Middle and Upper Devonian Palaeoenvironments in and around the Rodheim - Bieber Carbonate Complex (Lahn Syncline, West Germany)

by

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(p = photographs)

Abstract

A description and environmental interpretation is presented of the various facies both within and surrounding the Rodheim-Bieber Carbonate Complex, an isolated limestone unit some 400m. thick at most and , at present,about 5km. in diameter.

The limestones of the complex were deposited on a local volcanic seamount produced by eruption of pyroclastic material in the late Givetian within the Rhenish trough of the Variscan Geosyncline, the extrusion of volcanic material near to or above sea level creating a greater range of environments on the summit of the seamount than are seen in the neighbouring, deeper water basin and submarine rise areas. In these latter areas sedimentation was slow, the limestones developed on submarine ridges and the non-carbonates deposited in the intervening basins being both fine grained and pelagic in nature, evidence of any benthonic organic contribution to sediment formation being virtually absent. Sediments of this type are developed on the slopes of the Rodheim - BieberRise.

On the summit of the rise, carbonate sedimentation began with the formation of limestones considered to have formed under intertidal to shallow subtidal conditions (Schwelm Facies). In the early Upper Devonian, the limestones deposited show a great deal of lateral variation resulting from the development of a reef rim along the eastern edge of the submarine rise. Mapping of the facies patterns within this l limestone group (Dorp Facies) indicates that at that time the

carbonate body was atoll-like both in general outline and in environmental character, with lagoonal sediments developed in the central area of the seamount plateau behind a biologically constructed reef rim bordering the eastern edge of the seamount. The atoll stage in the development of the complex was abruptly brought to an end in the M. Adorf by updoming of the volcanic basement resulting in emergence and consequent destruction of the reef-rim biota. The limestones were extensively fissured and karst eroded at this time. Renewed subsidence is recorded in the M. - U. Adorf by the abrupt appearance of crinoidal calcarenites over karst eroded Dorp limestones.These limestones are replaced gradually by nodular limestones of pelagic type as the summit of the seamount sank below the critical depth below which shallow-water limestones could no longer develope. Only with renewed volcanicity and associated tectonic activity in the L. Carboniferous were shallow water conditions reestablished briefly on the Bodheim-Bieber Schwelle (Oolitic Facies) before the whole area was affected by the influx of the Culm greywackes from the south immediately preceding orogenic deformation of the area.

Few if any of the sediments have remained unaffected by diagenetic changes. Particularly striking in the case of the massive limestones are the changes brought about by late (?Tertiary) metasomatism. The textural affects and consequences of these changes are dealt with in the text.

Acknowledgements

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I have composed this thesis which describes my own work. Where results of other authors are referred to, this is clearly indicated.

CHAPTER 1

Introduction

Regional Setting

The Middle and Upper Devonian rocks of the Eastern Rheinisches Schiefergebirge show a great deal of lateral variation. Continental clastic sediments (lying just north of the main mass of the Schiefergebirge) pass southwards into impure mixed detrital sediments rich in bivalves, brachiopods and other shallow-water marine benthos ("Rhenish Magnafacies" of Erben, 1964; "External Shelf" of Krebs, 1971) in the synclinal area of the Sauerlands (Figs 1.1 and 1.2). South of the Siegen Anticline, composed almost entirely of Lower Devonian sediments, in the Lahn-Dill Syncline, the character of the sediment changes to a "mixture" of fine grained limestones and shales, with common "reef" limestones and submarine volcanics. These rocks are thought to have been deposited in the deeper trough or eugeosynclinal area in the Rhenish geosyncline, the shales in the basin areas ("Becken"), the limestones on submarine rises ("Schwellen"). These rises are commonly of volcanic origin, though tectonic horsts do also occur (Schmidt 1926; Rabian, 1956) (Fig 1.3). The nature of the limestone developed on these rises was dependent on the depth of the summit below sea level. ("Hercynian (Bohemian) Magnafacies" of Erben, 19641 "Axial Trough" of Krebs, 1971. There is

