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

FIELD GUIDEBOOK

ISBN : 0251-4249

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Organized by :
Institut Scientifique, Mohammed V University– Agdal, Rabat
International Subcommission on Devonian Stratigraphy (SDS)
International Subcommission on Carboniferous Stratigraphy (SCS)
IGCP 596 on "Climate Change and Biodiversity patterns in the Mid-Paleozoic"

in memory of Dr. Volker EBBIGHAUSEN

22nd to 29th March 2013

FIELD GUIDEBOOK

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

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Cover : Devonian succession at El Khraouia (day 2), Eastern Tafilalt (photo A. El Hassani)

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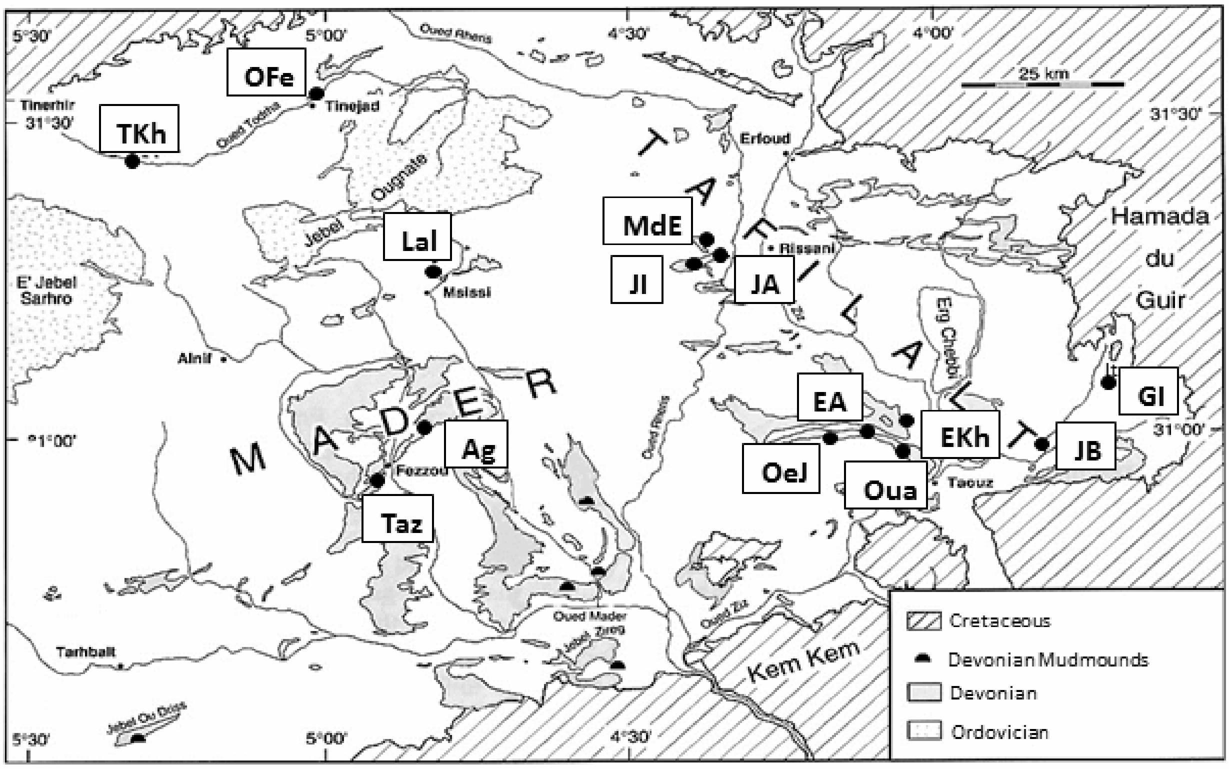
Drive to Ouarzazate

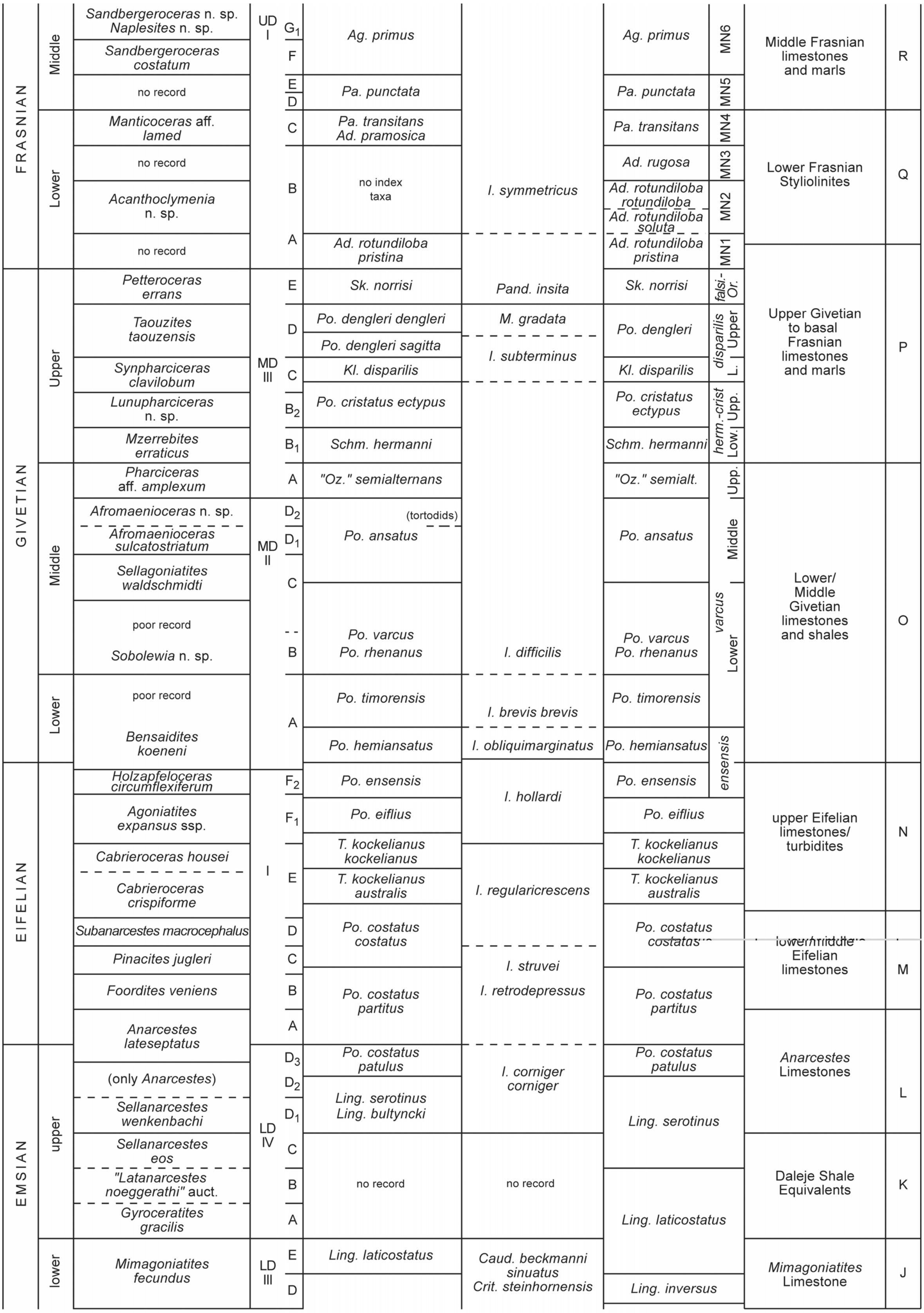
Fig. 1. Geographic position of Devonian/Carboniferous field localities in the SE Anti-Atlas:

TKh = Taourirt n´Khellil
OFe = Oued Ferkla
EKh = El Khraouia
Oua = Jebel Ouaoufilal
EA = El Atrous (South)
OeJ = Oum el Jerane
JB = Jebel Begaa region
GI = Gara El Itima region
Lal = Lalla Mimouna
Ag = Aguelmous Syncline
Taz = Tazoult
JA = Jebel Amelane
JI = Jebl Ihrs
MdE = Mdoura-East



(map base on KAUFFMANN 1998)

		Tafilalt Ammonoids	key	Tafilalt Conodont Zonations: deeper	shallow	standard zones		Lithological Units	key
FAMENNIAN	uppermost	no record	UD VI	no record	no record	*Pr. kockeli*	*praesulcata* U.	Ouaoufilal Formation	Y
		Ac. (Stockumites) n. sp. aff. *subbilobatum*	F_2			*costatus-kockeli-Interregum*	Middle		
		no record	E						
		GAP	D	GAP	GAP	*Siphonodella preasulcata*	Lower	GAP	
		Parawocklumeria paradoxa	C_2	*Bi. ultimus*	*"siphonodelloids"*			uppermost Famennian limestones and marls	X
		Effenbergia falx	C_1						
		Effenbergia lens	B						
		Linguaclymenia similis *Kalloclymenia subarmata*	A			*Bi. ultimus*	Upper		
	upper	*Medioclymenia aguelmousensis*	UD V C	*Bi. costatus*		*Bi. aculeatus*	*expansa* Middle	*Gonioclymenia* limestones and marls	W
		Gonioclymenia hoevelensis	B						
		Gonioclymenia subcarinata	A_2						
		Endosiphonites muensteri	A_1	*Bi. aculeatus*					
		Sporadoceras orbiculare	UD IV C	*Bi. stabilis stabilis*		*Bi. stabilis stabilis*	Low.	*Platyclymenia* limestones and marls	V
		Procymaclymenia pudica	B	*velifer-stabilis-Interregnum*		*Pa. gracilis manca*	*post.* U.		
		Prototyclymenia sp.				*Po. styriacus*	L.		
		Platyclymenia annulata	A			*trachytera-styriacus* Int.	*trachytera* Upp.		
	middle	*Sulcoclymenia sulcata*	UD III C_2	*Sc. velifer*		*Ps. granulosus*		middle Famennian limestones and marls	U
		Afrolobites sp.	C_1			*Pa. rug. trachytera*	L.		
		Planitornoceras euryomphalum	B			*Sc. velifer*	*marginifera* Upm.		
		Sporadoceras equale	A	*Pa. marginifera marginifera*		*Pa. marginifera utahensis*	Upp.		
		Polonites n. sp.	I H						
		Maeneceras latilobatum *Maeneceras meridionale*	G			*Pa. marginifera marginifera*	Low.		
	lower	*Paratonoceras lentiforme*	UD II F	*Pa. rhomboidea*		*Pa. rhomboidea*	*rhomboidea* Upp.	dark-grey lower Famennian limestones and marls	T
		Ch. (Compacteras) rotundum *Praemeroceras dahmanii* *Praemoceras petterae*	E			*Pa. poolei*	Low.		
		Paratorleyoceras aff. *globosum*	D	*Pa. glabra pectinata*		*Pa. gl. pectinata*	*crepida* Um		
		Ch. (Cheiloceras) subpartitum	C	*Pa. glabra prima*		*Pa glabra prima*	Upp.		
		Ch. (Compactoceras) verneuili	B						
		Phoenitixes frechi	A	*Pa. termini*		*Pa. termini*	M		
				Pa. tenuipunctata		*Pa. crepida*	L		
						Pa. tenuipunctata	*triang.* U		
				Pa. clarki		*Pa. clarki*	M		
				Pa. triangularis		*Pa. ultima*	L		
FRASNIAN	Upper	*Crickites holzapfeli* *Crickites rickardi*	UD I L	*Ag. ubiquitus*	*I. alternatus*	*Pa. linguiformis*	MN 13 b/c	Kellwasser Facies	S
		Archoceras varicosum	K	*Pa. bogartensis*		*Pa. bogartensis*	a		
		Manticoceras carinatum	J	*Pa. winchelli*		*Pa. winchelli*	MN 12		
		Sphaeromanticoceras rhynchostomum	I	*Pa. feisti*		*Pa. feisti*	MN 11		
	Middle	*Beloceras tenuistriatum*	H	*Pa. plana*	*I. symmetricus*	*Pa. plana*	MN 10	Middle Frasnian limestones and marls	R
				Pa. ljaschenkoae		*Pa. proversa*	MN 9		
		Mesobeloceras kayseri	G_{2b}	*Ag. coeni*		*Pa. housei*	MN 8		
		Mesobeloceras housei	G_{2a}	*"0z." nonaginta*		*"0z." nonaginta*	MN 7		


FRASNIAN
Middle
Lower
GIVETIAN
Upper
Middle
Lower
EIFELIAN
EMSIAN
upper
lower
Sandbergeroceras n. sp.
Naplesites n. sp.
Sandbergeroceras costatum
no record
Manticoceras aff. lamed
no record
Acanthoclymenia n. sp.
no record
Petteroceras errans
Taouzites taouzensis
Synpharciceras clavilobum
Lunupharciceras n. sp.
Mzerrebites erraticus
Pharciceras aff. amplexum
Afromaenioceras n. sp.
Afromaenioceras sulcatostriatum
Sellagoniatites waldschmidti
poor record
Sobolewia n. sp.
poor record
Bensaidites koeneni
Holzapfeloceras circumflexiferum
Agoniatites expansus ssp.
Cabrieroceras housei
Cabrieroceras crispiforme
Subanarcestes macrocephalus
Pinacites jugleri
Foordites veniens
Anarcestes lateseptatus
(only Anarcestes)
Sellanarcestes wenkenbachi
Sellanarcestes eos
"Latanarcestes noeggerathi" auct.
Gyroceratites gracilis
Mimagoniatites fecundus
UD I
G1
F
E
D
C
B
A
MD III
E
D
C
B2
B1
A
MD II
D2
D1
C
B
A
I
F2
F1
E
D
C
B
A
LD IV
D3
D2
D1
C
B
A
LD III
E
D
Ag. primus
Pa. punctata
Pa. transitans
Ad. pramosica
no index taxa
Ad. rotundiloba pristina
Sk. norrisi
Po. dengleri dengleri
Po. dengleri sagitta
Kl. disparilis
Po. cristatus ectypus
Schm. hermanni
"Oz." semialternans
(tortodids)
Po. ansatus
Po. varcus
Po. rhenanus
Po. timorensis
Po. hemiansatus
Po. ensensis
Po. eiflius
T. kockelianus kockelianus
T. kockelianus australis
Po. costatus costatus
Po. costatus partitus
Po. costatus patulus
Ling. serotinus
Ling. bultyncki
no record
Ling. laticostatus
I. symmetricus
Pand. insita
M. gradata
I. subterminus
I. difficilis
I. brevis brevis
I. obliquimarginatus
I. hollardi
I. regularicrescens
I. struvei
I. retrodepressus
I. corniger corniger
no record
Caud. beckmanni sinuatus
Crit. steinhornensis
Ag. primus
Pa. punctata
Pa. transitans
Ad. rugosa
Ad. rotundiloba rotundiloba
Ad. rotundiloba soluta
Ad. rotundiloba pristina
Sk. norrisi
Po. dengleri
Kl. disparilis
Po. cristatus ectypus
Schm. hermanni
"Oz." semialt.
Po. ansatus
Po. varcus
Po. rhenanus
Po. timorensis
Po. hemiansatus
Po. ensensis
Po. eiflius
T. kockelianus kockelianus
T. kockelianus australis
Po. costatus costatus
Po. costatus partitus
Po. costatus patulus
Ling. serotinus
Ling. laticostatus
Ling. inversus
MN6
MN5
MN4
MN3
MN2
MN1
falsi- Or.
disparilis Upper L.
herm.-crist Upp. Low.
Upp.
Middle
varcus
Lower
ensensis
Middle Frasnian limestones and marls
Lower Frasnian Styliolinites
Upper Givetian to basal Frasnian limestones and marls
Lower/ Middle Givetian limestones and shales
upper Eifelian limestones/ turbidites
lower/middle Eifelian limestones
Anarcestes Limestones
Daleje Shale Equivalents
Mimagoniatites Limestone
R
Q
P
O
N
M
L
K
J

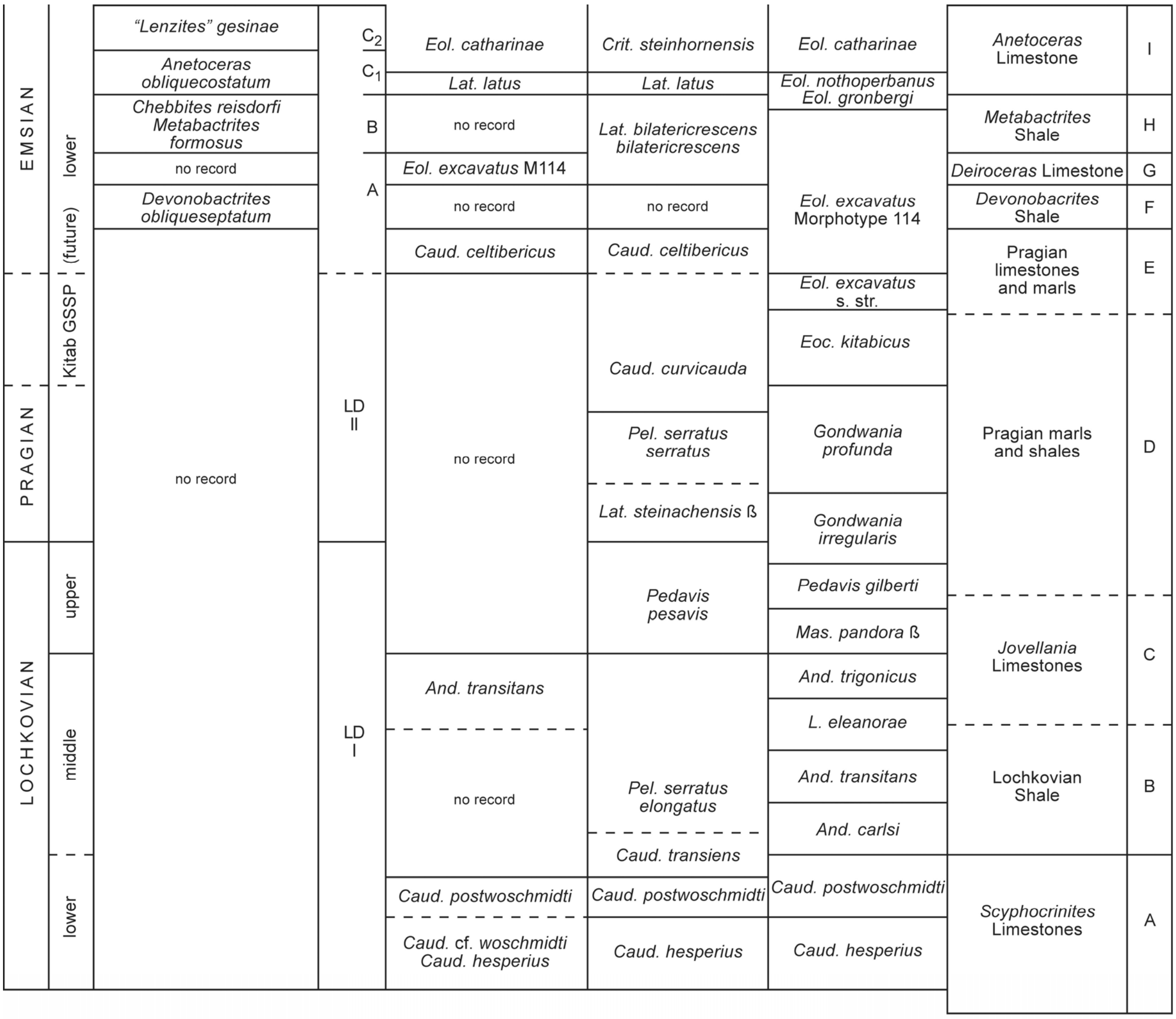


Fig. 2. Stratigraphic chart for the Devonian of the Tafilalt: chronostratigraphy, regional ammonoid stratigraphy in comparison with the international zonal key (LD, D, UD = Lower, Middle, Upper Devonian), regional deeper and shallow-water conodont successions (the latter still incomplete in the Famennian) in comparison with the international « standard », and lithostratigraphical intervals (Units A to Y) that should be recognized in a future formal formation/member terminology.

Introduction to the Fieldtrip

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Structural Overview

The geology of Morocco offers one of the more complete geological histories in the world; it shows sediments from the Archean to the Quaternary. It is, therefore, of major interest for the international scientific community.

Morocco has also several attractions, apart from its beautiful landscape, such as the Devonian mud-mounds of Hamar Lakhdad in the eastern Tafilalet, the peridotites of Beni Bouzera in the Rif belt, and so on. The majority of Moroccans, including managers of natural areas, especially the non-specialized scientific community, nowadays often ignores these resources. This encourages studies of a type "inventory", which geologists rarely indulge, but it has a fundamental role in raising public awareness and for the protection, enhancement, and wise use of this heritage.

Among the many values of geological sites is the currently prevailing commercial value in SE Morocco: It generates a massive exploitation or abuse of fossils and mineral wealth that will inevitably lead to their demise in the short or medium term. A flourishing trade (national and especially international) developed at the expense of fossil and mineralogical rarities. This situation will undoubtedly observed during our meeting in the Tafilalt.

During its geological history, Morocco has experienced significant marine and continental sedimentation, from the oldest Precambrian time (Archean, uo to 3.8 Billion years ago) until the Present. Very early signs of life on Earth are also known in Morocco, by the first appearance of eukaryotes in the Precambrian "P III" of silicified limestone, lacustrine hummocky stromatolitic constructions, described as *Collenia* (s.l.) or *Conophyton*. These are the oldest fossils observed in Morocco (~ 555 Ma).

The meeting and fieldtrip associated with the Annual Meeting of the International Subcommission on Devonian Stratigraphy, in association with the Subcommission on Carboniferous Stratigraphy, and the IGCP 596 on “Climate Change and Biodiversity patterns in the Mid-Palaeozoic” will be held in the Tafilalt area, which constitutes the eastern part of the Anti Atlas of Morocco. This meeting is devoted to the richly fossiliferous Devonian and Lower Carboniferous of the region, with some focus on the boundary between both systems.

The field meeting will start south of the High Atlas at what is considered as the *South front of the Meseta Domain: Tinerhir Carboniferous olistostrome and thrust faults,* then will continue through the Middle Palaeozoic of the Eastern Anti-Atlas, what is commonly named Tafilalt-Maider area.

There, we will observe the characteristic Quaternary gravel covering much of this arid area. All the sections that we will cross will be mostly Palaeozoic in age. Cretaceous outcrops preserved along the belt, which border to the East and South of the area, constitute the Hammadas and generally lie unconformably on the Palaeozoic. Structures were developed during the late Carboniferous orogeny and levelled before the Cretaceous transgression.

The Triassic and Jurassic crop out at the northern border of the Anti Atlas (Jurassic rocks are also represented by doleritic sills and dykes, which penetrated into the folded and/or faulted Palaeozoic sequences). Nice Triassic sections can be observed north of the main road just after passing the town of Tinghir and, of course, can be observed in the distance in the elevated mountains of the High Atlas. Later deformation is represented by the Cretaceous folds of Kem-Kem and by isolated folds affecting the Hammadas Pliocene cover. They are the result of the Alpine orogenic deformation, which, in the North of the Anti Atlas, produced the Rif and Atlas belts. Unconformities in the region have been used to subdivide the Atlas belt and to trace the main tectonic Atlasic events.

Summarized stratigraphy of the Palaeozoic series

The Anti Atlas domain consists of Precambrian cores belonging to the West African Craton. They are generally metamorphic, crystalline, and have recorded several tectonic phases, in particular the Pan-African Orogeny (600 Ma). At this time, erosion produced a major conglomerate lying just beneath the Adoudounian limestones (early Cambrian age in its upper part).

In the Anti Atlas, shales and sandstones, the “Grès terminaux” containing trilobites, end the early Cambrian. The Middle Cambrian (1000 to 2000 m of thickness) is represented by "*Paradoxides* shales" and by sandstones with *Conocoryphe* of Tabanite (we will observe them on the last day in the area of Alnif). These sandstones form a very regular relief, underlining the structures and wrapping around the Anti Atlas along its southern margin. In the area of Agdz (between Ouarzazate and Zagora), exposures of Late Cambrian shales and associated beds, immediately preceding the Ordovician, were dated by trilobites (DESTOMBES&FEIST1987).

The Ordovician is particularly well represented in the Anti Atlas by siliciclastic sediments. The system is divided into four groups:

- External "Feijas" group
- First Bani group
- Ktaoua Group
- Second Bani Group

They are often quite distinct in the landscape because of their different resistance to erosion (Appalachian topography).

During the Silurian, Gondwana migrated towards the north, and this is characterised by a change in the type of sedimentation and by the development of warm water faunas. The Silurian is characterised by black nautiloid-rich limestones ("*Orthoceras* Limestone") intercalated with graptolite shales.

The Devonian outcrops conformably overlie the Silurian, without any gap and without any notable change in the type of sedimentation. It is a period, which shows a clear differentiation of facies within the Anti Atlas. So it becomes necessary to distinguish between the Western zones, the Draa Valley, and eastern areas, Tafilalt and Maider. A detailed overview of Devonian research in these areas is developed in the next chapters and will be seen during the fieldtrip.

Concerning the Carboniferous, the best sections are those of the western Anti Atlas (Draa Valley), where we can follow sections for tens of kilometers. Nevertheless, there are important Carboniferous outcrops in the eastern Anti Atlas, too. We can observe a stratigraphic continuity with the Devonian. There is certainly a change in the type of sediments that passes from a Devonian limestone dominance to a prevalence of clay and quartzites during the Carboniferous. The Devonian-Carboniferous boundary is one of the main goals of our meeting and, surely, detailed explanations will be given during the fieldtrip, with special focus on this boundary.

Regarding the deformations in the Anti Atlas, the Precambrian structures and, in particular, "Panafrican's" were reactivated several times during the Paleozoic sedimentation and the Variscan compression.During the Hercynian orogeny, old faulting, who guided partly the sedimentation, replay.

Despite a now long research history in the eastern Anti Atlas and a wealth of recent publications, there is still much more to discover, to observe, and to sample. It is certain that the participants will find new fossils and their expertise will add to our knowledge. The rocks, their fossils, the landscape, and the Moroccan people should make our field symposium a remarkable experience

THE ALLOCHTHONOUS SILURIAN-DEVONIAN IN OLISTOSTROMES AT "THE SOUTHERN VARISCAN FRONT" (TINERHIR REGION, SE MOROCCO) – PRELIMINARY DATA

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1. INTRODUCTION

The Lower/Middle Palaeozoic of Morocco belongs to two strongly different plate tectonic realms: the mildly deformed Anti-Atlas at the northwestern margin of the stable Gondwana Craton, and the Moroccan Hercynides at the southwestern margin of the Variscides, which were strongly tectonized during the middle Carboniferous final collision of Gondwana and Laurussia. The main suture between both is formed by the South Atlas or Tizi n´Test Fault. Eastwards, in the Tinerhir area, a southernmost Variscan thrust belt consists of allochthonous sediment slices, which were pushed southwards onto the autochthonous Lower Palaeozoic of the Jebel Saghro (MICHARD et al. 1982). FERONI et al. (2010) developed for this "Southern Variscan Front" a model of transpressional tectonics and nappe stacking.

Prior to the deposition of thick Upper Tournaisian to Upper Visean shallow-shelf clastics (HINDERMEYER 1954, GRAHAM & SEVASTOPULO 2008) of the Jebel Asdaf and Jebel Tisdafine, a steep palaeoslope must have formed at the northern margin of the Gondwana autochthon. It received large olistostromes during an interval of block faulting, with massive erosion on uplifted areas. The size of olistoliths varies strongly. The lower part consists of conglomerates/breccias with small-to median sized (up to a few decimeters), rounded or angular clasts. In the upper part, house-sized limestone glide blocks should not be mistaken as autochthonous Devonian units bound by faults. The timing of the major regional interval of Eovariscan tectonics can be inferred from the age of the youngest eroded sediments and of the oldest overlying Carboniferous strata.

The locality to be visited is Taourirt n´Khellil (= Rhellil on the topographic map, Figs. 1, 3), sheet Alnif, 1/100 000, Feuille NH-30-XIX-2. It lies ca. 1.5 km north of Aït Issa (= Ait Aïssa), in the southern plain below the Jebel Tisdafine. It includes the hill to the south and isolated limestone blocks (samples OL) just north of the road between Tinerhir (= Tineghir, ca. 20 km to the west) and Tinejdad (= Tindjdad, ca. 30 km to the east). GPS coordinates are N31° 26' 49,344" W5° 20' 57,948". Coming from Tinerhir the locality is easily marked by the high telecommunication masts on the following hill and by the subsequent branch of the road to Alnif.

Fig. 1. Hill with the main exposure of the conglomeratic-brecciated Taourirt n´Rhellil Formation, N of Ait Issa and SW of Tinerhir.

Fig. 2. Conglomerate/breccia with Devonian limestone clasts at Taourirt n´Khellil.

The existence of Devonian blocks within a Carboniferous "conglomerate" at Khellil (Fig. 2) was first noted by CLARIOND & TERMIER (1933), who observed blocks with Givetian corals. Subsequently, HINDERMEYER (1955) listed trilobites, brachiopods, nautiloids, ammonoids, and corals that indicated clasts of upper Emsian/Eifelian, Givetian, and Frasnian age. On the geological map, sheet Todrha-Ma´der (DESTOMBES et al. 1988), the Khellil outcrop is mapped as an isolated eastern occurrence of the "Flysch schisto-greseux (S. de Tinerhir)" but the Middle/Upper Devonian age (dm-s) refers to the age of the clasts, not to the re-deposition age. There is an unclear distinction between clast and re-sedimentation ages on the revised geological maps for sheet 1/50 000 Imtir (Tikkedarine Formation with partly large blocks versus the similar olistoliths in the Ait Yalla Formation, SCHIAVO et al. 2007) and Taghazout (massive conglomerate of Taourirt n´Rhellil Formation versus the olistoliths in the Ait Yalla

Formation, DAL PIAZ et al. 2007). GRAHAM & SEVASTOPULO (2008) were obviously unaware of the new lithostratigraphic terminology on the revised maps and assigned the conglomerate to the base of their new Isfoul Member of a new Jebel Tisdafine Formation. The latter is recognized at group level on the revised maps and the Isfoul Member is a synonym of the Ait Yalla Formation.

We attempted to record as many as possible different lithofacies that were reworked and re-deposited in the olistrostromes. Some samples (AI OL2) were taken from a smaller outcrop to the east. The analysis of mikrofacies and biostratigraphy aimed to reconstruct the facies development in the now completely eroded source area at the southern Variscan suture. A complete reconstruction, however, is difficult since thin, condensed units and unconsolidated shales/marls had a low preservation potential.

Palaeogeographic clues come from the comparison of faunas and facies with the adjacent autochthonous Silurian/Devonian of the Tinejdad region to the east and the (northern) Maider and Tafilalt to the SE/ESE. The latter regions were supposedly separated in the Devonian by a partly emergent Rheris-Ougnate Platform (KAZMIERCZAK & SCHRÖDER 1999) or Ougnate High (DÖRING 2002). In this context it is significant that some clasts yielded completely new faunal records for southern Morocco. But our data suggest that there was no significant barrier between the Devonian of the Tinerhir region and the main Anti-Atlas at least until the end of the Middle Devonian. It is likely that the eroded cover of the Jebel Sahgro-Jebel Ougnate formed the olistostrome source.

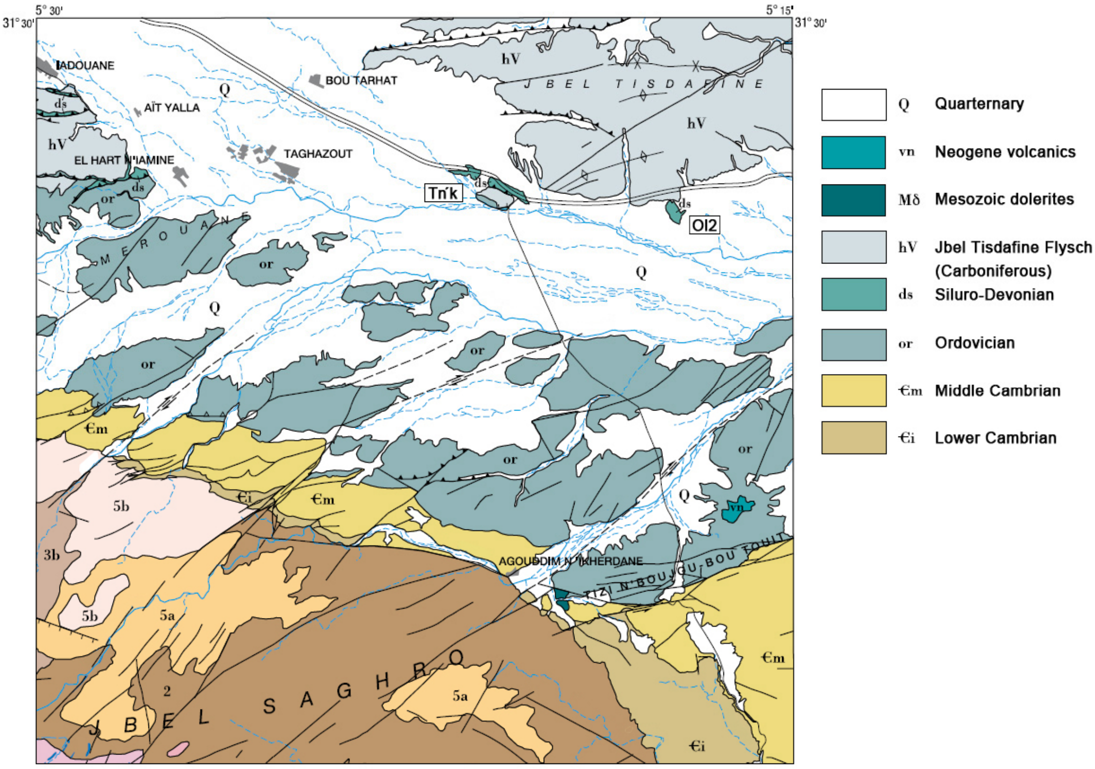

Fig. 3. Geological overview of the area SW of Tinerhir, between the Carboniferous at the Jebel Tisdafine and the Ordovician of the Jebel Saghro, showing the position of the allochthonous Silurian/Devonian locality at Taourirt n´Khellil (Tn´K) and of small other Devonian outcrops to the west and east (OL2). Redrawn from DAL PIAZ et al. (2007: fig. 3).

2. FACIES TYPES AND THEIR AGES

A. Pelagic platform/ramp facies

A1. Orange/brownish weathering micrite (Sample AL 2011-1)
Mikrofacies: Strongly bioturbated mudstone with few ostracods and variable, very fine Fe-content.
Fauna: Common agglutinating foraminifers (*Tolypammina*).
Age: Unclear (no conodonts).

A2. Yellowish weathering micrite (Sample AI Ol2 P2)
Mikrofacies: Bioclastic wackestone with dacryoconarids, shell filaments, crinoid remains, ostracods (partly with geopetal filling), trilobites, and (secondarily altered) large pyrite grains.
Fauna: *Eoctenopolygnathus pireneae* (first record for southern Morocco), *Caudicriodus celtibericus, Caud. curvicauda*.
Age: Upper Pragian, *pireneae* or *celtibericus* Zone. The direct association of both marker species supports the range in the type Praha Limestone (e.g., SLAVIK

2004) but is in conflict with the younger entry of *celtibericus* in Spain (CARLS et al. 2002).

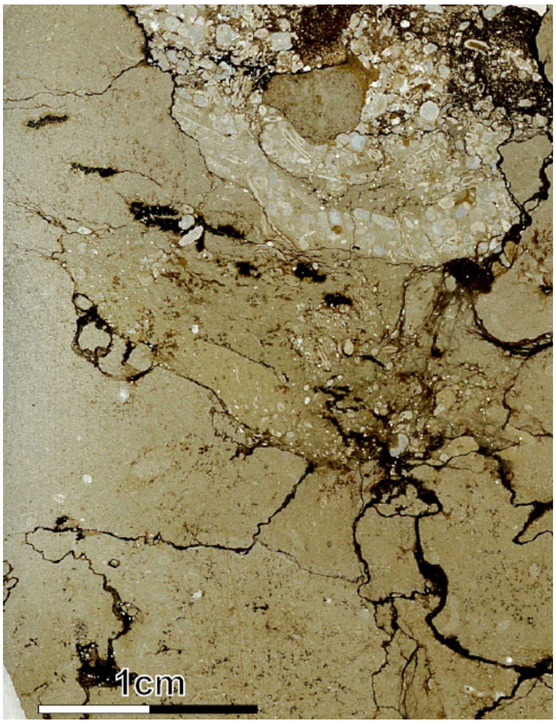

Fig. 4. Facies A3 (Sample AI K)

A3. Reddish/greenish micrite (Sample AI K, Fig. 4)
Mikrofacies: Mudstone, truncated by a two-phased channel with crinoid wackestone/packstone; abundant iron mineralisations on dissolution seams.
Fauna: Dominant *Palmatolepis perlobata schindewolfi* and *Polygnathus nodocostatus nodocostatus*, associated with *Pa. perlobata maxima* (Pl. 2, Fig. 4), *Pa. rugosa* aff. *ampla* (Pl. 2, Fig. 5), *Pa. glabra distorta* (Pl. 2, Fig. 6), *Pa.* cf. *marginifera* (Pl. 2, Fig. 7, intermediate to *quadrantinodosa inflexa*), *Po. nodocostatus ovatus* (Pl. 2, Fig. 9), *Po. fallax* (Pl. 2, Fig. 8), *Po. pseudotenellus, Po. semicostatus* “Central Morphotrend” (Pl. 2, Fig. 11), *Po. pennatuloideus* (Pl. 2 , Fig. 10), *Scaphignathus velifer velifer* (Pl. 2, Fig. 13), *“Icriodus” cornutus* Gp. (Pl. 2, Fig. 12), and others (17 taxa).
Age: Middle Famennian. The association of *Sc. velifer velifer* and *Po. pseudotenellus* suggests the *velifer* Zone (= Uppermost *marginifera* Zone) but transitional forms between *Pa. quadrantinodosa inflexa* and *marginifera* normally occur much earlier (basal *marginifera* Zone). The record of *Po. fallax* confirms its upper range extension to the *velifer* Zone in DREESEN & DUSAR (1974).

A4. Brownish-grey dacryoconarid limestone (Sample AI Ol Top2, Fig. 5)
Microfacies: Alternating peloid mudstones, partly microsparitic, and current-oriented dacryoconarid grainstones with sparitic matrix.
Fauna: *Criteriognathus steinhornensis* (abundant).
Age. Lower Emsian, *steinhornensis* Zone.

Fig. 5. Facies A4 (Sample AI Ol Top2, 1 = peloids)

A5. Light-grey to reddish dacryoconarid limestone (Sample AI L17)
Microfacies: Strongly bioturbated to intraclastic dacryoconarid packstone with some pyrite in microsparitic matrix and with many subrounded to angular clasts of micritic dacryoconarid wackestone.
Fauna: *Eolinguipolygnathus gronbergi* (Pl. 1, Fig. 5), *Crit. steinhornensis* (Pl. 1, Fig. 3), *Crit. miae* (Pl. 1, Fig. 8), *Latericriodus beckmanni beckmanni* (Pl. 1, Fig. 6), *Lat. bilat. bilatericrescens* (Pl. 1, Fig. 7), and *Caud. sigmoidalis* (Pl. 1, Fig. 4).
Age: Lower Emsian, higher *gronbergi* or *steinhornensis* Zone.

A6. Reddish to medium-grey, sparitic dacryoconarid limestone (Sample AI Ol2 P9)
Microfacies: Bioturbated dacryoconarid wacke- and packstone with numerous sparite fenestrae.
Fauna: *Crit. steinhornensis*, *Crit. miae, Lat. bilatericrescens bilatericrescens*.
Age: Lower Emsian, lower part of *steinhornensis* Zone.

A7. Reddish dacryoconarid-*Metabactrites* Limestone (Sample E-Tin 2; very rare)
Microfacies: Dacryoconarid packstone with fragmentary goniatites, fine hematite, and partly sparitic matrix.
Fauna: *Metabactrites* sp., no conodonts.
Age: Lower Emsian, with a very primitive, monospecific ammonoid assemblage (LD III-B).

A8. Grey to brownish, micritic “*Orthoceras* Limestone” (Sample AI Br+Ort)
Microfacies: Orthocone packstone/rudstone with ostracods, few crinoid remains, and shell filaments in

micritic matrix. The cephalopods are partly filled by geopetals.
Fauna: *Ancyrodelloides carlsi* (Pl. 1, Figs. 1-2). KRÖGER (2008) identified orthocones of this interval as *Plagiostomoceras* and *Hemicosmorthoceras*.
Age: Basal middle Lochkovian, *carlsi* Zone (see revised zonation in CORRADINI & CORRIGA 2012).

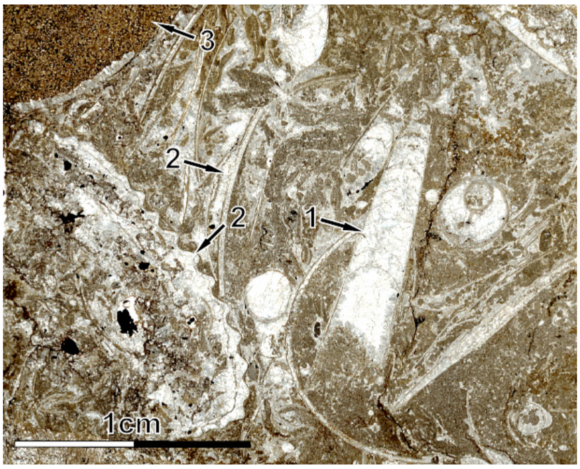

Fig. 6. Facies A9 (Sample E-Tin 10, 1 = small orthocone, 2 = ribbed or smooth bivalves, 3 = rounded extraclast).

A9. Dark-grey, bioclastic “*Orthoceras* Limestone” (Sample E-Tin 10, Fig. 6)
Microfacies: Orthocone rudstone with gastropods, ostracods, large bivalves, other shell filaments, microsparitic extraclasts, early diagenetic pyrite and micritic, microsparitic or peloidal matrix. Geopetal fillings are common; the orthocones are often current-orientated.
Fauna: *Polygnathoides siluricus* (Pl. 1, Figs. 14-15). According to KRÖGER (2008), *Temperoceras, Arionoceras*, and *Kopaninoceras* are associated.
Age: Upper Ludlow, lower Ludfordian, *siluricus* Zone.

A10. Middle-grey, micritic goniatite limestone (Sample AI L21)
Microfacies: Bioclastic wackestone with goniatites, orthocones, bivalves, shell filaments, brachiopods, trilobites, crinoid ossicles, and fine pyrite.
Fauna: *Maeneceras meridionale meridionale, Armatites planidorsatus, Buchiola* sp., dominant *Pa. glabra lepta* Late Morphotype (Pl. 2, Fig. 17), common *Pa. perlobata schindewolfi*, and associated *Pa. quadrantinodosa inflexoidea, Pa. glabra pectinata* Morphotype 2 (Pl. 2, Fig. 15) and specimen with trend to *glabra acuta* (Pl. 2, Fig. 16), *Po.* cf. *vetus* (Pl. 2, Fig. 19), *Polylophodonta gyratilineata* (Pl. 2, Fig. 21), *Polylo. confluens* (Pl. 2, Fig. 20), *Polylo.* cf. *triphyllata* (Pl. 2, Fig. 22), *Po. fallax* (Pl. 2, Fig. 18), *Bispathodus stabilis vulgaris* (Pl. 2, Fig. 14), and others (17 taxa).
Age: Basal middle Famennian, *meridionale* Zone (UD II-G), (Lower) *marginifera* Zone, based on the association of *Pa. quadrantinodosa inflexoidea*, the Late Morphotype of *Pa. glabra lepta*, and *Polyl.* cf. *triphyllata*. The oldest *Bi. stabilis* are commonly supposed to enter in the Upper *marginifera* Zone (DREESEN & DUSAR 1974, ZIEGLER & SANDBERG 1984). Our older specimens confirm the mostly overlooked downwards range extension by CAPKINOGLU (2005).

A11. Reddish, micritic goniatite limestone (Sample AI 2-5)
Microfacies: Bioturbated cephalopod wackestone with sparite-filled goniatites and orthocones, gastropods, ostracods, and shell filaments in a micritic to peloidal matrix.
Fauna: Goniatites indet., *Pa. glabra lepta* Late Morphotype, *Pa. glabra prima* M3, *Pa. glabra pectinata* Morphotype 2, *Pa. perlobata schindewolfi, Pa. gracilis gracilis,* and *Polygnathus* sp.
Age: Basal middle Famennian, *marginifera* Zone.

A12. Dark-grey, fine crystalline cephalopod limestone (Sample AI Clym)
Microfacies: Bioturbated bioclastic wackestone with goniatites, clymeniids, trilobites, bivalves, brachiopods, shell filaments, and crinoid remains in a mostly microsparitic matrix.
Fauna: *Platyclymenia annulata, Platy.* cf. *subnautilina, Trigonoclymenia* cf. *protacta, Protactoclymenia* sp., ?*Prionoceras* sp., *Erfoudites ungeri, Guerichia* sp., dominant *Bi. stabilis vulgaris* (Pl. 2, Fig. 1) and *Pa. gracilis gracilis*, in association with *Branmehla inornata, Br. ampla, Po. perplexus, Pa. gracilis manca* (Pl. 2, Fig. 3), *?Clydagnathus* sp. (Pl. 2, Fig. 2), and others (11 taxa).
Age: The ammonoids are typical for the *annulata* Zone (UD IV-A). The first record of *Pa. gracilis manca* for Morocco (see HARTENFELS 2011, p. 45) prove at least the *gracilis manca* (= Upper *postera*) Zone. The questionable clydagnathid is a form with advanced ornamentation, typical for an even higher level. This gives a correlation with at least UD IV-B, but probably with the part below *Procymaclymenia* (see Aguelmous Syncline, KORN et al this vol.).

B. Neritic facies

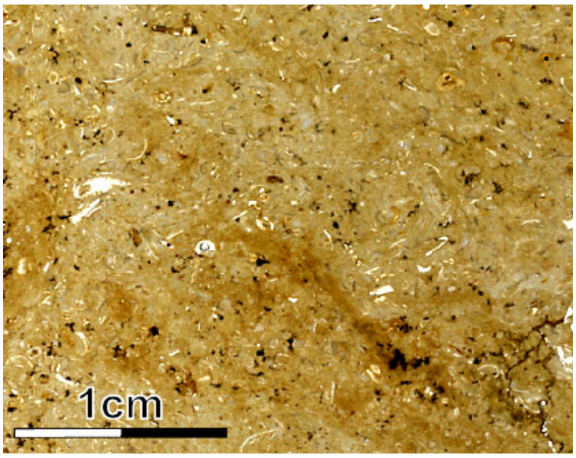

Fig. 7. Facies B1 (Sample E-Tin 1).

B1. Yellowish gastropod limestone (Sample E-Tin 1, Fig. 7)
Microfacies: Strongly bioturbated bioclastic wackestone with many gastropods, ostracods, shell filaments, some trilobites, and fine pyrite in micritic matrix.
Fauna: *Po. zieglerianus* (Pl. 1, Figs. 12-13), *Linguipolygnathus cooperi* (Pl. 1, Fig. 11).
Age: Lower Eifelian, *partitus* to basal *costatus* Zones.

B2. Yellowish-grey bioclastic limestone (Samples AI 2011-3 and AI Ol Top1)
Microfacies (2011-3): Partly recrystallized, strongly bioturbated bioclastic wackestone to packstone/rudstone with many dacryoconarids, gastropods, small crinoid ossicles, tabulate corals, and fine pyrite in micritic to coarse sparitic matrix.
Fauna: *Caudicriodus sigmoidalis* (2011-3), *Caud. celtibericus* (Ol Top 1), *Crit. steinhornensis* (Ol Top 1), abundant branching, *Cladochonus*-type tabulate corals (2011-3).
Age. Lower Emsian, *steinhornensis* Zone (Ol Top1).

B3. Laminated, yellowish-grey brachiopod-gastropod limestone (Sample E-Tin 11)
Fauna: Strophomenids, *Straparollus* sp., *Flajsella schulzei, Wurmiella tuma* (new records for southern Morocco).
Age: Upper part of middle Lochkovian, *eleanorae* to *trigonicus* Zones (see CORRADINI & CORRIGA 2012).

B4. Bluish-grey bioclastic limestone with crinoids (Sample E-Tin 6)
Microfacies: Crinoid grainstone with dacryoconarids (in specific layers), partly dolomitized and with sparite matrix.
Age: Unclear (probably Lower Devonian).

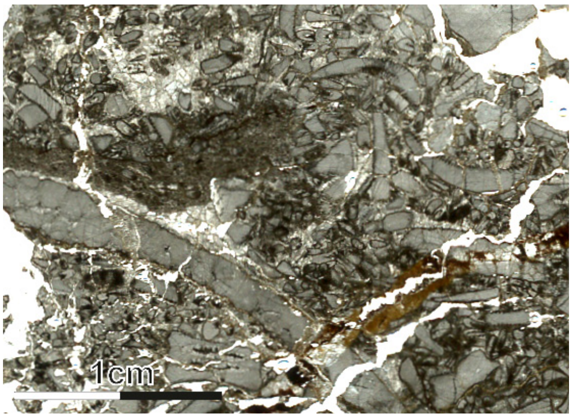

Fig. 8. Facies B6 (Sample AI L7Var) with lobolite fragments.

B5. Brownish-grey, medium-grained crinoidal limestone (Sample AI 3-5)
Microfacies: Thin-bedded alternation of fine, silty dacryoconarid-crinoid grainstone with microsparite matrix and crinoid-dacryoconarid wackestone with brownish micrite/microsparite matrix.
Age: Unclear (probably Lower Devonian, no conodonts).

B6. Coarse-grained, light to medium grey crinoidal limestone (Samples AI 2011-2 and Al L7Var, Fig. 8)
Microfacies: Coarse, recrystallized and partly dolomitized crinoidal packstone (2011-2) or coarse crinoidal pack-rudstone (L7Var) with small pockets of micrite matrix.
Fauna (AI L7Var): *Pseudooneotodus beckmanni*, *Panderodus* sp., *Caud. woschmidti, Caud.* cf. *woschmidti* (transitional to *postwoschmidti), Zieglerodina remscheidensis,* silicified ostracods, proetid remains, tabulate corals, crinoid lobolites.
Age: Basal Lochkovian (*woschmidti* Zone, AI L7Var).

B7. Coarse-grained, dark grey oolite (Samples AI LI)
Microfacies: Grainstone with many dolomitized ooids, crinoid ossicles and variably-sized, rounded micritic extraclasts, few tabulate corals, agglomerates of rounded quartz grains, and some early diagenetic pyrite.
Age: Unclear (no conodonts).

B8. Sandy, dark-grey crinoidal limestone with brachiopods (Sample E-Tin 4)
Microfacies: Silty crinoid grainstone with micritized crinoid ossicles, brachiopods, corals, gastropods, and micrite extraclasts; micritic matrix partly preserved but mostly washed out.
Age: Unclear (no conodonts).

B9. Reddish, coarse-grained crinoidal limestone (Sample AI L7)
Microfacies: Coarse crinoid grainstone with partly reworked thin interlayers of crinoidal wackestone and with iron-rich micrite matrix.
Fauna: *Caud. postwoschmidti, Ziegl. remscheidensis.*
Age: Upper part of lower Lochkovian, *postwoschmidti* Zone.

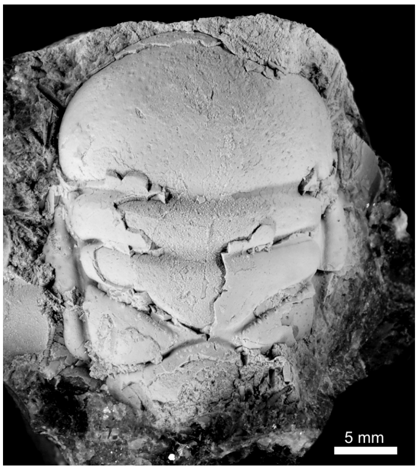

Fig. 9. *Cheirurus* (*Crotalocephalina) gibbus auster* from facies B10 at Taourirt n´Khellil, a typical Pragian species of both the Tafilalt and the Meseta.

B10. Grey, coarse-grained crinoid-trilobite limestone (Sample E-Tin 9)
Microfacies: Rudstone with many crinoids, dacryoconarids, some trilobites, orthocones, and gastropods; micritic matrix partly washed out and replaced by sparite.
Fauna: *Cheirurus (Crotalocephalina) gibbus auster* (Fig. 9), *Pragoproetus tafilaltensis, Eremiproetus*? sp., *Reedops maurulus*, odontopleurids, scutelluids, *Platyceras* sp., *Orthonychia* sp., and other gastropods.
Age: Pragian.

B11. Reddish, sparitic limestone with branching Tabulata (Sample E-Tin 3)
Microfacies: Grainstone with strongly micritized crinoid ossicles and gastropods; matrix sparitic.
Fauna: Common tabulate coral with divergent polypar bundles.
Age: Unclear (probably Lower Devonian).

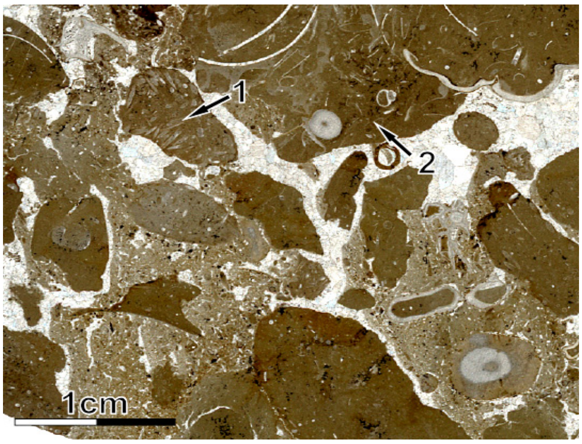

Fig. 10. Facies B12 (Sample AI SP FL, 1 = dacryoconarid packstone intraclast, 2 = wackestone intraclast).

B12. Yellow-brownish, sparitic limestone breccia (Sample AI SP FL, Fig. 10)
Microfacies: Limestone breccia with subrounded to subangular brownish clasts of mud-/wackestone, bioturbated dacryoconarid packstone, bioturbated bioclastic wackestone with gastropods, trilobites, and crinoid ossicles, sitting in a matrix of coarse orthosparite or of light-grey, partly washed out, microsparitic wackestone with crinoid debris, tabulate corals, dacryoconarids, ostracodes, and silt.
Fauna: *Wurmiella wurmi* (Pl. 1, Figs. 9-10).
Age: Middle to upper Lochkovian (see CORRADINI & CORRIGA 2012).

C. Reefal facies
C1. Grey brachiopod-coral limestone (Sample E-Tin 13)
Fauna: Atrypids and solitary Rugosa.
Age: Probably Givetian (no conodonts).

C2. Dark-grey *Thamnopora*-stromatoporoid limestone (Sample E-Tin 12, Fig. 11)
Microfacies: Float-/rudstone with tabulate corals, solitary Rugosa, stromatoporoids, trilobites, and crinoid debris in a micritic, peloidal or sparitic matrix.
Fauna: *Thamnopora* sp., *Stringophyllum (Stringophyllum)* sp. (det. A. MAY), *Ling. linguiformis, Ling. klapperi, Po. pseudofoliatus.*
Age: Probably Lower Givetian.

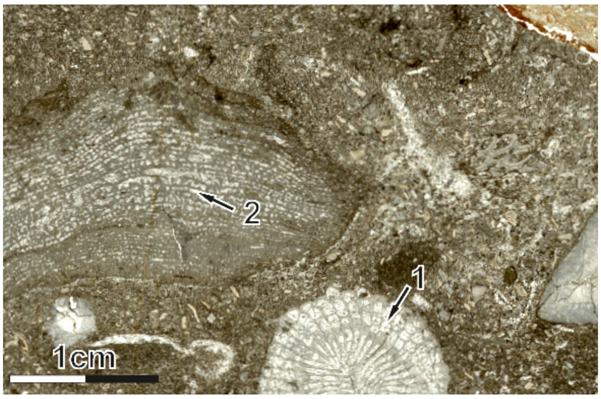

Fig. 11. Facies C2 (Sample E-Tin 12, 1 = *Stringophyllum (Stringophyllum)* sp., 2 = stromatoporoid).

C3. Dark-grey coral limestone (Sample AI L20)
Microfacies: Coral-brachiopod float-/rudstone with tabulate and solitary rugose corals, fragmented crinoid ossicles, gastropods, and micritic extraclasts sitting in a mostly micritic to peloidal matrix.
Fauna: *Thamnopora* sp., *Po. limitaris*.
Age: Upper Givetian (see range of *Po. limitaris* in ABOUSSALAM & BECKER 2007).

C4. Grey, micritic rugose coral limestone (Sample E-Tin 5)
Microfacies: Coral floatstone with tabulate corals, rounded clasts of strongly recrystallized colonial rugose corals, and some crinoid debris sitting in a mostly micritic to peloidal matrix
Fauna: *Phillipsastrea* sp., *Alveolites* sp., favositids, *Po. ordinatus*.
Age: Upper Givetian; *Po. ordinatus* ranges in the Tafilalt from the *dengleri sagitta* Subzone to the *norrisi* Zone (ABOUSSALAM & BECKER 2007).

D. Ordovician quartzite (very common)

3. RECONSTRUCTION OF THE FACIES HISTORY AND PALAEOGEOGRAPHIC RELATIONSHIPS

The analyzed clast population shows a very distinctive distribution of facies types in specific time intervals, with a surprising record lack for several episodes.

Silurian

A very common to partly dominant facies is the Ludfordian "*Orthoceras* Limestone" (Facies A9), which occurs all over the Anti-Atlas, from the western Dra Valley to the eastern Tafilalt (LUBESEDER 2008), and beyond. This marker of a significant upper Silurian regression episode has been re-named after its characteristic orthoconic genus as *Temperoceras* Limestone (KRÖGER 2008). The lack of any other Silurian clasts can be easily explained by its predominant shale/marl facies.

Lochkovian

The equally wide-spread *Scyphocrinites* Limestones (e.g., HOLLARD 1977, LUBESEDER 2008) are represented by coarse crinoidal limestones (Facies B6 and B9), partly with fragmented plate lobolites (Fig. 8). They span most of the lower Lochkovian but uppermost Silurian samples, as known from the Tafilalt (e.g., HAUDE & WALLISER 1998), are lacking.

The basal middle Lochkovian “*Orthoceras* Limestone” (Facies A8) can be correlated with a poorly known marker limestone of the Tafilalt (Bed PK in KRÖGER 2008). It marks a second, short, eutrophic interval of sea-level fall in the Lochkovian. Almost nothing is known about the Lochkovian of the northern Maider and northern Tafilalt. FRÖHLICH (2004) noted the presence of Lochkovian/Pragian limestones at the northern tip of the Jebel Rheris but provided no details. This prevents any detailed comparison. This is equally true for the somewhat younger brachiopod-gastropod limestone (Facies B3) and limestone breccia (Facies B12). They indicate for the middle/upper Lochkovian of the olistostrome source area the presence a shallow, neritic carbonate platform with high-energy reworking events. This setting contrasts with the pelagic, outer-shelf *Jovellania* Limestones of the southern Tafilalt (HOLLARD 1977, KRÖGER 2008).

Pragian

The trilobite-rich Facies B10 and the younger bioclastic limestones of Facies A2 suggest a drowning of the reconstructed shallow carbonate platform towards the end of the Pragian. The trilobite assemblage combines typical taxa of the Pragian Limestone of the Tafilalt with faunal elements of the Meseta (HELLING & BECKER, Abstract Volume). But the coarse and high-energy deposition of Facies B10 gives evidence for a continuing shallower environment than in most Tafilalt areas (apart from the Hamar Laghdad Mudmounds). The complete lack of *pireneae* faunas elsewhere in the Anti-Atlas underlines the distinction of the source area.

Emsian

The lower Emsian is mostly represented by various dacryoconarid limestones (Facies A4 to A7), which partly build the decameter-sized blocks north of the road and at locality Ol2. They document a shallow pelagic to outer neritic carbonate ramp that existed for a long time. Wackestones with many neritic faunal elements (Facies B2) and crinoidal limestones (Facies B6 and B9) indicate a range into the realm with crinoid forests and storm-influenced sedimentation. The general facies assemblage is closely similar to the Bou Tiskaouine Formation of the northern Maider (HOLLARD 1981) and to the contemporaneous *Deiroceras* and *Anetoceras*-*Mimagoniatites* Limestones of the Tafilalt. An exception is the unique block of reddish Dacryoconarid-*Metabactrites* Limestone, which has no known equivalents to the E/SE.

The absence of upper Emsian clasts at Taourirt n´Khellil is surprising, since this interval is thick and, in its upper part, well represented by nodular, pelagic limestones in all adjacent regions. Equivalents of the *Anarcestes* Limestone of the Tafilalt are well developed in the Tinejdad area (HINDERMEYER 1955, WARD et al., this volume) and in the northern Maider (FRÖHLICH 2004). A poorly preserved goethtic/limonitic anarcestid suggests that the upper Emsian of the source region was (partly) too marly for re-deposition as solid clasts.

Eifelian

The neritic limestone of Facies B1 reflects an Eifelian regressive trend in the source region but the record of *Pinacites* in HINDERMEYER (1955) gives evidence for the persistence of pelagic faunal elements. This mixture suggests palaeogeographic affinities with the northern Maider (HOLLARD 1974) whilst the Tinejdad region and northern Tafilalt show a more pelagic setting. FRÖHLICH (2004) showed that the upper Eifelian is incomplete at the Jebel Rheris and a corresponding trend may explain the restricted representation of Eifelian clasts at Taourirt n´Khellil. It is also possible that some of the conodont-free crinoidal limestones are Eifelian in age. More sampling may clarify this question. The change from hemipelagic to neritic or even reefal facies in the Eifelian of the northern Maider provides the evidence for a rising Rheris-Ougnate Platform N of the Maider Basin (SCHRÖDER & KAZMIERCZAK 1999).

Givetian

All dated biostromal limestones at Taourirt n´Khellil fall in the Lower to Upper Givetian. The presence of conodonts in the reefal facies proves an open marine faunal influx and the absence of a reef barrier. This gives close similarities with the biostromes of the northern Maider, especially of the Jebel Rheris (FRÖHLICH 2003). In the Ouihlane area, the known successions end in the Lower Givetian (KAZMIERCZAK & SCHRÖDER 1999). There are no Givetian biostromes in the northern Tafilalt or Tinejdad area.

Frasnian

The lack of Frasnian olistoliths is as surprising as the absence of upper Emsian clasts. However, the record of *Manticoceras* and “*Eobeloceras*” (probably = *Naplesites*) by HINDERMEYER (1955) provides firm evidence that a Middle Frasnian pelagic facies was developed in at least parts of the source region. This indicates similarities with the Middle Frasnian of the northern Tafilalt but the blackish Lower and Upper Frasnian limestones of that region have not yet been observed. The Frasnian of the northern Maider has partly been eroded during an upper Famennian regression and is best preserved in the eastern part of the Jebel Rheris (FRÖHLICH 2004). There, the shallow carbonate platform drowned in the Lower Frasnian, but neritic limestones, minor last biostromes, conglomerates and high siliciclastic input point to an inner-ramp, near-shore setting, with calcareous and clastic debris coming from the N. This is the best evidence for synsedimentary tectonics and the further rise of an erosive Ougnate High, which started to separate the Tinerhir and northern Maider regions. Its northward extension and non-sedimentation may have contributed to the rarity of Frasnian clasts. The Middle Frasnian N of Tinejdad is developed in a very different, microsparitic distal debris flow facies that is almost devoid of fossils.

Famennian

So far there is no evidence for lower Famennian clasts at Taourirt n´Khellil although this is in most parts of the Anti-Altlas an interval of very high sea-level with extensive pelagic sedimentation. At the Jebel Rheris of the northern Maider, however, shallow-water carbonates with distinctive conglomerates, debris flows and sandstones continue from the Frasnian (FRÖHLICH 2004). Therefore, it is possible that the deposition in the source area continued to be disrupted by the Ougnate High.

The middle and early upper Famennian (upper UD II/top UD IV) at Taourirt n´Khellil is characterized by iron-rich, condensed pelagic cephalopod limestones as in the Tafilalt. There is no record of similar facies in the northern Maider. EL BOUKHARI et al. (2007) noted Famennian limestones with *Pa. gracilis gracilis* from a small outcrop ca. 20 km to the east on sheet Taroucht. This suggests a middle/upper Famennian extension of the pelagic Tafilalt Platform to the NW, north of the Ougnate High. The facies extended further to the west since middle Famennian pelagic limestones of the *gracilis sigmoidalis* Subzone (UD III) occur also as reworked clasts in the Ait Yalla Formation of sheet Imtir, just S of Tinerhir (SCHIAVO et al. 2007).

The absence of upper/uppermost (UD V/VI) Famennian limestones is intriguing since there is a good record of shallow-water limestones of this age in the northern Maider, where they transgress partly much older strata (FRÖHLICH 2004, KORN et al. 2004, BECKER et al. 2012). This proves a drowning of southern parts of the Ougnate High. There, the Hangenberg Event Interval is represented by shales and brachiopod-rich sandstones, partly overlain by topmost Devonian (*kockeli* Zone) crinoidal limestones and subsequent pelagic Tournaisian siliciclastics with ammonoids.

4. AGE OF EOVARISCAN EROSION AND RE-SEDIMENTATION

The timing of the Eovariscan uplift event is bracketed by the so far youngest, upper Famennian clasts (top UD IV) at Taourirt n´Khellil and the Upper Tournaisian age of the lower Ait Yalla Formation at Jebel Asdaf (HINDERMEYER 1954, Ichem Member in GRAHAM & SEVASTOPULO 2008). The uplift must have been considerable since the erosion cut from deeper-water pelagic Famennian into the Ordovician. Further efforts are required to search for uppermost and Lower Tournaisian Famennian limestones or D/C boundary brachiopod sandstones. Synsedimentary tectonics created fast subsiding troughs in the southern Tafilalt and southern Maider during the Hangenberg Event Interval (KAISER et al. 2011). It is well possible that they reflect the same crust movements, which created the olistostromes of the Tinerhir region.

5. REFERENCES

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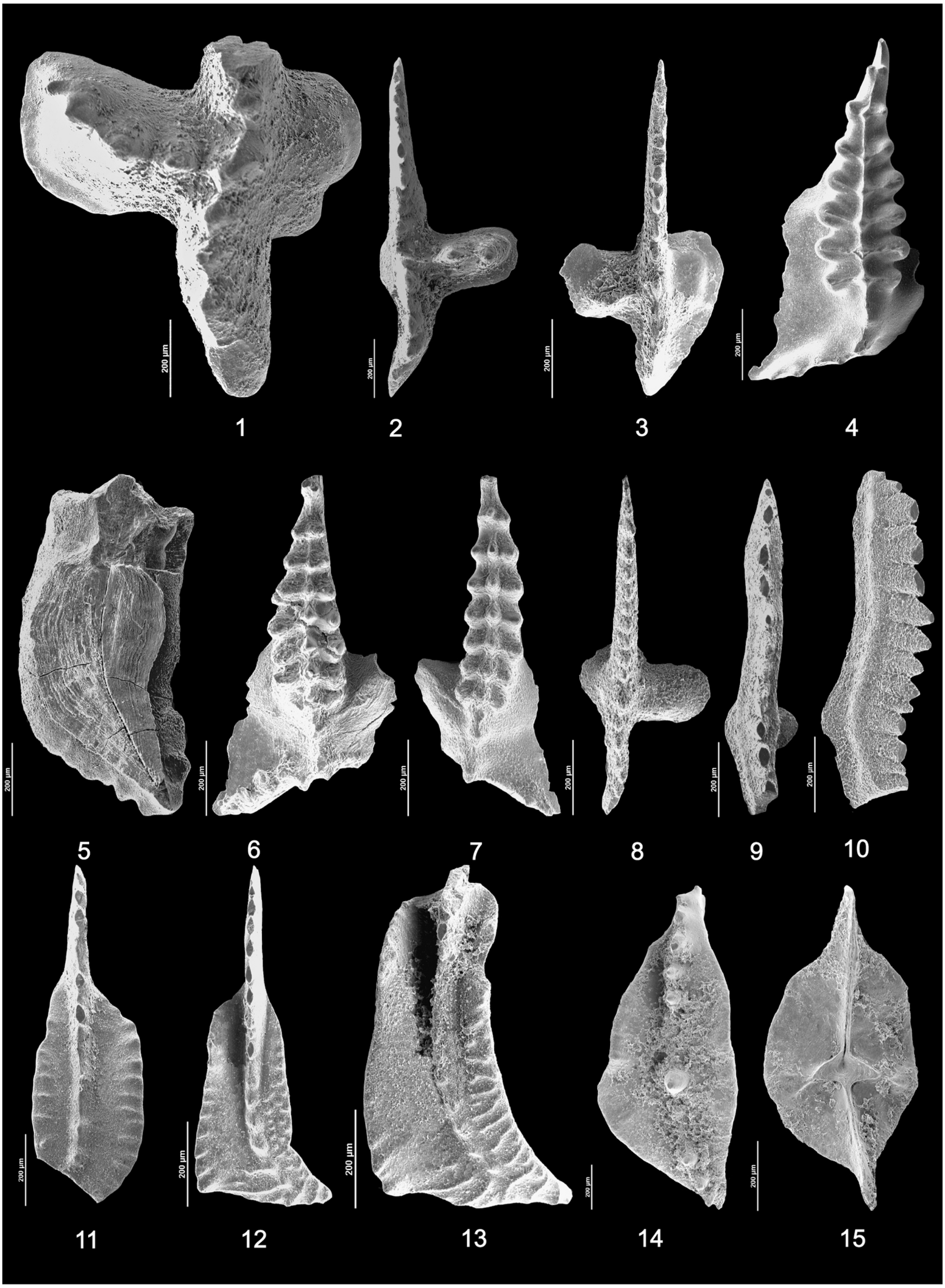
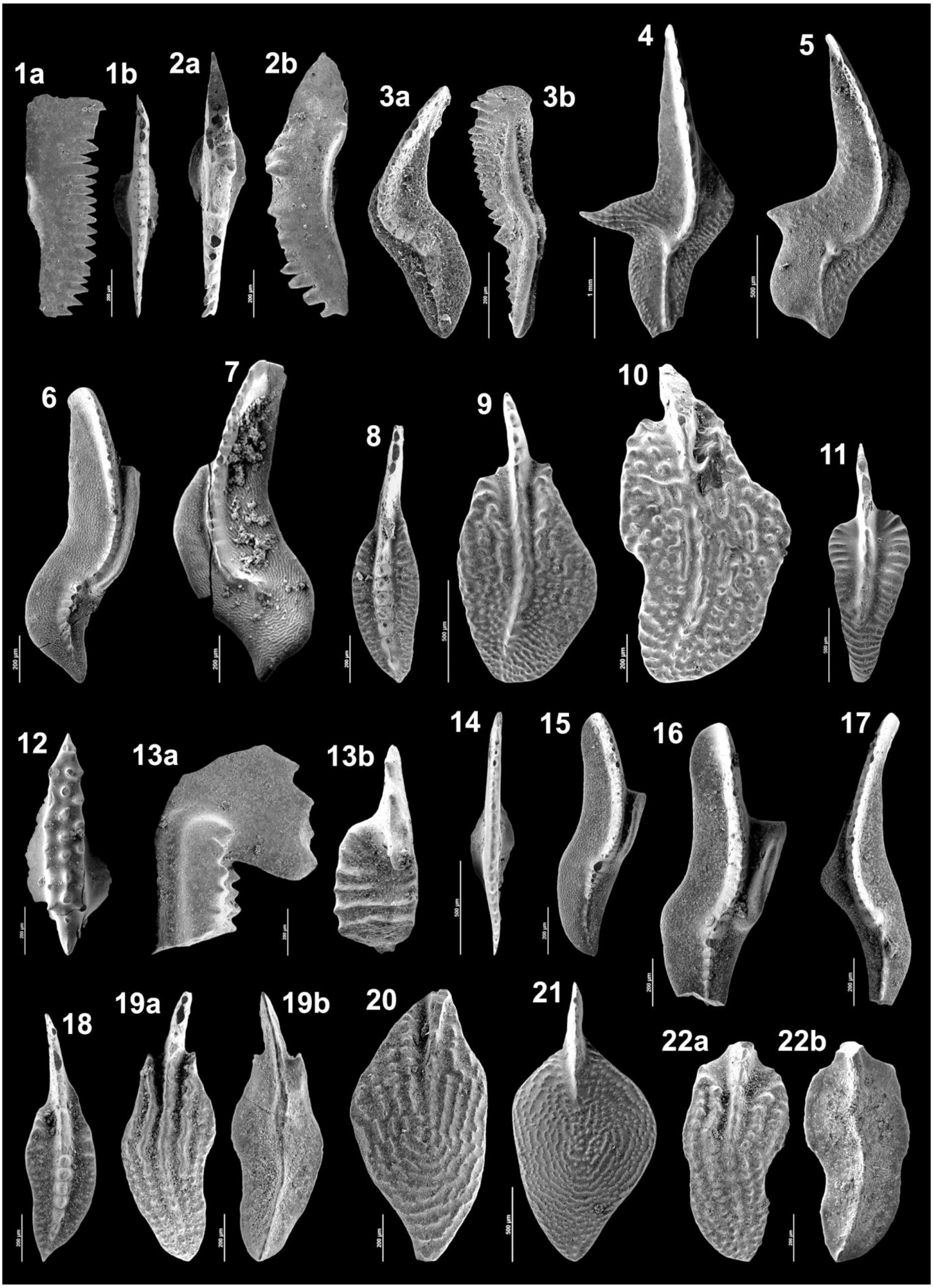


Pl. 1. Conodonts from Taourirt n´Khellil. **1-2**. *Ancyrodelloides carlsi*, Sample AI Br+Ort, basal middle Lochkovian, **3**. *Crit. steinhornensis*, Sample AI L17, lower Emsian, **4**. *Caud. sigmoidalis*, Sample AI L17, lower Emsian, **5**. *Eoling. gronbergi*, Sample AI L17, lower Emsian, **6**. *Lat. beckmanni beckmanni*, Sample AI L17, lower Emsian, **7**. *Lat. bilatericrescens bilatericresens*, Sample AI L17, lower Emsian, **8**. *Crit. miae*, Sample AI L17, lower Emsian, **9-10**. *Wurm. wurmi*, Sample AI SP FL, middle/upper Lochkovian, **11**. *Ling. cooperi*, Sample E-Tin 1, lower Eifelian, **12-13**. *Po. zieglerianus*, Sample E-Tin 1, lower Eifelian, **14-15**. *Polygnathoides siluricus*, Sample E-Tin 10, Ludfordian.



