9. GENESIS OF GIVETIAN BIOSTROMES

During Givetian and Frasnian times, favourable conditions for reef growth prevailed on a global scale, the largest expansion and the highest number of build-ups compared to other periods of the Phanerozoic existed in this time (Copper 1989, Kiessling et al. 1999). Givetian reef builders were diverse, with microbialites, calcified cyanobacteria, stromatoporoids, tabulate corals, and calcified algae as the dominant biota (Wood 1999). These organisms erected buildups in warm water areas all over the world, e.g. the about 350 km long reef complex in the Canning Basin, Australia (Playford 1980), various buildups in North America (e.g. Moore 1988), in Central Europe (e.g. Krebs 1974) and Asia (e.g. Yu & Wuyi 1988). Bio-constructions of Devonian age also exist in NW-Africa, for example in the Western Meseta of Morocco: Gendrot et al. (1969) and Gendrot (1973) provide an overview over Devonian reef environments for this area, Cattaneo et al. (1993) investigated especially in Givetian limestones of the north-western Meseta, Kempf (1993) and Gonschior (1994) studied Devonian deposits in the Rehamna massive (Western Meseta). Reef structures of the Western Meseta are mostly of the biostrome type, but some bioherms also developed. In the Anti-Atlas and further south, coral-stromatoporoid biostromes and bioherms of the Middle Devonian, often together with mud- and reef-mounds, were investigated for instance by Dumestre & Illing (1967), Wendt (1988), Wendt et al. (1997), Kaufmann (1998), Wendt & Kaufmann (1998).

There is no distinct difference between Moroccan buildups and those further north concerning their fossil content, but constructions like large barrier reefs did not develop in NW-Africa. Continuously grown bioconstructions are limited to a maximum thickness of about 30 m (reef-mounds in Spanish Sahara, see Dumestre & Illing 1967), only biostrome successions reach higher thicknesses. With about 200 m thickness, the Givetian biostrome / crinoidal-grainstone alternation at the eastern edge of the Jebel Rheris is unique in NW-Africa. Objectives of the current chapter are, to determine the reasons for biostromal growth at the Jebel Rheris and to reconstruct the colonisation pattern of the biostrome constructors (this chapter is already published, see Fröhlich 2003).

9.1 Biostrome classification

Kershaw (1994) proposed a classification of bi-ostromes, based on the internal arrangement of the biota, especially the percentage of in-place versus allochthonous components is considered. This approach can well be utilised in the present study, applying the terms auto-, autopara-, para- and allo-biostromes for > 60% in place, 20-60% in place, < 20% in place and 0% in place components. In section 1, 11 biostromes have been studied (Fig. 10) considering fossil content, texture, substrate and range of transportation of the components.

9.1.1 Allobiostromes

Biostromes B 8 and B 11 can be classified as allobiostromes. The basis is irregular, parts of the crinoidal grainstone layer below were reworked during deposition. The corals and stromatoporoids are mostly fragmented and randomly distributed, their size varies between 9 x 3.5 cm and 3 x 1 cm in diameter (measured on a 2-dimensional surface). About 40 % of the allobiostromes consists of internal sediment, which is mostly made up of crinoid ossicles, peloids, intraclasts and few shell fragments.

9.1.2 Parabiostromes

Biostromes B 1, B 2, B 4, and B 5 are parabi-ostromes. The basis is often erosive (Pl. 4/2), most of the biota are partially destroyed. But some flat tabulate corals and stromatoporoids remained in life position. Average size of the components is about 10 x 6 cm, the internal sediment has a portion of 30 %.

9.1.3 Autoparabiostromes

One autoparabiostrome (B 10) was recognised by its flat substrate, composed of crinoidal debris with undestroyed corals and stromatoporoids, situated in place above it. About 60 % of the components were fragmented and randomly distributed throughout the biostrome, their size varies between 13 x 10 cm and 9 x 6 cm. 30 % of the layer consists of internal sediment.

9.1.4 Autobiostromes

Two types of autobiostromes can be identified at the Jebel Rheris at section 1: Biostrome B 9 is 60 cm thick and consists mostly of large stromatoporoids (up to 100 x 60 cm) and some corals that often grew

on each other and are to 80 % in place. The substrate consists of crinoidal debris; the internal sediment is 20-30 %. The second type of autobiostrome (B 6, B 7) has almost no vertical extension, only one generation seems to be preserved (see below). The corals and stromatoporoids grew close to each other in a laterally restricted zone, they have a patchy distribution. Thickness of the layer is about 30 cm.

A composite biostrome (B 3) can be observed at section 1, consisting of an autoparabiostromal facies in the lower part and an allobiostromal facies in the upper part (Pl. 4/3). The former comprises about 30 % in place organisms that can mostly be found at the bottom of the layer, towards the top, the amount of disrupted stromatoporoids and tabulate corals and the share of the internal sediment increases. The average size of these biota is 6 x 3 cm. Thereon, a 40 cm thick succession of dolomitised crinoidal debris without corals or stromatoporoids can be noticed, and then an allobiostromal facies follows with a sharp and horizontal contact. Only fragments of stromatoporoids and corals can be observed here with average diameters of 2 x 2 cm. The internal sediment has a portion of about 40 % and is completely dolomitised. So this 1.4 m thick layer developed through 2 phases: The gradual growing lower part, probably storm-influenced and buried by crinoidal debris, and the transported and resedimented upper part.

9.2 Biostrome constructors

9.2.1 Method

The composition of 11 biostromes in section 1 was determined in a rectangle that was laid on the vertical section of the layers (where possible on the bedding plane). The length of the sides of the rectangle was determined by defining the average diameter of the organisms in centimetres and multiply this value with ten, so the size of the rectangle depends on the average size of the determined biota. Thus the total amount of determined specimens should remain more or less equal in each biostrome, the relative quantity of representatives of one group against another will be emphasised. The class Stromatoporoidea is not further subdivided into orders, but the size and the outline form is described using the classification of Kershaw (1998) and, where possible, the ornament and growth patterns. The subclasses Rugosa and Tabulata are split into subgroups according to their

skeletal fabric, except for solitary Rugosa.

9.2.2 Results

Due to the dendroid growth form of Thamnoporids, the determined specimens were hardly complete but generally fragmented, so Alveolitids can be regarded as the dominant coral in the biostromes, followed by Phillipsastreids, Favositids, solitary Rugosa and Heliolitids (Table 13). One specimen of *Hexagonaria* sp. was found by Erbacher (1991) at the Jebel Rheris, but these colonial rugose corals seem to be extremely rare. The ratio corals / stromatoporoids changes considerably between single biostromes (Table 13), but both groups are equally important constructors at the Jebel Rheris. It is also noteworthy that in contrast to other Middle Devonian bioconstructions, almost no algae or microbes contributed to biostromal growth.

Root (1967) introduced the 'guild concept', which helps to distinguish "groups of species that exploit the same class of environmental resources in a similar way ... without regard to taxonomic position". This was applied in characterising fossil communities (e.g. Bambach 1983) and determining the ecological function of the frame-builders, and might serve as a useful tool in the present study. Fagerstrom (1988) subdivided reef communities into five guilds (constructor, baffler, binder, destroyer, and dweller), each contributing in a distinct way to the development of bioconstructions. But there are also some weaknesses concerning the application of the concept, which are discussed by Fagerstrom & Weidlich (1999).

The constructor guild is by far the most important group contributing to the Jebel Rheris biostromes. Most of the stromatoporoids and probably all of the colonial corals belong to this guild, building a more or less stable framework. The baffler guild mostly is represented by crinoids, to a lesser extend by solitary rugose corals and bryozoans. These organisms were obstructions to currents, but, except for the corals, are rarely preserved as complete skeletons. Almost no representative of the binder guild was found apart from few stromatoporoids, which laterally overgrew some constructors. Encrusting organisms that united internal sediments with the framework were not noticed. Biologic destruction of the constructors by boring and rasping organisms of the destroyer guild also was not obvious. Brachiopods and gastropods were members of the dweller guild, which neither destroyed nor contributed to the biostromal growth;

Biostrome B 1 [2.25 m²]

37 stromatoporoids (5x2 - 45x20 cm)				12 tabulate corals (1x1 - 4x2 cm)	1 rugose coral
16	12	7	2	6 Alveolitids, 5 Thamnoporids, 1 Favositid	solitary
tabular,	bulbous,	domical,	uncertain		
3 x coalescence, mamelons: rarely				often fragmented	
ratio stromatoporoids / corals ~ 3 : 1					

Biostrome B 2 [1.44 m²]

41 stromatoporoids (2x2 - 40x17 cm)				6 tabulate corals (4x3 - 20x20 cm)	2 rugose corals
1	2	10	28	3 Alveolitids, 3 Favositids	Phillipsastreids 20x15, 40x17cm
tabular,	bulbous,	domical,	uncertain		
ratio stromatoporoids / corals ~ 5 : 1					

