesagrumosus* (KEDO) BYV., *Auroraspora rugosiuscula* (JUSCH.) BYV., etc. (C1t).

The Ognenskaya « Series » is composed of interbedded carbonaceous, silty limestone, and calcareous siltstone. It includes the following Early Carboniferous miospores: *Hymenozonotriletes pusillus* (JBR.) JSCH., *Diatomozonotriletes albus*KEDO, *Spinospirites carbonicus*BYV., *Remispora perforatus*BYV., *Trematozonotriletes irregularis* (ANDR) JSCH., *Chaetosphaerites pollenisimilis* (HER.) RUT., and *Grumosisorites cerebrum*BYV., etc.

The upper subformation of the Syryhskaya Formation predominantly contains the miospores *Leiotriletes subparvus*ISCH., *Lophotriletes subverrucosus*ISCH., *Archaeozonotriletes macroreticulata* (NAUM.) KEDO, *Hymenozonotriletes conformis*NAUM., *H. auranthicus*NAUM., *Dictyotriletes graniformis*KEDO, *Densosporites undatus*BYV. & N.UNM. etc. (C1t). The miospores *Trachytriletes parvus* (IBR.) ISCH., *Lophotriletes paululus*ISCH., *Leiotriletes pennatus* (ISCH.) KEDO, *Brachytriletes gyratum*ISCH., *Zonomonoletes stenomarginatus*ISCH., *Dictyotriletes commalatus*ISCH., *Monoptica unica*ISCH. etc., most characteristic for the upper part of the Lower Carboniferous, dominate the upper subformation of the Syryhskaya Formation.

The miospores *Verrucosisporites macroreticulatus* (KEDO) BYV., *V. phaleratus*(LUBER) LUB., *Grumosisorites cerebrum*(BYV.) BYV., *Archaeozonotriletes macrospinus* var *punctatus*KEDO., *A. attenuata* (BYV.) BYV., *Retusotriletes septalis*ISCH. var *concesus*ISCH., *Hymenozonotriletes explicatus* KEDO., *H. undulatus*ISCH., *H. subtenelus*ISCH., *H. mentitus*ISCH., *Eurizonotriletes trivialis*KEDO & ISCH., *H. subgranulatus*KEDO, *Acanthotriletes squamosus*KEDO (C1t) are prevail in the Amatkanskaya Formation.

The miospores *Brochotriletes faveolatus*NAUM. var. *minor*NAUM., *Acanthotriletes denticulatus*NAUM., *Hymenozonotriletes mancus*NAUM., *Archaeozonotriletes angulosus* (NAUM.) NEKR., *H. multigulatus* KEDO (D3 fm) are found in the Himgildinskaya Formation (upper subformation). The miospores *Cyclogranisporites aureus* (LOOS.) POT. & KR., *Psilohymena mirabilis* (LUB.) HART., *Laevigatisporites vulgaris* (IBR.) ALP. & DOUB, *Tuberculispora paulula* (ISCH.) OSHURK. (C1-2) are found in the lower subformation of the Himgildinskaya Formation.

The miospores *Simozonotriletes intirtus* (WALTZ) ISCH. var *trigonus*KEDO, *Punctatisporites platyrugosus* (WALTZ) SULLIV. var. *giganteus*WALTZ, *Chomotriletes concentricus* (BYV.) OSHURK., *Vallatisporites genuinus* (IUSCH.) BYV., *Hymenozonotriletes minimus*KEDO, *Verrucosisporites mesogrumosus* (KEDO) BYV., *Diaphanospora submirabilis* (KEDO.) BYV. (C1) are found in the Ulzutuyskaya « Series ».

The Panofskaya and Muhtunnaya Formations, and the Ognenskaya « Series » include lenses and layers of organo-clastic limestones and calcarenites, with mixed assemblage of Cambrian - Devonian organic residues.

Thus, at the end of the Upper Devon there was a change of sedimentary environments in the region. Carbonate and terrigenous-carbonate sediments of a stable shelf were replaced by terrigenous (greywacke) series of the Upper Devonian – Lower Carbonian. At that time a large deflection of greywacke was formed on the southern periphery of the shelf in the rear suprasubduction volcanic zone.

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## (P) Preliminary data on plicathyridines (Brachiopoda) from the Frasnian of southern Belgium and the Middle East (Afghanistan, Iran)

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### 1. Introduction

Plicathyridines are distinctive athyridide brachiopods, which are restricted to the Devonian (Lockhovian–Frasnian), although TAN (1986) described two species assigned to *Anathyris* from the Lower Carboniferous of China (Hunan). According to ALVAREZ & RONG (2002), the generic assignment of TAN (1986)'s species is uncertain. After a period of decline during the Middle Devonian, mainly in the course of the Givetian, the plicathyridine brachiopods display great abundance and diversification during the Frasnian on the Russian Platform and in the Altai-Sayan province (see ALVAREZ 1990 and RZHONSNITSKAYA & MODZALEVSKAYA 1996 for references), where many endemic species have been recognized, till their disappearance during the late Frasnian biological crisis (GRUNT & RACKI 1998, ALVAREZ 2003, ALVAREZ & MODZALEVSKAYA 2001). Nevertheless, plicathyridines are also known from Frasnian-aged rocks of Western Europe (southern Belgium) and the Middle East (central Iran and eastern Afghanistan). They only occur sporadically in these three regions where they constitute a minor part of the Frasnian brachiopod assemblages. Due to the small number of specimens available, only preliminary results are presented herein.

### 2. Frasnian plicathyridines from southern Belgium

Plicathyridines were reported on several occasions within the Devonian succession of southern Belgium (e.g. MAILLIEUX 1941), but they have never been illustrated, except some poorly preserved specimens from the Pépinster Formation (late Eifelian–early Givetian), which were tentatively assigned to *Hexarhytis ferronesensis* (DE VERNEUIL & D'ARCHIAC) by KAYSER (1895, pl. 3, figs. 5–6). However, they badly need to be revised as KAYSER (1895)'s identification was rejected by ASSELBERGHS (1923) who considered that these specimens have nothing in common with DE VERNEUIL and D'ARCHIAC'S species from the Pragian of the Cantabrian Mountains (Spain).

In southern Belgium, Frasnian plicathyridines are known from the south and southeastern borders of the Dinant Synclinorium (Fig. 1A). The latter correspond to the distal part of the Namur–Dinant Basin which recorded the development of carbonate mud mound levels separated by argillaceous episodes. Plicathyridines were quoted in the former stratigraphic units "F2b" and "F2e" by MAILLIEUX (1941), which now correspond to the shaly La Prée (Nismes Formation) and Ermitage (Moulin Liénaux Formation) members, respectively (Fig. 1B). On the basis of the available data, the range of *Anathyris* (in terms of conodont zones) in southern Belgium spans the interval of the *Palmatolepis falsiovalis* to the *P. punctata* zones (Fig. 1B). Up to now, plicathyridines are only known by few specimens, which are part of old collections housed at the University of Liège (ULg) and the Royal Belgian Institute of natural Sciences (IRScNB).

Two Frasnian species, belonging to the *phalaena–helmerseni* groups as defined by RZHONSNITSKAYA & MODZALEVSKAYA (1996), are provisionally identified as *Anathyris* (*A.*) sp. A [= *A. helmerseni* (von BUCH) in MAILLIEUX (1941)] (Figs 1B, 2.1–2.3) and *A.* (*A.*) sp. B (= *Athyris* (*Terebratula*) *hispanica* DE VERNEUIL in DEWALQUE 1881, Fig. 2.4) until larger collections are obtained.

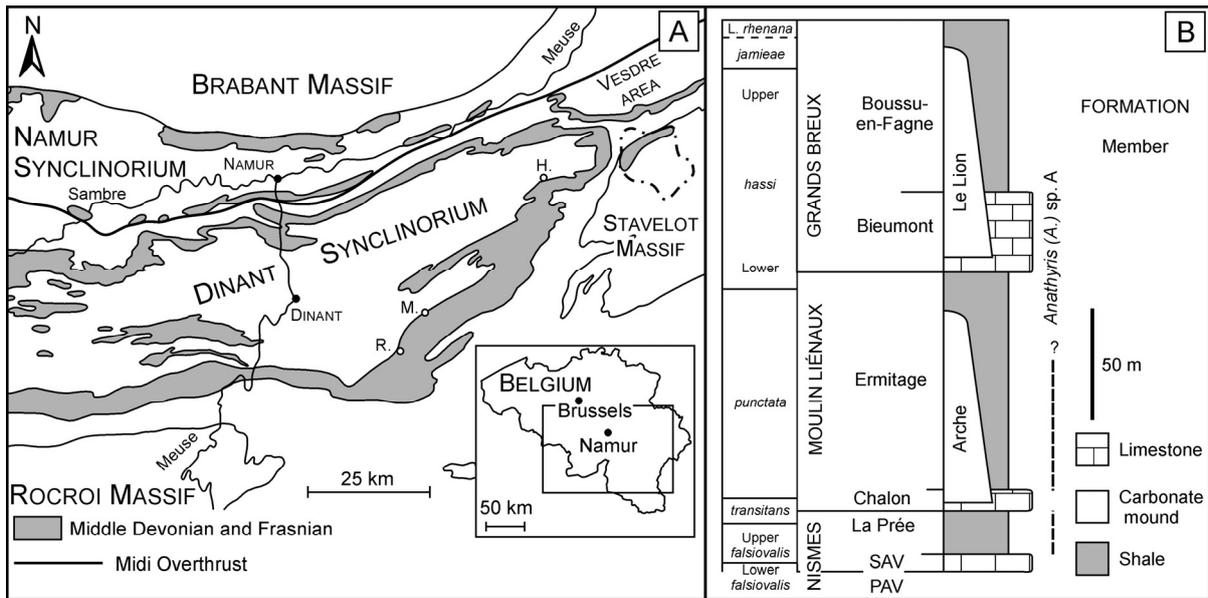
### 3. Frasnian plicathyridines from the Middle East

In Iran, some plicathyridines (Figs. 2.5–2.7) were recovered from a section located approximately 3 km west of the village of Chahriseh in the Kuh-e Kaftar mountains, ca. 55 km northeast of Esfahan (Esfahan Province) by BRICE et al. (in MISTIAEN et al. 2000). There, they occur in the upper part of the second biostromal horizon, which is developed in the upper carbonate unit of the Shistu Formation and apparently located just below the *Palmatolepis jamieae* conodont Zone (MISTIAEN et al. 2000). This particular horizon consists of several metres of marly limestone rich in phaceloid and massive rugose corals, associated with abundant lamellar tabulates. From the brachiopod viewpoint, they are associated with rhynchonellides and spiriferides. These few poorly preserved specimens, deposited in the Faculté Libre des Sciences & Technologies de Lille (France) and catalogued with the prefix GFCL, may be assigned to *Plicathyris* on the basis of their external morphology (e.g., small size, lateral folds more or less developed), but further material is required to reach a better identification.

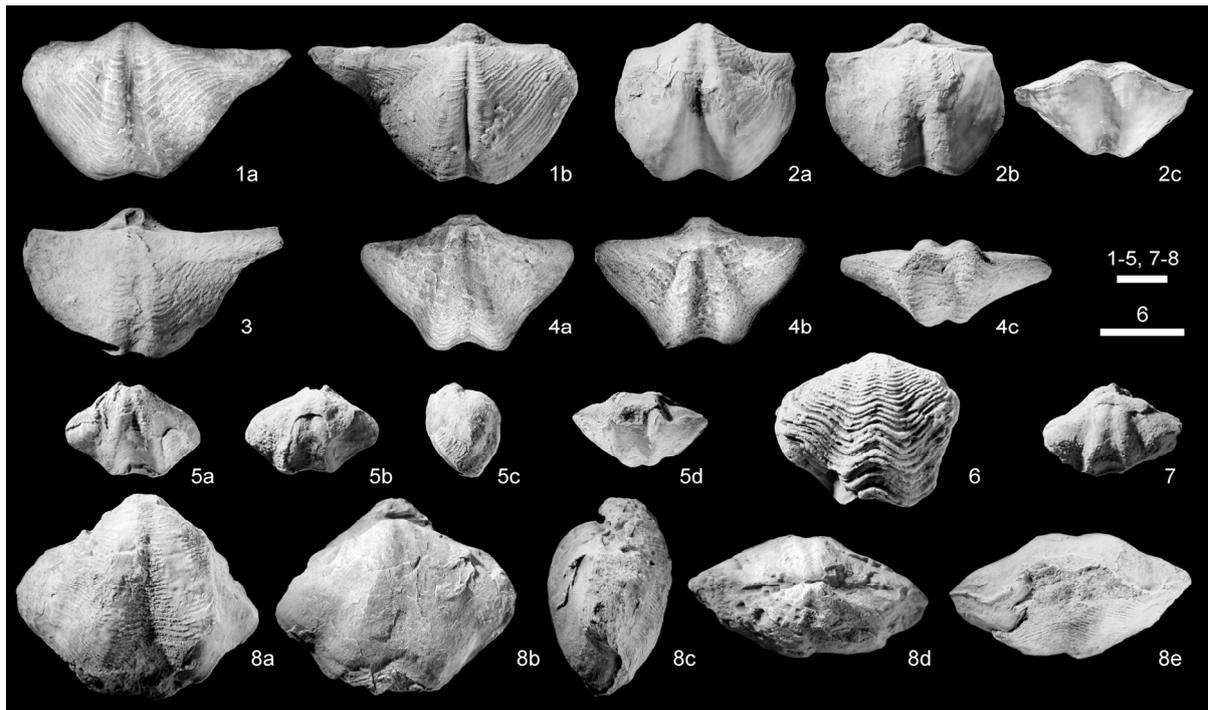
Frasnian plicathyridines from Afghanistan come from the Dacht-e-Nawar area, southwest of Kabul, where three localities (Bokan, Doni Yarchi and Saïd Habib) have yielded some badly preserved specimens (BRICE 1971). They were previously assigned to *Anathyris* e.g. *helterseni* (von BUCH) by BRICE (1971) (Fig. 2.8). In the Dacht-e-Nawar area, their range in terms of conodont zones is unknown. However, they are associated with an unidentified species of *Cariniferella*, which probably belongs to the *C. tioga* group. These orthides may indicate the *P. punctata* conodont Zone in comparison with their occurrence in the Frasnian succession of the Boulonnais (northern France), but further research are needed.