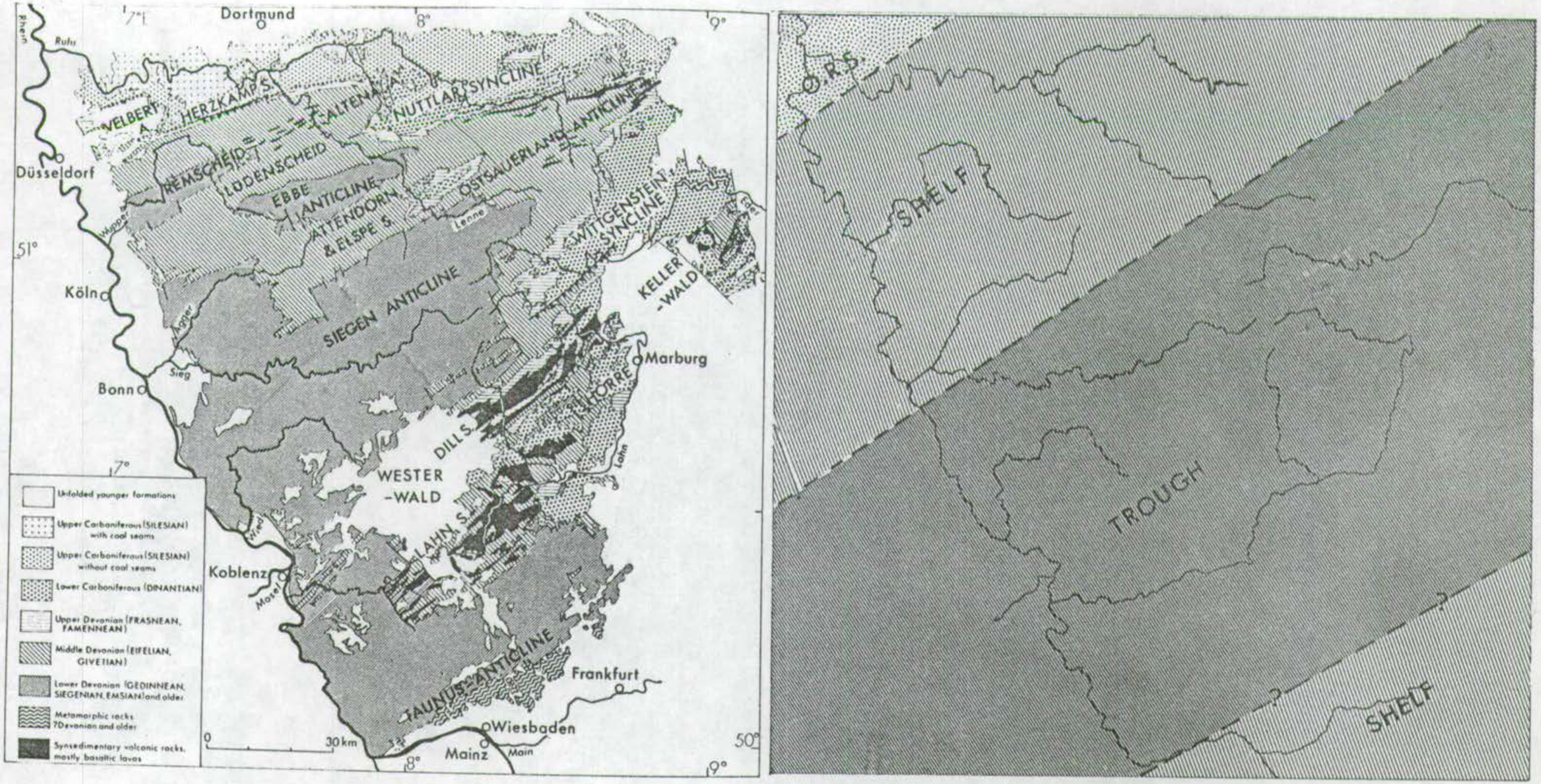


Fig. 1·1 a. Simplified geological map of the Rheinish Schiefergebirge (from Meischner, 1971).
b. Major palaeogeographic zones for the Middle Devonian superimposed.