Biostrome B 3 [0.04 m²]

39 stromatoporoids (1x1 - 15x5 cm)				60 tabulate corals (1x1 - 25x10 cm)	1 rugose coral
10	8	1	21	6 Alveolitids, 53 Thamnoporids, 1 Favositid	solitary
tabular,	domical,	laminar,	fragments		
				fragmented	
ratio stromatoporoids / corals ~ 2 : 3					

Biostrome B 4 [0.06 m²]

6 stromatoporoids (2x1 - 10x6 cm)	42 tabulate corals (2x1 - 7x5 cm)	1 rugose coral
fragmented	7 Alveolitids, 30 Thamnoporids, 5 Favositid	Phillipsastreid (fragment)
	fragmented	
ratio stromatoporoids / corals ~ 1 : 7		

Biostrome B 5 [1 m²]

17 stromatoporoids (4x3 - 20x15 cm)			20 tabulate corals (1x1 - 4x2 cm)	1 rugose coral
6	4	7	8 Alveolitids, 7 Heliolitids, 3 Favositids, 2 Thamnoporids	Phillipsastreid (fragment)
domical,	bulbous,	fragmented		
ratio stromatoporoids / corals ~ 1 : 1				

Biostrome B 8 [0.06 m²]

6 stromatoporoids (2x2 - 9x5 cm)	39 tabulate corals (1x1 - 6x5 cm)	1 rugose coral
fragmented	2 Alveolitids, 35 Thamnoporids, 1 Favositid, 1 Heliolitid	solitary
	fragmented	
ratio stromatoporoids / corals ~ 1 : 7		

Biostrome B 9 [6.25 m²]

25 stromatoporoids (100x60 - 12x10 cm)	2 tabulate corals (2x1, 18x10 cm)	2 rugose corals
2 x domical, 3 x coalescence 20 x uncertain	1 Alveolitid, 1 Thamnoporid	1 Phillipsastreid (20x18 cm), 1 solitary
ratio stromatoporoids / corals ~ 6 : 1		

Biostrome B 10 [1 m²]

8 stromatoporoids (27x25 - 5x4 cm)	25 tabulate corals (2x1, 20x13 cm)	6 rugose corals
4 x domical, 4 x uncertain	23 Alveolitids, 2 Thamnoporids	5 Phillipsastreids (18x7 - 5x4 cm), 1 solitary
ratio stromatoporoids / corals ~ 1 : 4		

Biostrome B 11 [0.5 m²]

4 stromatoporoids (7x4 - 3x2 cm)	10 tabulate corals (2x3, 10x4 cm)	17 rugose corals
fragmented	9 Alveolitids, 1 Thamnoporid	10 Phillipsastreids (3x7 - 1x2 cm), 7 solitary
ratio stromatoporoids / corals ~ 1 : 7		

Biostrome B 6 [0.25 m²]

16 stromatoporoids (4x3 - 22x14 cm)		
8	2	6
domical,	tabular,	uncertain
16 tabulate corals (2x1 - 10x5 cm)		
10 Alveolitids, 5 Thamnoporids 1 Favositid		

Biostrome B 7 [1 m²]

34 stromatoporoids (4x3 - 45x20 cm)		
14	5	15
domical,	bulbous,	uncertain
2 x coalescence		
14 tabulate corals (2x1 - 10x5 cm)		
10 Alveolitids, 3 Favositids, 1 Heliolitid		

	Stro.	Cor.
B 1	37	13
B 2	28	8
B 3	39	61
B 4	6	43
B 5	17	21
B 6	16	16
B 7	34	15
B 8	6	40
B 9	25	4
B 10	8	31
B 11	4	27

Biostromes B 1 - B 11

234 stromatoporoids		
51	23	19
domical,	bulbous,	tabular,
141 uncertain / fragmented		
253 tabulate corals		
92 Alveolitids, 9 Heliolitids, 18 Favositids, 134 Thamnoporids		
33 rugose corals		
21 Phillipsastreids, 12 solitary		
ratio stromatoporoids / corals ~ 1 : 1		

Tab. 13: Fossil content of selected biostromes, determined within defined areas (size in brackets within the grey bar) on the vertical section. That of biostromes B 6 and B 7 was determined on the bedding plane. The total amount of biota, determined within defined areas of 11 biostromes in section 1 is presented at the right margin, furthermore the amount of stromatoporoids (Stro.) and corals (Cor.) of each biostrome.

they were not prevalent.

9.3 Colonisation

At section 1, the pioneer stage of biostromal growth is preserved and can be examined both in the vertical section and on bedding planes. Here, the organisms are mostly in situ and formed clusters that are laterally surrounded by crinoidal debris and some shell fragments. The stage of vertical accretion of stromatoporoids and corals is not fully developed. Biostromes B 6 and B 7 represent this kind of initial stadium and were investigated in detail (Table 13).

About 70 % of the biota in biostrome B 6 seem to be in place, the portion of the internal sediment increases from 20 % in the centre of the cluster (Fig. 30) rapidly towards the margin. Constituents of the latter are besides the dominant crinoidal debris few brachiopods and small fragments of corals and stromatoporoids. The substrate consists of coarse crinoidal debris.

As in B 6, about 70 % of the stromatoporoids and corals are in place in B 7, some are overturned but mostly not transported. A vertical section of this 30 cm thick layer (Fig. 30) shows a 5 cm thick basis made up of crinoidal debris, overlain by a coarse 5 cm thick layer of fragments of stromatoporoids, corals, crinoids, and brachiopods. On this debris in situ growing stromatoporoids and corals occur, which are surrounded and sometimes overlain by crinoidal debris. Pioneer communities of reefs have been described as simple in biotic composition and consisting mostly of fragile, twiglike forms that started growing in quiet water (Nicol 1962, Copper 1974). The succeeding stages towards the climax or mature community grew in shallower water and became more complex and diversified. Such changes in a community through time are termed 'ecological succession' by Walker & Alberstadt (1975), who distinguish a basal stabilisation zone, an overlying colonisation zone, then a diversification zone and finally a domination zone in ancient reefs. An initial stabilisation zone is not developed at the Jebel Rheris, probably because of the coarse and stable substrate; the colonisation zone can be subdivided into three stages (Fig. 31): (1) cluster settlement; the pioneer communities, consisting of stromatoporoids and corals, started growing in distinct areas, which might be called nucleation points. (2) lateral dispersion; no patch reefs developed from these growth centres. Colonisation continued mainly laterally (Pl.
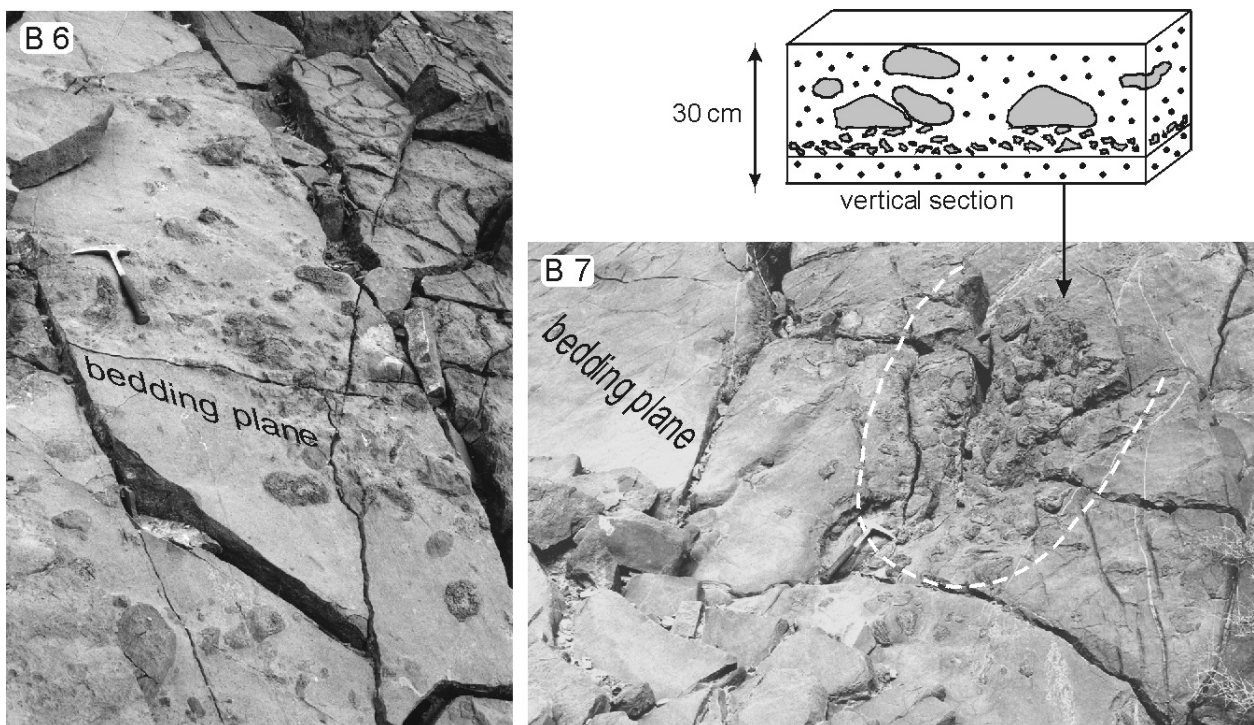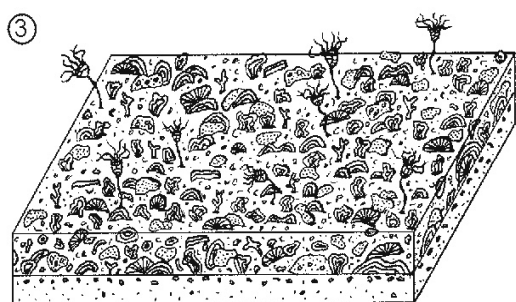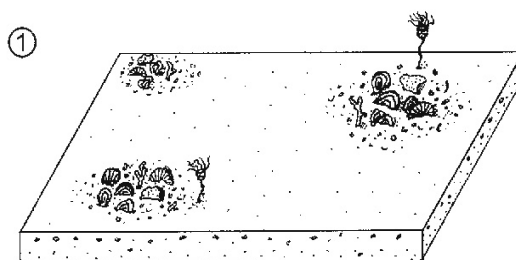