Pl. 2. Famennian conodonts from Taourirt n´Khellil. **1.** *Bi. stabilis vulgaris*, **2.** ?*Clydagnathus* sp., **3**. *Pa. gracilis manca* (all sample AI Clym, *stabilis stabilis* Zone), **4.** *Pa. perlobata maxima*, **5.** *Pa. rugosa* aff. *ampla*, **6.** *Pa. glabra distorta*, **7.** *Pa.* cf. *marginifera* (transitional from *quadrantinodosa inflexa*), **8.** *Po. fallax*, **9.** *Po. nodocostatus ovatus*, **10.** *Po. pennatuloideus*, **11.** *Po. semicostatus* “Central Morphotrend”, **12.** “*Icriodus*” *cornutus* Gp., **13.** *Sc. velifer velifer* (all sample AI K, *velifer* Zone), **14.** *Bi. stabilis vulgaris*, **15.** *Pa. glabra pectinata* Morphotype 2, **16.** *Pa. glabra pectinata* Morphotype 2 with trend to *Pa. glabra acuta*, **17.** *Pa. glabra lepta* Late Morphotype, 1**8.** *Po. fallax*, **19.** *Po*. cf. *vetus*, **20.** *Polylo. confluens* , **21.** *Polylo. gyratilineata*, **22.** *Polylo*. cf. *triphyllata* (all sample AI L21, (Lower) *marginifera* Zone).

THE DEVONIAN AT OUED FERKLA (TINEJDAD REGION, SE MOROCCO)

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1. INTRODUCTION AND LOCATION

The Oued Ferkla, a tributary to the Oued Todhra, exposes ca. 3 km N of Tinejdad an autochthonous Lower/Middle Devonian succession (Fig. 1). Coming from Tinejdad, the studied section lies just past the bridge over the Oued Ferkla at GPS N 31.53716° W 05.01069°. There are natural outcrops along the northern river bank (Fig. 2) and a road cut with a low, steep cliff (Figs. 3, 5). Following an initial survey in spring 2010, the section was investigated in detail in March 2012. This survey allocated 156 beds within a 45 meter thick sequence. It is of special palaeogeographic interest since it is separated by ca. 40 km distance from the Devonian outcrops of the northern Maider. It is part of a largely unstudied Devonian outcrop belt further to the N and NE, which forms the northernmost preserved Devonian of stable Gondwana. The area allows comparisons with the allochthonous Devonian at the southern Variscan Front of the Tinerhir area (see RYTINA et al., this. vol.) and with the main Anti-Atlas. Especially interesting is the record of Daleje, Choteč, Kačák, and the two *Pumilio* Events, ca. 50 km ENE from their nearest good outcrops of the NW Tafilalt.

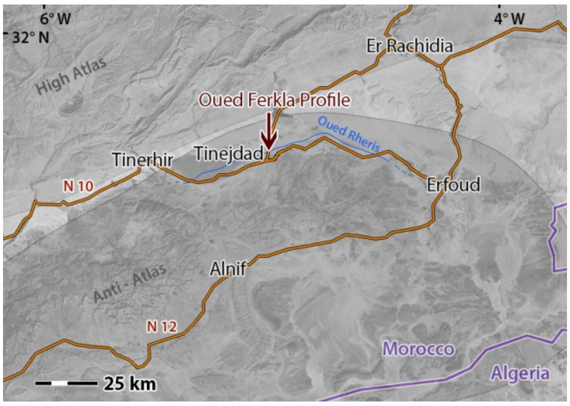

Fig. 1. Geographic position of the Oued Ferkla section

The Lower Emsian to Middle Givetian succession is dominated by grey, thin to thick-bedded (5-40 cm) and partly nodular limestones with deeper neritic to pelagic macrofauna. Intercalated are poorly fossiliferous, beige to grey, sometimes greenish, silty shales/marls. The strata dip to the N-NW with a tilt of 35-45°. They are sometimes intersected by small normal faults (Fig. 3), with displacements at the dm to m scale.

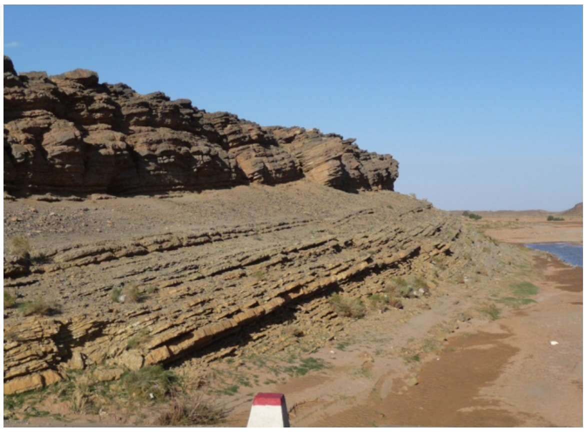

Fig. 2. Northern bank of the Oued Ferkla with natural exposure of Unit II at the base (Daleje Shale Equivalent, partly covered by sand), the nodular Unit III in the middle (*Anarcestes* Limestone), and the Chotec Event level as an incision at the base of the cliff-forming, Eifelian Unit III.

Fig. 3. Undulating normal fault within the Eifelian on the eastern side of the road cut.

2. DETAILED SECTION DESCRIPTION

The entire section is shown in Fig. 10. Bed packages were combined to units/subunits with boundaries at facies changes or events.

Unit I (Beds 1a-1b): Two beds (28 cm, Fig. 4) of medium-gray, slightly laminated to bioturbated mudstone almost without fauna. There are a few brachiopod remains and ostracods (Fig. 11a), no conodonts. A top Lower Emsian age is inferred from its position below the local Daleje Shale Equivalents.

Fig. 4. The initial Daleje Event as a sharp contact between Bed 1 and the overlying greenish, silty shales with some limestone concretions.

Unit IIa (Bed 2): 315 cm greenish shales with some limestone nodules but without macrofauna. The age of the overlying limestones shows that this interval correlates with the (lower part) of the Daleje Shale Equivalent (BECKER & ABOUSSALAM 2011) of the Tafilalt and with the lower Er Remlia Formation of the Maider (HOLLARD 1974). The transgressive Daleje Event, which has been proposed to mark the base of the upper Emsian, is probably represented by the sharp change from limestone to shale at the base (Fig. 4).

Unit IIb (Beds 3-14): 281 cm grey-beige shales with 10-30 cm large limestone nodules/concretions, which exhibit some orthocones (Bed 9) and rugose corals in the uppermost beds (Fig. 10, Beds 10, 12). They alternate with 10-38 cm thick grey limestone beds with rare macrofauna (primarily crinoids, dacryoconarids, goniatites, and shell detritus) and some (altered) pyrite. Bed 11 and 13 show additionally corals and trilobites. Unit IIb correlates probably with the higher Er Remlia Formation of the northern Maider (e.g., BULTYNCK 1985).

Unit III (Beds 15-30): The interval starts with 45 cm of grey mudstone (Bed 15), almost without shell detritus and with just a few ostracods (Fig. 11b), dacryoconarids, and foraminifers (*Tolypammina*); there are no conodonts. it is overlain by 1048 cm of slightly lighter grey, nodular limestone with some pyrite nodules (characteristic in Bed 23), sometimes intercalated by thin (ca. 0.5 cm) layers of shale/marl (Fig. 10). The nodular limestones can be classified as bioturbated wacke- and packstones (Fig. 11c) with crinoids, dacryoconarids, shell detritus, and ostracods. Some beds yielded orthocones (Beds 21, 30), phacopid trilobites (Bed 17, 28), brachiopods (Bed 27), bivalves (*Panenka*; Beds 29, 30), solitary Rugosa (*Oligophyllum*), and agglutinating foraminifers. Loose *Sellanarcestes applanatus* and *Anarcestes* provide an upper Emsian age for the main part of the succession and evidence for hemipelagic deposition. This contrasts with the conodont sample from Bed 27, which only contains the shallow-water species *Belodella resima* in association with acrotetrid brachiopods, ostracods (including spinose forms of the “Thuringian Ecotype”), and foraminifers (*Psammosphaera, Webbinelloidea*). The Chotec Event is not distinctive but marked by a more argillaceous nodular interval (top of Bed 30), which forms a slight platform on the slope (Fig. 2).

Fig. 5. Western cliff of the Oued Ferkla road cut with exposure of the (Lower) Kačák Event (position of hammer = Bed 101), marked by an interval of alternating dacryoconarid marls and laminated dark limestones (first thick bed above hammer = Bed 104).

Unit IV (Beds 31-93): This sequence of 1139 cm nodular to solid limestone is more fossiliferous than below, with a higher diversity of biota. There are thin-bedded limestones at the base (Beds 31-34), followed by some massive beds, 18 to 88 cm thick (e.g., Beds 36, 37, 44, 65, 66), which form the main cliff (Fig. 2). The beds are slightly darker than below and detrital. The dominant microfacies are bioturbated wacke-/packstones with crinoids, orthocones, trilobites, brachiopod fragments, and ostracods. Bed 31 also shows rugose corals (Fig. 11d), dacryoconarids, agglutinating foraminifers (*Ammodiscus, Tolypammina, Hemisphaerammina*), acrotetrids. *Linguipolygnathus bultyncki, Ling. pinguis*, *Icriodus corniger corniger*, and *Bel. resima* give a lower Eifelian age (upper part of *partitus* Zone). This confirms the position of the Chotec Event at the cliff base. Goniatite cross-sections occur rarely in the lower part (*Cabrieroceras* in Bed 40) but are more frequent above (Beds 58, 68, 69, 71, 90, 96, *Subanarcestes* in Beds 92-93). Other macrofauna includes phacopids (Bed 47), orthocones (e.g., Beds 48, 53, 65, 76, 82, 84, 89, 96), ribbed brachiopods (Bed 58), *Panenka* sp. (Bed 69, ?Bed 98b), and gastropods (Beds 48, 84, 98b). Unit IV suggests a deep neritic setting with increasing pelagic influx (slight deepening trend) in some higher beds.

Unit V (Beds 94-99): 219 cm medium to dark grey, solid, micritic limestone with common goniatites (*Subanarcestes macrocephalus* and others), orthocones, shell detritus, and a few crinoid remains (Fig. 10). The top (Bed 99) is a light grey wackestone (Fig. 11e) with some ostracods, many dacryoconarids, crinoids, acrotetrids, coral fragments, and foraminifers (*Webbinelloidea, Hemisphaerammina, Thurammina*). The conodont fauna is composed of *Ling.*

linguiformis Morphotypes γ1 and γ2 (sensu WALLISER & BULTYNCK 2011), *Tortodus kockelianus, Po. eiflius, Po. angusticostatus, Po. parawebbi, Po. praetrigonicus, Po. robusticostatus,* and *I. struvei.* This is a typical association of the upper Eifelian *eiflius* Zone (or upper subzone of the *kockelianus* Zone; see BELKA et al. 1997 and WALLISER & BULTYNCK 2011).

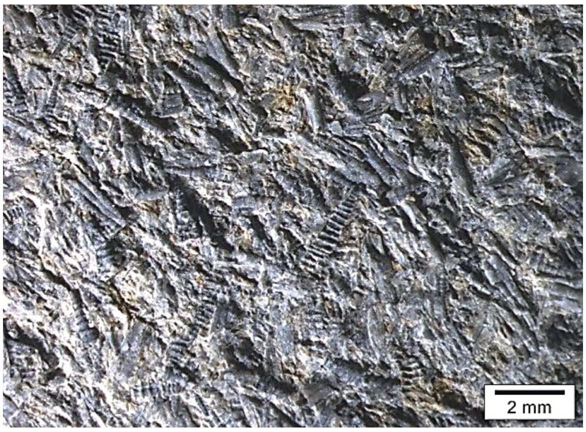

Fig. 6. Mass occurrence of nowakiids and styliolinids on a bedding plane within the basal Kačák Event Interval (marl layer within the middle part of Bed 101).

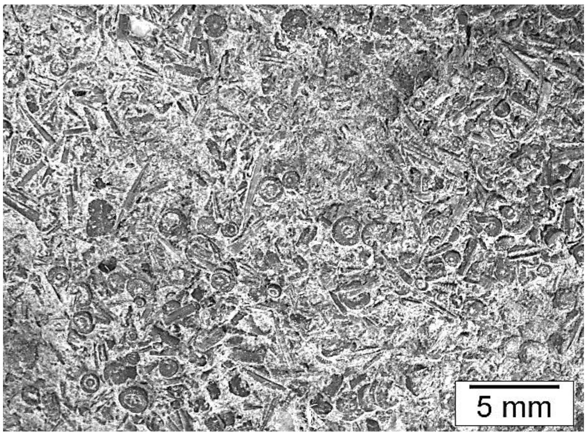

Fig. 7. Coquina of dacryoconarids and minute crinoid ossicles in the lower Kačák Event Interval (Bed 106).

Unit VI (Beds 100-125): Alternation of strongly weathered, dark shales (10-46 cm) with thin- to medium-bedded (5-16 cm), dark-grey, laminated limestone (e.g., Fig. 8). Bed 100 at the base is a dark grey bioturbated dacryoconarid wacke-/packstone (Fig. 11f) with orthocones, trilobites, a few ostracods, coral fragments, foraminifers (*Tolypammina, Hemisphaerammina*), and shell debris, laterally grading or changing abruptly into a strongly recrystallized dacryoconarid grainstone. *Ling. linguiformis, Po. eifelius, Po. robusticostatus, Po. parawebbi,* and *Po. trigonicus* suggest the basal *ensensis* Zone, locally defined by the extinction of *T. kockelianus*. It signals the base of the main or Lower Kačák Event (see WALLISER & BULTYNCK 2011). This is locally characterized by marly dacryoconarid coquinas of Bed 101 (Fig. 6). Higher parts of the transgressive event interval include mass occurrences of dacryoconarids in association with small crinoid ossicles (e.g., Bed 106, Fig. 7). Such thin levels reflect short episodes of maximum eutrophication and plankton blooms during a prolonged phase of maximum flooding. The dark-grey Bed 125 at the top carries a few centimeter large nodules of dark limestone and reflects a final transgressive peak. The slightly regressive Upper Kačák Event (see WALLISER & BULTYNCK 2011) is not distinctive within the interval of laminated limestones below (Figs. 5, 10).

Fig. 8. Amalgamation of distal debris flows causing the lamination of Bed 108 (12 cm thick) within the Kačák Event Interval.

Unit VII (Beds 126-144): 448 cm medium grey, sometimes laminated limestone, occasionally with undulating bedding surfaces and nodules. Bed 126 at the base indicates a gradual transition from the Kačák Event Interval. It is a recrystallized, microsparitic, laminated alternation of dacryoconarid grainstones and wackestones with minor ostracods, proetid debris, foraminifers (*Hyperammina*), and crinoids. One layer (Fig. 11g) shows convolute bedding, which suggests deposition by a distal debris flow. A maenioceratid cross-section on the outcrop gives a topmost Eifelian to basal Givetian age (WALLISER et al. 1995). The conodont fauna includes *Bel. resima*, *Ling. linguiformis* Morphotypes γ1-3, *Po. pseudofoliatus,* and *Po. ensensis.* Either it falls in the top *ensensis* or basal *hemiansatus* Zone; but it lacks the Givetian index species. Macrofauna from the main part mostly consists of orthocones (Beds 129-134; Beds 141-142). Bed 143 near the top is a light-grey, bioturbated wackestone (Fig. 11h) featuring ostracods, dacryoconarids, crinoid debris, foraminifers (*Psammosphaera, Webbinelloidea*), and some brachiopods. *Neopanderodus perlineatus, Ling. linguiformis, Ling. mucronatus, Po. timorensis, Po. pseudofoliatus, Po. varcus, I. arkonensis walliserianus, I. regularicrescens,* and *I. difficilis* give a basal Middle Givetian age, (*rhenanus-varcus* or *difficilis* Zones, BULTYNCK 1987 and BELKA et al. 1997; compare composite ranges in GOUWY & BULTYNCK 2002).

Unit VIII (Beds 145-156): 427 cm alternation of thin-bedded, light grey nodular limestones and medium-bedded, somewhat darker, more massive limestones. The sequence is interrupted by a 20 cm thick, silty, strongly weathered, grey-brown shale (Bed 151). Compared to Unit VII, the succession displays fewer fossils; common fauna is restricted to crinoids, dacryoconarids, and shell detritus (Fig. 10). The 18 cm thick Bed 145 at the base stands out as a dark grey limestone with a mass occurrence of micromorphic brachiopods (*Ense pumilio*). This eutrophic Lower *Pumilio* Event Bed (LOTTMANN 1990) yielded *Ling. linguiformis, Ling. mucronatus, Po. timorensis,* and *I. difficilis*, indicative of higher parts of the *varcus-rhenanus* Zone. Bed 147 (Fig. 11i) is a slightly bioturbated, dark grey wackestone with brachiopods, crinoids, dacryoconarids, a few gastropods, and foraminifers (*Hemispaerammina*, rare *Hyperammina*). It contains *Ling. linguiformis, Ling. weddigei, Po. timorensis, Po. varcus, Bel. resima, T. bultyncki,* and *Tortodus* sp., still of the same zone as above.

Fig. 9. The dark-grey Lower *Pumilio* Event Bed (Bed 145) at the base of Unit VIII.

The Upper *Pumilio* Event at the top of Bed 152 (Fig. 10) is less distinctive than the lower level. It is a medium-grey brachiopod packstone (Fig. 11j) with abundant, micromorphic *Ense pumilio*, shell detritus, some dacryoconarids, and crinoid debris. Foraminifers are absent. The conodont fauna includes *Ling. linguiformis, Ling. mucronatus*, *Po. ansatus, Po. varcus, Po. timorensis, I. brevis eslaensis, Neop. perlineatus,* and *Bel. resima.* This confirms the stratigraphic position of the Upper *Pumilio* Event at the base of the *ansatus* Zone, as in the Tafilalt (BULTYNCK 1987). There is also the characteristic, event-controlled, sudden incursion of *Latericriodus latericrescens latericrescens* (e.g., LOTTMANN 1990, ABOUSSALAM 2003). The sequence ends with Bed 156, a grey, 25 cm thick, massive, slightly bioturbated wackestone with crinoid debris, dacryoconarids, shell fragments, and goniatites (Fig. 11k). The conodont assemblage consists of *Ling. linguiformis, Ling. mucronatus,* and *Ling. weddigei* (still *ansatus* Zone).

3. PALAEOGEOGRAPHIC RELATIONSHIPS

Due to the common carbonates, the basal upper Emsian Daleje Shale Equivalents (Unit II) at Oued Ferkla have more similarities with the northern Maider than with the Tafilalt. The facies of the higher upper Emsian Unit III is somewhat closer to the nodular *Anarcestes* Limestone of the pelagic Tafilalt Platform. The contemporaneous Tazoulait Formation of the northern Maider consists partly of more solid limestone (HOLLARD 1974, BULTYNCK 1985, *Sellanarcestes* Limestone or Depophase 1b of DÖRING 2002). But the deepening trend of the overlying lower El Otfal Formation can be recognized at Oued Ferkla by the marly Bed 29.

The Eifelian of Oued Ferkla contains fewer goniatites than most Tafilalt sections but there are more differences to the northern Maider. That region shows very solid goniatite limestones and crinoidal limestones, often alternating with marls, distinctive marls with large, black limestone concretions (Tarherat, HOLLARD 1974), unfossiliferous red marls (Tarherat), neritic to biostromal limestones, and coarse debris flows (Ouihlane, SCHRÖDER & KAZMIERCZAK 1999). The succession at the Jebel Rheris is partly incomplete (FRÖHLICH 2004). This suggests a lower to middle ramp setting, in contrast with the hemipelagic platform facies of the Tafilalt and Oued Ferkla. The northern Maider has also no similar record of the Kacak Event Interval and its specific faunal blooms. Its Givetian is even more different, with the wide spread of partly thick coral-stromatoporoid biostromes (e.g., LE MAITRE 1947, SCHRÖDER & KAZMIERCZAK, 1999, FRÖHLICH 2004).

In summary, the Oued Ferkla Devonian can be interpreted as a distant NW-extension of the Tafilalt Platform. The Middle Devonian rise of the Ougnate High gradually diminished similarities with the northern Maider. The palaeogeographic situation, however, changed again in the Upper Devonian. The still poorly investigated Frasnian N of Oued Ferkla consists of ca. 130 m, strongly deformed, poorly fossiliferous, recrystallized, microsparitic and dolomitized mudstones and marls. A conodont sample from the top yielded *Ancyrodella curvata* Late Form, *Ancyrognathus tsiensi, Palmatolepis hassi, I. alternatus alternatus,* common *Avignathus decorosus* (including its diagnostic Pb element), and others. This suggests the lower part of the Upper Frasnian (MN 11 Zone, see KLAPPER 1997). So far, nothing is known about possible Famennian sediments of the Tinejdad region.

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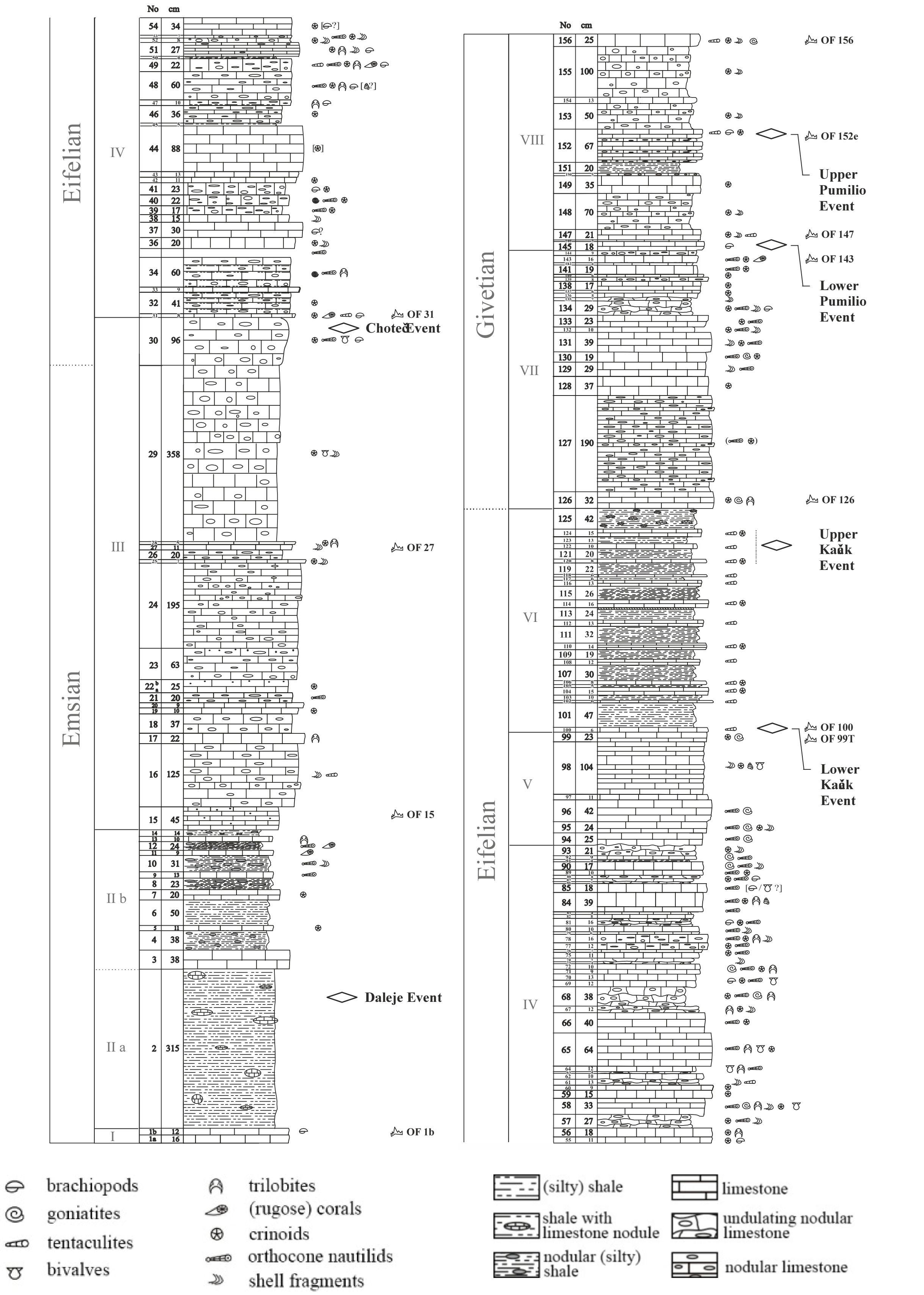

Fig. 10 (next page)**.** Lithological log, faunal content, chronostratigraphy, and the position of global events at Oued Ferkla (for biostratigraphical dating see text).

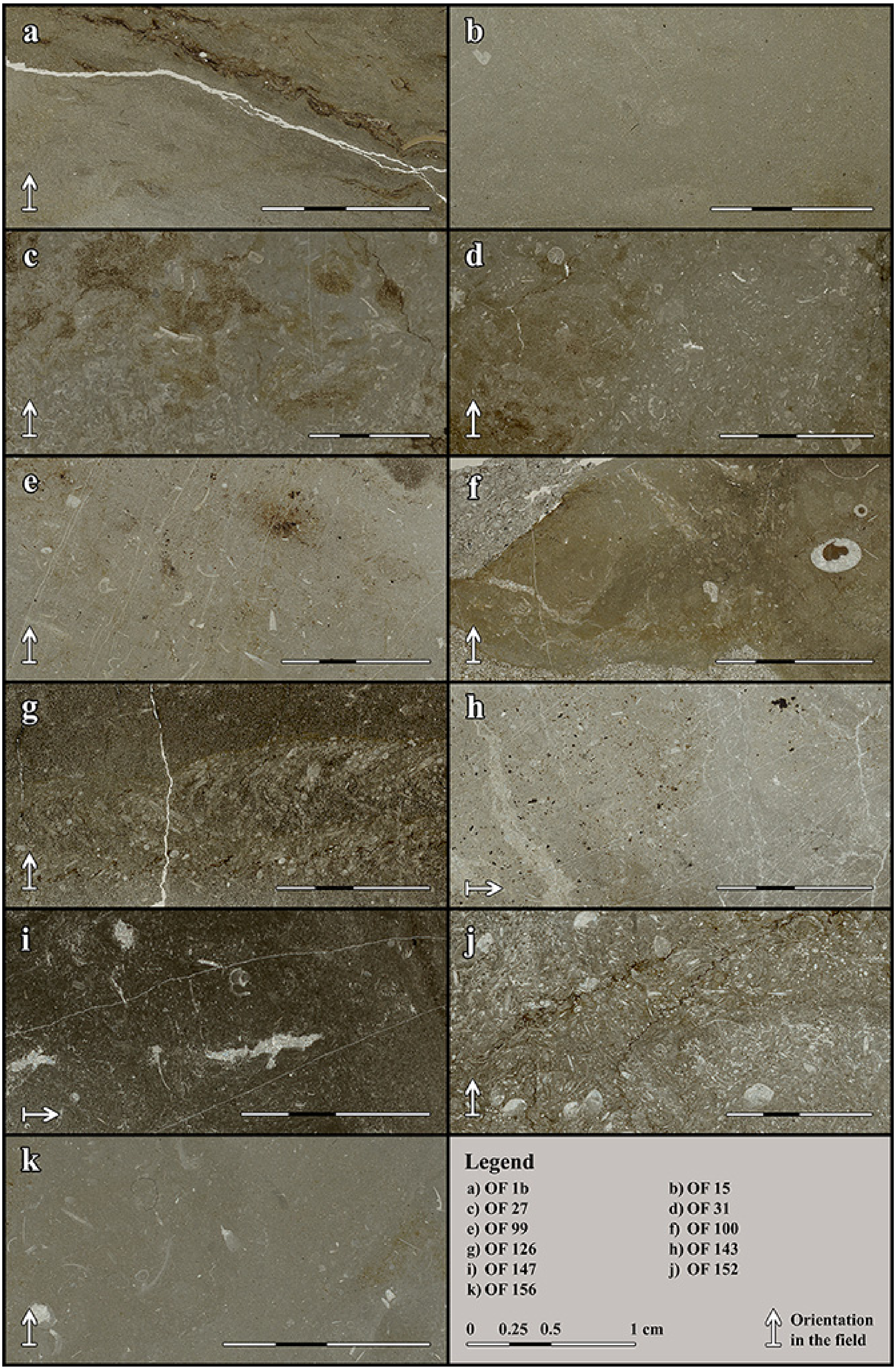

Fig. 11. Microfacies of representative Emsian to Givetian beds at Oued Ferkla.

70THE LOWER AND MIDDLE DEVONIAN AT EL KHRAOUIA (SOUTHERN TAFILALT)

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1. INTRODUCTION

El Khraouia lies in the southern Tafilalt on the topographic sheet Taouz-Ouest (NH-30-XIV-4), just north of the sharp bend at the eastern end of the northern limb of the Amessoui Syncline, ca. 11 km NNW of Taouz and 4 km E of the abandoned El Atrous settlement (Fig. 1). It has been included in a cross-section by WENDT et al. (1984: fig. 8), who noted its palaeogeographical position at the transition from the southern Tafilalt Platform to the eastern Tafilalt Basin. A succession ca. 2-3 km to the N has been named in WENDT & BELKA (1991) and KLUG (2002) as Rich Tamirant. KRÖGER (2008) described the Lower Devonian stratigraphy and nautiloid faunas at Filoun Douze at the southern limb of the Amessoui Syncline, 4 km to the south.

The Devonian sequence dips with ca. 40-50° to the NE. The GPS position for Bed 9b is W 4° 4´ 20´´, N 31° 0´ 24´´. The Lower Devonian is dominated by thick, poorly fossiliferous shale/marl, interrupted by fossiliferous marker limestones (Pl. 1, Fig. 1). The Middle Devonian (Pl. 1, Fig. 8) consists of limestone with pelagic to neritic fauna (up to middle ramp facies). The Eifelian forms a high ridge. From the Taghanic Event level on, basinal shales return. The visited succession has several advantages: 1. Its large outcrop size, 2. Due to its relative remoteness it has been less collected than easy accessible sections, 3. Due to its basinal setting it is complete and most units are relatively thick, 4. It has recently been logged during the revised geological mapping of the Taouz region. However, research is still at an initial stage.

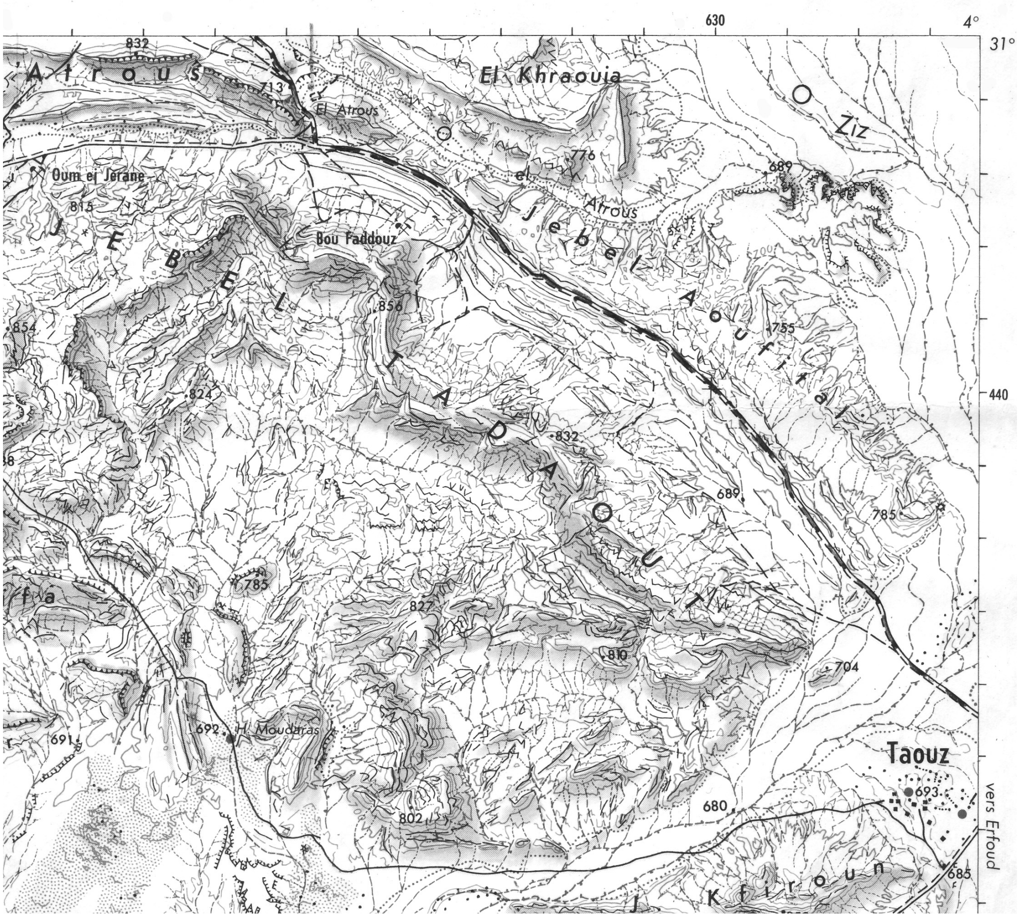

Fig. 1. Geographic position of the El Khraouia ridge E of El Atrous, N of the Jebel Ouaoufilal (= Aoufital), W of the Oued Ziiz, and NNW of Taouz, topographic sheet Taouz-Ouest (the main piste Rissani-Taouz is marked).

2. LOWER/MIDDLE DEVONIAN SUCCESSION

A. *Scyphocrinites* Limestone

Two 12 and 20 cm thick beds of solid, bluish, coarse crinoidal limestone (Beds 1 and 3a) are separated by ca. 4-5 m deeply weathered marl (Bed 2). Bed 1 yields abundant scyphocrinitids, including crowns and plate lobolites (Pl. 1, Fig. 2). The latter are known to be associated with the genera *Camarocrinus* and *Marhoumacrinus* (HAUDE & WALLISER 1999). A third crinoidal limestone (Bed 3c) follows ca. 1.7 m above Bed 3a, a fourth (Bed 4b) yet 1.2 m higher up. Bed 4b (Pl. 2, Fig. 1) is a representative crinoid rudstone with a recrystallized lobolite of 33 mm diameter, and a dark, organic-rich, slightly peloidal micrite matrix. The widespread disarticulation of the crinoids in Unit A suggests that they were hit by occasional storms.

Age: So far there is a very poor local conodont record; a sample from Bed 1 was barren. According to HOLLARD (1977) and HAUDE & WALLISER (1998), Unit A spans the Silurian-Devonian boundary. So far, we have only obtained basal Lochkovian conodonts, including *Caudicriodus woschmidti* (see its revised range in CARLS et al. 2008), from the lower part of the unit at other sections. This agrees with data in BELKA et al. (1999). Bed 4 yielded a juvenile *Caud. alcolae* (Pl. 3, Fig. 1) and acodiniform elements (*Acodina, Distacodina*). Based on data in GARCÍA-LOPEZ et al. (2002) and CORRADINI & CORRIGA (2012), *Caud. alcolae* suggest a middle Lochkovian age. However, relatives were already observed high in the lower Lochkovian (higher *transiens* Zone) of Cantabria. The top part of Unit A ranges in any case at least into the upper part of the lower Lochkovian, based on *Caud. transiens* from Jebel el Mrier, 13 km to the SW.

B. Lochkovian Shale

A subsequent ca. 200 m wide plain covers poorly exposed shales, which may be 70-100 m thick (Bed 5). They indicate a major late lower/middle Lochkovian deepening episode. Near the top (Bed 6a), there is a laminated, silty mudstone with foraminifers and rare dacryoconarids in a dark, pyrite- and organic-rich micrite matrix (Pl. 2, Fig. 2). It grades upwards into a layer of silty ostracod-crinoid wacke-packstone with some orthocones, probably a rare distal storm layer. Even 1.5 to 2 m higher there is a level of siderite platelets (Bed 6c), which indicate a sedimentary break during hypoxic conditions.

Age: Bed 6a yielded no conodonts, only rare ostracods and foraminifers (*Thurammina, Psammosphaera, Tolypammina*). According to HOLLARD (1977) the middle to upper parts of Unit B contain *Monograptus praehercynicus* and *Neomonograptus hercynicus* of the middle/early upper Lochkovian.

C. *Jovellania* Limestone (sensu KRÖGER 2008)

Ca. 80 cm thick grey, nodular limestone with abundant orthocones (Bed 7), lying in the plain.

Age: HOLLARD (1977) reported from this level of the Taouz area the early upper Lochkovian *Homocteno-wakia bohemia*. KRÖGER (2008) added *Paranowakia intermedia,* the index of the next higher nowakiid zone. However, a neighboring locality yielded the middle Lochkovian (CORRADINI & CORRIGA 2012) *Ancyrodelloides transitans*. This contradiction is deepened by an association of *Homoct. bohemia* and *Ancyrodell*. cf. *transitans* in HOLLARD (1977).

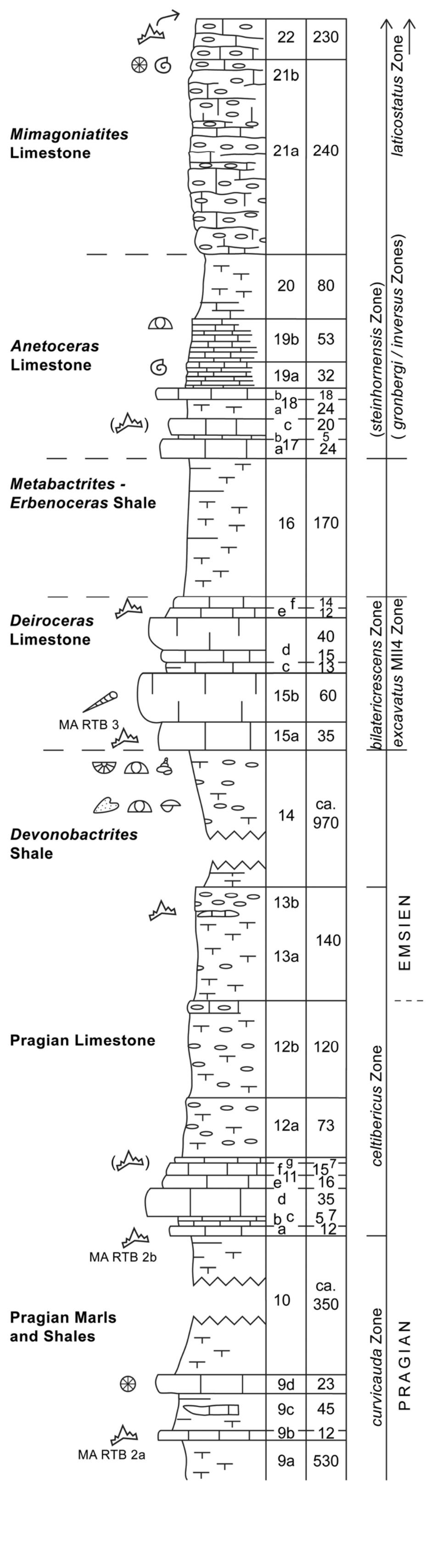

Fig. 2. Upper Pragian to lower Emsian lithology, and conodont biostratigraphy at El Khraouia.

D. Pragian Marl and Shale

Two intervals of deeply weathered marl/shale (Beds 8a and 9a, ca. 8.5-9 m thick), interrupted by a thin nodular limestone (Bed 8b), which is a strongly bioturbated dacryoconarid-crinoid packstone with many fragmented trilobites, mollusk debris, ostracods, some extraclasts, and Bryozoa (Pl. 2, Fig. 3). The micrite matrix is rich in very fine pyrite. The many neritic faunal elements indicate a regressive trend but the benthic environment remained dysoxic. The crinoidal Beds 9b and 9d in the upper part of Unit D support the regressive trend. Bed 9b is a bioturbated bioclastic wacke-packstone with crinoid, trilobite and mollusk debris, dacryoconarids and ostracods. Some sparite fenestrae may represent microbial mats.

Age: HOLLARD (1977), ALBERTI (1981), and KRÖGER (2008) place the base of the Pragian at a color change just above the *Jovellania* Limestone. ALBERTI (1998) reported in agreement with this view *Now. (Turkestanella) acuaria* cf. *prisca* from the basal Unit D of the central Tafilalt, followed higher by the typical subspecies. Sample MA RTB 2a from Bed 9b yielded a flood of *Belodella* and a few *Caud.* cf. *curvicauda* (Pl. 3, Fig. 2). This suggests an upper Pragian age (SLAVIK 2004; regional *curvicauda* Zone).

E. Pragian Limestone

Unit E begins with a low, ca. 1 m thick ridge of solid, thin- to medium-bedded, light grey limestone (Beds 11a-g). Bed 11d is a nodular, bioturbated bioclastic packstone with abundant crinoid, dacryoconarid, mollusk, and trilobite debris, and ostracods. It represents a storm-influenced, deeper neritic, lower carbonate ramp. Reddish, hematite-rich, diagenetically overprinted seams represent condensation intervals between depositional events. The higher part of the Pragian Limestone is more nodular and less condensed (Beds 12a-13b, ca. 4 m). The microfacies of Bed 13b (Pl. 2, Fig. 4), a crinoid-mollusk packstone, resembles Bed 11d. Gastropods, small brachiopods and nautiloids add to the deep neritic setting.

Age: The base and top of the unit (Beds 11a, 13b) are dominated by *Belodella* (Pl. 3, Fig. 3), which supports a neritic setting, but there are some associated *Caud. celtibericus* (Pl. 3, Fig. 4). This regional *celtibericus* Zone appears to correlate with the lower Emsian Conodont Step 17 of CARLS & VALENZUELA-RÍOS 2002, the level of the first *Eolinguipolygnathus excavatus* Morphotype 114, the proposed future basal Emsian index taxon. However, the direct association of *Eoctenopolygnathus pireneae* and *Caud. celtibericus* in the allochthonous Devonian of the Tinerhir area (RYTINA et al. this volume) proves an upper Pragian range of *Caud. celtibericus*, as suggested for the Bohemian type region (SLAVIK 2004). The Pragian Limestone probably falls in the lower Emsian of its current (Zinzilban) GSSP definition but the future, revised Emsian base will lie somewhere in its middle/upper part. Such an interpretation is in accord with nowakiid data from other Tafilalt sections (ALBERTI 1981, 1998: last acme of *Now. (Turkestanella) acuaria acuaria* and first peak of *Guerichina africana*).

F. *Devonobactrites* Shale

Ca. 9-10 m thick marl with many limestone nodules and a rich neritic fauna in the lower part, including trilobites (phacopids, scutelluids), small, smooth brachiopods, crinoids, and tabulate corals (*Thamnopora* and others). The upper half is less fossiliferous (some orthocones) and rich in weathered pyrite. This indicates a deepening upwards.

Age: In the adjacent Amessoui Syncline, for example at El Atrous North (= Takkat ou el Heyene) and Jebel Ouaoufilal (KLUG et al. 2008), there is a diverse fauna, including the oldest bactritids, which define the basal Emsian cephalopod zone LD III-A. ALBERTI (1998) recorded from the base of Unit F of the central Tafilalt the last *Now. (Turk.) anteacuaria* and *Guerichina*. Therefore, the base of the unit may correlate with the transgressive basal Emsian *atopus* Shale of Bohemia (SLAVIK 2004), slightly below the Lower Zlichov Event level sensu CHLUPÁČ & KUKAL (1986).

G. *Deiroceras* Limestone

Ca. 2 m massive, solid, grey limestones (Beds 15a-f), forming a small, prominent cliff. The base (Bed 15a) is a bioturbated bioclastic packstone with many styliolinids, crinoid ossicles, ostracods and mollusk debris. Bed 15f is very similar (Pl. 2, Fig. 5) but also includes gastropods. The microfacies is typical for a shallow hemipelagic carbonate platform. The base reflects a sharp regression and probably a sequence boundary, followed by a LST.

Age: The very rich conodont fauna from the base (Sample MA RTB 3) is, again, dominated by *Belodella*. But there are also frequent *Criteriognathus miae* (Pl. 3, Fig. 7) and icriodids, including *Caud. celtibericus, Caud. sigmoidalis* (Pl. 3, Fig. 5), and *Latericriodus bilatericrescens multicostatus* (Pl. 3, Fig. 6). The latter characterizes the basal Emsian *bilatericrescens* Zone. A single *Eol. excavatus* Morphotype 114 (Pl. 3, Figs. 8-9) confirms the basal Emsian age. Bed 15f is dominated by the three subspecies of *Lat. bilatericrescens*, in association with *Crit. miae, Caud. sigmoidalis, Eol. excavatus* (s.str. and Morphotype 114), and rare *Eol.* n. sp. aff. *pannonicus* (sensu BECKER & ABOUSSALAM 2011, Pl. 3, Fig. 10). This assemblage, especially the last species, is regionally typical for the top part of the *excavatus* M114 Zone. ALBERTI (1981) found in the central Tafilalt *Now. (Now.) zlichovensis maghrebiana* and *Now. (Now.) praesulcata* in Unit G and just above.

H. *Metabactrites* Shale

1.7 m of poorly fossiliferous shale/marl (Bed 16), which represent a significant deepening episode

(TST/HST, Chebbi Event sensu BECKER & ABOUSSALAM 2011). Unit H is in the Amessoui Syncline generally much less fossiliferous than Unit F (KLUG et al. 2008 and own data).

Age: In the central and eastern Tafilalt the rich oldest ammonoids (*Metabactrites, Chebbites, Erbenoceras,* etc.) of Unit H define the lower Emsian zone LD III-B (see KLUG 2001 and KLUG et al. 2008).

I. *Anetoceras Limestone*

At the base there are ca. 50 cm solid limestone with poor macrofauna (Bed 17), followed by 24 cm marl (Bed 18a), another solid limestone with few goniatites (Bed 18b), and ca. 85 cm platy limestone (Bed 19) with some phacopids (Fig. 2). The base marks a sharp regression, turning into condensed LST (Bed 17) and subsequent TST deposits.

Age: *Erb. solitarium* (Pl. 1, Fig. 5) indicates the mid-lower Emsian *Anetoceras obliquecostatum* Zone (LD III-C). A conodont sample from Bed 17c was unexpectedly barren. At Jebel el Mrier to the south, and elsewhere in the Tafilalt, *Crit. steinhornensis* is typical for Unit I but polygnatids are always rare at this level. In the central Tafilalt, the base of Unit I has *Now. (Now.) praesulcata* and *Now. (Now.) tafilaltana*, followed by *Now. (Now.) praecursor* and, near the top, *Now. (Now.) barrandei* (ALBERTI 1981, 1998).

J. *Mimagoniatites* Limestone

Above ca. 80 cm deeply weathered marl (Bed 20, ?late TST), Unit J consist of ca. 4.7 m middle grey nodular limestone, which is bioclastic and somewhat darker than the *Anetoceras* Limestone in the lower part (Bed 21, with goniatites) but light-grey at the top (top Bed 22). The top of Unit J is a slightly bioturbated dacryoconarid packstone, with cone-in-cone stacking of nowakiids and styliolinids, abundant shell debris, some fragmentary crinoids, and an upwards decreasing pyrite content of the fine micrite matrix. This suggests an improved oxygenation and circulation upwards despite a slight deepening, as indicated from the influx of deeper-water conodonts.

Age: *Mimagoniatites cf. fecundus* from Bed 21 is the index of the top lower Emsian zone LD III-D. The conodont fauna from the top of Bed 22 is rich in *Belodella* and *Neopanderodus* but also includes *Linguipolygnathus laticostatus (*Pl. 3, Figs. 11-12), *Ling. vigierei, Ling. inversus (*Pl. 3, Fig. 13), and *Caud. ultimus.* This association is typical for the laticostatus Zone at the top of the lower Emsian. In the central Tafilalt, *Now. (Now.) elegans* enters low in Unit J and *Now. (Now.) cancellata* (unrevised) at its top (ALBERTI 1981).

K. Daleje Shale Equivalents

100–120 m silty, greenish-grey, poorly fossiliferous shales (“Bed” 23), which are only well exposed in small, steep gullies (Pl. 1, Figs. 1, 6). The (main) transgressive Daleje Event occurred at the base, above a minor discontinuity surface.

Age: In the central and eastern Tafilalt, Unit K carries rich goethitic (originally pyritic) ammonoid faunas of the early upper Emsian LD IV-A to IV-C (BECKER & HOUSE 1994, KLUG 2002, WEBSTER et al. 2005). There are no conodonts.

L. *Anarcestes* Limestone

Several meters of yellowish weathering, light-grey, marly nodular limestone (“Bed” 24), with abundant goniatites (*Sellanarcestes, Anarcestes, Achguigites*), and phacopid trilobites. Outcrops are heavily covered by debris from the Eifelian cliff above. Unit L represents a HST.

Age: The goniatites are typical for the higher upper Emsian *Anarcestes* Zone (LD IV-D), which correlates with the *serotinus* to *patulus* Zones (e.g., BELKA et al. 1999, KLUG 2002). Polygnathids, however, are rare in the *Anarcestes* Limestone. The Emsian/Eifelian boundary lies elsewhere in the Tafilalt (e.g., BECKER & ABOUSSALAM this vol.) near its top.

M. Lower/Middle Eifelian Limestone

At the base there is a ca. 1.4 m thick interval of dark-grey, solid limestone (Bed 25a) with common goniatites, including *Fidelites, Werneroceras*, and (loose) early *Subanarcestes* (Pl. 1, Fig. 7), as well as bivalves (*Pterochaenia*) that are typical for pelagic low-oxygen facies. The transgressive and eutrophic Chotec Event is expected in this interval but styliolinites have not yet been seen, perhaps due to the restricted outcrop (wide cover by debris from above). Above, there is a more massive, ca. 80 cm thick *Subanarcestes* Marker Limestone (Bed 25b), which occurs widely on the Tafilalt Platform (e.g., BULTYNCK 1985, BECKER & HOUSE 1994, KLUG 2002). It represents a short regressive interval (late HST) and yielded *Pinacites* sp., *Suban. sphaeroides*, and *Fidelites*. It is overlain by ca. 4.5 m nodular limestone with *Subanarcestes*, other goniatites, orthocones, brevicones, and *Panenka* sp. (Beds 25c-26). 1.3 m below the top thin and fine distal turbidites commence. The top of Bed 26 is a bioturbated bioclastic packstone with many styliolinids, crinoids, and mollusk debris, indicative of a shallow pelagic carbonate platform/ramp.

Age: Bed 25a falls in the basal Eifelian *Fidelites* or *Foordites* Zone (MD I-B), which correlates with the *partitus* Zone (BULTYNCK 1985, BECKER & ABOUSSALAM, this vol.). Bed 25b can be placed in the higher part of the *Pin. jugleri* Zone (MD I-C), which equals the basal part of the *costatus* Zone (BULTYNCK 1985, BECKER & HOUSE 1994, KLUG 2002*).* The minor subsequent deepening of the lower *costatus* Zone seems to have high correlation potential in North Africa. *Ling. linguiformis, Ling. pinguis, Icriodus regularicrescens* (Pl. 3, Fig. 15), and *Polygnathus angusticostatus* (Pl. 3, Fig. 14) from the last nodular level of Bed 26 fall in the upper part of the *costatus* Zone.

N. Upper Eifelian Turbidites

Ca. 3 m dark-grey, solid, laminated turbiditic limestones forming the cliff top, sometimes with convolute bedding. The top of Bed 27 is a representative laminated and recrystallized (microsparitic) limestone with grading from dacryoconarid wackestone with some trilobite and mollusk debris into silty mudstone (Pl. 2, Fig. 7). The identical turbidites of the adjacent eastern slope (Bed 28, ca. 3 m) show large *Zoophycos* but other macrofauna is rare. Downslope, vertically bedded, ca. 10 m alternating dark-grey marls and thin turbidites follow (see Pl. 1, Fig. 8), which end with two more solid turbidite beds (Beds 29d and 29d). The turbiditic interval reflects both a deepening and steepening of the slope, which suggests a regional tectonic trigger. However, the rough correlation of its base with the Bakoven Event sensu DE SANTIS & BRETT (2011) may be more than a coincidence.

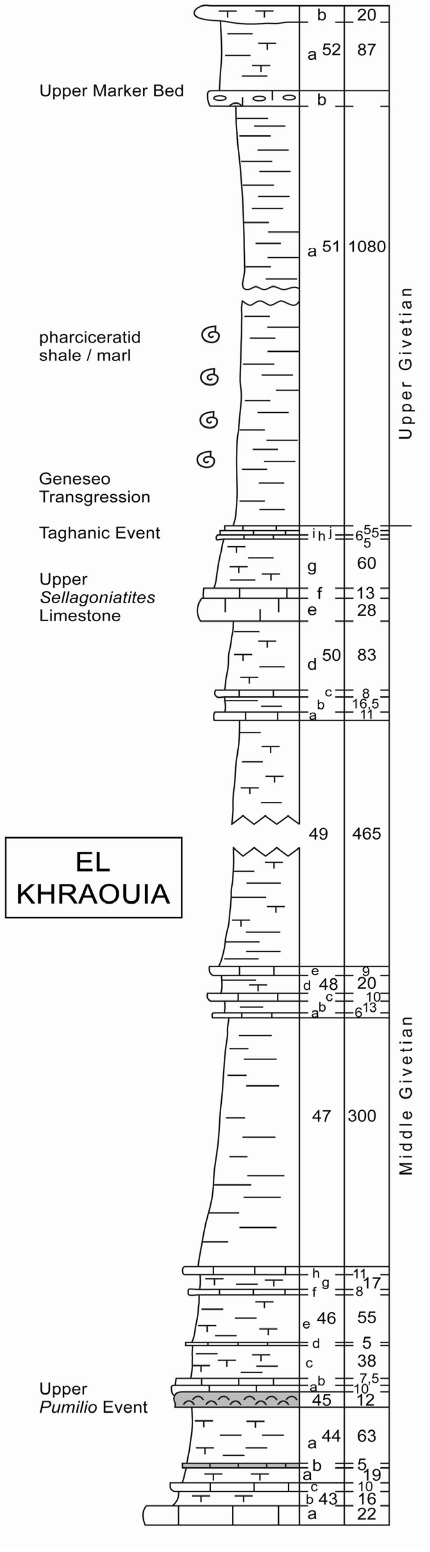

Fig. 3. Middle/Upper Givetian lithostratigraphy at El Khraouia across the Upper *Pumilio* and Taghanic Events (Middle and Upper Givetian).

Bed 30 (ca. 60 cm) is defined by a return to light-grey marls and nodular goniatite limestone with *Holzapfeloceras* and *Agoniatites*. The main Kačák Event Interval may be (partly) represented by an overlying marl unit (Bed 31, ca. 4 m). Its coincidence with the end of wide-spread turbidite shedding, which is also true for the NW (Ottara), E (Hassi Nebech) and SW (Jebel el Mrier), is remarkable.

Age: The top of Bed 27 produced a rich conodont fauna with *Tortodus kockelianus kockelianus* (Pl. 3, Fig. 16), *Po. angusticostatus, Po. robusticostatus, Po. pseudofoliatus, Po. angustipennatus, Ling. linguiformis* Morphotypes γ1-2, and *I. anterodepressus*. This is a typical assemblage of the (main) *kockelianus* Zone (compare BELKA et al. 1997). Bed 30b contains the index species of the latest Eifelian *Holz. circumflexiferum* Zone (KLUG 2002a, MD I-F2 of BECKER & HOUSE 1994), which characterizes the Kačák Event Interval.

O. Lower/Middle Givetian Limestone

At the base there is an alternation of thin-bedded grey limestone and marl with some phacopid remains (Bed 32). These are overlain by a ca. 1 m high cliff composed of more solid limestone beds (Bed 33). At the base is a fining upwards bioturbated bioclastic limestone with many dacryoconarids, some ostracods and mollusk debris. This microfacies is characteristic of a calm, hemipelagic carbonate ramp. On the main eastern slope there is a still poorly studied, more than 15 m thick alternation of thin- to thick-bedded solid limestone and deeply weathered marl with some rugose and tabulate corals. They indicate a shallower, neritic mid-ramp setting.

Bed 43 is a laminated, dark-grey limestone, which sandwiches a mass occurrence of minute brachiopods (Ense). The thin section of this Lower *Pumilio* Event bed (LOTTMANN 1990) shows a strongly recrystallized brachiopod-dacryoconarid-ostracod packstone, which is interpreted to be the result of a sudden eutrophication event. It is followed by a ca. 10 m thick, almost vertical succession of bioclastic limestones (up to 50 cm thick) and marls. Bed 45 is the Upper *Pumilio* Event level and consists of two brachiopod coquinas. The thin-shelled brachiopods are strongly recrystallized, imbricated and associated with some dacryoconarids (Pl. 2, Fig. 8). The marls just below (Bed 44) mark the base of the transgressive

Depophase If-UPum sensu BECKER & ABOUSSALAM (2011). Further downslope there are more, mostly thin-bedded bioclastic limestones alternating with marls. The 3 m thick Bed 47 and the 4.65 m thick Bed 9 (Fig. 3) are tentatively correlated with phases of the *Maenioceras* Marl (BECKER & ABOUSSALAM 2011, Depophase If-Win). A sequence of solid limestones forms Bed 51, which includes a 41 cm thick marker unit (Beds 51e-f) and a bundle of three thin limestones (Beds 51g-i) at the top.

Age: The base of Bed 33 yielded *Po. varcus* (Pl. 3, Fig. 17), *Ling. linguiformis, I. difficilis* (Pl. 3, Fig. 18), and others. Therefore, it falls already in the basal Middle Givetian. Consequently, there is a strong condensation of the Lower Givetian, in large contrast to the Middle Givetian to Frasnian. The Upper *Pumilio* Event layer yielded *Po. ansatus* (Pl. 3, Fig. 19), *Po. varcus, Po. xylus, I. brevis brevis*, and others, indicative of the basal *ansatus* Zone, as at Bou Tchrafine (BULTYNCK 1985). There is also the sudden influx of *Latericriodus* (Pl. 3, Fig. 20), as at Oued Ferkla (WARD et al this vol.), which suggests an immigration pulse from eastern North America. The apparent breakdown of a palaeobiogeographic barrier by the Upper *Pumilio* Event is currently not understood at all. Bed 50i yielded a diverse conodont association, including *Ling. weddigei, Ling. mucronatus, Po. ansatus, Po. varcus*), *Tortodus caelatus* (Pl. 3, Fig. 1), *T.* aff. *weddigei*, *I. brevis brevis,* and others. *T.* aff *weddigei* indicates the upper part of the *ansatus* Zone, probably within the Taghanic Event Interval. Therefore, the massive limestone within Bed 50 is correlated with the regionally widespread Upper *Sellagoniatites* Limestone (ABOUSSALAM 2003, ABOUSSALAM & BECKER 2011). The marly Bed 51f is thought to mark the base of Depophase IIa-Tagh but more detailed data are required.

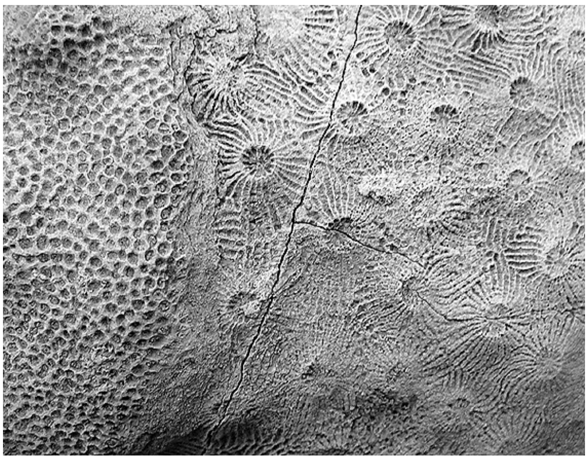

Fig. 4. *Phillipsastrea*, partly overgrown by an alveolitid, Bed. 51a, basal upper Givetian (ca. x 0.9).

P. Upper Givetian Marl and Limestone

The termination of the Middle Givetian neritic limestone succession is abrupt. The subsequent, ca. 11 m thick, deeply weathered marls contain a mixture of neritic (colonial *Phillipsastrea*, Fig. 4, tabulate corals, crinoid ossicles, and gastropods) and pyritic pelagic fauna (tornoceratids, *Pharciceras, Stenopharciceras,* and two species of *Pseudoprobeloceras,* Fig. 5). This sharp break reflects the significant basal Upper Givetian eustatic rise (Geneseo Transgression, base of Depophase IIa-Gen of ABOUSSALAM & BECKER 2011). Bed 51b is a 18 cm nodular limestone (bioturbated mudstone with rare styliolinids and ostracods) with some small brachiopods and increasing fine siliciclastic detritus at the top. It is separated by ca. 90 cm marl (Bed 52a) from a vertical, laterally variably thick prominent marl unit (Bed 52b).

Fig. 5. *Pseudoprobeloceras* cf. *praecox* (left, 17 mm diameter) and *Ps. pernai* (right, 21 mm dm) from the Upper Givetian marl (Bed 51a).

Age: The goniatites from Bed 51a correlate straight away with the famous *Pharciceras* Fauna of Hassi Nebech to the east (BENSAID 1974, BOCKWINKEL et al. 2013). Since this assemblage comes mostly from the middle part of the Upper Givetian (MD III-D), its lower part must be represented by the lower part of the thick marl. Locally, there is no evidence for the regional Lower Marker Bed (sensu BECKER & HOUSE 1994, 2000; see ABOUSSALAM 2003), which is normally a rather massive goniatite limestone. However, Bed 51b can be correlated with the regional Upper Marker Limestone (*dengleri dengleri* Subzone, = upper part of previous Upper *disparilis* Zone, ABOUSSALAM & BECKER 2007), based on the presence of *Po. dengleri dengleri* (Pl. 3, Fig. 22), *Po. ordinatus*, *Po. tafilensis* (Pl. 3, Fig. 23), *Po. paradecorosus*, and *Schmidtognathus peracutus*. The observed shallowing upwards is rather characteristic for the unit. The Givetian/Frasnian boundary lies close to Bed 52b or within the subsequent wide plain.

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Plate 1


Fig. 1


Fig. 2


Fig. 3


Fig. 4


Fig. 5


Fig. 6


Fig. 7


Fig. 8

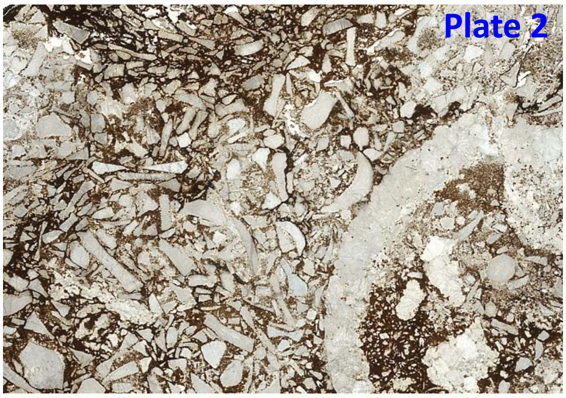


Fig. 1. Bed 4b, top of *Scyphocrinites* Limestone.

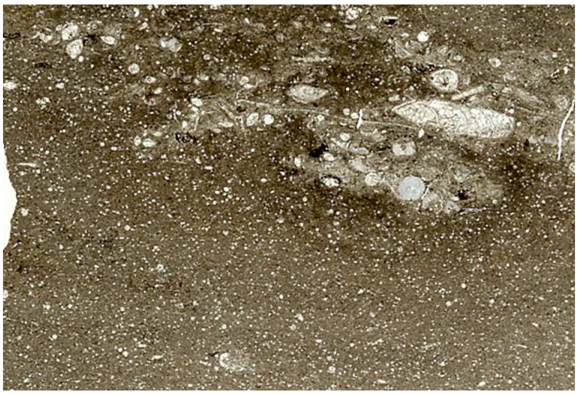


Fig. 2. Bed 6a, within upper Lochkovian Shale.

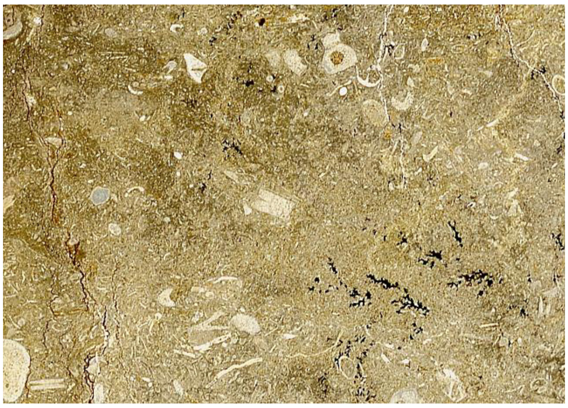


Fig. 3. Bed 8b, within Pragian Marl/Shale .

Fig. 4. Bed 13b, top of Pragian Limestone.

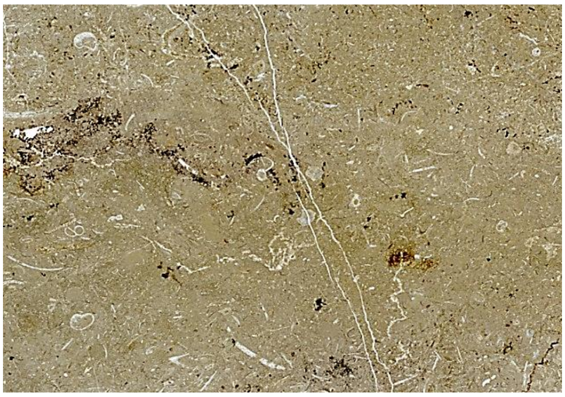


Fig. 5. Bed 15f, top of *Deiroceras* Limestone.

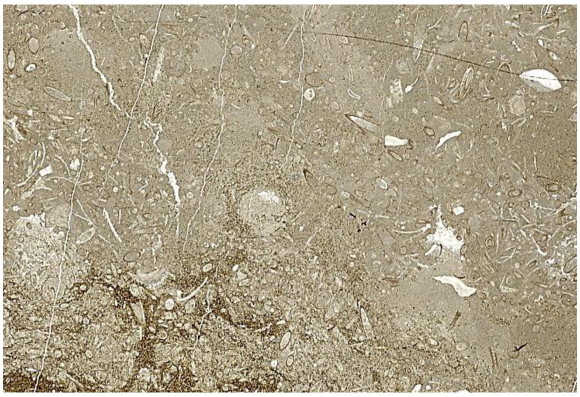


Fig. 6. top Bed 22, top of *Mimagoniatites* Limestone.

Fig. 7. Top of Bed 27, upper Eifelian turbidite.

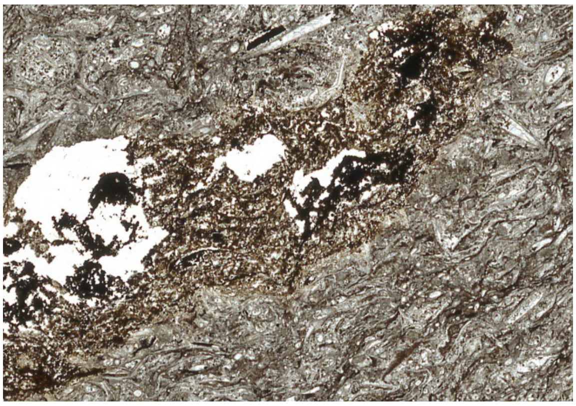
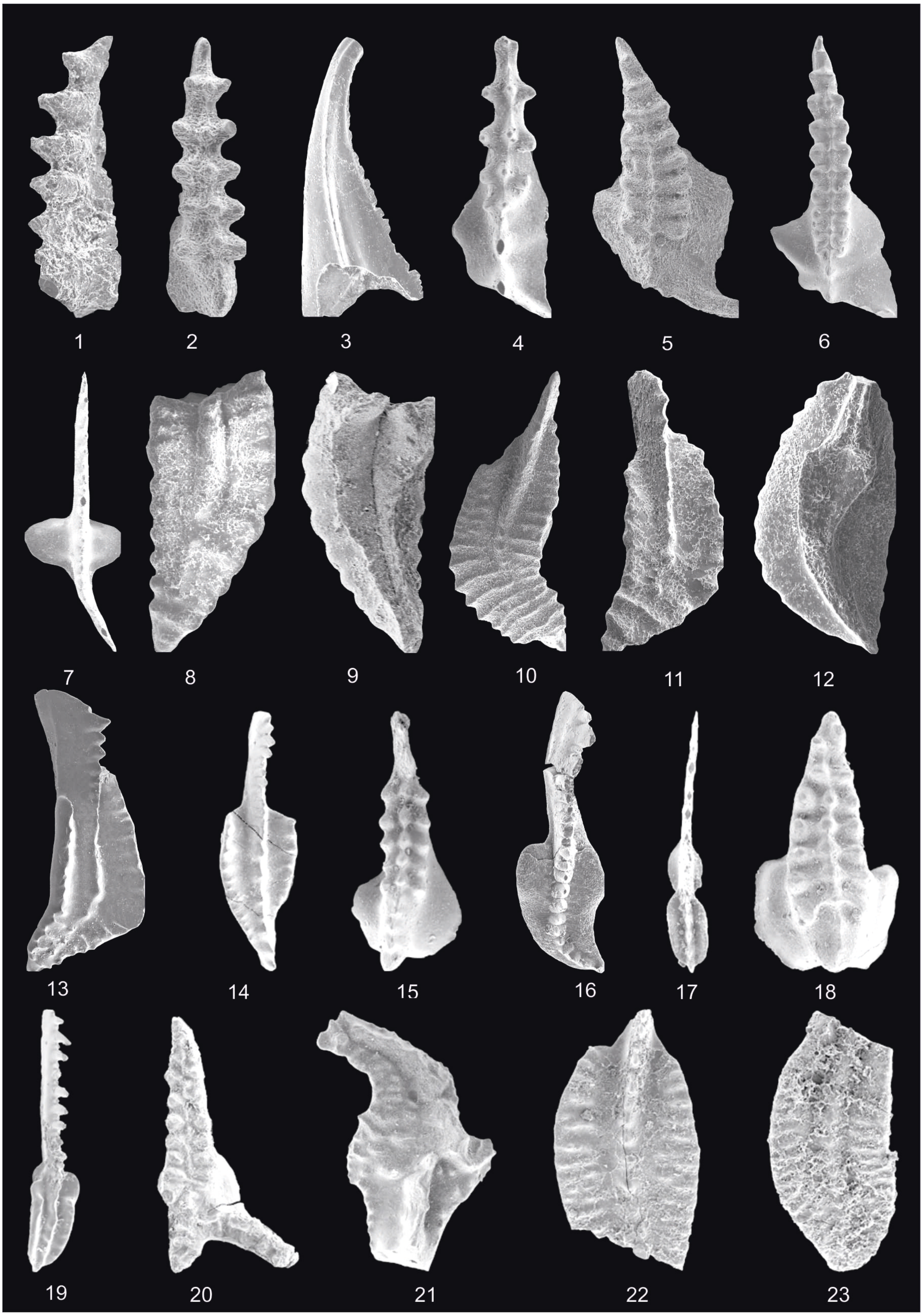


Fig. 8. Upper *pumilio* Bed (Bed 45).



Pl. 3. Lower/Middle Devonian conodonts from El Khraouia. **1.** *Caud. alkolae* juv., oblique view of poorly preserved specimen, Bed 4, **2.** *Caud.* cf. *curvicauda*, Bed 9b, **3.** *Belo. triangularis*, Bed 9b, **4.** *Caud. celtibericus*, Bed 11a, **5.** *Caud. sigmoidalis*, Bed. 15a, **6.** *Lat. bilatericrescens multicostatus*, Bed 15a, **7.** *Crit. miae*, Bed 15a, **8-9.** *Eol. excavatus* Morphotype 114, Bed 15a, **10.** *Eol*. n. sp. aff. *pannonicus*, Bed 15f, **11-12.** *Ling. laticostatus*, Bed 22top, **13.** *Ling. inversus*, Bed 22top, **14.** *Po. angusticostatus*, top Bed 26, **15.** *I. regularicrescens*, Bed 26, **16**. *T. kockelianus kockelianus*, top Bed 27, **17.** *Po. varcus*, Bed 33, **18.** *I. difficilis*, Bed 33, **19** *Po. ansatus*, Bed 45, **20.** *Lat. latericrescens latericrescens*, Bed 45, **21.** *T. caelatus,* Bed 50i, **22.** *Po. dengleri dengleri*, Bed 52, **23.** *Po. tafilensis*, Bed 52.

THE UPPER DEVONIAN AT EL KHRAOUIA (SOUTHERN TAFILALT)

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1. INTRODUCTION

The Upper Devonian of El Khraouia has not been studied by previous authors. WENDT et al. (1984) noted its peculiar palaeogeographic position at the western margin of the Tafilalt Basin, with very thick Upper Frasnian beds and increasing condensation towards the middle Famennian, which suggests a sea-level controlled mid-Famennian progradation of the hemipelagic Tafilalt Platform. For the geographic position see the Lower/Middle Devonian section description in this volume. The GPS position for the basal Famennian (Sample MA RTB 12) in the middle of the wide plain is W 4° 3´ 58´´, N 31° 0´ 35´´. After an initial survey in spring 2011, the top lower to basal upper Famennian was logged and sampled in more detail in March 2012. The sedimentary and faunal succession can be summarized as follows:

2. FRASNIAN

2.1. Lower Frasnian

Upper Givetian marls with pharciceratids and the regional Upper Marker Bed (sensu BECKER & HOUSE 1994b, Upper *Pharciceras* Horizon in EBERT 1993) form the western margin of a wide plain. A good part of the subsequent ca. 65 m thick, deeply weathered, unfossiliferous shale/marl sequence ("Bed 53") must represent the Lower Frasnian since the first limestone above falls already in the middle part of the Middle Frasnian. Unlike as in other basinal sections (e.g., Hassi Nebech, BOCKWINKEL et al. 2013), the Frasnes Event Interval is not easily marked by styliolinites.

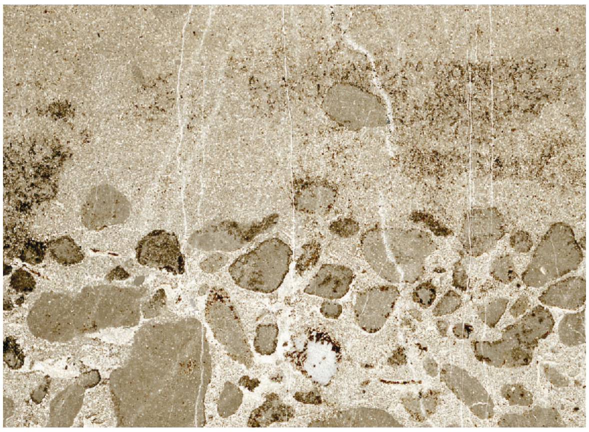


Fig. 1. Reworking level with partly Fe-coated extraclasts at the base of Bed 54.

2.2. Middle Frasnian

Bed 54 is a peculiar 26 cm thick, middle grey marker limestone (slightly bioturbate, peloidal mudstone with rare trilobite remains), which conglomeratic base gives evidence for a reworking event and unconformity. Many darker-grey mud-wackestone pebbles were rounded and iron-coated before re-sedimentation (Fig. 1). They testify a mid-Frasnian regressive episode, with significant bottom turbulence reaching the deeper basinal facies, with estimated 150-200 m water depth. The poor sorting and grading speaks against the transport by a single tempestite or turbidite.

Following a 50 cm marl (Bed 55a), there is a single light-grey nodular limestone (Bed 55b, bioturbated, strongly dolomitized mudstone with rare styliolinids, Fig. 2). All subsequent thin limestones are darker brownish-grey, laminated and represent distal calcareous turbidites. The irregularly bedded Bed 56d, a goethite/hematite-rich crinoid wacke-packstone with some styliolinids, is a thick (up to 42 cm) example. Most other limestones are only 5-8 cm thick. The interval from Bed 56a to Bed 58 is ca. 35 m thick.

Fig. 2. Strongly bioturbated nodular limestone (mudstone) of Bed 55b.