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**Fig. 1.** A. Schematic geological map of southern Belgium. Abbreviations for the fossiliferous localities: H., Hamoir; M., Marche-en-Famenne; R., Rochefort. B. Frasnian lithostratigraphic units of the southeastern border of the Dinant Synclinorium and distribution of *Anathyris (A.) sp. A.*



**Fig. 2.** 1–3. *Anathyris (A.) sp. A.*: 1a–b, incomplete specimen (IRScNB Invert 09340-0014) in ventral and dorsal views, Moulin Liénaux Formation, Marche-en-Famenne (Belgium); 2a–c, incomplete specimen (IRScNB Invert-05408-0001) in ventral, dorsal and anterior views, Nismes Formation, Rochefort (Belgium); 3, almost complete specimen (IRScNB Invert-03031-0025) in dorsal view, Nismes Formation, Rochefort (Belgium). 4a–c, *Anathyris (A.) sp. B.*, complete specimen (ULg 2012-12-21) in ventral, dorsal and anterior views, precise stratigraphic level unknown (Nismes or Moulin Liénaux Formation), Hamoir (Belgium). 5–7. *Plicathyris?* sp. from the Shistu Formation (central Iran, Chahriseh area), 5a–d, complete specimen (GFCL 2745) in ventral, dorsal, lateral, and anterior views; 6, incomplete specimen (GFCL 2746) in dorsal view; 7, incomplete specimen (GFCL 2748) in dorsal view. 8a–e, *Anathyris* eg. *helterseni* (von BUCH), southern Afghanistan (Dacht-e-Nawar area, Bokan), specimen GFCL 3559 in ventral, dorsal, lateral, posterior and anterior views. Scale bars are 5 mm.

## **(P) Revision of the conodont species *Icriodus orri* KLAPPER & BARRICK and its significance for the Eifelian biostratigraphy**

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*Icriodus orri* was established by KLAPPER & BARRICK (1983) to include elements differing from *I. retrodepressus* by an exceptionally wide posterior basal cavity. According to the original concept the species embraces two forms that can have either a relatively short and wide spindle or a narrower and longer spindle. In some specimens the lateral-row denticles can be isolated, in others they are connected with the middle-row denticles forming characteristic narrow transversal ridges. The transversal denticle rows are notably more separated in the anterior part of the spindle than in the posterior part. Within this apparently morphologically diverse taxon the authors distinguished two morphotypes. The Morphotype I displays a distinct depression in the posterior part of the spindle where the denticles of the middle row, if present, are lower than others. In the Morphotype II, to which the holotype was attributed, the posterior depression is absent, and the middle-row denticles are at the same level than lateral ones, and can be linked to the lateral-row denticles.

A closer study of the forms originally attributed to *I. orri* reveals that the morphology of the P<sub>1</sub> elements of the species is morphologically disparate. Size and outline of the basal cavity are variable. In the Morphotype I the posterior expansion of the cavity is smaller than in the Morphotype II, and more symmetrical with a less pronounced and generally rounded spur, while in the Morphotype II it is asymmetrical, with a sharply outlined spur on the inner side of a basal cavity. The spur is located considerably below the abruptly widening margin of the outer basal cavity in the anterior part of the spindle. Both morphotypes were compared to the conodonts obtained by the authors from the intracratonic Michigan (USA) and Belarussian (Europe) basins, and from literature data. The American material was derived from the Dundee and Rogers City formations and from the lowermost part of the Bell Shale, while the Belarussian conodonts are from the lower part of the Kostyukovichi Horizon of the Pripyat Graben area. For this comparative analysis we selected only the specimens that clearly possess or lack the median posterior depression. Within our broadly concept of *I. orri* we were able to distinguish typical representatives similar to the holotype, and two new species: one with a narrow elongated spindle, and another similar to the Morphotype I sensu KLAPPER & BARRICK (1983). The three taxa differ in the spindle outline, number and shape of denticles, and presence/absence of the median depression in the posterior part of the spindle.

*I. orri* (Figs. 1-5) is characterized by an elongated, wide spindle with convex margins, widest in the posterior part. The spindle axis is almost straight and only its anterior extremity is slightly bent inwards. It is narrow and pointed, whereas the posterior end is wide. The denticles of the middle row are connected with the lateral-row denticles forming 7-8 transversal, clearly separated denticle rows showing an almost straight alignment. The lateral-row denticles are elongated transversally, with sharp upper margins and are markedly bigger than the rounded middle-row denticles. The latter are linked by a longitudinal ridge, particularly well-visible in the posterior part of a spindle. All denticles are of the same height. The basal cavity is asymmetrical and wide with a clearly marked spur.

*Icriodus* sp. A (Figs. 6-10). Representatives of this species have elongated narrow spindle with an axis that is distinctly bent inwards. The outer margin is convex, while an inner one is slightly concave. Both extremities are pointed but the posterior one is slightly wider. The middle row is composed of 6 rounded, isolated and slightly anteriorly displaced denticles. They are smaller than those from lateral rows with which they can be linked. In the lateral rows there are 8 denticles, with an oval outline, tending to display

sharp upper margins. At least the three posteriormost transversal denticlerows are shevron-like. All denticles are of the same height.

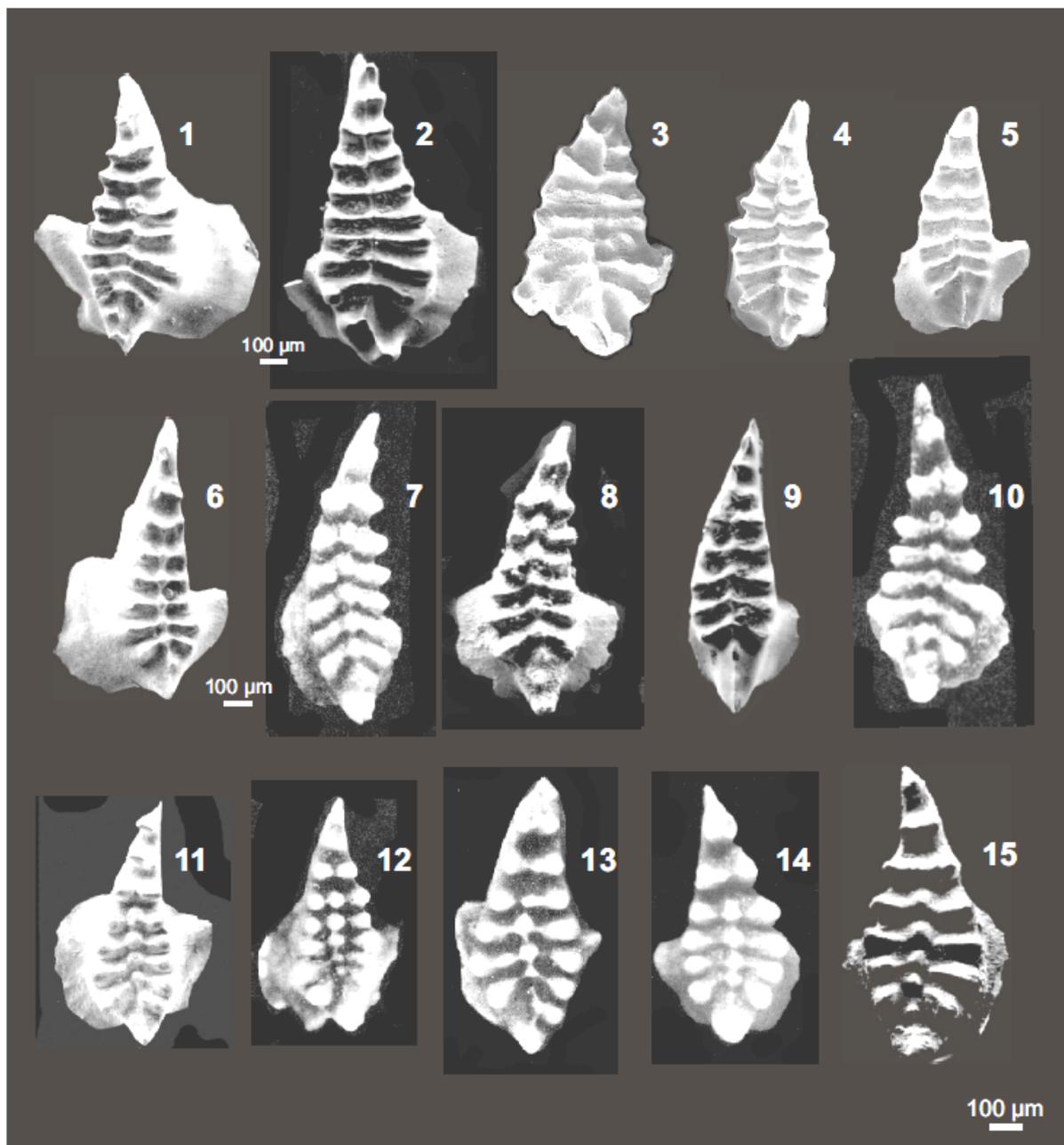
*Icriodus* sp. B (Figs. 11-17) is a stout form with a relatively short and wide spindle. Its posterior part is wide with convex margins and a straight axis, and ends with a short, wide, pointed or slightly rounded extension of the spindle. The anterior part is elongated and gradually tapers towards the pointed anterior extremity which is bent inwards. The middle row consists of 4-5 rounded isolated denticles, smaller than those from lateral rows that are aligned with the median denticle row. The middle row denticles are absent from the anteriormost, narrow part of the spindle. The two last median-row denticles in the posterior part of the spindle are located in a distinct depression, and are lower than the other denticles. The lateral rows are composed of 6-7 oval, or, less frequently, rounded denticles. In the wider part of a spindle these denticles are much bigger and very close to each other.

Representatives of *I. orri* and *Icriodus* sp. A appear later than *Icriodus* sp. B. In Iowa *I. orri* occurs in the uppermost *kockelianus* and lower part of the *ensensis* zones (KLAPPER & BARRICK 1983). In the Michigan Basin it appeared probably near the boundary between the *kockelianus* and *ensensis* zones (BRETT et al. 2011), whereas in Belarus – in the *ensensis* Zone (NARKIEWICZ & KRUCZEK 2009). In Iowa and in the Michigan Basin *Icriodus* sp. A was found in the same strata as *I. orri*, whereas in Wisconsin it is known from deposits assigned to the lower part of the *ensensis* Zone (DESANTIS & BRETT 2011). In Iowa *Icriodus* sp. B was found probably in the uppermost part of the *kockelianus* Zone (KLAPPER & BARRICK, 1983). The specimens from Ohio are from the *kockelianus* Zone (DESANTIS & BRETT 2011), and those from Ontario (BRETT et al. 2011) and Michigan Basin from the *costatus* Zone.

The above review suggests that *I. orri* and *Icriodus* sp. A appear in the uppermost part of the *kockelianus* Zone and range into the *ensensis* Zone. On the other hand, *Icriodus* sp. B. appears in the *costatus* Zone and ranges up to the *kockelianus* Zone, where it co-occurs with representatives of *I. orri* and *Icriodus* sp. A.

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**Figs. 1-5.** *Icriodus orri* KLAPPER & BARRICK; **1** – reproduced from KLAPPER & BARRICK (1983, fig. 9AF, holotype); **2** – Michigan, lowermost Bell Shale; **3-4** – Belarus, Pińsk 54 well, Kostyukovichi Horizon; **5** – Belarus, Żytkowicze 2 well, Kostyukovichi Horizon.

**Figs. 6-10.** *Icriodus* sp. A; **6** – reproduced from KLAPPER & BARRICK 1983, fig. 9AD (= *Icriodus orri*); **7** – reproduced from SCHUMACHER 1971, pl. 9, fig. 14 (= *Icriodus nodosus* HUDDLE); **8** – Michigan, lowermost Bell Shale; **9** – Michigan, top of Rogers City Fm.; **10** – reproduced from SCHUMACHER 1971, pl. 9, fig. 18 (= *Icriodus nodosus* HUDDLE).

**Figs. 11-15.** *Icriodus* sp. B; **11** – reproduced from KLAPPER & BARRICK 1983, fig. 9X (= *Icriodus orri*, Morphotype I); **12** – reproduced from SCHUMACHER 1971, pl. 9, fig. 2 (= *Icriodus nodosus* HUDDLE); **13** – reproduced from UYENO et al. 1982, pl. 3, fig. 16 (= *Icriodus* aff. *I. retrodepressus* BULTYNCK); **14** – reproduced from SPARLING 1983, pl. fig. 13AS (= *Icriodus* sp. aff. *I. retrodepressus* BULTYNCK); **15** – Michigan, Dundee Fm.

The scale bar shown in the lower left corner applies to all specimens except those shown in Figs. 1, 2, 6, 7

## (P) Eifelian ammonoids from the Central Kyzylkumy Desert (Uzbekistan)

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No Eifelian ammonoids have been described from the Kyzylkumy desert, although they have been briefly mentioned in association with conodonts in several geological reports and publications (BOGOSLOVSKY 1982, CHERKASHENKO et al, 1989). These associations come from the Izaskaya Formation of the Aristantau Mountains (Central Kyzylkumy, Uzbekistan), where ammonoids are found along with conodonts and tentaculites. The ammonoid collection no. PIN 5440 is housed in the Paleontological Institute, Russian Academy of Sciences. The Izaskaya Formation is composed of a thick series of bedded dark-grey and black carbonates and calcareous breccia alternating with dolomites and fissile silty limestones with cherty lenses and nodules. The Izaskaya Formation (over 700 m thick) contains four members. The lower member is composed of an alternation of limestones, cherts and laminated calcareous siltstones with tentaculites. The second member is composed of an alternation of carbonates and siltstones, with subdominant beds of gradational breccia 1.5 to 7 m thick, with fragments 1 to 40 cm in size. The third member is composed of dolomites interbedded with limestone, calcareous siltstone, and cherty beds. Fossils include brachiopods and conodonts. The fourth member is composed of micritic limestones alternating with calcareous siltstone with subdominant cherts. The fourth member contains abundant fossils, which include brachiopods, ammonoids, tentaculites and conodonts, dating the assemblage relative to the conodont-based Emsian-Eifelian boundary. The Aristantau assemblage contains *Polygnathus costatus costatus*, *P. costatus partitus*, *P. costatus patulus* (lower *costatus* Zone). Beds immediately below the ammonoid-bearing horizon contain the tentaculites *Nowakia sulcata*.