Fig. 30: The initial state of biostromal growth is preserved in B 6 and B 7. Stromatoporoids and corals simultaneously started settling on a stable substrate made up of crinoidal debris (black dots), fragments of stromatoporoids, corals, and brachiopods. The growth started in distinct zones, creating clusters of organisms ('cluster settlement').

Fig. 31: Colonisation of the sea floor proceeded on a more or less stable substrate, mostly made up of crinoidal debris, and can be divided into three phases: (1) Cluster settlement; pioneer communities, which mainly consisted of corals and stromatoporoids, evolved in laterally delimited areas. (2) Lateral dispersion; the biota spread out on the sea floor, starting from the clusters, but did not establish mounds or patch reefs. (3) Vertical accretion; corals and stromatoporoids grew on each other, but a more or less planar sea floor was maintained.

nities from time to time was presumably caused by periodically occurring storms, which is confirmed by sedimentary structures of the crinoidal grainstone facies. Slumping structures can also sometimes be observed, arguing for an at least partially inclined sea floor, which might additionally have caused gravity induced transportation of components in the biostromes.

9.4.2 Stromatoporoid growth forms

The dominating growth form of the stromatoporoids is domical and bulbous (see Pl. 4/5, 8). Some authors documented a distinct succession of these organisms in bioconstructions, where low profile laminar forms dominate at the basis and stabilise the substrate, towards the top, more erect forms developed (Bjerstedt & Feldmann 1985, Kershaw 1990). But the substrate at the Jebel Rheris mostly was made up of coarse crinoidal debris, so there was probably no need for initial stabilisation. This might be the reason for the absence of laminar forms. In general, stromatoporoid growth forms were influenced both by environmental and genetic factors (references in Kershaw 1998), but whether genetic factors affected the growth forms at the Jebel Rheris cannot be decided. As no stromatoporoids with ragged outline forms occur, indicative for periodical sedimentation (Kershaw 1998), deposition rates during biostromal growth were probably low and homogenous.

9.4.3 Biostrome genesis

Biostromal growth started on a stable substrate, which was made up of crinoidal debris, sometimes additionally stabilised with fragments of corals, stromatoporoids and brachiopods. The cluster settlement (see chapter 9.3) requires locally more favourable growth conditions, as these biota usually preferred a

4/6), until large areas of the sea floor were covered. (3) vertical accretion; biostrome constructors more and more grew on each other (Pl. 4/7), but a mostly planar sea floor was maintained.

9.4 Discussion

9.4.1 Depositional environment

Givetian rocks at the Jebel Rheris show all types of biostromes from the transported allobiostrome to the autobiostrome with mostly in situ organisms. Most of the fragmentation and transportation of organisms can probably be related to events: As some biostromes appear not to be affected by wave action or currents, a water depth below fair weather wave base is inferred; so the disruption of fossil commu-

'standoff' style and tried to avoid competitive inter-actions (Kershaw 1998). But one can only speculate about slight differences of the substrate, local changes of the current regime or nutrient concentrations. The crucial point, however, in the development towards a biostrome and not towards a bioherm or reef-mound in the lateral dispersion phase is probably the guild structure. As mentioned above, the constructor guild with mostly domical and bulbous growth forms dominated the community, but the binder guild was more or less absent, which did not lead to a patch reef, stable enough to withstand periodically occur-ring storms. So the laterally protected growth on the sea floor without erecting a distinct topography was presumably energetically favoured. It has to be mentioned that this interpretation cannot be applied to biostromes, whose flat upper surface is due to ero-sion (cf. Kershaw 1993).

Other factors also influenced the growth of bi-ostromes. Several authors discussed, why extensive sheet-like accumulations of marine organisms some-times did not swell into a reefal structure or mound-like bioherm. After Brouwer (1964), Middle and Late Devonian biostromes in the Cantabrian mountains developed on a relatively stable shelf in a shallow sea under uniform conditions. Embry & Klovan (1970) proposed that a slowly subsiding shelf generates bios-tromes, a rapidly subsiding basin promotes bioherms. For Silurian biostromes on Gotland, Kershaw (1993) and Kershaw & Keeling (1994) assumed a relatively stable sea level, low clastic supply, laterally extensive stabilised substrate and a more or less flat sea bed in shallow water.

The extreme thickness variations of the Givetian succession, ranging from 250 m thickness in section 1 over a lateral distance of 1700 m to 80 m thickness at section 4, are attributed to differential subsidence, creating different amounts of accommodation space. The development of biostromes and sedimentation of interbedded crinoidal grainstones kept up with the subsiding basin and maintained mid-ramp conditions, favourable for the growth of corals and stromatoporoids, throughout the Jebel Rheris, apart from its southwestern edge. Thus, variation in the rate of subsidence was not an influencing factor during the establishment of biostromes at the Jebel Rheris. Local siliciclastic input can prevent the development of large biostromes, as it hinders the growth of corals and stromatoporoids, but such an input is generally very low in Givetian facies in the investigated area.

9.4.4 Termination of biostromal growth

At section 1, 49 biostromes alternate with the cri-noidal grainstone facies, so the stromatoporoids and corals repeatedly stopped growing. No erosion surfaces, caused by regressions, were observed on top of the biostromes as Kershaw (1993) reported from the Silurian Hemse biostromes of Gotland. A reason for the repeated termination of biostromal growth could have been a sudden covering of the biota with bioclastic (mostly crinoidal) detritus by storm events, but it is improbable that they regularly covered the sea floor over several kilometres to create flat topped biostromes. Climatic changes might have affected the growth of corals and stromatoporoids. The palaeogeographic position of Gondwana during the Devonian is still matter of debate (e.g. McKer-row et al. 2000, Tait et al. 2000), but it is likely that the modern NW-Africa was located during Middle Devonian times at the southern limit of warm wa-ter shelf carbonate production, between a southern latitude of 30° and 40° (Scotese 2002). Therefore, slight changes in temperature would have had ma-jor effects on the life of corals and stromatoporoids: These organisms would have stopped growing more or less isochronously throughout the shelf and left behind bioconstructions with flat upper surfaces. Slight increases in temperature again would provide fertile conditions. The palaeolatitude might also be responsible for the lack of large and continuously grown reef complexes in the present NW-Africa.

But the most probable reason for the repeated termination of biostromal growth at the Jebel Rheris were several drowning events (sensu Kendall & Schlager 1981). Although no hardgrounds or shales occur above the biostromes, the general facies model, discussed above, indicates that crinoidal grainstones can be related to deeper water conditions than bios-tromes. So the Givetian succession presumably was controlled by cyclic sea-level changes.

9.4.5 Ecological succession

The Givetian strata at the Jebel Rheris can be called an 'allogenic succession' (after Odum 1971, p. 255), because the succession is not community controlled. Similarities to the 'short-term succession' of Walker & Alberstadt (1975) can be noticed, concerning the periodic wipe out of a community, which is then reconstructed. But the reason is different: In their

case study, periodic influx of mud changed the environment.