Age: It is likely that parts of the early Middle Frasnian have locally been reworked. Bed 54 yielded *Icriodus* cf. *vitabilis* (Pl. 1, Fig. 3), *Ancyrodella curvata* Early Morphotype (Pl. 1, Fig. 2), *Ancyrognathus coeni*, (Pl. 1, Fig. 1), *Polygnathus paradecorosus*, and *Po.* aff. *paradecorosus* (Pl. 1, Fig. 5, transitional to *Avignathus decorosus*). This assemblage without *Palmatolepis* is typical for the regional *coeni* Zone and correlates with the Middle Frasnian MN Zone 8. Bed 55b belongs to ca. the same interval, based on *Ad. curvata* Early

Morphotype in association with abundant *I. symmetricus, Po. paradecorosus*, and *Polygnathus* div. sp. But a single juvenile *Pa. plana* signals the regional *plana* Zone (or MN Zone 10) higher in the Middle Frasnian. Bed 56 produced concordantly very abundant *Ad. curvata* Early Morphotype in association with *Ag. coeni, Pa. plana* (Pl. 1, Fig. 7), and *Pa. hassi* (Pl. 1, Fig. 6).

2.2. Upper Frasnian

Bed 58t is slightly darker than most of the older turbidites and, in accord with its conodonts, is taken as the base of the Upper Frasnian. It is followed by ca. 40 m deeply weathered shale/marl, interrupted by four thin, distal turbidites. Bed 59d (Fig. 3) is a representative, laminated, pyrite-rich calcisiltstone with shell filaments and abundant homoctenids in some layers. Bed 60b (Fig. 4), by contrast, hardly has any fauna. It consists of a pyrite-rich and laminated alternation of silty, somewhat peloidal mudstone, which contains rare shell filaments and foraminifers, with thinner, unfossiliferous calcisiltites.

Fig. 3. Distal turbidite (calcisiltite) of Bed 59d.

Age: Bed 58t yielded marker conodonts for the basal part of the Upper Frasnian (MN Zone 11), *I. alternatus alternatus* (Pl. 1, Fig. 8), and *Ag. triangularis* (Pl. 1, Fig. 9), in association with *Ad. curvata* Late Morphotype and *Po. evidens* (Pl. 1, Fig. 10). It correlates with the oldest Kellwasser facies of the Tafilalt Platform, which is only locally preserved beneath the true Lower-Upper Kellwasser interval (e.g., WENDT & BELKA 1991) and always separated by an unconformity (BECKER & HOUSE 2000, BELKA et al. 2002). Bed 59d falls in the top part of the Upper Frasnian (MN Zone 13), based on the association of fragmentary *Pa. bogartensis, Pa.* cf. *hassi, Pa.* cf. *winchelli, I. alternatus alternatus*, and *Ad. curvata.* The thick, rather uniform and poorly fossiliferous sequence prevents the recognition of equivalents of the Lower and Upper Kellwasser levels. But it is intriguing that the terminal Frasnian (conodont) extinction event occurred within a rather uniform, low-oxygen facies, without obvious break. The conodont- and dacryoconarid/homoctenid-free Bed 60b seems to represent the "Dead Zone" of the pelagic extinction peak. The local lack of sea-level, impact and oxygenation signatures leave temperature, salinity stratification, sea water chemistry, and/or trophic changes as potential extinction agents.

Fig. 4. Bed 60b, the Frasnian-Famennian boundary "Dead Zone", very poorly fossiliferous, laminated mudstone.

3. FAMENNIAN

3.1. Lower Famennian

The base of the Famennian is marked by the sudden, opportunistic proliferation of *Phoenixites frechi*, the only unaffected goniatite survivor of the Upper Kellwasser Extinction (BECKER 1993). It enters in marly concretions (Bed 60f), mostly recrystallized, microsparitic, pyrite-rich mudstones. A few additional specimens were observed just above in a thin, brownish weathering, calcareous siltstone (Bed 60h). Subsequently, the wide plain consists of deeply weathered shale/marl, intercalated by 15 levels of up to 28 cm thick marly concretions, calcisiltites or calcareous siltstones without any fauna (Beds 61a to 63a). The marls of Bed 63a (or Beds Fa 3a-5a, Fig. 5) yield loose, minute hematitic goniatites, dominated by tornoceratids (*Falcitornoceras* aff. *bilobatum, Armatites planidorsatus, Arm*. aff. *planidorsatus*), *Lobobactrites paucesinuatus*, rare *Cheiloceras*, and smooth brachiopods.

Bed 63b (= Fa 6b of the new numbering from 2012) is a distinctive goniatite coquina (Pl. 2, Fig. 1) or cephalopod rudstone (Pl. 2, Fig. 1) with masses of cheiloceratids (Figs. 6a-b), including the *Cheil. (Compactoceras) verneuilii* and *undulosum* Groups, and other goniatites (*Falcitornoceras*). It indicates a sudden peak of eutrophication, which improved the living conditions for goniatites during decreased clastic input. This reflects an initial phase of the Upper Condroz Event (BECKER 1993, *rhomboidea* Zone regression in WENDT & BELKA 1991). The ca. 8 m thick marls of the subsequent slope are interrupted by three thin intervals of calcareous siltstone/calcisiltite (Beds Fa 7b, 9a-d, $10b_{1\text{-}5}$) and by one thin goniatite limestone (Bed Fa 8b, Fig. 5). Again there is a minute hematitic fauna (loose from Beds Fa 7a-10a), which

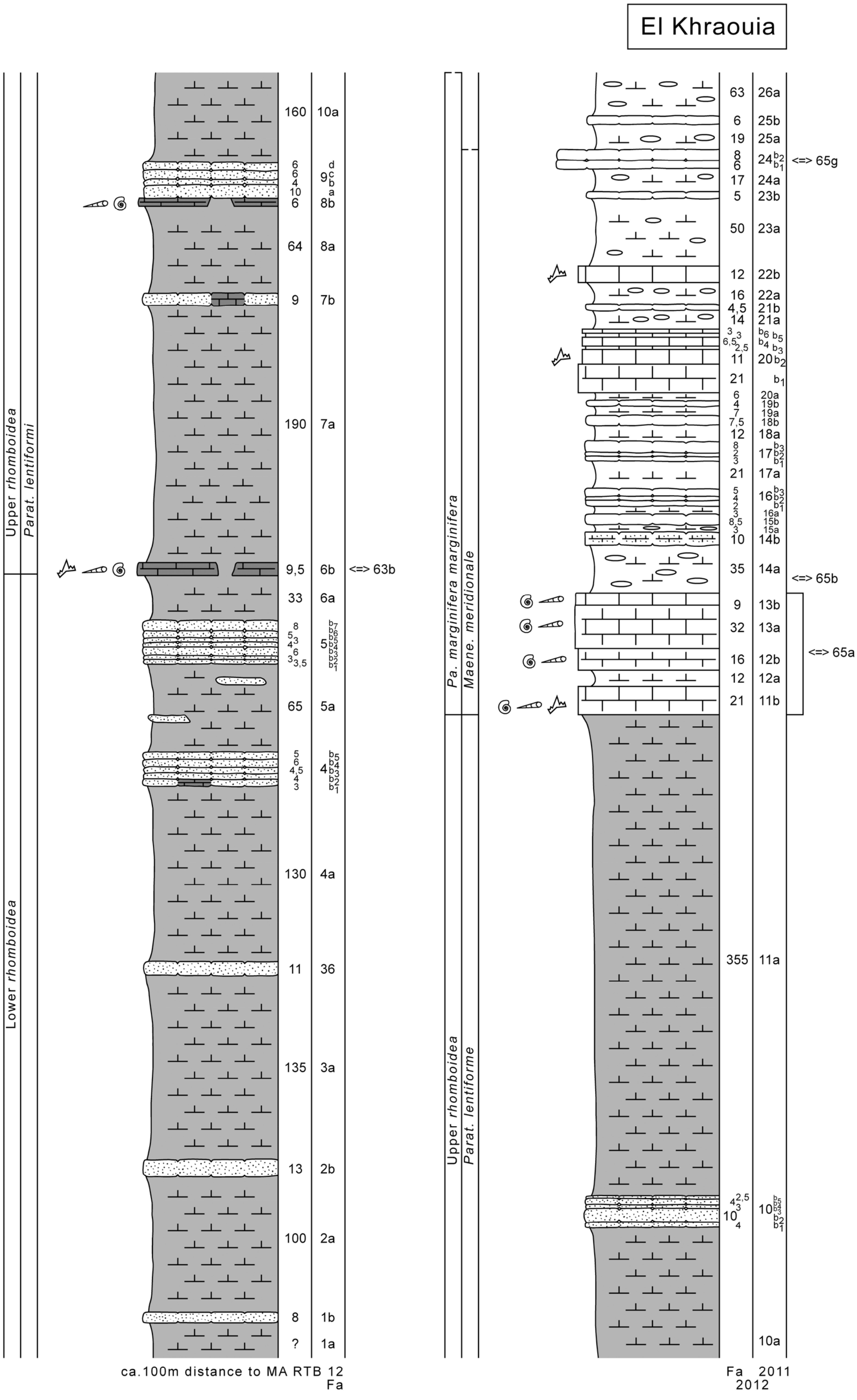

El Khraouia
Upper rhomboidea
Parat. lentiformi
Lower rhomboidea
Pa. marginifera marginifera
Maene. meridionale
Upper rhomboidea
Parat. lentiforme
<=> 63b
<=> 65g
<=> 65b
<=> 65a
ca.100m distance to MA RTB 12
Fa
Fa 2011
2012

Fig. 5 (part 1).

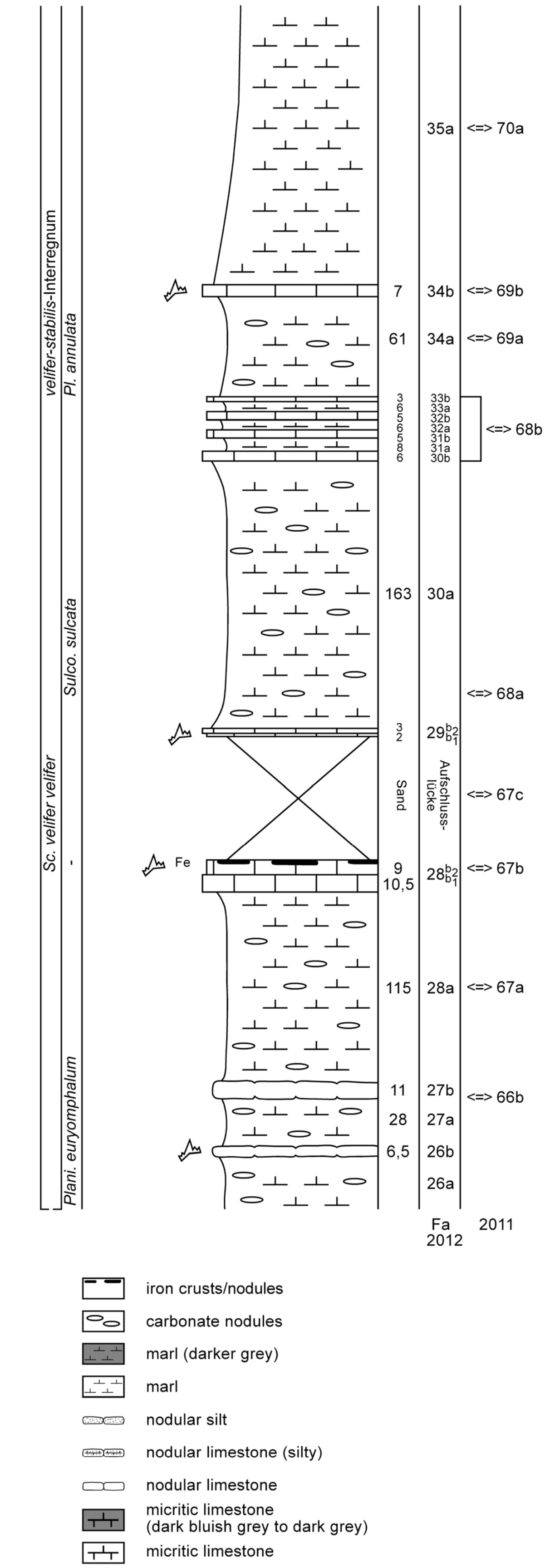
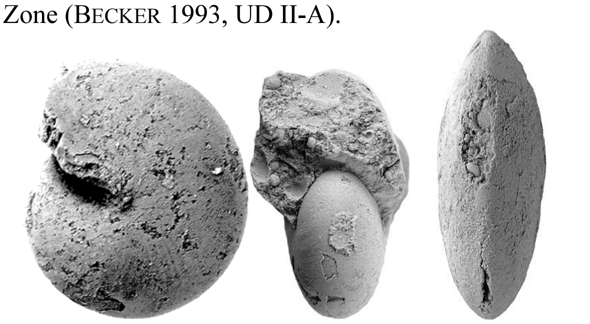


Fig. 5 (part 2). Top lower to basal middle Famennian lithological log, position of conodont samples, and biostratigraphy at El Khraouia.

includes the same species as below and, in addition, rare *Cheiloceras (Cheil.)* cf. *amblylobum* and *Paratorleyoceras globosum*.

Age: Bed 60f contains a typical basalmost Famennian conodont association (Lower *triangularis* Zone), which consists of *Pa. ultima, Pa. subperlobata* (pl. 1, Fig. 11), and rare *I. alternatus alternatus* (Pl. 1, Fig. 12). Despite the calm environment, which speaks against reworking, there is a single, last *Ad. curvata*. It may have survived the main conodont extinction for a very short time. A similar, rare basal Famennian occurrence was previously reported from the Montagne Noire (BECKER & HOUSE 1994a: section MP-SE-a) but there explained by reworking. In the goniatite zonation, Beds 60f and 60h fall in the *frechi* Zone (BECKER 1993, UD II-A).



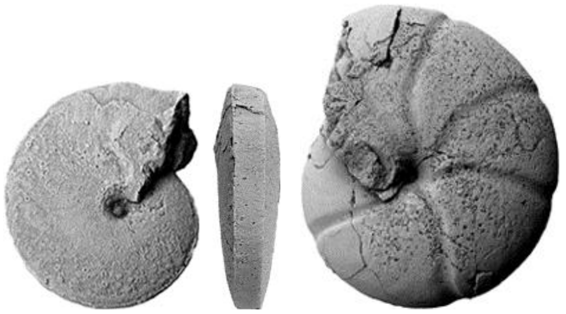
a. b. c.

Fig. 6. Goniatites from Bed 63b (= Fa 6b). **a.-b**. *Cheiloceras* (*Compactoceras*) *verneuilii* Gp. (46 mm dm), **c.** ventral view of *Parat*. cf. *lentiforme* (53 mm dm), index of UD II-F.

The goniatite association of Beds Fa 3a-5a is typical for the *globosum* Zone (UD II-D, BECKER 1993); the UD II-B/C interval must lie in Beds 61/62. Bed Fa 6b is the widespread *Cheiloceras undulosum* Bed of the Tafilalt (BECKER 1993). The new collection from El Khraouia, supported by unpublished data from Rich Haroun (central Tafilalt Platform), includes as a minor faunal element *Paratornoceras* cf. *lentiforme* (Fig. 6c). This re-dates the regional marker bed as UD II-F, instead as top II-E1, as in the last zonation review of BECKER et al. (2003). The position high in the lower Famennian is confirmed by a rich conodont fauna of the Upper *rhomboidea* Zone (Tab. 1), with characteristically abundant *Pa. rhomboidea* (Pl. 3, Fig. 14). The dominance of the *I. cornutus* Group (almost 70 % of the fauna, Pl. 3, Fig. 3) contrasts with the overall basinal setting and supports a regressive episode. *Po. vetus* is a new record for North Africa. The loose goniatites from Beds Fa 7a-10a closely resemble the fauna from below and include the index species of the *globosum* Zone (UD II-D); they may not come from levels above Bed Fa 6b.

3.2. Middle Famennian

The middle Famennian begins with a ca. 90 cm thick package (Beds Fa 11b-13b) of predominant cephalopod limestone (*Maeneceras* Limestone). Bed Fa 11b is a cephalopod-rhynchonellid rudstone, with evidence of polyphase reworking and increasing influence of bottom currents at the base (Pl. 2, Figs. 2-3). The environment has rather suddenly changed from deeper basin/slope to a deeper carbonate platform (see WENDT et al. 1984). The *Maeneceras* Beds and platform progradation continue further eastwards to the Hassi Nebech region (see BOCKWINKEL et al. 2002). As elsewhere in the Tafilalt, the base of this “IIß Limestone” is a sequence boundary (final phase of Upper Condroz Event, BECKER 1993), followed by the global eustatic rise that characterizes the base of the middle Famennian (see sea-level curve in WENDT & BELKA 1991).



a. **b.** **c.**

Fig. 7. Goniatites from the *Maeneceras* Beds (Beds 11b-13b, UD II-G). **a-b.** *Armatites planidorsatus* (28 mm dm), **c.** *M. meridionale meridionale* (31.5 mm dm).

The middle part of the middle Famennian is less fossiliferous and consists of ca. 4.5 m alternating solid limestone (beds up to 21 cm, Fa $20b_1$) and nodular marls (Beds Fa 14a to 27b, Fig. 5; thicker than thought during the rough first survey in 2011, = Beds 65e-66b). Bed Fa 22b is a representative, recrystallized wacke-packstone with considerable fine siliciclastic influx (Pl. 2, Figs. 4-5), which suggests a shallowing of the platform. Towards the west (Amessoui Syncline) the quartz content increases further in brachiopod-rich limestones (WENDT et al. 1984).

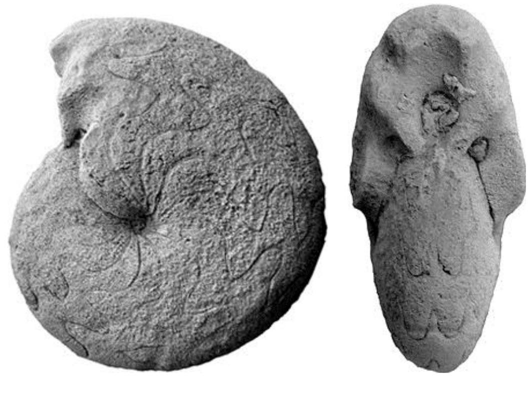


Fig. 8. Loose *Sporadoceras angusisellatum* from Bed 67 (= Fa 28a-29a), UD III-B/C, 46.5 mm dm.

An upper part of the middle Famennian consists of nodular marl intervals (Beds Fa 28a and 29a, = 67a and 67c) sitting in the upper part of the low Famennian cliff. These units with *Sp. angustisellatum* (Fig. 8), *Xenosporadoceras spiriferum*, orthocones and gastropods are separated by 20 cm limestone with an iron crust at the top (Beds Fa $28b_{1-2}$). One 1 km to the north, a slightly different fauna with *Enkebergoceras* is preserved as goethitic/hematitic moulds (altered pyrite). This suggests that a hypoxic setting developed during a minor deepening phase, which also can be recognized in the central and northern Tafilalt (e.g., Jebel Erfoud, HARTENFELS 2011, Bine Jebilet). The subsequent marls (Bed Fa 30a) and limestones (Beds 30b-33b) have not yet been sampled sufficiently for ammonoids.

Age: Beds Fa 11b (= 65a) to 13b contain the index goniatite of the *M. meridionale* Zone (UD II-G, Fig. 7, previously named as *M. subvaricatum*, BECKER et al. 2002), and abundant *Arm. planidorsatus* (Fig. 7). The long-known (BECKER 1993) correlation with the (Lower) *marginifera* Zone is confirmed by a diverse conodont fauna from Bed 11b with 16 species (Tab. 1), including two zonally diagnostic subspecies of *Pa. quadrantinodosa* (Pl. 3, Fig. 13) and the late morphotype of *Pa. glabra lepta* (Pl. 3, Fig. 8). *Pa. glabra prima* M3 is dominant (Pl. 3, Fig. 10), *Polylophodonta* rare. *Pa. marginifera marginifera* enters in Bed $20b_2$ (Pl. 3, Fig. 11) together with the local oldest records of *Mehlina* and *Branmehla* (Pl. 3, Fig. 2). In this sample *Pa. gracilis gracilis* is relatively dominant. Moderately common *Bispathodus stabilis vulgaris* indicate the Upper *marginifera* Zone (ZIEGLER & SANDBERG 1984) since this species occurs only very rarely below (CAPKINOGLU 2005, RYTINA et al. this vol.). The disappearance of *Pa. quadrantinodosa* lends further support for this interpretation. *Bi. stabilis vulgaris* increases in abundance in Bed 22b (Pl. 3, Fig. 1) and above. Only an undivided extended *marginifera* Zone is recognizable in the Anti-Atlas (HARTENFELS 2011).

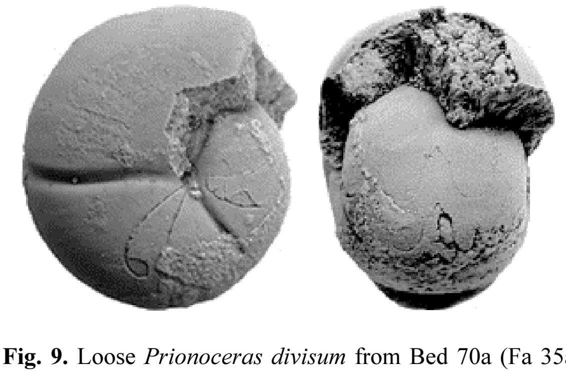


Fig. 9. Loose *Prionoceras divisum* from Bed 70a (Fa 35a), probably *annulata* Zone (UD IV-A), 25 mm dm.

Two subspecies of the nominate species (see Pl. 3, Fig. 24) prove that Bed 26b falls in the lower part of the *Scaphignathus velifer* Zone (see HARTENFELS 2011). *Pa. perlobata schindewolfi* and *“Po.” diversus*, an unusual homoeomorph of *Skeletognathus* (Pl. 3, Figs. 16-18), are very common. In Bed $28b_2$ the latter even becomes the dominant taxon, followed in abundance by *Br. inornata*; this association gives a rather unusual conodont biofacies. *Sp. angustisellatum* from Beds Fa 28a-29a is a marker species of the regional *Planitornoceras euryomphalum* Zone (UD III-B/C, BECKER et al. 2002). A single *Pa. gracilis semisigmoidalis* from Bed 29b provides a correlation with the *trachytera* Zone of the standard succession (Hartenfels 2011), which index species is so far unknown from the Anti-Atlas. The goniatite-conodont correlation agrees with Germany (Korn & Ziegler 2002) and NW Australia (Becker & House 2009).

3.3. Upper/uppermost Famennian

Our tentative placing of the middle/upper Famennian substage boundary follows the proposal of Hartenfels et al. (2009). Bed Fa 34a (= Bed 69a) may represent the Lower *Annulata* Event Interval, based on loose platyclymeniids and prionoceratids. A thin marker limestone at the top (Bed Fa 34b = 69b) is a bioturbated, silty mud-wackestone (Pl. 2, Fig. 6) and indicates a drowning of the hemipelagic carbonate platform. This local deepening trend continues in the subsequent ca. 14 m thick marls (Bed 70a = Fa 35a), which yielded a diverse ammonoid fauna in (altered) pyritic or limestone preservation. The same facies continues for 16 m after a thin limestone (Bed 70b) and ends with a thin nodular limestone (Bed 71b). Overlying marls (“Bed 72”) with rare clymeniids are hardly exposed and mostly covered by debris from the siliciclastics of the Aoufital Formation, which has not yet been studied. The Devonian-Carboniferous boundary should lie within the ridge formed by (pro)deltaic clastics.

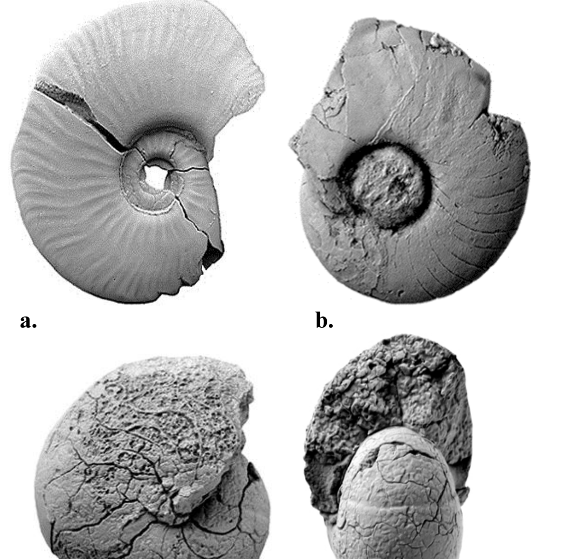


Fig. 10. Loose ammonoids from Bed 70a (= Fa 35a)**. a.** *Procym. pudica* (47.5 mm dm), **b.** *Protacto*. aff. *pulcherrima* (40 mm dm), **c-d.** *Sp. orbiculare* (42 mm dm).

Age: The ammonoid fauna from around Bed Fa 34a falls in the *annulata* Zone (UD IV-A). Accordingly, the conodont assemblage of Bed 34b, without *Scaphignathus* (Tab. 1), characterizes the regional *velifer-stabilis* Interregnum (Hartenfels 2011). The subsequent loose, pyritic *Prionoceras divisum* (Fig. 9), *Pr. frechi, Platy.* cf. *annulata, Erfoudites ungeri*, *Protactoclymenia* sp. (Fig. 10b), and *Protoxyclymenia* sp. indicate UD IV-A/B for Bed 70a (= Fa 35a). This fauna may have been derived partly from the Upper *Annulata* level. Calcareous *Procymaclymenia pudica* (Fig. 10a), *Prot. ventriosa*, and *Sp. orbiculare* (Fig. 10c-d) are index species of UD IV-B/C (upper part of *Platyclymenia* Stage or upper Hembergian). It seems that the *orbiculare* marker bed of the Tafilalt (UD IV-C, Becker et al. 2002) is locally developed. Loose *Gonioclymenia* and *Cymaclymenia* from Bed 71 represent UD V. *Medioclymenia aguelmousensis* (Fig. 11a) and *Muessenbiaergia bisulcata* (Fig. 11b) show that the succession reaches at least the top Dasbergian and lower Wocklumian (UD V-C - VI-B, Becker et al. 2002). It is unlikely that the uppermost Famennian is complete since the *Wocklumeria* Zone (UD VI-D) is missing beneath the Hangenberg Event all over the Tafilalt (Kaiser et al. 2012).

Fig. 11. Loose, fragmentary clymeniids from “Bed 72”. **a.** *Muess. bisulcata* (UD VI-B), fragment length = 41 mm, **b.** *Medio. aguelmousensis* (UD V-C), length = 44 mm.

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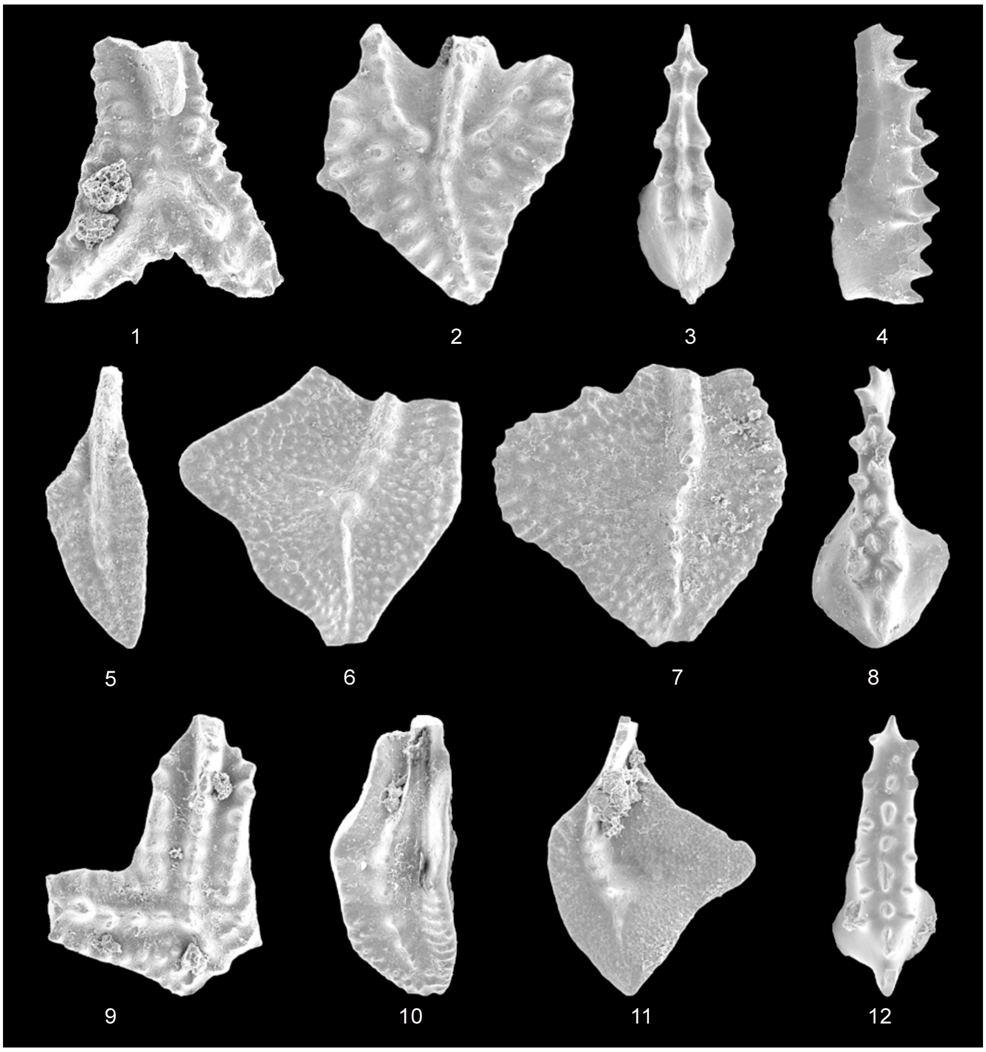
T. FÄHRENKEMPER produced the section logs, E. KUROPKA processed most of the conodont samples.



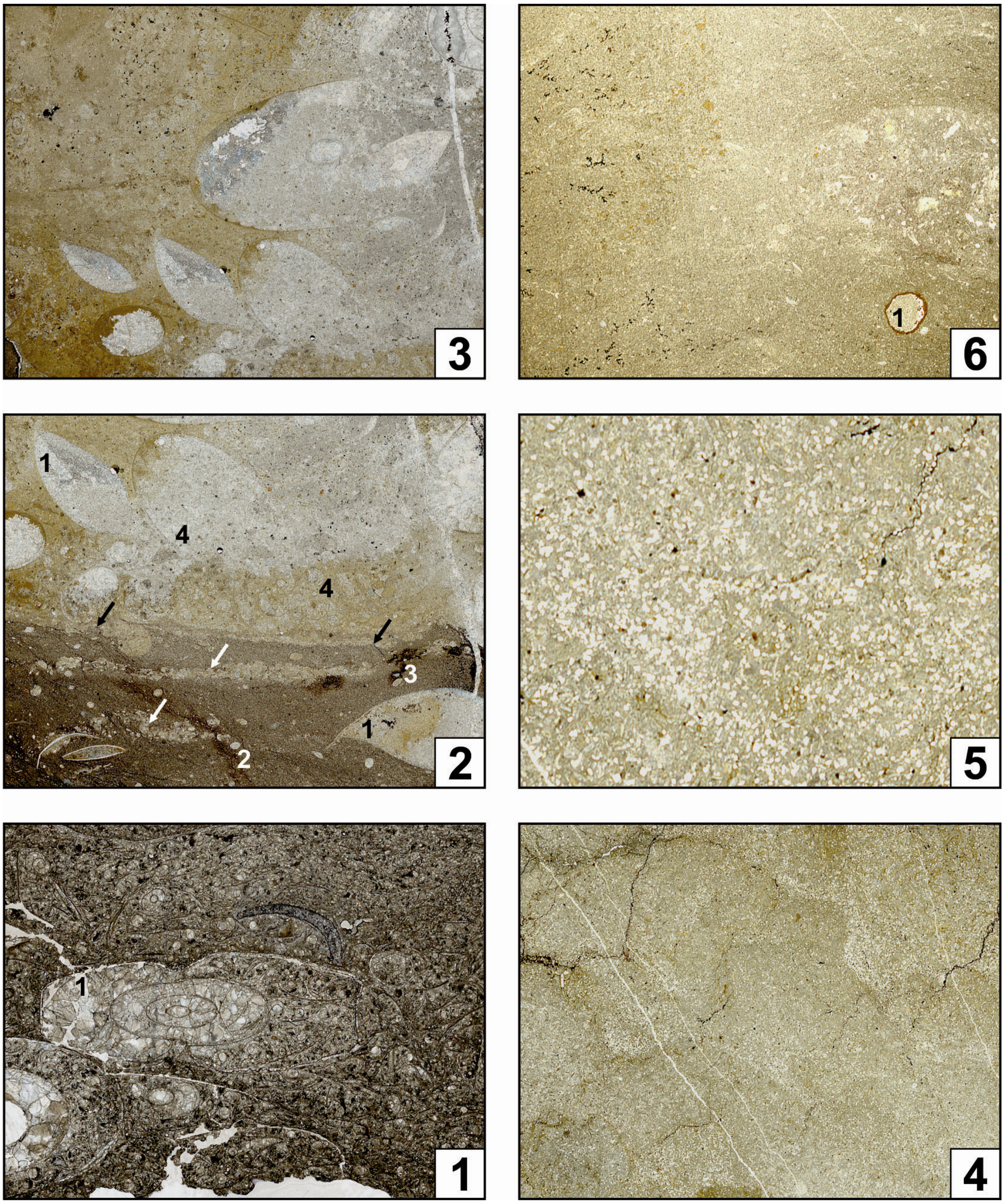
Pl. 1. Frasnian to basal Famennian conodonts from El Khraouia, 1-5: Bed 54 (regional *coeni* or MN 8 Zone), 6-10: Bed 56d (regional *plana* or MN 10 Zone, 11-12: Bed 60f (Lower *triangularis* Zone). **1.** *Ag. coeni*, **2.** *Ad. curvata* Early Morphotype,

3-4. *I.* cf. *vitabilis*, **5.** *Po*. aff. *paradecorosus*, **6.** *Pa. hassi*, **7.** *Pa. plana*, **8, 12.** *I. alternatus alternatus* (two different morphotypes), **9.** *Ag. triangularis*, **10.** *Po. evidens*, **11.** *Pa. subperlobata*.

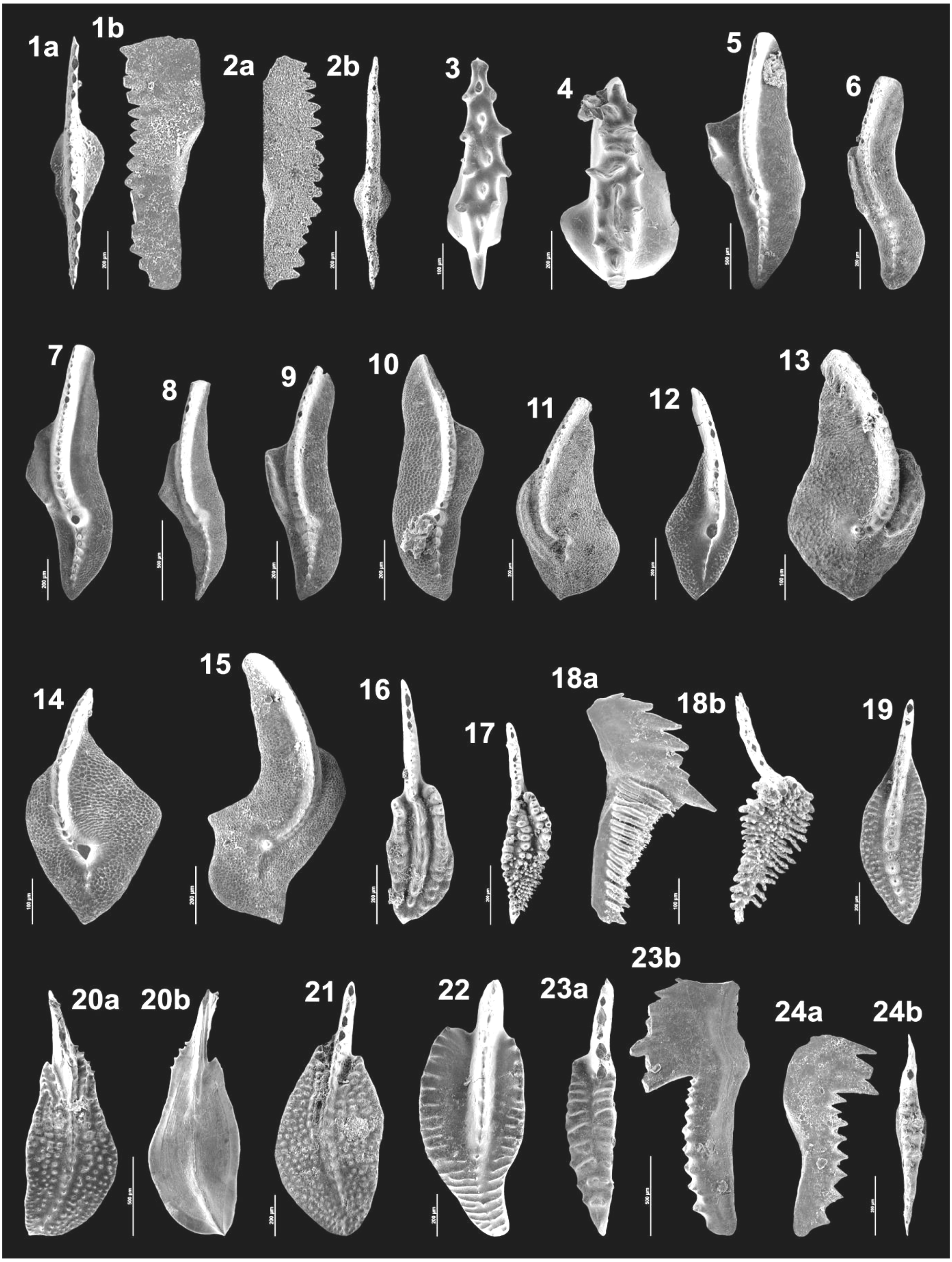
regional conodont zonation	Upper rhom.	*Pa. marginifera marginifera*			*Sc. velifer velifer*			*v.-st.-Inter.*
Bed-No.	6b	11b	$20b_2$	22b	26b	$28b_2$	$29b_1$	34b
I. iowaensis Group	10							
Pa. glabra lepta Early Morphotype	8							
Pa. rhomboidea	42							
Po. transitus	1							
Pa. glabra acuta	2	22						
Pa. glabra prima Morphotype 3	17	110	11					
Pa. glabra pectinata Morphotype 2	3	40	8					
Po. nodocostatus ovatus	2	-	-	6				
Po. nodocostatus nodocostatus	26	-	8	32	5			
Po. semicostatus "Central Morphotrend"	2	-	5	4	2	6		
Po. semicostatus "Morphotrend 3"	2	-	-	-	1	6		
Po. semicostatus "Morphotrend 8"	2	-	-	-	-	1		
Po. glaber glaber	1	-	1	1	1	5		
Pa. minuta minuta	11	3	-	-	-	2	1	
"*I.*" *cornutus* Group	319	4 (3juv)	16	28	-	-	-	3
Pa. perlobata schindewolfi	11	26	38	79	143	22	26	8
Caenodontus sp.	1 (?)	-	-	-	-	-	-	1
Po. fallax		2						
Pa. quadrantinodosa quadrantinodosa		8						
Pa. quadrantinodosa inflexoidea		1						
Po. pseudotenellus		20						
Polylo. concentrica		1						
Polylo. confluens		1						
Pa. perlobata helmsi		1	6	2	5			
"*Po.*" *diversus*		3	11	12	115 (8Pb)	129 (1Pb)		
Pa. glabra lepta Late Morphotype		7	-	-	-	11		
Neo. communis		10	-	-	-	-	-	9
Po. glaber medius			2					
Pa. marginifera marginifera			7	1	-	3		
Pa. minuta schleizia			1	-	-	15	74	
Br. inornata			33	6	57	64	28	
Pa. gracilis gracilis			75	-	2	8	201	69
M. strigosa			18	5	2	7	-	1
Bi. stabilis vulgaris			33	50	231	8	3	15
Br. ampla			62	9	5	3	2	1
Pa. glabra distorta				1	-	18		
Pa. perlobata maxima				7	5	1	1	
Pa. rugosa aff. *ampla*					4			
Po. aff. *semicostatus*					1	5		
Sc. velifer velifer					10	25	23	
Sc. velifer leptus					3	7	10	
Po. semicostatus "Morphotrend 6"						2		
Po. semicostatus "Central Morphotrend" → "Morphotrend 8"						1		
Ct. brevilaminus						1	-	1
Pa. gracilis semisigmoidalis							1	
Po. perplexus							2	3

El Khraouia

Tab. 1. Ranges and abundance of conododont taxa and morphotypes in the top lower to basal upper Famennian at El Khraouia, based on samples from spring 2012.



Pl. 2. Microfacies of Famennian beds from El Khraouia. **1.** Bed 6b, *undulosum* Bed, Upper *rhomboidea* Zone, width ca. 30 mm; cephalopod rudstone with *Falcitornoceras* (1), cheiloceratids (left corner), crinoid remains, fragmented thin-shelled bivalves, and few ostracods; micritic matrix generally washed out, only preserved in some protected interspaces. **2.** Bed 11b, basal part, *marginifera* Zone, width ca. 28 mm; irregular karst surfaces (black arrows) document the multiphase character of a microsparitic limestone, which basal part is a brownish wackestone with some silt, abundant ostracods (2), crinoid remains (3), and shell fragments. Two layers (white arrows) of small-sized crinoid remains and siliciclastics document episodic reworking and accumulation by bottom currents. Above the karstic surface follows a cephalopod (4) – rhynchonellid (1) rudstone with shell detritus and ostracods in a microsparitic matrix. **3.** Bioturbated middle part of Bed 11b, width ca. 33 mm; with cross-sections of brachiopods (*Perrarisinurostrum*), of an orthocone, and *Maeneceras*. **4.** Bed 22b, *marginifera* Zone, width ca. 29 mm; bioturbated microsparitic wacke-packstone with fine silt, fragmented bivalves, ostracods, crinoid remains, and rare gastropods. **5.** Details of Bed 22b, showing the silty microsparite matrix. **6.** Bed 34b, regional *velifer-stabilis*-Interregnum, width ca. 18 mm; microsparitic mud-wackestone with fine silt, ostracods, fragmented cephalopods, thin-shelled bivalves, and *Frutexites*-type (1) brownish to golden encrustions of bioturbation structures.



Pl. 3. Famennian conodonts from El Khraouia. **1.** *Bi. stabilis vulgaris*, Bed 22b, **2.** *Br. ampla*, Bed 20b$_2$, **3.** “*I.*” *cornutus* Gp, Bed 6b, **4.** *I. iowaensis* Gp, Bed 6b, **5.** *Pa. glabra acuta*, Bed 6b, **6.** *Pa. glabra distorta*, Bed 28b$_2$, **7.** *Pa. glabra lepta* Early Morph., Bed 6b, **8.** *Pa. glabra lepta* Late Morph., Bed 11b, **9.** *Pa. glabra pectinata* M2, Bed 11b, **10.** *Pa. glabra prima* M3, Bed 6b, **11.** *Pa. marginifera marginifera*, Bed 20b$_2$, **12.** *Pa. minuta minuta*, Bed 6b, **13.** *Pa. quadrantinodosa quadrantinodosa*, Bed 11b, **14.** *Pa. rhomboidea*, Bed 6b, **15.** *Pa. rugosa* aff. *ampla*, Bed 26b, **16-17.** ”*Po.*” *diversus*, Bed 26b, **18.** ”*Po.*” *diversus*, Pb element, Bed 26b, **19.** *Po. fallax*, Bed 11b, **20.** *Po. nod. nodocostatus*, Bed 6b, **21.** *Po. nod. ovatus*, Bed 22b, **22.** *Po. semicostatus* “M8”, Bed 6b, **23.** *Sc. velifer velifer*, Bed 28b$_2$, **24**. *Sc. velifer leptus*, Bed 26b.

THE LOCHKOVIAN TO EIFELIAN SUCCESSION OF THE AMESSOUI SYNCLINE (SOUTHERN TAFILALT)

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1. INTRODUCTION

The outstanding quality of Palaeozoic outcrops in Morocco is well known and is reflected by the amount of publications on Moroccan Palaeozoic fossils in general and of Devonian fossils in particular. Palaeontological pioneer work on the Devonian sedimentary successions of the Tafilalt was carried out by CLARIOND (1934a, b), ROCH (1934), TERMIER & TERMIER (1950), PETTER (1959), MASSA et al. (1965), and particularly HOLLARD (1967, 1974, 1981). Much of today’s stratigraphical research in this region is founded on their contributions.

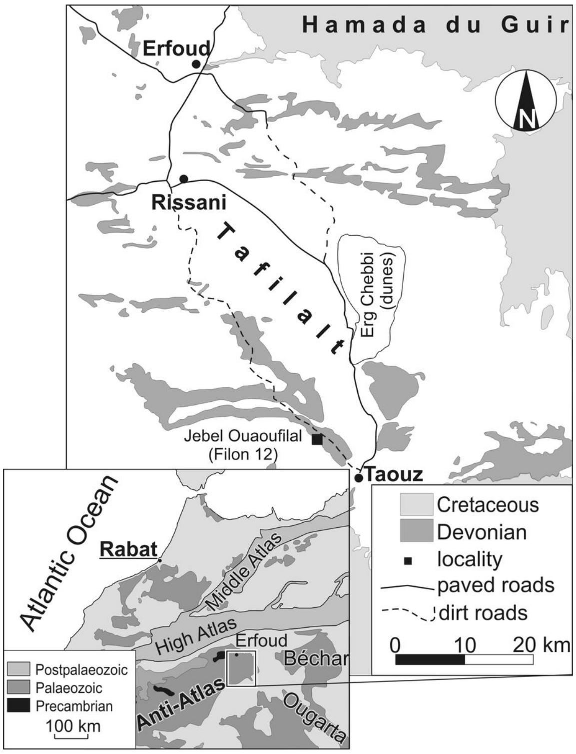


Fig. 1. Map of the Tafilalt with the location of the outcrop.

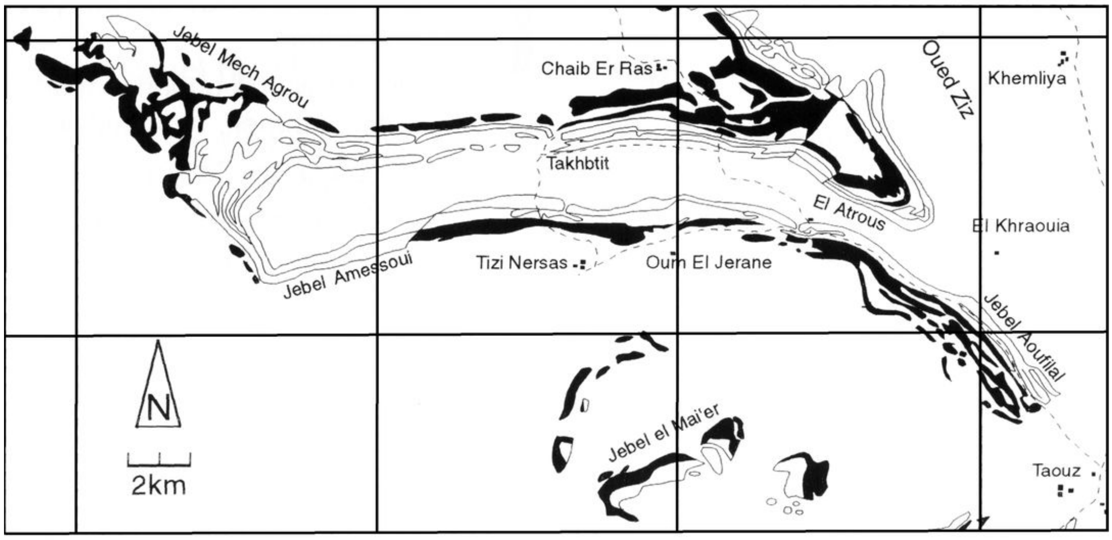
In the Tafilalt, Devonian sedimentary rocks crop out in a series of East-West trending synclines (Fig. 1), which are partially cut off by reverse faults with strike-slip components at their southern flanks (TOTO et al. 2008). Usually, these faults mainly affected the shaly parts of the sequence from the Silurian and parts of the Early Devonian. This fact, in combination with the more shaly composition of the Early Devonian part of the sequence, accounts for the scarcity of complete Early Devonian exposures. At least the claystone parts of the succession are almost always covered by thin Neogene deposits. One exception is the Ouidane Chebbi section East of Erfoud, which has been published in detail by BELKA et al. (1999). Nevertheless, some of the aforementioned synclines west of the Ziz Valley also expose the Early and Middle Devonian succession well. Sedimentological processes have been examined by WENDT (1985, 1988) and WENDT et al. (1984).

Here, we will focus on the south-eastern part of the largest of these synclines, named Amessoui Syncline after the Jebel Amessoui (979 m) near the western end of the syncline (Fig. 2). In the region between Jebel El Atrous, the mine of Filon 12, El Khraouia, the Jebel Ouaoufilal, and Taouz, much of the Early and Middle Devonian rock succession is exposed and quite fossiliferous (Fig. 2). Early and Middle Devonian fossils of this area have been published by, e.g., TERMIER & TERMIER (1950), PETTER (1959), BECKER & HOUSE (1994), KLUG et al. (2000, 2008), KLUG (2002a, 2002b, 2007), KLUG & KORN (2002), KRÖGER (2008), and by DE BAETS et al. (2010).

2. LOCHKOVIAN

The *Scyphocrinites* Beds are well-exposed at the Silurian/ Devonian boundary; partially, they can be seen in trenches, where the articulated crowns of this planktonic crinoid were excavated by the locals. The exact position of the boundary is disputed. HAUDE & WALLISER (1998) found *Scyphocrinites* loboliths within beds with *detorta - eosteinhornensis* (Late Pridoli) to *remscheidenensis - woschmidti* conodont faunas (Late Pridoli - ?Early Lockovian) east at Chaib-er-Ras in the Amessoui Syncline (Locality 477). They pointed out the report of the graptolite “*Monograptus* plutôt *uniformis* que *praehercynicus*” SE of Bou Faddouz by WILLEFERT in HOLLARD (1977) in beds, which correlate to just above their conodont samples interval and might still contain *Scyphocrinites*. They were therefore unsure where to place the boundary within or above the *Scyphocrinites* beds. According to BELKA et al. (1999), the boundary is actually below the *Scyphocrinites* Beds and therefore, these limestones would be entirely of

Early Devonian outcrops in the Amessoui-Syncline



Middle Devonian outcrops in the Amessoui-Syncline

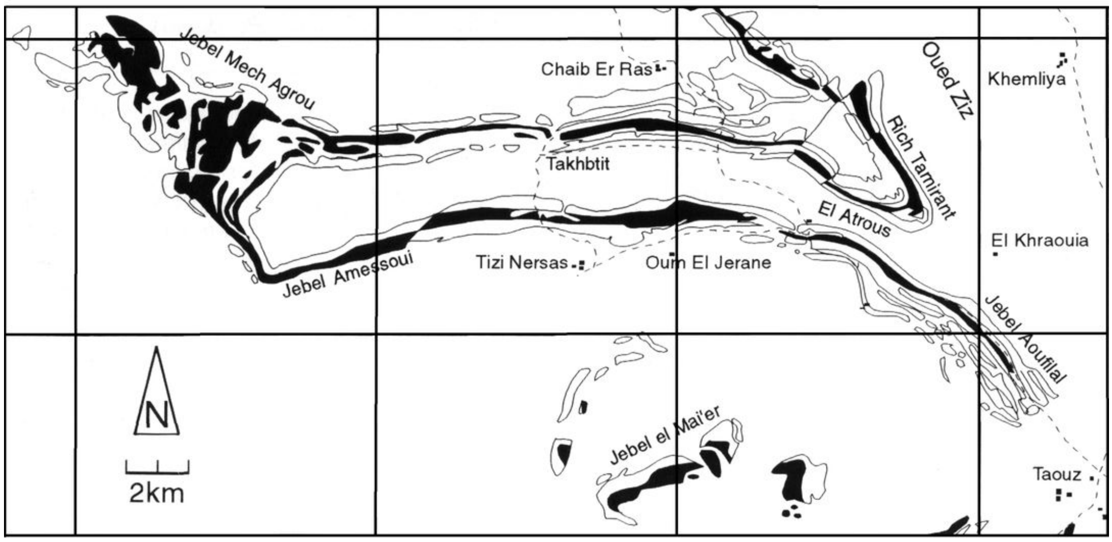
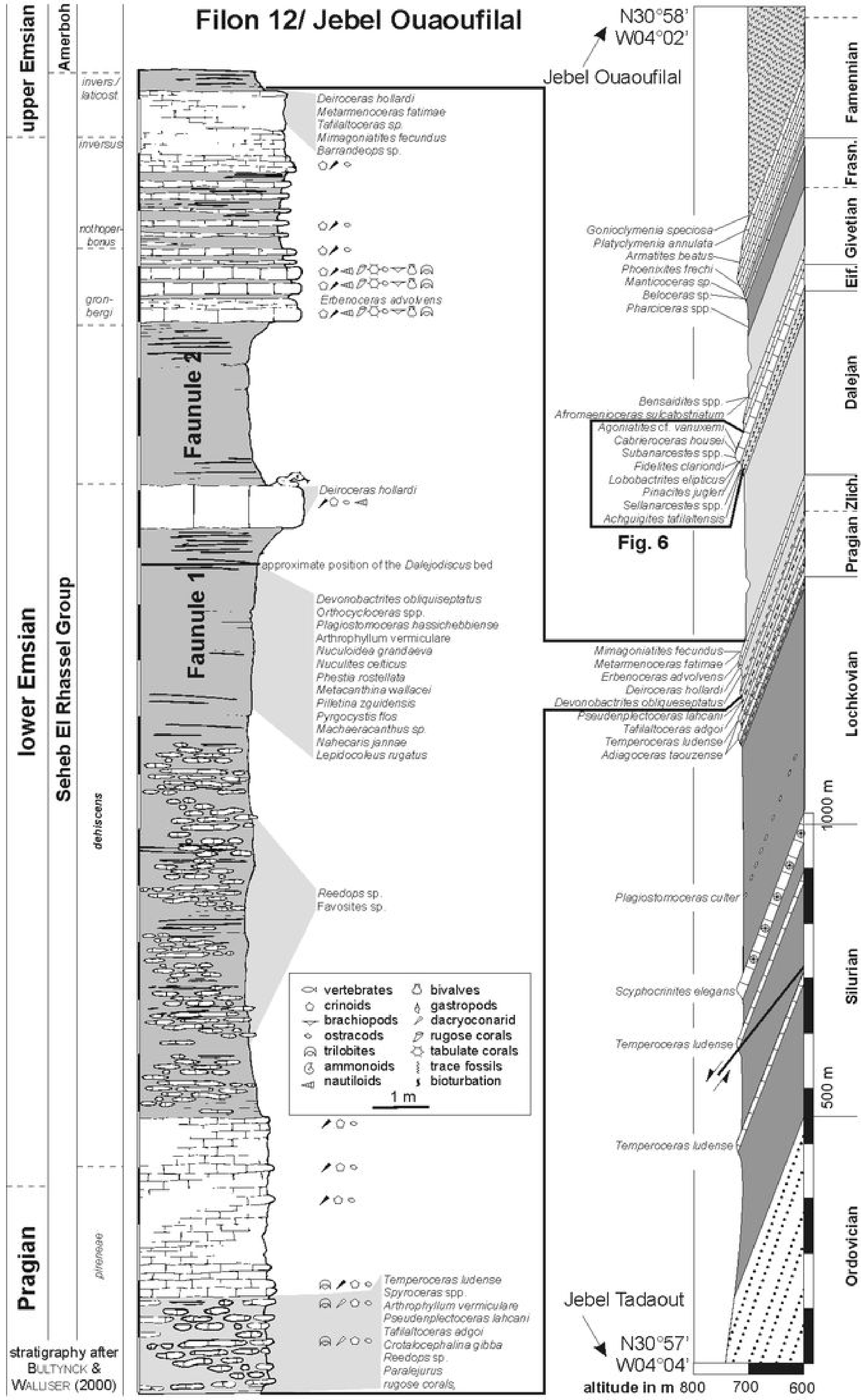


Fig. 2. Maps of the Amessoui-Syncline with the distribution of Early and Middle Devonian outcrops.

Fig. 3. The Pragian part of the Jebel Tadaout to Jebel Ouaoufilal section near Filon 12 (image courtesy Richard HOFMANN, Zürich).

Fig. 4. Left: The Pragian to early Emsian part of the section (modified after KLUG et al. 2008). **Right:** Generalised section of the Ordovician to Tournaisian sediments between Jebel Tadaout and Jebel Ouaoufilal (near Filon 12 and Taouz). The strata dip uniformly 20° to the NE. The Silurian orthocone limestone is duplicated, probably due to a thrust fault.



Lochkovian age in Ouidane Chebbi. BELKA et al. (1999) found *Latericriodus postwoschmidti*, *Ozarkodina remscheidensis remscheidensis*, and *Panderodus* cf. *semicostatus* in the lower part of the carbonate beds. This stratigraphic interpretation needs to be confirmed, however, for the Filon 12-El Atrous-Jebel Ouaoufilal-area; KRÖGER (2008) drew the boundary within the *Scyphocrinites* Beds.

The *Scyphocrinites* Beds consist of 1 to 20 cm thick limestone beds, in which complete crowns, loboliths, and more or less articulated stems of this crinoid occur in rockforming numbers. The crinoids are accompanied by a few bivalves, platyceratid gastropods and some orthocerids. The crinoid limestones are interbedded with marly limestone beds with abundant cephalopods with orthoconic shells; KRÖGER (2008) listed the following cephalopod taxa: *Orthorhizoceras*, *Ankyloceras*, *Subormoceras*, *Kopaninoceras*, *Merocycloceras*, *Orthocycloceras*, *Parakionoceras*, *Hemicosmortho-ceras*, and *Akrosphaerorthoceras*.

The *Scyphocrinites* Beds are overlain by dark claystones, which appear pink when weathered. About 40 m above the *Scyphocrinites* Beds, some marls are locally exposed; they contain nodules with abundant *Plagiostomoceras culter* (KRÖGER 2008) and small brachiopods. 100 m higher in the section, several more carbonatic layers occur, the first of which are full of orthocones and rare accompanying fauna. This increase in carbonate, clastic components and benthic fauna reflects a regression, which KRÖGER (2008) correlated with the “*pesavis* Bioevent” or “Lochkov/Prag Event” of SCHÖNLAUB (1996, CHLUPÁČ & KUKAL 1996). In the area between El Atrous and Taouz, the following sequence of latest Lochkovian, Pragian and early Emsian age forms four to five low ridges. The stratigraphically oldest part has been named “*Jovellania* limestone” by KRÖGER (2008). The early Emsian (Zlíchovian) ‘*Jovellania*’ limestone of HOLLARD (1981) and BULTYNCK & WALLISER (2000a, 2000b) was later renamed “*Deiroceras* limestone” by KRÖGER (2008). KRÖGER’S (2008) *Jovellania* Limestone supposedly contains the Lochkovian-Pragian boundary, while BECKER & ABOUSSALAM (2011) place the boundary just above the *Jovellania* Limestone. At the base, there is a nodular layer of black limestone full of *Adiagoceras taouzense*, *Hemicosmorthoceras semi-imbricatum*, and *Temperoceras ludense*, together with *Jovellania* as well as bivalves. About 3 m higher in the section, a layer contains large *Temperoceras ludense*, *Adiagoceras taouzense*, *Bohemojovellania adrae*, and nowakiids as well as also some gastropods, bivalves, and other benthic fauna. The abundance of benthic organisms is slowly increasing, although the low diversity in benthos in combination with the dark colour of the limestone bed point at low oxygen-conditions. KRÖGER (2008: fig. 7) figured two sandstone beds above the aforementioned orthocone limestone, which he interpreted as the top of the Lochkovian part of the section.

3. PRAGIAN

The various definitions of the Pragian boundaries are differing (e.g,. BULTYNCK & WALLISER 2000a, b). Herein, we simply follow KRÖGER (2008) and leave the argument about the position of the boundary to other authors. The Pragian part of the succession in the Filon 12 area (Figs. 3, 4) consists of sedimentary rocks with lighter colour than in the Lochkovian succession (compare BELKA et al. 1999). Additionally, it is more carbonatic and contains more benthic organisms as well as sand (KRÖGER 2008). About 15 m above the Lochkovian sandstone beds, four limestone layers follow, with 2 to 5 m thick marly sequences separating them. Especially the third of the limestone beds (“K3” of KRÖGER 2008) and the subsequent marls are rich in trilobites (*Crotalocephalina*, *Reedops*, *Paralejurus*), ostracods, dacryoconarids, echinoderms, gastropods, tabulate and rugose corals, as well as cephalopods (*Temperoceras ludense*, *Spyroceras* spp., *Arthrophyllum vermiculare*, *Pseudenplectoceras lahcani*, *Tafilaltoceras adgoi*). Additionally, bryozoans, hederelloids, hyolithids and macroscopic remains of other groups of Metazoans can be found. Especially in the marls, the shells are locally slightly silicified and thus weather nicely out of the sediment.
Above the last of the four limestone beds, about 6 m of marly nodular limestones follow. The fauna is similar to that of the preceding units, but usually slightly less diverse and less well preserved. In the subsequent greenish claystones, the Pragian-Emsian boundary is located.

4. EMSIAN

Again, the Pragian/ Emsian boundary has been discussed a lot over the past decades and it is not the purpose of this chapter to review all possibilities, opinions and fantasies about this boundary. The interested reader will find more information in, e.g., BULTYNCK & WALLISER (2000a, b), CARLS et al. (2008), BECKER & ABOUSSALAM (2011), which contains *Nowakia zlichovensis maghrebiana* in the eastern Tafilalt (Ouidane Chebbi). The marly beds of the Pragian grade into greenish claystones, which contain a rich limonitic fauna comprising probably over 100 species, described as “faunule 1” by KLUG et al. (2008) and by DE BAETS et al. (2010). These claystones can be seen near the right edge in Fig. 4. Large pseudomorphoses of limonite after pyrite weather out of the fine-grained sediment. Body chambers of *Devonobactrites obliquiseptatus* dominate this fauna, accompanied by orthocones such as *Orthocycloceras* spp., *Plagiostomoceras hassichebbiense*. Furthermore, a wealth of palaeo-taxodont and other bivalves, several gastropods, abundant trilobites (mainly *Metacanthina wallacei* and

Pilletina zguidensis), edrioasteroids (*Pyrgocystis flos*), some corals, hyolithids, spines of *Machaeracanthus*, rare phyllocarids (*Nahecaris jannae*), as well as machaerids (*Lepidocoleus rugatus*) can be found (KLUG et al. 2008; DE BAETS et al. 2010).

Fig. 5. *Deiroceras* limestone with two weathered specimens of *Deiroceras hollardi*.

The claystones are overlain by the "*Deiroceras* limestone" *sensu* KRÖGER (2008), which is about 1 m thick and contains abundant large *Deiroceras hollardi* with conch diameters of up to 10 cm (Fig. 5). In the eastern Tafilalt, this limestone unit has yielded conodonts of the Gronbergi Zone (BECKER & ABOUSSALAM 2011). This limestone is covered by 3 m of claystones with an increase in carbonate content towards the top. Limonitized elements of "faunule 2" described from correlative claystones overlying the *Deiroceras* Limestone in the eastern Tafilalt (KLUG et al. 2008) have only rarely been found in the Amessoui Syncline (DE BAETS et al. 2010). KRÖGER (2008) mentions abundant rugose corals and trilobites from this interval. The overlying *Erbenoceras* Limestone begins with three to four 20 cm-thick limestone layers rich in macrofossils; this wackestone contains abundant dacryoconarids, crinoids, corals, phacopid trilobites, ostracods, *Panenka* sp., brachiopods, and the ammonoids *Erbenoceras advolvens* as well as *Anetoceras* sp. (DE BAETS et al. 2013). The following layers are thinner and the carbonate content appears to decrease. About 1 m below the top of the limestone unit, a dark grey crinoid limestone can be seen; this is locally rich in Prasinophyceae and also yields phacopid trilobites. The topmost layers contain some *Mimagoniatites fecundus* (usually poorly preserved, hence the erroneous assignment to *Centroceras* in KRÖGER 2008), abundant large orthocones of the species *Deiroceras hollardi* and *Metarmenoceras fatimae*. The nothoceratid *Tafilaltoceras* sp. is also moderately common in these strata. In the eastern Tafilalt, dacryoconarids of the *Nowakia barrandei* and *Nowakia elegans* zones have been found in these beds and associated with *Mimagoniatites fecundus* (DE BAETS et al. 2010, BECKER & ABOUSSALAM 2011). Locally, articulated *Barrandeops* sp. are abundant. Some of the marly limestone layers are strongly burrowed.

Above the carbonates, a distinct facies change is visible, which can be correlated with the Daleje Event (MASSA et al. 1965, HOLLARD 1974, WALLISER 1984, 1985, KAUFMANN 1998, BELKA et al. 1999, FERROVÁ et al. 2011). According to JOHNSON et al. (1985, 1996), this coincides with a global transgression. These light coloured late Emsian limestone beds measure about 100 m in total thickness. Towards the top, the carbonate content is slowly increasing again.

In the slopes of the ridges (Fig. 4, right), which are capped by the Middle Devonian carbonates, this carbonate increase is visible in marly nodules near the base of the slope and the subsequently increasing abundance of marly nodule layers as well as macrofossils. Some of the lower layers contain many phacopids (HOLLARD 1960) and solitary rugose corals. These latest Emsian deposits yielded a diverse silicified ostracod fauna with the genera *Ctenoloculina*, *Ampuloides*, *Cryptophaga*, *Microcheilinella*, *Kirkbyella*, *Healdianella* and *Jenningsina* (kindly determined by H. BLUMENSTENGEL, Jena; see also FEIST & GROOS-UFFENORDE 1979). In the more limy nodular marls on top, cephalopods are very abundant, namely *Achguigites hidens, A. tafilaltensis*, *Sellanarcestes applanatus*, *S. ebbighauseni*, *S. draensis*, *S. tenuior*, *S. wenkenbachi*, *Anarcestes lateseptatus*, *An. latissimus*, *Mimagoniatites bohemicus*, *Arthrophyllum vermiculare*, and, a little higher in the section, the spiny nautilid *Hercoceras mirus* (see also, e.g., KLUG 2002, EBBIGHAUSEN et al. 2011, MONNET et al. 2011). The accompanying benthos is similarly rich; it contains various gastropods, bivalves, phacopids, harpetids, and other trilobites. The Emsian-Eifelian boundary is near the top of these yellowish, light coloured nodular marls.

5. EIFELIAN

Above these nodular marls, about 1 m of slightly marly shales follow, which are very poor in fossils. In the overlying 3 to 4 m, the carbonate content increases stepwise (Fig. 6). In the first metre, limestone nodule bands occur, which still have a rather light grey colour. In the subsequent metre of the section, the nodules are more or less fused, forming 5 to 10 cm thick limestone layers. These are nearly black, consist of dacryoconarid wacke- to packstones and contain abundant agoniatitids of the taxa *Fidelites fidelis*, *F. clariondi* and, at the top of this part, the index ammonoid *Pinacites jugleri*. The fauna comprises also orthoconic cephalopods, large bivalves, rare trilobites, brachiopods and gastropods. Although precise correlation is still needed, the Choteč Event-levels need to be sought near this part of section, close to the first occurrence of *P. jugleri* ("*jugleri* Event" *sensu* WALLISER 1985, REQUADT & WEDDIGE 1978, CHLUPÁČ 1985, HENN 1985, BERKYOVÁ & MUNNECKE 2010, KOPTÍKOVÁ 2011)

The next layer is conspicuous; it can be found in nearly the entire Tafilalt, sometimes as one massive limestone layer, which locally splits into two to three massive limestone beds (KLUG 2002a, b). In the area of the Tafilalt pelagic ridge, this layer shows several indications for condensed sedimentation such as iron bands, eroded fossils, micritised shells, shell accumulations, and questionable stromatolites (Fig. 7E). It contains many cephalopod species such as the ammonoids *Crispoceras tureki*, *Fidelites fidelis*, *Fidelites clariondi*, *Mimagoniatites bohemicus*, *Pinacites jugleri*, *Subanarcestes marhoumense*, the bactritid *Lobobactrites ellipticus*, orthocones such as *Spyroceras* sp., brevicones such as *Cerovoceras* and *Brevicoveras*, and cyrtocones (KLUG 2002a, 2002b, KRÖGER 2008). Various macrofossils of benthic organisms have also been found including bivalves, gastropods, trilobites, brachiopods etc. (Fig. 7D)

The overlying 5 to 10 cm thick bedded nodular limestone bed contains a quite similar fossil content, only *Subanarcestes macrocephalus* and *Wendtia/ Clarkeoceras ougarta* have their first appearances in these horizons. Above these 1.5 m, one or two massive limestone beds follow, which are also rich in cephalopod fossils including early *Cabrieroceras crispiforme*. These beds are overlain by 1 m of thin bedded nodular limestones with *Agoniatites vanuxemi*, *Diallagites* spp., and *Cabrieroceras housei*. The subsequent massive limestone bed and the first thin limestone layers above contain large *Agoniatites* (Fig. 7B) and *Cabrieroceras* (Fig. 7C), large rutoceratids, abundant brevicones such as *Cerovoceras* and *Brevicoceras*, and large cyrtocones. *Subanarcestes* is still abundant. The Eifelian part of the section ends with thin bedded nodular limestone and marl. These sediments are usually found near the top of the Middle Devonian ridge near the top of the North-East-facing slope. They yield fewer fossils and can be correlated with the Kačák transgression ("*otomari*" and "*rouvillei*" Events *sensu* WALLISER 1985, HOUSE 1985, WALLISER et al. 1995, SCHÖNE 1997), "Great Gap", "Late Eifelian Events 1 + 2", "Odershausen Events" (WEDDIGE 1988, 1996, WEDDIGE & STRUVE 1988).

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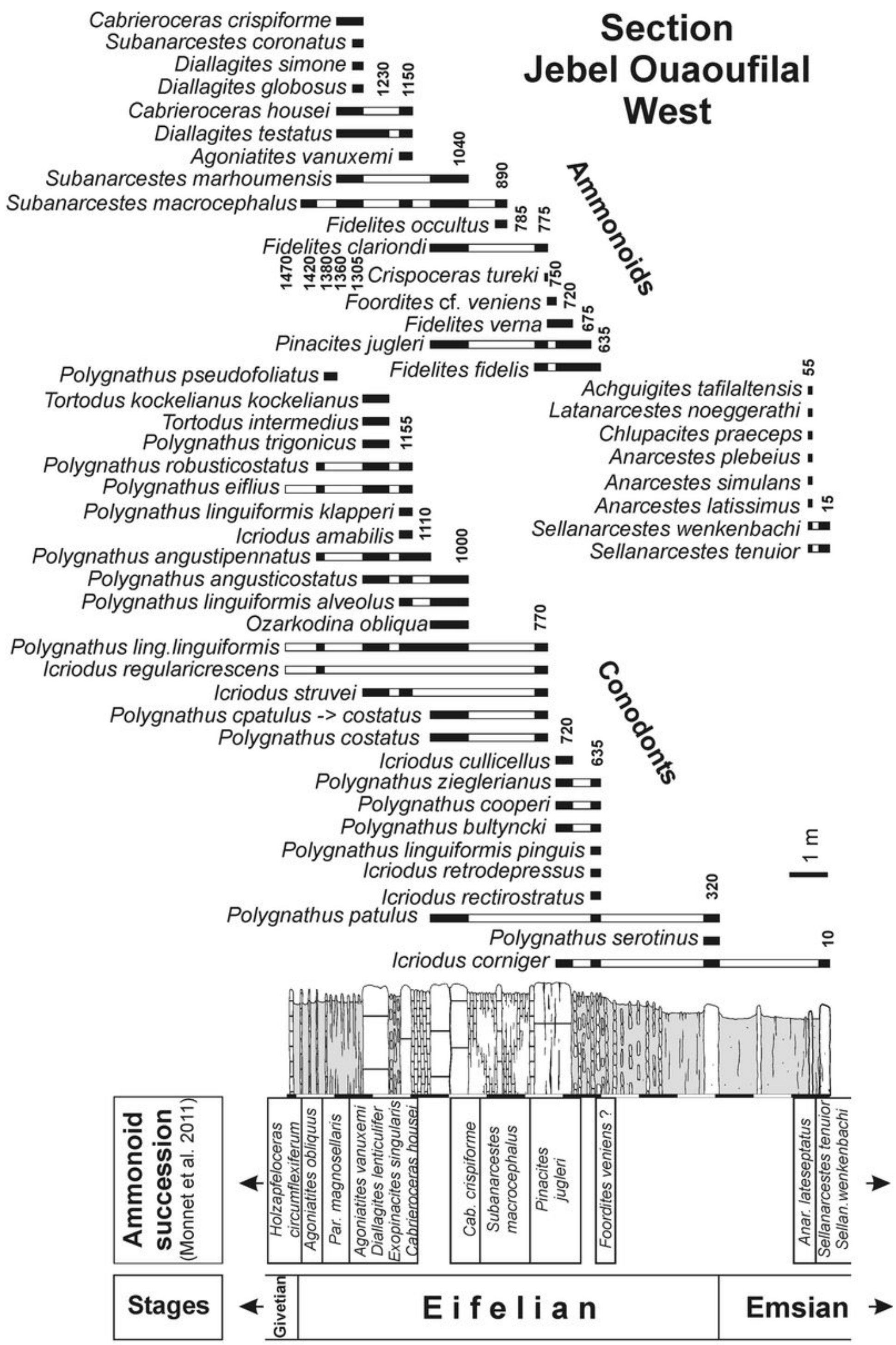

Fig. 6. The late Emsian to basal Givetian part of the section (from KLUG et al. 2008). Ammonoid stratigraphy according to MONNET et al. (2011).