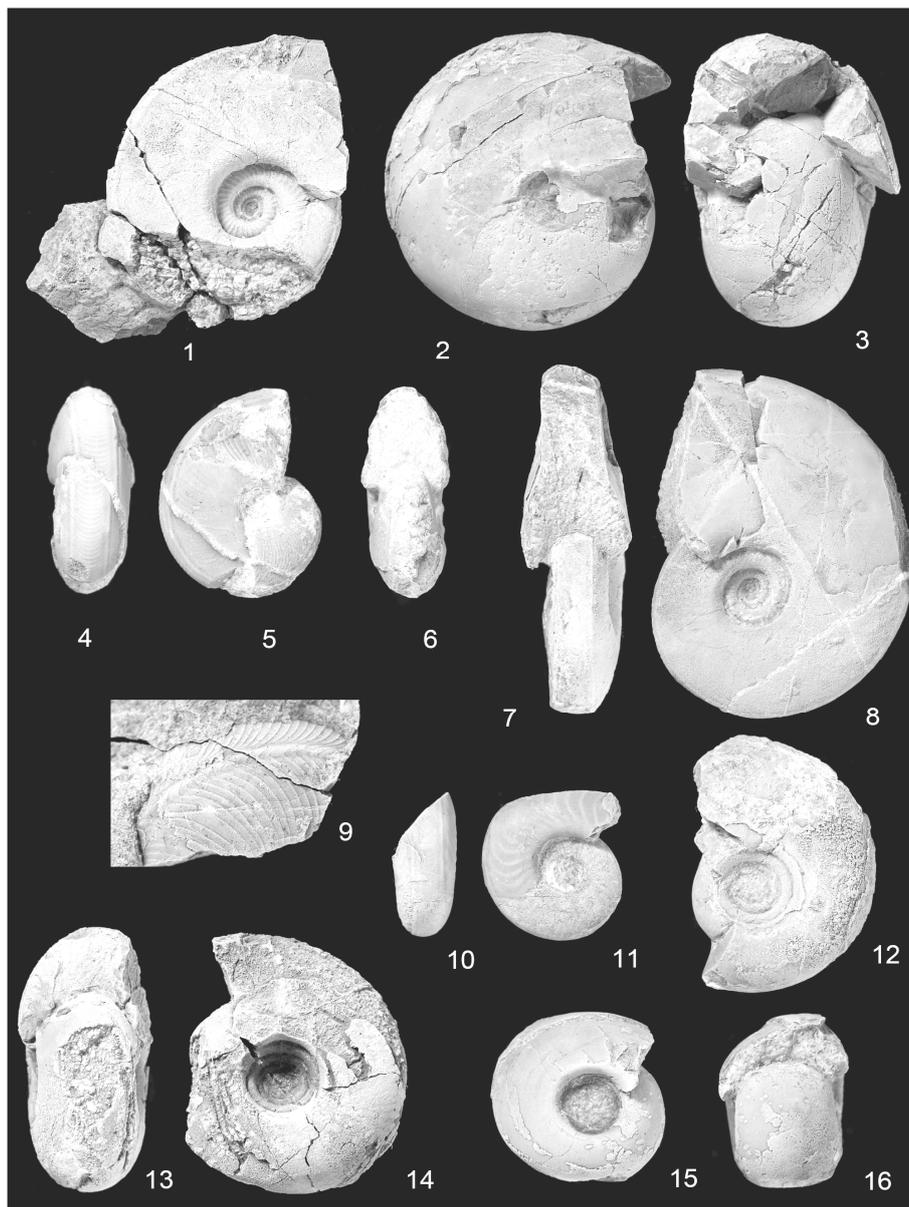
The ammonoids come from several outcrops of the middle part of the fourth member as mass accumulations of moulds and fragments of moulds. Many shells are recrystallized, with shell matrix rarely preserved. The assemblage contains *Fidelites occultus*, *Fidelites fidelis*, *Fidelites cf. clariondi*, *Fidelites cf. verna*, *Subanarcestes macrocephalus*, *Subanarcestes* sp. nov., *Cabrieroceras* sp., etc. This assemblage correlates with that from the Lower Eifelian *Subanarcestes macrocephalus* Zone in Morocco (MD-1D) (BECKER & HOUSE, 1994, 2000, KLUG 2002, EBBIGHAUSEN et al. 2011 and others), with the assemblage from the red carbonates near the village of Pokrovskoe, Bobrovka River, eastern slope of the Middle Urals (BOGOSLOVSKY, 1969) and with the assemblage from the second member of the Afonino Formation, Elets River (TSYGANKO 2008). Faunas of similar age have also been recorded from the Saoura Valley, Algeria (PETTER 1959, GÖDDERTZ 1987), in the Cantabrian Mountains, Spain (Montesinos, 1987, 1988), in the Armorican Massif (BABIN 1989). *Fidelites occultus* and *Subanarcestes cf. macrocephalus* have been recorded in Guangxi, China (the latter was originally identified as *Werneroceras* sp.) (BAI et al. 1982, BECKER & HOUSE 1994). The assemblage from Aristantau differs from most of the above assemblages in the absence of the genus *Pinacites*, which in many other regions enters at the very base of the Eifelian (MD-1C Zone = *P. jugleri* Zone) and continues into the *S. macrocephalus* Zone. In some regions, e.g., in Bohemia and Tafilalt, the higher horizons contain *Exopinacites*.

Like in Germany and Bohemia, the depositional conditions in the Aristantau basin in the early Eifelian became more marine, with a considerable increase in pelagic faunal elements, with ammonoids and tentaculites becoming more abundant. This increase, which probably corresponded to the approaching Choteč event, marked worldwide by the deposition of black shale in the *Pinacites jugleri* and *Subanarcestes macrocephalus* zones. However, only the latter zone can so far be established in Aristantau. The absence of Pinacitidae in the Kyzylkumy, in an otherwise diverse assemblage, is difficult to explain. It

is possible that the family Pinacitidae was on the whole poorly represented in the Devonian Turkestan Ocean. Interestingly, in the Aristantau section, the appearance of *Cabrieroceras* is recorded in the *S. macrocephalus* Zone, as in some Moroccan sections (KLUG 2002).

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**Plate 1**

All specimens are from Aristantau, Central Kyzylkumy Desert, Uzbekistan. **Figs. 1, 7-9.** *Fidelites occultus*: (1) specimen PIN 5440/13, x 1, sample 63/1; (7-8) specimen PIN 5440/30, x 1, sample 63/1; (9) specimen PIN 5440/5, x 3.5, sample 59. **Figs. 2-3.** *Subanarcestes macrocephalus*, specimen PIN 5440/1, x 1, sample 59. **Figs. 4-6.** *Fidelites cf. verna*, specimen PIN 5440/3, x 4.6, sample 59. **Figs. 10-11.** *Fidelites cf. occultus*, specimen PIN 5440/18, x 1.2; sample 63/1. **Fig. 12.** *Cabrieroceras* sp., specimen PIN 5440/31, x 1, sample 63/1. **Figs. 13-16.** *Subanarcestes* sp. nov. (13-14) specimen PIN 5440/11, x 1, sample 63/1; (15-16) specimen PIN 5440/14, x 1.5, sample 63/1.

## **(P) Facies development and death of the Devonian Wülfrath Reef (northern Rhenish Massif, Germany)**

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Reefs flourished during the Middle and lower Upper Devonian (Frasnian) more than during any other parts of Earth history. The Rhenish Massif, a segment of the northern, external Variscides, contains numerous Devonian reef complexes. Limestones of the Wülfrath Reef Complex crop out on the specially folded, NE plunging Velbert Anticline in the northern Bergisches Land. It was situated on the flat, distal shelf south of the Old Red Continent and falls, unlike many other reefs to the south, completely into the Frasnian. The studied area is in the northernmost part of the outcrop area (see KARRENBERG 1954), in the eastern part of the Rohdenhaus-North Quarry, located north of the city of Wülfrath. Here, the (primarily southern) rock face was examined, as well as an adjacent drill core (Well Ro9913). The well's maximal depth was about 166 m, of which the lower 110 m were analysed, including the upper boundary of the reefal limestone at about 59 m depth. Bedding planes show an inclination of about 40-60° relative to the drilling direction. No single bed could be correlated between the well core and the quarry or between different parts of the quarry. 14 microfacies types (MF) were defined and conodonts were sampled from core and outcrop.

The lower and major part of the succession represents a restricted back-reef (lagoonal) environment. The fauna is poor except for stromatoporoids, which appear in various forms and in high numbers. Dendroid stromatoporoids of the genera *Amphipora* and *Stachyodes* are generally dominant (MF 3, Fig. 1). Tabular and massive growth forms are also abundant in some layers, constituting bind- and framestones and forming small patch reefs (MF 4, Fig. 2). Other faunal elements present are gastropods, ostracods, foraminiferans, tabulate corals (mostly *Thamnopora*), and unidentified shell fragments. Peloids constitute the main component and intraclasts of various sizes are frequent, especially in layers with less abundant macrofossils (MF 1). Tempestites are common. The water was shallow to very shallow, with a few intertidal intervals (MF 2). This phase represents the Dorp Magnafacies (e.g., KREBS 1974), i.e. a developmental stage of a reef complex with a discrete division into fore-reef, reef core and back-reef.

The back-reef sediments are capped by an erosional disconformity (Fig. 3), above which there is no longer evidence of a reef core acting as a barrier. The sediments directly under- and overlying this surface indicate low water depth (MF 2, 3, 5, 6), including a microbialithic bindstone (MF 5) that was only found in a small sink on the erosion surface. The reef core seems to have been eroded during a regression, without recovery. Instead, an open-marine carbonate platform with a high biological and ecological diversity developed (MF 5-14). Typical open-marine organisms occur, especially conodonts, cephalopods, trilobites and elasmobranchs. In contrast, the formerly dominant and varied stromatoporoids became limited to lamellar/tabular growth forms and are only present in the lower part of the succession, where they constitute bindstones and small framestones together with alveolitic tabulate corals (MF 6, Fig. 4). *Stachyodes australe* (WRAY, 1967) is the predominant binder in this assemblage.

The most common facies of the platform sediments (MF 8) is marked by abundant crinoids and brachiopods and a wide range of other biogenic components, including stromatolitic crusts, calcareous algal tubes, foraminiferans, sponge spicules and frameworks (*Hexactinellida*), *Thamnopora* and other tabulate corals, as well as solitary and (broken) colonial rugose corals, bivalves, gastropods, cephalopods (goniatites and orthoconic forms), bryozoans (various kinds), trilobites, ostracods, echinoids, conodonts,

and shark teeth (mostly *Protacrodus*). Variability in the content and fabric is high in this facies. The presence of micritic layers and tempestites indicates sedimentation between the normal wave base and the storm wave base. A very similar facies near the top of the succession (MF 13) is distinguished by the presence of laminoid Alveolitida, a generally higher mud (micrite) content, and the absence of rudstones. It was probably produced close to the storm wave base.

The platform sediments show an overall deepening trend towards the top. This is evident first from a change of composition in conodont assemblages, from Icriodus/Polygnathus-dominated (Fig. 6e-f) to Palmatolepis-dominated (Fig. 6a-d) assemblages, and on a larger scale from the change of the basic sedimentation structure from pack- and grainstones to wacke-/mud- and floatstones with increasing mud content. This is accompanied by the disappearance of the locally last stromatoporoids and blooms of microbial mats (MF 10). The uppermost facies (MF 14) was probably deposited below the storm wave base.

Other, more rare MF of the platform include one facies bearing dasycladaceans in considerable amounts (MF 9), one consisting almost purely of crinoid ossicles (MF 11), and one rich in sponge spicules (MF 12). There are also microbialithic Zebra Limestones (MF 7), named for conspicuously alternating layers of dark wackestones and light zement, that seem to be more common in some parts of the quarry to the north.

The platform phase bears only a slight resemblance to the Iberg Magnafacies (e.g. KREBS 1974), which is defined as a cap stage without a reef core or diversified zonation. In the Iberg Reef of the Harz Mountains, steep slopes led to a dominance of reef talus sediments (e.g., GISCHLER & ERKOÇ 2012), which is not the case at Rohdenhaus. The Schlupkothen Subfacies (of the Iberg Magnafacies), which occurs in the southern Velbert Anticline, is a less sloped extension of the fore-reef-like Hahnenfurth Subfacies towards the basin, but described as fossil-poor (BRINKMANN et al. 1970). The lack of stromatoporoids in the upper part of the Wülfrath Reef Complex is unusual and sets it apart from the other, mostly older complexes in the region.

The conodont assemblage from just above the top back-reef unconformity with *Icriodus alternatus alternatus* BRANSON & MEHL, 1934 (Fig. 6e) and *Ancyrodella curvata* (BRANSON & MEHL, 1934) Late Morphotype suggests an upper Frasnian age. *Palmatolepis winchelli* (STAUFFER, 1938) (Fig. 6a) from slightly higher marks the Montagne Noire (MN) 12 Zone. Near the top of the platform succession, *Pa. bogartensis* (STAUFFER, 1938) and others indicate the MN 13a Zone. There is a layer at the upper boundary of the platform sediments where conodonts from the Frasnian and Famennian are mixed. This includes *Pa. bogartensis* (Fig. 6b) and *Pa. eureka* ZIEGLER & SANDBERG, 1990 (Fig. 6c), which represent MN 13 Zone, and a peculiar giant form of *Palmatolepis* with affinities to *Pa. salibaevi* BARDASHEV, 2009 (Fig. 6d).

Where subsequent karstification hasn't destroyed the succession, the platform limestones are still mostly capped by an erosional discontinuity lasting from the early to middle Famennian and overlain by various early to middle Famennian sediments. Only in a very restricted area the sharp contact with dark-grey and goniatite-rich Upper Kellwasser Limestones has been observed (Fig. 5). It represents the uppermost Frasnian (MN 13a/b Zones) and is also capped by erosion and overlain by earliest Famennian limestones with many orthoconic cephalopods and the opportunistic goniatite *Phonexites frechi*, just as in sections from southern Morocco. Most likely the platform drowned finally during the hypoxic Upper Kellwasser Event.