A diversification zone, following the colonisation zone, was not noticed in the Jebel Rheris biostromes (data of the species level are not considered, see chapter 9.2). The ratio corals / stromatoporoids varies from biostrome to biostrome, non of these groups succeeded in dominating the community over a longer period of time, which was observed in other Middle Palaeozoic bioconstructions (e.g. Wilson 1975). In general, this is an important difference between a thick succession of biostromes and a continuously grown reef: Biostromes repeatedly stop growing and have to start again with the colonisation zone (or stabilisation zone, where necessary), so there is little chance for a community to reach a mature stage.

10. DIAGENESIS OF GIVETIAN BIOSTROMES

The diagenetic history of Devonian bioconstructions has been studied all over the world (e.g. Mattes & Mountjoy 1980, Smosna 1984, Kerans et al. 1986, Machel 1990, Kaufmann & Wendt 2000). In the eastern Anti-Atlas (southern Mader), the diagenesis of carbonate mounds was investigated in detail (Kaufmann 1997). The palaeogeographic position of these mounds is located in a slightly deeper environment (mid-ramp and outer ramp) than the coral-stromatoporoid biostromes and crinoidal grainstones of the Jebel Rheris (mid-ramp), furthermore different facies types (e.g. no micrite and distinct microbial communities in the biostromes) occur. Together with local variations in subsidence history, this might lead to different diagenetic patterns, which will be examined in the following. The diagenesis of stromatoporoid bioconstructions is also of general interest for petroleum exploration, because these buildups are known to be oil-producing in Canada (e.g. Stearn et al. 1987). So it is the aim of this chapter to unravel the timing and nature of cementation, dolomitisation, compaction, and dissolution of the Givetian biostrome succession of the Jebel Rheris.

10.1 Cement types

10.1.1 Fibrous calcite
This cement type is extremely rare in the Givetian facies and occurs only in some pores of thamnoporoids in section 4 (Pl. 5/1; Pl. 6/1). The crystal length is commonly around 150 µm, the width around 25 µm, i.e. they have an acicular form. They grew perpendicular to the substrate and point into the centre of the pore, establishing isopachous rims. As these crystals show unit extinction and straight twin plains, they can be termed radial-fibrous (Kendall 1985). Cathodoluminescence (CL) shows a pattern of bright yellow luminescence at the bottom of the crystals, then a zone of alternating moderate and bright luminescence and again a bright zone at the top.

10.1.2 Syntaxial calcite spar
Echinoderm fragments are ubiquitous in Givetian rocks, therefore syntaxial cement can be found in almost every thin section. The crystals are turbid and sometimes are up to several millimetres in size. Zonation cannot be observed under CL, a moderate to bright yellow luminescence is common. Turbidity is mostly derived from microdolomites (Pl. 5/3, 4) which are up to 12 µm in diameter and sometimes consist of euhedral rhombs. They show a bright to dull luminescence.

10.1.3 Scalenohedral calcite
This type of cement (also called 'dentate' or 'dog-tooth' cement) was only found in section 9 in biostromal facies, forming the first cement generation around components like shell fragments (Pl. 6/4) and in intraskeletal pores of bryozoans (Pl. 5/2). The crystal shape is mostly pyramidal, the length is up to 150 µm. A clear appearance is common. Under CL, this cement typically shows three distinct zones: A thick non-luminescent inner zone, followed by a thin bright luminescent and a thin non-luminescent zone.

10.1.4 Blocky calcite spar
This is by far the most common cement type, observable in nearly all of the thin sections from Givetian strata. Equant calcite cement occurs in ostracod and brachiopod shells, in pores of tabulate corals, and it fills shelter porosity. A drusy mosaic occasionally can be noticed, crystal boundaries are sometimes irregular. Staining of thin sections with potassium ferricyanide (method of Dickson 1966) and CL examinations revealed the iron and manganese content of the carbonate, which allowed the differentiation of two types of blocky spar: Blocky spar I is non-ferran and mostly bright yellow to zoned moderate yellow, sometimes with non-luminescent appearance (Pl. 5/1, 2); blocky spar II (Pl. 6/3) is ferran, with a mostly

unzoned, dull, orange appearance (Pl. 5/8). The latter can only be found as filling in fractures.

10.2 Dolomite petrography and texture

Replacement dolomite is the only type that was observed at the Jebel Rheris. No dolomite cements were noticed. Crystals are euhedral to subhedral, with a unimodal size distribution of mostly large rhombs (up to 1,2 mm). From fully dolomitised to partly or not dolomitised areas a transition from a hypidiotopic mosaic with planar compromise boundaries to a idiotopic mosaic (Pl. 5/5, 6) developed (classification after Sibley & Gregg 1987); also isolated rhombs occur. Crystals show unit extinction and straight faces. CL revealed a distinct zonation: about 10 to 15 more or less dull zones can be distinguished in single rhombs, the colour varies between dark red to orange (Pl. 5/7). A high iron content of the dolomites is further confirmed by a blue staining that occurred after contact with potassium ferricyanide. In outcrop, weathered dolomite therefore can well be recognised due to a beige / orange colour. Dolomitisation is mostly fabric selective, i.e. larger components like stromatoporoids and corals often remained calcitic, while the matrix including smaller allochems is dolomitised (Pl. 6/5). Dolomitisation often appears patchy in a tens of centimetres to tens of meters scale, sometimes crossing the original bedding (Pl. 6/6).

10.3 Dissolution and compaction

Large scale dissolution features can be observed especially in the central and western part of the mountain, where cavities up to 20 cm in diameter occur in a largely dolomitised carbonate rock (Pl. 6/7). Here, presumably former calcitic relics like corals and stromatoporoids were selectively dissolved. The occurrence of stylolites indicates chemical compaction, which was caused by overburden and tectonic stress, as vertical and horizontal dissolution seams can be noticed.

10.4 Discussion

10.4.1 Early diagenesis
The first diagenetic step was the precipitation of fibrous calcite. The crystals can only grow in supersaturated sea-water (e.g. Alexandersson 1972), their isopachous fibrous nature is a further criterion for marine origin (Longman 1980). This cement type only occurs in few skeletal cavities of thamnoporoids. The bright to moderate yellow luminescence of the fibrous calcite indicates anoxic conditions, where Mn^{2+} was incorporated into the calcite lattice and Fe^{2+} was eliminated from the pore-water as sulphide (Tucker & Wright 1990). Active water circulation, usually a prerequisite for the precipitation of this cement type, therefore is improbable. Bacteria might have contributed to the precipitation of this cement, which was observed by Berner (1971) in some skeletal cavities, where decaying organic matter was present.

Syntaxial cements, which are common in Givetian rocks of the Jebel Rheris, generally grow in the marine environment as well as in subsurface and subaerial zones (e.g. Burgess 1979, Perkins 1985, Walker et al. 1985), the latter authors also see a dependence on variations in marine water geochemistry. The timing of the precipitation of these cements at the Jebel Rheris cannot easily be determined; mostly bright to dull luminescence and the lack of zonation indicates a growth under homogenous anoxic burial conditions.

Scalenohedral calcite (SC), the first cement generation in biostromes in section 9, precipitated mostly in oxic pore-water, as it is non-luminescent, apart from one thin zone that incorporated Mn^{2+}-ions. After Reinhold (1999), the formation of SC is not restricted to a specific diagenetic environment. In the Middle Devonian carbonate mounds of the southern Mader (Kaufmann 1997) and in the Ahnet Basin (Algerian Sahara, Kaufmann & Wendt 2000), nonluminescent and bright-/banded scalenohedral cements are interpreted as shallow burial precipitates. A meteoric origin of SC in Givetian limestones of the Jebel Rheris is improbable, because it occurs within a mid-ramp facies without indications of subaerial exposure. Therefore it is also interpreted as shallow burial cement.

Blocky calcite spar generally either grows in deep burial or in the meteoric phreatic environment, exceptions are rare (Given & Wilkinson 1985). The meteoric environment, however, is precluded in this study for the reasons mentioned above. Blocky spar I grew on top of marine cements. Where these are lacking, it fills the whole pore space and is generally bright luminescent, indicating precipitation in the shallow burial environment, where Fe^{2+} is not incorporated into the calcite lattice. The rare occurrence of partly non-luminescent blocky spar I

in the centre of the pores (Pl. 5/1) might indicate that the pore space was closed very rapidly, still in the zone of shallow burial, where oxic conditions can sometimes prevail. In contrast, the majority of diagenetic successions shows a non-luminescent – bright – dull zonation, which indicates cementation while increasing burial (e.g. Grover & Read 1983, Frykman 1986). Also in carbonate mounds of the southern Mader, dull luminescent blocky spar fills the centre of pores (Kaufmann 1997). The lack of dull-luminescent blocky spar I in Givetian rocks of the Jebel Rheris and the implicated rapid cementation thus would explain the absence of mechanical compaction.