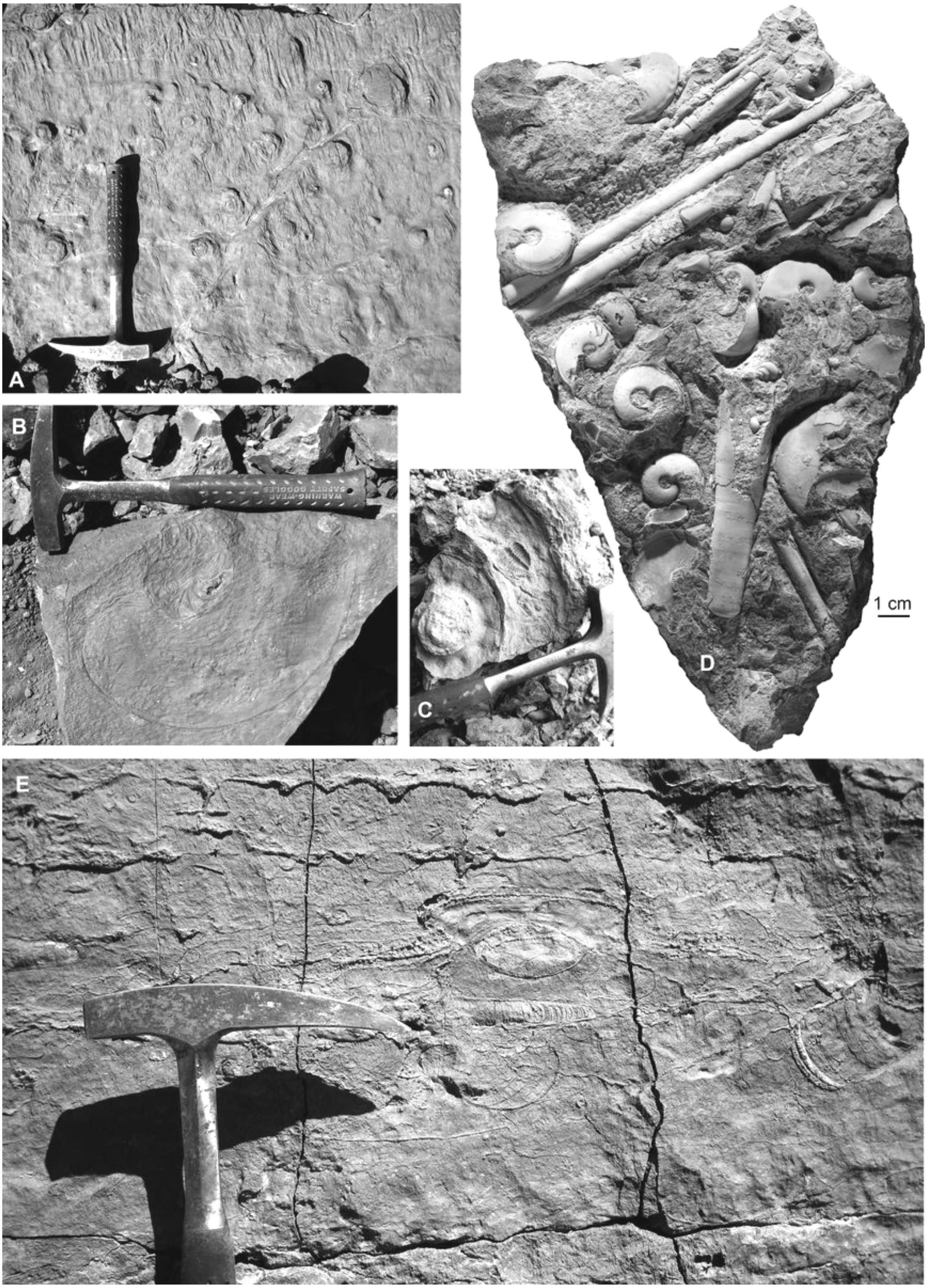

Fig. 7. Eifelian fossils of the Filon 12/ Jebel Ouaoufilal-region. **A.** late Eifelian mass occurrence of *Cabrieroceras* and *Subanarcestes*. **B.** 30 cm adult *Agoniatites* cf. *vanuxemi* from the late Eifelian *Agoniatites vanuxemi* Zone. **C.** *Cabrieroceras* cf. *housei*, probably also almost 30 cm in diameter (from KLUG 2002). **D.** Effects of condensation: bedding plane with *Lobobactrites ellipticus*, *Fidelites clariondi*, *Pinacites jugleri*, *Spyroceras* sp., a gastropod and a pygidium of a phacopid. **E.** Weathered surface of a section through the massive limestone bed of the *Pinacites jugleri* Zone in the early Eifelian, showing cross sections through an orthocone, a cyrtoconic nautiloid (right), a stromatolite-dome-liken structure (above the center), and two iron crusts (top).

THE GIVETIAN - FAMENNIAN AT OUM EL JERANE (AMESSOUI SYNCLINE, SOUTHERN TAFILALT)

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Fig. 1. Overview of the Givetian to Famennian succession at Oum el Jerane, with the Givetian slope in the mid-ground, a marker limestone and biostromes at the slope base, a wide plain caused by the deep weathering of upper Middle Givetian to Middle Frasnian marls, a ridge formed by lower Famennian limestones, ending to the left with the *Gonioclymenia* Trench and subsequent, greenish-yellow, fine siliciclastics of the lower Aoufital Formation. Background = Ordovician quartzites.

1. INTRODUCTION

The Amessoui Syncline in the southern Tafilalt provides superb and fossil-rich Devonian outcrops for ca. 35 km on strike, ca. in WNW-ESE direction. It has been celebrated for its intercalations of neritic to reefal and pelagic facies in the first surveys of MASSA (1965) and HOLLARD (1974). The gradual transition from deep (basin slope) to shallow (inner carbonate ramp) facies settings offers a huge potential for the reconstruction of Devonian palaeoecological gradients, which by far has not yet been exploited. Our investigations at Oum el Jerane (x = 620.25, y = 444.0; upper left corner on topographic map = fig. 1 in BECKER et al. this vol.), ca. 4 km W of the abandoned El Atrous settlement, have concentrated in the last decade on the Givetian to Famennian succession. It is of particular significance because it contains a good record of biostromes close to the Taghanic and Frasnes Events, a unique basal Upper Givetian goniatite fauna, and a fine exposure of the Dasberg Crisis Interval (Fig. 1). It is also intriguing that, unlike as at the close northern limb of the Amessoui Syncline (e.g., BECKER et al. 1989, WALLISER et al. 1989), there is no clear Kellwasser Facies around the Frasnian-Famennian boundary. Previous publications concerning Oum el Jerane are HOLLARD (1974: section El Atrous II), KORN et al. (2000: Dasberg Event), GINTER et al. (2002: shark teeth of the *Gonioclymenia* Limestone), ABOUSSALAM & BECKER (2007: Upper Givetian), KLUG et al. (2008: Lower Emsian), ABOUSSALAM & BECKER (2011: Taghanic Event), HARTENFELS & BECKER (2009, 2012) and HARTENFELS (2011: Dasberg Crisis and higher beds).

2. GIVETIAN

2.1. Middle Givetian

HOLLARD (1974) assigned a succession of ca. 20 m dark-bluish limestone at the northern slope of the main Middle Devonian ridge, his Units 4a-b, to the Givetian. Our detailed section (Fig. 2) begins at the top of this sequence with nodular limestone (Bed -7), followed by a massive, middle-grey, bioclastic marker limestone with rugose and tabulate corals (*Thamnopora*) and crinoid debris (Bed -6). Towards the east (especially at the Jebel Ouaoufilal) the same level forms the stratum for larger biostromes. Above, there is recessive nodular marl (Bed -5) and then a minor cliff of five, middle-grey limestones (Beds -4 to 0) with poor macrofauna and iron mineralisations at some bed tops, indicating minor sedimentary breaks. This small cliff at the foot of the slope correlates with the Middle *Sellagoniatites* Bed of lateral sections (e.g., El Atrous North, ABOUSSALAM 2003, ABOUSSALAM & BECKER 2011) and represents an episode of low sea-level throughout the Tafilalt. The index goniatite has not yet been found locally.

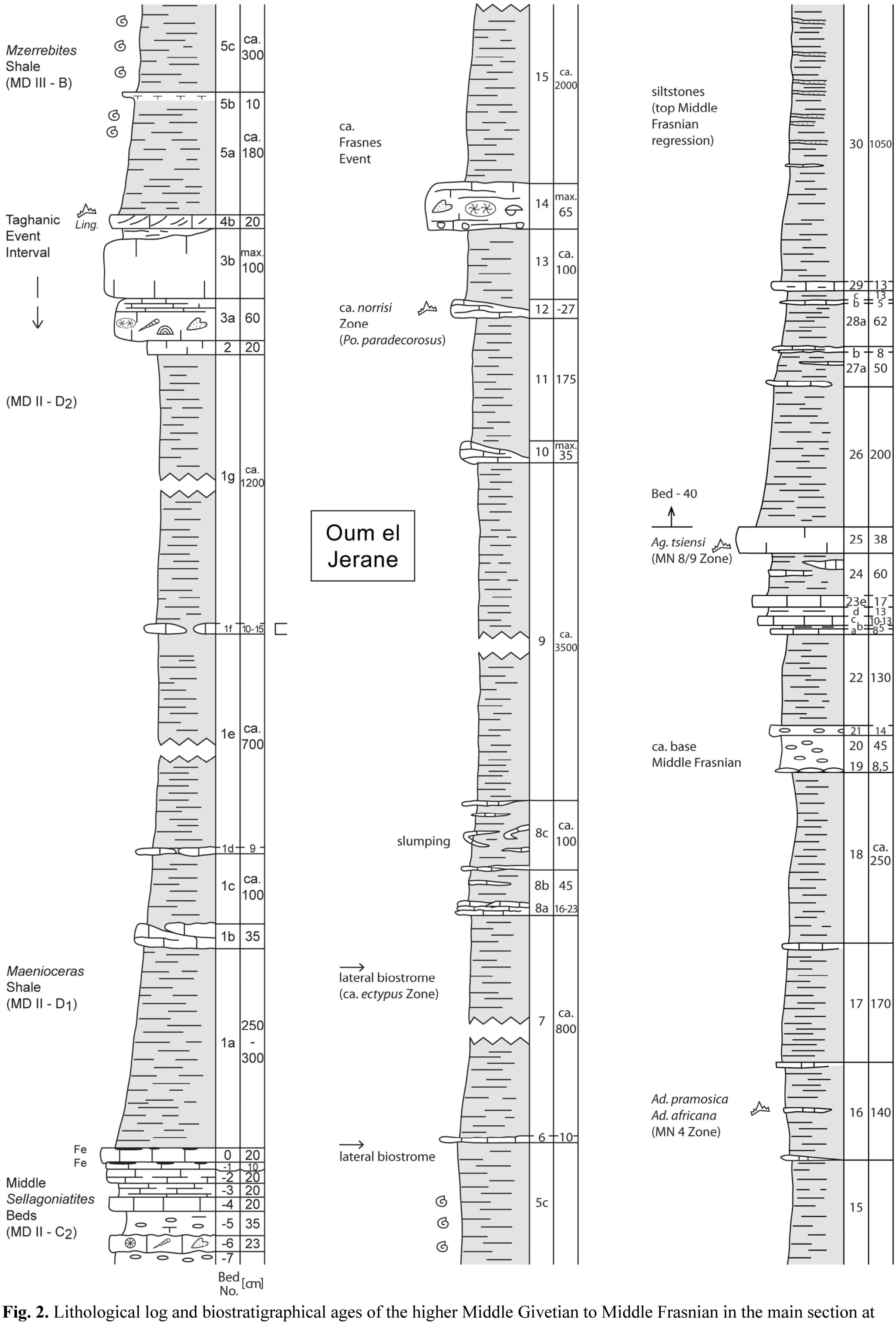

Fig. 2. Lithological log and biostratigraphical ages of the higher Middle Givetian to Middle Frasnian in the main section at Oum el Jerane.

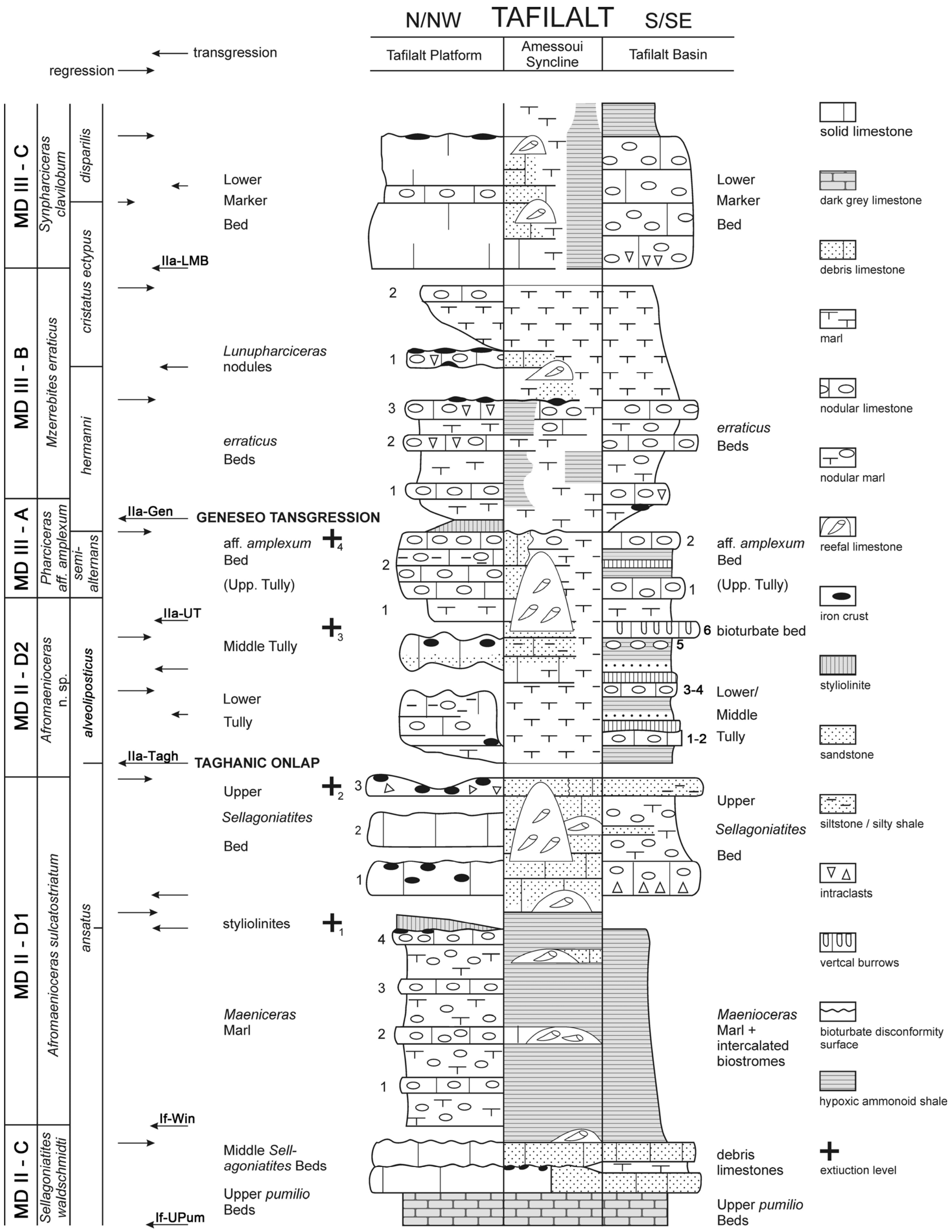

Fig. 3. Biostratigraphy, extinction levels (+$_{1-4}$), faunal blooms, episodes of biostrome growth and the revised depophase terminology for the central Tafilalt Platform, Amessoui Syncline, and Tafilalt Basin (ABOUSSALAM & BECKER 2011, fig. 6)

A subsequent wide plain consist of thick, deeply weathered, hypoxic shale/marl interrupted by two irregularly bedded lenses of fine crystalline, middle to dark grey limestone (Beds 1b and 1f). The base marks a transgressive surface (TST drowning unconformity), which correlates with the base of the *Maenioceras* Marls of the central Tafilalt and with Depophase If-Win sensu ABOUSSALAM & BECKER (2011; Fig. 3). Above (Bed 1g, Fig. 2), there are rare (altered) pyritic goniatites, including *Maenio. terebratum* n. ssp. (Pl. 1, Fig. 1), *Maenioceras* n. sp (Pl. 1, Fig. 2)., *Afromaenioceras* n. sp. (Pl. 1, Fig. 3), *Afro. sulcatostriatum*, *Sobolewia ?virginiana*, *Wedekindella brilonense, ?Sellagoniatites* sp., and tornoceratids (Gen. aff. *Phoenixites*), gastropods, and lycopsid plant remains (Pl. 1, Fig. 4).

Beds 2-3 compose a low, several decameter wide, patch-like biostrome with strong fragmentation, displacement and silification of reef builders and with lamination at the top of Bed 3a (Fig. 4). The bluish-grey solid limestone ranges from bindstone in the lower part to rud- and floatstone above. It yields large stromatoporoids, many Tabulata (mostly *Alveolites, Roseoporella,* and *Pachyfavosites*), solitary Rugosa, and *Phillipsastrea* (partly growing on large intraclasts). The reefal limestone (Fig. 5) wedges out along strike. Its small-scale intergrading with marls with pyritic fauna suggests that the adjacent basinal facies was not too deep and not too far laterally from the lower photic zone. The rud-floatstones may represent proximal tempestites or debris flows. A wide lateral transport is unlikely since many large corals are well-preserved. Following a thin marl interval that fills the irregular upper palaeotopography (Bed 4a), a cross-bedded massive, bluish limestone (Bed 4b) forms the top of a very low morphological elevation (“Taghanic Biostrome”).

Fig. 4. Block of thick coral limestone with upper lamination (Bed 3a), which suggests deposition by storm or debris flow.

Fig. 5. Block of coral rud-floatstone from the “Taghanic Biostrome” (Bed 3).

Age: The Middle *Sellagoniatites* Beds belong to the lower part of the *ansatus* Zone (Fig. 3). The goniatites from Bed 1g are indicative of the top Middle Givetian. Based on its range in the Dra Valley (BECKER et al. 2004), *Afromaenioceras* n. sp. suggests an inclusion of the lower/middle Taghanic Event Interval. Loose early *Pharciceras* (Pl. 1, Fig. 5) represent the Upper Taghanic Interval (ABOUSSALAM & BECKER 2011) or are possibly derived from higher beds. Bed 3 falls in any case in the Taghanic Crisis Interval. Bed 4b has only very few conodonts but *Linguipolygnathus linguiformis* proves that it still belongs to the Middle Givetian.

Fig. 6. Transition from the “Taghanic Biostrome” (Bed 3, left) to the deeply weathered *Mzerrebites* Shale (center).

2.2. Upper Givetian

A wide stretch of shale/marl just N of the “Taghanic Biostrome” (Fig. 5) is unique for all of the Tafilalt because of its distinctive assemblage of goniatites, with many new taxa, abundant gastropods, and other fossil groups:

Epitornoceras mithracoides
Nebechoceras eccentricum
Tornoceras typum
Tornoceratidae n.gen.I (Pl. 1, Fig. 13)
Tornoceratidae n. gen.II (*Cheiloceras* homoeomorph)
N. Gen. aff. *Phoenixites* div. sp. (Pl. 1, Fig. 14)
Pharciceras tridens Gp. (Pl. 1, Figs. 6-7, common)
Pharciceras aff. *amplexum*
Pharciceras div. n. sp. (e.g., Pl. 1, Fig. 8)
Extropharciceras div. n. sp. (e.g., Pl. 1, Fig. 9)
Pharciceratinae n. gen. (Pl. 1, Fig. 10, multilobate)
Mzerrebites juvenocostatus (Pl. 1, Fig. 11, very common and compressed)
Mzerrebites n. sp. (Pl. 1, Fig. 12, thicker)
orthocones and brevicones
Phillipsastrea sp. and *Thamnopora* sp.
bellerophontids and *Platystoma* sp. (very common)
nuculid bivalves
rhynchonellids, atrypids, and smooth brachiopods
crinoid stem pieces
phacopids
lycopsid remains

This rich goethitic fauna can only be collected in a small area NE of the biostrome (Fig. 6), which suggests a minor depression (puddle) with very local good conditions for bacterial sulfate reduction and early diagenetic pyritization. Laterally, just a few decameters to the east, the *Mzerrebites* Shale grades into a thinner alternation of dark, platy limestone and marl without macrofauna. It represents a sudden deepening and local eutrophication phase of the basal Upper Givetian Geneseo Transgression or Depophase

IIa-Gen sensu ABOUSSALAM & BECKER (2011; Fig. 3). It has the most diverse ammonoid assemblage of its age on a global scale and requires more detailed taxonomic treatment. Contemporaneous faunas of the Tafilalt Platform contain just a few species.

Bed 6 consists of lenses of dark, laminated distal tempestites/turbidites, variably 6-10 cm thick and alternating with shale/marl. Slightly to the east there is an up to 40 cm thick lens of biostromal limestone (coral-crinoid rudstone, BECKER & ABOUSSALAM 2011; Fig. 7) with abundant tabulate corals (alveolitids and thamnoporids), solitary Rugosa, and crinoid debris, but apparently without stromatoporoids. The differences between Middle and Upper Givetian coral assemblages of the Amessoui Syncline require more work, including taxonomic studies. A second, up to 25 cm thick biostrome layer with Tabulata and *Phillipsastrea* projects into a thick marl unit (Bed 7) of the measured succession (Fig. 2). It is followed by alternating bluish-grey, laminated limestone and shale/marl (Beds 8a-c) with large, partly isoclinal slump folds. These underline a slope setting with synsedimentary gliding of sediment packages and suggest that the laminated limestones are distal turbidites. In this context, the isolated, lenticular biostromal and intraclastic limestones probably represent debris flows from an adjacent, now eroded carbonate ramp, probably just to the south.

Fig. 7. Coral-crinoid rudstone, bed lateral to Bed 6 (east of main section).

The top Givetian is represented by a 70-80 m wide stretch of unfossiliferous shale/marl (Bed 9). Two thick-bedded but laterally thinning distal turbidites (Beds 10 and 12) are each overlain by unfossiliferous shale (Beds 11 and 13). The youngest known biostromal limestone of the Tafilalt (Bed 14, Fig. 8) is a crinoidal coral debris flow with lamination both at the base and at the top. There are many phillipastreids, thamnoporids, alveolitids, and large atrypids. The latter are not known from the lower coral limestones and document a palaeoecological change to brachiopod-richer facies in the Upper Devonian.

Age: The goniatite assemblage of Beds 5a-c is typical for the *Mzerrebites* Zone at the base of the Upper Givetian (MD III-B, Fig. 3). Advanced pharciceratids (Synpharciceratinae) and oxyconic Taouzitidae (*Darkaoceras*) are still lacking. The second lateral coral limestone yielded *Icriodus difficilis, Tortodus* sp., and *Polygnathus limitaris*, a species that ranges regionally throughout the Upper Givetian (ABOUSSALAM & BECKER 2007). Bed 12 produced a poor conodont fauna with *Belodella resima, I. difficilis,* and *Po. paradecorosus*, including its *Polygnathus*-type ramiform elements The last species enters in the topmost Givetian *norrisi* Zone. Therefore, it is likely that the last coral limestone shortly above marks the regressive terminal Givetian.

Fig. 8. Crinoidal coral limestone (Bed 14), probably a debris flow, top of the Givetian (thin bed at right margin = Bed 12).

3. FRASNIAN

3.1. Lower Frasnian

The Lower Frasnian is mostly represented by a ca. 20 m thick shale/marl (Bed 15) without macrofauna. A thin, bluish-grey, distal turbidite within the following Bed 16 yielded an unexpectedly rich and diverse conodont assemblage with many ancyrodellids (*Ad. africana*, Pl. 4, Fig. 1*, Ad. pramosica,* Pl. 4, Fig. 2, *Ad. recta,* Pl. 4, Fig. 3, *Ad. rugosa,* Pl. 4, Fig. 5, *Ad. alata*, Pl. 4, Fig. 4), icriodids (*I. symmetricus, I. subterminus,* Pl. 4, Fig. 11, *I. tafilensis,* Pl. 4, Fig. 7, *I. expansus*), *Mesotaxis asymmetrica, Mes. guanwushanensis* (Pl. 4, Fig. 10, = *falsiovalis*), *Zieglerina ovalis* (Pl. 4, Fig. 6), *Po. paradecorosus, Po. webbi* (Pl. 4, Fig. 8), *Po. ordinatus* (Pl. 4, Fig. 12), *Po. alatu*s, *Po. pennatus* (Pl. 4, Fig. 9), *Ctenopolygnathus angustidiscus*, and *Tortodus* sp. Beds 17-18 continue the basinal facies with more than 4 m of shale/marl, interrupted by a thin turbidite (Bed 17b). The Frasnes and Timan Events of the central Tafilalt Platform (ABOUSSALAM & BECKER 2007), with their characteristic black styliolinites, cannot be recognized in the Amessoui Syncline.

Age: *Ag. pramosica* and *Ad. africana* date the conodont fauna of Bed 16 as top Lower Frasnian *transitans* Zone (MN 4 Zone). However, there is a facies-controlled absence of *Palmatolepis*, which indicates a neritic source area. Some faunal elements (e.g., *Ad. recta,* a single *Klapperina*) appear to have been reworked on the eroded platform from the top Givetian to Lower Frasnian (MN 2/3 Zones). It is intriguing that the Jebel el Mrier south of the

Amessoui Syncline complete lacks the whole Upper Devonian, due to non-deposition and/or erosion.

3.2. Middle Frasnian

The base of the Middle Frasnian is tentatively placed at the base of Bed 19, a grey nodular limestone, followed by nodular marl (Bed 20) and a slightly nodular, more solid limestone (Bed 21). They document an episodic transition from a basinal slope setting to the lower part of a hemipelagic carbonate ramp. Above a ca. 1.3 m thick shale/marl (Bed 22) and thin limestone (Beds 23a), Bed 23c is a bluish-grey, solid bioclastic limestone with crinoid debris. Bed 23e is lighter grey than the beds below. Separated by 60 cm marl and thin limestone (Bed 24), Bed 25 is a massive local marker bed (light-grey bioclastic limestone). It records a relative sea-level fall. The conodont biofacies supports a neritic setting; the fauna consists of the early morphotype of *Ad. curvata, Ancyrognathus tsiensi*, *I. symmetricus, Bel. resima*, *Cteno. angustidiscus*, *Po. paradecorosus,* and a polygnathid close to *Polygnathus* n. sp. G sensu KLAPPER & LANE (1985). *Palmatolepis* is missing. The marker bed was used to continue the section logging laterally on the next hill to the west (Fig. 9). The base of the lateral Bed -40 correlates with the base of Bed 26. The subsequent succession (Fig. 10) allows us to align the two sections with the higher Famennian described by HARTENFELS (2011; Fig. 11).

Following the interval with some distal turbidites of Beds 26-29 (within -40), the major change in the higher part of the Middle Frasnian is the increasing, fine siliciclastic influx. Thin, laminated calcareous siltstones are probably also distal turbidites that reflect a regressive trend in the source area, with a change from carbonate ramp to near-shore clastics. Bed -39 is the first calcisiltite within the almost 38 m thick silt-marl package of Beds -40 to -32. It is a bioturbated, very unfossiliferous grainstone (Pl. 2, Fig. 1). A similar bed (Bed -37) follows ca. 16 m higher, whilst Bed -33 is an alternation of calcareous siltstones and thin calcisiltites yet ca. 8 m higher.

Age: *Ag. tsiensi* from Bed 25 is an alternative marker for the regional *coeni* Zone (ca. MN 8/9 Zone) in the middle part of the Middle Frasnian. *Polygnathus* n. sp. G is so far known from the MN 9 Zone of western Canada (KLAPPER 1997). Therefore, the lower part of the Middle Frasnian (MN 5-7 Zones) appears to be relatively thin, especially in comparison with the thick Upper Givetian. Bed -39 probably still belongs to the Middle Frasnian but it is poor in diagnostic taxa. *Pa. ljaschenkoae* (pl. 4, Fig. 15) ranges from the upper MN 8 into the lower part of the MN 11 Zone (KLAPPER 1997).

3.3. Upper Frasnian

The package of eight thin calcisiltite layers forming Bed -31 is arbitrarily taken as the base of the Upper Frasnian. The conodont fauna from the top (Bed -31h) includes the late morphotype of *Ad. curvata* (Pl. 4, Fig. 16), *Ag. tsiensi* (Pl. 4, Fig. 17), *Ag. amplicavus* (Pl. 4, Fig. 13), *Pa. hassi* (Pl. 4, Fig. 18), juvenile palmatolepids resembling *Pa. hani, I. symmetricus, I. alternatus alternatus* (Pl. 4, Fig. 19), *I. vitabilis* (Pl. 4, Fig. 14), and various polygnathids. The still poor representation of palmatolepids indicates that conodonts continued to be transported downslope from a deeper neritic setting. The next limestone interval (Bed -29a-f) consists of middle-grey crinoidal brachiopod limestone of an even shallower source. The Frasnian/Famennian boundary must follow somewhat higher, within the thick and rather homogeneous alternation of marls and thin calcisiltites of Beds -28 to ca. -23. A clarification requires further sampling. In any case, locally no Kellwasser Beds are developed.

Age: The basal Upper Frasnian age of Bed -31h is based on *I. alternatus alternatus*. The associated *Ag. amplicavus* is normally a late Middle Frasnian species (MN 8/9 Zones) but most likely the ancestor of *Ag. triangularis*, which enters slightly above the base of the MN 11 Zone (KLAPPER 1997). Therefore, Bed -31 is placed close to the Middle/Upper Frasnian boundary. Bed -29 may correlate with the brachiopod limestone noted by SCHINDLER (1990) high in the Upper Frasnian at his section El Atrous II, also at the southern limb of the Amessoui Syncline.

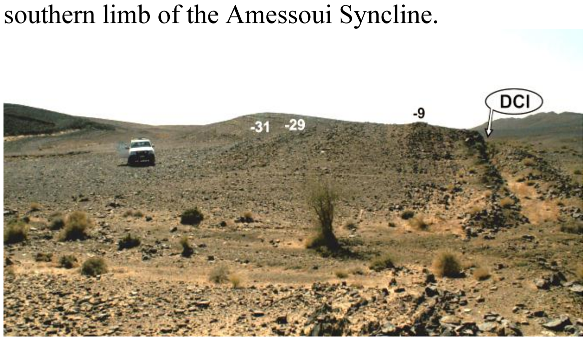

Fig. 9. Overview of the lateral (western) section with the position of marker beds and of the Dasberg Crisis Interval at the steep margin to the *Gonioclymenia* Trench.

4. FAMENNIAN

4.1. Lower Famennian

The not yet demarcated lower Famennian continues the marl-calcisiltite alternation of the Upper Frasnian. A conodont sample from Bed -19, a distal turbidite with convolute bedding (Pl. 2, Fig. 3), yielded a deep-water, palmatolepid assemblage with *I. alternatus* Gp. *Pa. minuta minuta, Pa. tenuipunctata, Pa. perlobata perlobata, Pa. quadrantinodosalobata, Pa. sandbergi, Pa. termini*, and *Neopolygnathus communis*. The absence of other polygnathids is remarkable. The *Palmatolepis*-dominated biofacies suggests that the Frasnian neritic source area has been drowned in the course of the global lower Famennian eustatic rise. There is no change of lithofacies in the subsequent ca. 3.5 m (Beds -18 to -10, Fig. 10).

Starting with Bed -9 middle-grey crinoid-brachiopod limestones appear. This local shallowing trend culminates higher up (Bed -3) first in bioturbated, silty pack-grainstone with some goniatites, brachiopods,

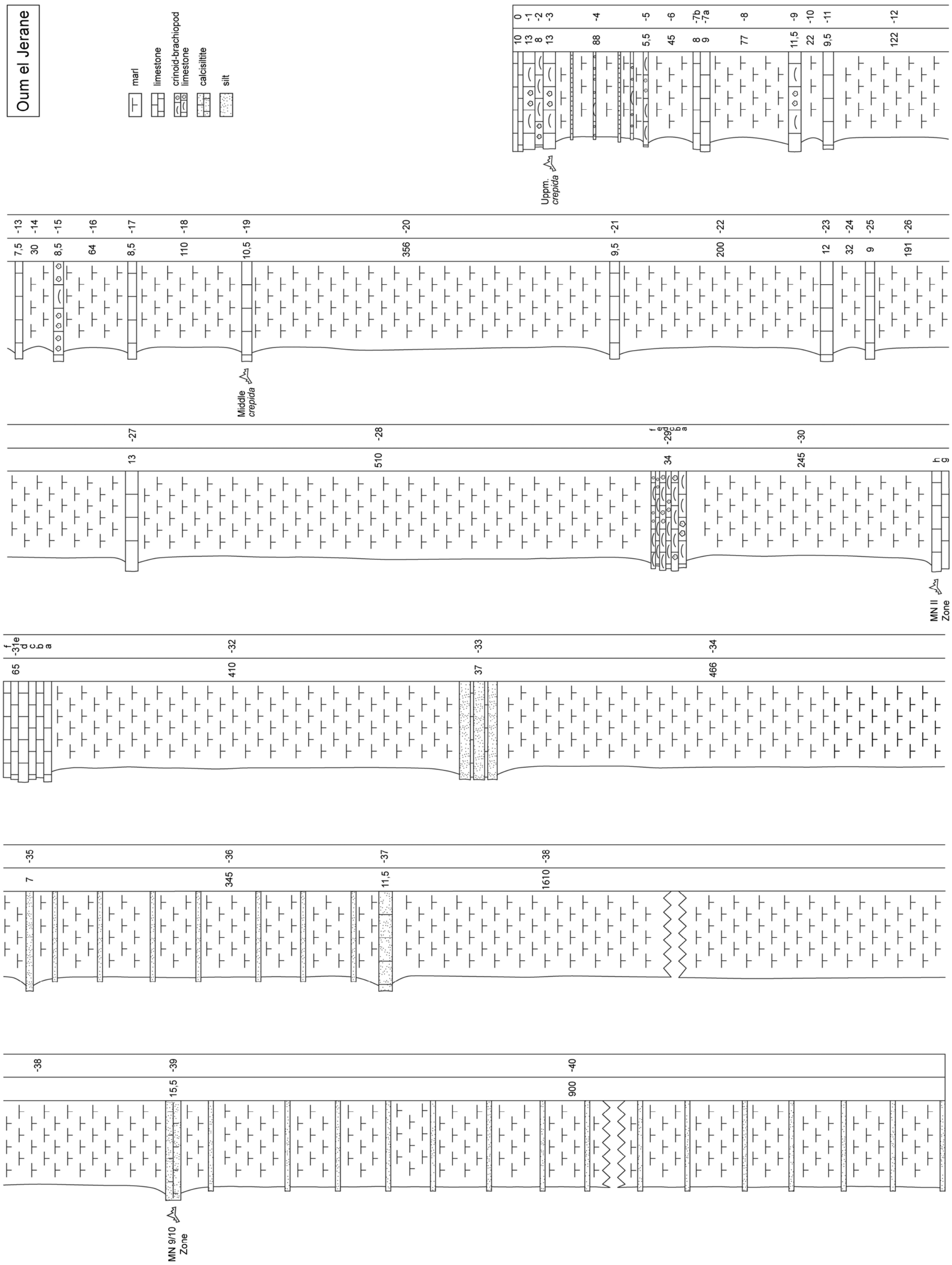

Fig. 10. Lithological log and the position of conodonts samples in the higher Frasnian to lower Famennian of the measured lateral section (below the Famennian section of HARTENFELS 2011).

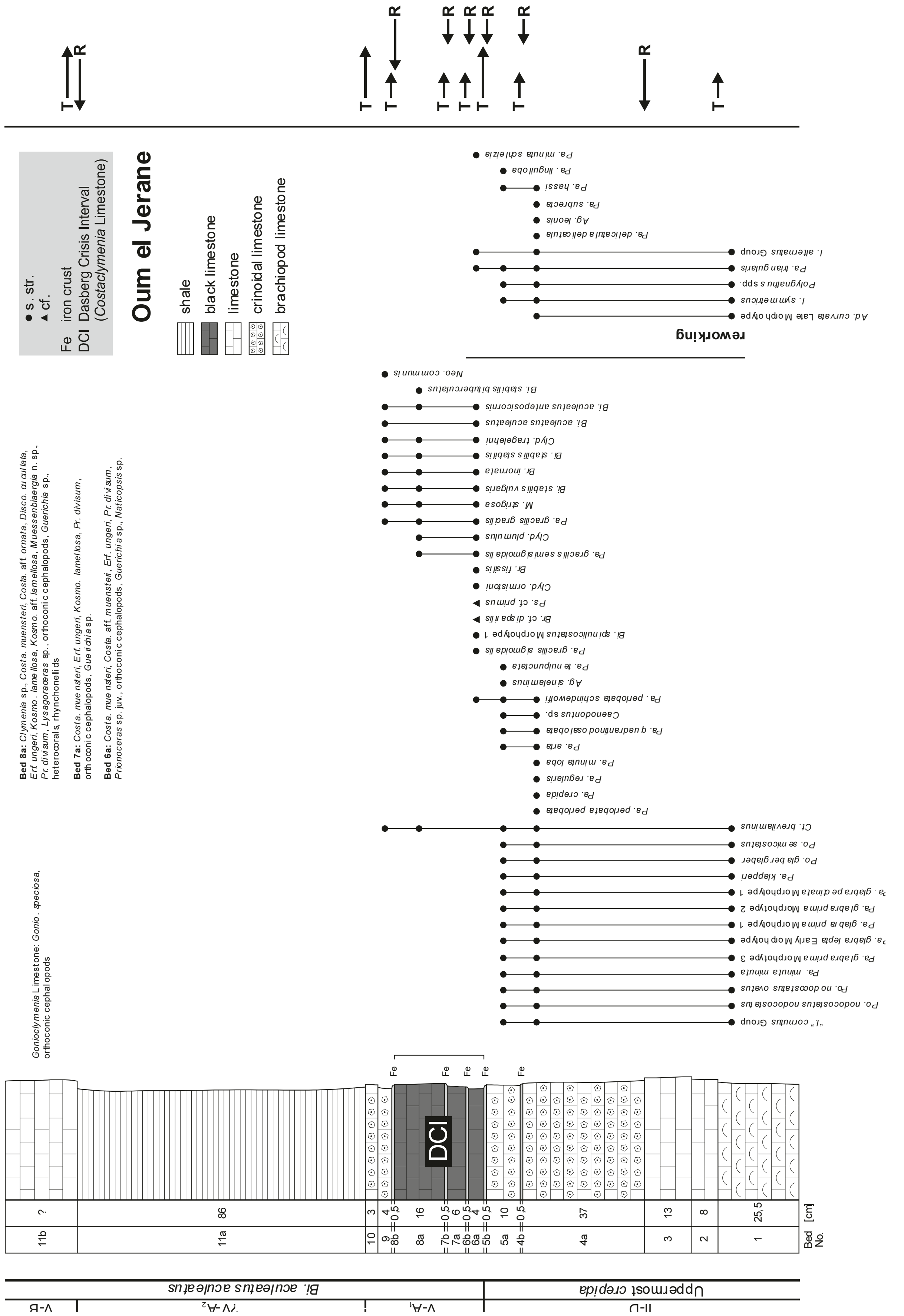

Fig. 11. Lithological and faunal succession, relative sea-level changes, and conodont zonation around the Dasberg Crisis at Oum el Jerane (from HARTENFELS 2011).

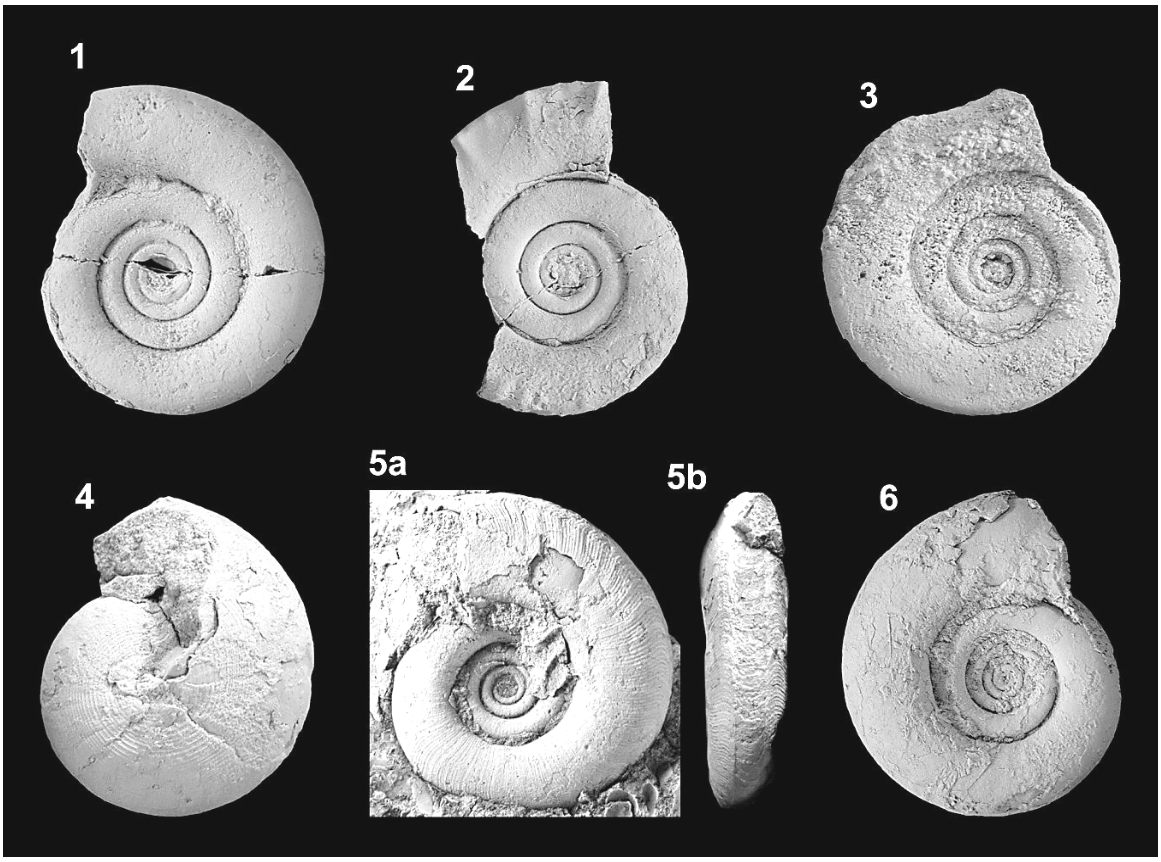

Fig. 12. Ammonoids from Bed 8a, upper *Costaclymenia* Limestone, *muensteri* Zone (UD V-A1, *Bi. aculeatus aculeatus* Zone), **1.** *Clymenia* sp., 32 mm diameter, **2.** *Costa. ornata*, 38 mm dm, **3.** *Costa. muensteri*, 22 mm dm, **4.** *Erf. ungeri*, 17,5 mm dm, **5a-b.** *Kosmo. lamellosa*, 23,6 mm dm, **6.** *Kosmo*. aff. *lamellosa*, 22 mm dm.

abundant crinoid remains and extraclasts (Pl. 2, Fig. 4). The diverse conodont fauna (more than twenty taxa) contains palmatolepids (*Pa. glabra prima* M1 and M3, *Pa. glabra lepta* Early Morphotype, *Pa. glabra pectinata* M1, *Pa. minuta minuta, Pa. minuta loba, Pa. crepida, Pa. perlobata perlobata*, *Pa. quadrantinodosalobata*), icriodids (*I. alternatus* Gp., first North African record of *I. mawsonae*), and polygnathids (most morphotrends of *Po. semicostatus*, two subspecies of *Po. nodocostatus*, and others). The assemblage is completed by reworked *“Ag.” cryptus* and Frasnian conodonts, such as *Ad. curvata* Late Morphotype and ancyrognathids. Specimens that are indistinguishable from *Ling. linguiformis* Morphotype γ1b (sensu WALLISER & BULTYNCK 2011) either reflect the reworking of Lower/Middle Givetian carbonates or are perfect homoeomorphs within the *Po. semicostatus* Gp. (see HARTENFELS 2011). The absence of reworked Lower Frasnian and Upper Givetian conodonts favors the second interpretation. The recycling of Upper Frasnian strata supports the regression in conjunction with an adjacent synsedimentary uplift.

The shallowest Famennian setting (neritic middle ramp) is reached with the coarse crinoidal and brachiopod limestones of Beds 0 to 5 (Fig. 11, GPS N 30° 59´39.3´´ W 4°8´18.7´). They were first described by WENDT et al. (1984). The prevailing brachiopod grain-rudstones may include *Stromatactis*-fabrics (Bed 2, Pl. 2, Fig. 5), which were formed by microbial mats during slow sedimentation and rapid cementation. There are abundant, complete rhynchonellids (Pl. 2, Fig. 6) and coated extraclasts (Pl. 3, Fig. 1), which were partly exhumed from the Frasnian. This led to frequent Upper Frasnian to lower Famennian reworked conodonts in samples from Beds 1, 4a, and 5a (see HARTENFELS 2011; Fig. 11 and Pl. 5, Figs. 1-2, 8-11). The mixed *Palmatolepis-Polygnathus* biofacies, with 7-18 % *Po. semicostatus* and up to 8 % icriodids, is typical for a deeper, storm-influenced neritic setting. It contrasts strongly with the black cheiloceratid coquinas and pelagic carbonate platform of the northern Amessoui Syncline, just a few km to the north (e.g., BECKER et al. 1989, BECKER 1993).

Age: *Pa. termini* from Bed -19 defines the Middle *crepida* Zone, ca. in the middle of the lower Famennian. *Pa. glabra pectinata* M1 and *Pa. perlobata perlobata* from Bed -3 are index species of the Uppermost *crepida* Zone (better *pectinata* Zone). *“Ag.” cryptus* is normally restricted to the *triangularis* Zones and, therefore, indicates some reworking of basal Famennian strata. Beds 1-5a still fall in the Uppermost *crepida* Zone, which is based on the upper ranges of “*Ag.” sinelaminus* (Pl. 5, Fig. 3), *Pa. tenuipunctata* (Pl. 5, Fig. 12), *Pa. angusta*, and *Pa. perlobata perlobata*. *Pa. klapperi* occurs in all three samples (see Pl. 5, Fig. 19).

4.2. Middle Famennian

The Middle Famennian (upper UD II/III) and lower part (UD IV) of the upper Famennian are completely missing in an unconformity at the top of Bed 5b. An iron crust reflects extreme condensation during prolonged non-deposition, which lasted for at least eight conodont zones (of the standard succession).

4.3. Upper Famennian

Three coarsely detrital, black cephalopod coquinas, the *Costaclymenia* (previously *Endosiphonites*) Limestone (Beds 6a, 7a, 8a, cephalopod rudstone, Pl. 3, Figs. 2-4) form the hypoxic Dasberg Crisis Interval. The transgressive base is locally the main Dasberg Event. A re-naming is inevitable since BARTZSCH & WEYER (2013) just showed that *Endosiphonites* was introduced as an invalid (junior) name replacement for *Clymenia* and that it is not an older synonym of *Costaclymenia*, which has to be re-instated. The dark cephalopod limestones (Fig. 13) are separated by thin, reddish iron crusts, which mark short sedimentary breaks. The rather uniform and very abundant ammonoid fauna consists of often complete specimens of *Costa. muensteri* (Fig. 12.3), *Costa. ornata* (Fig. 12.2), *Clymenia* sp. (Fig. 12.1), *Kosmoclymenia lamellosa* (Fig. 12.5a-b), *Kosmo*. aff. *lamellosa* (Fig. 12.6), *Muessenbiaergia* n. sp., *Prionoceras divisum, Discoclymenia cucullata*, and *Erfoudites ungeri* (Fig. 12.4, showing the characteristic spiral ornament). The youngest known *Prionoceras* are rare short-term survivors from the otherwise extinct *Platyclymenia* Faunas (of UD IV). Associated are orthocones, brevicones, bivalves of the pelagic realm (*Guerichia*), heterocorals, and rhynchonellids. The dominance of single-rowed bispathodids is in accordance with a hemipelagic setting. But, as below, there are reworked Upper Frasnian and lower Famennian conodonts and the coarse matrix and abundant crushed shells testify a rather high-energy deposition.

Fig. 13. Dark-grey upper *Costaclymenia* Limestone (Bed 8a) with orthocone, fragmentary *Costa. ornata* (upper part), *Erfoudites* (left middle), and *Kosmoclymenia lamellosa*.

The overlying Beds 9-10 are reddish-brown, iron-rich crinoid pack-rudstones without ammonoids (Pl. 3, Fig. 5). They mark a post-crisis regression and have a poorer conodont fauna than the *Costaclymenia* Limestones. There is a facies-controlled entry of *Neo. communis*. Bed 11a is a greenish, unfossiliferous shale. The originally massive *Gonioclymenia* Limestone has been completely exploited in km long, deep trenches, without any remaining natural outcrop. Fragments of the large *Gonio. speciosa* can still be found occasionally; better collecting is in the Erfoud rockshops. The microfacies of an isolated specimen from the trench is a crinoidal cephalopod rudstone with micrite matrix (Pl. 3, Fig. 6). The transgressive nature of the *Gonioclymenia* Limestone is evident from widespread unconformities below along the Amessoui Syncline (KORN et al 2000, HARTENFELS & BECKER 2009, HARTENFELS 2011). The top Dasbergian (UD V-C, *Medioclymenia* level) and the lower part of the uppermost Famennian (UD VI-A to D) are practically missing in a period of non-deposition and extreme condensation.

The plain and slope to the north is formed by the thick, greenish-grey, unfossiliferous siltstones of the lower Aoufital Formation, which accumulated during the early phase of the Hangenberg Regression (e.g., KAISER et al. 2012). Equivalents of the Hangenberg Black Shale are not visible. Sandstones of the regressive peak form the cliff top.

Age: The ammonoid association of Beds 6-8 is characteristic for the *muensteri* Zone of the Dasberg Crisis Interval (UD V-A1). Besides various costaclymeniids, the entries of *Discoclymenia* and of the oldest kosmoclymeniids are equally characteristic. The associated conodonts, notably *Bispathodus aculeatus aculeatus* (Pl. 5, Fig. 4)*, Bi. aculeatus anteposicornis* (Pl. 5, Fig. 5), *Bi. spinulicostatus* M1, and *Clydagnathus plumulus* (Pl. 5, Fig. 6), are index species of the *aculaeatus aculeatus* (Sub)Zone (= basal part of Middle *expansa* Zone). Beds 9-10 fall in the same zone. Both limestones and Bed 11a may correlate with the *Gonio. subcarinata* Zone (UD V-A2) of the central Tafilalt (BECKER et al. 2002). The transgressive *Gonioclymenia* Limestone (see WENDT et al. 1984) is characterized by the giant index species of UD V-B. New data (HARTENFELS & BECKER 2012) show that its conodonts fall in the *costatus* Subzone of the *aculeatus* Zone, not in the Upper *expansa* Zone, as suggested in GINTER et al. (2002). We have no evidence for the youngest Dasbergian (UD V-C) and Wocklumian (UD VI-A/D). However, the record of *Bi. ultimus* in GINTER et al. (2002) shows that some thin uppermost Famennian limestone may be developed very locally.

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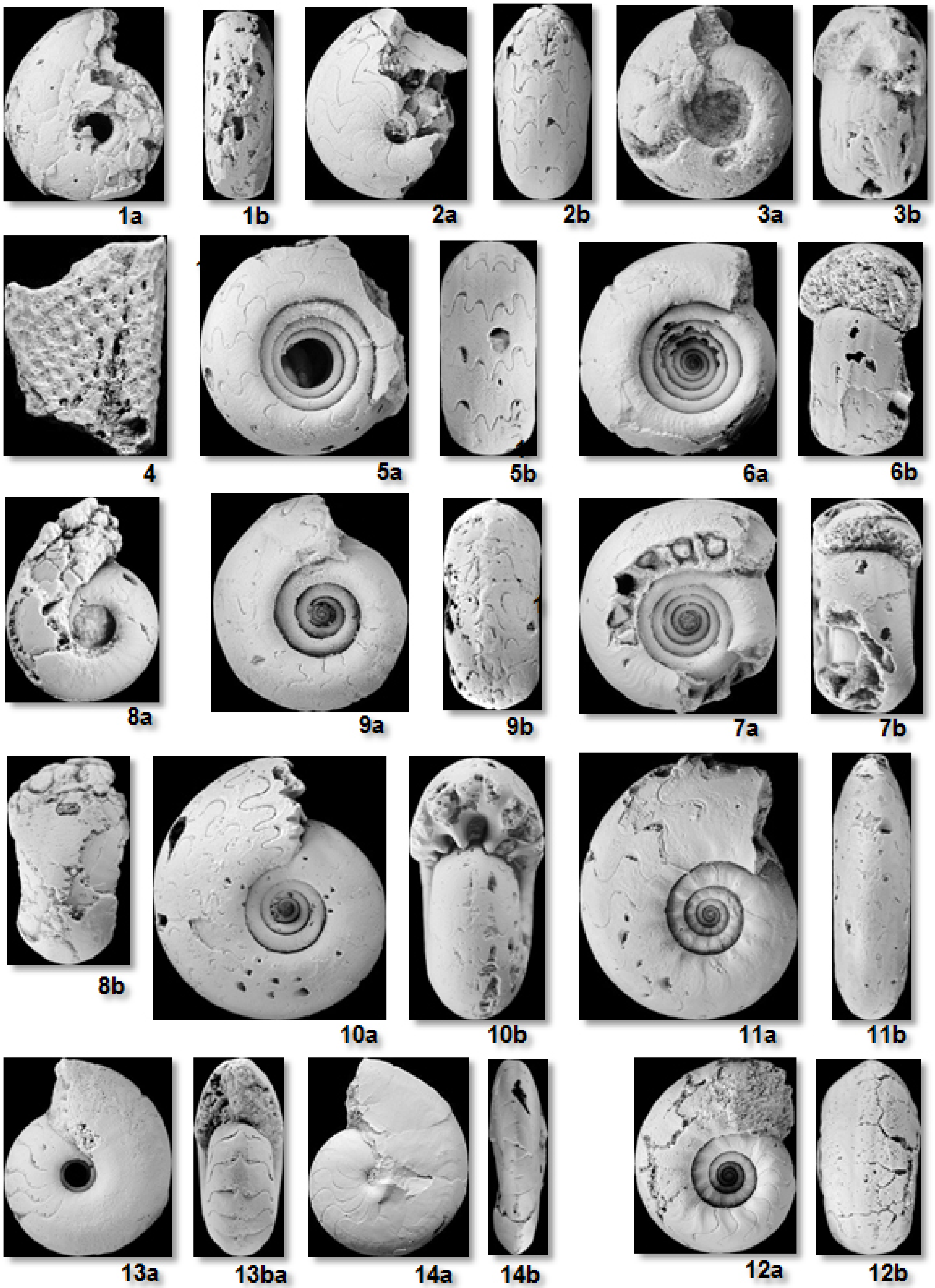
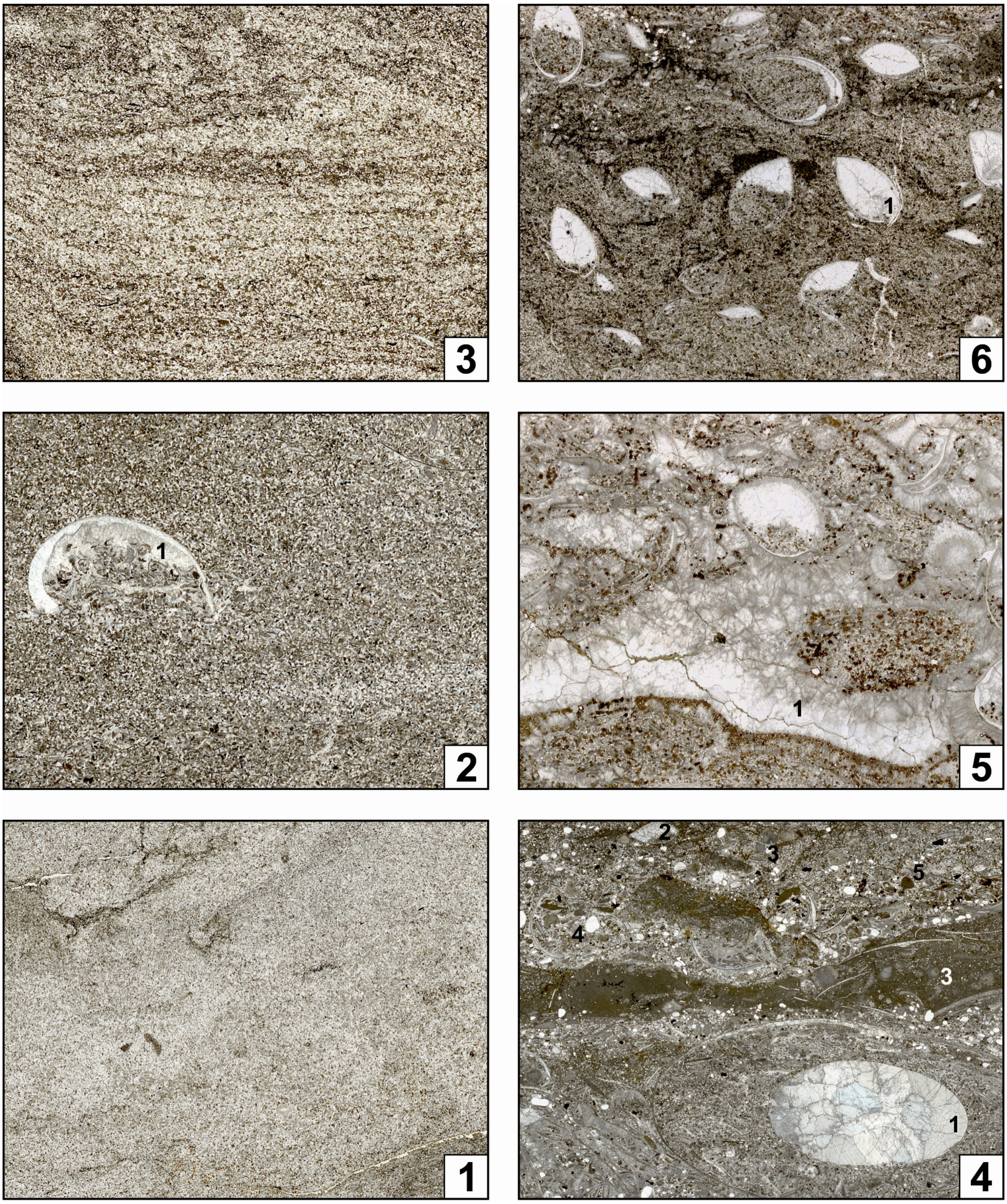
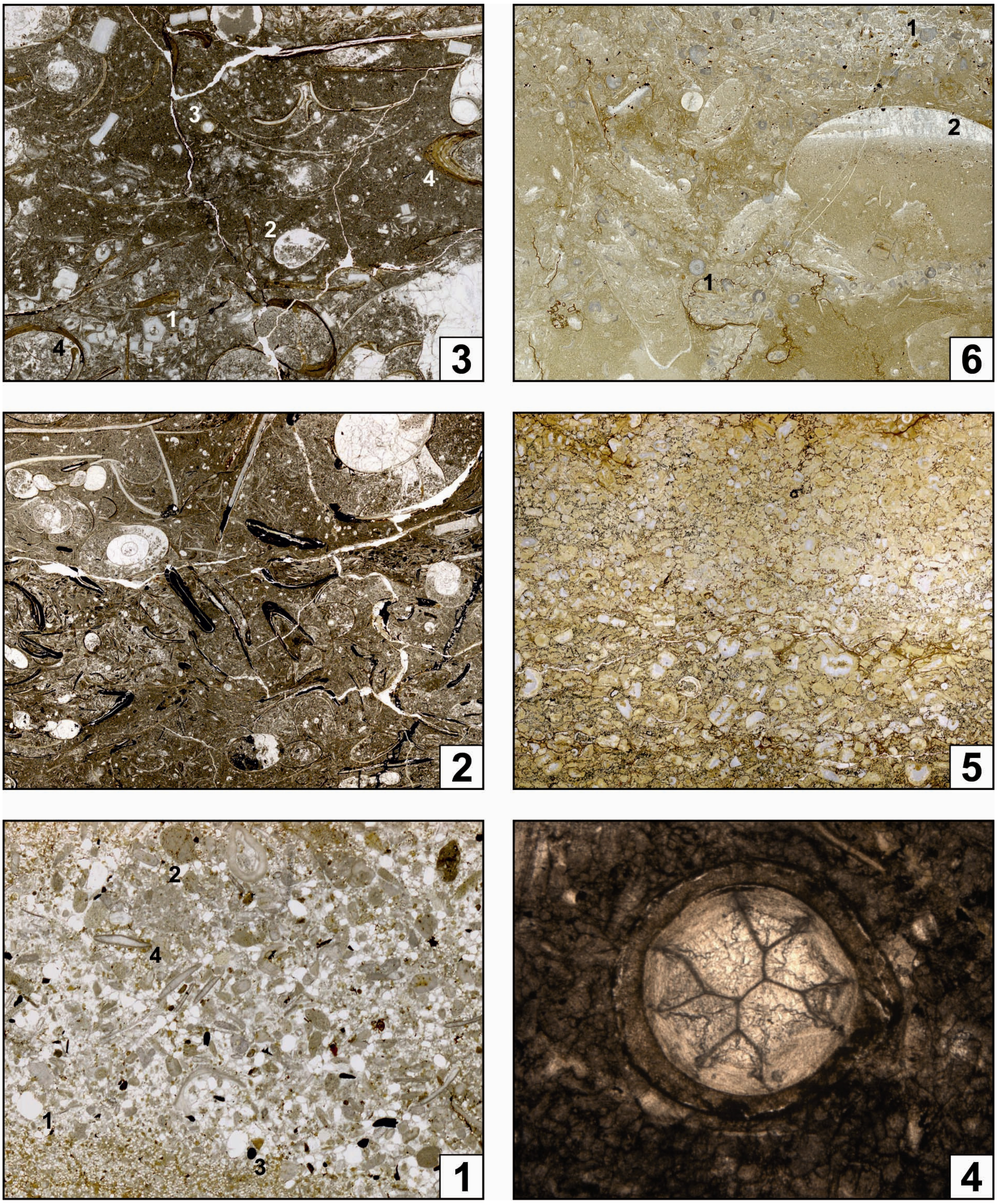


Pl. 1. Fossils from the *Maenioceras* Shale (1-5, loose collection) and *Mzerrebites* Shale (6-15) at Oum el Jerane. **1.** *Maenio. terebratum* n. ssp., 12.1 mm dm, **2.** *Maenioceras* n. sp., ca. 13.5 mm dm, **3.** *Afromaenioceras* n. sp., 10.7 mm dm, **4.** Lycopsid plant, length 26 mm, **5.** *Ph*. aff. *amplexum*, 17.7 mm dm, **6-7.** *Ph. tridens* Gp., 15 and 12 mm dm **8.** “*Pharciceras*” n. sp., 11.9 mm dm, **9.** *Extropharciceras* n. sp., 12.8 mm dm, **10.** Pharciceratinae n. gen. n. sp., 17.8 mm dm, **11.** *Mz. juvenocostatus*, 17 mm dm, **12.** *Mzerrebites* n. sp., 10.4 mm dm, **13.** Tornoceratidae n. gen. I n. sp., 11.7 mm dm, **14.** N. Gen. aff. *Phoenixites* n. sp. 1, 15 mm dm.

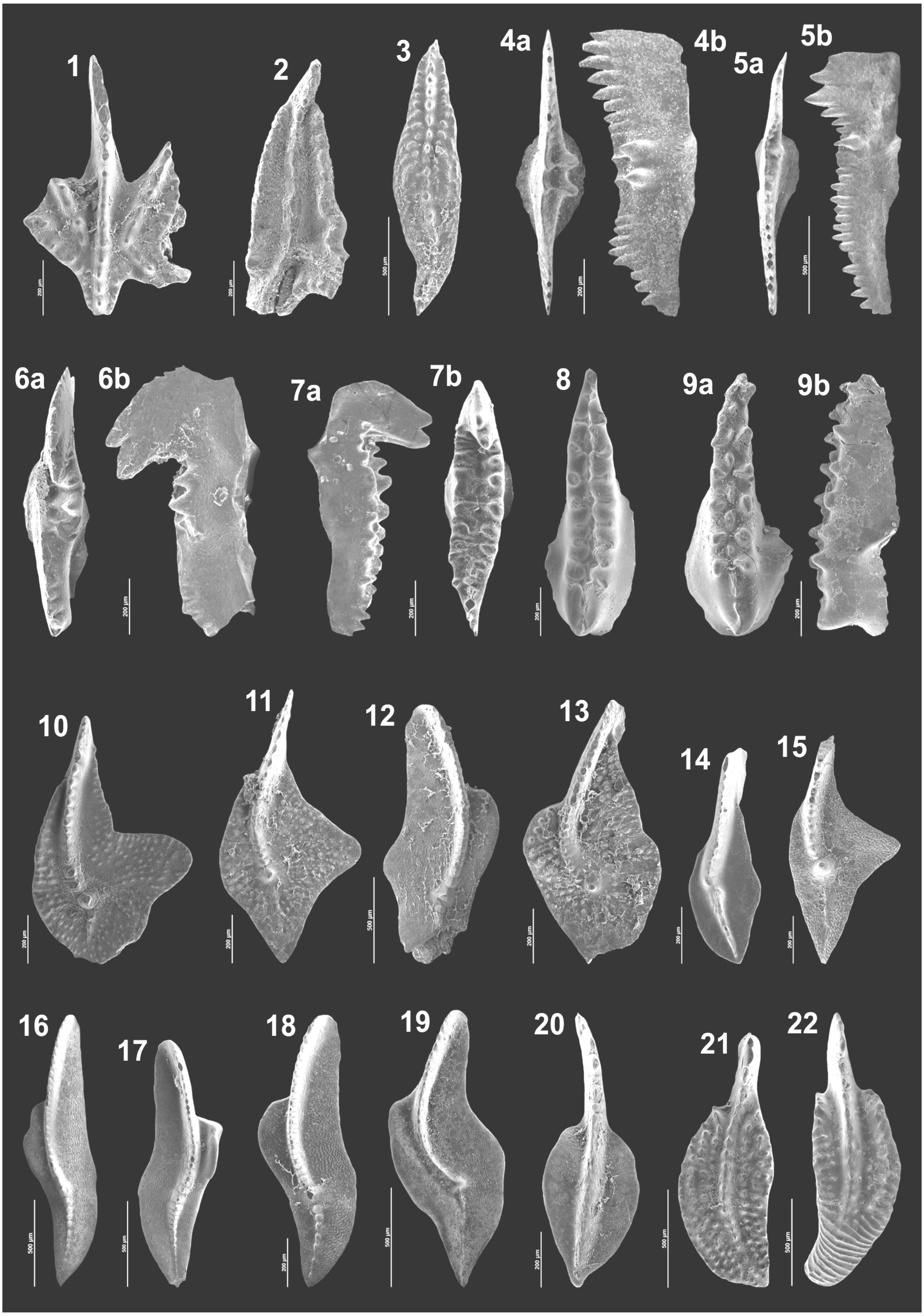


Pl. 2. Frasnian-Famennian microfacies examples from Oum el Jerane. **1.** Bed -39, upper part of Middle Frasnian (ca. MN 9 Zone), bioturbated, silty grainstone (calcisiltite, poor bio- and lithoclastics as in Bed -31h); width ca. 36 mm. **2.** Bed -31h, basal Upper Frasnian (lower MN 11 Zone), silty calcisiltite (peloidal, partly microsparitic grainstone) with brachiopod valve (1), ostracods, fine crinoid and mollusk debris, and partially preserved micrite matrix; width ca. 20 mm. **3.** Bed -19, laminated, silty, unfossiliferous calcisiltite with convolute bedding, suggesting a turbiditic origin; width ca. 29 mm. **4.** Bed -3, lower Famennian, Uppermost *crepida* Zone, bioturbated, silty pack-grainstone with cheiloceratid cross-section (1, phragmocone filled by coarse orthosparite), brachiopods (2), abundant crinoidal remains (3), rounded, partially coated micrite extraclasts (5), and horizontal burrow with finer, darker filling in the middle; width ca. 32 mm **5.** Bed 1, lower Famennian, Uppermost *crepida* Zone, brachiopod grain-rudstone with *Stromatactis*-type fenestral (microbialithic) fabric in the lower part; width ca. 23 mm. **6.** Higher part of Bed 1, showing gradation into a silty brachiopod float-rudstone with small in-situ rhynchonellids (1) and crinoid debris; width ca. 41 mm.



Pl. 3. Microfacies of Famennian strata from Oum el Jerane (see HARTENFELS 2011). **1.** Bed 4a, lower Famennian, Uppermost *crepida* Zone, siliciclastic crinoidal grain-rudstone with angular quartz grains (1), micritic (2) and phosphatic (3) extraclasts, some with marginal coating (4); width ca. 17 mm. **2.** Bed 6a, upper Famennian, basal *Costaclymenia* Limestone (Dasberg Crisis Interval, UD V-A1, *Bi. aculeatus aculeatus* Zone), cephalopod rudstone with strongly encrusted ammonoids and crushed mollusk shells. Geopetal fabrics are partly oblique and document exhumation and resedimentation prior to final cementation; width ca. 43 mm. **3.** Bed 8a, top layer of *Costaclymenia* Limestone (*Bi. aculeatus aculeatus* Zone), cephalopod rudstone with often strongly fragmented ammonoids and orthocones, crinoid remains (1), ostracods, brachiopods (2), rare heterocorals (3), and microbial encrustions (4); width ca. 35 mm. **4.** Detail of heterocoral from Bed 8; width ca. 2 mm. **5.** Bed 9, higher *Bi. aculeatus aculeatus* (Sub)Zone, crinoid pack-rudstone showing hematite impregnation and diagenetic dissolution of particle margins; width ca. 45 mm. **6.** Bed 11b, *Gonioclymenia* Limestone (UD V-B, *Bi. costatus* Subzone), cephalopod (2) rudstone with micritic matrix, ostracods and common crinoid remains (1); width ca. 41 mm.

Pl. 4. Frasnian conodonts from Oum el Jerane, 1-12, Bed 16 (MN 4 Zone), , . **1.** *Ad. africana*, **2.** *Ad. pramosica*, **3.** *Ad. recta*, **4.** *Ad. alata*, **5.** *Ad, rugosa*, **6.** *Ziegl. ovalis*, **7.** *I. tafilensis*, **8.** *Po. webbi*, **9.** *Po. pennatus*, **10.** *Mes. guanwushanensis* (with unusually irregular ornament), **11.** *I. subterminus*, **12.** *Po. ordinatus*, **13.** *Ag. amplicavus*, **14.** *I. vitabilis*, **15.** *Pa. ljaschenkoae*, **16.** *Ad. curvata* Late Morphotype, **17.** *Pa. hassi*, **18.** *I. alternatus alternatus*.



Pl. 5. Conodonts from the Famennian of Oum el Jerane. **1.** *Ad. curvata* Late Morphotype, Bed 4a (reworked), **2.** *Ag. leonis*, Bed 4a (reworked), **3.** *“Ag.” sinelaminus*, Bed 5a, **4.** *Bi. aculeatus aculeatus*, Bed 6a, **5.** *Bi. aculeatus anteposicornis*, Bed 6a, **6.** *Clyd. plumulus*, Bed 8a, **7.** *Clyd. ormistoni*, Bed 6a, **8.** *I. symmetricus*, Bed 1 (reworked), **9.** *I. alternatus alternatus* M1, Bed 6a (reworked), **10.** *Pa. hassi*, Bed 4a (reworked), **11.** *Pa. triangularis*, Bed 5a (reworked), **12.** *Pa. tenuipunctata*, Bed 5a, **13.** *Pa. crepida*, Bed 4a, **14.** *Pa. minuta minuta*, Bed 5a, **15.** *Pa. minuta loba*, Bed 4a, **16.** *Pa. glabra lepta* Early Morphotype, Bed 4a, **17.** *Pa. glabra pectinata* M1, Bed 4a, **18.** *Pa. glabra prima* M3, Bed 5a, **19.** *Pa. klapperi*, Bed 4a, **20.** *Po. glaber glaber*, Bed 4a, **21.** *Po. nodocostatus nodocostatus*, Bed 4a, **22.** *Po. semicostatus* “Central Morphotrend”, Bed 4a.

MIDDLE FAMENNIAN TO MIDDLE TOURNAISIAN STRATIGRAPHY AT EL ATROUS (AMESSOUI SYNCLINE, SOUTHERN TAFILALT)

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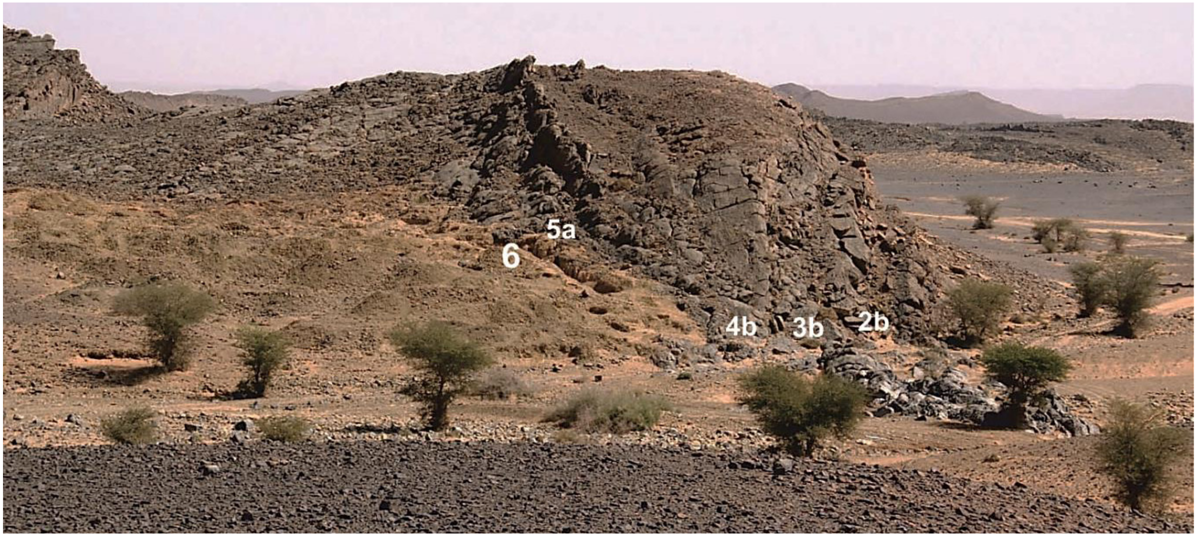


Fig. 1. The middle Famennian cliff of crinoidal limestone and debris flows (bed numbering from HARTENFELS 2011), overlain (6) by the deep trench and excavations of the *Gonioclymenia* Limestone (DU V-B).

1. INTRODUCTION

The investigated succession lies ca. 12 km NW of Taouz at the southern limb of the Amessoui Syncline, just south of the abandoned El Atrous village and along the piste from Taouz to Rissani. The GPS position for the base is N 30°59′5.5′′ W 4°5′47.0′′ (see also map in El Khraouia chapter, BECKER et al. this vol.). The locality became famous when HOLLARD (1960) discovered the regionally first *Gattendorfia* fauna, which was re-discovered much later independently by KORN et al. (2002) and the Bochum-Münster Group (see KAISER 2005, BECKER et al. 2006, and BECKER 2010). HOLLARD (1974), who concentrated on the Middle Devonian, labelled four sections as El Atrous but all are located on strike to the west. KORN et al. (2000) investigated middle/upper Famennian ammonoid localities of the Amessoui Syncline. Our section is in the middle between their sections El Atrous West and East, but it is identical with El Atrous East in HARTENFELS (2011). GINTER et al. (2002) and EICHHOLT (2007) studied upper Famennian shark teeth from our locality. The D/C boundary interval was investigated in detail by KAISER (2005) and KAISER et al. (2011), which results are summarized here.