#### Acknowledgements:

The Rheinkalk AG kindly allowed access to their quarry and provided the cores for this study, which was a part of the "Eovariscan" DFG Project granted to RTB and ZSA. It is also a contribution to IGCP 595.

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### Figures:

**Fig. 1.** MF 3 – Rudstone with *Amphipora*, *Stachyodes* and *Thamnopora*. Karst filling (dark) on the left side. Well sample 9913-131, bar = 1 cm.

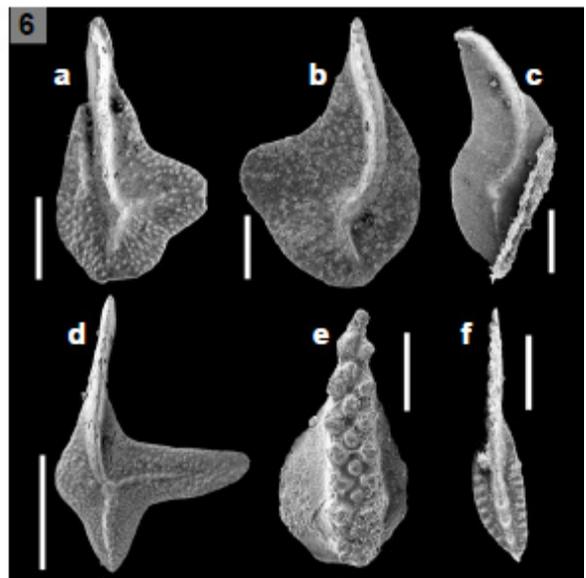
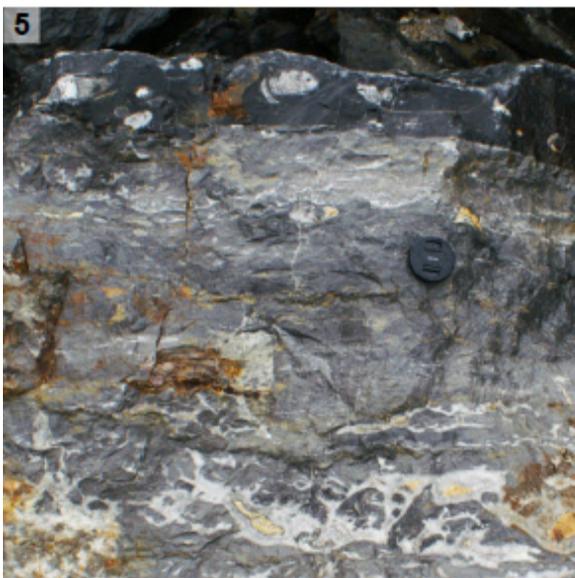
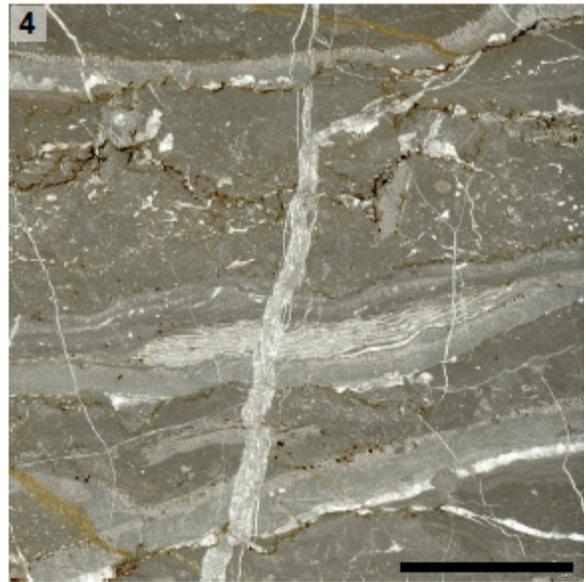
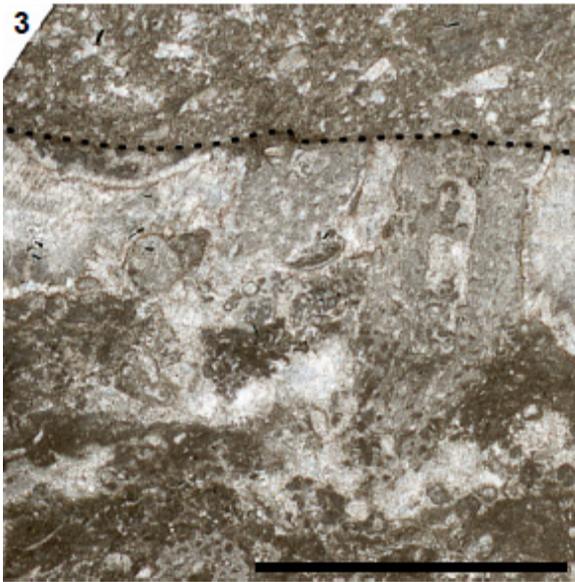
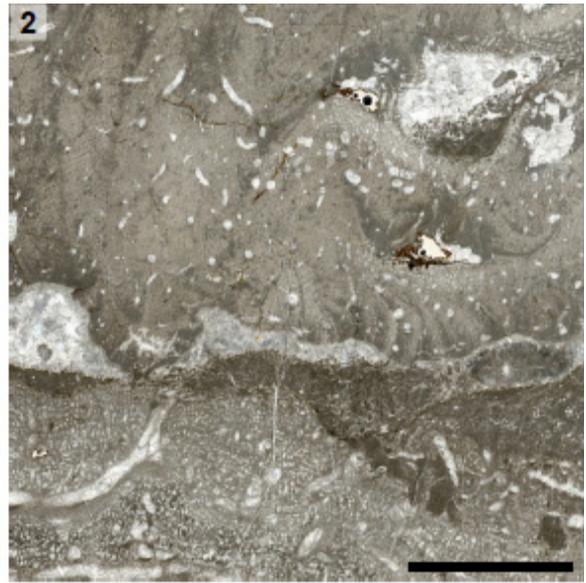
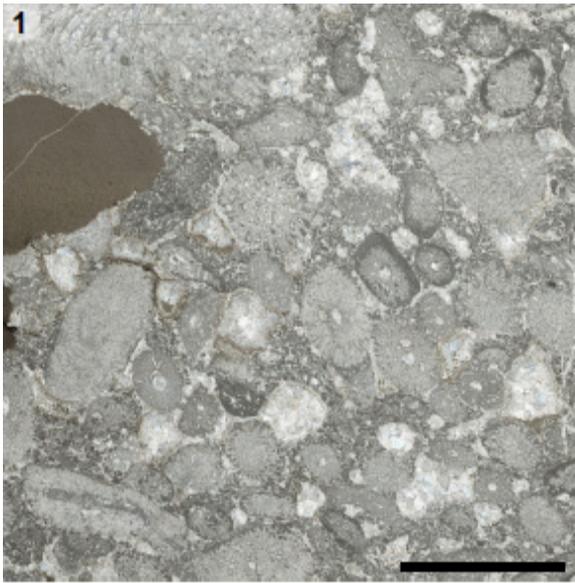
**Fig. 2.** MF 4 – Stromatoporoid framework. Gaps geopetally filled with peloidal packstone (MF 1) and cement. Well sample 9913-123B, bar = 1 cm.

**Fig. 3.** Top of back-reef sediments; erosional disconformity (dotted line) truncating *Amphipora*-stems, later hollowed. Well sample 9913-122B, bar = 5 mm.

**Fig. 4.** MF 7 – Stromatoporoid-tabulate-coral-bindstone. well sample 9913-117, bar = 1 cm.

**Fig. 5.** Contact between last microbialithic reefal limestone (with some *Manticoceras*) and Upper Kellwasser Limestone (black, at the top) with many *Crickites holzapfeli*. Isolated block at Rohdenhaus-North. Cap diameter = 6.1 cm.

**Fig. 6.** Conodonts examples. (a) *Pa. winchelli* (STAUFFER, 1938), outcrop sample 18, bar = 500 µm; (b) *Pa. bogartensis* (STAUFFER, 1938), well sample 9913-59A, bar = 200 µm; (c) *Pa. eureka* ZIEGLER & SANDBERG, 1990, well sample 9913-59A, bar = 500 µm; (d) *Palmatolepis* aff. *salibaevi* BARDASHEV, 2009, well sample 9913-59A, bar = 1 mm; (e) *Icriodus alternatus alternatus* BRANSON & MEHL, 1934, well sample 9913-122, bar = 200 µm; (f) *Polyganthus decorosus* STAUFFER, 1938, outcrop sample 27, bar = 200 µm.



## Devonian-Carboniferous Radiolarians and Conodonts from the South of the Char Ophiolite Zone

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New data were obtained on radiolarians and conodonts from the Char Ophiolite Zone, east Kazakhstan. The Char Ophiolite Zone or Char Suture-Shear Zone is one of the units of the Central Asian orogenic belt (ZONENSHAIN et al. 1990, ERMOLOV et al. 1980, SAFONOVA et al. 2012). This strongly tectonically dislocated structures represent remains of the NE branch of the Paleo-Asian Ocean that was closed due to the collision and amalgamation of the Kazakhstan and Siberian continents during the Late Carboniferous (DOBRETSOV 2003, BUSLOV et al. 2004, SAFONOVA et al. 2012). The sedimentary rocks from the melanges contain Late Devonian to Early Carboniferous radiolarians. A first report on radiolarians from the Char Zone was made by B.B. NAZAROV (1975). Later Late Devonian radiolarian associations from cherts of the Karabaev and Urumbaev Formations and Early Carboniferous radiolarians from siliceous sediments of the Verochar Formation were reported (IWATA et al. 1997). Detailed studies of the territory, its geodynamics and Paleozoic serpentinite mélange revealed new data on basalts and sedimentary sequences associated with them (SAFONOVA et al. 2012). The resampling of old localities as well as new samples showed the presence of relatively well preserved radiolarian associations; their Late Devonian – Early Carboniferous age was proved by findings of conodonts in the same samples.

The Karabaev and Urumbaev Formations are volcano-sedimentary units comprised of basalts, tuffs, tuffaceous sandstones, siliceous-aleurolites, siliceous mudstones, vari-colored cherts (sometimes bedded), and rare limestones. Radiolarians and conodonts were obtained mainly from red, greenish-grey and black cherts, and conodonts also from limestones, which crop out near Nikolaevka village, the dam of the Char water reservoir, and to the NW of Char village. The Verochar Formation consists of volcanic-carbonate-terrigenous deposits: limestones intercalated with basalts, siliceous mudstones, cherts, and tuffaceous sandstones. Radiolarians and conodonts were collected from grey cherts to the SW of Igorevka village.

The Late Devonian radiolarian assemblage recovered from the Urumbaev Formation (Char dam) is diverse and represented by the spherical forms *Trilonche minax* (HINDE), *Tr. echinata* HINDE, *Tr. davidi* (HINDE), *Trilonche* sp., *Astroentactinia stellata* NAZAROV, *A. cf. paronae* (HINDE), *Borisella* ?sp., *Archocyrtium* cf. *typicum* CHENG, *Arocyrtium* sp., *Tetraentactinia* cf. *barysphaera* FOREMAN, *Stygmospaerostylus* sp. 1, and *Stygmospaerostylus* sp. All identified species are characteristic for the Late Devonian worldwide. Moreover, *Tetraentactinia barysphaera* and *Archocyrtium typicum* were described as typical Famennian species (FOREMAN 1963, CHENG 1986, MALETS 2011). The conodonts *Palmatolepis rhomboidea* SANNEMANN and *Palmatolepis* sp., recovered from limestones, are characteristic for the middle Famennian *rhomboidea*-Lower *marginifera* Zones.

From the several outcrops of the Karabaev Formation two localities were sampled - (1) to the SE and (2) to the NW of Char village. Diverse and relatively well preserved radiolarians along with rare conodonts have been found. The first assemblage contains *Trilonche minax* (HINDE), *Tr. davidi* (HINDE), *Tr. echinata* HINDE, *Trilonche* sp., *Stygmospaerostylus* sp., *Tetraentactinia* aff. *barysphaera* FOREMAN, and the conodonts *Palmatolepis* sp. and *Polygnathus* sp. The presence of *Palmatolepis* spp. along with *Tetraentactinia* aff. *barysphaera* suggests a Late Devonian (Famennian) age for the source strata. The assemblage from the second locality yielded *Trilonche* cf. *vetusta* HINDE, *Tr. echinata* HINDE, *Tr. davidi* (HINDE), *Stygmospaerostylus* sp., *Archocyrtium* cf. *reidelli* DEFLANDRE, *Ar. cf. ormistoni* CHENG, *Archocyrtium* sp., *Duplexia* sp., and *Helioentactinia* sp. Among conodonts *Polygnathus* sp. and *Siphonodella* cf. *bella* KONONOVA & MIGDISOVA were recovered. *Siphonodella bella* was first found

from the Tournaisian (Early Mississippian) *duplicata* Zone (BARSKOV et al. 1984). The recovered species of *Archocyrtium* are also characteristic for the Early Carboniferous (CHENG 1986).

It should be mentioned the previous age of the Karabaev Formation was supposed to be Frasnian? – Famennian; moreover it was believed to underlie the Urumbaev Formation (ERMOLOV et al. 1980, IWATA et al. 1997). Our new findings of radiolarians and conodonts allow to specify its age as Late Devonian–Earliest Carboniferous. However, the age and stratigraphic position of both the Urumbaev and Karabaev Formations require further investigation.

The Verchar Formation yielded typical Lower Carboniferous *Albaillella* cf. *paradoxa* DEFLANDRE, *Albaillella* sp., and *Polyentactinia* sp. The conodont *Gnathodus* cf. *punctatus* (COOPER) occurs at the same locality. The radiolarian *Albaillellaparadoxa* is the index-species for the Tournaisian *paradoxa* Zone (DEFLANDRE 1952, NAZAROV 1988). The recovered conodonts are characteristic for the *Siphonodella isosticha*–Upper *S. crenulata* - lower *Gn. typicus* Zones. This confirms the Tournaisian age.