10.4.2 Late diagenesis

Further diagenesis of the Givetian rocks occurred during deep burial. Dolomitisation under subsurface conditions is shown by: 1) Dolomitised areas cross the original bedding and show a patchy distribution. 2) The growth of coarse dolomite crystals, often more than 1 mm in diameter, also is strongly suggestive of burial diagenesis (Zenger 1983). 3) Dolomite is ferroan, proved by potassium ferricyanide staining and a dull orange/red luminescence.

Arguments for shallow burial conditions during dolomitisation are the mostly smooth crystal surfaces and the euhedral to subhedral dolomite mosaic. After Gregg & Sibley (1984), rough crystal surfaces and thus a xenotopic mosaic is energetically favoured above a 'critical roughening temperature', which exists for dolomite above ca. 50°C. Furthermore, baroque or saddle dolomite, which develops during deeper burial above ca. 60°C (Radke & Mathis 1980),

was not noticed at the Jebel Rheris. In addition, some dolomite crystals were truncated by fractures, and none were found associated with stylolites. It seems that dolomitisation occurred during one diagenetic stage, as only one type of dolomite was observed, in contrast to studies of Mattes & Mountjoy (1980), who distinguished five dolomite types from different diagenetic stages.

Source of the Mg^{2+} for dolomitisation at the Jebel Rheris was probably the about 70 m thick Emsian shale succession. It was frequently suggested that the conversion of smectite to illite during increasing burial releases Mg^{2+} (as well as Fe^{2+} and further ions) (e.g. Mattes & Mountjoy 1980, McHargue & Price 1982), which expulse during compactional dewatering of shales into overlying strata.

Later diagenetic stages include fracturing of the lithified and dolomitised rocks, where fractures are up to 2 mm wide, and subsequent precipitation of the dull-luminescent ferroan blocky spar II in the newly created cavities. These cemented fractures were later cut by stylolites. Also after dolomitisation, some of the remaining calcitic macro-fossils were dissolved, which lead to a considerable porosity in some areas of the Jebel Rheris (Pl. 6/7). Few of these cavities were again filled by calcite spar (Pl. 6/8), similar to observations of Mattes & Mountjoy (1980). Dissolution and reprecipitation of such large areas requires a certain amount of porosity and permeability for the movement of interstitial waters, which was probably created by the dolomitisation.

DIAGENETIC EVENT	marine	shallow burial	deep burial
Fibrous calcite	——		
Non-luminescent scalenohedral calcite		—	
Bright-luminescent blocky spar I		——	
Ferroan replacement dolomite		——	
Fracturing			—
Dull-luminescent blocky spar II			——
Pressure solution			—

Fig. 32: Timing and environments of diagenetic events in a Givetian biostrome succession.

10.5 Summary of diagenetic features

After the deposition of Givetian sediments, the first diagenetic step was the precipitation of marine fibrous calcite in few skeletal cavities (Fig. 32). Under shallow burial conditions, some scalenohedral calcite crystals grew, succeeded by rapid precipitation of mostly bright-luminescent blocky spar I, which closed the whole pore space. Thereafter, ferroan replacement dolomite grew in a depth, where the temperature probably did not exceed 50°C - 60°C. Dolomitisation obscured considerable areas of the Givetian rocks and crosses the original bedding, showing an irregular, patchy distribution. Smectite to illite conversion in Emsian shales probably provided Mg^{2+} for the dolomitisation. During deeper burial, the dolomitised rock was fractured, and the fractures were subsequently cemented with dull-luminescent blocky spar II. The filled fractures later were cut by stylolites.

Biostrome constructing macro-organisms sometimes were selectively not dolomitised and later dissolved, which created cavities and considerable porosity in the rock. Rarely, some cavities were filled again with calcite cement.

11. FAMENNIAN PHOSPHATIC BLACK PEBBLES

The term 'black pebble' was defined by Strasser (1984) as "reworked carbonate lithoclasts with a distinct dark grey to black penetrative coloration", Leinfelder (1987) defined 'black pebbles' as "blackened lithoclasts or fossil fragments". Generally, these stained components are considered as indicators of emergence in carbonate successions, as they are commonly associated with mud cracks, caliche, root-casts, or primary vadose cements (Strasser 1984). Black pebbles occur in transgressive lag deposits (Garzanti 1991) and may be concentrated in back-stepping parasequences within transgressive systems tracts (Baum & Vail 1988). They are poorly known from Palaeozoic settings, where the oldest were described by Sardeson (1914) from the Ordovician, but more abundant in the Mesozoic (e.g. Seyfried 1980, Strasser & Davaud 1983) and still form in Holocene carbonate sediments (Kendall & Skipwith 1969, Strasser 1984).

In Famennian deposits of the Jebel Rheris, several black pebble containing layers occur, sometimes with conglomeratic texture. These levels can be traced over several kilometres. Geochemical data reveal that the black pebbles of the present study are reworked phosphorites. Occurrences of phosphatic sedimentary rocks in Middle and Upper Ordovician deposits of the central and eastern Anti-Atlas have briefly been mentioned by Saadi (1986), but no detailed studies of Palaeozoic phosphorites of the eastern Anti-Atlas were published so far. In general, phosphatic black pebbles have not been described, so it is the aim of this chapter to discuss the age, mode of formation, and relevance for the detection of sea-level changes of phosphatic black pebbles.

11.1 Previous work

Black pebbles in limestones are widely known, but the causes of staining are divers. (1) Organic matter derived from decayed algae or terrestrial plants is frequently inferred, it may impregnate the sediment in anoxic and alkaline environments (Häfeli 1966, Kendall & Skipwith 1969, Ward et al. 1970, Bechstädt in Seyfried 1980, Strasser & Davaud 1983, Strasser 1984, Leinfelder 1987, Bernier & Strasser 1988, Lang & Tucci 1997). (2) Pyrite sometimes acts as a blackening agent: Sugden (1966) described pelletal limestones, which were stained grey to black by dispersed pyrite. Wright (1986) reported allochems, blackened after the drowning of a palaeosol, whereby iron from the soil and sulphate from sea water caused pyrite formation during anoxia. (3) Penetrative blackening of limestones by iron and manganese possibly took place on the sea floor during periods of non-deposition (Sardeson 1914). Reworking of these corrosion surfaces produced black limestone pebbles. Iron and manganese from terrigenous influx is shown to be responsible for the staining of carbonate sand at the Great Barrier Reef (Maiklem 1967), where organic material of foraminifera acted as nucleus for Fe-/Mn-precipitation in the reducing zone of the sediment. (4) Shinn & Lidz (1988) and Vera & Jimenez de Cisneros (1993) found selective blackening of limestones through forest fires. Their assumption was supported by heating experiments which showed that carbonates blacken at temperatures between 400° and 500° C.

Phosphatisation occurs mostly in the shallow marine environment. The bulk of contemporaneous phosphorites occurs in nutrient rich upwelling regions

off the South American west coast and off SW-Africa, where high biological productivity causes strong concentrations of organic matter on the sea floor. The bottom sediments get enriched in phosphorous due to the decay of plankton and fish scales, which leads to the precipitation of apatite in anoxic pore waters (Veeh et al. 1973, Manheim et al. 1975, Burnett 1977, Jahnke et al. 1983, Thomson et al. 1984, Froelich et al. 1988, Baturin 2000). Significant phosphorite deposits occur also in non-upwelling environments. O'Brien & Veeh (1980) reported phosphorites on the east Australian continental margin, which possibly formed due to low sedimentation rates, preventing the dilution of precipitated apatite. Ruttenberg & Berner (1993) discovered apatite formation in the Mississippi Delta and Purnachandra et al. (2002) showed that phosphorous from continental sources was utilised by bacteria to construct phosphate stromatolites off the southeast coast of India. Phosphorites occasionally form in terrestrial environments as products of wheathering (Flicoteaux et al. 1977, Southgate 1986) or bird excrements (guano) (Flicoteaux & Melfi 2000). Redeposited phosphorites were described by Summerhayes et al. (1972), who found sand-sized phosphatic detritus in unconsolidated sediments on the shelf off northwest Africa, which was eroded from Cretaceous and Tertiary phosphatic rocks.

In the geological record, periods of preferential phosphate accumulation alternate with others, where almost no phosphorites of economic importance were formed. Cook & McElhinny (1979) and Slansky (1980) concluded that favourable conditions for phosphorite genesis prevailed in the late Proterozoic, Early Cambrian, Ordovician, Permian, Late Cretaceous – Eocene, and Miocene. Poor conditions predominated during the Silurian, Devonian, Carboniferous, and Triassic. Nonetheless, Devonian phosphates of economic importance rarely exist, for instance in Iran (Movahhed et al. 1968).