2. MIDDLE FAMENNIAN

As first noted by KORN et al. (2000), the middle Famennian cliff just north of the piste along the El Atrous ridge (Fig. 1) is characterized by thick, coarse debris flows. They were derived from a now eroded inner shelf carbonate platform just to the south. Our section begins with 73 cm crinoidal limestone (pack-rudstone, Beds 1 and 2a), which are rich in probable green algae (*Issinella*), a typical form of middle Famennian crinoid mounds in Belgium (DREESEN et al. 1985). This facies gives evidence for deposition on or from a shallow carbonate platform within the photic zone (Fig. 2.1). Conodonts are diverse (Fig. 4, 17 taxa) and include marker species, such as *Scaphignathus velifer velifer*, *Sc. velifer leptus, Palmatolepis gracilis semisigmoidalis, Polygnathus pseudotenellus* (Fig. 5.12-13), and *Po. subirregularis*. The assemblage falls in the *trachytera* Zone of the “standard zonation” and in the extended regional *velifer* Zone of the Tafilalt. The overlying thick debris flow (Bed 2b) cuts into Bed 2a (Fig. 3). A similar sedimentation, with two more massive debris flows (Beds 3b and 4b), continues for another 1.7 m (Fig. 4). Bed 3a is strongly recrystallized, with syntaxial overgrowth of the crinoid fragments (Fig. 2b). The conodont assemblages become poorer, but with the first local entries of *Alternognathus regularis continuus* and *Po. perplexus* in Bed 4a The whole interval belongs to the (undivided) *trachytera* Zone. The presence of reworked Upper Frasnian (e.g., *Ancyrodella curvata* Late Morphotype, Fig. 5.1, and *Pa. winchelli*, Fig. 5.8) and lower Famennian (e.g., *Pa. quadrantinodosalobata*, Fig. 5.7) conodonts in all samples shows a deep erosion in the source region, during a significant, long episode of low sea-level. It resulted from a mixture of the general regressive eustatic trend in the higher middle Famennian (JOHNSON et al. 1985) and uplift (block faulting). Variable high proportions of *Icriodus*, of the shallow-water (Ardennes Shelf) "*Po.*" *semicostatus* Group, and of *"Po. diversus"* support a neritic biofacies. Eroded blocks of the debris flows include both crinoidal and cephalopod limestones.

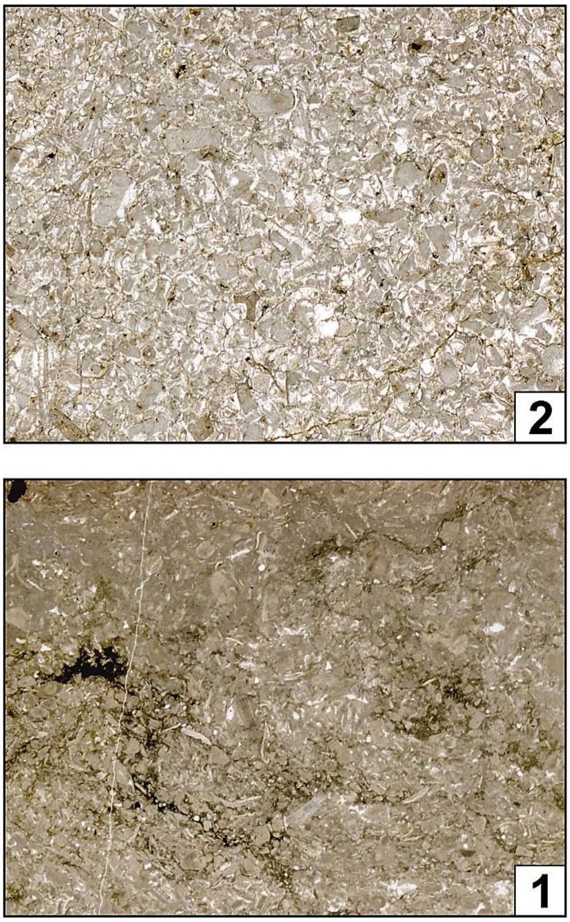


Fig. 2. 1. Bed 1, crinoid pack-rudstone with strongly fragmented crinoid debris and abundant tubular fossils, probably the green algal genus *Issinella* (compare DREESEN et al. 1985); figure width ca. 17 mm. **2.** Bed 3a, crinoid grain-rudstone with syntaxial echinoderm overgrowth; figure width ca. 20 mm.

3. UPPER FAMENNIAN

Bed 5a is a 7 cm thick, middle-grey crinoidal limestone, which is similar to the beds below. Rare *Protactoclymenia* and the conodont fauna, however, prove a significant hiatus at the base, which spans probably the top of the *trachytera* Zone and the main part of the *Platyclymenia* Stufe (UD IV-A/B), including the *Annulata* Event Beds. Marker condonts are *Bispathodus stabilis bituberculatus* (Fig. 5.2), *Bi. stabilis stabilis*, both morphotypes of *Pa. gracilis expansa* (Figs. 5.5-6), *Po. delicatulus* (Fig. 5.9), *Po. homoirregularis* (Fig. 5.11), *Po. margaritatus* (Fig. 5.10), and *Pseudopolygnathus inordinatus* (Fig. 5.15). The bed clearly falls in the *stabilis* Zone (= Lower *expansa* Zone). Records of *Sporadoceras orbiculare* from the same level at El Atrous East (KORN et al. 2000) are in full agreement and confirm an UD IV-C age (regional *orbiculare* Zone). Again, there are reworked Upper Frasnian to lower Famennian conodonts. A hematitic, red crust at the top (Bed 5b) indicates another sedimentary break, which spans the lower Dasbergian (UD V-A), including the dark Dasberg Event Beds that are so well exposed at Oum el Jerane (HARTENFELS 2011), only a few km to the west.

Fig. 3. Erosive contact between the crinoidal Bed 2a and the brecciated debris flow of Bed 2b.

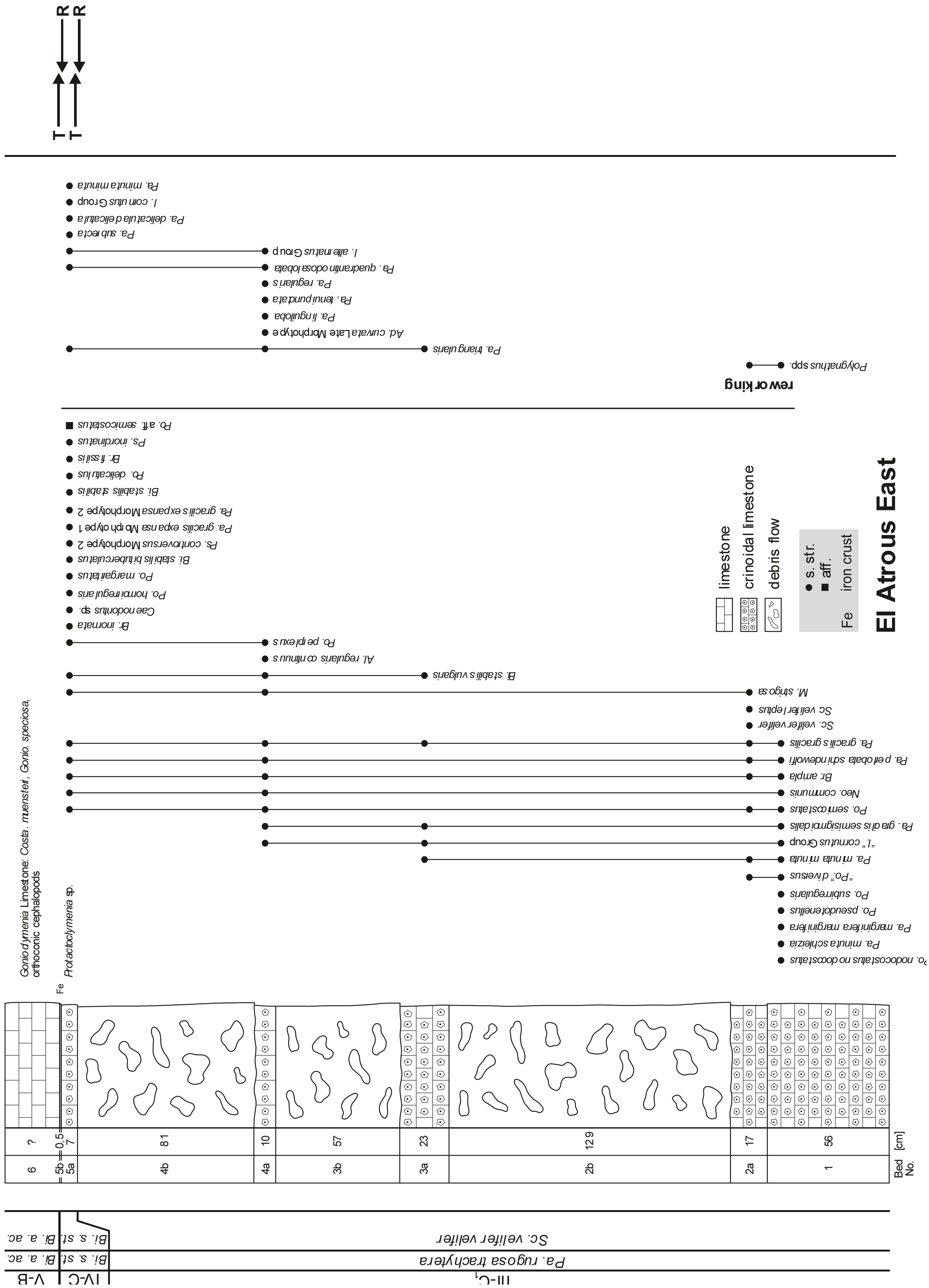
The subsequent *Gonioclymenia* Limestone (Bed K1, numbering of KAISER 2005) has been quarried deeply and completely, creating dangerous holes and tunnels that may collapse (**Attention** !). Occasionally, fragments of the large *Gonio. speciosa*, a marker of UD V-B (*hoevelensis* Zone), can be found. There are also some trilobites and solitary Rugosa. The conodonts fall in the *aculeatus aculeatus* (= Middle *expansa* Zone, HARTENFELS & BECKER 2012).

4. UPPERMOST FAMENNIAN

Topmost Erfoud Formation

The quarrying has largely destroyed the outcrop but the presence of condensed uppermost Famennian strata can be deduced from loose blocks of nodular limestone and of green or pink marl with proetids, large-eyed phacopids, gastropods, small brachiopods, and solitary rugose corals. *Parawocklumeria paradoxa*, the zonal marker for UD VI-C_2, occurs in Bed K2 together with *Bispathodus ultimus* M2 (Fig. 6), the index for the *ultimus* or Upper *expansa* Zone. The top of the Erfoud Formation is incomplete. Ammonoids and conodonts of the youngest pre-Hangenberg Event beds (*Wocklumeria* Zone, UD VI-D) are missing, as everywhere in the Tafilalt (BECKER et al. 2002). This indicates non-deposition due to pre-Hangenberg Event shallowing and increased episodic bottom turbulence. This unconformity represents a sequence boundary.

Fig. 4. Middle Famennian section log and conodont faunas at El Atrous (HARTENFELS 2011: fig. 43).



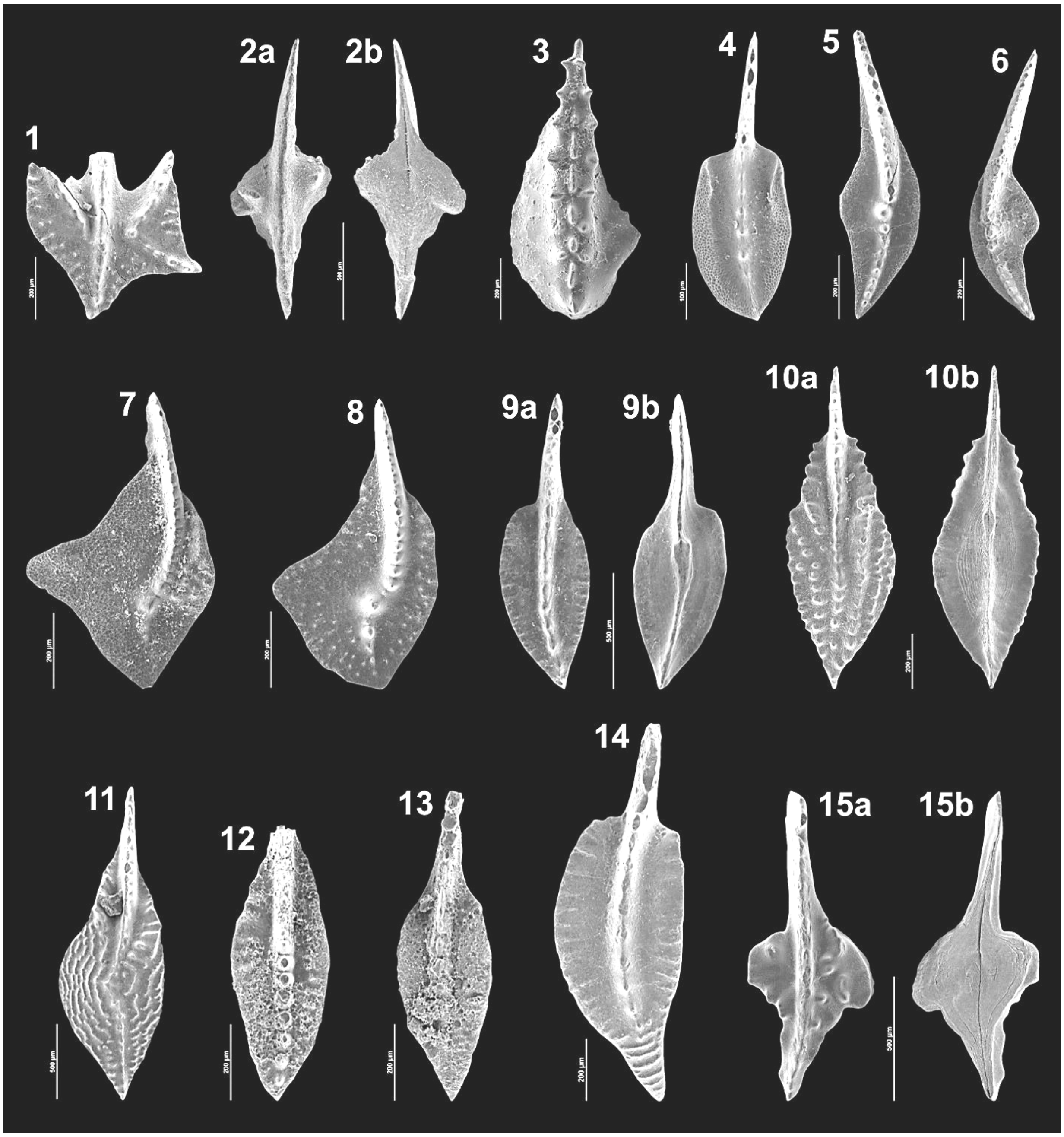


Fig. 5. Conodonts from the middle/upper Famennian of El Atrous (Bed 1-4b = undivided *trachytera* or regional *Sc. velifer velifer* Zone, Bed 5a = *Bi. stabilis stabilis* or Lower *expansa* Zone). **1.** *Ad. curvata* Late Morphotype, Bed 4a (reworked), **2a-b.** *Bi. stabilis bituberculatus*, Bed 5a, **3.** *I. alternatus alternatus* M1, Bed 4a (reworked), **4.** *Neo. communis*, Bed 5a, **5.** *Pa. gracilis expansa* M1, Bed 5a, **6.** *Pa. gracilis expansa* M2, Bed 5a, **7.** *Pa. quadrantinodosalobata*, Bed 4a (reworked), **8.** *Pa. winchelli*, Bed 5a (reworked), **9a-b.** *Po. delicatulus*, Bed 5a, **10a-b.** *Po. margaritatus*, Bed 5a, **11.** *Po. homoirregularis*, Bed 5a, **12-13.** *Po. pseudotenellus*, Bed 1, **14.** *Po semicostatus* “Morphotrend 3”, Bed 5a, **15a-b.** *Ps. inordinatus*, Bed 5a.

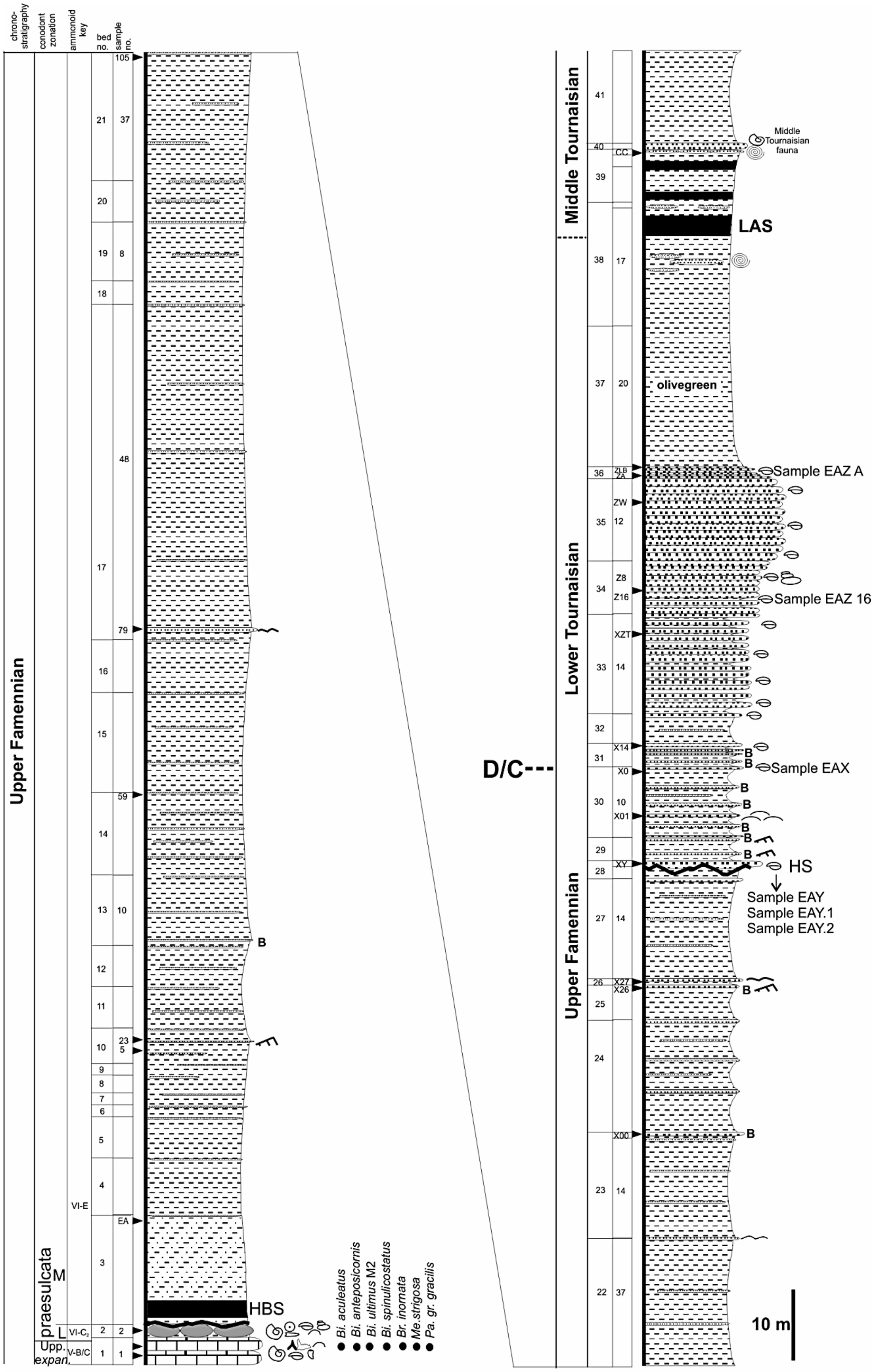


Fig. 6. Lithostratigraphy and uppermost Famennian conodonts at El Atrous (KAISER 2005, fig. 33).

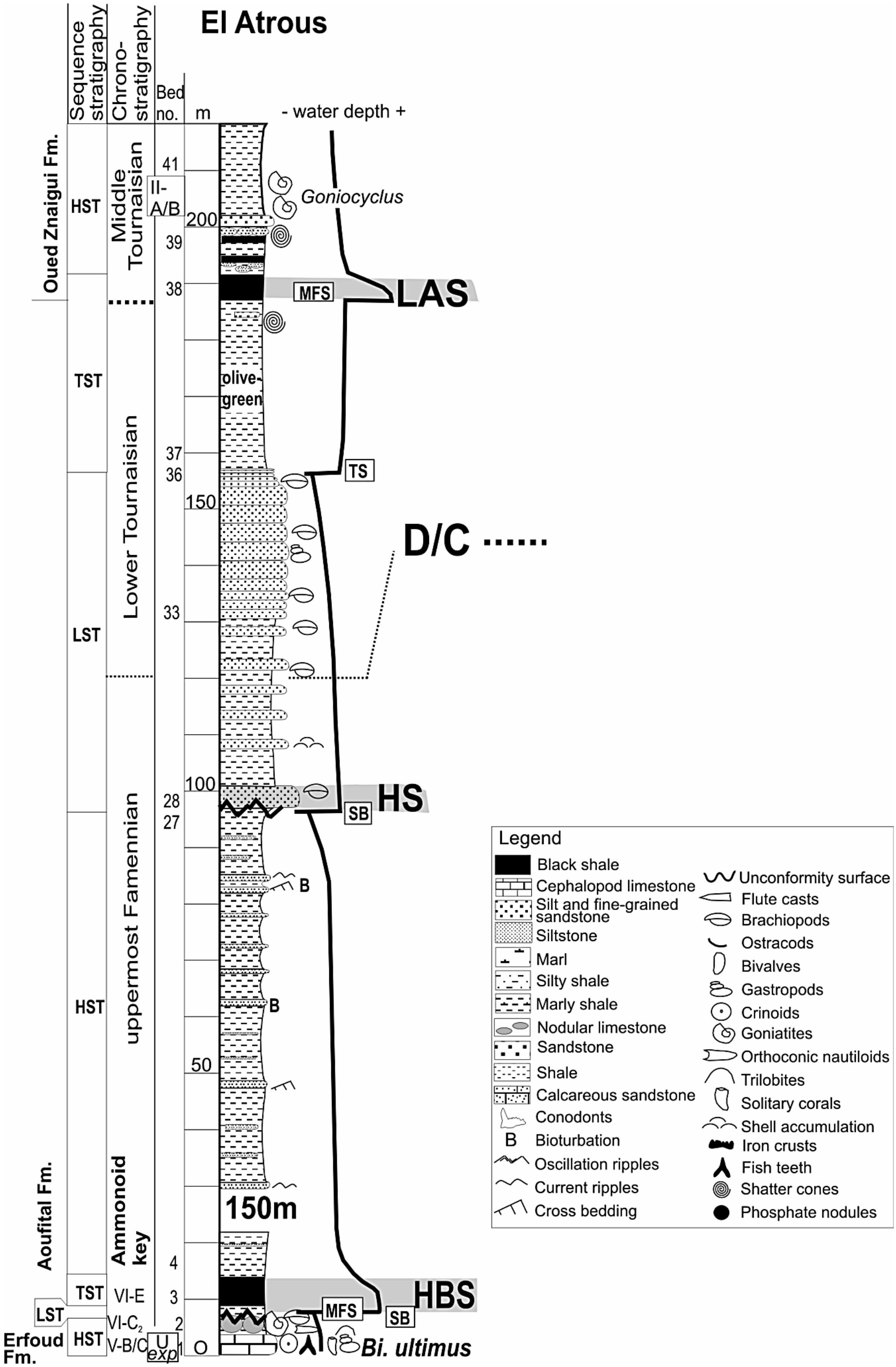


Fig. 7. Lithology, black shale events (HBS = Hangenberg Blackshale equivalents, LAS = Lower Alum Shale equivalents, sequence stratigraphy, and sea-level history at El Atrous (extracted from KAISER et al. 2011, fig. 5).

Hangenberg Black Shale Equivalents

Reddish-white weathering, unfossiliferous shales (lower Bed K3), probably with sulphates from pyrite weathering, follow above the limestone succession. The base of this poorly exposed anoxic level is a sharp marine flooding surface (drowning unconformity).

Aoufital Formation

The lower part of the Aoufital Formation (upper Bed K3 to K27) consists of a more than 260 m thick succession of alternating unfossiliferous, greenish shales and thin-bedded, silt- to fine-grained sandstones. Laterally to the east, at Jebel Ouoaoufilal, a thin sideritic layer with the oldest *Acutimitoceras (Stockumites)* proves a lower UD VI-F age. Parallel and convolute bedding as well as ripple cross lamination of thin sandstone beds are typical for low-density turbidity currents (LOWE 1982) of a deeper pro-deltaic setting, which partly filled up a slowly subsiding basin. It reflects a highstand phase (HST or FSST caused by forced regression) and correlates with the main, green-grey Hangenberg Shale of the Rhenish Massif (BLESS et al. 1993). Turbiditic prodeltaic facies are most likely to be deposited when the rate of sea-level fall is high and when humid conditions enable a rapid discharge of clastics from the terrestrial to the open marine realm. The thickness of Beds K3 to K27 proves that the HST/FSST accomodation space of El Atrous was more than 200 m in a relatively short time span; much of this was a consequence of rapid infilling of the subsiding deep basin with fine-grained siliciclastics. The amount of synsedimentary subsidence and eustatic fall is difficult to assess at El Atrous but the combined effects of rapid sedimentation and sea-level drop resulted in more than 100 m of bathymetric change (from deeper pelagic to shallow subtidal/intertidal). The similar base level fall in other regions, such as in the more condensed sections at M´Karig or Ouidane Chebbi (KAISER et al. 2011), with almost no sediment infilling, suggests a significant eustatic sea-level change.

This rather monotonous unit passes upwards into more proximal, non-turbiditic sandstones and terminates with an erosive channel filled by a laterally wedging mass-flow deposit (Bed K28). Although a role of synsedimentary tectonics is likely, we interpret the much more proximal clastics as an equivalent of the Rhenish Hangenberg Sandstone. The mass flow can be interpreted as a basin slope channel or incised valley fill (LST) above a sequence boundary. It contains univalved, laterally transported, ribbed and small-sized shallow-water brachiopods (*Centrorhynchus* (?) sp., Sample EAY.1, and *Hemiplethorhynchus* sp., Sample EAY.2), which indicate an uppermost Famennian to basal Tournaisian age (BRICE et al., 2005, 2007). Normally the two taxa should not overlap but they have been found together in the same position at Lalla Mimouna (see BECKER et al. this vol.).

5. D/C BOUNDARY INTERVAL

The overlying succession (Beds K29–K36) starts with shales and intercalated thin sandstone beds without shelly benthos. They signal a minor deepening above the LST. This unit terminates with proximal sandstone deposits, yielding masses of brachiopods and occasionally gastropods. The rhynchonellids from the uppermost sandstone beds (Samples EAX, Fig. 8.1, and EAZ 16, Fig. 6) belong to *Hemiplethorhynchus* (?) sp. cf. *allani*. They prove shallow-water conditions and, in the absence of *?Centrorhynchus*, possibly a Tournaisian age (compare BRICE et al. 2005, 2007). Thin sections reveal abundant brachiopod shells and compositionally mature, texturally immature sandstones. They indicate the progradation of a shoreline and a minor regression (a parasequence boundary), as it is known from the D/C boundary of Germany (BLESS et al. 1993). Therefore, the latter is tentatively placed at the beginning of the brachiopod-rich interval (Bed K30).

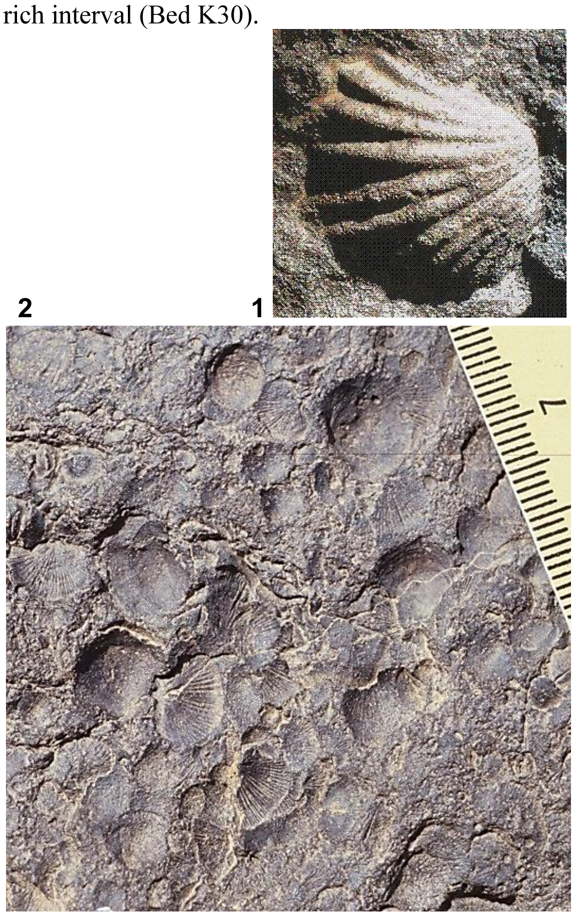


Fig. 8. 1. Rhynchonellid *Hemiplethorhynchus* from Bed K30 (Sample EA X, det. D. BRICE), **2.** Chonetid plaster from the top of the El Atrous ridge (Bed K36).

The overall LST terminates with thin-bedded sandstones (Beds K35-36) with chonetid plasters (Fig.

8.2) and gastropods (Fig. 6, Sample EAZ A), typical for a maximum lowstand (HAQ et al., 1988). This unit forms the top of the steeply dipping ridge (Fig. 9). The subsequent basal Carboniferous early HST begins with thin-bedded siltstones, shales (Bed K37) or sandstones. They are overlain by retrogradational, transgressive, hemipelagic to pelagic deposits. Thus, the sea level rose rapidly in the early Tournaisian and the accumulation of shales reflects an offshore deposition without coarse clastic influx in a low-energy depositional environment. The late Lower Tournaisian *Gattendorfia* faunas of the Mfis area (BOCKWINKEL & EBBIGHAUSEN 2006) have not yet been found in the Amessoui Syncline.

Fig. 9. View of the steep cliff of the upper Aoufital Formation (Beds K35-36) and poor exposure of the lower Oued Znaigui Formation in the foreground.

6. MIDDLE TOURNAISIAN

Poorly exposed, reddish-white weathering, probably originally pyrite-rich shales (Beds K38- 39) without any fauna define the base of the argillaceous Oued Znaigui Formation. They are interpreted as an anoxic equivalent of the German Lower Alum Shale at the base of the Middle Tournaisian, which represents a significant transgressive and eutrophic interval (e.g., SIEGMUND et al. 2002). Occasionally, sandstone lenses with cone-in-cone structures are intercalated. Most likely, they represent dewatering structures of seismite layers and provide evidence for synsedimentary tectonic movements near the Lower/Middle Tournaisian boundary (see TAHIRI et al. this vol.).

Immediately above the multicolored shales decalcified siltstone nodules appear, which sometimes contain three-dimensionally preserved or squashed goniatites (KORN et al. 2002, BECKER et al. 2006), orthocones, and ribbed bivalves. Phragmocones were mostly dissolved and leave hollows. Therefore, it is not easy to get specimens with sutures. Involute forms are better preserved than evolute ones. The ornament is impressed on the inner moulds, which bear negative traces of endoliths. The first fauna was collected between the ridge slope and the track to Rissani. The assemblage is characterized by *Goniocyclus ammari*, as local zonal index, Gen. aff. *Goniocyclus* n. sp. (with rounded instead of sharply angular ventral ribs), abundant *Protocanites* n. sp., two new species of *Imitoceras*, a new species of *Ac. (Stockumites),* a new relative of "*Gattendorfia*" *costata*, and rare specimens of the oldest ceratitoid ammonoid (Prodromitidae n. gen., see BECKER 2010). This fauna falls in the lower part of the Middle Tournaisian *Goniocyclus* Genozone (cu II-A, BECKER 1993).

The same *Gonio. ammari* assemblage can be found for another ca. 15 m and along a low S-N running hill with flat top. Separated by ca. 40 m of shale, the richest fauna was collected on the other side of the track. It contains a range of different *Goniocyclus* species, including *Go. elatrous* as the index for a second local Middle Tournaisian goniatite zone. *Proto. hollardi, Protocanites* n. sp (both mostly as negative imprints)., and *Imitoceras* n. sp. 1 are very common. *"Globimitoceras" rharhizzense,* the new prodromitid, the constricted "*Imitoceras*" *abundans* Gp., and relatives of *Gatt. costata* are much rarer. A new, strongly asymmetric bivalve genus is very distinctive. It is accompanied by three other bivalve species. HAHN et al (2012) described recently an isolated (loose) new *Moschoglossis* laterally from ca. this interval in open nomenclature. More trilobite remains would be very useful.

The dominance of goniatites and the very restricted benthos (few bivalve) signal a deep outer shelf habitat, with deposition at larger depths than the condensed Famennian pelagic limestones and marls. The upper level with *Go. elatrous* has more benthos than the basal concretions with *Gonio. ammari*.

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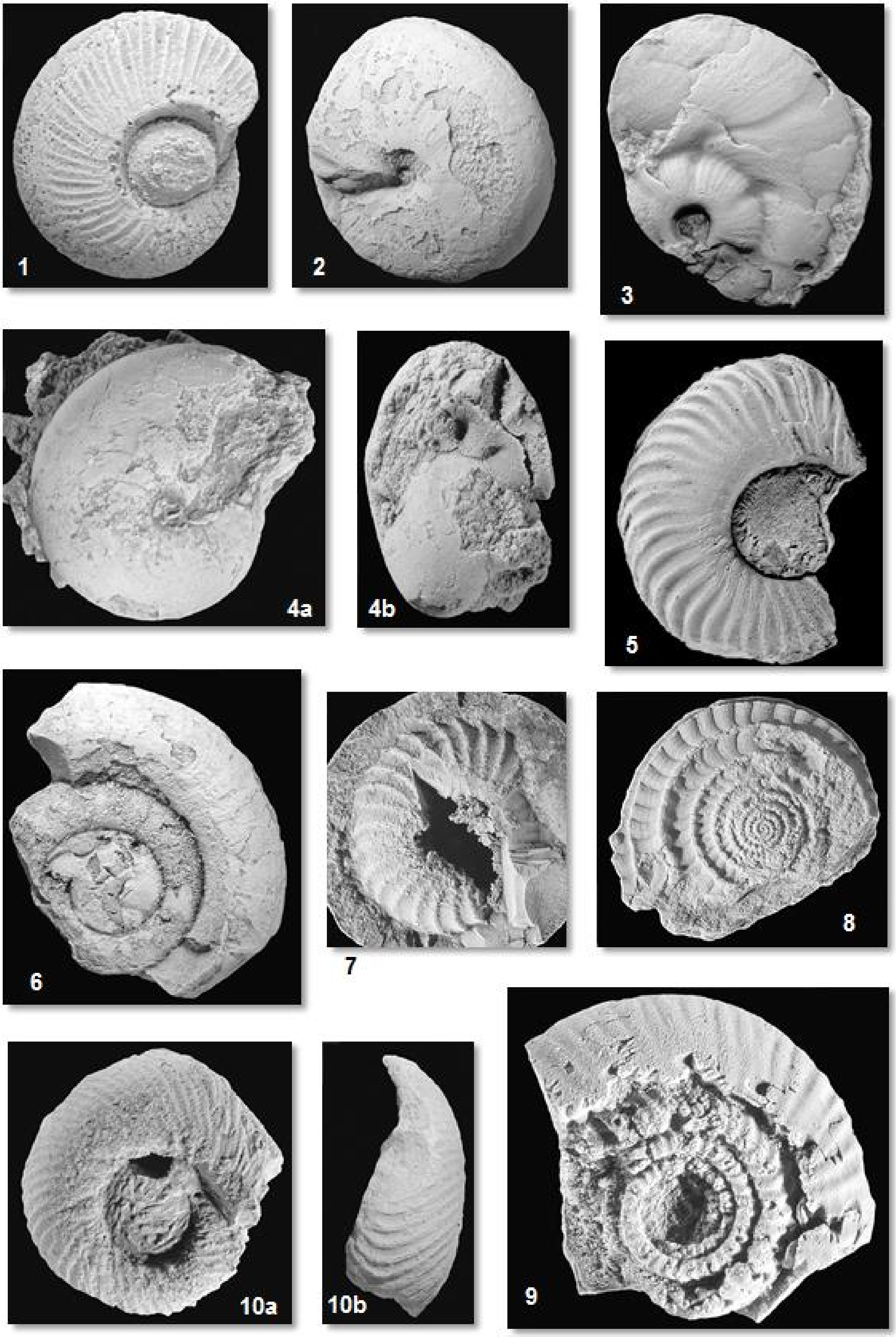

Pl. 1. New goniatites from the lower Oued Znaigui Formation (*Goniocyclus ammari* and *Go. elatrous* Zones) of El Atrous. **1.** “*Gattendorfia*” n. sp. aff. *costata*, **2.** *Acut. (Stockumites)* n. sp., **3.** *“Globimitoceras” rharhizzense*, **4a-b.** *Imitoceras* n. sp. 1, **5.** *Goniocyclus* aff. *elatrous*, **6.** *Protocanites* n. sp., **7.** *Goniocyclus ammari*, **8-9.** Prodromitidae n. gen. n. sp., juvenile and median-sized specimen (future holotype), **10a-b.** Gen. aff. *Goniocyclus* n. sp.

PRELIMINARY DATA ON VISEAN (CARBONIFEROUS) CORALS AND BRACHIOPODS FROM THE STRATA BETWEEN THE DJEBEL BEGAA AND THE GARA EL ITIMA (EASTERN TAFILALT, MOROCCO)

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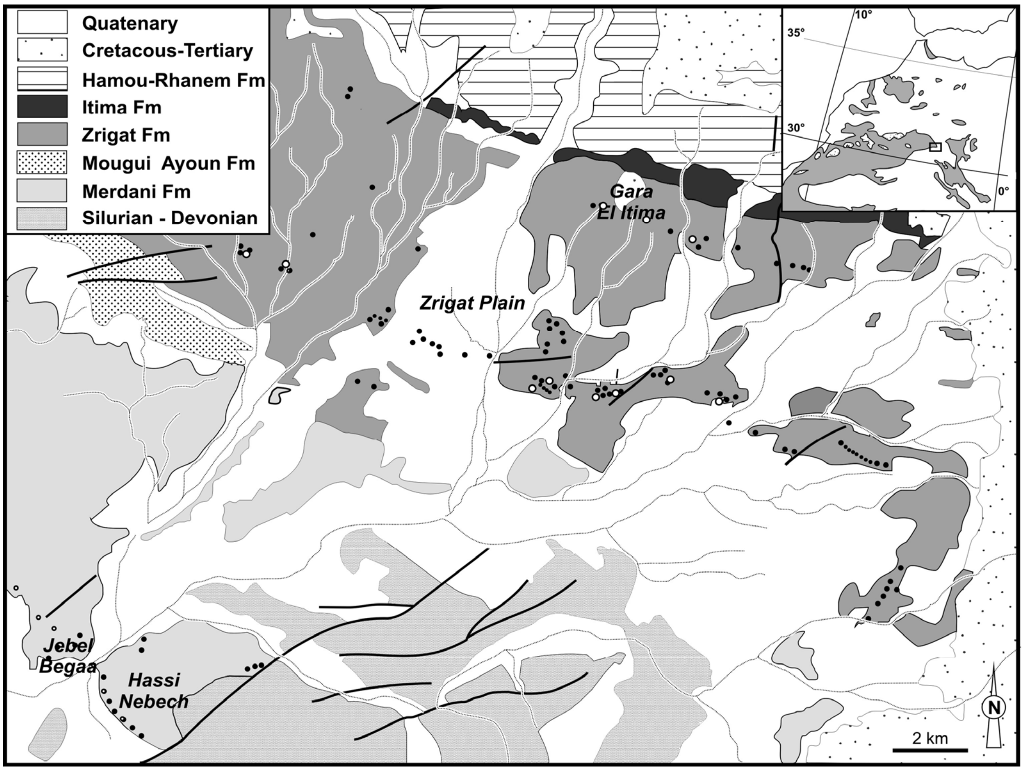

Fig. 1. Geological map of the study area after WENDT et al. (2001). Black points show the position of the mud mounds sensu WENDT et al. (2001). White points show the main sample locations.

1. INTRODUCTION

The abundance of data for the spatial and temporal distribution of Mississippian corals and brachiopods in North Africa is variable. Some regions are well studied (e.g. Moroccan Meseta: SAID et al. 2007, 2013, ARETZ 2010, ARETZ & HERBIG 2010,), whereas data are relatively scarce for others (e.g. Mouydir [Algeria]: Mottequin & Legrand-Blain 2010; Aretz 2011).

South-eastern Morocco (eastern Anti-Atlas) is among the second category, although Carboniferous macrofauna has been already known since the works of CLARIOND (1932, 1934) and DELÉPINE (1939, 1941). This lack of data is even more striking when compared to well-known faunas of other Palaeozoic strata in that area. The pioneer workers established faunal lists containing mainly ammonoids, brachiopods and corals. It is only in recent years that this fauna has regained some attention; today modern data are available for ammonoids (KORN et al. 1999, KLUG et al. 2006) and gastropods (HEIDELBERGER et al. 2009). WENDT et al. (2001) provided some determinations of corals (family and genus level) and brachiopods from outcrops around mud mounds. The Viséan ammonoid assemblages will be separately presented in this volume (KORN et al.).

It is the aim of this contribution to present preliminary data on corals and brachiopods recently (late 2011) collected in the eastern Tafilalt on a transect from the Djebel Begaa to the Gara El Itima, and to place them into a facies and stratigraphic frame. This results also in a brief description of the so-called reefal facies in that area. The later have already been mentioned at the Djebel Begaa and in the Zrigat plain by PAREYN (1961) in his monumental study of the

Carboniferous succession of the Béchar Basin and adjacent regions. WENDT et al. (2001) have mapped over 100 of these reefal structures in three NW-SE trending mud-mound belts (Fig. 1).

2. SUCCESSION AND LITHOSTRATIGRAPHY

Earlier workers started to establish a lithostratigraphical frame (Fig. 2) of the area (e.g. DÉLÉPINE 1939, PAREYN 1961), but formalized formations were not established until the geological map of DESTOMBES & HOLLARD (1986). And even then, detailed descriptions or logs of these formations were missing for this several kilometre thick mixed siliciclastic-carbonate succession, which overall is dominated by shales.

However, the oldest formation, the Oued Znaïgui Formation consists of several hundred of meters of greyish to reddish shales. Conglomerates and sandstone beds are intercalated. Cone-in-cone structures can be locally abundant. This formation is topped by dark grey shales of the Merdani Formation (400-500 m thick) in which relatively thin siltstone and fine-grained sandstone beds are intercalated. In its lower part, the formation contains limestone lenses (‘reefal structures’) in the Jebel Begaa area (Fig. 3, A). The following Mougui Ayoun Formation (ca. 300m thick) is dominated by coarse siliciclastic rocks, often channelized sandstones and some conglomerates. Shales are intercalated and can locally attain several decimetres of thickness. The Zrigat Formation (more than 500 m thick) is again largely dominated by shales, but periodically decimetre thick sandstone and bioclastic or oolithic limestone beds occur. Characteristic is the presence of two intervals (the middle and upper mud mounds levels of WENDT et al. 2001), which contain limestone lenses of up to 15 m in thickness and several tens of meters in diameter (Fig. 3, A, F). At the top of the formation the succession is locally dominated by limestone and sandstone beds. The succeeding Itima Formation (200 m thick) is dominated by often channelized fluvial sandstones and intercalated finer-grained detritic rocks. It can be difficult to differentiate this formation from the shales succeeding Hamou-Rhanem Formation (200 m thick?). The youngest Carboniferous rocks correspond to a monotonous series of bedded micritic limestones with shale interbeds of the Hassi Mrheimine Formation (several tens of meters thick).

The difficulties to differentiate several of the formations defined by DESTOMBES & HOLLARD (1986), lateral facies changes and new conodont data resulted in a revised stratigraphic scheme (WENDT et al. 2001), in which the Mougui Ayoun and Itima formations become coarse-grained lateral facies equivalents of the Zrigat and respectively Hamou-Rhanem formations (Fig. 2). WENDT et al. (2001) also reduced the extension of the Merdani Formation in excluding the levels with reefal limestones from the formation. They were included in the Zrigat Formation. KLUG et al. (2006) proposed sequence stratigraphic interpretations for the middle and upper parts of the Zrigat Formation.

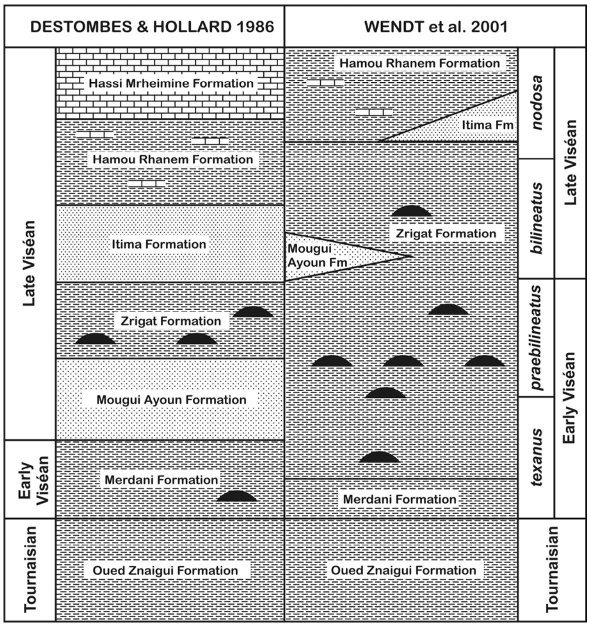

Fig. 2. Overview on the stratigraphic nomenclature of the study area showing the differences between the subdivision used for the geological map 1:200.000 (DESTOMBES & HOLLARD, 1986) and the modifications of WENDT et al. (2001). Modified from WENDT et al. (2001). The figure is not to scale. Approximate thicknesses for the formations in the definition of DESTOMBES & HOLLARD (1986): Oued Znaïgui Formation: several hundred meters, Merdani Formation: 400-500 m, Mougui Ayoun Formation: ca. 300m, Zrigat Formation: more than 500 m, Itima Formation: 200 m, Hamou-Rhanem Formation: 200 m thick?, Hassi Mrheimine Formation: several tens of meters.

3. SAMPLE LOCATIONS

More than 300 coral specimens have been recovered from a series of surface outcrops on a south-north transect starting at the Jebel Begaa and ending at the Gara El Itima. Loose specimens from the surface as well as specimens enclosed in the limestones have been sampled. The corallites of loose specimens are often abraded. The corals have mainly been found in the carbonated horizons of the Merdani and Zrigat formations, fewer samples are from more siliciclastic facies of the lithostratigraphical units.

3.1. Jebel Begaa

A rich and diverse coral assemblage has been recovered from a few meter-thick interval west and northwest of the Jebel Begaa (Fig. 3, A). There a conglomerate (locally rich in corals, Fig. 3, B) is topped by bedded to massive limestone. Lateral facies changes are relatively common in this interval, and well seen in the composition and rounding of the clasts in the conglomerate. Further towards the Jebel Begaa massive limestones are exposed. These limestones seems to be allochtonous (scarp breccia?) as seen in a small gully on the hill flank, where several tens of meters thick series of greenish shales and intercalated thin sandstone beds form the lateral equivalent of the limestones. The top of the Jebel Begaa is dominated by faintly bedded to medium-bedded bioclastic and crinoidal grainstones (Fig. 3, C). Very few corals and brachiopods have been found in this interval.

3.2. Hassi Nebech.

This locality corresponds to a NW-SE trending ridge SE of the Jebel Begaa. Strata are dipping towards the NE. The upper part of the ridge, forming several cliffs, is dominated by limestone resembling to massive reefs bodies surrounded by bedded facies (Fig. 3, D). The lower part of the ridge consists of shales. Some of these limestone cliffs are massive, faintly bedded limestone breccias (Fig. 3, E), whereas others correspond to bedded bioclastic, grainy limestones. Between the cliffs, bedded limestone and/or greyish-brownish shales occur. The view from the ridge crest towards the NE clearly shows the chaotic distribution of massive, bedded and shaly facies in this part of the succession. It is interesting to note that as on the Jebel Begaa mud-dominated carbonate facies, interpreted as mud mounds by WENDT et al. (2001) occurs only rarely, and that few large cliffs are, at best, locally rich in carbonate mud and spar-filled cavities. Thus the number of mud mounds indicated by Wendt et al. (2001) has to be questioned and some of the limestone blocks even might be olistoliths.

3.3. Zrigat plain

Corals have been collected from several small hills of the Zrigat plain. These are the so-called mud mounds and intermound facies of WENDT et al. (2001), which also have numbered them. Most samples have been collected from the 'main belt of Zrigat mounds' sensu WENDT et al. (2001), which crosses the central part of the Zrigat plain in WNW-ESE direction (Fig. 3, A ,F). This belt is composed of several tens of small hills of up to 20 m in height and 200 m in diameter. These hills are mainly composed of massive to faintly bedded limestones and surrounded by shales, locally carbonated. WENDT et al. (2001) defined four types of mounds: (A) massive crinoidal wacke- or packstones without stromatactis; (B) massive crinoidal wacke- or packstones with rare *stromatactis* (Fig. 3, G), (C) similar to (B), but allochtonous; and (D) biodetrital grainstone mounds (Fig. 3, H). The widely missing contacts between mound and intermound facies are one of the important problems when interpreting these mounds. WENDT et al. (2001) showed the allochtonous nature of some mounds (their type C), but it remains unclear if these are isolated cases or if also mounds showing the overall regional dip could be allochtonous. KLUG et al. (2006) underlined that some of these mud mounds are in fact channel fills and olistolith-like structures. For the moment it is difficult to estimate how many carbonate mud mounds are preserved in the Zrigat plain, but a considerable lower number than that of WENDT et al. (2001) is most likely. Or all are allochtonous, several being in a geometric position coherent with the regional structure? The co-existence of allochtonous and autochtonous mounds has also to be elucidated further.

3.4. Gara El Itima.

The upper part of the Zrigat Formation was sampled south of the Gara El Itima in the so-called 'northernmost (mound) belt' of WENDT et al. (2001). Here the formation seems to be richer in carbonates and sandstones. This corresponds most likely to the time interval for which KLUG et al. (2006) proposed the maximum progradation of a carbonate platform. A rich brachiopod fauna has been collected near WENDT et al. (2001)'s locality mound 100 (biodetrital mound) (Fig. 3, H).

4. FAUNA

4.1. Coral assemblages

Jebel Begaa

The fauna in the most western outcrops comprises a series of small solitary rugose corals like "*Pentaphyllum*" sp., *Rylstonia* sp., *Sychnoelasma* ssp. (Fig. 4, A), *Zaphrentites* s.l. sp., and *Cravenia* ssp., but also medium-sized solitary rugose corals like *Siphonophyllia* sp., Caninimorpha indet., *Cyathoclisia* sp., *Koninckophyllum*? sp., and *Merlewoodia* sp. The only colonial taxa are among tabulate corals, and all belong to the genus *Michelinia*. This coral assemblage is considered being of early Viséan age.

The massive limestones (scarp breccias?) at the bottom of the Jebel Begaa contain *Sychnoelasma* sp., and for the first time colonial rugose corals (*Siphonodendron scaleberense* NUDDS & SOMERVILLE, 1987 (Fig. 4, G)). Higher up, near to the peak of the Jebel, but still in the supposedly interval of green shales, large-sized siphonophyllids, provisionally attributed to *Siphonophyllia samsonensis* (SALÉE 1913) (Fig. 4, C) have been found. This fauna is most likely already late Viséan in age. 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.

Hassi Nebech.

A relatively rich coral fauna has been recovered from a block just west of the ridge and the ridge itself. This fauna comprises small and large-sized solitary and colonial rugose corals (*Amplexizaphrentis* ssp., *Cravenia* sp., *Siphonophyllia samsonensis*?, *Siphonophyllia* sp. 1, *Rylstonia sguilmensis* SEMENOFF-TIAN-CHANSKY, 1974, *Amygdalophyllum* sp., *Siphonodendron* gr. *kleffense* SCHINDEWOLF 1927 (Fig. 4, E), *Solenodendron horsfeldi* (SMITH & YU, 1943) and tabulate corals (*Michelinia* ssp.). Most fauna has been collected from the limestones, but some of the smaller solitary rugose corals and tabulate corals are from more shaly intervals. The corals indicate a late Viséan age, most likely Asbian, for all limestones in the Hassi Nebech area. The zaphrentids, caniniids and micheliniids from the shales just below the limestones do not allow a precise dating. However, a middle to late Viséan age is proposed.

Zrigat plain

The 'main belt of Zrigat mounds' *sensu* WENDT et al. (2001) contains a diverse Asbian coral fauna. Small solitary corals, mostly from shaly facies, are represented by *Rylstonia benecompacta* HUDSON & PLATT, 1927, *Rylstonia* sp., *Bradiphyllum* sp., *Cyathaxonia cornu* Michelin, 1847, *Cyathaxonia rushiana* VAUGHAN, 1906, and *Zaphrentites* s.l. sp. Large solitary corals are abundant in carbonate and siliciclastic facies. So far identified are *Merlewoodia* sp., *Siphonophyllia samsonensis*?, *Siphonophyllia*? sp., *Pseudozaphrentoides alloiteaui* SEMENOFF-TIAN-CHANSKY, 1974, Caninimorpha indet., *Dibunophyllum bipartitum* (MCCOY, 1849) (Fig. 4, B), *Koninckophyllum* cf. *destitum*, *Clisiophyllum*? sp., *Archnolasma*? sp., *Pareynia* sp., and *Axophyllum pseudokirsopianum* SEMENOFF-TIAN-CHANSKY, 1974. Colonial rugose corals are diverse and fasciculate forms dominate; *Siphonodendron irregulare* (PHILLIPS, 1836), *Siphonodendron pauciradiale* (MCCOY, 1844), *Siphonodendron martini* (MILNE-EDWARDS &HAIME, 1851), *Lithostrotion decipiens* (MCCOY, 1849) (Fig. 4, D), and *Solenodendron furcatum* (SMITH, 1925). There are also abundant tabulate corals (*Michelinia* ssp. (Fig. 4, H), *Multithecopora* sp.).

Gara El Itima

The coral assemblage from the highest carbonate levels of the Zrigat Formation comprises a diverse range of forms. It contains *Amplexizaphrentis* 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. (Fig. 4, F), *Michelinia* ssp., and *Palaecis* sp. This coral assemblage is late Viséan in age. It lacks typical markers for the Brigantian as colonial axophyllids, although Brigantian ammonoids have been found in this stratigraphic level (KLUG et al. 2006).

4.2. Brachiopod assemblages

Brachiopod data are highly biased on the transect and further material is required to reach better identifications. Only very few brachiopods have been collected from the successions of the Jebel Begaa (*Pustula* sp.), Hassi Nebech (Echinoconchoidea indet., *Latiproductus*? sp., *Spirifer [Mesochorispira?]* [Fig. 5, C]) and in the 'main belt of Zrigat mounds' sensu WENDT et al. (2001) (*Latiproductus* sp.) (Fig. 5, A-B). A richer assemblage has been recovered near mound 100 next to the Gara El Itima. It comprises *Marginatia* sp. (Fig. 5, G-H), *Buxtonia* sp., *Lamellosathyris lamellosa* (LÉVEILLÉ, 1835) (Fig. 5, I-J), *Tylothyris*? sp., *Angiospirifer*? sp. (Fig. 5, D-E), Brachythyridoidea indet. and Syringothyridoidea indet. *Dictyoclostus* cf. *pinguis* (MUIR-WOOD, 1928) (Fig. 5, F) was found in a similar stratigraphic position further to the east (around mound 98). It is presently difficult to use these brachiopods for a precise dating, but these preliminary data confirm the late Viséan age given by ammonoids and corals for the assemblages recovered from the Hassi Nebech, the Zrigat plain and the Gara El Itima.

5. DISCUSSION

5.1. *Stratigraphy*

The new data only partly confirm the previous biostratigraphic work. The data for upper part of the studied transect are largely in agreement with the existing data on ammonoids (KORN et al. 1999, KLUG et al. 2006). As already observed elsewhere in Morocco (ARETZ 2010), typical European coral markers for the Brigantian do not occur although ammonoids indicate this age in the upper part of the Zrigat Formation.

However, there is an important discrepancy in the lower and middle part of the transverse. The limestones forming the peak of the Jebel Begaa and the Hassi Nebech ridge contain clearly a late Viséan coral fauna including several typical markers. Thus the early Viséan age proposed by DESTOMBES & HOLLARD (1986) and WENDT et al. (2001) is too old. The same is true for the 'main belt of Zrigat mounds' sensu WENDT et al. (2001) that is early Viséan in age according to WENDT et al. (2001), but the coral data clearly show a late Viséan age. These new data result in the question for discrimination of the processes and sources responsible for the very thick and fast accumulation of fine-grained siliciclastic rocks during the late Viséan time in the eastern Tafilalt, at least on the studied transect.

Besides this, the question of the autochtonous or allochtonous nature of the carbonate bodies around the Jebel Begaa but also in the Zrigat plain is another important point for the reconstruction of the basin evolution of the eastern Tafilalt during Mississippian times. The area may be tectonically less stable as often thought.

5.2. Coral assemblages and facies

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 undissepimented solitary corals are often 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.

The taxonomic diversity and composition of the coral fauna found in the Zrigat plain resembles contemporaneous coral faunas found in Béchar Basin

(Algeria, SEMENOFF-TIAN-CHANSKY 1974) and the Jerada Basin (NE Morocco, ARETZ 2010). It is interesting to note that in the Jerada Basin (ARETZ 2010) the youngest coral assemblage indicates an older age than other organisms as observed at the Gara El Itima.

5.3. Palaeobiogeography

From a palaeobiogeographical viewpoint, the coral assemblages are typical Viséan faunas of the western Palaeotethys or Western European Province (SANDO 1990), which includes North Africa. The first results indicate that the eastern Tafilalt assemblages show not surprisingly strong similarities to the nearby Viséan basins of the Sahara (SEMENOFF-TIAN-CHANSKY 1974, ARETZ 2011) characterizing the northern margin of Gondwana. This is well seen in the number of species, found in the eastern Tafilalt, which have often been considered being endemic to the Béchar Basin. Their occurrences are so far the main difference with the faunas of the mobile Variscan belt towards the north, and thus foster reconstructions in which the Tafilalt Basin is thought to be in prolongation to the Béchar Basin.

As already mentioned above, the typical colonial axophyllid corals characterizing the basal Brigantian in Western Europe and Iberia) did not occur in the assemblages in the eastern Tafilalt. This may be the result of unfavorable regional facies, overall rareness of these forms in Northern Africa (only scarcely known in the Tindouf Basin and later in the Béchar Basin, SEMENOFF-TIAN-CHANSKY 1985), or a bias in the sampling. However, this absence in combination with the presence of some taxa specific for the Tafilalt and Béchar basins point to a minor palaeobiogeographical barrier between the northern Gondwana margin and the mobile Moroccan Variscan belt.

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Fig. 3. A. View from mound 100 in southern direction. The low hills in the middle are part of the ‘main mound belt’ in the Zrigat plain (distance to the photographer, about 6 km). The large hill is the Jebel Begaa (distance to the photographer, about 15 km). **B.** Conglomerate containing mainly limestone pebbles and a diversified coral fauna, northwest of the Jebel Begaa, Merdani Formation, early Viséan. **C.** Crinoidal facies dominating the massive limestones forming the peak of the Jebel Begaa, Merdani Formation, late Viséan. **D.** View on the Hassi Nebech ridge from the South. Note the presence of several limestones cliffs forming the ridge. **E.** Limestone breccias forming one of the cliffs in picture D, height of the picture about 10 m. **F.** Alignment of small hills in the ‘main mound belt’ in the Zrigat plain. **G.** Network of stromatactoid cavities in one of the mounds in the Zrigat area. Note the different orientation of the cavity system and the bedding. **H.** Biodetrital mound (mound 100) of WENDT et al. (2001) southwest of the Gara El Itima. The ‘mound’ may be in fact a channel filling.

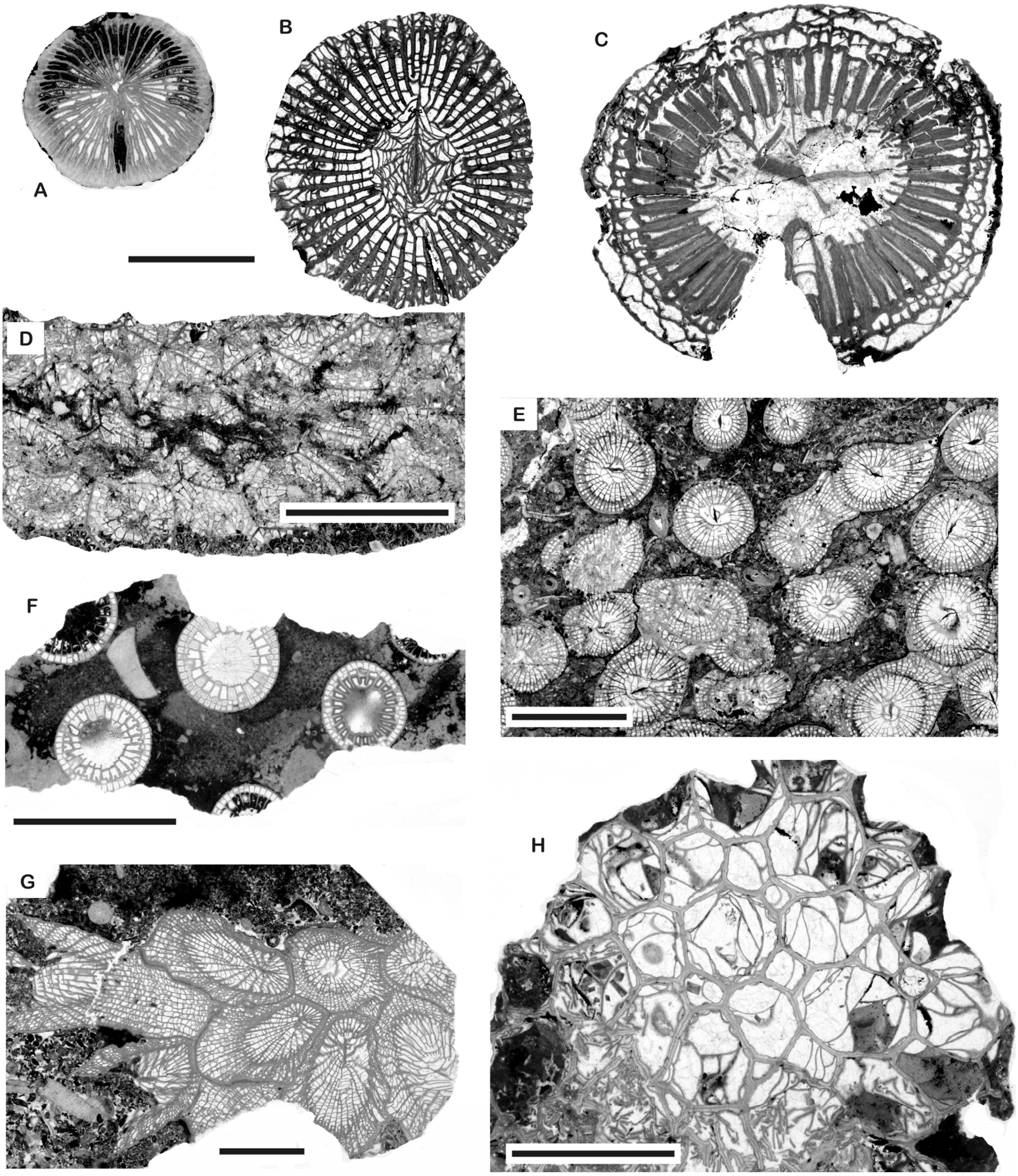

Fig. 4. A. *Sychnoelasma* sp.; from conglomerate northwest of the Jebel Begaa, Merdani Formation, early Viséan. **B.** *Dibunophyllum bipartitum* (McCoy, 1849); from the biodetrital mound 100 south of the Gara El Itima, Zrigat Formation, late Viséan. **C.** *Siphonophyllia samsonensis*?; from limestone block just below the peak of the Jebel Begaa, Merdani Formation, late Viséan. **D.** *Lithostrotion decipiens* (McCoy, 1849); mound 20, Zrigat plain, lower Zrigat Formation, late Viséan. **E.** *Siphonodendron* gr. *kleffense* Schindewolf 1927; Hasi Nebech, Merdani Formation, upper Viséan. **F.** *Diphyphyllum* sp.; southeast of the Gara El Itima, top Zrigat Formation, late Viséan. **G.** Subcerioid colony of *Siphonodendron scaleberense* Nudds & Somerville, 1987; massive limestone (scarp breccias?) northwest of Jebel Begaa, Merdani Formation, late Viséan. **H.** *Michelinia* sp.; mound 20, Zrigat plain, lower Zrigat Formation, late Viséan. Scale bars for all specimens: 10 mm.

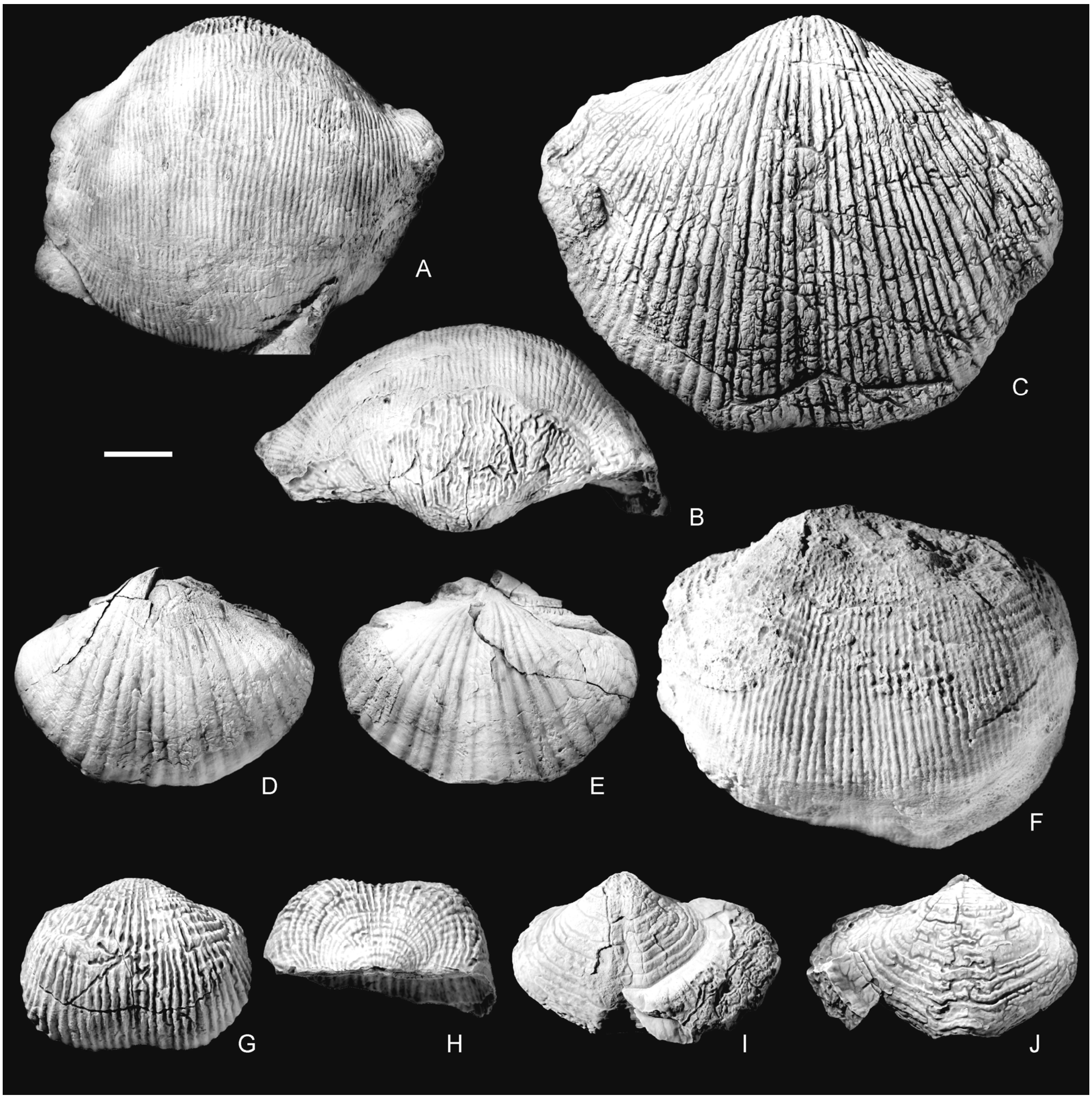

Fig. 5. A-B. *Latiproductus* sp., 5, ‘main belt of Zrigat mounds’ sensu WENDT et al. (2001), mound 20, Zrigat Formation, late Viséan, incomplete ventral valve in ventral and posterior views. **C,** *Spirifer* (*Mesochorispira?)* sp., Hassi Nebech, Merdani Formation, late Viséan, incomplete ventral valve. **D-E,** *Angiospirifer*? sp., mound 100, Zrigat Formation, late Viséan, almost complete specimen in ventral and dorsal views. **F.** *Dictyoclostus* cf. *pinguis* (MUIR-WOOD, 1928), mound 98, Zrigat Formation, late Viséan, ventral valve. **G-H,** *Marginatia* sp., mound 100, Zrigat Formation, late Viséan, almost complete specimen in ventral and posterior views. **I-J.** *Lamellosathyris lamellosa* (LÉVEILLÉ, 1835), mound 100, Zrigat Formation, late Viséan, complete specimen in ventral and dorsal views. Scale bar: 10 mm.

THE EARLY CARBONIFEROUS SUCCESSION IN THE VICINITY OF GARA EL ITIMA

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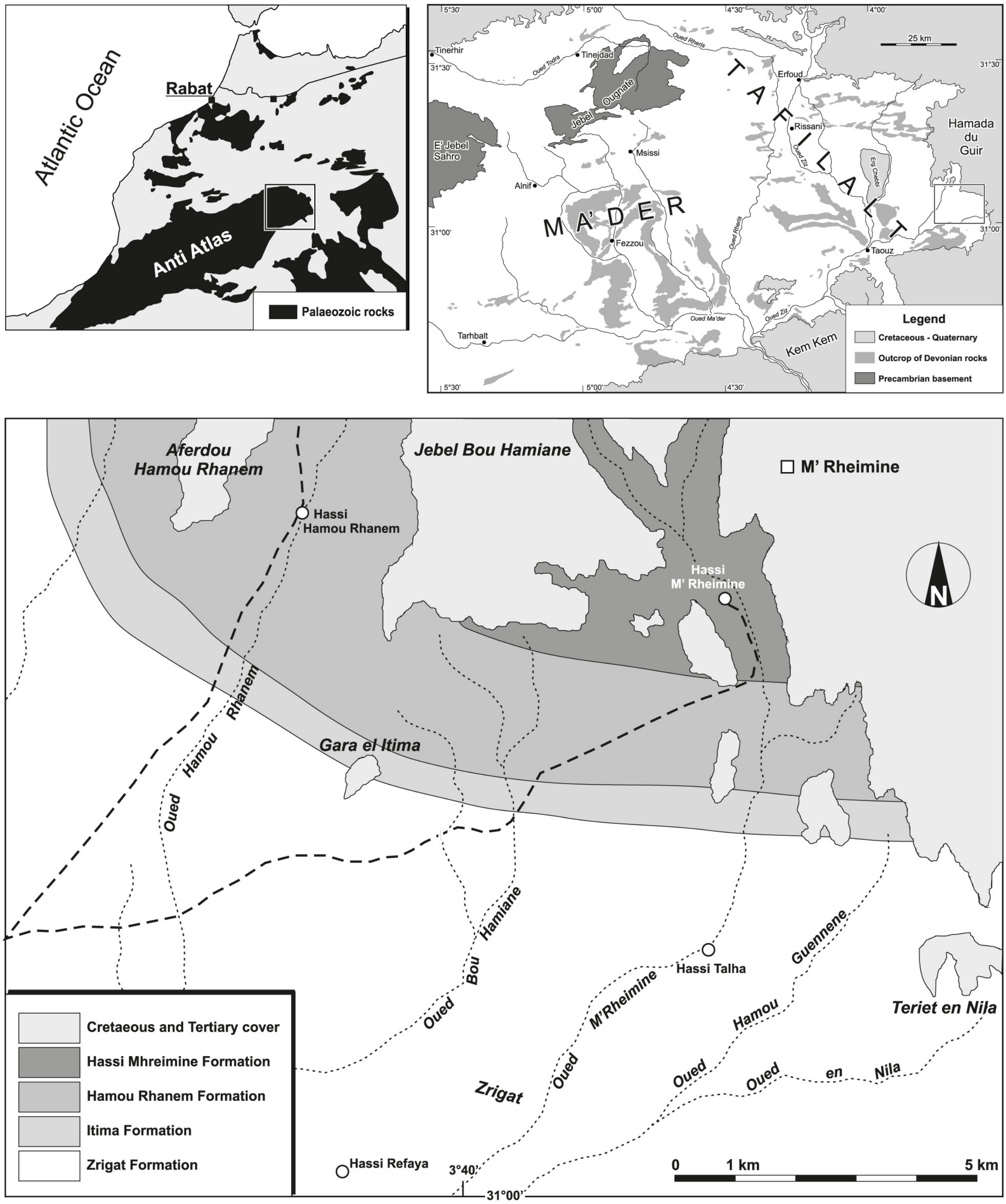

Fig. 1. Geographic position and geological map of the vicinity of the Gara El Itima in the Anti-Atlas (Morocco); after KLUG et al. (2006).

1. INTRODUCTION

The late Palaeozoic successions of Morocco are particularly well known for a rich Devonian fossil content. By contrast, faunas of Carboniferous age have been described much less frequently and their stratigraphic succession is not as well documented. This may be due to the more scattered occurrences of Carboniferous fossil localities, but it may be due to the discontinuous fossil record of marine faunas, which occur within thick siliciclastic rock sequences.

Carboniferous ammonoids from the Anti-Atlas of Morocco have been known for a long time (e.g., DELÉPINE 1941). Since the late 1990s, we studied a number of newly collected ammonoid faunas from various places in Morocco. These investigations were published in a series of articles (KORN et al. 1999, 2002, 2003, 2005, 2007; Bockwinkel & EBBIGHAUSEN 2006, KLUG ET AL. 2006, EBBIGHAUSEN & BOCKWINKEL 2007, KORN & EBBIGHAUSEN 2008), which focused on documenting of the North African ammonoid diversity. By these investigations, the Anti-Atlas of Morocco became one of the key regions for Tournaisian and Viséan ammonoid stratigraphy.

2. GEOGRAPHICAL AND GEOLOGICAL SETTING

The youngest Palaeozoic sedimentary rocks of the eastern Anti-Atlas are exposed in a wide syncline between the mesa of Gara El Itima and the military post Hassi M'Rheimine (Fig. 1) near the Moroccan-Algerian frontier, 33 km ENE of Taouz. The general inclination of the strata is very low and the rocks are mainly composed of siltstone and shales, therefore outcrop conditions are generally poor to moderate. However, there are conspicuous ridges, which allow for a sedimentological and palaeontological investigation and description of the Carboniferous rock succession in this area. The rocks are folded on a kilometre scale with fold axes mostly extending E-W with a slight inclination to the south. Carboniferous strata east of the Gara El Itima strike E-W and dip between 5 and 20° to the N. Near the centre of the syncline at Hassi M'Rheimine, these strata are almost horizontal (compare DESTOMBES & HOLLARD 1986). Undeformed, nearly horizontal Late Cretaceous and Tertiary sedimentary rocks unconformably overlie the Carboniferous rocks. A transgressive conglomerate is usually present above this angular unconformity.