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## **An attempt of time calibration of the Lower Tournaisian (Hastarian Substage) based on orbitally forced sequences**

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### **1. Introduction**

The Hastarian Substage correspond to the lower division in Belgium and adjacent areas (CONIL et al. 1977, HANCE & POTY 2006). Third-order sequences are well marked in the Tournaisian and Viséan of the Belgian Namur-Dinant Basin and were the base for the revised sequence stratigraphy of the Mississippian of HANCE et al. (2002). Investigations in two sections in the Dinant vicinity displaying continuous well weathered sections - the railway cutting at Gendron-Celles and the Nutons quarry in Spontin – show that the lower Tournaisian part of the third-order sequences 1, the sequence 2 (Fig.1) and the lower part of the sequence 3, are divided in shorter sequences. Because of their relative regular distribution, their thickness, and the fact they are not grouped in bundles, these sequences are considered as due to only one orbital parameter corresponding to precession cycles.

### **2. Description and distribution of orbitally forced sequences**

The sequences vary from alternations of shale and calcareous shale (mainly in the Pont d'Arcole Formation [Fm]) to alternations of calcshale and limestone (mainly in the Hastière Fm), and to limestone bed dominated (mainly in the upper part of the Landelies Fm). Their thickness varies from about 0.2 m to 1 m, and is strongly influenced both by the sediment production and by the compaction in the argillaceous levels and precession dissolution in the limestone levels. The distribution of their sedimentary nature follows the sedimentary evolution of the third-order sequences 1 to 3 described by HANCE et al. (2002). Truncations at the top of the sequences indicating emersion can occur during the low-stand (LST) and the falling-stage system tracts (FFST) of the third-order sequences. They are not marked during the transgressive (TST) and high-stand system tracts (HST), suggesting relatively low eustatic variations at the scale of these sequences.

The lower part of the Gendron-Celles section allows to study the orbitally-forced sequences in the Hastière Fm (HST and FSST of the third-order sequence 1, and LST of the sequence 2), almost from the base of the Tournaisian. The later is placed just above the 1.76 m-thick limestone bed marking the base of the Hastière Fm. That section is completed by the Spontin section which allows to study the sequences in the upper member (mbr) of the Hastière Fm (LST of the sequence 2), the Pont d'Arcole Fm (TST of the sequence 2), the Landelies Fm (HST and FSST of the sequence 2), and the Yvoir Fm (LST and TST of the sequence 3) in which is situated the base of the Ivorian Substage (base of Upper Tournaisian, base of *Polygnathus communis carina* Zone).

In the Gendron-Celles section, the 1.75 m-thick thin-bedded limestone unit situated just above the Devonian-Carboniferous boundary does not show any trace of rhythmic deposits. Those ones develop above it and 31 sequences were counted in the rest of the lower mbr of the Hastière Fm. None was recognized in the 3.65 m-thick massive limestone of the middle mbr which marks the top of the sequence 1 (FSST). Both the Gendron-Celles and the Spontin sections show 15 sequences in the upper mbr corresponding to the LST of the sequence 2. In the Spontin section, 24 sequences compose the Pont d'Arcole Fm (TST 2), 28 the lower part of the Landelies Fm (HST 2), and 26 the upper part of the formation (FSST 2). 31 compose the lower part of the Yvoir Fm (LST and TST 3) until the Hastarian/Ivorian boundary (Lower/Upper Tournaisian boundary).

### 3. Attempt of calibration

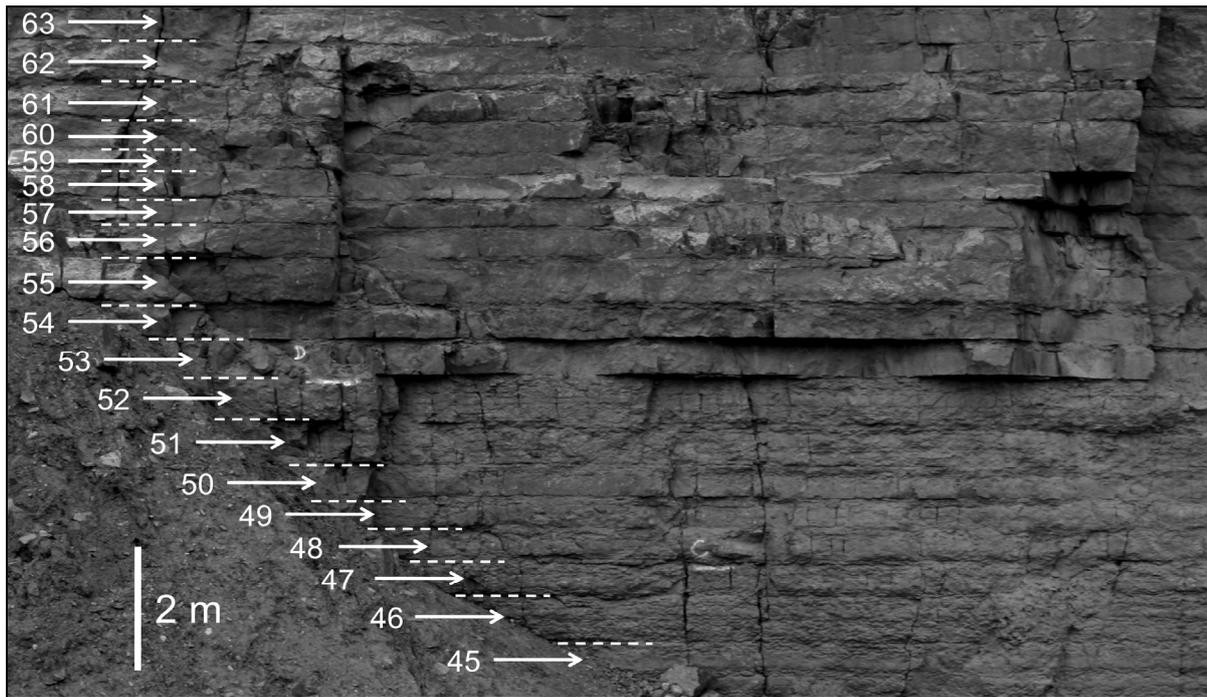
The precession cycles vary through time and are usually considered as being not useful for a time calibration. But according to BERGER et al. (1989), the precession cycles for the Lower Tournaisian are about 17 and 20.2 Ka, and it is possible to consider 18.6 Ka as a rough average for their duration. On another hand, according to GILES (2009) who calibrated the three last third-order sequences of the Viséan to 2.4 Ma and considered that they could correspond to excentricity cycles, we consider that the Tournaisian ones also could last about 2.4 Ma. Therefore the 93 precession sequences recognized in the third-order sequence 2 could represent 1.73 Ma, suggesting that the erosion surfaces marking the sequence boundaries between sequence 1 and 2, and 2 and 3, could correspond to gaps as long as about 0.67 Ma if we consider a similar number of precession sequences in both sequences 1 and 3.

So, considering the duration of the 31 precession sequences in the Hastarian part of the third-order sequence 1, i.e. 0.577 Ma (not including the unknown time corresponding to the 1.75 m-thick unit at the base and the 3.65 m-thick unit at the top), + 0.67 Ma, + 1.73 Ma, + 0.67 Ma, + 31 sequences in the Hastarian part of the third-order sequence 3, i.e. 0.577 Ma, we obtain 4.224 Ma as a possible duration for the Hastarian. Note that, according to MENNING et al. (2001), the Hastarian lasted for about 6 Ma.

A refinement of that method and the extension of the recognition of orbitally forced sequences to the Ivorian Substage (Upper Tournaisian) possibly could contribute to a better definition of its length and of the one of the Tournaisian Stage.

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**Fig. 1.** Orbitally forced sequences 45 to 63 in the HST of the third-order sequence 2, lower part of the Landelies Fm, Hastarian; Nutons quarry, Spontin.

## **The use of conodont and vertebrate biostratigraphy in determining key boundaries in Late Devonian to Early Carboniferous sections, Canning Basin, Western Australia**

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The northern, onshore section of the Canning Basin, in northwestern Western Australia, preserves an extensive Devonian Reef Complex (PLAYFORD et al. 2009). The reef complexes saw continual building over some 30 Myr from the Givetian through to the Frasnian before a sudden halt towards the end of the Famennian. Extensive studies into the geochemistry, sedimentology, diagenesis and palaeontology have been undertaken (PLAYFORD et al. 2009 and references therein); however, correlation between facies, particularly the back-reef, and core to outcrop remains problematic.

This current study examines the Givetian Cadjebut Formation, the Frasnian Gogo Formation, the Frasnian to Famennian Virgin Hills and Napier formations as well as the Famennian–Carboniferous Fairfield Group. The F/F boundary is located within the Virgin Hills Formation, but locating the precise position of this major extinction event has been difficult and relies heavily on goniatite and conodont biostratigraphy, as the facies lack the distinct anoxic horizon found in Europe and north Africa and typically associated with the event (e.g., PLAYFORD et al. 2009). However, constraining the boundary in the back-reef facies is even more difficult due to the absence of index fossils. Similar problems are encountered in constraining the exact position of the Devonian–Carboniferous Boundary (PLAYFORD et al. 2009).

The usefulness of microvertebrates as a biostratigraphic tool was recognised early on, especially as many of the microvertebrate groups occur in restricted environments, where conodonts and other invertebrate index fossils do not (TURNER 1997). TRINAJSTIC & GEORGE (2009) were able to correlate thelodont and chondrichthyans against the standard conodont zonations for the Frasnian in the Canning Basin (TRINAJSTIC & GEORGE 2009). It is the aim to extend the microvertebrate scheme from the Givetian to Carboniferous in the Canning Basin.

The incorporation of microvertebrates into the established conodont schemes will aid in the identification of major geological boundaries in the Canning Basin, and facilitate correlation, especially between restricted facies, across northern Gondwana. The Devonian reefs contain a diverse vertebrate fauna, which show gradational change from less derived form in the Givetian to larger dinichthyid placoderm predators in the Famennian. There is a major shift in the fauna with chondrichthyans becoming more numerous in the Late Famennian and a major radiation in the early Carboniferous. In addition, the youngest record of thelodonts, originally thought to have become extinct at the end of the Frasnian (TURNER 1997), have been recovered from the Late Famennian. These new finds indicate a closer faunal relationship between Western Australia, Iran and North Africa than previously realised.

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## **(P) The brachiopod reference collections of the Science Museum (Coimbra, Portugal): a legacy for the advance of Devonian stratigraphy and global correlation**

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The Science Museum of the University of Coimbra is housing one of the oldest paleontological collections of Portugal. Among these are the KRANTZ and SAEMANN Brachiopod collections which were bought at the beginning of the 20th century as teaching material for the students of Natural Sciences at the University of Coimbra. Due to re-organization of the museum collections, movement of the exhibition halls and the inauguration of the "new" Science Museum, these classical and ancient collections were put aside. In the frame of our actual project these brachiopod collections were now revised, catalogued, new labels printed and specimens adequately stored in new and closed boxes.

The Krantz and Saemann Brachiopod collections consist each of 178 lots, of which the majority are German and Czech Devonian specimens. Both collections include specimens from the Gedinnian to the Framennian, however, the KRANTZ Collection concentrates on Siegenian and Emsian material, whereas the SAEMANN Collection is yielding especially Gedinnian and Middle Devonian material. The Late Devonian is represented by only a few lots in each collection. The revision of the brachiopod taxa shows that many of them represent type species of different genera, most of them even collected at topotype localities. As a matter of fact these collections are still of high importance for actual stratigraphical and paleobiogeographical studies, even though were bought more than a decade ago and never served for scientific paleontological studies. An example for their scientific importance is the current debate of the faunal and geographical relationship of today's Portugal with Gondwana and Laurentia during the Devonian Period.

In order to distribute and to conserve the content of the KRANTZ and SAEMANN brachiopod collections, two catalogues have been elaborated (SCHEMM-GREGORY & HENRIQUES in press a, b). They are written in popular scientific style, explaining the biological and paleontological background of brachiopods, their evolution during the Phanerozoic, and the paleogeographical position of the collecting localities. The chapters of the catalogue are ordered by period and within the period by brachiopod phylogeny following the revised "Treatise on Invertebrate Paleontology". One representative brachiopod of each lot is figured and all information (name, inventory number, detailed stratigraphy and collecting locality information) are provided. Both catalogues are written in Portuguese and English so that information is available for scientific researchers from all over the world.

### **Acknowledgements:**

The authors thank Miguel M. RAMALHO and Rita SILVA (both LNEG, Lisbon) and Pedro CASALEIRO (Science Museum, Coimbra) for giving access to the brachiopod collections under their care. José PIÇARRA (LNEG Beja) is acknowledged for discussion about the Paleozoic Geology and stratigraphy of Portugal. M. S.-G. thanks Ursula MÜLLER-KRANTZ (Bonn, Germany) for providing historical information about the history of the "Dr. F. KRANTZ, Rheinisches Mineralien-Kontor" and its relationship to the University of Coimbra and the Geological Commission of Portugal.

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This is a contribution to the IGCP Project 596 “Climate Change and biodiversity Patterns in the Mid-Paleozoic”.

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## The Lower Devonian of Portugal – an attempt for brachiopod biostratigraphical correlation

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Devonian brachiopods are found in four regions within two Portuguese tectono-stratigraphic domains: Ossa Morena Zone (Barrancos area) and Central Iberian Zone (Rates area, resp. Valongo Syncline; Dornes and Portalegre synclines) (Fig. 1). Since Nery DELGADO (1835-1908), one of the first pioneers of the geology in Portugal, extensive brachiopod samples have been collected, all of them accurately labeled and stored in the LNEG Geological Museum or at the archives of the former Geological Survey, today LNEG, in Lisbon, however, the identification needs in most cases a modern revision. Even though these collections are extensive in paleontological and lithological aspects and are still studied by various scientists (e.g., PIÇARRA et al. 1999, OLIVEIRA et al. 2000), a detailed (bio)stratigraphical correlation of the Devonian of Portugal is still lacking.