11.2 Black-pebbles and nodules of the Jebel Rheris

In the upper part of the Famennian succession (Middle *expansa* – Middle *praesulcata* Zone) of section 1, black stained particles occur in a 22 m thick interval, which decreases to 3 m in section 10. These particles fall into two easily distinguishable types – types 1 and 2. In 13 examined thin sections, 413 type 1 black pebbles were counted, which are com-

posed of well sorted and rounded fine sand (average diameter 0.07 mm) in a brown to black matrix (Pl. 7/3). In addition, 74 black pebbles without internal components were counted. Only 7 type 2 black nodules were found which contain carbonate bioclasts (crinoids, bryozoans, calcimicrobes, ostracods, and shells) (Fig. 33). The brown to black matrix slightly penetrated the carbonate particles, causing a dark staining. In 4 black pebbles vertebrate remains were observed (Pl. 7/3) and a single one contained ooids with about 0.2 mm diameter (Pl. 7/6). Type 1 black pebbles and type 2 black nodules are often dispersed (Pl. 7/1), sometimes concentrated in discrete layers; their diameter varies between 0.3 millimetres and 3 centimetres.

The host rock consists mostly of crinoid ossicles and fragments of bryozoans, some brachiopods, molluscs, calcimicrobes, ostracods, and rare solitary rugose corals. Vertebrate remains are quite abundant (Pl. 7/7). Well rounded and sorted quartz grains have a portion of about 5 % of the rock. Their diameter is usually 0.5 mm, sometimes up to 3 mm. Horizontal lamination, trough cross bedding, and wave ripples sometimes occur.

The high amount of fragmented benthic organisms and the scarcity of fine grained sediment indicates that the black pebbles were deposited in a shallow marine, near-shore environment, influenced by storms.

11.3 Material and methods

As the type 1 black pebbles at the Jebel Rheris are more resistant to weathering than the surrounding limestone, they can be easily sampled, which was carried out at four locations in section 1, 2, and 4 (Fig. 15). From each location, 10 black pebbles with diameters of about 5 mm were ground and the mineral composition of the powder was determined by a *Philips X-ray Diffractometer PW 1040* with PC-controlling and secondary monochromator CuK-alpha radiation 40kV/20mA. Quantitative analysis was accomplished by *Siroquant Ver. 2.5*.

The same sample powder was decarbonised and subsequently processed with a *Rosemount Dohrmann TOC Boat Sampler, Model 183*, to determine the total organic carbon (TOC) of the insoluble residue. Simultaneously, the carbonate content of these samples was determined by colour-indicator titration to measure the TOC of the complete pulverised black pebble samples.

Sample	11	13	40	43
Quartz	40.2	39.0	44.6	37.7
Calcite	10.8	13.5	10.0	7.1
Apatite	41.8	35.4	39.0	53.6
Kaolinite	0.5	0.4	2.4	
Albite	6.7	11.7	4.0	
Microcline				1.6
TOC	0.076	0.075	0.099	0.243

Tab. 14: Mineral composition and total organic carbon (TOC) content of type 1 black pebbles (wt.%). Each sample represents the average composition of 10 black pebbles.
Error of mineral composition is 0.3 %.

To analyse the major element composition, a scanning electron microscope (SEM) type *Leo 1450 VP* was used, equipped with an EDX detector (*INCA energy 400*). All samples were sputter-coated with carbon, acceleration voltage amounted to 10 kV and 15 kV respectively. The major element composition of freshly fractured surfaces of 11 type 1 black pebbles was determined within 16 mm². Major elements of 3 type 2 black nodules, made up of carbonate bioclasts, were measured on the surface of thin sections within areas of about 4 mm² (Fig. 33).

11.4 Mineral composition and geochemistry

11.4.1 Type 1 black pebbles

A summary of the *mineral composition* of type 1 black pebbles is presented in Tab. 14. There is no significant difference between the samples of all four locations. As already observed in thin sections, quartz is a major constituent with about 40 wt.%. However, apatite often is the dominant mineral with up to 53.6

wt.%, it is obviously the major phase of the dark matrix between the quartz grains. Calcite, kaolinite and feldspar are of minor importance.

Organic carbon is an insignificant constituent of the black pebbles (see Tab. 14), varying between 0.075 wt.% in sample 13 and 0.243 wt.% in sample 43. In addition, the organic carbon content of two samples of the host rock was determined: Sample 13e contained 0.04 wt.%, sample 13d 0.02 wt.%.

The *major element composition* of 11 type 1 black pebbles is summarised in Tab. 15. As an exception, sample 13/1 contains no phosphorous, whereas the other samples have an average of 9.49 wt.% phosphorous. Fluorine is only present in three samples. In all measured black pebbles, calcium is the dominant element apart from ubiquitous oxygen; silica mostly ranks third. Aluminium and iron are always present, with an average amount of 2.15 wt.%. Sodium, potassium, and magnesium are common ingredients, but usually below 1 wt.%.

11.4.2 Type 2 black nodules

Type 2 black nodules also contain phosphorous (Fig. 33). P contents between 2.0 and 3.3 wt.% represent the average amount of the whole surface area within ca. 4 mm², i.e. including carbonate particles and quartz grains. Fluorine only exists in the blackened area of sample 13g (Fig. 33/A').

Additionally, five measurements within ca. 4 mm² were carried out in areas of two thin sections (13g and 13k), which are not blackened. The results (Tab. 16) show that carbonate is the major, and quartz a minor constituent; phosphorous was not measured.

Sample	11/1	11/2	11/3	13/1	13/2	13/3	40/1	40/2	40/3	43/1	43/2
Na	0.69	0.40	0.62		0.67	0.81	0.41				
Mg	0.67	0.41	0.55		0.49	0.24	0.63	0.33	0.46		
Al	2.94	2.40	2.66	0.92	2.47	2.12	3.93	1.21	1.84	1.59	1.53
Si	17.40	11.44	10.15	2.20	12.27	15.43	17.67	10.84	18.59	12.06	17.64
P	9.73	6.21	11.38		9.24	10.24	9.01	12.07	7.19	10.69	9.20
K	0.75	0.82	0.57		0.63	0.45	0.46			1.03	1.17
Ca	21.12	37.68	27.98	58.81	30.30	25.44	19.55	30.19	23.00	29.44	23.94
Fe	1.88	1.42	2.01	1.37	2.36	1.47	3.33	3.18	4.75	0.89	0.98
F			3.02							3.13	2.13
S				3.80							
Ti								0.20	0.30		
Zn									0.58		
O	44.81	39.21	41.06	32.89	41.57	43.80	44.81	42.18	43.29	41.17	43.41

Tab. 15: Major element composition of type 1 black pebbles (wt.%).
C concentrations are not considered, as samples were sputter-coated with carbon.

Sample	13g/1	13g/2	13k/1	13k/2	13k/3
Ca	31.0	34.3	31.8	29.2	27.7
Si	2.0	1.5	0.5	2.1	2.9
Mg	0.4		0.5	0.3	0.4
Al	0.4	0.3		0.3	
Fe	0.4				
O	65.8	63.9	67.2	68.1	69.0

Tab. 16: Major element composition (wt.%) of areas, which were not blackened.
C concentrations are not considered, as samples were sputter-coated with carbon.

11.5 Discussion and conclusions

Quartz and feldspar, which are common in the examined type 1 black pebbles, are detrital components, calcite may be derived from bioclasts or cements. Microcrystalline apatite surrounds the detrital components in the matrix and thus was probably precipitated authigenically after deposition. This conclusion is supported by the occurrence of phosphate ooids (Pl. 7/6). Similar ooids are known to grow authigenically at the sediment / water interface off South West Africa during the Late Miocene (Birch 1980). Phosphatisation must be responsible for the black staining of the pebbles at the Jebel Rheris, because clay minerals and organic material are only present in insignificant amounts (Tab. 14), and because areas without a dark matrix contain no phosphorous (Tab. 16). Still phosphatisation does not necessarily cause a dark staining. O'Brien & Veeh (1980) describe phosphate nodules on the east Australian continental margin, ranging in colour from dark grey to near white.

Black pebbles at the Jebel Rheris contain between 35.4 – 53.6 wt.% of apatite, so the strata, from which the black pebbles were reworked, consist at least partially of phosphorites. After Bushinsky (1969), sedimentary rocks containing more than 50 % apatite minerals (i.e. more than 18 % P_2O_5) are called phosphorites; this portion of apatite in rocks may be of economic importance (e.g. Rooney & Kerr 1967).