3. SEDIMENTARY SUCCESSION

The Late Viséan (Early Carboniferous) strata in the south-eastern Tafilalt around the mesa of Gara El Itima represent a mixed carbonate-siliciclastic system (Fig. 2). The sedimentary succession of more than 1000 metres thickness with its fauna, trace fossils, and sedimentary structures indicates a mostly slope to base-of-slope environment, except for the uppermost part. For this part a position on the shelf is suggested because of the presence of almost 100 m of well-bedded limestones. Probable lowstands are documented by mixed siliciclastic-calcareous turbidites whereas highstands are mainly characterised by carbonate deposits. Highstand deposits display a distal facies of reworked carbonates or the distal portions of carbonate platforms. The most distal parts of the lower slope and the base-of-slope environments are represented by three thick shale intervals (KLUG et al. 2006).

Zrigat Formation.

Immediately south and south-east of the Gara El Itima, a section of the Zrigat Formation is exposed with a thickness of about 550 metres. The formation is mainly composed of shales with a number of sandstone and limestone intercalations, which form ridges and are thus better exposed for detailed investigations. The sandstone intervals are usually fine-grained and slightly micaceous; they occasionally contain plant remains up to 20 cm in length (*Lepidodendron*). Intervals with carbonates are usually restricted to thin beds of marly or sideritic limestone or small sideritic, fossiliferous nodules. The Zrigat Formation shows a number of transgressive and regressive cycles, which are expressed by the striking cyclicity of siliciclastic and carbonate sediments.

The most conspicuous sedimentary unit is a complex of two carbonatic sub-units, which occurs in the middle of the Zrigat Formation. The lower of the two limestone sub-units is characterised by lateral variations in thickness and composition. Occasionally, it is developed as a crinoid or oolitic bioclastic wackestone, grainstone, or packstone, sometimes with stromatactis cavities. These rocks contain crinoids, corals, sponges, bryozoans, brachiopods, and gastropods as the main biogenic components. These limestone beds contain relatively few marine invertebrates and form isolated carbonate bodies (Fig. 3). These were interpreted as "mudmounds" by Wendt et al. (2001), who mapped and classified more than 100 mud mounds in an area of 400 km2 south of the Gara el Itima (Fig. 4).

The Zrigat Formation hosts at least four ammonoid horizons, the lower three of which are characterised by species of *Goniatites*, and the upper one contains the *Dombarites granofalcatus* assemblage (KORN et al. 1999; KLUG et al. 2006; Fig. 5). Particularly the Zrigat Formation yielded rather rich invertebrate faunas of which, besides the ammonoids, only the gastropods have been described so far (HEIDELBERGER et al. 2009).

Itima and Hamou-Rhanem Formations

The Itima and Hamou-Rhanem Formations consist of 500 m of poorly exposed shales overlying the capping sandstones of the Zrigat Formation. North-east of the Gara El Itima, this thick unit yielded moderately rich ammonoid faunas in three horizons.

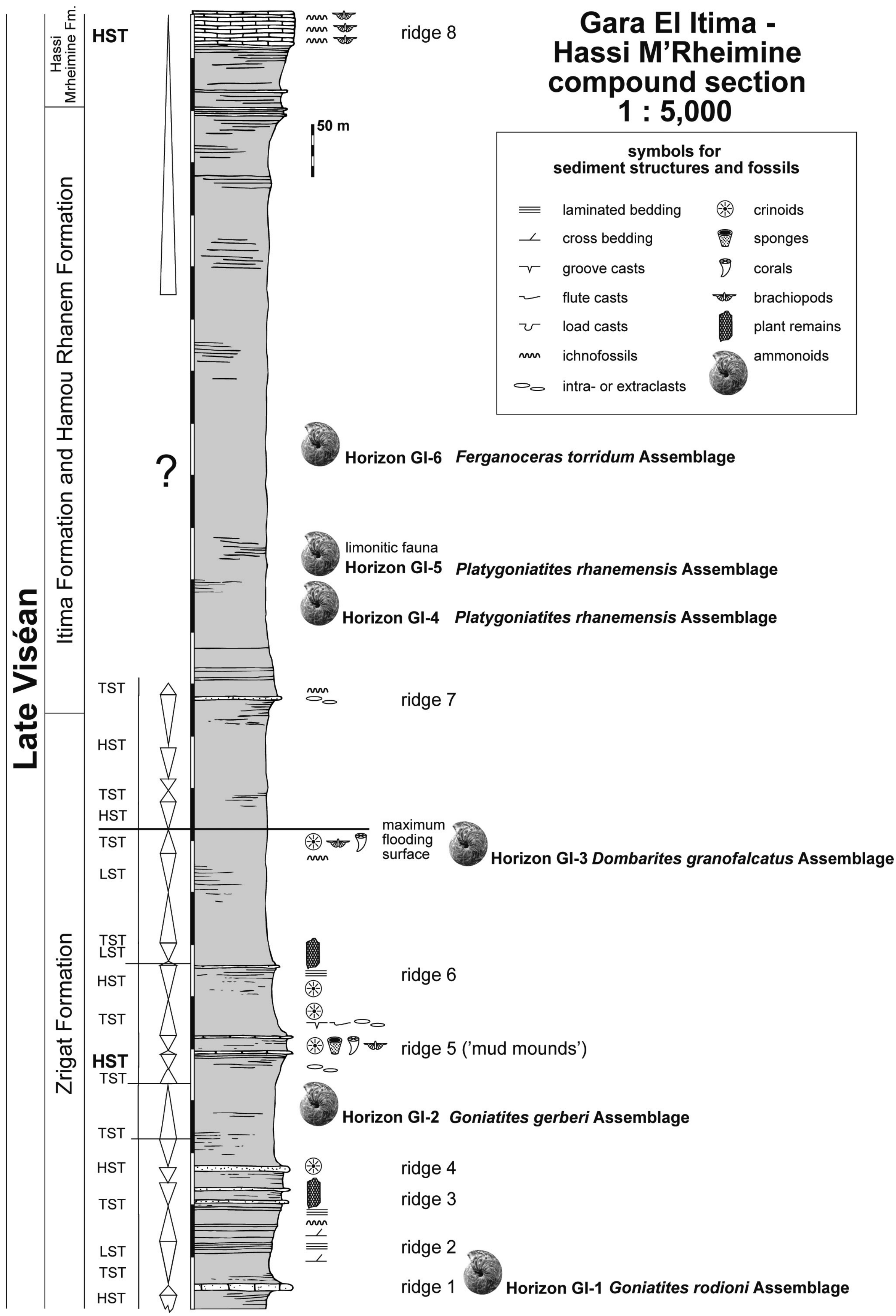

Fig. 2. Compound section of Gara El Itima and Hassi M'Rheimine. Note that this section contains information from various sections East of Gara El Itima; after KLUG et al. (2006).

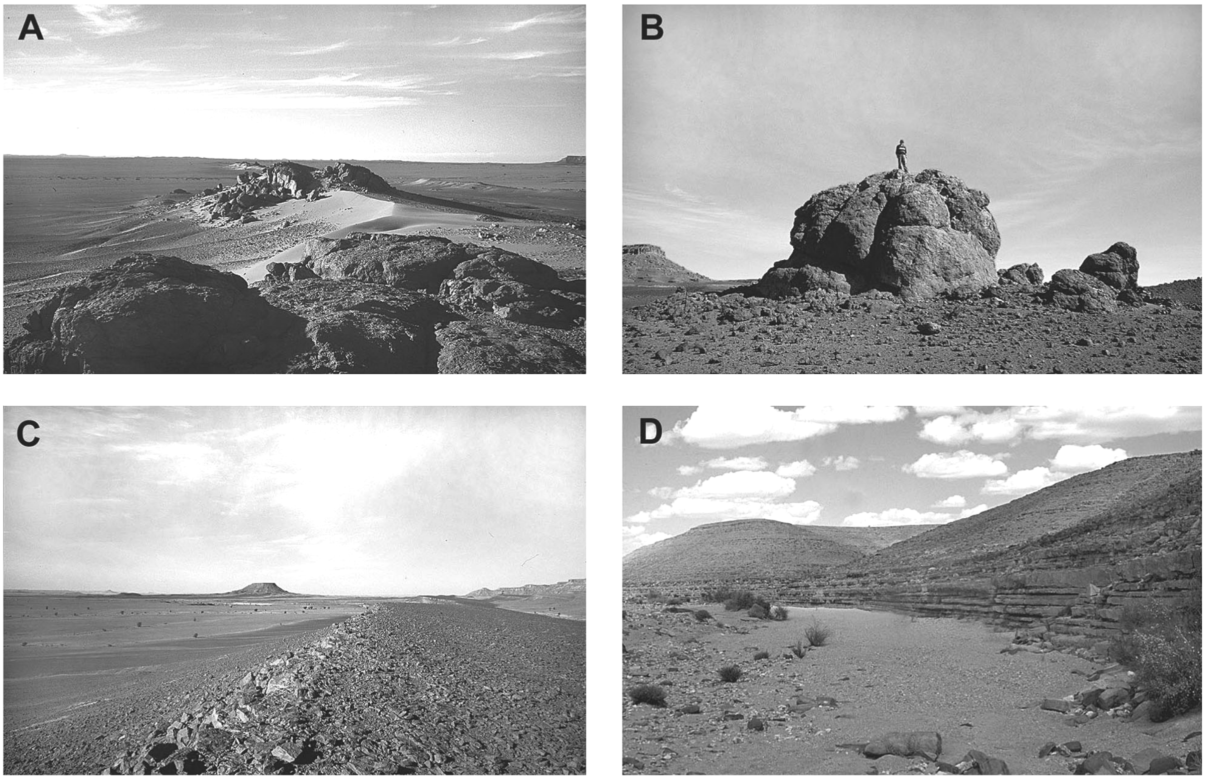

Fig. 3. Outcrop photographs showing some of the investigated rock formations. **A.** Zrigat Formation, lower part with 'mudmound' facies. **B.** Zrigat Formation, higher part with crinoid grainstone facies. **C.** Base of the Hassi Mhreimine Formation with the Gara El Itima (Cretaceous sediments) in the background. **D.** Well-bedded limestone of the Hassi Mhreimine Formation; from KLUG et al. (2006).

The rocks of the Itima and Hamou-Rhanem Formation represent a thick transgressive succession consisting of monotonous shales (ca. 500 m). The water depth obviously increased after the Zrigat Formation, as there are fewer intercalations of coarse clastic sediments. DESTOMBES & HOLLARD (1986) named these intervals the Itima and Hamou Rhanem Formations. Both formations are of Late Viséan age. WENDT et al. (2001) dated these two formations as *nodosa* Zone.

Hassi Mrheimine Formation

The uppermost part of the section lies close to Hassi Mrheimine. There, a monotonous succession of approximately 100 m of a moderately thin-bedded limestone-marl succession is exposed (Fig. 3). This interval contains more clay and siliciclastics at its base. The carbonate content rises significantly in the upper half of the section. The upper 30 m of the formation consist of a series of mudstone layers that are 5 to 50 cm thick and that contain abundant trace fossils (*Planolites*) and occasional spiriferid brachiopods.

Within this interval, abundant brachiopods and some trace fossils were found, indicating a relatively shallow water depth during the deposition of this interval, thus suggesting a regression (late highstand). The Hassi Mrheimine Formation has most probably a Late Viséan or an early Serpukhovian age. Since biostratigraphic data are poor and fossil remains and sedimentary structures are rare, the interpretation of the palaeoenvironment and the definition of the accurate stratigraphic position of this succession remain uncertain.

4. BIOSTRATIGRAPHY

The area of the Gara el Itima is the key region for the Early Carboniferous ammonoid stratigraphy in the Anti-Atlas (KORN et al. 2007). At least seven clearly separable ammonoid horizons have been discovered so far (Fig. 6, 7), ranging from the early Late Viséan *Entogonites-Maxigoniatites* Assemblage to the latest Viséan *Ferganoceras torridum* Assemblage.

The lower four of these contain species of the genus *Goniatites*, with the succession *G. lazarus*, *G. tympanus*, *G. rodioni*, and *G. gerberi*. The latter assemblage has probably two horizons, the lower with *G. gerberi* and the upper with *G. stenumbilicatus*. This is the most detailed succession of the genus *Goniatites* worldwide.

The upper three assemblages are characterized by the genera *Dombarites* and *Platygoniatites*, i.e. genera typical for the latest Viséan ammonoid assemblages of the South Urals (RUZHENCEV & BOGOSLOVSKAYA 1971). These genera allow for a correlation of successions in the Cantabrian Mountains of Spain and the South Urals.

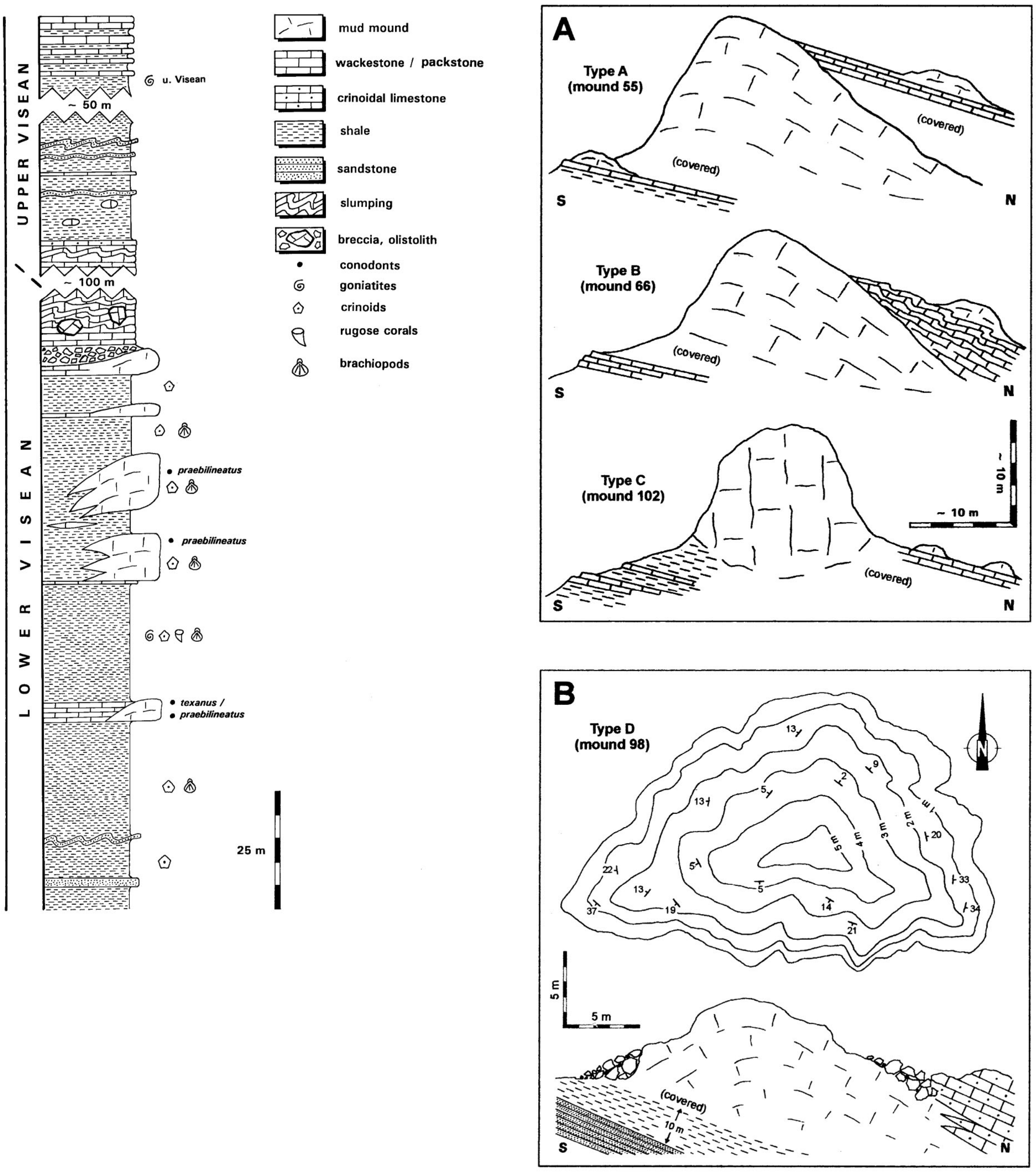

Fig. 4. Left side: Section of part of the Zrigat Formation near mounds 19– 20; right side: **A.** Simplified sketches of type A, B, and C mounds. **B.** Topographic map and profile of mound 98 (type D). Note interfingering and onlapping of crinoidal grainstone and more steeply dipping eastern/western than northern/southern flanks; from WENDT et al. (2001).

5. PALAEOGEOGRAPHY

The ammonoid faunas described from the Gara el Itima have been shown to be new with only a few exceptions (KLUG et al. 2006, KORN et al. 2007). All species were new except for two *Goniatites stenumbilicatus* and *Dombarites granofalcatus*, which had already been described by KULLMANN (1961) from the Cantabrian Mountains of Spain. The occurrence of genera, however, gives a better insight in palaeogeographic relationships between the regions.

The palaeogeographic patterns have been outlined by KORN et al. (2012), who analysed various time intervals including the *Goniatites*-dominated and *Dombarites*-*Platygoniatites*-dominated ssemblages from the Gara el Itima. The analyses led to different results for the two time intervals:

- The *Goniatites*-dominated assemblages show rather similar compositions worldwide; a distinctive pattern can hardly bee seen.
- The *Dombarites*-*Platygoniatites*-dominated assemblages show a very distinctive pattern (Fig. 8), which point at the separation of major realms or faunal provinces (KORN et al. 2012):
- A north-western realm consists of the four North Variscan regions (Rhenish Mountains, North England, South Portugal, and Moroccan Meseta, which form a dense cluster in the scatter plot) as well as the

American Midcontinent and the Antler Foreland Basin.

- The second realm includes the South Urals, the Jadar Block of Serbia, the Cantabrian Mountains of Spain, the Anti-Atlas of Morocco, and the Saoura Valley of Algeria.

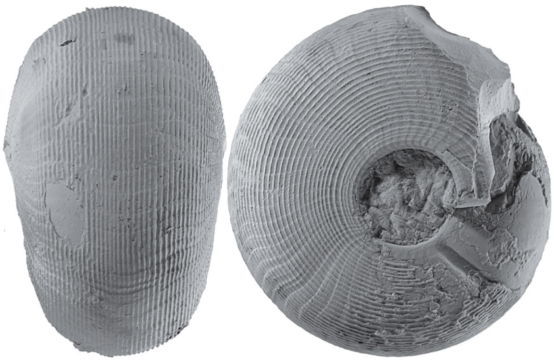

Fig. 5. *Dombarites granofalcatus* from the upper part of the Zrigat Formation (Gara El Itima); from KLUG et al. (2006).

This means that the increasing provincialism during the Late Viséan caused rather sharply separated assemblages. Of particular interest is the comparison of the Late Viséan ammonoid faunas from the Gara el Itima (Anti-Atlas) and the Chebket el Hamra (Jerada Basin). While the latter yielded ammonoid faunas very closely related and, to some degree, even identical with regions of the Rhenohercynian and Subvariscan basins (e.g., South Portuguese Zone, British Isles, Rhenish Mountains), relationships with the Anti-Atlas are very weak.

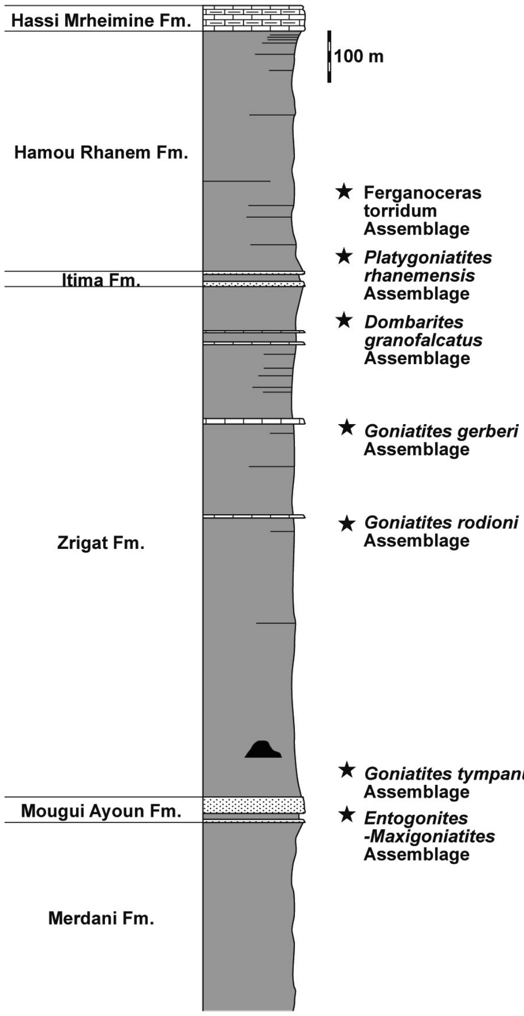

Fig. 6. Generalised columnar section of the Carboniferous succession in the region of the Gara el Itima (Anti-Atlas, Morocco) with the position of the ammonoid assemblages (from KORN et al. 2007).

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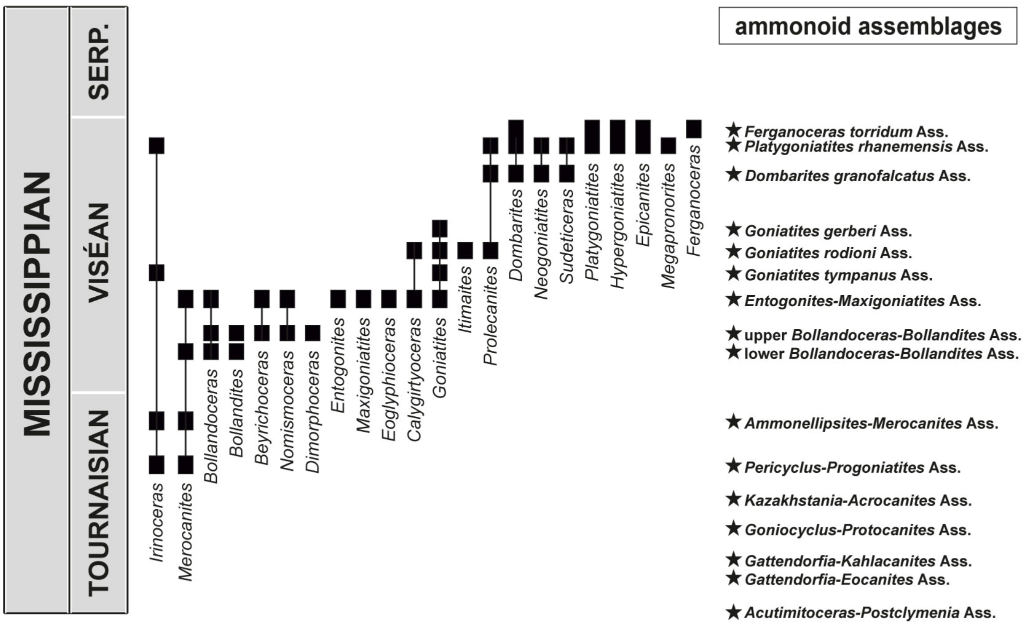

Fig. 7. The stratigraphic distribution of Viséan ammonoid genera in North Africa (from KORN et al. 2007).

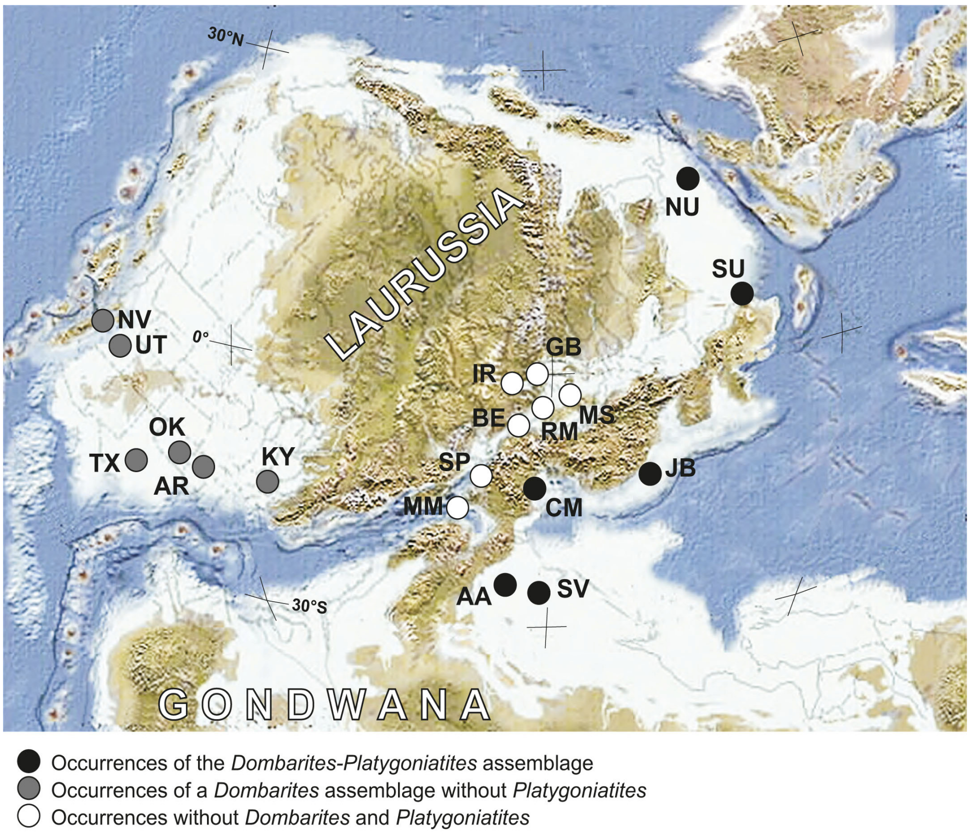

Fig. 8. Palaeogeographic map for the North Atlantic region of the Viséan-Serpukhovian boundary, after SCOTESE 1997; image (modified) by Ron BLAKEY, Flagstaff, Arizona. [NV – Nevada; UT – Utah; TX – Texas; OK – Oklahoma; AR – Arkansas; KY – Kentucky; MM – Moroccan Meseta; SP – South Portugal; IR – Ireland; BE – Belgium; RM – Rhenish Mountains; GB – England; MS – Moravia and Silesia; SU – South Urals; NU – North Urals; CM – Cantabrian Mountains; AA – Anti-Atlas; SV – Saoura Valley; JB – Jadar Block]; after KORN et al. (2012).

CHAOTIC DEPOSITS IN THE LOWER CARBONIFEROUS FORMATIONS OF THE MERZOUGA AREA (TAFILALET, EASTERN ANTI ATLAS, MOROCCO): GEODYNAMIC IMPORTANCE

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1. INTRODUCTION

In the Tafilalet, Merzouga area, Eastern Anti Atlas, Morocco (Fig. 1), the Lower Carboniferous Series is constituted by several sedimentary formations dated as Middle Tournaisian to Upper Visean (CLARIOND 1934, 1944, DELÉPINE 1939, 1941, HOLLARD 1960, DESTOMBES & HOLLARD 1986, WENDT et al. 1991, KLUG et al. 2006, BOCKWINKEL & EBBIGHAUSEN 2006, KORN et al. 2005, 2007, HEIDELBERGER et al. 2009). The synthetic succession consists of seven sedimentary formations, which are each hundred to several hundred meters thick (DESTOMBES & HOLLARD 1986, WENDT et al. 1991; Fig. 2): (i) The Oued Znaïgui Formation (Middle to Upper Tournaisian): shales, conglomerates and sandstone; abundant cone-in-cone structures. (ii) Merdani Formation (Lower Visean): siltstone and fine grained sandstone; cone-in-cone structures; a reefal limestone crops out in the Jebel Begaa area. (iii) Mougui Ayoun Formation (Upper Visean): sandstones, conglomerates and shales; (iv) Zrigat Formation (Upper Visean): shales, sandstone, bioclastic or oolithic limestones and mud mounds (often aligned along an approximate West-East trend Wendt et al. 1991); (v) Itima Formation (Upper Visean): deltaic sandstones and detritic rocks; (vi) Hamou-Rhanem Formation (Upper Visean): shales and fine sandstones; (vii) Hassi Mrheimine Formation (Upper Visean): micritic limestones and shale. The Lower Tournaisian has partly been assigned to the Devonian-Carboniferous Aoufilal Formation (e.g., KAISER et al. 2011).

Recently, during a geological mapping program at the Tafilalet (Taouz and Merzouga areas), new data on the Lower Carboniferous series were obtained: (i) lateral facies changes in the Oued Znaigui Formation; (ii) the presence of chaotic deposits levels in the Oued Znaigui and Moug Ayoun formations. These data suggest a new sedimentary and geodynamic frame for the Lower Carboniferous of the Tafilalet area.

2. LATERAL FACIES CHANGES OF THE OUED ZNAIGUI FORMATION

The Oued Znaïgui formation (Middle-Upper Tournaisian; DESTOMBES & HOLLARD, 1986, WENDT et al. 1991, KORN et al. 2002, KAISER et al. 2011) shows lateral facies differentiations between northern (OuidaneChebbi area, Figs. 1, 3) and southern (Taouz and Mfis area, Fig. 1) regions. At Ouidane Chebbi, the basal Znaïgui Formation, ferruginous, microconglomeratic thin sandstone and shale (≈100m thick) follow, with minor unconformity, on the Aoufilal Formation. In the Mfis area, the ferruginous microconglomeratic levels are more numerous and slightly thicker. In both areas (Ouidane Chebbi and Mfis) undated volcanic sills (basaltic lavas) are set up in parallel to the stratification of the upper part of the Oued Znaïgui Formation.

In the Taouz area, the lower Oued Znaigui Formation shows chaotic deposits, which are unknown at Ouidane Chebbi and Mfis. Three main localities are distinguished (Fig. 1):

North of Jebel Ardane (south of Taouz, Figs. 1, 4a-b): In the lower part of the Oued Znaïgui Formation (Fig. 2) there are chaotic deposits (≈100m thick) formed by metric to decametric blocks of Upper Silurian limestone (with fractured big orthocones, Figs. 4a-b), Lower and Middle Devonian goniatite limestones and some blocks of quartzite and sandstone probably of Ordovician age. These various blocks are packed in a greenish grey slaty matrix. Actually, the contact between chaotic deposits and the quartzitic Ordovician substratum is of tectonic nature, due to Variscan and post-Variscan removements of a major strike slip zone (BAIDER 2007, RADDI et al., 2007, BAIDER et al. 2008).

North of Jdaid (west of Taouz, Figs. 1, 4c): Similar chaotic deposits crop out, with essentially quartzitic (probably Ordovician in age), rounded blocks (2 m at maximum size) and Upper Silurian “*Orthoceras* Limestone” blocks. The matrix is shaly. The contact between chaotic deposits and the quartzitic Ordovician substratum seems to be more of sedimentary nature.

SE of Jebel Begaa (east of Taouz, Fig. 1): Chaotic deposits are formed essentially by rounded limestone blocks (1 to 3 m, containing locally goniatite and orthocone faunas). The relationship between these chaotic deposits and the subjacent (?) dolomitic limestones (Devonian or Carboniferous?) are still unknown.

The Jebel Kfiroun Zone (immediately southwest of Taouz, Fig. 1): The zone is formed by different sized, mixed megablocks of Upper Ordovician (quartzite and shales), Silurian (pelites and limestones), and Devonian (limestones and shales). Additional studies are necessary to know if this zone is formed by an olistostrome belonging to the lower part of the Oued Znaïgui Formation, which is constituted here by shales and thin sandstones.

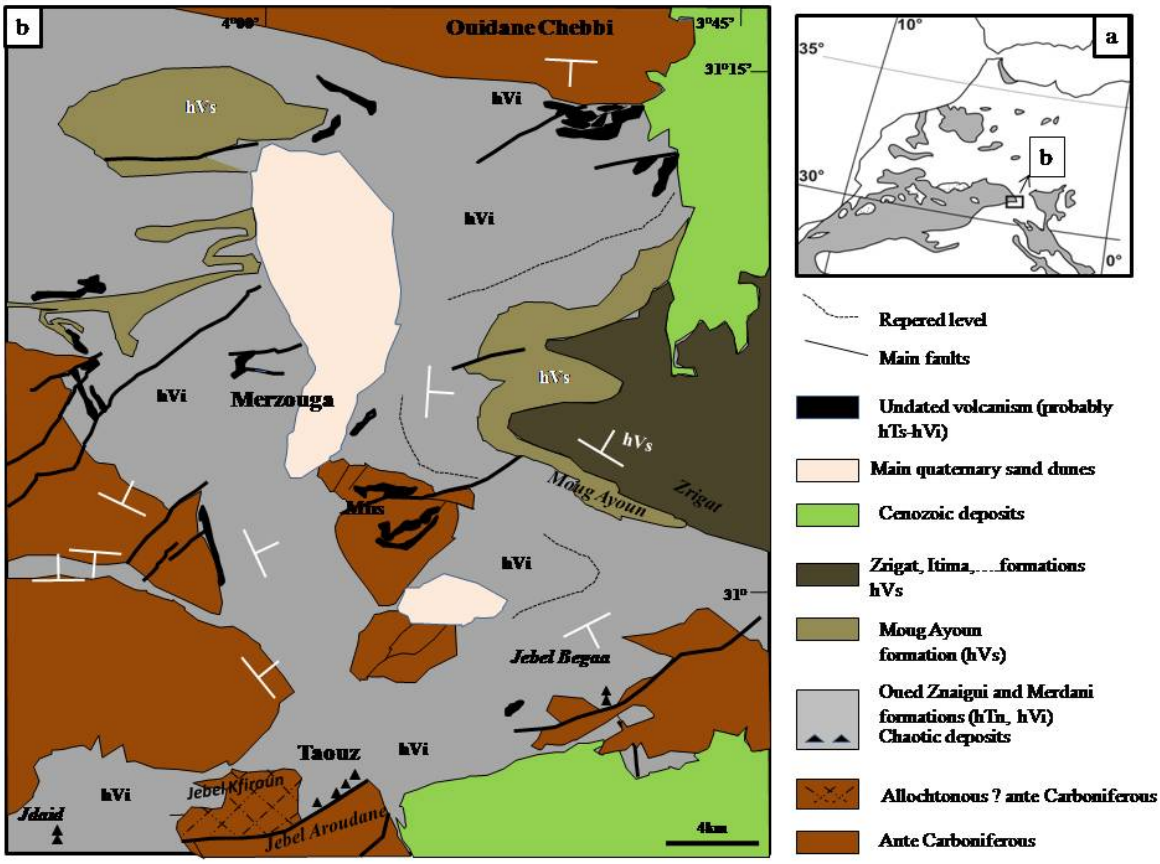

Fig. 1. Sketch map, extract of the geologic map of Morocco, 1/200 000, Sheet of Tafilalet-Taouz, (Notes et Mémoires, n° 244, DESTOMBES & HOLLARD, 1986).▲: Chaotic deposits in the Oued Znaïgui Formation (Tournaisian).

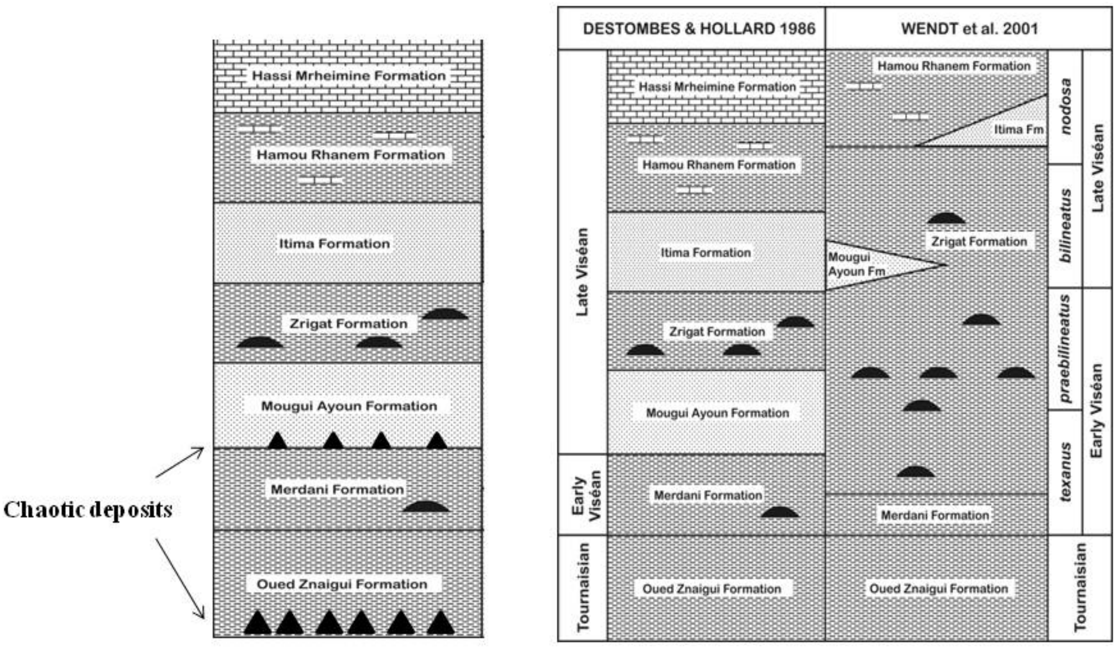

Fig. 2. Emplacement of chaotic deposits in the stratigraphical subdivisions used for the geological map 1:200.000 (DESTOMBES & HOLLARD, 1986) and in WENDT et al. (2001).

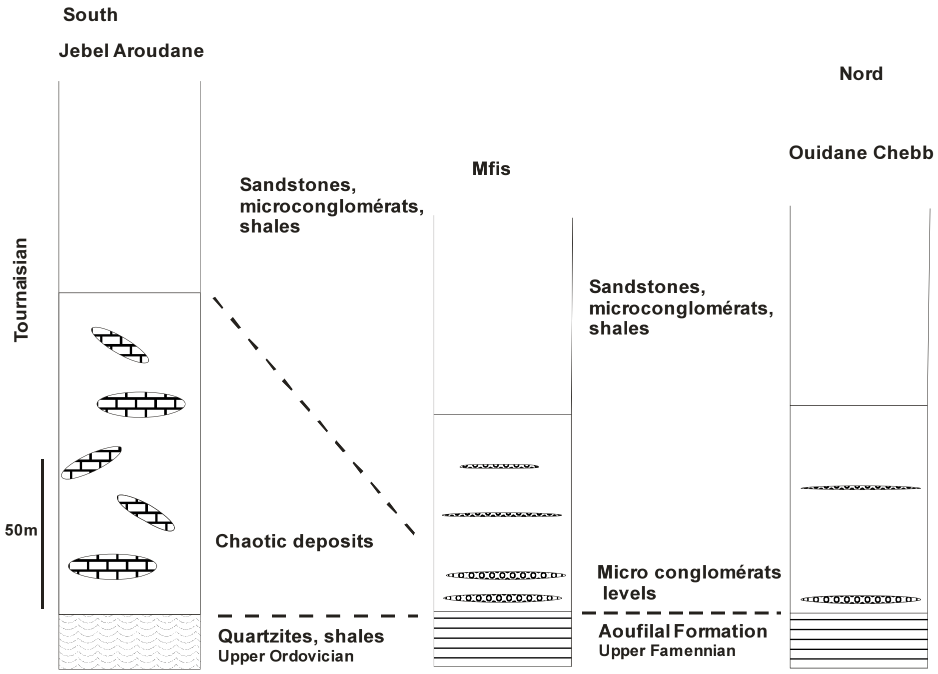

Fig. 3. Lateral facies variations at the base of the Oued Znaïgui Formation.

3. CHAOTIC DEPOSITS IN MOUG AYOUN FORMATION

In all of the Tafilalet area (on both sides of the Merzouga Quaternary sand dunes, Fig. 1), above the essentially clayey Merdani Formation (Lower Visean, DESTOMBES & HOLLARD 1986, WENDT et al. 1991, top Tournaisian to Middle Visean, KORN et al. 2007), the Moug Ayoun Formation (basal Upper Visean, DESTOMBES & HOLLARD 1986, KORN et al. 2007, Fig. 2), starts with a lower level of chaotic deposit (50 m thick), mainly formed by sandstone and quartzite blocks (0,5 -10 m), rare crinoidal limestone blocks (5 m), and small (0,10-0,30 m) “*Orthoceras* Limestone” blocks. The matrix is slaty. Several slumps occur at the base of blocks. An upper level of chaotic deposits (100-120 m) contains fewer and smaller blocks, mainly sandstone and quartzite.

4. DISCUSSION, GEODYNAMIC FRAME

The chaotic deposits of the lower part of the Oued Znaïgui Formation are considered as rock falls deposited in an approximately E-W, tectonically active (Jdaid-Taouz-Begaa) zone (with several Variscan and post-Variscan fault removements), at the border of the ***Merzouga Basin.*** They indicate that the Taouz (Jdaid-Jebel Aroudane-SE of Jebel Begaa) zone was the southern border of the deepening northeastward (north to north-east) basin. This basin was limited to the south by the Jebel Aroudane horsts, emerged or immersed, bordered by faults and with material that fed the chaotic deposits. This is affirmed by the absence of chaotic deposits to the north. Chaotic material came from the dislocation of the neighbouring pre-Carboniferous horst containing upper Neoproterozoic (?) up to Devonian strata (DESTOMBES & HOLLARD 1986). These deposits suggest synsedimentary tectonic activity at the beginning (Tournaisian-Lower Visean) of the genesis of the basin. Essentially there were tilted tectonic block, with northeastward dipping collapse structures (Fig. 5). The responsible type of an extensive (or transtensive?) tectonic regime started in theTafilalet already at the Upper Devonian (WENDT 1985). This active southern part of the Merzouga Basin was used as a corridor for the advance from the east of the Lower Carboniferous transgression via the Bechar Basin (Algeria; KAZI-TANI et al., 1991) to the Maader area (Fezzou Basin, Fig. 6). The synsedimentary tectonic movements continued during the basal Upper Visean, as confirmed by chaotic deposits of the Merdani Formation. However, the “chaotic material” is not varied at this level.

Similar syn-tectonic deposits are also found in some parts of the High Atlas Palaeozoic massifs: Ait Tamelil, Skoura (JENNY & LE MARREC 1980.), Tisdafine Basin (near Tineghir, Fig. 6, SOUALHINE et al. 2003). These types of Carboniferous chaotic deposits were also described farther north in the Meseta Domain (IZART et al. 1997, PIQUÉ 2001, EL KAMEL & EL HASSANI 2006). It means that in this northwestern part of Gondwana, the western advance of the Lower Carboniferous “tethyal” transgression, is synchronous with an extensive regime, which was possible and facilitated by the tectonic opening (remobilizing of old substratum faults) of new sedimentary basins.

Fig. 4. a. View of chaotic deposits of the Oued Znaïgui Formation north of Jebel Aroudane; **b.** Block of Upper Silurian limestone; **c.** chaotic deposits of Oued Znaïgui formation at Jdaid; **d.** chaotic deposits of Moug Ayoun Formation; **e.** sandstone block in the lower level of the chaotic deposits in the Moug Ayoun Formation.

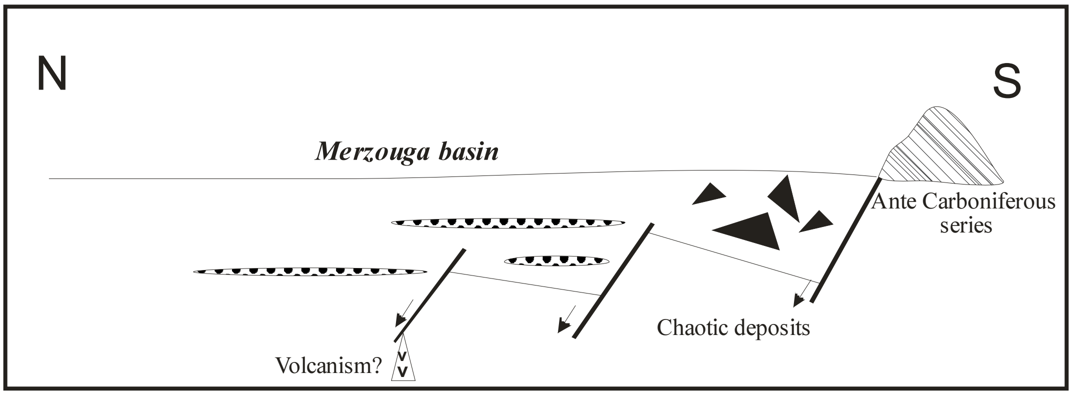

Fig. 5. Sketch of tectonic block movements at the southern margin of the Merzouga Basin.

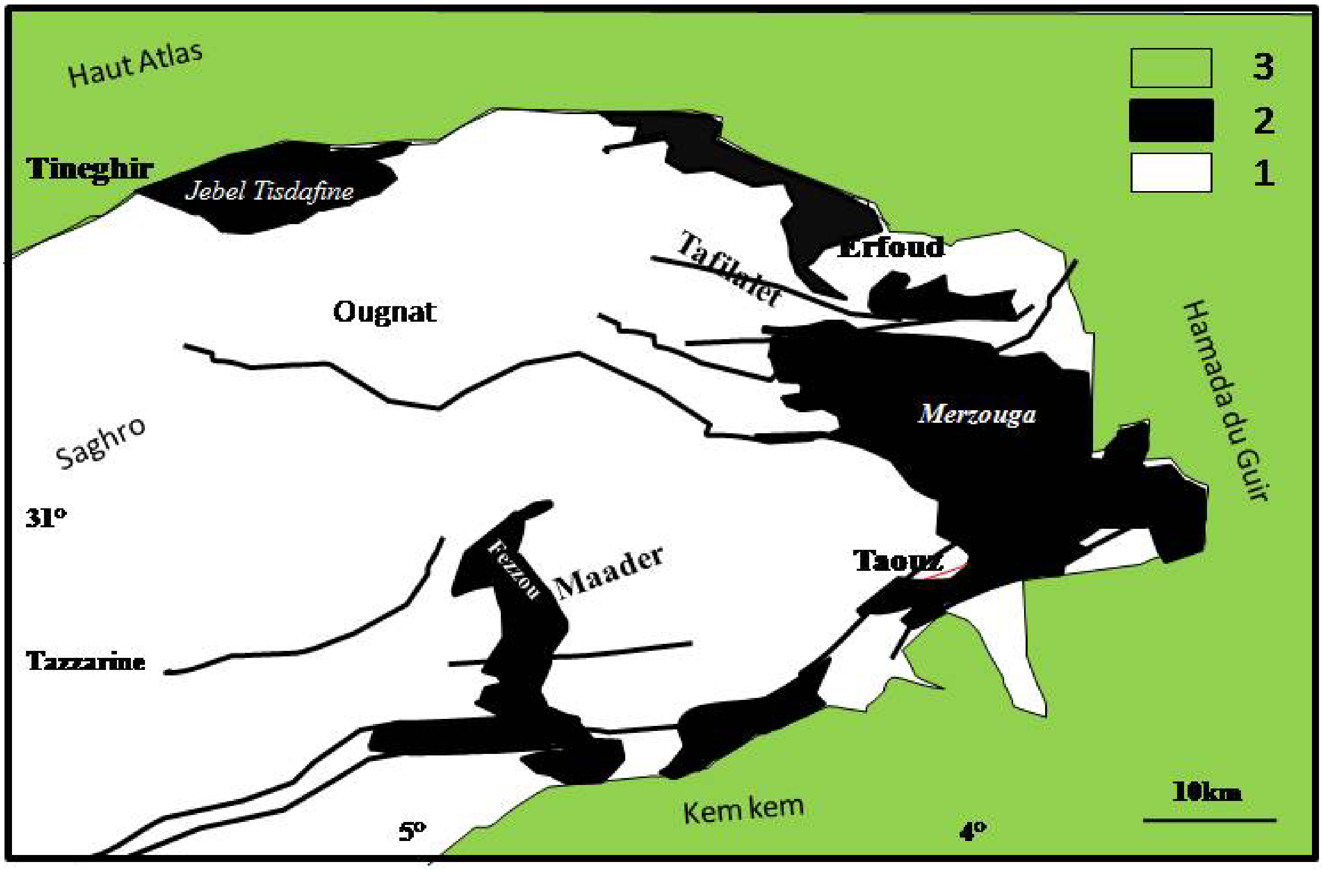

Fig. 6. Sketch map of the Carboniferous in the Eastern Anti Atlas (Tafilalet and Maader).1, Ante Carboniferous; 2, Carboniferous; 3, Post Carboniferous.

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THE DEVONIAN – CARBONIFEROUS BOUNDARY AT LALLA MIMOUNA (NORTHERN MAIDER) – A PROGRESS REPORT

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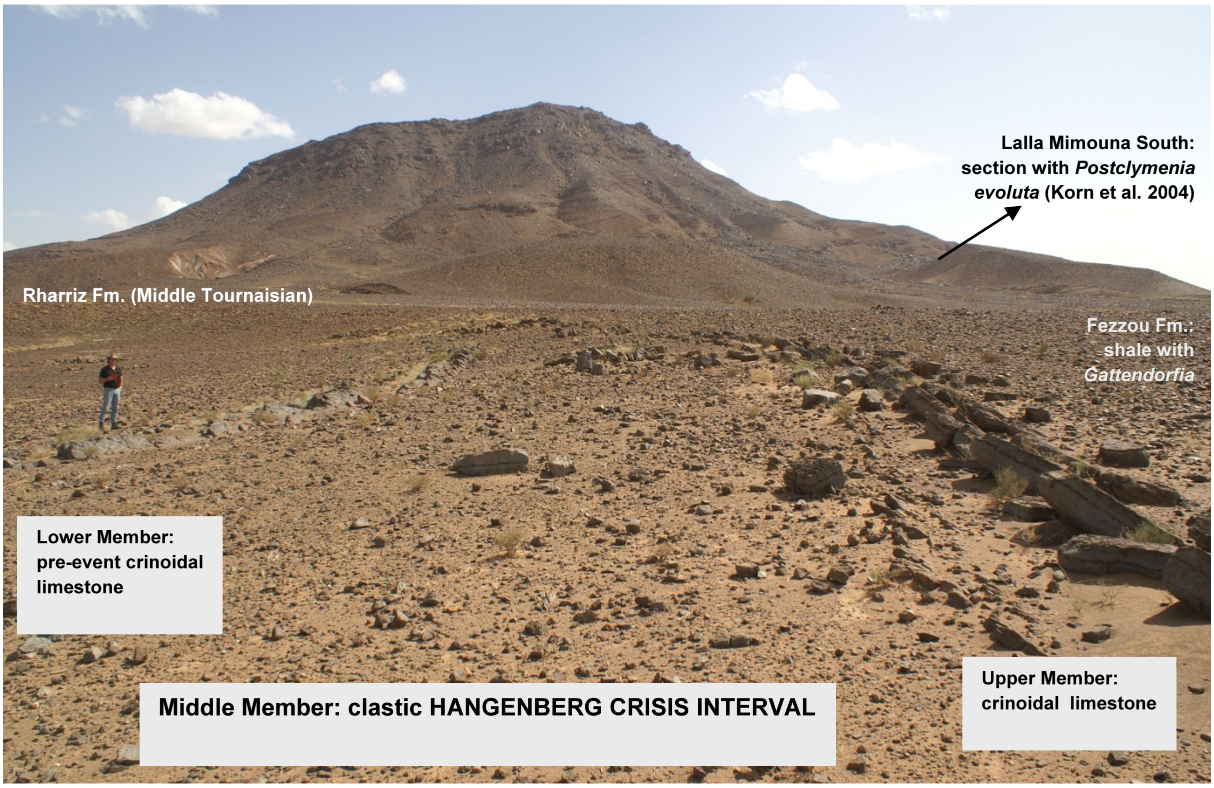

Fig. 1. Overview of the D/C boundary section at Lalla Mimouna North (view to the south, main mountain built by Ordovician sandstones) with the three members of the Lalla Mimouna Formation (new), overlain by goniatite shales of the Fezzou Formation (left margin) and the Middle Tournaisian Rharriz Formation (background). The position of the southern section of KORN et al. (2004), hidden behind a gravel hill, is also marked.

1. INTRODUCTION

The re-sampling of the Devonian/Carboniferous stratotype at La Serre, Trench C (Montagne Noire), by KAISER (2009) confirmed the view of ZIEGLER & SANDBERG (1996) that siphonodellids with the morphological characteristics used to place the base of the *Si. sulcata* Zone at the base of Bed 89 occur much lower down in the succession, near the base of the Upper Oolite (within Bed 84). This rejected the idea of a gradual morphological (FLAJS & FEIST 1988) or even phylogenetic transition within a *Si. praesulcata-sulcata* lineage in the stratotype. Since there is no other criterion, a different marker fossil or geochemical signal, that allows us to correlate the base of Bed 89 with precision into any other known boundary section, it was decided that the La Serre GSSP and base of the Carboniferous need to be revised. The consequently installed D/C Boundary Task Group is searching globally for new boundary sections. A main goal is to increase our knowledge of the stratigraphy, sedimentology, biodiversity and geochemical changes around the global Hangenberg Crisis. The future GSSP level is completely open for discussion. Improvements in conodont taxonomy (e.g., CORRADINI et al. 2011, KAISER & CORRADINI 2011) play a pivotal role in this revision process, especially since the lost type of the current basal Carboniferous index, *S. sulcata*, is from the upper part of the Lower Tournaisian (EVANS et al. 2013) and morphologically not close to the siphonodellids that have been used to define the problematical current GSSP.

The SE Anti-Atlas, especially the southern Maider, southern and eastern Tafilalt, include a range of important D/C boundary sections with different facies and faunas that were deposited in the Maider Basin (e.g., Aguelmous Syncline), on the Tafilalt Platform (Amessoui Syncline), and in the Tafilalt Basin (Ouidane Chebbi-Mfis region). These have recently been described and correlated with the German (Rhenish) succession by KAISER et al. (2011). Ammonoids, event and sequence stratigraphy provide the main regional time framework. Conodonts are sparse in the pre-Hangenberg Event nodular limestones of most sections and only few Lower Tournaisian specimens could be obtained since limestone deposition almost ceased with the main Hangenberg Event until late in the Upper Tournaisian. In addition, the upper part of the pre-event beds (at least the *Wocklumeria* Zone, UD VI-D) is missing in an unconformity just below the Hangenberg Black Shale equivalents all over the Tafilalt.

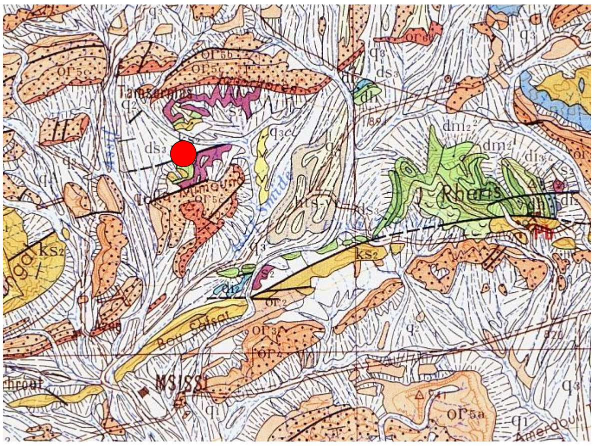

Fig. 2 . Position of Lalla Mimouna North (red dot) N of Msissi (geological map, sheet Todrha-Ma´der).

Based on a detailed survey of the Jebel Rheris area at the northern margin of the Maider (FRÖHLICH 2004), KORN et al. (2004) described an important locality (their Section A, here named as Lalla Mimouna South) from the northern slope of the Lalla Mimouna Mountain N of Msissi (Figs. 1-3). Vertically-bedded, dark, recrystallized limestone of the topmost Devonian *kockeli* Zone yielded an association of *Postclymenia evoluta* and *Acutimitoceras (Stockumites) hilarum*, which correlates with the Lower Stockum Limestone of Germany. Stable isotope data record a very large negative spike ($\delta^{13}C$ = -17 $^{o}/_{oo}$) in the goniatite bed, which is explained by extensive, early diagenetic recycling of organic matter.

Our field work concentrated since 2009 on section Lalla Mimouna North (Figs. 1, 3-4), which is identical with Section B of KORN et al. (2004). The GPS coordinates for the main section are N 31°16.502´ W 4°49.092´. A lateral section was measured just 30 m to the south because it has a better exposure of some upper beds and since some beds change thickness at short distance. Preliminary accounts of both were provided by BECKER et al. (2011, 2012) in the Newsletters of the Carboniferous and Devonian Subcommissions. Based on new samples from spring

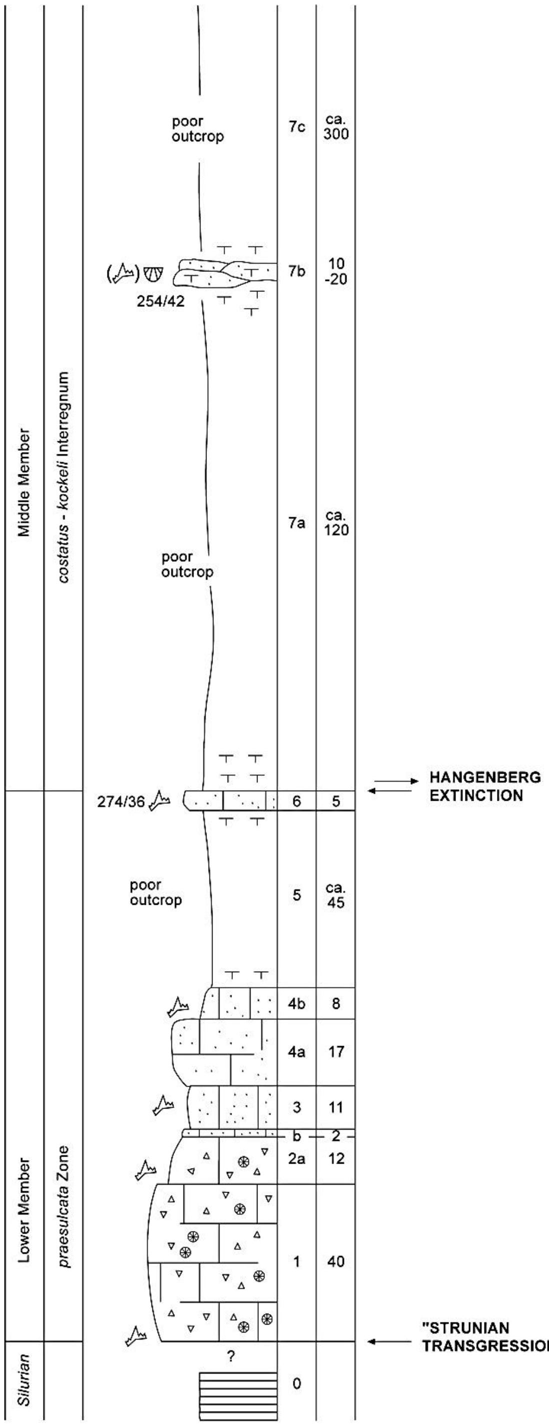

Fig. 3. Lithological log, position of conodont samples and events in the Lower and lower part of the Middler Member of the Lalla Mimouna Formation,

2012, this progress report includes new ammonoid, brachiopod, and conodont data, which result in significant changes of biostratigraphical dating and correlation. The main advantages of the section are:

1. It is the only known Anti-Atlas section with rich conodonts from the uppermost Famennian, right to the main Hangenberg Extinction level.
2. Fossiliferous Hangenberg Event clastics allow a correlation with the North African D/C boundary brachiopod succession (e.g., BRICE et al. 2005, 2007).

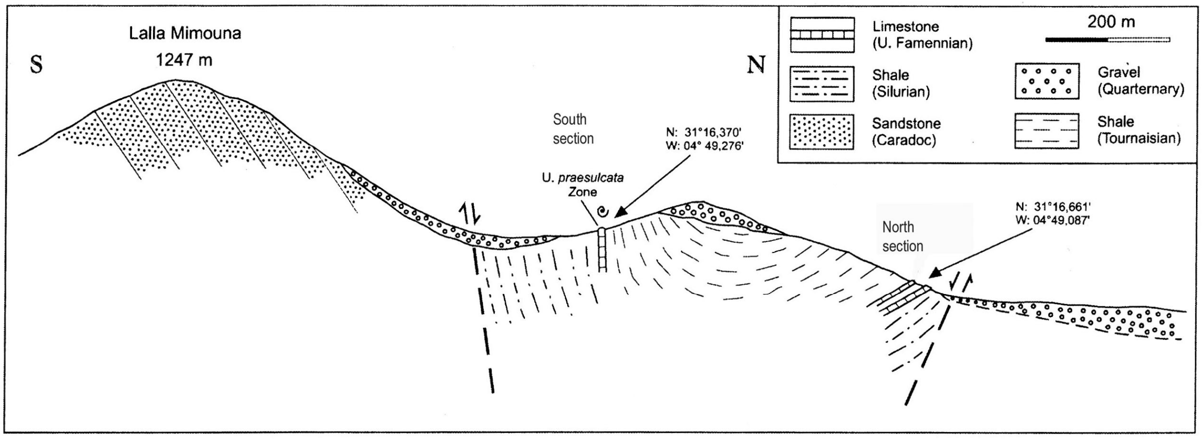

Fig. 4. Cross-section through the northern slope of Lalla Mimouna after FRÖHLICH (2004: fig. 20).

3. Conodont faunas include a wealth of intermediates between *Polygnathus* and early siphonodellids (TRAGELEHN 2010 and in prep.), including two new genera, *Si. (Eosiphonodella)* JI (1985), and rare, true *Si. (Siphonodella)*. This morphological and taxonomic complexity strongly pleads against the simplistic use of a supposed *Si. praesulcata-Si. sulcata* lineage for chronostratigraphic definition.

4. New protognathodid and polygnathid records suggest that the upper event interval (*costatus-kockeli* Interregnum of KAISER et al. 2009 = upper part of Middle *praesulcata* Zone) is represented by limestone with sufficient conodonts.

5. The section contains the only abundant Lower Tournaisian conodonts of the Anti-Atlas.

6. Goniatite shales provide a correlation with the diverse Lower Tournaisian goniatite faunas of the southern Maider (EBBIGHAUSEN & BOCKWINKEL 2007).

2. SEDIMENTS AND FAUNAS

The Lalla Mimouna Mountain is formed by resistant Ordovician siliciclastics. On the northern slope they are supposedly overlain by Silurian shales (Fig. 3) but new records of *Protocanites* sp. (Fig. 11) and nautiloids confirm that a part of the shales belong in fact to the Middle Tournaisian Rharriz Formation. The Silurian age of the shales below the studied D/C boundary succession is not proven biostratigraphical data. On the other hand there is no evidence for any pre-uppermost Famennian (Devonian) sediments in the Lalla Mimouna region.

The examined uppermost Famennian and Lower Tournaisian crinoidal limestones are so different from contemporaneous strata of the main part of the Maider that they are assigned to a new **Lalla Mimouna Formation**, which is ca. 7 m thick in its presented type locality, Lalla Mimouna North. The (pre-Hangenberg) Lower Crinoidal Limestone of BECKER et al. (2011, 2012) forms a Lower Member (ca. 1.4 m thick, Fig. 3), the intercalated, brachiopod-rich, clastic tongue of the lower Fezzou Formation a Middle Member, and the ca. 1.3 m thick post-event Upper Crinoidal Limestone an Upper Member.

2.1. Lower Member

Beds 1-4b (Figs. 3, 5) consist of coarse-grained, light-grey to reddish-brown (4a-b) crinoidal limestone without other macrofauna, which become thinner and sandier upwards. They represent crinoid forests which grew on a shallow, nearshore neritic carbonate platform around the Lalla Mimouna Island (FRÖHLICH 2004). Storms caused a disarticulation and angular fragmentation during lateral transport. Thin-sections show a strong recrystallization of crinoid pack- to grainstones, with many angular quartz grains and brachiopod and trilobite fragments in some layers. The base of the unit is sharp, without reworking of older strata. Above a poorly exposed silty/marly interval (Bed 5), the sandy, reddish last pre-Hangenberg limestone (Bed 6, Fig. 6) continues the microfacies and recrystallization from below. Abundant brachiopod fragments are present.

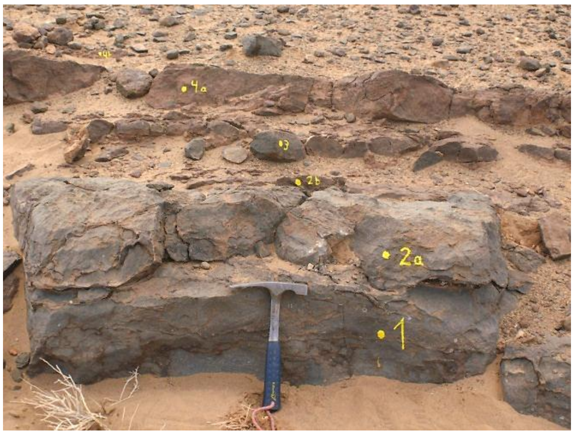

Fig. 5. Main part of the Lower Member of the type Lalla Mimouna Formation (Beds 1-4b).

2.2. Middle Member

Bed 6 is sharply overlain by a deeply weathered succession of silty shale/marl with intercalations of thin-bedded, greenish-grey, bioturbated, calcareous brachiopod-rich siltstones (Bed 7b, Fig. 6). Processing for conodonts was not successful. The brachiopod assemblage consists mostly of rhynchonellids. A preliminary account of taxa (det. D. BRICE) includes ?*Centrorhynchus* sp. (two morphotypes; Figs. 10.1-2), cf. *Sedenticellula* sp. 1-2 (see Fig. 10.3), ?*Hemiplethorhynchus* sp. (Fig. 10.4), ?*Hadyrhyncha* sp. (Fig. 10.6), rare ?*Araratella* sp., and orthids (Fig.10.5). *Araratella* is wide-spread but restricted to the uppermost Famennian (SARTENAER & PLODOWSKI 2003). *Centrorhynchus* (see BRICE et al. 2007) and *Hadyrhyncha* (see SARTENAER 1998) appear earlier in the middle and upper Famennian but are not known from the Carboniferous. *Sedenticellula* ranges from the Hangenberg Event Interval of La Serre (LEGRAND-BLAIN & MARTÍNEZ CHACÓN 1988) at least to the Middle Tournaisian. *Hemiplethorhynchus* is normally a Tournaisian genus but it has been recorded from sandstones near the Devonian-Carboniferous boundary at El Atrous (BRICE et al. 2007). In summary, the brachiopod assemblage of the Middle Member is in accord with a position at the top of the Devonian (despite the poor preservation).

Fig. 6. The deeply weathered main Hangenberg Crisis Interval (Middle Member) starting above Bed 6; small exposure of Bed 7b near the end of the measuring stick; background = Upper Member.

The Middle Member represents the glacially controlled global Hangenberg Regression/Lowstand (see review in KAISER et al. 2011). It correlates with the much thicker lower part of the Fezzou Formation in the Aguelmous Syncline to the south. Despite some digging at the base, we could not find any evidence for an equivalent of the Hangenberg Black Shale. The anoxic and transgressive initial part of the Hangenberg Crisis may have been removed during the subsequent sudden regression, as it has been observed locally at Lambidia (NW Aguelmous Syncline, KAISER et al. 2011). In the upper part of the Middle Member there is a distinctive, poorly lithified, coarse crinoid sand (Bed 7f, Fig. 8). The laminated to cross-bedded, calcareous siltstone below (Bed 7d) yielded no conodonts and similar brachiopods as Bed 7b, notably cf. *Sedenticellula* sp. and ?*Centrorhynchus* sp. Beds 7i-j (Fig. 8) form a calcareous transition to the Upper Member. This suggests a gradual, slow sea-level rise in the upper event interval.

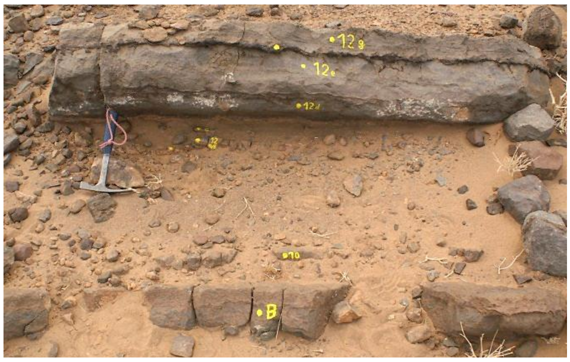

Fig. 7. The three subunits of the Upper Member of the type Lalla Mimouna Formation (main section).

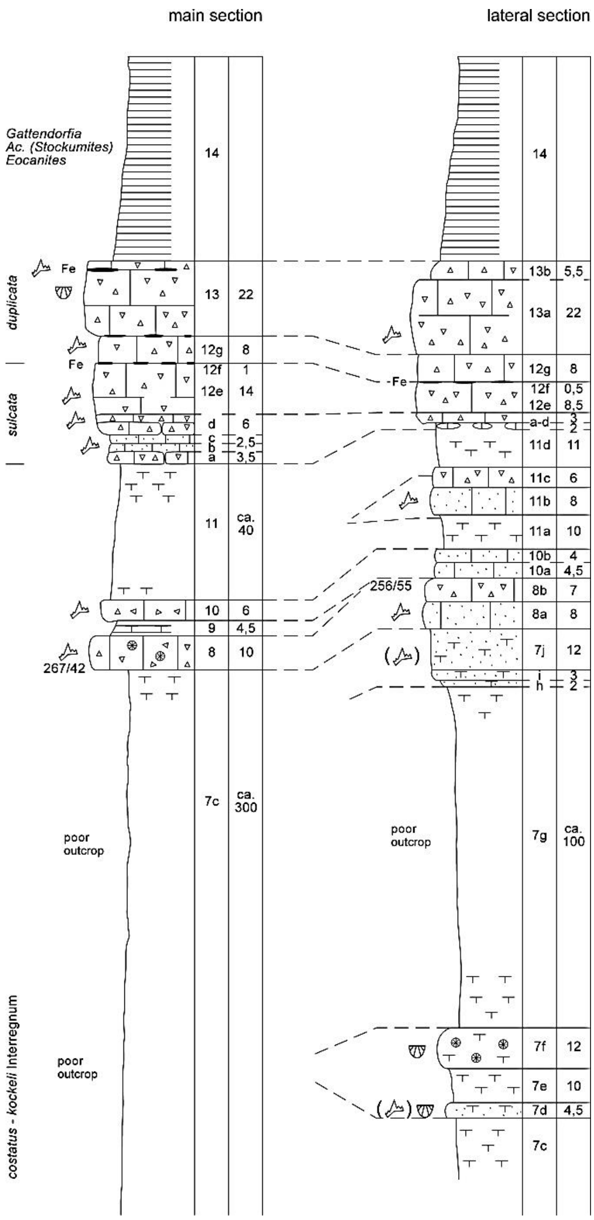

Fig. 8. Correlation of the upper part of the main and of the lateral section, ca. 30 m to the south, showing the wedging out of some crinoidal debris beds, and the gradual transition from the Middle to the Upper Member of the Lalla Mimouna Formation.

2.3. Upper Member

The Upper Member (Upper Crinoidal Limestone in BECKER et al. 2011, 2012) consists of three subunits (Fig. 7), which show a re-establishment of the shallow, near-shore carbonate platform. Beds 8-10 are rather sandy, crinoid pack-rudstones with intercalated, laminated to cross-bedded siltstone layers. There are coarse-grained and finer, partly strongly recrystallized beds, which wedge out at short distance (Fig. 8). In the main section there are layers with many brachiopod fragments and washed out matrix. The mostly angular crinoid fragments have micritized margins. It is clear that the bed is composed of several depositional events.

The middle subunit (Bed 11) is a deeply weathered calcareous siltstone in the main section, with intercalated, sandy or silty crinoid pack-rudstones in the lateral section (Beds 11b-c). Bed 11b contains siltstone extraclasts and ooids. This gives an interesting parallel to the topmost Devonian (upper Hangenberg Event Interval) oolites of La Serre and the Rhenish Massive (e.g., KOCH et al. 1970).

The upper subunit (Beds 12-13) is a sequence of solid, partly recrystallized crinoid pack-rudstones with a much reduced silt/sand content, apart from some discrete siltstone layers. The micrite matrix is variably light-grey, darker or washed out. The reduced clastic influx suggests a slight deepening. Below Bed 12g, Bed 13a, and Bed 13b there are iron crusts that formed during minor sedimentation breaks. There are some brachiopods and possibly fine, dark brownish phosphate clasts in the matrix.

2.4. Fezzou Formation

The last crinoidal packstone is sharply overlain by deeply weathered, thick greenish silty shales with rare, small, mostly poorly preserved goniatites and other fauna in the lower part. There are two new species of *Gattendorfia* (Figs. 10.1-2), the first Moroccan *Eocanites* of the *supradevonicus* Group, *Kazakhstania nitida*, *Acutimitoceras (Stockumites)* aff. *endoserpens* (Fig. 10.4), *Acut. (Stock.) ?intermedium* (juv.), *“Imitoceras”* n. sp. (Fig. 10.3), nuculoids, crinoid stem pieces, a new bivalve genus (somewhat homoemorphic to the Fammenian *Loxopteria*, e.g., NAGEL-MYERS et al. 2009), orthocones, and rare gastropods (Fig. 10.5). The *supradevonicus* Group is restricted in the Rhenish Massive (VÖHRINGER 1960) to the upper part of the Lower Tournaisian (*Gatt. crassa* Zone, LC I-C/D of BECKER 1996). This, and the presence of *Kaz. nitida*, enable a correlation of the Lalla Mimouna fauna with the main goniatite level of the Aguelmous Syncline (EBBIGHAUSEN & BOCKWINKEL 2007, with recent age comments by BECKER in HAHN et al. 2012). However, the differences between the goniatite assemblages of the two neighbouring regions suggest a strong facies influence (deeper and offshore in the Aguelmous Syncline) on the faunal composition. The revised age for the local basal Fezzou Formation is in accord with the new conodont data (see below). The significant transgression and reduced sea floor oxygenation in the upper half of the Lower Tournaisian does not correlate with an established eustatic pulse but it was also noted in the stratotype section (FLAJS & FEIST 1988).

2.5. Rharriz Formation

Even higher, to the S on the slope, and separated by a long outcrop gap, there are deeply weathered shales with few brownish sideritic nodules. These contain the Middle Tournaisian index goniatite *Protocanites* sp. (Fig. 11) and orthocones.