Our actual project is the revision of the classical Devonian brachiopod collections of Portugal and the elaboration of a brachiopod biostratigraphy. We have visited the outcrops by Nery DELGADO and collected further material with special attention to the stratigraphy. The strong tectonic deformation and the lack of external moulds of brachiopods of the collections are complicating the identification. With the use of recent collected material, we have identified the specimens of the museum collection under modern taxonomical aspects. In our work we paid special attention to the Siegenian and Emsian strata: Dornes Formation (Dornes Syncline), São Mamede Formation (Portalegre Syncline), Telheiras Formation (Valongo Syncline) and Monte de Russianas Formation (Barrancos area), because brachiopods of older Devonian strata have already been studied in the last decade (e.g., GOURVENNEC et al. 2008, 2010; LE MENN 2002).

We present here a correlation of these 4 regions based on brachiopod biostratigraphy. In our data the increase of endemism is recognizable within the Portuguese Emsian brachiopod faunas (SCHEMM-GREGORY 2011) and affinities with northern Gondwanan terranes (today's Morocco, Spain and France) and partly with *Laurentia* are seen.

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**Fig. 1:** Topographical map of Portugal with undivided Silurian to Lower Devonian strata highlighted in grey and brachiopod collecting localities indicated.

## The Devonian of the Turkish Taurides – new biostratigraphic insights

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Three Devonian successions in the Central and Eastern Taurides (Eceli, Halevikdere, and Kocadere) have been studied in detail (and recently extended stratigraphically) with respect to biostratigraphy, sedimentology, facies and global events. These sections are 1050 to 1350m thick; an additional section (Akkuyu, 170 m) close to the Eceli Section has also been studied. After some biostratigraphical data from the Eastern Taurid region were reported a few years ago (GÖNCÜOĞLU et al. 2004 and literature therein), sufficient data have now been recovered to allow the subdivision of the successions at stage level, permitting their approximate correlation and comparison with the regional lithostratigraphy; and this enables reconstruction of the temporal development of sedimentary processes and facies changes in the individual sections and comparison to some of the Devonian global events (WEHRMANN et al. 2010). Data gained after publication of the summarising article by WEHRMANN et al. (2010) have recently been presented in parts for one of the sections (Halevikdere; YALÇIN et al. 2012) and will be shown in some detail.

Two stratigraphic levels are of special interest in these sections: the Silurian/Devonian boundary interval and the interval with newly discovered equivalents of the Frasnian/Famennian boundary. In the first case, two very different groups of organisms contribute to knowledge of the Silurian/Devonian boundary beds: scyphocrinoids (loboliths as well as parts of the calyxes and arms) and palynomorphs. The finds of loboliths of the Camarocrininae (determination by R. HAUDE, Göttingen, is gratefully acknowledged) which belong to the plate loboliths (for more information see HAUDE 1972, 1989, 1992, 1998, 1999; HAUDE et al. 1994, HAUDE & WALLISER 1998) indicate a position adjacent to the Silurian/Devonian boundary (Figs. 1 and 2). This is the first record of scyphocrinoids from Turkey. Support for the Silurian/Devonian boundary interval also comes from the newly studied palynomorph samples which contain acritarchs, chitinozoans, and spores. New data from palynomorphs are also "responsible" in respect to the Frasnian/Famennian boundary interval. Very rich assemblages (again representing different groups of palynomorphs) are present in a unit which may be regarded as equivalent of the late Frasnian Kellwasser Event. Brachiopods collected from the Halevikdere and Kocadere sections (both Eastern Taurides) also have biostratigraphical potential in the Lower Devonian parts and the Frasnian portions of the sections.

In addition to these more specific levels, some newly sampled brachiopods from the new Akkuyu Section (Central Taurides) may help in age determination when investigated more closely. At first glance, a Frasnian age can be suggested, at least for the upper part of the section.

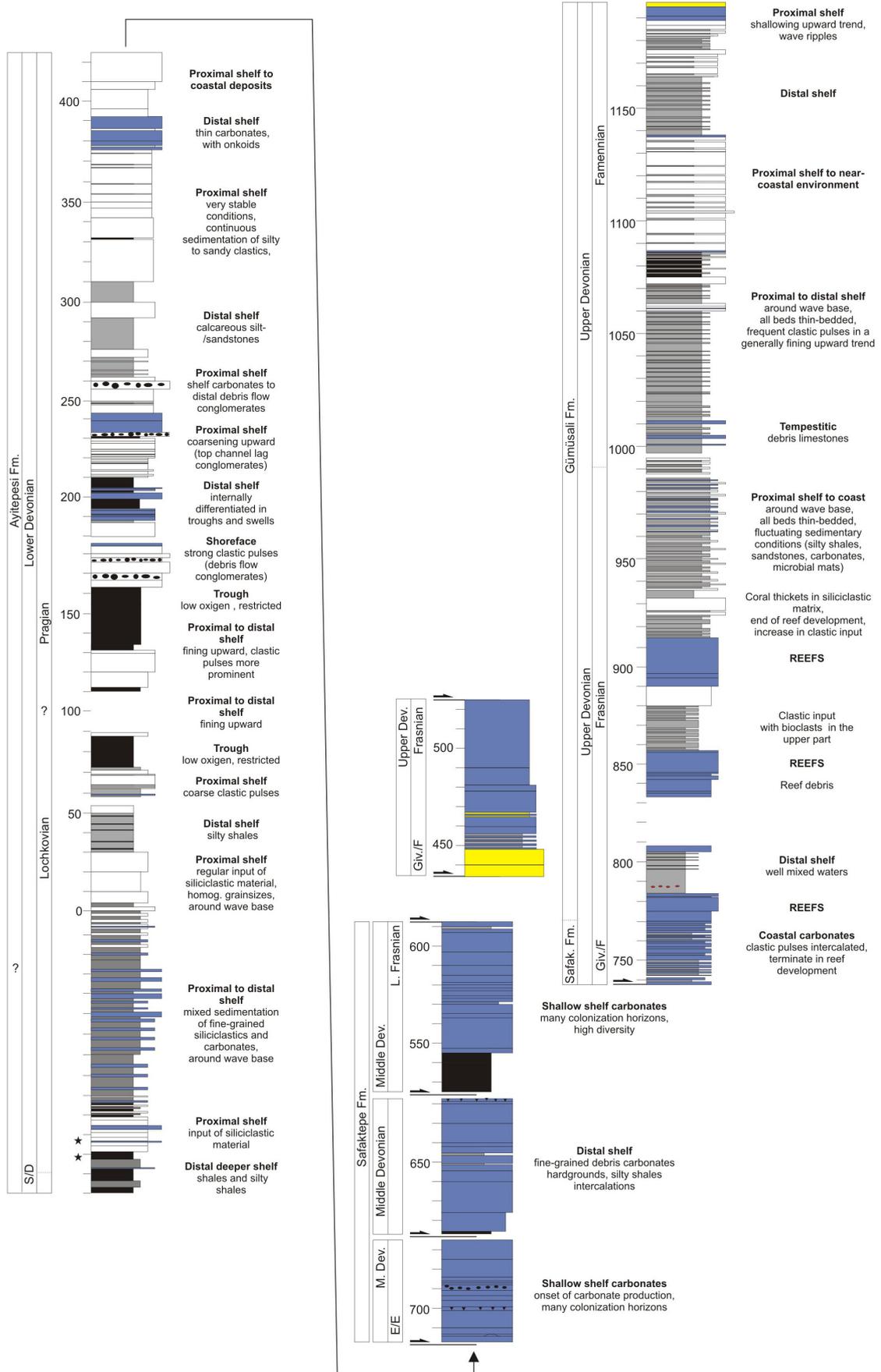
The biostratigraphic framework of the studied sections now allows quite accurate interpretations of the evolution of sediments and facies in the Devonian of the Taurides including the option of widespread comparison and correlation.

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## Fig. 1:

Lithology, stratigraphy, and depositional environment interpretation of the Halevikdere Section, Eastern Taurides (modified after WEHRMANN et al. 2010 and YALÇIN et al. 2012). Note that meter numbering does not necessarily reflect the stratigraphic order; lower part added after recent fieldwork (without numbering). \* = position of two horizons with "plate loboliths" close to the Silurian/Devonian boundary.





**Fig. 2.** Lobolith of a scyphocrinoid Camarocrininae from the lower lobolith bed (19.6 m above base of newly studied section shown in Fig. 1). The lobolith was found *in situ* (compare YALÇIN et al. 2012, fig. 9 – specimen visible further to the left from two other specimens mentioned in that figure).

## Subdivision of the "Original Pragian" - new data from the type area

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The pending problem of the position of the prospective basal Emsian boundary largely affects the duration of the Pragian Stage. Combination of inaccuracy and deviations in radiometric data and the basal Emsian GSSP caused large time-proportional problems in the Lower Devonian: 1. practical vanishing of the entire stage (Pragian), 2. extremely long duration of the Emsian, 3. unknown duration of the Lochkovian.

The Pragian Stage has been originally proposed to correspond with the complete succession of well lithologically and biostratigraphically constricted Praha Formation. The correlation is based on extensive faunal datasets that have been developing since the later half of 19th century but also on other methods – e.g., magnetostratigraphy (MS), gamma-ray spectrometry (GRS). The stratigraphic correlation by means of the DTW (Dynamic Time Warping) method has been successfully applied within the Prague Synform (HLADIL et al. 2010) but also for distant correlation between sections in Barrandian and Kitab State Reserve, Uzbekistan (HLADIL et al. 2011). If we consider the proportions based on high-frequency cycles provided by CHLUPÁČ (2000) then duration of the "original Pragian" based on the Praha Formation can be estimated between 6 to 7.5 Ma; it may thus represent more than one fourth of the Lower Devonian time, considering duration of the Lower Devonian between 25 to 27 Ma (KAUFMANN 2006). This considerable time-span is of course drastically reduced (up to only one third) due to the basal Emsian GSSP (cf. e.g., SLAVÍK et al. 2007, CARLS et al. 2008).

The Praha Formation ("original Pragian") can be subdivided into three parts based on: 1. the biostratigraphic data – i.e., dactyloconarid ranges and conodont zonation (SLAVÍK 2004a, b), 2. eustatic and sedimentary features of the key-sections, 3. geo-chemo-physical properties of the carbonate rocks (KOPTÍKOVÁ et al. 2010, HLADIL et al. 2011). In the presentation, the suggested three-fold subdivision of the "original Pragian" will be confronted with the "official Pragian" (i.e. Pragian confined by the GSSP boundaries). The newly obtained radiometric datum from the Lochkovian/Pragian boundary will be presented and the relative duration of the Lower Devonian stages will be commented.

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## **Sea level versus climate in the Famennian deposits of the southern Poland – an insight from rock magnetic measurements**

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Two Famennian sections located in the Holy Cross Mountains (Janczyce 1 borehole and Kowala Quarry) and Goczałkowice IG-1 borehole from the Silesia region (southern Poland) were selected for detailed magnetic susceptibility (MS) and rock magnetic properties analyses. In order to distinguish MS carriers, several tests were performed, such as: isothermal and anhysteretic remanent magnetization (IRM, ARM), S-ratio, hysteresis measurements. Natural gamma ray record was tested together with inorganic geochemistry measurements for selected samples.

Two hundred and fifty meters interval of the Famennian part of the Janczyce 1 borehole was sampled. Clear, large-scale magnetic susceptibility changes mirror total gamma ray record. Moreover, distinct correlation of selected elements (proxies of a detrital input into the basin), with magnetic susceptibility confirmed that MS signal is mostly of terrigenous origin, therefore can be indirectly correlated with postulated T-R cycles.

In the uppermost Devonian interval of the Kowala Quarry marls and marly shales are dominated by paramagnetic minerals, which contribute significantly to magnetic susceptibility signal. In wavy nodular limestones and micritic limestones the presence of hematite-rich horizons (related possibly to Permo-Triassic remagnetization episode of Devonian carbonates) was also confirmed by more detailed studies of rock magnetic properties.

Distinct increase in magnetic susceptibility in the Famennian/Tournaisian interval of the laminated grained limestones is visible in the Goczałkowice IG 1 section as well as in all of the studied sections. The obtained results were compared with the previously distinguished sedimentary transgressive-regressive cycles. However these data were not comparable due to possibly secondary MS signal, which might be connected with an intense diagenetic alteration.

The complex nature of magnetic susceptibility in the carbonate rocks makes it difficult to draw sea level or climate related conclusions, without a careful inspection of a secondary minerals influence.

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## Famennian conodont zonation in the Barma and Makarovo deposits at the Kuk-Karauk-I Section (South Urals)

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The Upper Devonian succession is well exposed in the Kuk-Karauk-I section (MARKOVSKY 1937, KONONOVA 1969, ABRAMOVA et al. 1990, SINITSYNA et al. 1995, ABRAMOVA 1999, ARTYUSHKOVA 2004, ARTYUSHKOVA et al. 2011). This succession mainly consists of packstones with corals, brachiopods, crinoids, conodonts, and fish remains. Here, we give a new detailed conodont stratigraphic subdivision of the *triangularis* - Upper *marginifera* Zones (Barma and Makarovo horizons; see Fig. 1).

**The *triangularis* Zone:** The lower boundary of the zone (the F/F boundary) lies at the bottom of the Barma Horizon inside the brachiopod coquinas (Bed 2 in Fig. 1). It is determined by the simultaneous appearance of the conodont *Palmatolepis triangularis* and the brachiopod *Parapugnax markovskii* (ABRAMOVA 1999, ABRAMOVA & ARTYUSHKOVA 2004). The *triangularis* Zone is represented by the interval of the joint Lower and Middle zones. The conodont association is extremely poor in taxa, where *Icriodus* and rare *Palmatolepis* specimens are recorded (see Fig. 1). The thickness of the Barma Horizon (Lower – Middle *triangularis* zones) is 0.6 m.

The Upper *triangularis* Zone corresponds to the base of the Makarovo Horizon (BARYSHEV & ABRAMOVA 1996). Deposits of this zone are absent in the studied section.