11.5.1 Age of phosphatisation

Two different phosphatized facies types can be distinguished in the study area. Black sandstone pebbles (type 1) prevail, though hardly any sandstones occur in the Devonian deposits of the mountain. Few thin sandstone layers are intercalated in lower Famennian strata, but they never show a dark staining and due to their rare occurrence, they can not be the source of the considerable amount of type 1 black pebbles. These must be derived from the emerged area north of the Jebel Rheris. First sandstones occur in the Ordovician (Llandeilo) (Destombes et al. 1985), so it is likely that the black sandstone pebbles are Ordovician in age and were phosphatized during that time. Chauff (1978) described a comparable situation in a Palaeozoic succession of Missouri (USA): He found grey to black phosphate pebbles within an Upper Devonian cross-bedded, conglomeratic oo-biosparite, interpreted as a tidal channel deposit. It was possible to recover Ordovician conodonts from the phosphate pebbles, which proved that they were reworked from Ordovician strata, probably shales. In the present study, however, the type 1 phosphatic pebbles are reworked sandstones, so the attempt to gain microfossils from the pebbles by dissolving them was not made, the determination of their age is merely based on lithofacies comparison.

In contrast to the type 1 black pebbles, type 2 black nodules of the present study consist of the same carbonate bioclasts as in the upper Famennian rocks. Sometimes, no sharp contact exists between phosphatic and non-phosphatic zones but a smooth transition, which shows that phosphatisation and staining occurred *in situ* (Fig. 33/A). This indicates that a patchy phosphatisation took place during the late Famennian at the Jebel Rheris.

11.5.2 Formation of phosphorites

The depositional environment of the Ordovician sandstones, from which the phosphatic type 1 black pebbles probably derived, is described by Destombes et al. (1985) as shallow marine. The amount of organic material, usually an important carrier of phosphorous, is insignificant (0.12 wt.%), compared to about 2 % in the Permian Phosphoria Formation (Gulbrandsen 1966) and a Miocene phosphorite in North Carolina (Rooney & Kerr 1967). Therefore upwelling, providing nutrients and causing excess primary bioproduction, which is a major reason for phosphatisation in modern seas (see chapter 11.1), is unlikely. Instead, rivers which are known to carry significant amounts into the ocean (Kolodny 1981), may have transported phosphorous onto the shelf.

Type 2 phosphorite nodules with carbonate bioclasts formed in a shallow marine, storm influenced, nearshore environment. A major source for the phosphorous may have been vertebrate remains, which are abundant in the surrounding sediment (Pl. 7/7).

	wt.%
O	36.85
F	1.21
Mg	0.28
Al	0.34
P	3.30
Ca	25.23
Si	32.79

	wt.%
O	46.93
P	2.11
Ca	9.76
Si	41.20

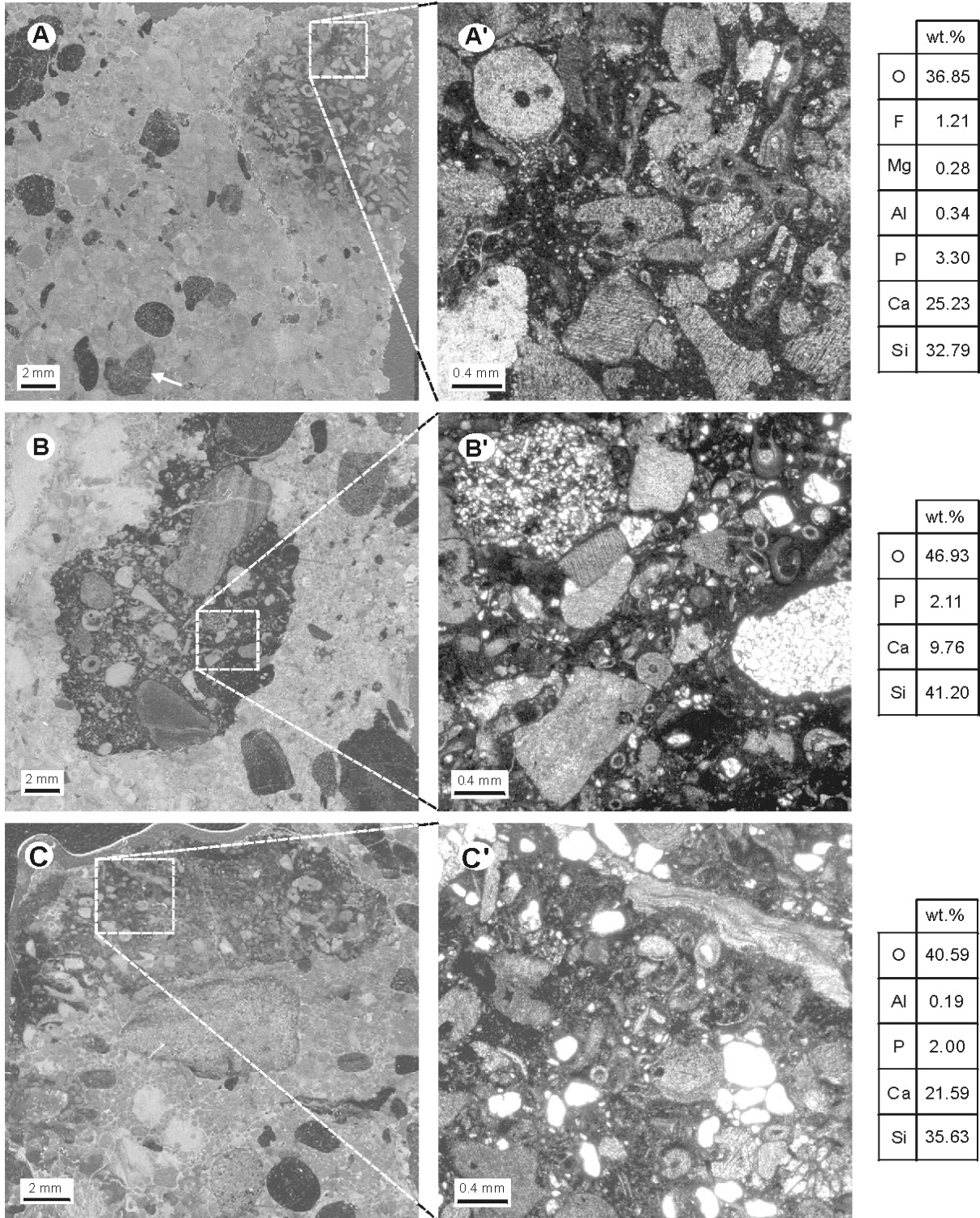|	wt.%
O	40.59
Al	0.19
P	2.00
Ca	21.59
Si	35.63

Fig. 33: Photographs of thin sections of type 2 black nodules, consisting of carbonate bioclasts. The major element composition of these areas, measured with an EDX detector, is shown on the right margin.
(A) Black pebbles and a vertebrate remain (arrow) in a crinoidal limestone (sample 13g, section 4). Notice diffuse blackened area in the upper right edge. (A') Close-up of the blackened area of A): Crinoid ossicles and fragments of bryozoans are surrounded by a dark matrix. The table aside shows the major element composition of a 3 x 1.5 mm large area within A'. (B) Type 2 black nodule, consisting of crinoid ossicles, fragments of bryozoans, and smaller type 1 black pebbles (sample 13k, section 4). (B') Close-up of B), wherein the major element composition was determined within 2 x 2 mm. (C) A 2 cm large type 2 black nodule at the top of the photograph contains fragments of crinoids, bryozoans and brachiopods as well as quartz grains (sample 13b, section 4). (C') Close-up of C). EDX measurements were carried out within a 2 x 1 mm large area of C'.
Carbon concentrations are not considered, as thin sections were sputter-coated with C.

The high amount of quartz suggests that continental input was another source of phosphorous. Recent phosphorite formation occurs in the following depositional environments: (1) Upwelling conditions off the American and African west coasts, characterized by high primary organic production, lead to the accumulation of organic-rich ooze in shallow waters (e.g. Burnett 1977, Jahnke et al. 1983, Thomson et al. 1984, Baturin 2000). (2) Off east Australia, phosphorites form in organic-poor sediments at about 400 m water depth due to low accumulation rates (O'Brien & Veeh 1980). (3) Apatite formation in nearshore carbonate sediments at Bermuda (Gaudette & Lyons 1980) depends on a high amount of organic matter. (4) Schwennicke et al. (2000) concluded that restricted water circulation and high bioproductivity may cause phosphogenesis in shallow marine environments from Cambrian to Recent. Non of these examples show similar conditions of phosphatisation described in the present study, which shows that phosphorite nodules may form in coarse bioclastic grainstones, without significant amounts of organic matter.

Phosphogenesis often is related to anoxic conditions in the marine environment (e.g. Flicoteaux & Trompette 1998, Schenau et al. 2000) and the time, when the type 2 phosphatic nodules developed at the Jebel Rheris correlates with the uppermost Famennian global anoxic Hangenberg Event. In low-latitude palaeogeographical positions around the globe, organic- and phosphatic-rich black shales were deposited in the Upper *expansa* to Upper *praesulcata* Zone (compilation in Caplan & Bustin 1999). Investigations in an upper Famennian carbonate ramp of Alberta, Canada, showed that the phosphate content already increased within a crinoidal and skeletal grainstone succession underneath an organic-rich Tournaisian black shale (Caplan et al. 1996), which is explained with progressive eutrophication. However, no organic-rich black shales exist in the Devonian succession of the Jebel Rheris, no indicators of an oxygen minimum zone or anoxia occur.