3. CONODONT STRATIGRAPHY

Bed 1 yielded both morphotypes of the index species of the *Bispathodus ultimus* Zone (= previous Upper *expansa* Zone), which is associated with longer-ranging uppermost Famennian taxa, such as *Bi. aculeatus aculeatus* (Pl. 3, Fig. 15), *Bi. aculeatus anteposicornis, Bi. costatus* (both morphotypes), *Bi. spinulicostatus* M1*, Bi. stabilis vulgaris, Mehlina strigosa, Branmehla inornata,* and *Neopolygnathus communis*. The absence of *Palmatolepis* is intriguing but can be explained by the peculiar, shallow litho- and biofacies. Most characteristic are intermediates between polygnathids and siphonodellids. There are two clearly separate groups, which represent two different lineages. Forms assigned to N. Gen. I resemble the *Po. symmetricus* Group, especially in the shape and ornamentation of the platform, but have started to invert the aboral surface. The developing pseudokeel is still incomplete and not yet flat, especially not under the posterior platform. *Po. spicatus* may belong to this group. N. Gen. 2 has a fully developed flat pseudokeel that is constricted under the central platform but it continues to the posterior tip. The (anterior) platform shape remains polygnathid not pseudopolygnathid, in having a sharply delimited anterior platform margin with marked, projecting nodose shoulders. Some forms, especially those with wide cavity (e.g,, the medium-sized specimen of Pl. 1, Fig. 1), resemble *Pseudopolygnathus* (e.g., *“Ps.” graulichi*) but the anterior platform is different. In Bed 1 there are one new species of N. Gen. 1 (n. sp. A, Pl. 1, Fig. 5) and four different morphs/taxa of N. Gen. II (Pl. 1, Figs. 1-4). One specimen (Pl. 1, Fig. 6) has to be assigned to a new species of *Si. (Eosiphonodella).* It is in fact disturbingly close to *Si. (Eo.) sulcata* (s.l., Group 7 of KAISER & CORRADINI 2011) if the minor anterior protrusion of the outer platform is neglected. But it is not really an unusual record since there are more specimens of the same form in the pre-Hangenberg uppermost Famennian of Franconia (TRAGELEHN in prep.). The entry of any *Siphonodella* is usually taken to place strata in the *praesulcata* Zone but *Si. (Eo.) praesulcata* does not occur in the pre-Hangenberg limestones of Lalla Mimouna.

The conodont assemblage of Bed 2a is very similar but there are different representatives of N. Gen. II (Pl. 1, Figs. 7-8), which are very close to the Tournaisian *“Ps.” scitulus* JI et al. (1985). A second form of N. Gen. I (n. sp. B) possesses an almost complete pseudokeel and is transitional to *Si. (Eosiphonodella)* n. sp. A (of Pl. 1, Fig. 6; compare lower surface of "*Polygnathus* sp. A" in WANG & YIN 1988). Representatives of associated *Bi. ultimus* (both morphs, Pl. 3, Figs. 20-22) and *Bi. spinulicostatus* M1 (Pl. 3, Fig. 17) are illustrated.

Beds 3 and 4a have a much more restricted conodont fauna that consists only of bispathodids. This faunal change correlates with the shift to reddish limestone. The last pre-event limestone, Bed 6, suddenly contains *Pa. gracilis gracilis* (Pl. 3, Fig. 13). N. Gen. 1 and 2 re-appear together with *Bi. stabilis vulgaris* and *Bi. aculeatus acleatus. Bi. costatus* M2 is illustrated on Pl. 3 (Fig. 18). There are one N. Gen. 2 (Pl. 1, Fig. 10, similar to Fig. 2) and two new forms (n. sp. C-D) in N. Gen I (Pl. 1, Figs. 11-12). The pseudokeel is almost complete in two specimens (Pl. 1, Figs. 13-14), which may be placed as a second new species (n. sp. B) in *Si. (Eosiphonodella).* They do not conform with any of the siphonodellid groups defined in KAISER & CORRADINI (2011).

The lower subunit of the Upper Member (Bed 8a) begins with a fauna that is still rather similar to the Lower Member. However, the sudden entry of *Protoganthodus meischneri* (Pl. 3, Fig. 5) and *Pr. collinsoni* is characteristic for the *costatus-kockeli* Interregnum (former Middle *praesulcata* Zone). *Po.* cf. *purus purus* (Pl. 3, Fig. 10; basal pit situated more anterior than in typical specimens) and first *Ps*. aff. *primus* indicate the beginning of the post-event recovery phase. Last specimens of *Pa. gracilis gracilis* (Pl. 3, Fig. 14) may have been reworked because there are also rare *Bi. ultimus* M2 and *Bi. costatus* M2 (Pl. 3, Fig. 19), which normally do not range above the main Hangenberg Event (see discussion in KAISER et al. 2009). The high-energy deposition and the strong reworking of crinoid clasts make conodont reworking likely. There are also *Bi. aculaeatus aculeatus, Bi. aculeatus anteposicornis, Bi. spinulicostatus* M1, *Bi. stabilis vulgaris, M. strigosa*, *Br. inornata, Neo. communis* (Pl. 3, Fig. 11). Representatives of N. Gen 2 (Pl. 1, Fig. 15) belong to forms that survived from the Lower Member. A *Si. (Eosiphonodella)* n. sp. B (Pl. 1, Fig. 16) is just slightly more advanced than the specimen from Pl. 1, Fig. 14. In summary, Bed 8a produced the most diverse conodont fauna from the Hangenberg Crisis Interval (Middle *praesulcata* Zone) on a global scale. At least three, but possibly more, of a total of 17 taxa appear to have been reworked. For comparison, there are only nine identified taxa in the known second richest assemblage at Trolp (Austria, KAISER et al. 2009) and never more than four at La Serre (last update in KAISER 2009).

Conditions for conodonts deteriorated in the sandy, higher part of the lower subunit of the Upper Member. Beds 8 of the main section still yielded bispathodids (including a reworked *Bi. costatus* M2), *Branmehla, Mehlina*, and *Neopolygnathus*, Bed 10 has only *Neo. communis*.

Bi. aculaeatus aculeatus, Bi. aculeatus anteposicornis, and *Bi. stabilis vulgaris* return in the intercalated limestone in the middle subunit (Bed 11b) but the fauna remains restricted (five taxa). There is a nice specimen (Pl. 2, Fig. 1), which combines a siphonodellid platform with a polygnathid basal pit. There are two possibilities: it is either a transitional form between *Si. (Eo.) sulcata* s.l., (Group 7 of KAISER & CORRADINI 2011) and *Si. (Si.) duplicata* (s.str. = Morphoptype 2, see JI et al. 1985, pl. 19, figs. 14-15), or it is a N. Gen. 1 (with some relationships to *Si. (Eosiphonodella)* n. sp. B, which, however, has a much better developed pseudokeel).

The thin limestones at the base of the upper subunit of the Upper Member (Bed 12b-c) produced several Carboniferous-type conodonts (*Neo*. cf. *carina* and *Ps. primus*), in association with various long-ranging survivors, such as bispathodids, *Neo. communis* and *M. strigosa*. *Si. (Eosiphonodella)* n. sp. B is represented by several variants (Pl. 2, Figs. 2-5). There are several slightly curved *Si. (Eo.) praesulcata* (Pl. 2, Figs. 7-8), which fall in Group 3b sensu KAISER & CORRADINI (2011). A single, narrowly lanceolate specimen (Pl. 2, Fig. 6) has a curvature of more than 15° and, therefore, falls in *Si. (Eo.) sulcata* (s.l.) sensu FLAJS & FEIST (1988). It resembles critical specimens from Bed 85 at La Serre (FLAJS & FEIST 1988, pl. 6, figs. 9-10), which were variably regarded as transitional morphotypes between *praesulcata* and *sulcata* or as *sulcata* (ZIEGLER & SANDBERG 1996). Anyway, the new record from Lalla Mimouna North suggests a correlation of Bed 12b-c with the basal part of the Upper Oolite at La Serre.

Bed 12d lacks so far the *praesulcata/sulcata* Gp. but there is yet another morph of N. Gen 2 (Pl. 2, Fig. 9). Representatives or relatives of *Si. (Eosiphonodella)* n. sp. B also show variable morphologies (Pl. 2, Figs. 10-11), which continues in Bed 12e (Pl. 2, Figs. 12-14). The same sample yielded an intermediate between *Si. (Eo.) praesulcata* and *sulcata* s.l. with rather posterior platform curvature (Pl. 2, Fig. 15). A *Si. (Eo.) sulcata* s.l. (Pl. 2, Fig. 16) falls in Group 5 of KAISER & CORRADINI (2011) and suggests that Bed 12e should be assigned to the *sulcata* s.l. Zone. A single *Pa. gracilis gracilis* has been reworked.

Above the level of iron encrustations, Bed 12g yielded the only *Si. (Siphonodella)* of the whole succession, *Si. (Si.) carinthiaca* (Pl. 3, Fig. 4), a marker species for the higher Upper *duplicata* Zone (not *hassi* Zone, since *Si. hassi* JI 1985 is a junior homonym of *Si. cooperi hassi* THOMPSON & FELLOWS 1970). It is associated with a second *praesulcata-sulcata* s.l. intermediate (Pl. 3, Fig. 3, resembling the specimen from Pl. 2, Fig. 15), *Si. (Eo.) praesulcata* (Pl. 3, Fig. 2; cf. Group 2 sensu KAISER & CORRADINI 2011) and a triangular N. Gen. 2 (Pl. 3, Fig. 1). There is no evidence so far for the Lower *duplicate* Zone, which could be missing in the crust at the base of Bed 12g.

In Bed 13a there is a strange change of conodont biofacies, which resulted in the sudden disappearance of all siphonodelloids and the re-appearance of *Protognathodus*, with *Pr. meischneri* (Pl. 3, Fig. 6), *Pr. collinsoni* (Pl. 3, Fig. 7), and *Pr. kockeli*, accompanied by the bradytelic *Bi. stabilis vulgaris*

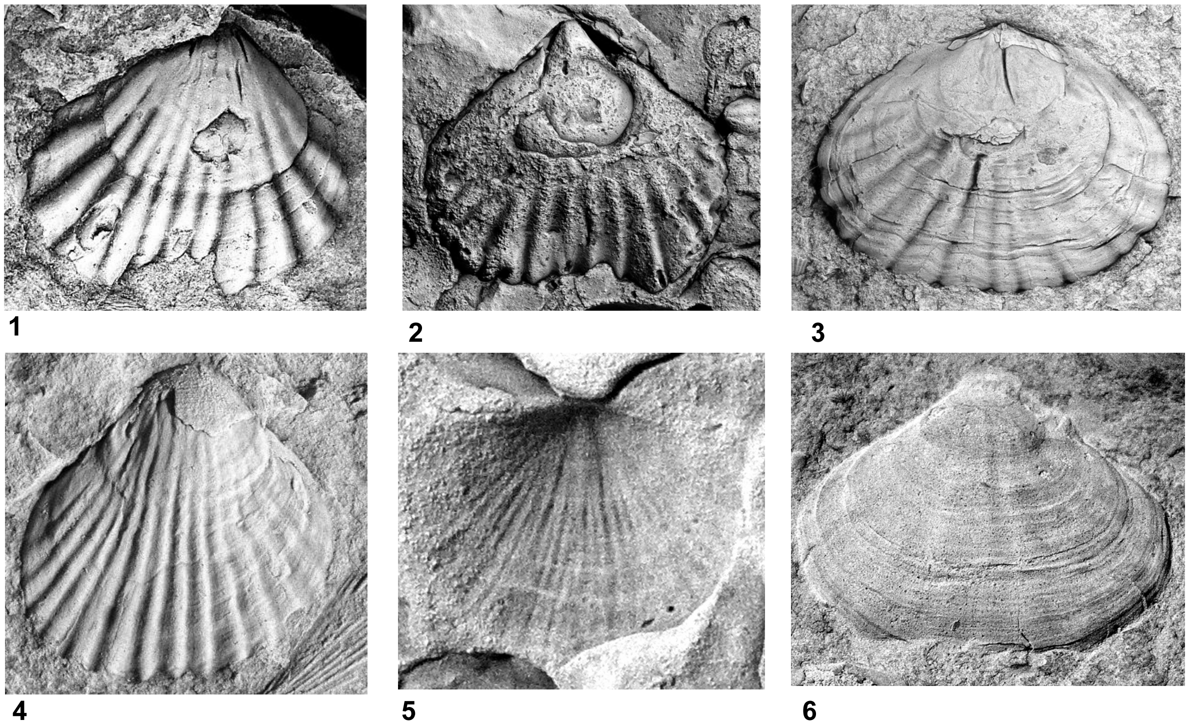

Fig. 9. Brachiopods from the Middle Member of the Lalla Mimouna Formation. **1-2.** ?*Centrorhynchus* sp. (Morphotype 1), **3.** cf. *Sedenticellula* sp., **4.** ?*Hemiplethorhynchus* sp., **5.** orthid, **6.** ?*Hadyrhyncha* sp.

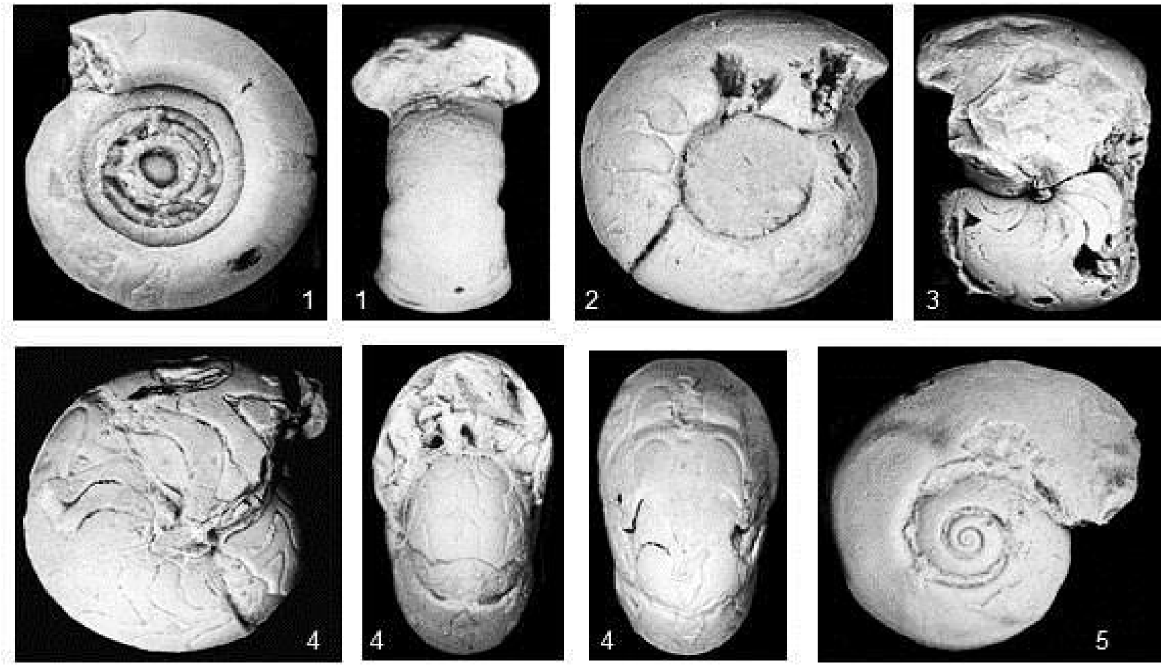

Fig. 10. Goethitic (primarily pyritic) fossils from the basal Fezzou Formation of Lalla Mimouna South. **1.** *Gatt.* aff. *jacquelinae*, **2.** *Gattendorfia* n. sp., **3.** **“***Imitoceras”* n. sp., **4.** *Acut. (Stock.)* aff. *endoserpens*, **5.** oxyconic euomphalid gastropod with constrictions.

and *Neo. communis*. At the top (Bed 13b), only the latter two taxa and advanced *Pr. kockeli* remain (Pl. 3, Figs. 8-9). The previous incomplete sampling of the section, without knowledge of the various Carboniferous-type true siphonodellids below, resulted in the false dating of Bed 13b as topmost Devonian *kockeli* Zone. A similar late return of the *Protognathodus* biofacies is known from La Serre (KAISER 2009), which gives an astonishing parallel despite the considerable distance. The similarities continue with the subsequent deepening of the local Fezzou Formation and corresponding Upper Member of the Griotte Formation (see BECKER & WEYER 2004).

4. CONCLUSIONS

The new faunas from Lalla Mimouna North are crucial for the understanding of the evolution of siphonodellids, their ancestors and of a parallel group (N. Gen. 2) with similar development of a pseudokeel. Their complexity is much higher than previously thought. Morphological transitions from very early (pre-Hangenberg) to more advanced (post-crisis) *Si. (Eosiphonodella)* could be used in chronostratigraphy, but somewhat below the current GSSP level. More taxonomic precision is required. Lalla Mimouna North is significant because of its diverse conodonts from the upper event interval (Beds 8-10). Currently the base of the *sulcata* Zone s.l. is placed at the base of Bed 12(b-c), which seems to correlate with the base of the Upper Oolite and revised base of the *sulcata* s.l. Zone at La Serre. The locality is not useful if the D/C boundary is defined by the Hangenberg Black Shale Event or by the onset of *Pr. kockeli*.

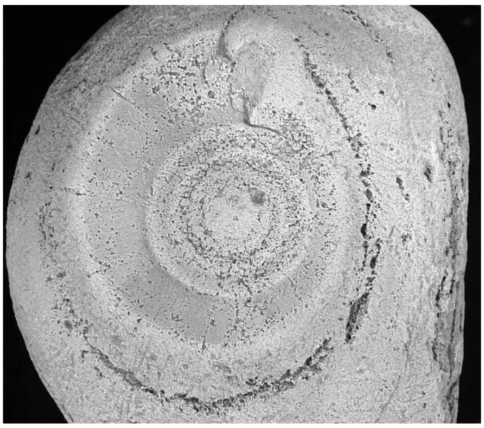

Fig. 11. Negative of a *Protocanites* sp. in a sideritic concretion from shales south of Lalla Mimouna North (Rharriz Formation)

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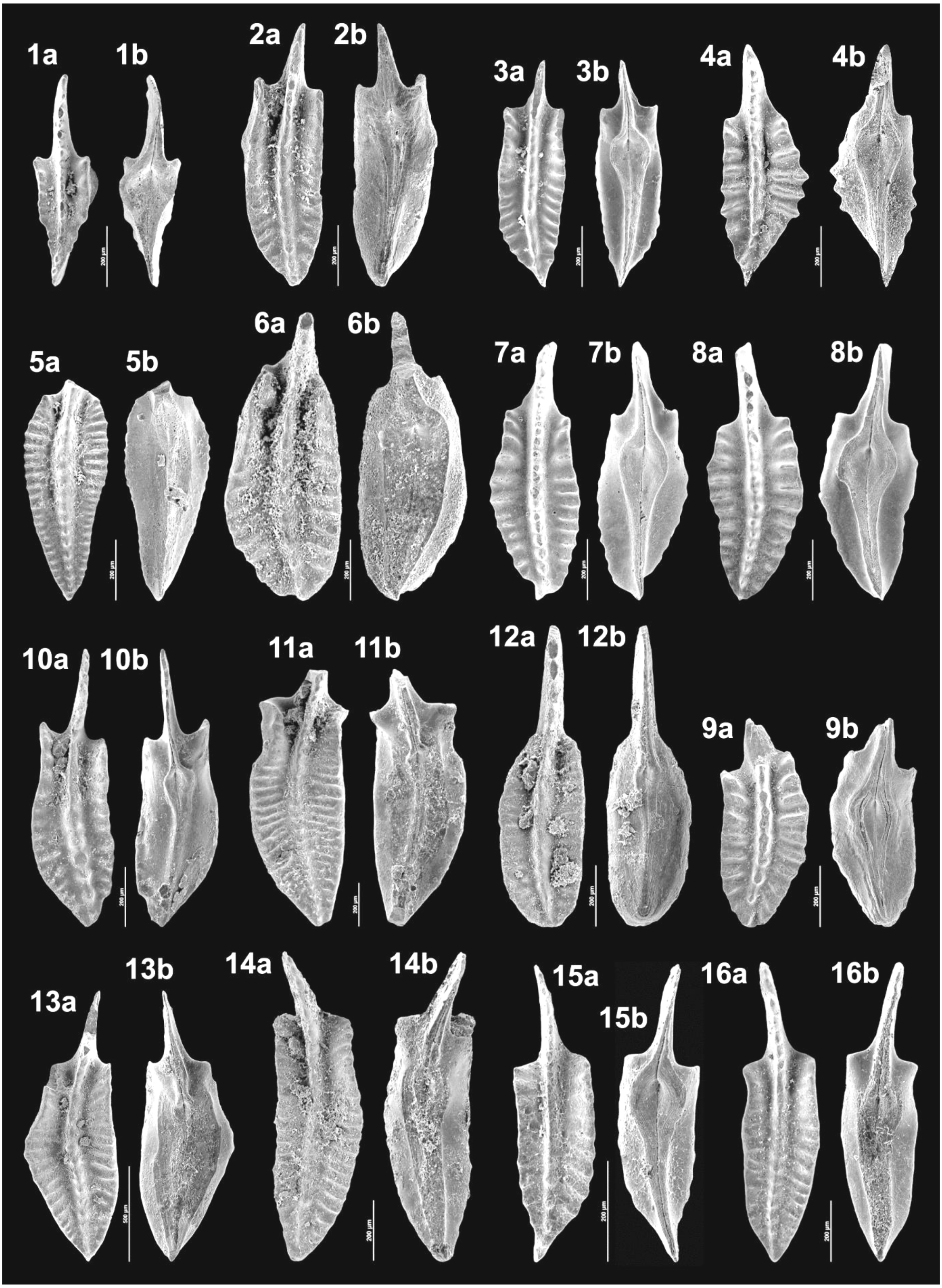
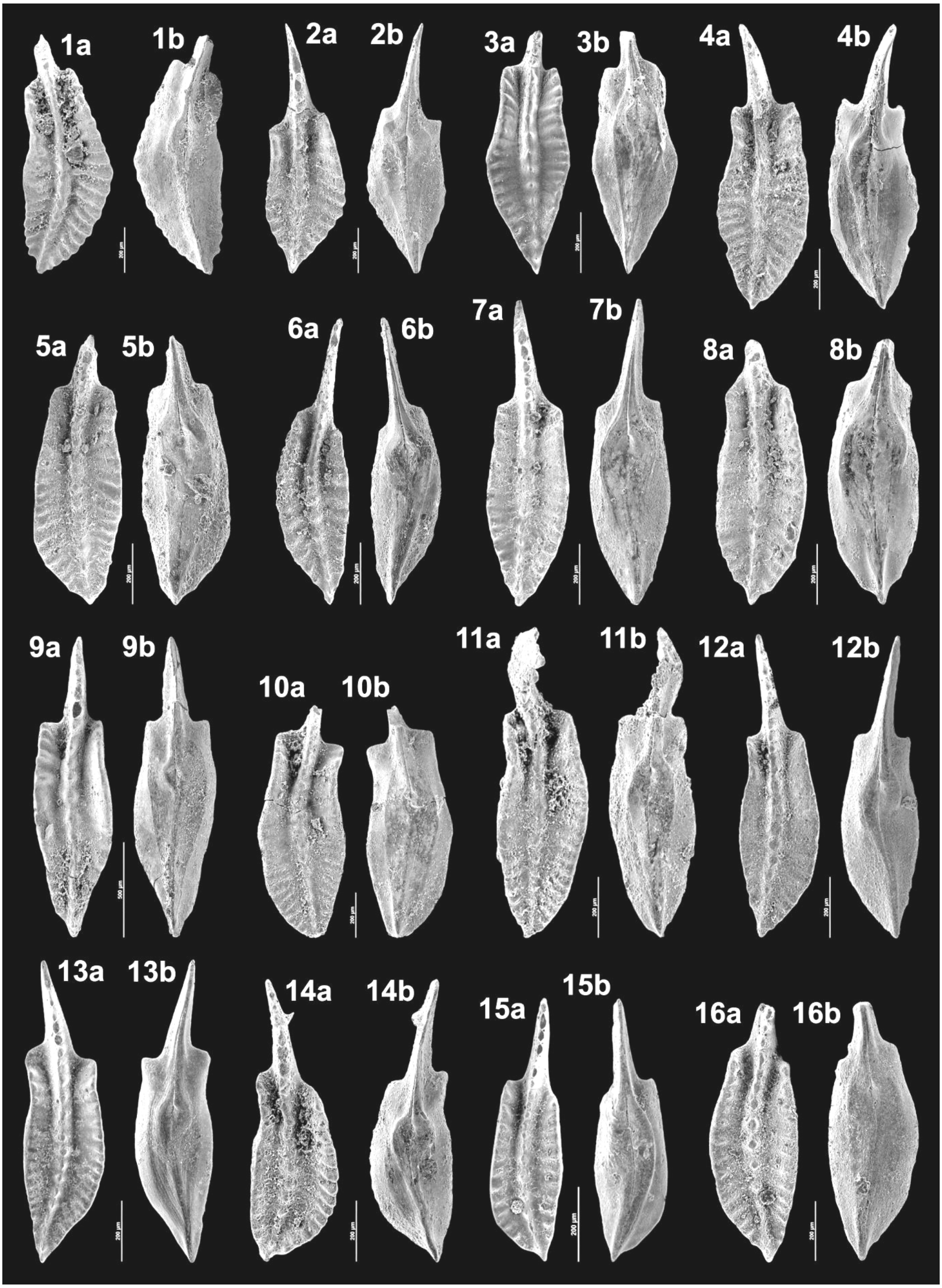
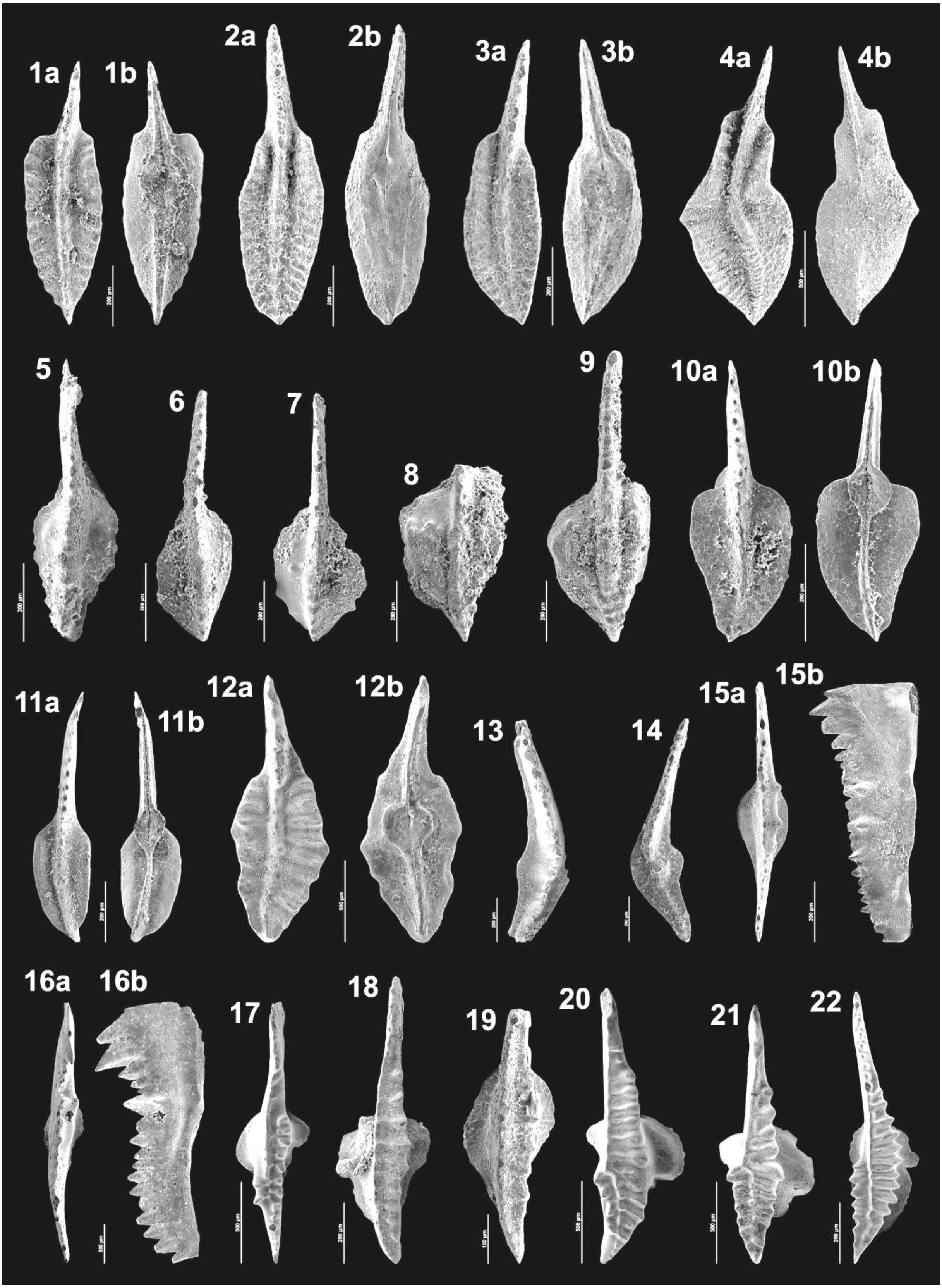


Pl. 1. Siphonodelloids from the Lower (1-14) and upper part of Middle Member (15-16). **1-4.** N. Gen. 2 div. sp. (four morphs/taxa, Fig. 4 with some similarity of platform shape, but not of the pseudokeel, to *Ps. primus* s.l.; compare Pl. 3, Fig. 12)., Bed 1a, **5.** N. Gen. 1 n. sp. A, Bed 1a, **6.** *Si. (Eosiphonodella)* n. sp. A, homoemorphic to *Si. sulcata*, Bed 1a, **7-8.** N. Gen. 2 sp. (possibly *"Ps." scitulus*), Bed 2a, **9.** N. Gen. 1 n. sp. B, transitional to *Si. (Eosiphonodella)* n. sp. A, Bed 2a, **10.** N. Gen. 2 n. sp. (similar to Fig. 2), Bed 6, **11.** N. Gen. 1 n. sp. C, with some resemblance to the *Po. symmetricus* Gp., Bed 6, **12.** N. Gen. 2 n. sp. D, Bed 6, **13-14.** *Si. (Eosiphonodella)* n. sp. B (two different morphotypes), still intermediate to N. Gen. 1, Bed 6, **15.** N. Gen. 2 n. sp., larger specimen of the same form as in Fig. 1, **16.**



Pl. 2. Siphonodelloids from the Upper Member of the Lalla Mimouna Formation. **1.** ?N. Gen 1 sp. (or an intermediate between *Si. sulcata* s.l. and *Si. duplicata* s. str.), Bed 11b, **2-5.** *Si. (Eosiphonodella)* n. sp. B (four variants), Bed 12b-c, **6.** *Si. (Eo.) sulcata* s.l. (advanced form of Group 3b sensu Kaiser & Corradini 2011), **7-8.** *Si. (Eo.) praesulcata* s.l. (typical Group 3b), Bed 12a-b, **9.** N. Gen. 2 n. sp. (different morph than below), Bed 12d, **10-11.** *Si. (Eosiphonodella)* n. sp. B (two different forms), Bed 12d, **12-14.** *Si. (Eosiphonodella)* n. sp. (three variiants), Bed 12e, **15.** *Si. (Eo.) praesulcata* s.l., (new variant intermediate to *sulcata* s.l.), Bed 12e, **16.** *Si. (Eo.) sulcata*, Group 5 sensu Kaiser & Corradini (2011), Bed 12e.



Pl. 3. Various conodonts from Lalla Mimouna North. **1.** N. Gen. 2 n. sp., Bed 12g (compare Pl. 1, Fig. 15), **2-3.** *Si. (Eo.) praesulcata,* s.l., cf. Group 2 sensu KAISER & CORRADINI (2011) and an intermediate to *sulcata* s.l. (see Pl. 2, Fig. 15), Bed 12g, **4.** *Si. (Si.) carinthiaca*, Bed 12g, **5-6.** *Pr. meischneri*, Bed 8a and 13a, **7.** *Pr. collinsoni*, Bed 13a, **8-9.** *Pr. kockeli*, poorly preserved and advanced specimens, Bed 13b. **10.** *Po*. cf. *purus purus*, Bed 8a, **11.** *Neo. communis*, Bed 8a, **12.** *Ps*. aff. *primus*, Bed 12b-c, **13-14.** *Pa. gracilis gracilis*, Bed 6 and reworked in Bed 8a, **15.** *Bi. aculeatus aculeatus*, Bed 1a, **16.** *Bi. aculeatus anteposicornis*, Bed 12e, **17.** *Bi. spinulicostatus* M1, Bed 2a, **18-19.** *Bi. costatus* M2, Bed 6 and reworked in Bed 8a, **20-21.** *Bi. ultimus* M2, Bed 2a, **22.** *Bi. ultimus* M1, Bed 2a.

FAMENNIAN AND TOURNAISIAN STRATA OF THE AGUELMOUS SYNCLINE

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1. INTRODUCTION

The name of the town Fezzou is world renown for well-preserved Late Devonian ammonoids in goethitic or limonitic steinkern preservation (Fig. 2), beginning with the extensive monographs by PETTER (1959, 1960). From the late 1970s such material was frequently sold in fossil and gemstone fairs, and specimens are stored in numerous collections worldwide. Unfortunately for the scientific community, commercial exploitation of the fossiliferous strata took placed before a detailed biostratratigraphic investigation has been achieved. However, nowadays it is still possible to collect nicely preserved ammonoids in a large outcrop. These faunas were the subjects of several research projects, and their succession is now rather precisely known (KORN 1999, BECKER & HOUSE 2000, BECKER et al. 2002, EBBIGHAUSEN & KORN 2007).

The Anti-Atlas of Morocco (Fig. 1) is one of the key regions for the Famennian ammonoid stratigraphy. Good outcrop conditions and fossiliferous sections are widespread over a wide area with various facies types, ranging from successions almost entirely composed of carbonates to sections dominated by fine-grained siliciclastics with only few limestone interbeddings. WENDT et al. (1984) and WENDT (1985, 1988) gave an overview for lateral facies distributions and provided the basis for later investigations. According to their studies, the Ma'der (or 'Maïder') and Tafilalt (or 'Tafilalet') regions of the eastern Anti-Atlas show a complex platform and basin distribution in the Famennian sedimentary succession. Characteristic for the platform-oriented sections are significant gaps in the successions, whereas the basin sections are much more complete and serve as reference sections of the area.

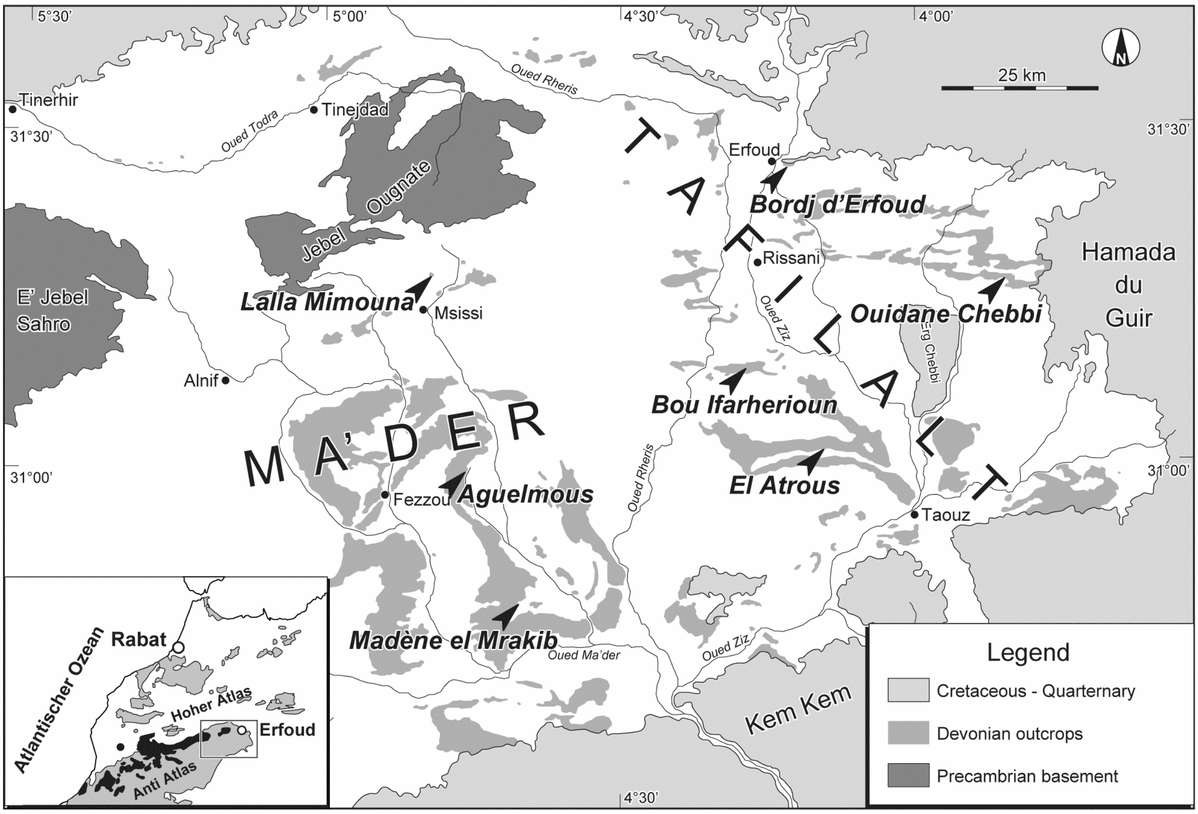

Fig. 1. The position of the Aguelmous Syncline in the Anti-Atlas of Morocco.

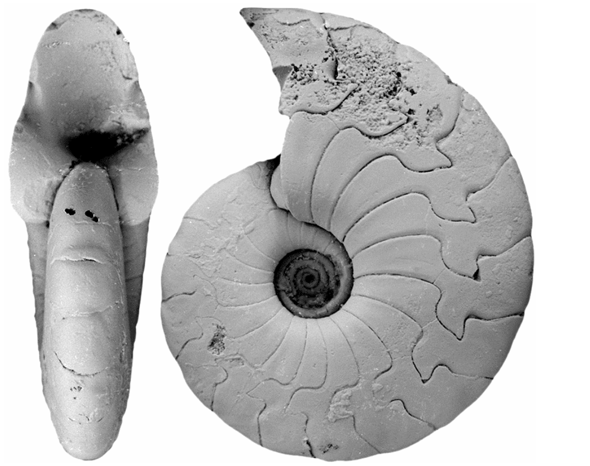

Fig. 2. Typical steinkern specimen of *Cymaclymenia* from the Aguelmous Syncline in the Anti-Atlas of Morocco.

2. THE LATE FAMENNIAN TO EARLY TOURNAISIAN SUCCESSION IN THE AGUELMOUS SYNCLINE

The Aguelmous is located northeast of Fezzou in the eastern Anti-Atlas of Morocco (Fig. 3). In the literature and on local maps, the name Aguelmous is used for two different topographic structures, (1) the entire area north-east of Fezzou, and (2) the ridge beginning directly behind the houses north-east of Fezzou and extending for approximately 16 km towards the north-east. This ridge forms the north-western flank of the Aguelmous Syncline and has a SW-NE axis, extending for 18 km, with an average width of 6 km. Late Devonian and Early Carboniferous strata have an inclination of 10-15º on both flanks of the syncline and show only little additional deformation. The section begins with fossil-rich Famennian shales, followed by latest Devonian sandstone beds of the Aoufital Formation forming a conspicuous topographic ridge (Aguelmous on the north-western flank, Rharrhiz and Lambidia on the south-eastern flank). The depression in the centre of the syncline is composed of Early Carboniferous shales and thin sandstone beds (EBBIGHAUSEN & BOCKWINKEL 2007).

Below the Aoufital Formation Sandstone, late Devonian grey shales with *Wocklumeria*-bearing assemblages (EBBIGHAUSEN & KORN 2007) are overlain by the 2 to 3 m thick equivalent of the Hangenberg Black Shale, first discovered and stratigraphically dated by KORN (1999) in the area of Madène el Mrakib southeast of Fezzou in the Southern Ma’der. At this place, KORN discovered poorly preserved specimens of *Acutimitoceras* sp. from the basalmost sandstone beds of the Aoufital Formation.

The Early Carboniferous sediments of the Aguelmous are mainly composed of siliciclastics (shales, siltstones) with very few carbonatic intercalations. The rock succession is deeply weathered and widely covered by debris. Sections can be measured only in few areas, and a combination of several sections led to the synthesis (Fig. 4). Lateral facies changes could not be observed in the study area. Although the faunas described in this study have been surface collected, a mixing of horizons can be ruled out. All assemblages are located in considerable distances from each other and are separated by sandstone ridges. KAISER (2005) provided a detailed description of the section.

2.1. Devonian

Late Famennian sections in the Aguelmous Syncline are claystone-dominated and represent the thickest sedimentary successions of this time interval in the Anti-Atlas. Most of the sections are composed of grey shales with occasional limestone intercalations, which are usually thin horizons with sideritic nodules. Ammonoid faunas occur throughout the Famennian succession; they are accompanied by orthocone cephalopods, brachiopods, rugose corals, crinoids, vertebrate remains, etc. However, particularly the ammonoids are, besides conodonts, the cardinal fossils for a biostratigraphic subdivision.

BECKER & HOUSE (2000) and BECKER et al. (2002) provided an ammonoid-based regional stratigraphic scheme for the Famennian sedimentary rocks of the Anti-Atlas. This scheme contains up to 26 units (zones or subzones), based mainly on sections of the Aguelmous Syncline. These authors compared this scheme with the zonation, which had been developed in the Rhenish Mountains; they found a number of common zones but also striking differences, the most critical of which have to be discussed on the basis of new investigations.

Platyclymenia interval

A major discrepancy exists in the description and stratigraphic interpretation of the *Platyclymenia* bearing beds of the Ma’der Basin. While, for the locality at Madène el Mrakib in the southern Ma’der, KORN (1999) proposed a zonation that largely correlates with the Rhenish Mountains, BECKER et al. (2000, 2002) published a more detailed and somehow deviating stratigraphic scheme for the sections in the Anti-Atlas. The differences mainly refer to the interpretation of species co-occuring with the genus *Platyclymenia*. A long list with more than 15 ammonoid species was reported by BECKER et al. (2000) and partly cited by HARTENFELS (2011) to co-occur (in beds O and P of the Madène el Mrakib section) with *Platyclymenia*. A possible contamination from younger beds outcropping above has been mentioned but not intensely discussed. In addition to characteristic species of the *Platyclymenia annulata* Zone, ‘*Cymaclymenia* n. sp.’ was reported being abundant in these horizons. These putative records were used to place beds O and P in the newly defined ‘*Cymaclymenia* n. sp. Zone’ or ‘*Cymaclymenia pudica* Zone’ in BECKER et al. (2002).

The collections by KORN (1999) and our new investigations between 2006 and 2010 (including about 2,000 in situ collected ammonoids) draw a picture significantly differing from the report provided by BECKER et al. (2000, 2002) and HARTENFELS (2011). While these stratigraphic schemes, regarding the shale intervals, are based on surface collections and thus strongly suffer from contamination by material from younger horizons, our new investigations are entirely based on trenches with in situ material. These investigations demonstrate that the fossils derive from a number of thin horizons and not from wide intervals and that some of the most important index fossils have a first occurrence much higher in the section, when compared with BECKER et al. (2000, 2002).

The new collections do not confirm the presence of '*Cymaclymenia pudica*' (now placed into *Procymaclymenia*) from beds O and P at Madène el Mrakib. In fact, only taxa characteristic for the *Platyclymenia annulata* Zone were collected, and not a single specimen of *Procymaclymenia* has been found in the course of our field investigations (Fig. 5). The 'abundant' occurrence of this genus, as stated by BECKER et al. (2000) must be doubted and regarded as contamination; the specimens referred to *Cymaclymenia* come from a distinct horizon with a mass occurrence of *Procymaclymenia* higher in the section. Therefore, beds O and P must be assigned to the *Platyclymenia annulata* Zone, and most of the '*Cymaclymenia pudica* Zone' is to be synomized.

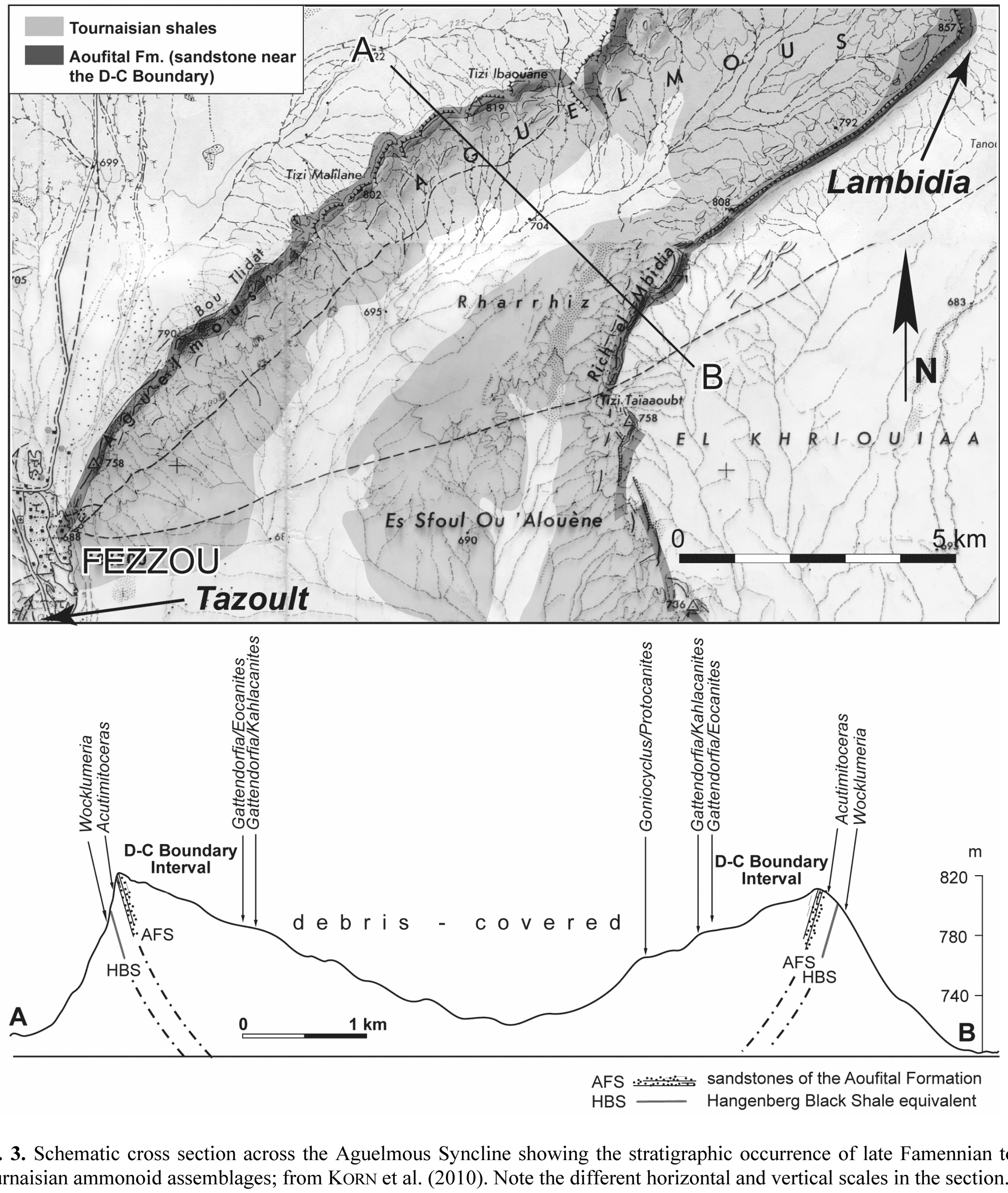

Fig. 3. Schematic cross section across the Aguelmous Syncline showing the stratigraphic occurrence of late Famennian to Tournaisian ammonoid assemblages; from KORN et al. (2010). Note the different horizontal and vertical scales in the section.

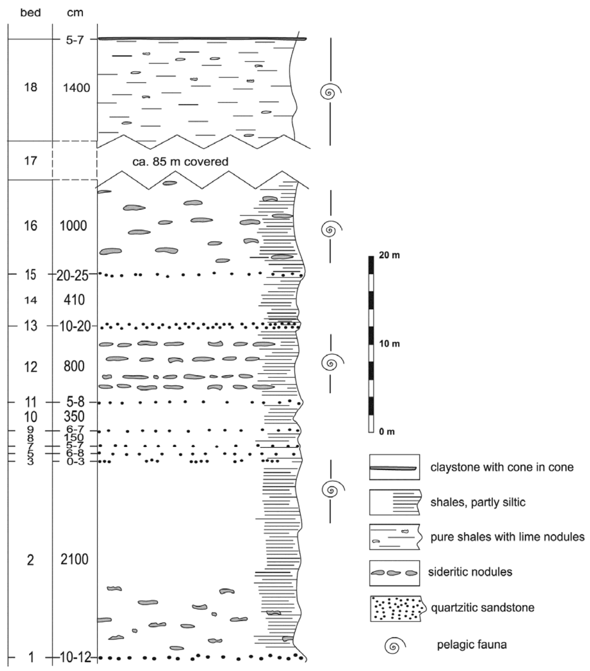

Fig. 4. The early Tournaisian sedimentary succession in the Aguelmous Syncline; from EBBIGHAUSEN & BOCKWINKEL (2007).

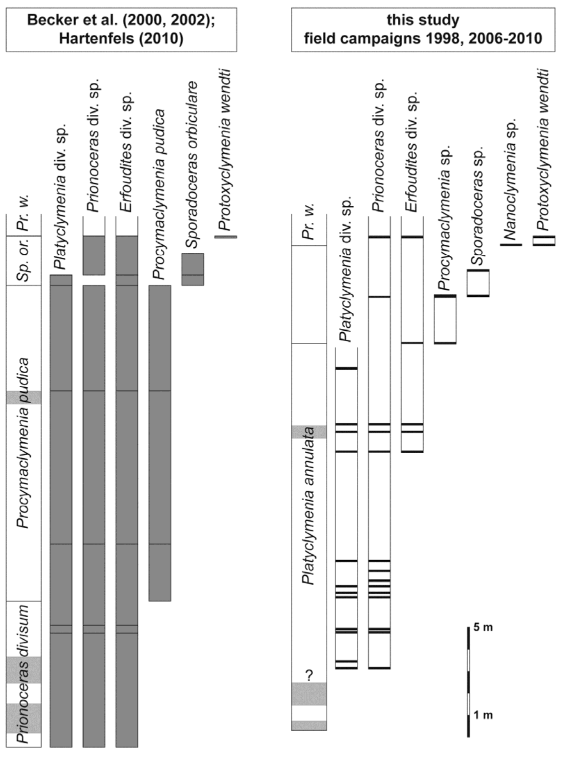

Fig. 5. The occurrence of most important ammonoid genera in the section of Madène el Mrakib (southern Ma’der).

Kalloclymenia interval

In a series of articles (BECKER et al. 2000, 2002; BECKER & HOUSE 2000) it was proposed that *Kalloclymenia subarmata* occurs, in the Eastern Anti-Atlas, in older horizons than in the Rhenish Mountains. As co-occurring genera, *Gonioclymenia* and *Clymenia* were mentioned, both of which characterise the *Clymenia* Stufe in the Rhenish sections.

Kalloclymenia was, right from the first definition of the ‘*Wocklumeria* Stufe’ by SCHINDEWOLF (1937), regarded as the main index genus for the base of this stage. KORN (1999) confirmed this based on investigations of sections in the Tafilalt. However, BECKER & HOUSE (2000) as well as BECKER et al. (2002) did not accept this view and correlated their ‘*Kalloclymenia subarmata* Zone’ in the Anti-Atlas with *Piriclymenia piriformis* Zone in the Rhenish Mountains, i.e. horizons containing the genus *Gonioclymenia*. There is no reason for such a correlation; these authors did not supply evidence for their hypothesis that *Kalloclymenia subarmata* occurs, in the Anti-Atlas of Morocco, one zone earlier than in the Rhenish Mountains. BECKER et al. (2002, p. 172) conceded that a co-occurrence of *Gonioclymenia* and *Kalloclymenia* ‘has only be observed in few strongly condensed Tafilalt sections’, but without providing information of which sections and fossil horizons were meant. During our field investigations (e.g., KORN 1999) in the Anti-Atlas of Morocco, the two genera *Gonioclymenia* and *Kalloclymenia* were never collected from the same bed. As a consequence, the two zones must be separated (HARTENFELS & BECKER 2012).

2.2. CARBONIFEROUS

The occurrence of Carboniferous ammonoids in the Ma’der, particularly at the Aguelmous ridge immediately north-east of Fezzou, has been known since the 1930’s, when CLARIOND (1935) mentioned “*Muensteroceras* sp.” and “*Aganides* sp.” from a locality a few kilometres east of Fezzou and proclaimed a Tournaisian age for this record. This view was adopted in the Carte Géologique du Maroc 1:200,000 (mapsheet Todhra-Ma’der), published by the Service Géologique du Maroc (DESTOMBES & HOLLARD 1988), in which Tournaisian sedimentary rocks are shown as exposed in the Aguelmous Syncline. Though it has long been known that “*Gattendorfia*” and “*Prionoceras subbilobatum*” can be found in this area (HOLLARD 1958), there has been no extensive collection of ammonoid faunas. During several field excursions in the years 2003 to 2006, the Carboniferous ammonoids of the Aguelmous Syncline were extensively collected from four horizons (Fig. 6), and more than 1,300 specimens from four horizons were available for their study (EBBIGHAUSEN & BOCKWINKEL 2007).

Age of the ammonoid fauna (from EBBIGHAUSEN & BOCKWINKEL 2007).

Ammonoid faunas were recorded from beds 2, 12, 16, and 18 (Fig. 4). The ammonoid assemblages of beds 2 and 12 show close similarities with the fauna described by BOCKWINKEL & EBBIGHAUSEN (2006) from Mfis in the southern Tafilalt of Morocco. The lack of index genera such as *Paprothites* and *Pseudarietites*, as characteristic for the *Gattendorfia*

Limestone of the Rhenish Mountains (VÖHRINGER 1960) and South China (RUAN 1981), but the presence of advanced forms such as *Imitoceras* speak for a position above the classical *Gattendorfia* Stufe. KORN et al. (2007) placed this fauna in the *Gattendorfia-Eocanites* Assemblage of the North African Carboniferous ammonoid succession. Bed 16 is dominated by the genera *Gattendorfia*, *Kahlacanites*, and *Hasselbachia* (Fig. 6). A very similar fauna is known from the Grès supérieur de Kahla from the Gara el Kahla near Timimoun in north-western Algeria (EBBIGHAUSEN et al. 2004). This fauna is placed in the *Gattendorfia-Kahlacanites* Assemblage (KORN et al 2007).

bed	2	12	16	18
Globimitoceras rharrhizense	□	□	□	■
Acutimitoceras hollardi	■	□	□	□
Acutimitoceras intermedium	■	□	□	□
Acutimitoceras occidentale	■	□	□	□
Acutimitoceras depressum	■	□	□	□
Acutimitoceras sarahae	■	□	□	□
Acutimitoceras mfisense	■	□	□	□
Acutimitoceras endoserpens	■	■	□	□
Acutimitoceras algeriense	■	■	■	□
Acutimitoceras sp. A	□	■	□	□
Acutimitoceras posterum	□	□	■	□
Acutimitoceras pentaconstrictum	■	□	□	□
Costimitoceras aitouamar	■	■	□	□
Hasselbachia gourara	□	□	■	□
Hasselbachia arca	□	□	■	□
Hasselbachia sp.	□	□	■	□
Kornia citrus	■	□	□	□
Imitoceras oxydentale	■	■	□	□
Imitoceras sp.	□	□	□	■
Gattendorfia jacquelinae	■	■	■	□
Gattendorfia debouaaensis	□	□	■	□
Gattendorfia lhceni	□	■	□	□
Gattendorfia gisae	□	□	■	□
Gattendorfia sp.	□	□	□	■
Kazakhstania evoluta	■	■	□	□
Kazakhstania nitida	■	□	□	□
Goniocyclus elatrous	□	□	□	■
Eocanites simplex	■	□	□	□
Eocanites sp.	□	□	□	■
Becanites sp.	□	□	■	□
Kahlacanites mariae	□	□	■	□
Kahlacanites meyendorffi	□	□	■	□
Protocanites hollardi	□	□	□	■
Gen. indet. 1 sp. indet	□	■	□	□
Gen. indet. 2 sp. indet	□	□	■	□

Fig. 6. Distribution of the Early Carboniferous ammonoid species in the Aguelmous sections; from EBBIGHAUSEN & BOCKWINKEL (2007).

The poorly preserved fauna from bed 18 is almost identical with the fauna known from a locality 75 km east of the Aguelmous in the Amessoui Syncline (KORN et al. 2002). It belongs to the *Goniocyclus-Protocanites* Assemblage (KORN et al. 2007).

3. THE OUTCROP NEAR LAMBIDIA

Among the multiple outcrops of late Famennian rocks, the area of Lambidia best exposes the succession from the *Platyclymenia annulata* Zone to the topmost Devonian sandstone unit (Fig. 7). The outcrop is composed of two tectonic units separated by a steep (45° and more), nearly N-S trending fault. The eastern unit, which represents a tilted block in which the beds are steeply inclined in western direction, shows the better exposure. The outcrop begins near the *Platyclymenia annulata* Zone, but these beds are poorly exposed. An almost continuous outcrop then exposes a thick succession predominantly composed of grey shales, which possess a sharp upper boundary at the contact to the Hangenberg Black Shale. This unit of about two metres thickness is strongly weathered and bleached. Above follow bedded and compact sandstone beds of the Aoufital Formation ("Lower Hangenberg Sandstone" of BECKER et al. (2002).

The western unit, which shows beds gently (10–15°) inclining to the west, shows a repetition of the upper portion of the eastern unit. BECKER et al. (2002) did not recognize the fault and repetition of strata; they suggested a continuous succession for the outcrop. Based on the occurrence of cymaclymeniids between their "Lower Hangenberg Sandstone" and "main Hangenberg quartzites", they introduced a "*Cymaclymenia involvens* Subzone", which, because of the reasons explained above, must be treated as invalid. Doubts on the validity of the succession were already casted by KAISER (2005), who regarded tectonic repetition as the most likely explanation. KAISER et al. (2011) interpreted the repetition as "post-sedimentary doubling of pre-event beds because of collapse of the steep palaeoslope". The first detailed description of the section was provided by BECKER et al. (2002), who published a columnar section of the beds representing four ammonoid zones of the late Famennian *Wocklumeria* Stufe (Fig. 7).

4. THE OUTCROP NEAR TAZOULT

The Devonian-Carboniferous boundary beds are poorly exposed on the shallow hill of Tazoult immediately south of Fezzou.

One of the Early Tournaisian sandstone horizons shows, at the interface with the shale horizon below, an arthropod trace, which is almost entirely covering the bedding surface. This conspicuous trace fossil has been figured by SEILACHER (2007, p. 196) as and has been attributed to the ichnospecies *Cruziana* cf. *reticulata* (Fig. 8). The bedding surface shows a large number of rather short tracks, which rarely reach a length of 50 cm. Their width is usually about 20 mm and their depth reaches 50 mm. In general the tracks have a shape of a trapezoidal cutting with sharp endopodal scratches. Currently the trace fossil is being studied in a research project by B. GUTWASSER (Berlin).

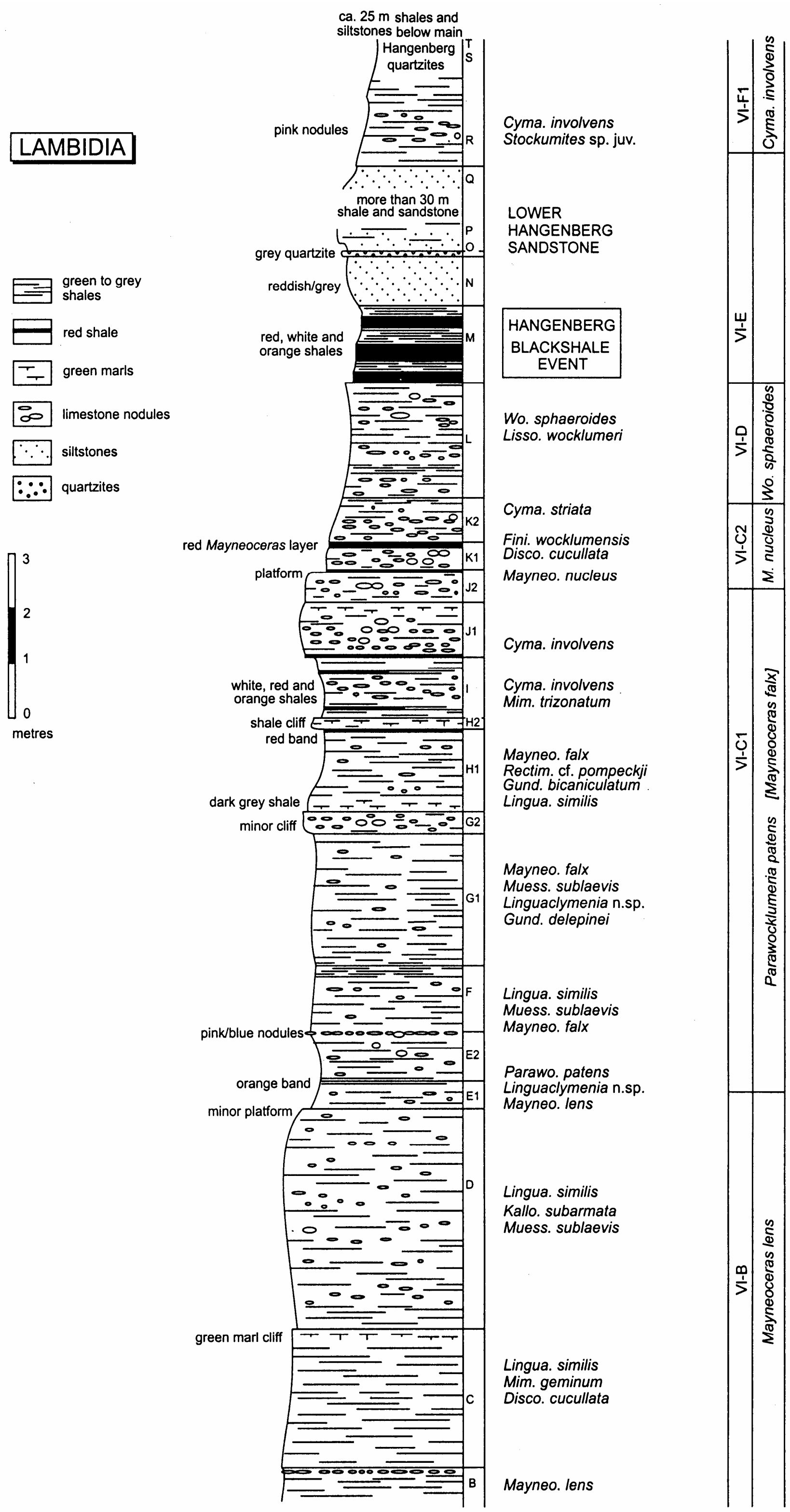

Fig. 7. The latest Famennian sedimentary succession in the outcrop at Lambidia; from BECKER et al. (2002).

According to the shape and ornament of the trace fossil it can be stated that the single tracks have not been formed in a single event. The same tracks must have been used again and again by similar-sized animals, as evident by the direction of scratches.

However, the originator of the trace is unknown. The morphology of the trace resembles those caused by Early Palaeozoic trilobites, but trilobites of the size necessary for the large traces are not known from the Early Carboniferous.

Fig. 8. The trace fossil *Cruziana* cf. *reticulata* from a basalmost Carboniferous sandstone bed of Tazoult.

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THE GLOBAL CHOTEC EVENT AT JEBEL AMELANE (WESTERN TAFILALT PLATFORM) – PRELIMINARY DATA

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1. INTRODUCTION

The term (Basal) Choteč Event was introduced byCHLUPÁČ, I. &KUKAL, Z. (1986, 1988) in Bohemia, where there is an abrupt change low in the Eifelian from the light-grey to reddish Třebotov to darker, more organic-rich ChotečLimestone. The intercalation of dark calcareous shales in some sections, ca. 1-2 m above the base, marks the peak of the transgressive event. It has recently been restudied in more detail by BERKYOVÁ (2009) and KOPTÍKOVÁ (2011), who supported the idea of a (probably eustatic) sea-level rise. The increased organic content is partly caused by sudden algal (prasinophyceen) blooms (BROCKE et al. 2009). CHLUPÁČ&KUKAL (1988) briefly summarized associated faunal changes and the possible global nature of the event. However, detailed studies in other regions are still almost lacking, with the exception of Bolivia (TROTH et al. 2011) and Nevada (ELRICK et al. 2009, PEDDER 2010). But in the latter region, the recognition of the event interval is seriously hampered by strong condensation and unconformities (VODRÁŽKOVÁ et al. 2011). For northern Spain there is a review by GARCIA-ALCALDE (1998) but more bed-by-bed data (in addition to, e.g., HENN 1985, MONTESINOS 1987, and GARCÍA-LÓPEZ&SANZ-LÓPEZ (2002) are needed. There is a detailed study in the Ossa Morena Zone of SW Iberia (MACHADO et al. 2010) but with ambiguities based on gaps in the section.

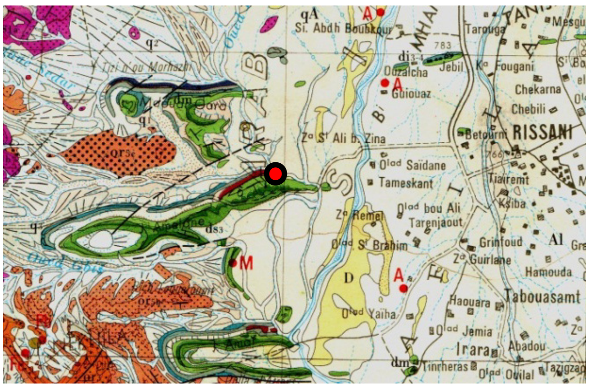

Fig. 1. Position of the Jebel AmelaneChoteč Event locality west of Rissani(red dot) on an excerpt from the geological map (sheet Tafilalt-Taouz). The red M to the south marks the Jebel MechIrdaneEifelian-Givetian Boundary Stratotype.

This preliminary report is the first detailed study on the Choteč Event in southern Morocco. The event has been first recognized in the Tafilalt by BECKER & HOUSE (1994, 2000), based on the BouTchrafine and

Fig. 2. Overview of the near vertical Choteč Event section near the eastern end of Jebel Amelane. Bed 3 is the solid bed to the right, Beds 11-12 form the minor cliff in the middle, Bed 13 the lower part of the lighter-grey rubbly limestone just left of it.

Jebel Amelane sections. The latter(at x = 600.4, y = 76.4) was chosen for a more detailed survey because of the very distinctive black styliolinites, marls and limestone, which interrupt an otherwise light-grey, well oxygenated hemipelagic platform setting. Emphasis was placed on the precise dating in terms of the conodont stratigraphy but modern dacryoconarid data would be desirable.KLUG (2002a, 2002b) outlined the regional significance of the Choteč Event for ammonoid diversity and morphology.

2. SEDIMENTS AND FAUNAS (Figs. 2-3)

The lower, upper Emsian part of the succession with abundant anarcestids (*Anarcestessimulans*, *Sellanarcestes* div. sp.), large bivalves (*Panenka* div. sp.), orthocones, and phacopids has been described by BECKER & HOUSE (1994). Sellanarcestids are restricted to the lower ca. 2 m, *An. lateseptatus* was found in Bed A_4. The upper part of these yellowish-weathering, nodular limestones form Beds 0-2 of the new section log. Bed 0 yielded *Panenka* and solitary

Rugosa, Bed 2 a *Fidelites* sp. Bed 3a (Fig. 4) is a solid marker limestone just below the Choteč Event Interval

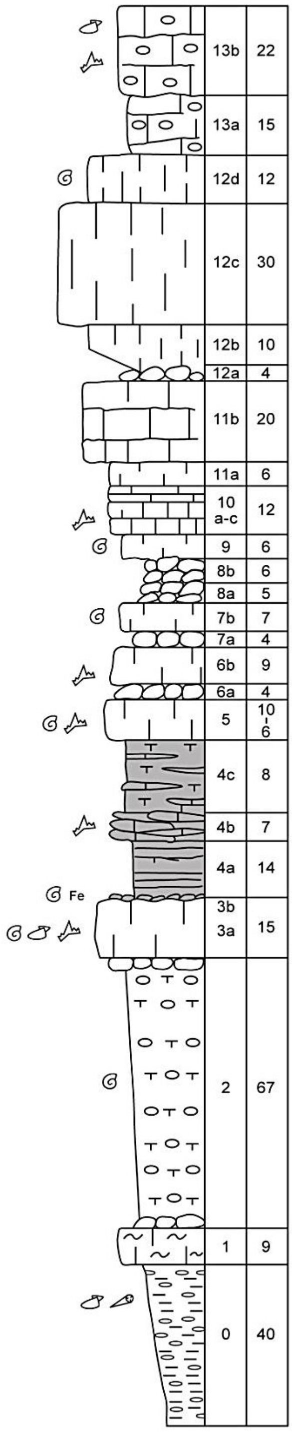

Fig. 3. Lithological log of the lower Eifelian at Jebel Amelane, showing current conodont sampling levels.

withsome more *Fidelites* and the oldest, large *Subanarcestes*cf. *marhoumensis*. A thin, nodular iron crust at the top (Bed 3b) indicates a condensation interval and shallowing upwards. The conodont assemblage from Bed 3a includes *Icriodus struvei* (Fig. 7.1), *Linguipolygnathus bultyncki* (Figs. 7.2-3), *Ling. pinguis* (Fig. 7.4), *Ling. linguiformis*, and *Ling. zieglerianus* (Fig. 7.5).

The event interval (Figs. 5-6) is marked by a sudden change to black, organic-rich, irregularly laminated, platy marls and styliolinites (*Styliolina*packstones, Beds 4a-c). This reflects a significant deepening and eutrophication pulse. So far there are no identifiable palynomorphs (R. BROCKE, oral comm.). The middle part is calcareous enough for conodont sampling. It yielded a restricted assemblage with *I. struvei, Ling. zieglerianus, Ling. coopericooperi* (Fig. 7.6), and a questionable *Po. costatuspartitus*.

Fig. 4. Thin-section of Bed 3a, a bioturbatedpackstone with abundant dacryoconarids, debris of bivalves and goniatites, gastropods, and trilobites.

Fig. 5. Details of the Choteč Event Interval at Jebel Amelane.

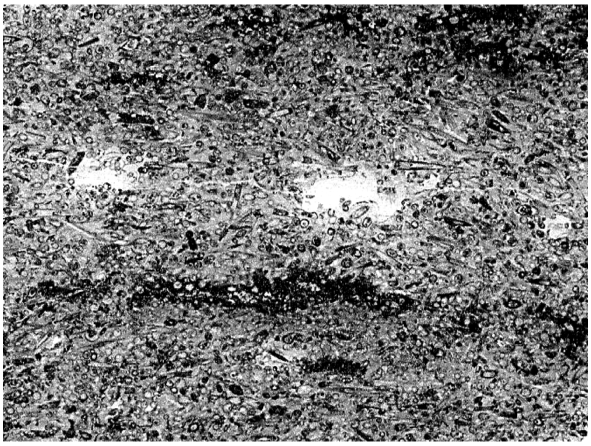

Fig. 6.Thin section of the dark, laminated styliolinites (styliolinidpackstone) of Bed 4b.

Above, there is an alternation of thin-bedded middle to light-grey nodular limestones, some with poorly preserved *Fidelites* (Beds 5, 7b, 9). Previously (BECKER & HOUSE 1994), a *Pinacites* sp. was found in Bed 5, which also contains *Ling. Pinguis* (common, Fig. 7.7)*, Ling. linguiformis, Ling. Zieglerianus* (Fig. 7.8),and a single *Ling. sogdianensis* (Fig. 7.22, new record for North Africa). Slightly higher, Bed 6b, a bioturbated, crinoidal dacryoconarid packstone with trilobites and partly washed out micrite matrix, produced intermediates between *Ling. zieglerianus* and *Po. robusticostatus*, still with a short lingua (Fig. 7.9, compare WEDDIGE 1977), in association with *Ling. linguiformis* (Fig. 7.16), *Ling. pinguis, Ling. zieglerianus,* and *I. struvei*. Bed 10a has the richest conodont fauna of all samples, with first entries of *I. amabilis* (very rare, Fig. 7.11), *I.* aff. *Regularicrescens* (Fig. 7.15), *I. anterodepressus* (common, Fig. 7.13), a new “*Ozarkodina*” (very rare, Fig. 7.12), and *Po. Angusticostatus* (Fig. 7.23). *Po.costatuspartitus* is the only subspecies of the *costatus* lineage. There are also re-appearances of long-ranging icriodids, such as *Caudicriodus culicellusculicellus* (Fig. 7.10) and common*I. cornigercorniger* (Fig. 7.14), and of *Ling. coopericooperi. I. struvei* is the most common species; the icriodid dominated (> 72 %) biofacies suggests a regressive trend.

More solid limestones start with Bed 11b and Beds 12b-f form a minor cliff, which represent the peak of regression. They display cross-sections of *Subanarcestes*. The massive limestone set correlates with the massive Bed 7 at BouTchrafine (BULTYNCK 1985, BECKER & HOUSE 1994) and equivalents can be found in many other sections (KLUG 2002a, 2002b). The nodular limestones on the southern or road side of the minor cliff resemble the upper Emsian facies in the presence of nautiloids and *Panenka*. Bed 13b is characterized by a sudden flood occurrence of *Po. costatuscostatus* (Fig. 7.18) in association with a few last *Po. costatuspartitus* (Figs. 7.19-20), common *Ling. linguiformis* and *I. anterodepressus*, rare *Po. praetrigonicus* (Fig. 7.21), some *Ling. zieglerianus*, *I.*aff. *struvei*, and last *Ling. bultyncki*(Fig. 7.17),and *Ling. pinguis*.

3. CONODONT STRATIGRAPHY

The precise dating of the Choteč Event Interval in the conodont zonation faces several problems: 1. Ambiguity concerning the distinction between the three index subspecies of *Po. costatus*, which are the zonal markers around the Emsian-Eifelian boundary (see comment by MACHADO et al. 2010). Variability and intermediate morphotypes are the main problem. 2. Rarity or absence of the marker lineage in the boundary and event interval (e.g., BERKYOVÁ 2009). 3. Unclear precise ranges of possible alternative index forms, for example of *Po. praetrigonicus* (= sp. aff. *Trigonicus*sensu KLAPPER et al. 1978). Here, the Ardenne-Moroccan composite of GOUWY & BULTYNCK (2002) is most helpful. 4. Imprecise correlation between the polygnathid and icriodid successions in order to date neritic successions. Our data indicate longer regional ranges for some taxa than given in the regional composite of BELKA et al. (1997).

Ling. pinguis and *zieglerianus* date Bed 3a clearly as *partitus* Zone but the zonal marker are absent. *I. struvei* and *Ling. linguiformis* have mostly been regarded as marker species for the *costatus* Zone (e.g., BELKA et al. 1997, KONONOVA & KIM 2003) but the composite of GOUWY & BULTYNCK (2002) showed their joint earlier appearance in the upper *partitus* Zone. There is no evidence that the event interval falls already in the *costatus* Zone. *Ling. coopericooperi* is mostly thought to disappear at the end of the *partitus* Zone (BELKA et al. 1997, BERKYOVÁ 2009) but there is a short overlap with the oldest *Po. costatuscostatus* at BouTchrafine (BULTYNCK 1985). However, more material from the event interval of Jebel Amelaneis desirable, especially of the *costatus* lineage.Bed 5 falls in the upperpart of the *partitus* Zone.This is in agreement with the record of *Po. sogdianensis*, which ranges from the *partitus* into the lower *costatus* Zone (BARDASHEV 1992).*Po.*cf. *robusticostatus* (intermediate to *Ling. zieglerianus*) of Bed 6b occurs in Germany in the *partitus* Zone (WEDDIGE 1977; compare the range of the *robusticostatus* Group in KLAPPER et al. 1978).