**The *crepida* Zone:** In the Kuk Karauk-I section this zone is observed for the first time. It is presented by the Lower *crepida* Zone only. The lower boundary is installed at the bottom of Bed 3, with the appearance of the typical species *Pa. crepida* SANNEMANN, *Pa. quadratinodosalobatasandbergi* JI & ZIEGLER, and *Pa. wolskajae* OVNATANOVA (see Fig. 1). It includes light-grey and brownish, layered packstones and granestones. The thickness of the Lower *crepida* Zone is 0.35 m. The conodont association is taxonomically diverse and consists of 3 genera and 17 species (see Fig. 1).

The interval of the Middle – Uppermost *crepida* zones corresponds to a hiatus.

**The *rhomboidea* Zone:** The lower boundary of the zone is established at the base of the Bed 4 by the appearance of the characteristic species *Pa. rhomboidea* SANNEMANN and *Pa. poolei* SANDBERG & ZIEGLER. Its whole thickness is 0.05 m only. This interval is marked lithologically by grey ferruginous grainstones. The conodont association consists of 3 genera and 11 species (see Fig. 1). The *rhomboidea* Zone is revealed in the Kuk-Karauk I section for the first time.

**The *marginifera* Zone:** The base of this zone is determined at the bottom of Bed 5 by the first appearance of the zonal species *Pa. marginifera marginifera* HELMS (see Fig. 1). The *marginifera* Zone deposits have a cyclic structure and are characterized by alternating packstones and wackestones.

The lower part of the Lower *marginifera* Zone, 0.30 m thick (Bed 5), is represented by grey-brownish, layered packstones (see Fig. 1). The conodont association includes 4 genera and 10 species (see Fig. 1). The upper part of the zone, 0.40 m thick (Bed 6), is characterized by grey wackestones (see Fig. 1). At the base of Bed 6 the conodont content drops sharply. In the conodont association there are only rare and longer-ranging taxa, such as *Pa. minuta minuta* BRANSON & MEHL, *Pa. glabra pectinata* ZIEGLER, and *Pa. perlobata schindewolfi* MÜLLER (see Fig. 1).

The base of the Upper *marginifera* Zone is installed at the base of Bed 7, by the first appearance of *Pa. utahensis* ZIEGLER & SANDBERG (see Fig. 1). The lower part of the zone is 0.40 m thick and represented by light-grey layered packstones. In the conodont association, *Pa. aff. perlobata sigmoidea* ZIEGLER and

*Polylophodonta* sp. are found together with transitional taxa of the Lower *marginifera* Zone (see Fig. 1). The conodont succession of overlying deposits is under study.

Thus, in the Kuk-Karauk-I section the conodont zonation of the Barma and Makarovo horizons of the Famennian Stage is incomplete. The Upper *triangularis* and Middle - Uppermost *crepida* zones are absent (see Fig. 1). If one takes the Lower/Middle Famennian boundary at the base of the *marginifera* Zone, then it falls in the Kuk-Karauk-I section at the base of Bed 5 (see Fig. 1).

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## Correlation of the middle Lochkovian (Lower Devonian) conodont successions in peri-Gondwana key localities

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Middle Lochkovian conodont faunas in neritic environments are mainly cosmopolitan and facilitate global correlations. They are chiefly composed of taxa belonging to the genera *Ancyrodelloides* and *Lanea* with minor widespread of other relevant taxa as *Flajsella*, *Masaraella*, *Pedavis* and *Kimognathus* (VALENZUELA-RÍOS & MURPHY 1997, MURPHY & VALENZUELA-RÍOS 1999). In three key areas of European peri-Gondwanan sections (Spain; Czech Republic and Carnic Alps) the *Icriodus* and *Pelekysgnathus* record is also remarkable and can help in increasing the detail of correlations (SCHÖNLAUB 1980, VALENZUELA-RÍOS 1994, SLAVÍK et al. 2012, CORRADINI & CORRIGA 2012).

The purpose of this report is to establish a common preliminary refined biostratigraphical subdivision for these key areas that can serve as a basis for a fine conodont-based correlation in the peri-Gondwana areas of Europe and North Africa.

*Ancyrodelloides* was traditionally the dominant ubiquitous middle-Lochkovian genus in neritic sediments, and the three-fold Lochkovian subdivision was mainly based on the occurrences and evolution within the genus (VALENZUELA-RÍOS & MURPHY 1997). After the subsequent introduction of the genus *Lanea* (MURPHY & VALENZUELA-RÍOS 1999), part of the previous *Ancyrodelloides* stock was included in the new genus *Lanea*, and these two genera form the basis for the detailed biostratigraphical subdivision of the middle part of the Lochkovian.

The Pyrenean set of sections exhibit the most complete and sequential record of these two genera and of the other relevant taxa afore-mentioned (*Flajsella*, *Masaraella*, *Pedavis*, *Kimognathus*, *Icriodus*, and *Pelekysgnathus*) that permitted VALENZUELA-RÍOS (1994, 2011) and VALENZUELA-RÍOS & LIAO (2012) the establishment of one of the finest biostratigraphical subdivision of the interval. Recent papers by CORRADINI & CORRIGA (2012) and SLAVÍK et al. (2012) confirm the validity of this subdivision in the Carnic Alps and in the Prague Basin respectively, although the sequences in these relevant areas are currently not as complete as in the Pyrenees. For instance, as yet no specimens of *Flajsella* have been identified in the Prague Basin and the specific diversity of *Ancyrodelloides* and *Lanea* is considerably minor there. The sequence in the Carnic Alps is more complete but still less diverse for the *Ancyrodelloides*, *Masaraella* and *Pedavis* genera.

Presence of more endemic genera (*Icriodus* and *Pelekysgnathus*) together with the more cosmopolitan faunas mentioned above supports interfacial correlations (benthonic-neritic) with other relevant areas as Celtiberia, Brittany and Morocco.

In summary, the combined study of sections in three key regions, Pyrenees, Carnic Alps and Barrandia, shows consistent stratigraphical distribution of cosmopolitan taxa, supporting the establishment of a fine biostratigraphical subdivision upon which detailed correlations can be based. This study amplifies the middle Lochkovian correlation net as well.

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## Conodonts around middle/upper Lochkovian (Lower Devonian) from the section Segre 4 in the Central Spanish Pyrenees

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The current conodont-based three-fold Lochkovian subdivision places the middle/upper boundary with the entry of *Masaraella pandora* beta (VALENZUELA-RÍOS & MURPHY 1997, MURPHY & VALENZUELA-RÍOS 1999). Below the boundary, specimens of the last part of the *Ancyrodelloides* and *Lanea* lineages characterized the upper part of the middle Lochkovian, the *trigonicus-pandora* beta Zone of VALENZUELA-RÍOS & MURPHY (1997). Above the boundary, the radiation of *Pedavis* and *Masaraella* distinguishes the lower half of the upper Lochkovian. This basic succession, together with sequential entries of members of *Flajsella* and *Kimognathus* conforms a detailed biostratigraphical net that allowed VALENZUELA-RÍOS & MURPHY (1997) the establishment of sound and exhaustive correlations between several Pyrenean and Nevadan sections for the middle Lochkovian. Since then, new sections have been investigated in the Pyrenees aiming at testing and improving this correlation proposal. One of these sections is Segre 4. Consequently, the two main goals of this report are to present the conodont succession from the section Segre 4 and to integrate it in the global Lochkovian correlation net. Additionally, we want to increase the knowledge on the regional spatial and time-distribution of relevant taxa.

The section Segre 4 is located in the west limb of a faulted anticline, while the section Segre 1 formed the east limb (VALENZUELA-RÍOS 1994, VALENZUELA-RÍOS & MURPHY 1997, MURPHY & VALENZUELA-RÍOS 1999, VALENZUELA-RÍOS & LIAO 2012). Due to tectonics, the thickness of the Segre 4 is considerably smaller, and the time-span is also limited to the upper part of the middle Lochkovian and the lower part of the upper Lochkovian. Therefore, this section is adequate for studying in detail the middle/upper Lochkovian boundary.

The conodont record is composed of 19 taxa grouped into eight genera: *Ancyrodelloides*, *Lanea*, *Kimognathus*, *Masaraella*, *Pedavis*, *Pelekysgnathus*, *Icriodus* and *Wurmiella*. The succession of occurrences permits identification of three zones that span the middle/upper Lochkovian boundary. The oldest one is the *trigonicus-pandora* beta Zone that is represented in Beds 1a-27. Besides the index of the zonal lower boundary (*A. trigonicus*) other four species of the genus are present (and became extinct) in this zone, *A. transitans*, *A. asymmetricus*, *A. kutscheri* and *A. murphyi*. Within this zone it is noteworthy the sequential occurrence of three species of *Lanea*, *L. omoalpha*, *L. eoeleanorae* and *L. telleri*. It is also notable the joint acme of *A. transitans* and *Icriodus angustoides alcoleae* in Bed 15, as this "acme-record" is also common in other Pyrenean sections and at comparable stratigraphical position. Finally the records of *Kimognathus limbacarinatus* (Bed 14a) and *Pelekysgnathus serratus gadarramensis* (Bed 27) are also important.

The base of the upper Lochkovian and of the *pandora beta-gilberti* Zone coincides with the entry of *M. pandora* beta in Bed 28. Within this zone (Beds 28-31) is remarkable the lowest records of four *Pedavis* species: *P. breviramus*, *P. pesavis*, *P. robertoi* morphs 1 and 2 of VALENZUELA-RÍOS (1994).

The entry of *P. gilberti* in Bed 32 indicates the beginning of the *gilberti-steinachensis* Zone. This zone is, however, only partly represented as a fault above Bed 33 disrupts the sequence.

In summary, the rich conodont record of Section Segre 4 permits the recognition of three Lochkovian zones *trigonicus-pandora* beta Zone, *pandora beta-gilberti* Zone and *gilberti-steinachensis* Zone, and the identification of the middle/upper Lochkovian boundary in Bed 28. The conodont record allows integration of this section into the global correlation net, and it is comparable with records of the main North American and European sections.

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## **The classic New York Devonian: Toward a new stratigraphic synthesis and chronostratigraphic chart**

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The classic Devonian succession of New York State (U.S.A.) comprises a key global reference section, referred to by geological researchers worldwide. Since publication of RICKARD'S (1975) New York Devonian correlation chart, various high-resolution stratigraphic analyses, employing sequence-, bio-, event-, and chemostratigraphic approaches, sometimes at bed-by-bed scale, have led to many new understandings of the succession. Combined with a more refined Devonian geochronology, including age dates from airfall volcanic tephra (K-bentonite) beds in New York, the authors are working toward publication of a new Devonian stratigraphic synthesis for New York.

The new publication will include a series of large graphic charts and accompanying text that outline the current understanding of New York Devonian stratigraphy. The overall New York Devonian rock succession will be plotted on two x-y charts, with geography along the x-axis. The y-axis of one will plot equal spacing for Lower, Middle and Upper Devonian strata, as used for the international Devonian Correlation Table. On the other, the y-axis will utilize a recent Devonian time scale. Intervals with high resolution stratigraphic data will be split out into separate charts, where finer detail will be visible. In addition, one or more additional charts will provide data on the timing and duration of various key geological and biological events reported from New York Devonian strata. These will include sequence stratigraphic third and fourth order cycles, changes in sediment composition/petrology, events in the Acadian orogeny as interpreted from foreland basin sediments and airfall volcanic tephra, major and minor biological extinctions and other global bioevents, ecological–evolutionary faunas, and faunal epiboles.

Many of the post-1975 changes in the stratigraphy and correlations occur within Givetian, Frasnian, and upper Eifelian strata. Additional revisions occur within the Lochkovian and Emsian. There is little to no change in some intervals, in part due to little significant stratigraphic work on them since RICKARD (1975).

A number of significant gaps remain in the knowledge of New York's Devonian record. A partial list includes: Devonian terrestrial strata; correlations between the terrestrial and marine strata; Upper Devonian "Chemung"-type shelf/ramp facies; Famennian (upper Upper Devonian) rocks in southwestern New York; the Pragian Stage (middle Lower Devonian), which is conformable only in the Port Jervis region of southeast New York; and biostratigraphy through numerous intervals of the Devonian succession (including palynological analyses).

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## **The Devonian and Carboniferous of Lena River Delta (Arctic Russia, Yakutia, Northern Kharaulakh)**

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Detailed lithological and biostratigraphic studies of key Upper Paleozoic sections (Upper Devonian, Carboniferous, and Permian) of the Northern Kharaulakh Ridge were carried out for the first time in the last several years.

Upper Paleozoic sediments compose almost all coastal outcrops on the right bank of the lower reaches of the Lena River and of a number of islands of its delta. Overturned to the west, fold structures of the western slope of the Kharaulakh Anticlinorium crop out. The anticlinal folds of the Kharaulakh Anticlinorium are built up of Devonian and Carboniferous deposits, and, in its synclines, Permian deposits. The studied sequences are tectonically highly dislocated and characterized by the presence of large monoclinal blocks and of blocks with complex, isoclinal-folded structures.

Earlier (ABRAMOV & GRIGORIEVA 1983, 1986; BOGUSH et al. 1965, BOGUSH & YUFEREV 1966) age interpretations of the allocated Late Paleozoic stratigraphic units were mostly based on brachiopods and foraminifers, whereas almost all stage boundaries of modern stratigraphic scales are defined by conodonts. Therefore, the study of microfossils (conodonts, radiolarians, ostracods), along with brachiopods and ammonoids, is a priority.

Among sections that are fundamental for the understanding of the Paleozoic stratigraphy of the Northern Kharaulakh are the Stolb Island Section and the so-called Southern Section.

Stolb Island is located in the delta of the Lena River, 2.2 km west of the Crest-Toomsa Cape, near the opening of the Olenyok, Bolshaya Trofimovskaya and Bykovskaya ducts. The Stolb Island Section was described along the north, west and south-western parts of the island. By provisional data, Late Devonian sequences are supposed to include the Frasnian/Famennian boundary interval.

The Southern Section is exposed along the right bank of Lena River extending for more than 10 km. It begins at 15.6 km south-southeast of Crest-Toomsa Cape, between the mouths of Taba-Bastakh-Yurege and Kysam Rivers. From north to south, almost continuous Carboniferous (complete) and Permian (up the Tatarian Stage) sequences including stratotypes of substantially all formations of the North Kharaulakh Ridge are exposed. The Carboniferous succession is represented by a number of formations: Bastakh (Tournesian), Atyrdakh (Visean), Tiksi (Visean-lower Bashkirian), Tugasir (Bashkirian-Gzhelian), lower Unguokhtakh (Gzhelian).