11.5.3 Relationship of black phosphatic pebbles / nodules to sea-level changes and sequence stratigraphy

As described in chapter 6.5, upper Famennian strata of the Jebel Rheris area represent a TST. In the northern and western parts of the working area, several unconformities occur, whereby conodont data from beds above the unconformities reveal four different upper Famennian ages. No regressions can be noticed in upper Famennian sections without stratigraphic breaks in the southern Jebel Rheris area, so the beds above the unconformities are interpreted as parts of different backstepping retrogradational parasequences, which belong to the same TST. The progressing coastal erosion during the late Famennian transgression reached the black phosphorites of probably Ordovician age farther north. The reworked type 1 black pebbles were subsequently shed onto the shelf and covered an area several kilometres wide. So the beds with type 1 black pebbles in the sections of the southern Jebel Rheris area (Fig. 15, 16) represent the distal parts of parasequences of a TST.

In contrast to type 1 black pebbles, type 2 black phosphorite nodules formed in situ and are probably not directly linked to sea-level changes.

12. CONCLUSIONS

In the northern Mader south of the emergent Jebel Ougnate, a carbonate shelf existed from early Givetian to late Famennian times, which developed on top of Emsian and Eifelian pelagic nodular limestones and shales. Homogeneous thickness and facies pattern of the latter show that stable tectonic conditions prevailed during that time. The deposition of coral-stromatoporoid biostromes and crinoidal grainstones onto the pelagic facies from early Givetian times on indicates a seaward-stepping of the facies belts (Fig. 34). A carbonate-ramp geometry during the Givetian is inferred from the gradual shift from biostromal limestones in the north, which were deposited below fair-weather wave-base, to crinoidal limestones in the south, deposited in a slightly deeper environment on the mid-ramp. The distally steepened part of this ramp consists of styliolinid-rich distal tempestites with convolute-bedding structures. Neither peritidal inner-ramp nor basinal Givetian deposits are preserved.

From the Givetian to the Upper Devonian, a seaward-stepping of the facies belts occurred, which caused the deposition of near-shore sediments (stromatolite limestone, lagoonal muds, and crinoid shoals) onto the Givetian mid-ramp facies. Thick shallow-marine conglomerates and some sandstone layers are further indicators that the coastline was not far to the N from the Jebel Rheris during the Late Devonian. The facies pattern indicates that a
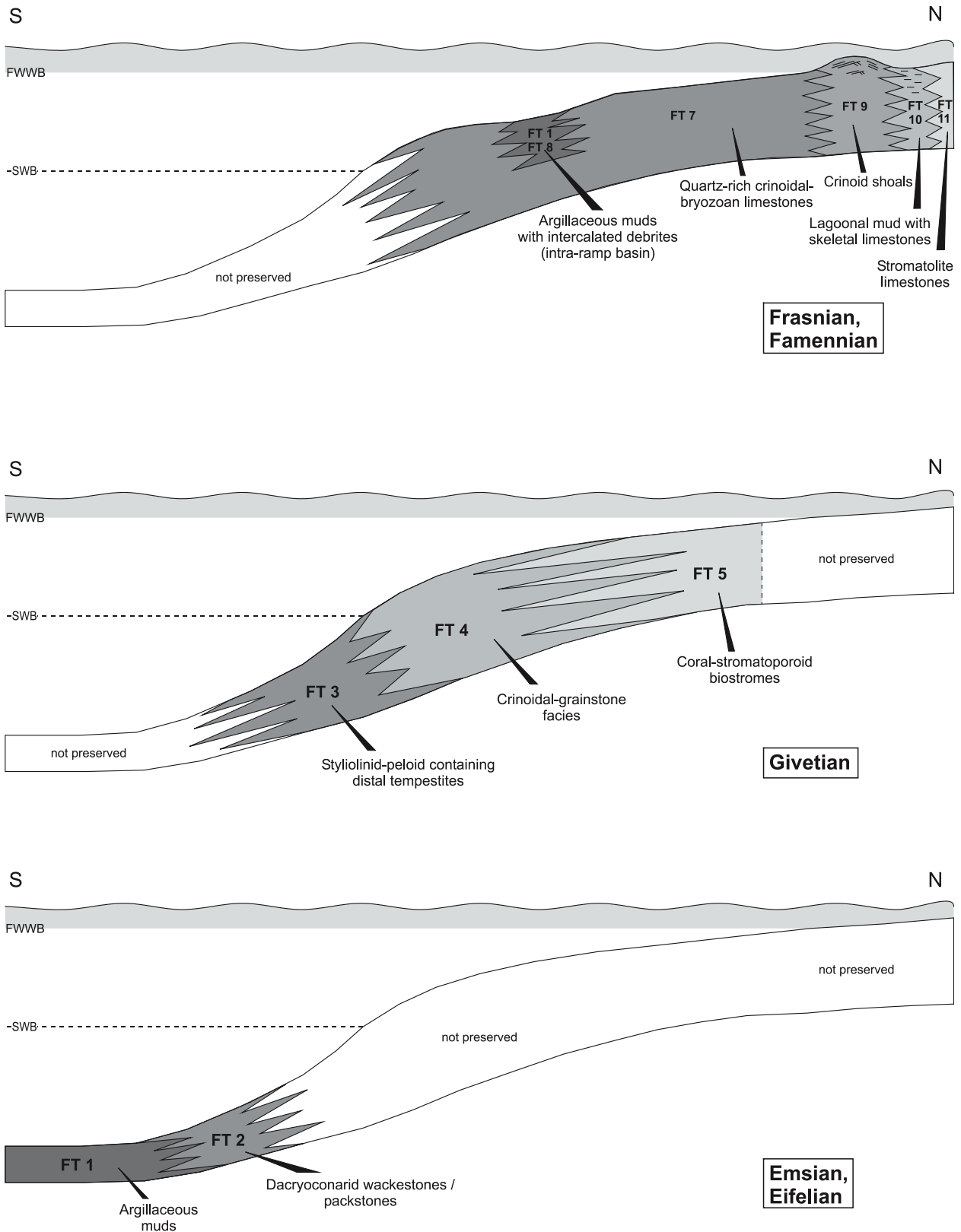
S   N

FWWB

-SWB-

not preserved

FT 1
FT 8

FT 7

FT 9

FT
10

FT
11

Quartz-rich crinoidal-
bryozoan limestones

Crinoid shoals

Argillaceous muds
with intercalated debrites
(intra-ramp basin)

Lagoonal mud with
skeletal limestones

Stromatolite
limestones

**Frasnian,
Famennian**

S   N

FWWB

-SWB-

not preserved

FT 5

FT 4

FT 3

Coral-stromatoporoid
biostromes

Crinoidal-grainstone
facies

Styliolinid-peloid containing
distal tempestites

not preserved

Givetian

S   N

FWWB

-SWB-

not preserved

not preserved

FT 2

FT 1

Dacryoconarid wackestones /
packstones

Argillaceous
muds

**Emsian,
Eifelian**

Fig. 34: Facies models for Emsian to Famennian deposits of the Jebel Rheris. Note the large-scale seaward-shift of facies belts, caused by a long-term relative sea-level fall from the Eifelian to the Famennian. FWWB: Fair weather wave base; SWB: Storm wave base; FT: Facies type.

carbonate ramp geometry continued, but only the inner ramp and mid-ramp part is preserved (Fig. 34), including a small intra-ramp basin.

While the Early and Middle Devonian were times of relatively stable conditions, deposition during the Late Devonian was considerably influenced by tectonic activity. It is expressed by synsedimentary faulting and neptunian dikes; moreover Late Devonian erosion and tilting caused hiatuses and angular unconformities. This tectonic activity is associated with the uplift of the hinterland, which probably influenced the evolution of the carbonate shelf since the Middle Devonian. A large-scale seaward-stepping of facies belts, i.e. a long-term relative sea-level fall, occurred from the Eifelian to the Famennian (Fig. 34). This relative sea-level fall during a time-span of approximately 30 Ma is related to a second order stratigraphic cycle.

The palaeolatitude of the investigated area during the Middle and Late Devonian is controversially discussed (see chapter 2 for references). However the organisms, composing the Givetian biostromes, are typical representatives of a 'Photozoan Association'. This term is used by James (1997) for warm-water, benthic calcareous communities. Furthermore ooids, which are considered to form only in warm-water environments (James 1997), occur in Upper Devonian rocks. Therefore, a warm-water depositional realm at about 30°S for the Middle and Late Devonian can be concluded.

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