*I. amabilis*and *Po.angusticostatus* date Bed 10a as *costatus* Zone. However, the first species has a slightly lower range in the composite of GOUWY & BULTYNCK (2002) and it is strange that the index taxon is still absent in a rich assemblage whilst *Po. costatuspartitus* does occur rarely. The presence of *Ling. coopericooperi* and *Caud.culicellusculicellus* prove a level that cannot be younger than the base of the *costatus* Zone. The stratigraphical significance of *I.*aff.*regularicrescens* is not clear; typical *I. regularicrescens* enter well above the base of the *costatus* Zone (BULTYNCK 1985, BELKA et al. 1997, GOUWY & BULTYNCK 2002).

The sudden dominance of *Po.costatus costatus* in the nodular Bed 13b provides evidence for a strong facies control on the distribution of this marker form. Associated *Ling. pinguis, Ling. bultyncki, Ling. zieglerianus,* and *Po.costatus partitus* prove that the so far investigated section top is still rather low in the *costatus* Zone (see BELKA et al. 1997).

4. GONIATITE STRATIGRAPHY

The basal, nodular part at Jebel Amelane (Bed A_1 in BECKER & HOUSE 1994) falls in the lower *An. Simulans* Zone with *Sellanarcestes* (LD IV-D1). The main, upper part of the cliff represents LD IV-D2 without that genus, which includes the *An. lateseptatus* Zone of KLUG (2002a). The first *Fidelites* is a marker of the lower Eifelian *Foorditesveniens* Zone (MD I-B). Currently, the Lower/Middle Devonian boundary has not been fixed locally The first *Subanarcestes* from Bed 3a extends the lower range of the genus in the Tafilalt (see range chart in KLUG 2002a).

The event interval is poor in goniatites but elsewhere (section El Atrous North) the first *P. juglerі*was found in equivalents of Bed 4c. This explains goethitic (originally pyritic) *Pinacites* faunas from a black shale level of the margin of the Tafilalt Basin to the east (e.g., section Tisserdimine). In agreement with data from other sections in KLUG (2002a), the lower Eifelian *Pinacites* Zone (MD I-C) ranges to the top of the sectionshown in Fig. 3.

5. CONCLUSIONS

The Jebel Amelane section provides impressive evidence for a sudden flooding and eutrophication of the otherwise oligotrophic western Tafilalt Platform during the brief Choteč Event Interval. The prevailing amorphous organic matter may indicate cyanobacteria as the main, suddenly blooming primary producers. Hypoxic to anoxic conditions arose from the biodegradation of the massive input of organic matter under conditions with low bottom circulation. The so far available conodont data suggest that this event occurred at the end of the *partitus* Zone, not in the lower *costatus* Zone. It is unlikely that the dark and shaly event phase of the Bohemian type region has a different (younger) age. Its tentative placing in the *costatus* Zone is based on the entry of *Po*. aff. *trigonicus*, a supposed alternative marker (KLAPPER et al. 1978, BERKYOVÁ 2009). However, it has been named as *Po. praetrigonicus* in BARDASHEV (19xx) and allegedly enters in Tadzhikistan in the *partitus* Zone. Additional collecting may resolve the question.

A second, smaller-scale deepening, above the regressive marker unit, enabled the bloom of the index taxon of the *costatus* Zone. This second lower Eifelian transgression has also been noted by KLUG (2002a) and correlated with the onset of Depophase 1d sensu JOHNSON et al. (1985).With respect to the pre-event shallowing upwards in Bed 3b, it seems important to recognize the Choteč Event as an equally important eustatic pulse (e.g., basal TST of Sequence Eif-1 in eastern North America, VERSTRAETEN2007).

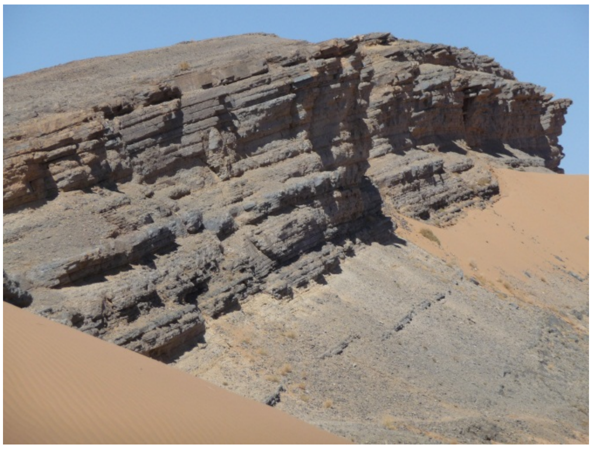

Fig. 8. LateralEifelian to Givetian cliff at Jebel Amelane, with the Kačak Event Interval as incision in the middle.

6. HIGHER STRATA

A complete succession through the higher Eifelian and Lower/Middle Givetian is exposed in the steep cliff just a few decameters to the west (Fig. 8, section of BECKER & HOUSE 1994, 2000; see also KLUG2002a). It includes the Kačak Event Interval but there are no associated dark shales or limestones. Similar as at the basal Givetian GSSP at Jebel MechIrdane just to the south (WALLISER et al. 1995), it is possible to collect the oldest maenioceratids (*Bensaiditeskoeneni*), with evidence that they co-occur with the youngest but very rare *Cabrieroceras*. Higher in the cliff and low in the Middle Givetian, there is a peculiar, thin goniatite coquina dominated by a new species of *Sobolewia*. The Taghanic Crisis Interval has been described by ABOUSSALAM (2003), with corrections in ABOUSSALAM & BECKER (2007). The Givetian-Frasnian boundary falls in an unconformity; the wide-spread Frasnes Event beds were mostly removed during a regression at the end of the Lower Frasnian. The Middle Frasnian is well exposed and partly rich in beloceratids (BECKER & HOUSE 2000). There is an impressive sharp base of the very fossiliferous Kellwasser Beds (Fig. 9), which top is an indistinctive but significant erosion surface. Late lower Famennian black goniatite limestones are directly encrusted; their stratigraphy was studied by BECKER (1993).

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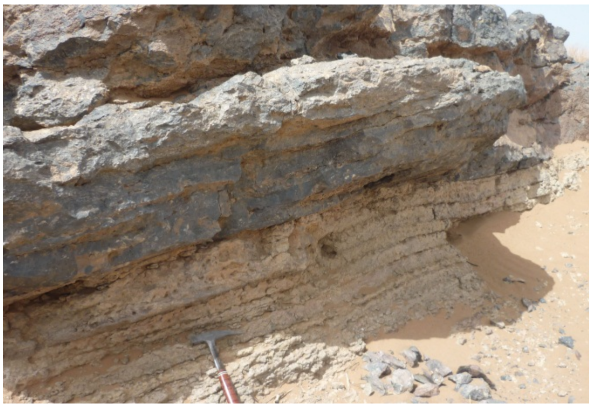

Fig. 9. Sharp contact between light-grey, nodular *Beloceras* Beds and the dark, bluish-grey Kellwasser Beds at Jebel Amelane.

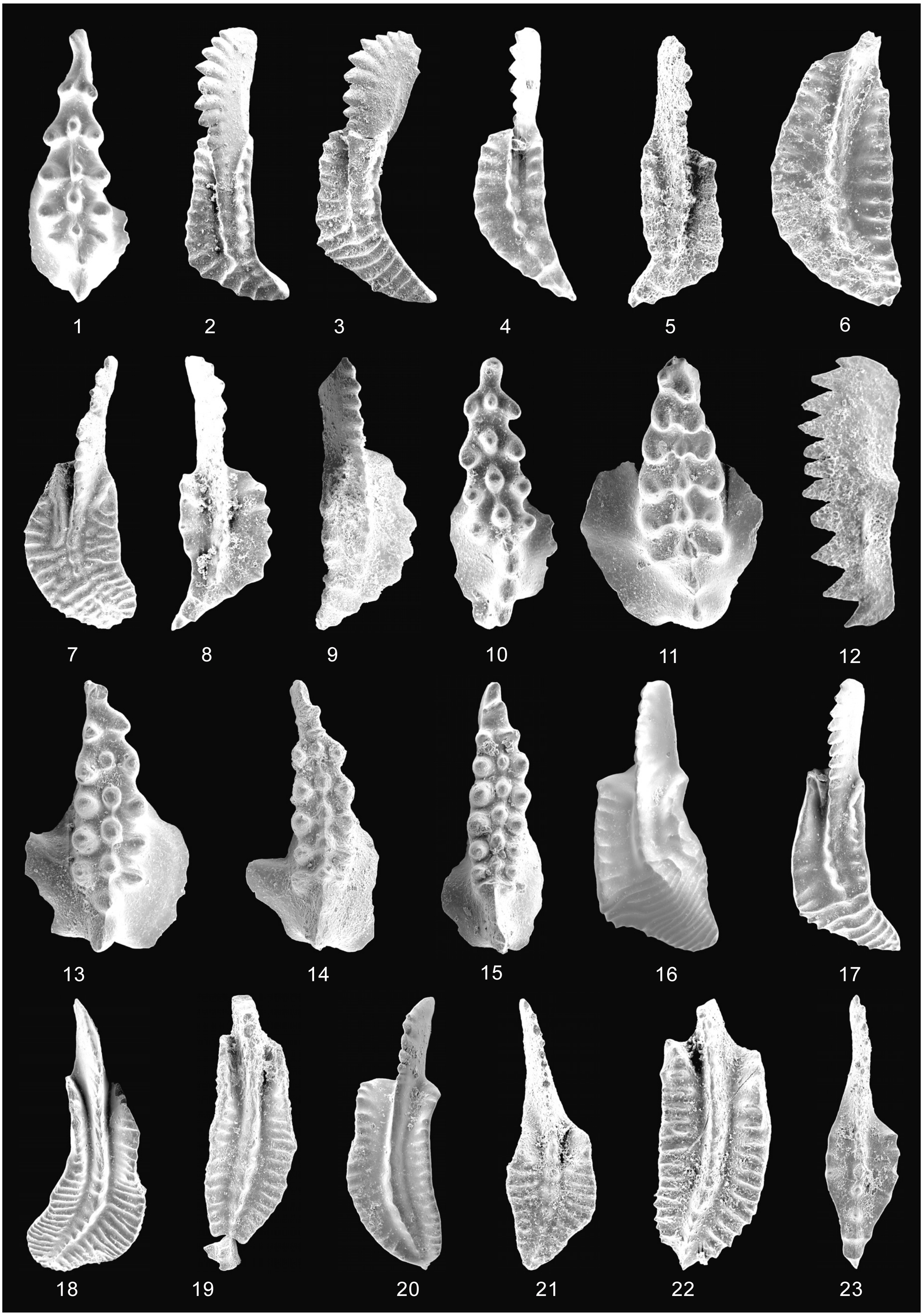

Fig. 7. Conodonts from around the Choteč Event at Jebel Amelane. 1-5 = Bed 3a, 6 = Bed 4b, 7-8 and 22 = Bed 5, 9 = Bed 6b, 10-16 and 23 = Bed 10a, 17-21 = Bed 13b. **1.***I. struvei*, **2-3, 17.***Ling. bultyncki*, **4, 7.***Ling. pinguis*, **5, 8.***Ling. zieglerianus*, **6.***Ling. coopericooperi*, **9.***Po.*cf. *robusticostatus*, **10.***Caud.culicellusculicellus*, **11.***I. amabilis*, **12.**"*Ozarkodina*" n. sp., **13.***I. anterodepressus*, **14.***I. cornigercorniger*, **15.***I.*aff.*regularicrescens*, **16.***Ling. linguiformis* $\gamma 2$, **18.***Po.costatuscostatus*, **19-20.***Po.costatuspartitus*, **21.***Po.praetrigonicus*, **22.***Po.sogdianensis*, **23.***Po.angusticostatus*.

LOWER EMSIAN STRATIGRAPHY AT JEBEL IHRS (WESTERN TAFILALT PLATFORM)

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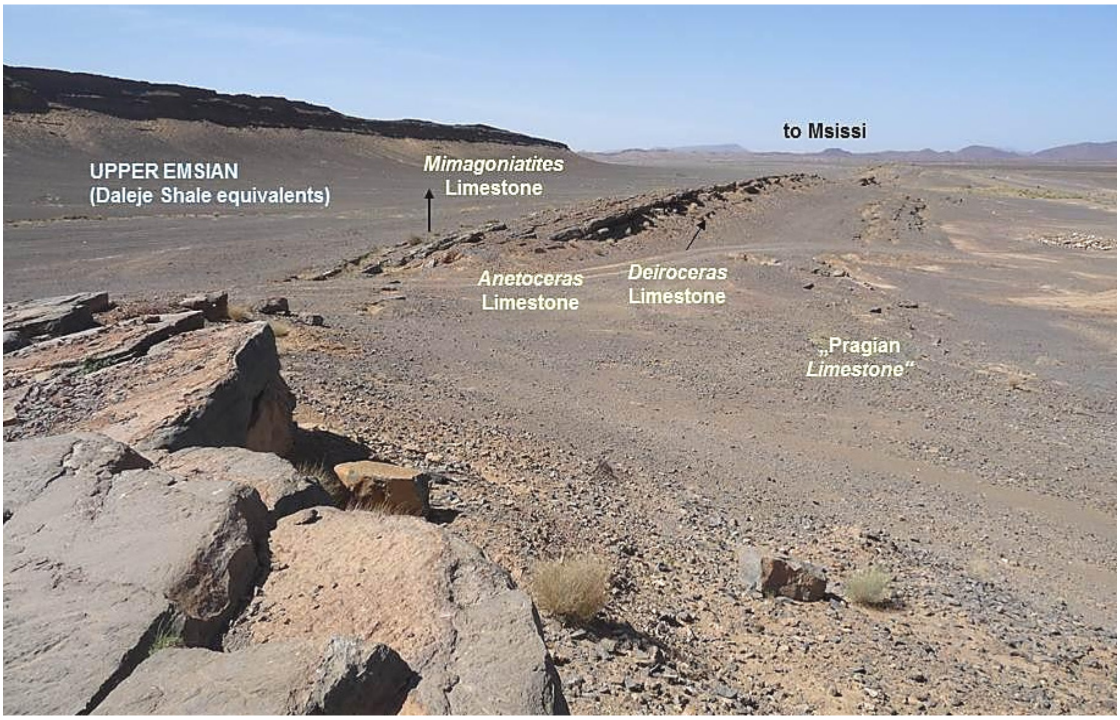

Fig. 1. View to the west, showing the extensive outcrop conditions and Emsian lithostratigraphy at Jebel Ihrs.

1. INTRODUCTION

The Jebel Ihrs lies in the western Tafilalt and is the continuation of the ca. W-E running Jebel Amelane, west of the pass formed by the main road from Rissani to Mississi (sheet Erfoud, NH-30-XX-2; Figs. 1-2), The local Pragian/Emsian litho- and biostratigraphy was first outlined by ALBERTI (1980, 1981) and correlated with the succession at Bou Tchrafine to the east. Because of its established dacryoconarid succession, last revised in ALBERTI (1998), it was chosen as one of the regional reference sections for a new investigation of lower Emsian conodont stratigraphy in the Tafilalt (ABOUSSALAM et al. in prep.). The strike of the exposed, gently dipping beds is almost parallel to the road (Fig. 1). The precise sampling spot can be recognized opposite to the tart-like Mdoura outcrop by a small hut at the road and a track that winds upwards through the ridge. The GPS coordinates at the section base are N 31° 16´ 11.4´´, W 4° 24´ 8,8´´. BECKER & ABOUSSALAM (2011) provided a preliminary summary of the lower Emsian litho- and conodont stratigraphy, which is updated here. Microfacies and the abundant dacryoconarids suggest in general a hemipelagic setting with alternating pelagic platform carbonates and slightly deeper deposited marls but the conodont biofacies suggests much shallower water. The exposed succession can be summarized as follows (Fig. 3), following the new regional unit terminology:

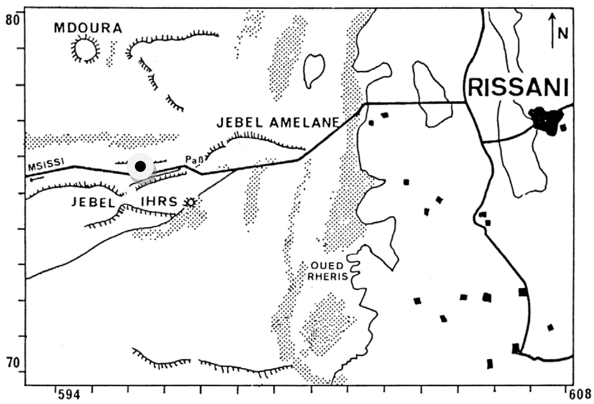

Fig. 2. Position of the measured section in the western Tafilalt, W of Rissani (BECKER 1993: fig. 48).

Unit E. "Pragian Limestone"

The wide plain beneath the road and towards Gara Mdoura has no outcrop of the Pragian Marls and Shales and of the basal "Pragian Limestone". Bed 2 at the base of the measured section is a solid, light-grey nodular limestone without macrofauna (Fig. 4).

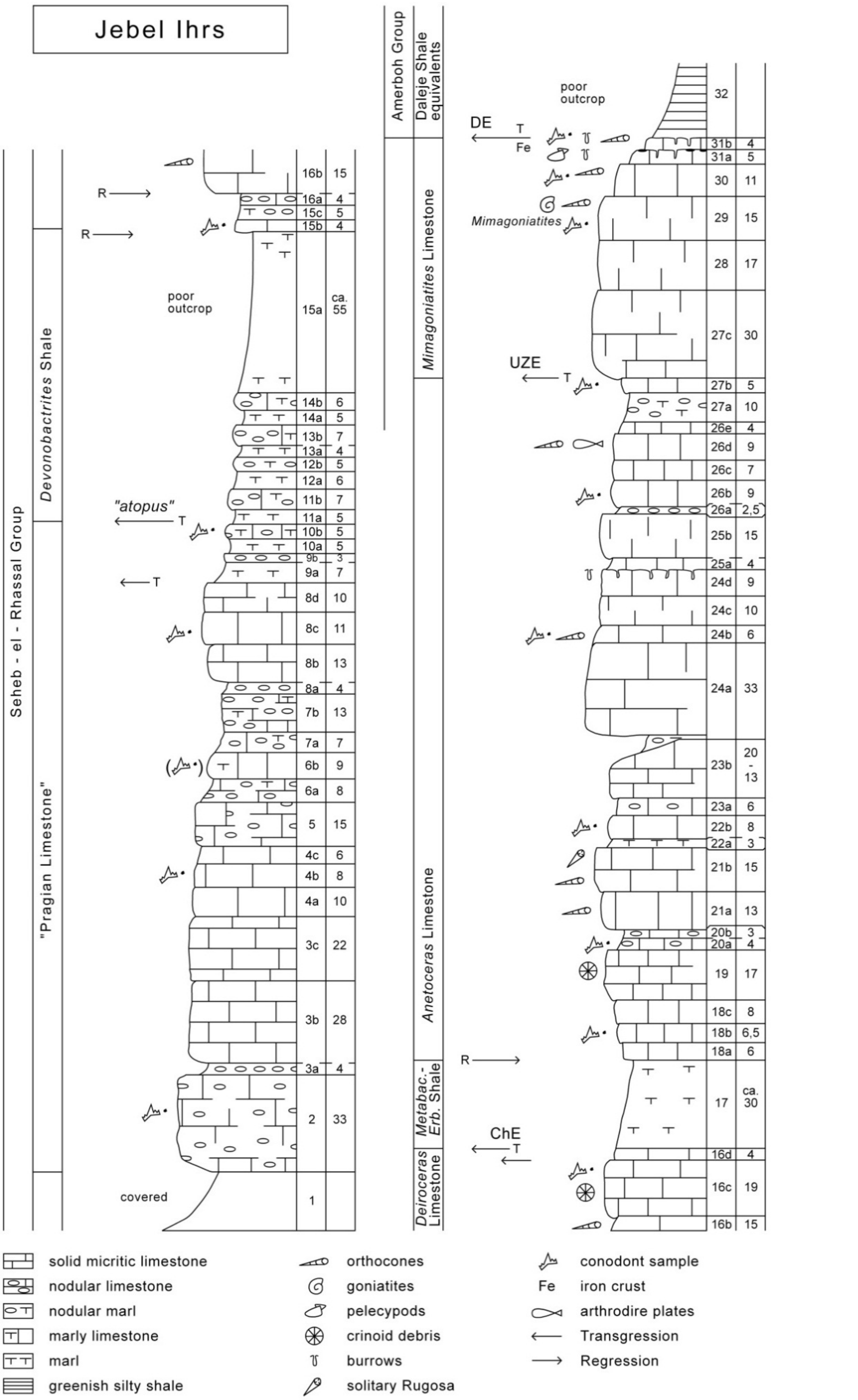

Fig. 3. Lithostratigraphy, position of conodont samples, and relative sea level fluctuations at Jebel Ihrs. "*atopus*" = supposed correlative of Bohemian "*atopus* Shale" with last monograptids, ChE = Chebbi Event, UZE = Upper Zlichov Event, DE = Daleje Event.

Fig. 4. “Pragian Limestone” at Jebel Ihrs (Bed 2 at the base), grading upwards into the locally unfossiliferous *Devonobactrites* Shale. The *Anetoceras* Limestone forms the platy limestone cliff in the mid-ground, Eifelian to lower Givetian limestones the main cliff in the background, with the Chotec Event level at the base of the solid rocks and the Kacak Event level as a small plateau.

It contains rare belodellids as the only conodonts (Fig. 7.1). These are typical in the Middle Devonian for neritic settings but obviously had a wider ecological range in the Lower Devonian. Beds 3-4 are rather similar, grading into more nodular limestone (Beds 5-7). Bed 8 is slightly darker and more solid. Overall, there are hardly any conodonts in the “Pragian Limestone”. Bed 4b yielded a few more belodellids and a single *Caud. celtibericus*. Thin-bedded alternations of marl and nodular limestone (Beds 9-14) form a gradual transition to the mostly covered *Devonobactrites* Shale. Despite this gradual deepening, the restricted belodellid biofacies continues in Bed 10b.

Age: It is reasonable to assign the “Pragian Limestone” to a local Lower *Belodella* Ecozone (BECKER & ABOUSSALAM 2011). Correlation with the upper Pragian-basal Emsian *celtibericus* Zone is provided by Bed 4b. CARLS & VALENZUELA-RÍOS (2002) used *Caud. celtibericus* as a marker of their Conodont Step 17, which is also the level of the first *Eolinguipolygnathus excavatus* Morphotype 114, the proposed future basal Emsian index taxon (CARLS et al. 2008). But the species has a lower range in Bohemia (SLAVIK 2004) and in a Tinejdad olistolith described by RYTINA et al. (this volume).

According to ALBERTI (1981, 1998) the oldest *Guerichina africana* overlap in the “Pragian Limestone” of Jebel Amelane with the last *Nowakia (Turkestanella) acuaria acuaria*. *Alainia?* cf. *hercyniana* is more typical for the upper part. These data suggest that the “Pragian Limestone” falls in the basal Emsian sensu the Zinzilban GSSP (see dacryoconarid-conodont correlation in KIM 2011) but the future, revised GSSP level may fall within it.

Unit F. *Devonobactrites* Shale

Not really exposed. It falls in the *Devonobactrites obliqueseptatum* Zone (LD III-A), based on the diverse “Faunule 1” sensu KLUG et al. (2008). ALBERTI (1998) noted the last *Now. (Turk.) anteacuaria* in the transitional interval from the “Pragian Limestone”. *G. africana* was said to reach the upper part of the unit. Therefore, its base, a significant regional deepening episode, is older than the Lower Zlichov Event level of Bohemia sensu CHLUPÁČ & KUKAL (1986), which lies above the range of *Guerichina*.

Unit G. *Deiroceras* Limestone

Three thin limestones beds (Beds 15b-16a) underlie the main, more massive and condensed (only 38 cm thick) *Deiroceras* Limestone (Beds 16b-d), which is rich in orthocones up to 2.5 cm in diameter. The base (Bed 15b) yielded only *Caud. celtibericus*. Bed 16c belongs to a radically different *Criteriognathus*-

Latericriodus-polygnathid biofacies; the sudden absence of *Caudicriodus* is remarkable. This assemblage suggests a deepening trend at the top of the unit. *Criteriognathus miae* and *Latericriodus bilatericrescens* (with the typical subspecies and *bil. multicostatus*) dominate. But there are also early polygnathids, notably *Eol. excavatus* (typical morphotype, Fig. 7.5, and Morphotype 114, Fig. 7.6), *Eol*. cf. *excavatus* (Figs. 7.7-8, = aff. *gronbergi* in BECKER & ABOUSSALAM 2011), and a new species with strongly curved, flatter platform, dense transverse ribbing, and linguiform posterior platform. This *Eol*. n. sp. aff. *pannonicus* (Figs. 7.9-11) resembles the somewhat older true *Eol. pannonicus* from Uzbekistan but lacks a small rostrum on the anterior platform.

Age: The basal part appears to fall locally still in the *celtibericus* Zone but the basal *Deiroceras* Limestone of El Khraouia (BECKER et al. this vol.) falls in the (higher) *excavatus* M114 Zone of the polygnathid and *bilatericrescens* Zone of the icriodid succession. The latter equals Conodont Step 19 of CARLS & VALENZUELA-RÍOS (2002). The top of the unit includes the marker species of both zones; the level of *Eol. gronbergi* has not yet been reached. The new, rather frequent *Polygnathus* has biostratigraphical significance in the upper part of the *excavatus* M114 Zone. This is also the level of the oldest *Now. (Now.) zlichovensis maghrebiana* and *Now. (Dimitriella) praesulcata* (ALBERTI 1998; lower *zlichovensis* Zone). This assemblage leaves the much lower range of the second species in Uzbekistan (through much of the *acuaria* Zone, KIM 2011) unresolved.

Unit H. *Metabactrite-Erbenocerass* Shale

Unit H is locally very thin (Bed 17) and unfossiliferous. At Bou Tchrafine (BECKER & HOUSE 2000) and many other sections (KLUG 2001, KLUG et al. 2008) it contains “Faunule 2” with the oldest goniatites (*Metabactrites-Erbenoceras* Zone, LD III-B). This episode of eustatic rise, the Chebbi Event Interval (see BECKER & ABOUSSALAM 2011), saw a nektonic revolution of the pelagic ecosystems after the extinction of the megaplanktonic graptolites (KLUG et al. 2010). *Now. (Now.) zlichovensis* disappears within Unit H (ALBERTI 1998).

Fig. 5. Bed numbering of the *Anetoceras* Limestone at Jebel Ihrs, with a distinctive incision below Bed 24a.

Unit I. *Anetoceras Limestone*

The *Anetoceras* Limestone begins with three thin limestones (Beds 18a-c), which belong (Sample 18b) to a *Latericriodus* biofacies with ca. 30 % *Caud. celtibericus* and rare *Crit. miae* (Fig. 7.2). The polygnathids have disappeared, in accordance with an overall regressive trend. The overlying more solid Bed 19 is followed by thin, nodular limestone with very abundant conodonts (more than 800 specimens in ca. 2.5 kg limestone of Bed 20b). *Lat. bilatericrescens latericrescens* is even more dominant (88.6 % of the assemblage) than below, with rare representatives of the two other subspecies (*multicostatus* and *gracilis*). *Crit. miae* is the second commonest species and there are very rare *Eol. excavatus* M114. The main innovation is the entry of *Lat. latus* and of very rare *Lat. beckmanni beckmanni*.

The middle part of the *Anetoceras* Limestone begins with solid to massive, light-grey limestone (Beds 21a-b). Bed 22b yielded a *Latericriodus-Criteriognathus* assemblage with somewhat more common *Lat. beckmanni beckmanni* and the first *Lat. beckmanni sinuatus. Crit. miae* is still much more frequent than the first *Crit. steinhornensis*. There is also a record of the oldest *Eol. catharinae*. Bed 23b forms a distinctive recession in the lower cliff (Fig. 5).

The upper part of the *Anetoceras* Limestone starts with the very massive, ledge-forming Bed 24a. Bioturbated, corroded surfaces, for example at the top of Bed 24d, indicate minor sedimentation breaks.

Some beds have many orthocones (Bed 24b). Subsequent limestones become thinner and there are large placoderm bones in Bed 26d. Conodonts from Bed 26b illustrate how the biofacies has changed further, into an almost pure *Criteriognathus* biofacies, but *Crit. miae* has disappeared. The top beds are thin or nodular and, without any evidence in the lithofacies, conodonts are almost lacking, leaving just a few belodellids.

Age: Bed 18b falls in the top part of the *bilatericrescens* Zone of the icriodid succession. A distinctive new zone with *Lat. latus* begins with Bed 20b. *Lat. beckmanni beckmanni* can serve as an alternative index species. This important level has also been recognized at Bou Tchrafine (BULTYNCK & WALLISER 2000) and further to the east at Ouidane Chebbi (BELKA et al. 1999). It equals Conodont Step 20 of CARLS & VALENZUELA-RÍOS (2002) and marks in the Amorican Massive (La Grange Limestone) an interval of the *gronbergi* Zone, just before the entry of *Eol. nothoperbonus* (BULTYNCK 1989). Bed 22b falls in the lower part of the *steinhornensis* Zone, which is Conodont Step 21 sensu CARLS & VALENZUELA-RÍOS (2002). At the same time it equals the Upper *gronbergi* Zone sensu BULTYNCK (1989), which should better be named as *catharinae* Subzone. The conodont-poor upper part of the *Anetoceras* Limestone represents a local Upper *Belodella* Ecozone.

ALBERTI (1981, 1998) noted that the lower part of the *Anetoceras* Limestone at Jebel Amelane (= Ihrs) has *Now. (Now.) praesulcata* and *Now. (Now.) tafilaltana*. *Now. (Now.)* cf. *praecursor* follows slightly higher and *Now. (Now.) barrandei* near the top of Unit I (ALBERTI 1981, 1998). This dense sequence suggests a strong condensation of Unit I.

Unit J. *Mimagoniatites* Limestone

The *Mimagoniatites* Limestone consists locally of massive, up to 30 cm thick (Bed 27c), bluish, middle-grey limestones. Large orthocones become very common in the upper part (Beds 29-30) and include actinoceratids and orthoceratids (Fig. 6C). The upper surface of Bed 29 displays several *Mimagoniatites* sp. (Fig. 6A). In sharp contrast to the distinctive goniatite influx there are no conodonts at all. This proves the ecological independence of both pelagic fossil groups. The thin two beds above the main *Mimagoniatites* Limestone (Beds 31a-b) are fossil-rich (many orthocones, large *Panenka*, placoderm plates, thamnoporid corals, Fig. 6B) and lighter-grey. Strong bioturbation and hematite incrustations testify a condensed and partly interrupted deposition. The pure polygnathid biofacies of Bed 31b suggests a deepening trend towards the top. A stepwise deepening from Zlíchovian into Dalejan strata is also typical for the Barrandian (FERROVÁ et al., 2013).

Age: Bed 29 falls in the *Mimagoniatites* Zone (LD III-D sensu BECKER & HOUSE 1994) high in the lower Emsian. The alleged record of *Mimosphinctes* at Jebel Amelane (= Ihrs), the index genus of the final lower Emsian (LD III-E), by MASSA (1965) could never be verified (see KLUG 2001). The still incomplete sampling and the shallow-water belodellid biofacies prevent the recognition of the *inversus* Zone. At the top of Unit J (Bed 31b), the *laticostatus* is typically

Fig. 6. Fossil cross-sections from corroded bedding surfaces of the *Mimagoniatites* Limestone at Jebel Ihrs. **A.***Mimagoniatites* cf. *fecundus*, Bed 29, **B.** Thamnoporid corals, Bed 31. **C.** Large actinoceratid (probably *Ormoceras*) and smaller orthoceratid (below), Bed 29.

developed, including the index species (Figs. 7.12-13) and close relatives of Ling. vigierei (Figs. 7.14-15). The boundary between the *Now. (Now.) barrandei* and *elegans* Zones seems to lie within the *Mimagoniatites* Limestone (ALBERTI 1981) but new and more detailed data are required. Rare *Now. (Now.) cancellata* were reported from the top limestone but have not been re-studied in the light of transitional forms.

Unit K. Daleje Shale Equivalents

Locally there is a sharp contact between the top *Mimagoniatites* Limestone and the thick, greenish-grey, partly silty shales, which are an equivalent of the Daleje Shale of Bohemia. The sudden sea-level rise at the base is taken as the main Daleje Event and as a marker for the future upper Emsian substage (see different view in FERROVÁ et al. 2012).

Age: At Jebel Ihrs there are no ammonoid faunas, which elsewhere in the Tafilalt consist of abundant *Gyroceratites, Rherisites,* and the oldest anarcestids (zones LD IV-A to IV-C, BECKER & HOUSE 1994, KLUG 2002, WEBSTER et al. 2005).

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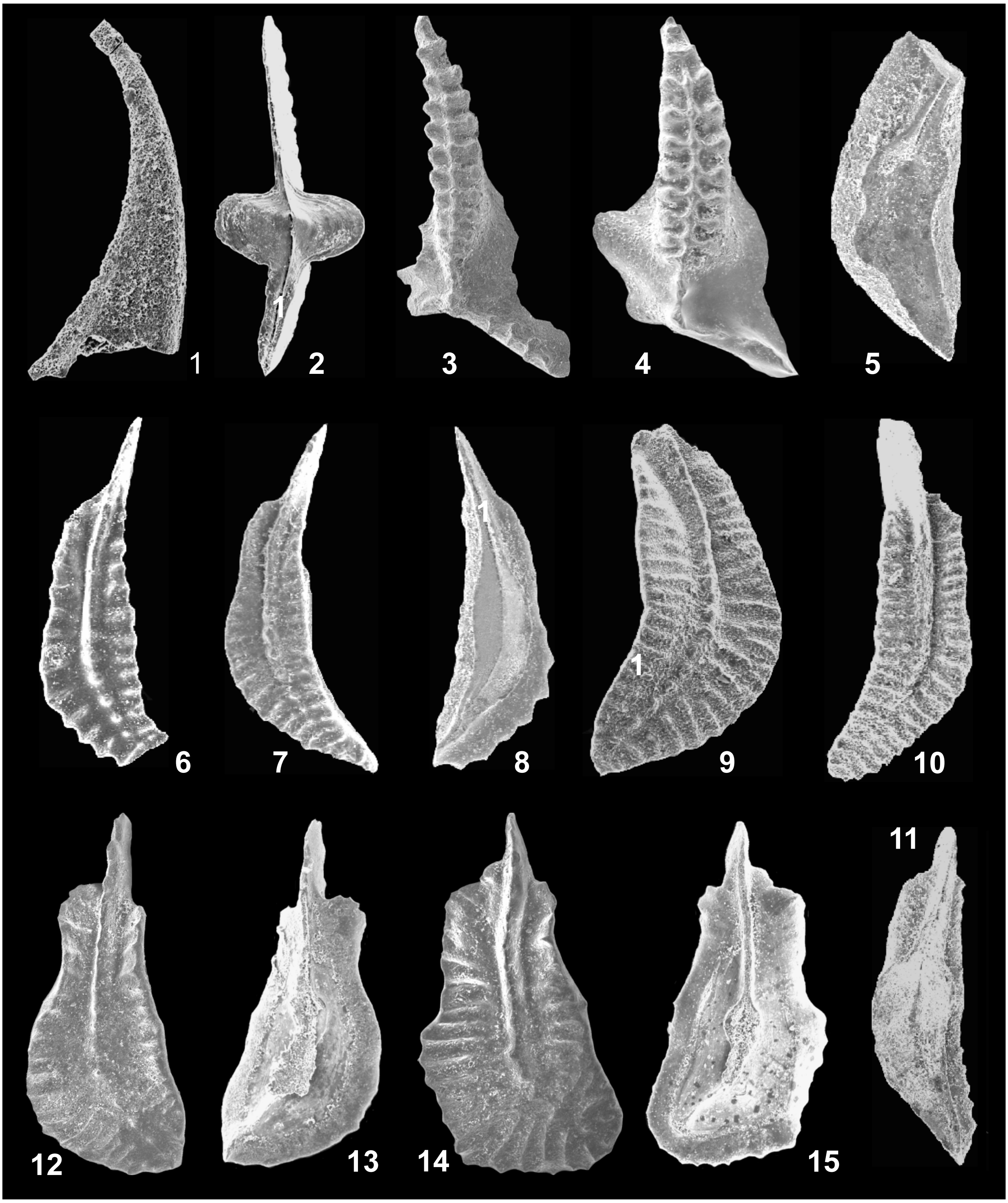

Fig. 7. Lower Emsium conodonts from Jebel Ihrs (extracted from BECKER & ABOUSSALAM 2011). **1.** *Belo. resima*, Bed 2, “Pragian Limestone”, **2.** *Crit. miae*, Bed 18b, lower *Anetoceras* Limestone, **3.** *Lat. bilatericrescens multicostatus*, Bed 18b, **4.** *Lat. bilatericrescens bilatericrescens*, Bed 18b, **5.-6.** *Eol. excavatus*, Bed 16c, top *Deiroceras* Limestone, lower view of wide specimen (5, Morphotype 114), upper view of specimen with broken posterior tip (6, possibly a typical morphotype), **7-8**. *Eol*. cf. *excavatus*, Bed 16c (“aff. *gronbergi*” in BECKER & ABOUSSALAM 2011), upper surface (7) as in Morphotype 114, shallow basal cavity (11) slightly intermediate to *Eol. abyssus*, **9-11.** *Eol*. n. sp. aff. *pannonicus*, Bed 16c, typical, wide morphotype (9) and a narrow morphotype (10-11, with upper and lower views), slightly transitional to *Eol. excavatus*, **12-13.** *Ling. laticostatus*, Bed 31b, lower and upper views, **14-15.** *Ling*. cf. *vigierei*, Bed 31b, upper and lower views.

MIDDLEGIVETIAN – MIDDLE FRASNIAN EVENT STRATIGRAPHY AT MDOURA-EAST (WESTERN TAFILALT)

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1. INTRODUCTION

The condensed Givetian to Frasnian succession of the hemipelagic Tafilalt Plattform is characterized by a succession of marker beds/unit, which can be well correlated through the region. But sea-level fluctuations in combination with the local palaeotopography led to non-deposition or post-sedimenary erosion during regressive episodes. Global event levels are marked by often dark-grey to black marls and fossil-rich limestones, which represent episodes of increased nutrient recycling, blooms of (mostly not preserved) primary producers and benthonic, planktonic or nektonic consumers, increased burial of organic matter, and reduced sea-floor oxygenation due to the bacterial degradation of the increased C_{org}-supply. The best examples are the *Pumilio* Events (LOTMANN 1990), Frasnes Event and Kellwasser Beds (e.g., BUGGISCH & CLAUSEN 1972, BENSAID et al. 1985, WENDT & BELKA 1991) but there are additional, less well-studied levels, for example in the late Lower to Middle Frasnian (Timan to Rhinestreet Events, e.g., BECKER et al. 1993, HOUSE et al. 2000). Eutrophic episodes can potentially be recognized in the carbon isotopes (e.g., ABOUSSALAM 2003).

Mdoura-East is the name used for a sequence of low hills just east of the distinctive "Mdoura Tart" and north of Jebel Amelane (see maps in BECKER & ABOUSSALAM and ABOUSSALAM & BECKER, this vol.). They expose a condensed succession from the upper Emsian to lower Famennian, with good representation of several event/crisis intervals. For parts of the Frasnian we present here a preliminary carbon isotope curve. Previous work dealt with the lower Famennian (BECKER 1993), Upper Givetian (ABOUSSALAM & BECKER 2007), and Taghanic Event (ABOUSSALAM & BECKER 2011). Results from these studies are summarized and combined with new data.

2. MIDDLE GIVETIAN / TAGHANIC CRISIS

The main Lower/Middle Givetian cliff has not yet been sampled in detail. The studied Middle Givetian starts just above the Upper *pumilio* level:

Middle *Sellagoniatites* Beds (MSell): Bed 0, 8 cm bluish to reddish-grey solid limestone.

***Maenioceras* Marl (MM)**: Beds 1-7, ca. 1m alternating nodular marl and nodular limestone (four subcycles), reddish-grey (Bed 1), light-grey (Bed 2), or yellowish weathering (Bed 7), with *Sell. waldschmidti* (Bed 5) and rare maenioceratis. The conodont fauna from the top is dominated by *Linguipolygnathous linguiformis* in association with various long-ranging Lower/Middle Givetian species. Deepening (Depophase If-Win) at the base.

(Local) **Marker Styliolinite**: Thin (max. 8 cm, Bed 8), ledge of dark-grey styliolinid packstone, which suggests an episode of increased trophication. It represents the initial "Taghanic Prelude". Records of *Polygnathus ovatinodosus* and *Tortodus*aff.*wedigei* from Jebel Amelane prove that the upper part of the *ansatus* Zone has been reached.

Upper *Sellagoniatites* Beds (USell): Solid, massive marker limestone (ca. 50 cm) with very sharp base and four subunits (Beds 9a-d, Fig. 1), which represent separate shallowing upwards cycles with gradual or sharp (Bed 9c) change from bioturbated styliolinid mudstone to wacke-packstone (Pl. 1, Fig. 1). Iron microstromatolites occur in Bed 9c, especially at the top. Conodont faunas (especially rich in Bed 9b) are still dominated by *Ling. linguiformis*, followed by *Po. Timorensis* (Pl. 2, Fig. 4) and *Po. ansatus* (Pl. 2, Fig. 1). Bed 9b yielded the globally youngest *Bipennatus* (Pl. 2, Fig. 2).The polyphase interval of the Upper *Sellagonitites* Beds is correlated with the sea-level low below the initial Taghanic Onlap level.

Fig. 1. The massive Upper *Sellagoniatites* Bed overlying sharply (with a minor unconformity) the recessive *Maenioceras* Marl. The covered interval above begins with the Upper Tully level (Upper Taghanic Crisis Interval).

Lower/Middle Taghanic Crisis Interval (LT/MT): Thin (4.5 cm) grey limestone encrusted directly onto the USell. It is a bioturbated styliolinid wackestone that grades upwards into a reddish (hematite-enriched) packstone with phacopids, microstromatolites, and mudclasts (Pl. 1, Fig. 2). The latter indicate strong condensation and the reworking of missing strata. This masks the initial Taghanic Onlap (DepophaseIIa-Tagh) in the western Tafilalt. The sudden incoming of various *Tortodus*-species (Pl. 2, Figs. 5-7) is regionally typical for the LT/MT (ABOUSSALAM 2003, ABOUSSALAM & BECKER 2011) but similar and contemporaneous *Tortodus* blooms are known from North America (“Prout Dolomite” sensu SPARLING 1999) and Spain (Pyrenees, LIAO & VALENZUELA-RÍOS 2008).A similar incrustation on top of the USell yielded at Seheb-el-Rhassel to the west very rare *“Po.” alveoliposticus*” (ABOUSSALAM 2003), the marker conodont that enters in the Lower Tully Limestone (e.g., ZIEGLER et al. 1976) and correlative strata of North America.

Fig. 2. Transgression of the Upper Tully Level with the oldest *Pharciceras* on the USell (Bed 9), followed by more solid limestone (main Lower Marker Bed, Bed 13) higher in the Upper Givetian.

Aff.*amplexum* Bed (am): Ca. 7-8 cm grey, slightly reddish weathering marl and nodular limestone (Beds 10a-c) with the oldest *Pharciceras* (*Ph*. aff. *amplexum*, Bed 10b) and *Epitornoceras mithracoides*, a fauna of the basal *Pharciceras* Stufe (MD III-A, topmost Middle Givetian). The conodont association is not as rich as in the LT/MT (10 versus 15 taxa; the tortodids have mostly disappeared) but *“Oz.” semialternans* (Pl. 2, Fig. 8), the index of the Upper Tully level (Upper Taghanic Crisis Interval), can be obtained in sufficient numbers. Very unusual is the rare occurrence of *Latericriodus latericrescens latericrescens* (Pl. 2, Fig. 9), which, outside North America, is normally almost exclusively found at the Upper *pumilio* level (see Oued Ferkla, WARD et al., and El Khraouia, BECKER et al., this vol.). It occurs in the Upper Tully Limestone, too (ZIEGLER et al. 1976), and also at the same level in Cantabria (Member C of Candas Formation, GARCÍA-LOPEZ 1987). The base of Bed 10 marks the transgressive beginning of DepophaseIIa-UT sensu ABOUSSALAM & BECKER (2011).

3. UPPER GIVETIAN

***Erraticus* Beds (err)**: Ca. 17 cm marls and nodular limestone (Beds 10d-h) with the basal Upper Givetian (sensu ABOUSSALAM & BECKER 2002) index goniatite *Mzerrebite serraticus* (Bed 10d, MD III-B1, Fig. 3). There are various marker species of the (Lower) *hermanni* Zone, such as *Schmidtognathus hermanni*(Pl. 2, Fig. 10), *Schm. pietzneri*(Pl. 2, Fig. 13), *Po. dubius*(Pl. 2, Fig. 12), *Po. limitaris*(Pl. 2, Fig. 11), and *Po*. aff. *pennatus*. The previously dominant *Linguipolygnathus* has disappeared locally completely; only *Po. ovatinodosus* continues from the beds below. This gives a very distinct conodont extinction at the top of the Taghanic Crisis Interval. *Schm. wittekindti* enters slightly higher, in Bed 10g. The relatively inconspicuous base of Bed 10d marks the eustatic Geneseo Transgression or the beginning of DepophaseIIa-Gen (sensu ABOUSSALAM & BECKER 2011).

Fig. 3. *Mz. erraticus*, the regional Upper Givetian (MD III-B) index goniatite, from Bed 10d (61 mm dm).

***Lunupharciceras* Bed (Lu)**: 4-5 cm platy, reddish limestone (Bed 11). The name-giving index goniatite(for MD III-B2) has not yet been found locally. Based on correlation with Seheb-el-Rhassel (ABOUSSALAM 2003), it should correlate with the *cristatus ectypus* Zone, but the *Po. Cristatus* Gp. has neither been found so far. But there is a return of some Middle Givetian species, such as *Po. xylus, Po. ansatus*, and *Icriodus difficilis*.

Fig. 4. Gen. nov.aff. *Extropharciceras* n. sp., loose from near the Lower Marker Bed (105 mm dm).

Lower Marker Bed (sensu BECKER & HOUSE 1994, LMB): 20-25 solid limestone (Beds 12-14, especially Bed 13, Fig. 2), nodular in the lower part (Bed 12) and towards the top. Large oncoceratids are characteristic and there are strange, large pharciceratids (*Pharciceras* n. sp., Gen. aff. *Extropharciceras* n. sp., Fig. 4), which differ from other faunas of MD III-C (see BOCKWINKEL et al. 2009). Conodonts from the base include *Po. cristatusectypus* and *Schm. peracutus* as marker species of the *cristatusectypus* Zone (= old Upper *hermanni* Zone). The re-appearence of *Po.varcus* is interesting. Bed 13 yielded *Klapperinavysotzkii* (Pl. 2, Figs. 14-15),*I. tafilensis* (= *arkonensis* in ABOUSSALAM & BECKER 2007, Pl. 2, Fig. 16, see NARKIEWICZ & BULTYNCK 2010), and *Schm. latifossatus*, but *Schm. pietzneri, Schm.wittekindti*, *Po. dubius, I. difficilis*, and *Po. xylus* are more common. The entry of *Klapperina* marks the (Lower) *disparilis* Zone; the more advanced *Kl. disparilis* itself has been found in the same bed of other sections. Forms resembling *Nothognathella* may represent the Pb element of the *Klapperina* apparatus (Pl. 2, Fig. 17).

Taouzites **Beds**: Ca. 15 cm laterally variable, grey to slightly reddish, nodular limestone (Beds 15-16) with abundant large oncoceratids and goniatites, including common *Synpharciceras, Lunupharciceras*, and *Extropharciceras* (but so far no *Taouzites*). *Po.denglerisagitta* (Pl. 2, Fig. 18), the index of the *sagitta* Subzone, enters in Bed 15. It is the main reason for correlation with the first *Taouzites* level of other sections. There are common associated *Kl. disparilis* and *Po. dubius*. *Kl. disparalvea* enters locally in Bed 16, where *Po. dubius* remainsthe most common species.

Upper Marker Bed (UMB): Distinctive, ca. 25 cm thick level of intensively red nodular marl with masses of mostly corroded goniatites. The loose assemblage includes *Darkaoceras*, the *Ph. Tridens* Gp., *Extropharciceras*, *Stenopharciceras*, *Synpharciceras, Taouzitestaouzensis*, *Pseudoprobeloceraspernai*, *Tornoceras*, and *Epit. mithracoides*. It will be interesting to compare this fauna of MD III-D with the contemporaneous deeper-water association of HassiNebech (BOCKWINKEL et al. 2013). Correlation with the much more massive, typical Upper Marker Bed (e.g., of BouTchrafine, BECKER & HOUSE 2000) is based on its position above the *sagitta* Subzone and below the *Petteroceras* Beds. The rich goniatite assemblage contrasts with the rarity of conodonts; dissolved nodules yielded just one specimen of *Po. denglerisagitta* and *Po. Cristatus ectypus*each. This is one more example for the palaeoecological independence of both fossil groups in pelagic settings.

Petteroceras **Beds (Pett)**: Three beds (total ca. 13 cm, Beds 18a-19) of yellowish, reddish or multi-colored nodular limestone with partly large-sized goniatites, including *Pett. errans, Petteroceras* n. sp. (Fig. 5), other rathermultilobed forms, *“Ponticeras” kayseri, Lunupharciceras*, orthocones, solitary Rugosa (Bed 18a), and *Panenka* sp. This assemblage defines the terminal Givetian *Petteroceras* Zone (MD III-E). Bed 19 is a bioturbated, bioclastic wacke- to packstone with many styliolinids, mollusk debris, and frequent early diagenetic iron mineralisations (Pl. 1, Fig. 3). The bed top can be strongly corroded, which indicates a submarine unconformity.

Fig. 3. *Petteroceras* n. sp. from MD III-E of Mdoura East (> 105 mm diameter)

The conodont association of Bed 18a contains very abundant *Po. xylus*, followed by *Po. dubius, Po. ovatinodosus, Po. dengleridengleri* (Pl. 2, Fig. 19), *“Oz.”sannemannisannemanni* (Pl. 2, Fig. 20), *“Oz.” sannemanniadventa*, and *Schm. peracutus* (Pl. 2, Fig. 21). *Po.aequidivisus* is locally rare. *Po.ordinatus* and *Po. Tafilensis* (Pl. 2, Fig. 22) were first found in Bed 18b, *Schm. gracilis* in Bed 19 (Pl. 2, Figs. 23-24). Correlation of the *Petteroceras* Beds with the *norrisi* Zone is based on very rare specimens of *Skeletognathus* at other sections (ABOUSSALAM & BECKER 2007).

4. LOWER FRASNIAN EVENTS

The Frasnian part of the succession was studied in a lateral section to the north and results have not yet been published. The analysis of conodont faunas is still incomplete and preliminary.

Frasnes Event Beds: The basal Frasnian of the Tafilalt is characterized by wide-spread, distinctive, dark-grey styliolinites and marls with several subunits. The lower part, the Frasnes Event Beds, has been dated as *rotundilobasoluta* Zone (= MN 2 Zone, ABOUSSALAM & BECKER 2007, update of BULTYNCK 1986). Conodonts from the base of Bed 20, however, show that there was a long period of non-deposition or subsequent erosion at Mdoura-East, resulting in an unconformity that spans the MN 1-3 Zones (3/4 of the Lower Frasnian).

Timan Event Beds:Ca. 7.5 cm black, organic-rich styliolinite at the base (Bed 20,Pl. 1, Fig. 4), followed by ca. 2 m, mostly covered dark marls and platy to laminatedstyliolinites with poor other fauna (Beds 21-24b). A conodont sample from Bed 20a yielded *Ancyrodella recta, Ad. pramosica, Ad. africana*, *Mesotaxisasymmetrica*, and *Belodellaresima* (Pl. 2, Fig. 25). This gives a MN 4 Zone age (see composite ranges in KLAPPER 1997), but *Palmatolepistransitans*, the main index species, is absent. The position of the Lower/Middle Frasnian has to be fixed by more samples in the upper part. Bed 20 correlates with the upper part of the styliolinites at BouTchrafine (BENSAID et al. 1985, BULTYNCK 1986). The long hiatus at the base underlines the regional significance of the Timan Event transgression.

5. MIDDLE FRASNIAN EVENTS

Middlesex Event Bed: Ca. 30 cm dark-grey, platy styliolinite without macrofauna. A conodont sample from the base produced, amongst others, *Belo. resima, I. symmetricus, Zieglerodinaovalis, Mes. asymmetrica, Mes. guanwushanensis*, and *Pa. punctata*. This fauna belongs to the base of the *punctata* Zone (MN 5 Zone), but without evidence of reworked Lower Frasnianconodonts, as reported fromBouTchrafine (BECKER & HOUSE 2000). A very peculiar novelty are the oldest known *Avignathus orthoptera*, which seem to belong to the apparatus of a polygnathid with very long free blade, which resembles *Po. brevilamiformis*. So far, *Avignathus* was supposed to enter at the top of the Middle Frasnian (see range of *Po. decorosus*, the well-known Pa element, in KLAPPER 1997). A sample from the top of the same bed was very poor in conodonts (just a few more "*Po. brevilamiformis*" = *Av*. aff. *decorosus*).

Lower *Sandbergeroceras* Beds (LSB):3 cm nodular limestone (Bed 24d), followed by a 8.5 cm thick, more solid bed, a light-grey, yellowish-grey weathering, bioturbated styliolinid wacke- to packstone(Bed 24e, Pl. 1, Fig. 5). The onset of bioturbation signals improved bottom ventilation, probably during a slight sea-level fall. Bed 24d again has *Avignathus*, together with *Mes. asymmetrica* and *Prioniodina* sp. Bed 24e yielded *Pa. punctata* and *Mes. guanwushanensis*. Both samples fall in the higher *punctata* or MN 5 Zone.

Middle *Sandbergeroceras* Beds (MSB):Above a 14 cm nodular marl (Bed 24f), the main MSB is a 15 cm thick, solid,greenish-grey nodular limestone (Bed 25a), which becomes more nodular towards the top (Bed 25b). The restricted conodont fauna from the base of Bed 25a includes *Pa. punctata*, together with *I.symmetricus*, various polygnathids, and two species of *Zieglerodina* (*ovalis*Gp.). At Jebel Amelane, however, the same unit yielded important Frasnianphacopids and index species of the MN 6 Zone (G. KLAPPERunpubl. coll.).

Pebble Bed and Red Marl:A very distinctive, up to 9 cm thick "conglomerate" (bioturbated mud-wackestone extra clastrudstone with small phosphate nodules, Pl. 1, Fig. 6), grading upwards into micritic, red marly limestone. This reworking unit has a wide distribution on the condensed Tafilalt Platform (e.g., BECKER & HOUSE 2000). Its condont fauna with *"T." ancyrognathoideus, Ad.lobata, Pa. punctata* (including *Pa. martenbergensis* as a subspecies), *Av.* aff. *decorosus*, and various other polygnathids is typical for the MN 6 Zone (regional *Ag. primus* Zone). There is evidence for a eustatic fall high in MN 6, based on comparisons with New York State (higher Cashaqua Shale, HOUSE & KIRCHGASSER 1993), the Russian Platform (top of Middle Domanik, HOUSE et al. 2000), or the Canning Basin of NW Australia (BECKER et al.1993, BECKER & HOUSE 1997).

Lower Rhinestreet EventInterval (LRh):The base of this hypoxic event interval is a thin, dark-grey, organic-rich styliolinid packstone without bioturbation and benthos (Pl. 1, Fig. 7). The sudden dacryoconarid bloom, which suggests strong eutrophication, ends abruptly. Bed 26b is a less organic-rich, but still middle-grey (not yellowish or reddish as the nodular limestones below and above), bioturbated mud- to wackestone with some shell debris (Pl. 1, Fig. 8). It is rich in benthic ostracods and conodonts: *Ancyrognathus primus, Ad. lobata, Ad. curvata* Early Form, *Ad. gigas* M1, "*Oz." nonaginta, Pa. punctata*, various polygnathids, including *Po.paradecorosus*, and icriodids. This fauna falls in the *nonaginta* or MN 7 Zone. It enables a correlation with the basal Rhinestreet Shale of New York (HOUSE & KIRCHGASSER 1993), the Middle Domanik Transgression of Russia (HOUSE et al. 2000), and the drowning of the Arche Reefs in the Ardennes.

Upper *Sandbergeroceras* Bed: 11 cm, greenish-grey, bioturbated, massive limestone (Bed 26c) with rare *Manticoceras*. At the adjacent Jebel Amelane section it falls in the MN 7Zone and the name-giving *Sandbergeroceras* and the oldest Beloceratidae (*Naplesites*) occur, which indicate UD I-G1.

***Acutiforme* Bed**: A partly quarried, very massive, light-grey micrite (Bed 28), underlain by 40-45 cm poorly exposed marl. The name of this widespread marker limestone stems from occasional *Mant. acutiforme*(see BECKER & HOUSE 2000). It is locally poor in conodonts, with a few *Po. paradecorosus* found in a sample from the top of Bed 28.

***Mesobeloceras* Beds**: Nodular marls just above (Bed 29, 30 cm) contain nice *Meso. housei*, the index species of UD I-G2a. The restricted conodont fauna of Bed 29a includes *Ad. lobata, I. symmetricus, Po. paradecorosus,* other polygnathids, and *Pa. punctata* (still MN 7 Zone).

Upper Rhinestreet Event Bed (URh): Bed 30 (26 cm) is another event bed with restricted oxygenation, a dark-grey, platy, bioclastic limestone with orthocones and, typical for eutrophic facies, *Buchiola*. There are still *"Oz." nonaginta*, *Po. paradecorosus* and *Pa. punctata* (*nonaginta* or MN 7 Zone).

***Beloceras* Beds**: The nodular marls above (Bed 31) still have (loose) *Mesobeloceras*. Bed 32 is a hard, 7 cm thick, micritic limestone with some trilobite debris.

Ag.coeni, associated with *Ad. curvata* Early Form, polygnathids and some *Palmatolepis* show that the

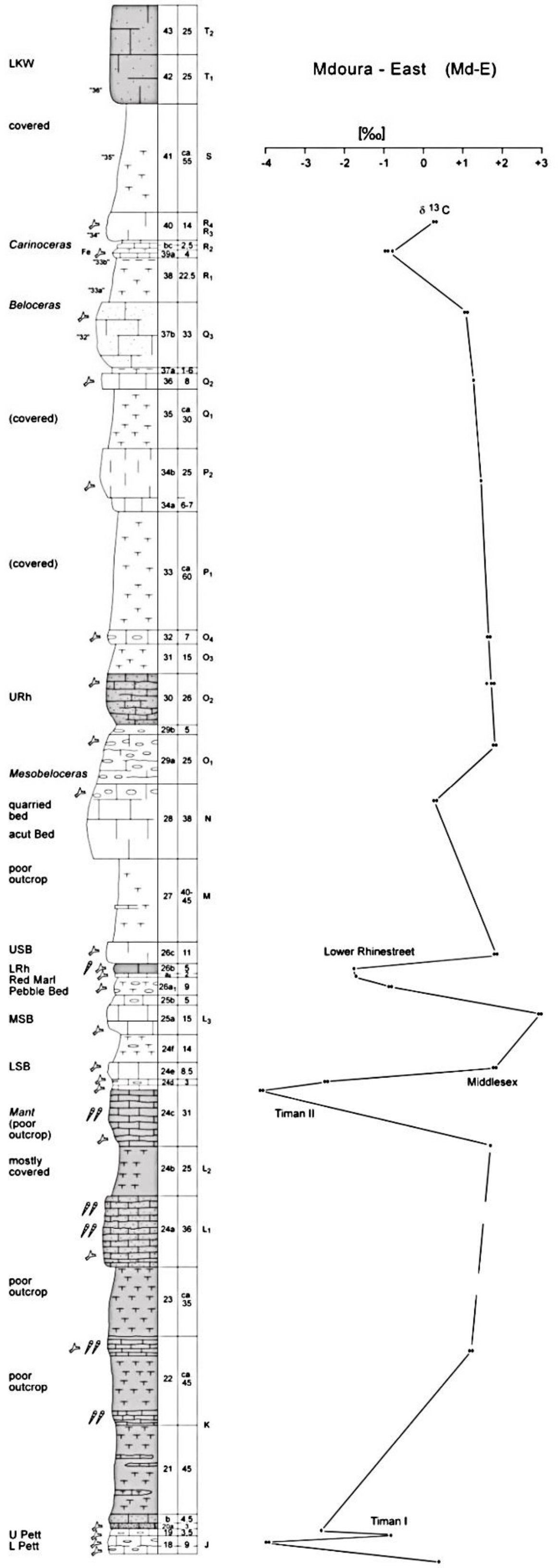

Fig. 4. Lower/Middle Frasnian section log and preliminary stable carbon isotope data for Mdoura-East.

regional*coeni* or MN 8 Zone has been reached. The subsequent ca. 2 m thick alternation of nodular marls with *Beloceras* and more solid micritic limestone (four cycles) fall in UD I-H of the ammonoid succession. As elsewhere in the Tafilal, it is not possible to recognize the MN 9 Zone. The oldest *Pa. hassi* from Bed 37b signal the MN 10 Zone. The thin Bed 39a is a distinctive, red (iron-rich) layer and is characterized by the return of probable *Avignathus* whose Pa elements resemble *Po. varcus*. Palmatolepids are lacking. Bed 40 yielded *Pa. hassi, Ad. lobata, Ag. coeni*, a *Tortodus, Po. webbi, I. symmetricus*, and a polygnathid homoemorphic to *Po. tafilensis*.

6. UPPER FRASNIAN

KellwasserFacies: The transition from the upper *Beloceras* Beds to the first massive, dark-grey limestone (Bed 42, 25 cm) is not exposed but presumably sharp as at Jebel Amelane. The Kellwasser Interval has locally not yet been sampled for conodonts or microfacies. Bed 43 (25-31 cm) is rich in *Manticoceras* and may represent the true Lower Kellwasser Limestone. Bed 44 (25 cm) is much less fossiliferous. *Crickitesholzapfeli*, the Upper Kellwassermarker (UD I-L), has been found at the top of Bed 45 (75 cm), which has irregularsubunits.

7. LOWER FAMENNIAN

Orthocone Bed: The top of Bed 46 is crowded by orthocones; this wide-spread marker unit (not yet separated in BECKER 1993), falls in the basal Famennian (UD II-A, *frechi* Zone). The overlying Beds 47-48 (= Fa 1 in BECKER 1993) are characterized by the mass occurrence of cheiloceratids and *Polonoceras* but specimens are very difficult to extract. This unit correlates with the transgressive *Polonoceras* Bed of Jebel Amelane, which representsalready the *Paratorleyoceras globosum* Zone (UD II-D). The main part of the lower Famennian is missing in an unconformity. Mdoura-East is special because of a thin subsequent limestone (Fa 2) with goniatites in very good preservation, including the best *Praemeroceras* (the UD II-E marker) of the Tafilalt. The red, crinoidalgoniatite limestone above, with masses of *Maeneceras meridionale, Armatites planidorsatus*, and *Acrimeroceras falcisulcatum,* falls in the transgressive basal middle Famennian (UD II-G, *meridionale* Zone).

7. CARBON ISOTOPE STRATIGRAPHY

Recent geochemical investigations in Belgium, Western Canada, Poland and South China provided evidence for major changes of the global carbon cycle in association with the Middlesex (= *punctata*) Event at the base of the Middle Frasnian. The evidence was summarized by PISARZOWSKA & RACKI (2012). Consequently we made an attempt to find these carbon isotope anomalies in sections of the Tafilalt Platform, including Mdoura-East. Data from whole rock carbonate (selected matrix of polished sections, Fig. 4) are still preliminary since additional samples are required. All beds were measured at least twice, with mostly very small variability

Results do not fit the established patterns elsewhere. The basal Timan Event Bed, the top of the Middlesex

Event Bed, and the Lower Rhinestreet Event Beds are all characterized by sharp negative excursion, which obviously resulted from the early diagenetic recycling of organic matter during recrystallization. The negative spikes interrupt a broadly positive plateau of the MN 4-10 but values hardly reach $2^0/_{00}$ (4-6 $^0/_{00}$ elsewhere). Only the Middle *Sandbergeroceras* Bed records a positive peak close to $3^0/_{00}$ but it falls already in the basal MN 6 Zone and, therefore, is not the main peak near the base of the *punctata* Zone.

The various event beds do not include detrital land plant material. There is no evidence for pulses of increased vegetation cover or increased weathering associated with the events. We therefore reject the idea of relationships between land plant evolution and offshore, deeper marine eutrophications, which allowed the bursts of dacryoconarid populations.

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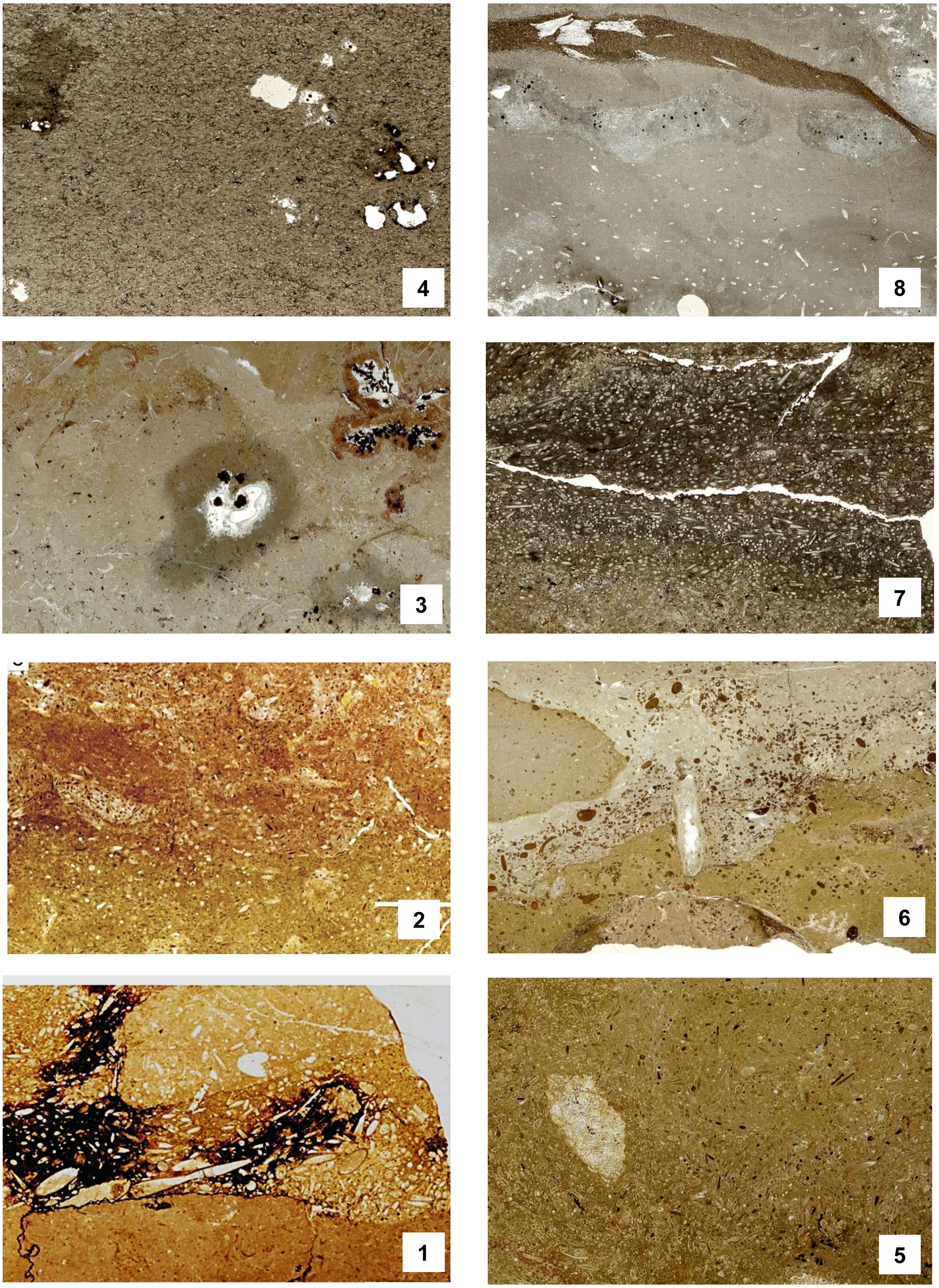
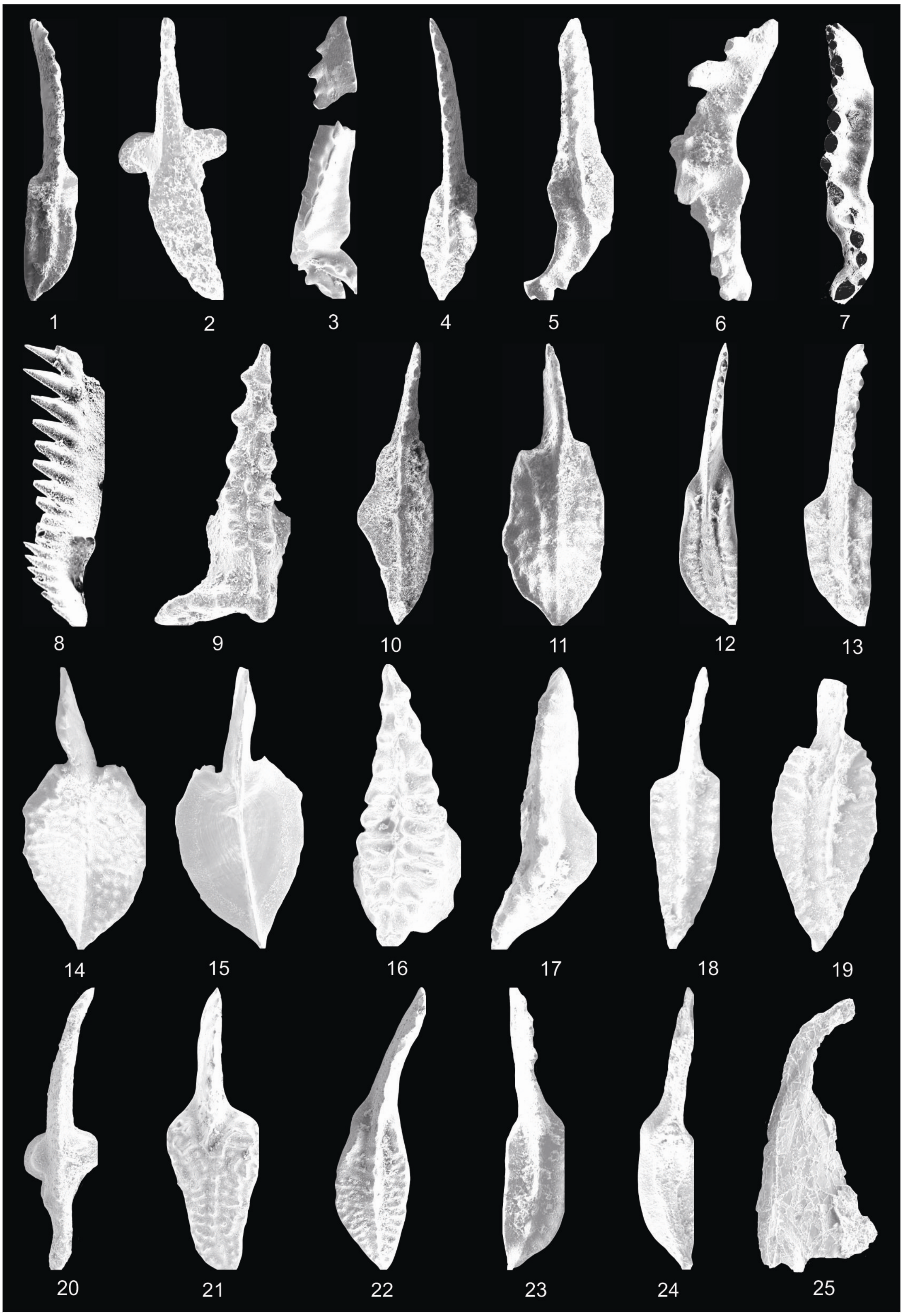


Pl. 1. 1. Bed 9c, Upper *Sellagoniatites* Limestone, mudstone with styliolinids, sharply truncated by styliolinid packstone with iron crusts. **2.** Bed 9e, Lower/Middle Taghanic Crisis Interval, bioturbated styliolinid wackestone grading upwards into iron-enriched packstone with mudclasts. **3.** Bed 19, Upper *Petteroceras* Bed, strongly bioturbated mudstone with styliolinids and iron encrustations. **4.** Bed 20, basal Timan Event Bed, dark-grey styliolinite without bioturbation. **5.** Bed 24e, Lower *Sandbergeroceras* Bed, bioturbated styliolinid wacke- to packstone. **6.** Bed 26a$_1$, Pebble Bed, strongly bioturbated lime mud “conglomerate” with enrichments of phosphate clasts. **7.** Bed 26a$_2$, Lower Rhinestreet Event, organic-rich, dark styliolinite without bioturbation. **8.** Bed 26b, Lower Rhinestreet Event, slightly bioturbated mudstone, interrupted by microsparitic vein.



Pl. 2. Givetian conodonts from Mdoura-East, 1-4. Bed 9b, lower half of Upper *Sellagoniatites* Bed, 5-7, Bed 9d, upper part of Upper *Sellagoniatites* Bed, 8-9, Bed 10b, aff. *amplexum* Bed, 10-13, Bed 10e, *erraticus* Bed, 14-16, Bed 13, Upper Marker Bed, 17, Bed 14, top of Upper Marker Bed, 18, Bed 16, *Taouzites* Beds, 19-21, Bed 18, Lower *Petteroceras* Bed, 22-24, Bed 19, Upper *Petteroceras* Bed, 25, Bed 20, Timan Event Bed. **1.***Po. ansatus*, **2.***Bip.bipennatusbipennatus*, **3.***Tortodus* sp., pathological (bifurcate), **4.***Po.timorensis*, **5.***T.*aff.*caelatus*, **6.***T. weddigei*, **7.***T.*aff.*sardinia*, **8.***“Oz.” semialternans*, **9.***Lat. latericrescens latericrescens*, **10.***Schm. hermanni*, **11.***Po.limitaris*, **12**. *Po.dubius*, **13**. *Schmpietzneri*, **14-15**. *Kl. vysotzkii*, **16.***I. tafilensis*, **17.**Gen. et sp. indet. (?Pb element of *Klapperina*, somewhat homoeomorphic to *Nothognathella*), **18**. *Po. denglerisagitta* (holotype), **19**. *Po. dengleridengleri*, **20**. *“Oz.” Sannemann isannemanni,***21**. *Schm. peracutus*, **22**. *Po. tafilensis*, **23-24.***Schm. gracilis*, **24**. *Belodellaresima*.