Lithological and complex (brachiopods, ostracods, ammonoids, conodonts, radiolarians, foraminifers) paleontological data for the Late Devonian and Carboniferous were obtained for the studied sequences. Sampling was made during the field trip of 2012, and most samples are still under study. Conodonts and radiolarians were found for the first time.

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## Evolution of Devonian brachiopods from the Altai-Sayan folded area and the Saltation Theory of Speciation

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During the revision of the taxonomy and stratigraphic ranges of the Devonian brachiopods in the western part of the Altai-Sayan Folded Area, the author faced a number of aspects that are difficult to explain by the "Darwinian" or by the synthetic theory of evolution. These contradictions could be removed by considering the phylogeny of Devonian brachiopods from the position of the Saltation Theory of Speciation.

In Devonian biostratigraphic charts of the Asian parts of Russia and neighboring regions for over 100 years units such as the "Conchidiella Beds" [beds with *Zdimir= Conchidiellabaschkiricus* (VERN.)], "Elythyna Beds" (beds with *Elythyna salairica* RŽON.), "Cheehiel Horizon" [deposits with *Euryspirifer pseudocheehiel* (HOU)], "Stringocephalus Beds" [strata with *Stringocephalus burtini* (DEFRANCE) and a number of closely related species and genera], "Phalaena Horizon" (deposits with *Anathyris supraphalaena* KHALF.), "Cyrtospirifer Beds" (beds with a lot of morphologically similar species of *Cyrtospirifer* NALIVKIN), "Monstrum Horizon" (strata with *Anathyrella monstrum* KHALF.), "Praelonga or Meisteri Beds" [beds with *Mesoplica praelonga* (SOW.) or the morphologically similar *Mesoplica meisteri* (PEETZ) and others] have been used. All of these biostratigraphically different units of different ranks are united by one fact: they contain species with wide to subglobal distribution. In fact, these species are biomarkers for the stratigraphic correlation of highly ranked units, often corresponding to the duration of stages and series of the Devonian System.

In addition to their extensive areas of distribution, many of these species are unified by the presence of an unusually wide range of intraspecific variations. This is logical - for the conquest of large areas species should have an ecological flexibility and a high adaptation potential to external, varying in time and space, environmental factors. This applies to many, but not all taxa, because for some there are problems of uncertain systematics and nomenclature.

So, we have a number of species (in some cases, for formal reasons, groups of closely related species) with huge areas and pronounced genetic polymorphism (which is directly correlated with the morphological variability). Within the theory of neo-Darwinism, according to the concept of quantum speciation, the analyzed species simply should have become a generator of many monomorphic species, formed on the basis of peripheral populations of ancestral species. But we see an opposite picture: all the analyzed species and groups of species are disappearing, leaving no closely related descendants, and are, thus, the final chains of phylogenetic branches.

Models of allopatric (geographic) speciation are very popular in neo-Darwinism and the possibility of sympatric speciation, i.e. speciation without a spatial separation of the original population, is practically excluded. But how can we explain the presence of continuous phylogenetic lineages (lineages of closely related species) in trilobites (YOLKIN 1983), ostracods (BAKHAREV & BAZAROVA, 2004), nektonic conodonts (YOLKIN & IZOKH 1999) and other groups of fossils within one paleobasin? The lineage of some monomorphic species belonging to *Protodouvillina* HARPER & BOUCOT can be considered as an example of sympatric speciation in the Lower Devonian brachiopods of the Altai-Salair paleobasin. In Section B-819, which belongs to a relatively deep-water, mono-facies and continuous (without visible breaks) Emsian sequence (YOLKIN et al. 1986), the succession of *Protodouvillina magna* YAZ. → *P. praedistans* (GRATS.) → *P. grandicula* (GRATS.) → *P. acorrugata* GRATS. (GRATSIANOVA & YAZIKOV, 1998) has been fixed

The whole group belongs to one trophic level; species-specific characteristics changed abruptly and transitional forms are absent.

The above findings are consistent with the characteristics of the Saltation Theory of Speciation, which have been summarized by V.N. STEGNIY (STEGNIY 1993). Using an integrated approach, STEGNIY succeeded in reconstructing the phylogeny of eight closely related species of malaria mosquitoes and to compare the chromosomal organization of these species at the beginning and at the end of the phylogenetic path. As a result of many years of research, he proved the assumption of a deep systemic reorganization of the chromosomal apparatus during speciation and substantiated the following observations:

1) In phylogenetically close, adaptively radiating species complex chromosomally monomorphic species usually are initial and polymorphic - the final link;

2) Species with an inversion polymorphism of adaptive rank (i.e., providing a broad ecological and climate adaptation) have a much lower potential for speciation than monomorphic narrowly adapted species.

To the observations outlined above we can add that the appearance of polymorphic species in the first stage leads to the extinction of ancestors, and subsequently, to the extinction of entire phylogenetic lineages.

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## The Middle Devonian Global Taghanic Biocrisis in the type region, northern Appalachian Basin, USA: New Insights from paleoecology and stable isotopes

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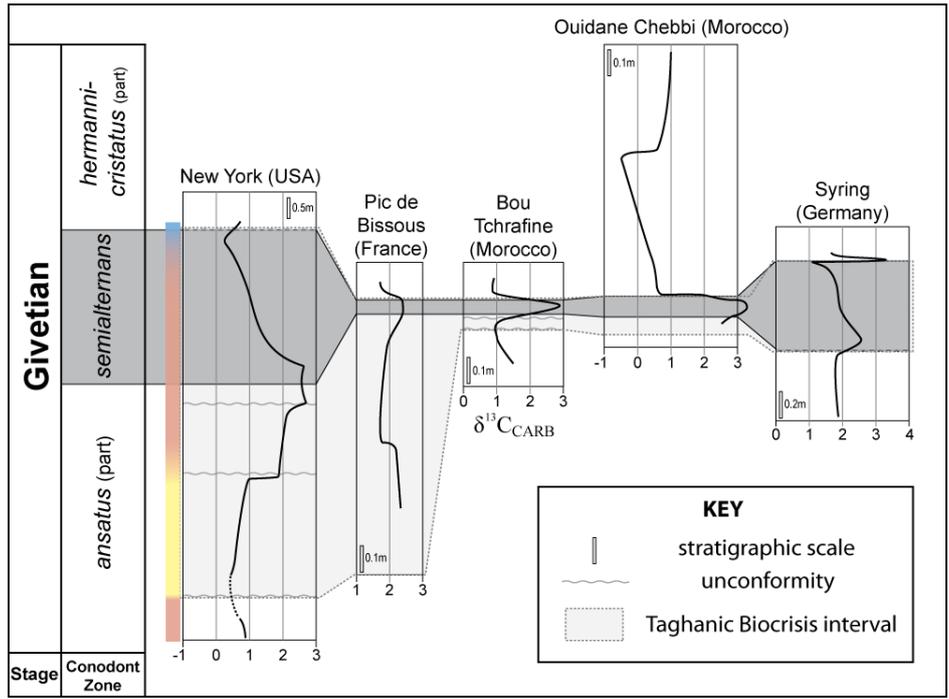
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During the Middle Devonian (latest Givetian) Taghanic Biocrisis, numerous taxa worldwide underwent a major extinction (HOUSE 2002, ABOUSSALAM 2003). This biocrisis also affected the composition of regional faunas, which ultimately resulted in the end of North American faunal provinciality (Johnson, 1970). Environmental changes associated with this biocrisis include episodes of warming, aridification, reduced oxygenation of epicontinental seas, and rapid eustatic sea level fluctuations (JOACHIMSKI et al. 2004, 2009; ABOUSSALAM & BECKER 2011, MARSHALL et al. 2011, BAIRD et al. 2012, ZAMBITO et al. 2012a). The Taghanic Biocrisis in the type region of the northern Appalachian Basin is recorded within the deposits of the uppermost Hamilton, Tully, and lowermost Genesee Groups over an interval of ~0.5 million years. The biocrisis includes three main sequential bioevent pulses: 1) the incursion of the warmer-water, equatorial Tully Fauna into eastern Laurentia and temporary loss of the Hamilton Fauna; 2) Tully Fauna extermination and replacement by a recurrent Hamilton Fauna, restocked from an unknown refugium; and, 3) extinction of large portions of the Hamilton Fauna and replacement by the more cosmopolitan Genesee Fauna (BAIRD & BRETT 2003, 2008; ZAMBITO et al. 2012a,b). Similar faunal patterns of incursion and recurrence have also been noted for other regions, albeit with somewhat different regional characteristics (ABOUSSALAM & BECKER 2011).

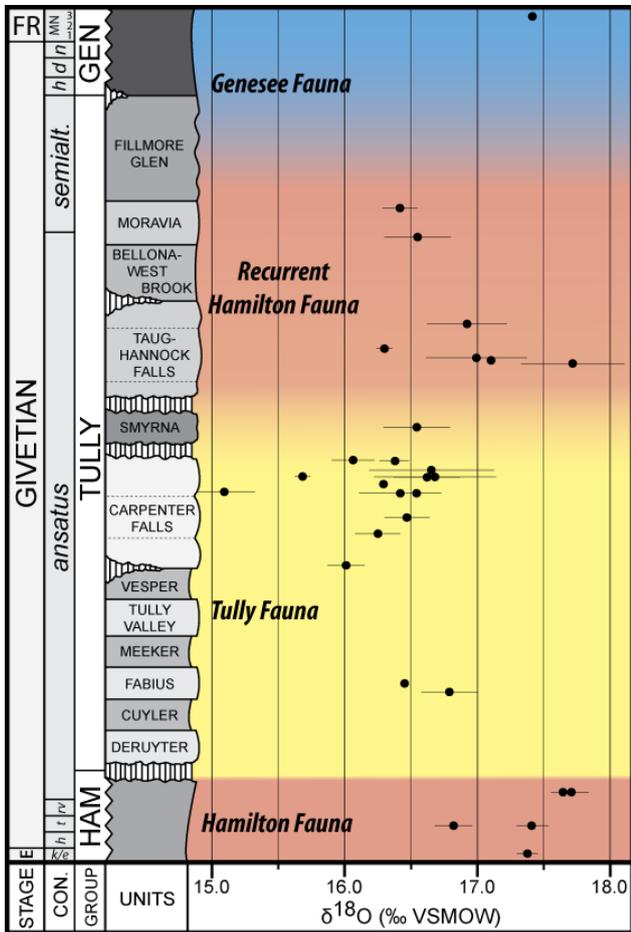
Detailed study suggests a complex interplay between global warming, sea level, and aridity, as well as local (tectonic) environmental changes. Global environmental changes during the biocrisis are reconstructed in the type region using  $\delta^{18}\text{O}$  of conodont apatite and  $\delta^{13}\text{C}_{\text{carbonate}}$ . The  $\delta^{13}\text{C}_{\text{carbonate}}$  record of the Appalachian Basin agrees well with that of other regions (Fig. 1). The Tully Fauna incursion is associated with a water temperature increase of up to 5°C, as recorded in  $\delta^{18}\text{O}_{\text{apatite}}$  (Fig. 2). Also, increased aridity at this time is suggested by faunal replacement patterns and inferred watermass dynamics (ZAMBITO et al. 2012a). During the subsequent recurrence of the Hamilton Fauna, in which only 60% of the taxa present before the Taghanic Biocrisis are observed,  $\delta^{18}\text{O}_{\text{apatite}}$  changes suggest that temperatures cooled to near pre-biocrisis conditions. The  $\delta^{13}\text{C}_{\text{carbonate}}$  record shows a stepped ~2‰ positive excursion that culminates in the lowermost *semialternans* Zone, and is associated with Hamilton faunal recurrence. This may be related to increased burial of organic carbon in oceanic settings; if so, the lowering of atmospheric carbon dioxide levels through Corg burial may have driven the observed cooling. During the third bioevent, eustatic sea-level rise was accentuated regionally by flexural basin downwarping associated with renewed Acadian tectonic activity, resulting in widespread low-oxygen conditions and an unprecedented increase in sediment flux into the Appalachian Basin. Quantitative paleoecological analysis demonstrates that the Hamilton taxa that survived to become incorporated in the post-biocrisis Genesee Fauna were generalized taxa adapted to nearshore, siliciclastic-dominated settings (ZAMBITO et al. 2012b). The  $\delta^{18}\text{O}_{\text{apatite}}$  record also suggests warming during the Hamilton to Genesee faunal transition, however, the increased sediment input and resulting thick, siliciclastic-dominated deposits, associated with the Acadian Orogeny, limits the sampling resolution of conodonts for  $\delta^{18}\text{O}_{\text{apatite}}$  analysis in this interval.

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**Fig. 1:** Biostratigraphic correlation of Taghanic Biocrisis sections with corresponding  $\delta^{13}C_{carb}$  reconstructions. New York reconstruction is from this study, all other sections adapted from ABOUSSALAM (2003). Note positive carbonate carbon isotopic excursion occurs, for the most part, within the *semialternans* Zone in all sections. Colors superimposed on New York section correspond to faunas in Fig. 2.



**Fig. 2:** Polygnathid  $\delta^{18}O_{apatite}$  values from samples before, during, and after the Global Taghanic Biocrisis from throughout the northern Appalachian Basin. Error bars are one standard deviation calculated from triplicate measurements of each sample. Samples without visible error bars means that one standard deviation is smaller than the size of the symbol used. Stratigraphic succession is a composite between Genesee and Moravia, NY, with vertical lines representing intra-Devonian erosion. Thickness of the Hamilton and Genesee groups is artificially condensed relative to the Tully Group in this stratigraphic reconstruction. Eifelian represented by 'E'. Conodont zones are as follows: *k./e.* = *kockelianus* and *ensensis* zones; *h.* = *hemiansatus* Zone; *timor.* = *timorensis* Zone; *r./v.* = *rhenanus/varcus* Zone; *semalti.* = *semialternans* Zone; *h.* = *hermanni* Zone; *d.* = *disparilis* Zone; *n.* = *norrisi* Zone; MN 1,2, and 3 = Montagne Noire Zones 1-3, respectively.