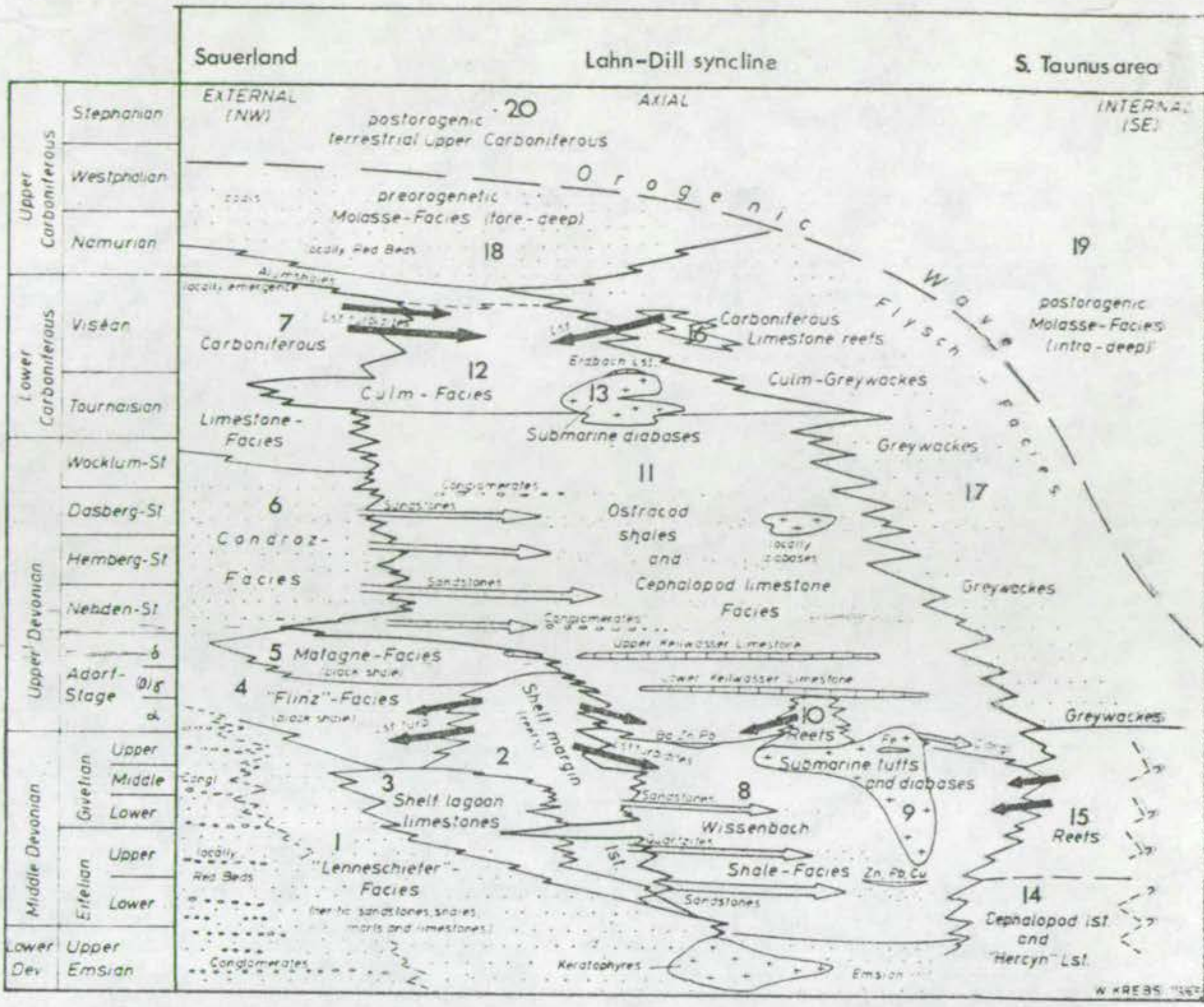


Fig. 1·2 Diagrammatic cross section through the Rhenish Trough of the Variscan Geosyncline from late Lower Devonian to Upper Carboniferous. 1-7= external shelf, 8-13= trough, 14-16= internal shelf, 17= flysch, 18= preorogenic molasse, 19-20= postorogenic molasse.
(slightly modified after Krebs,1971)

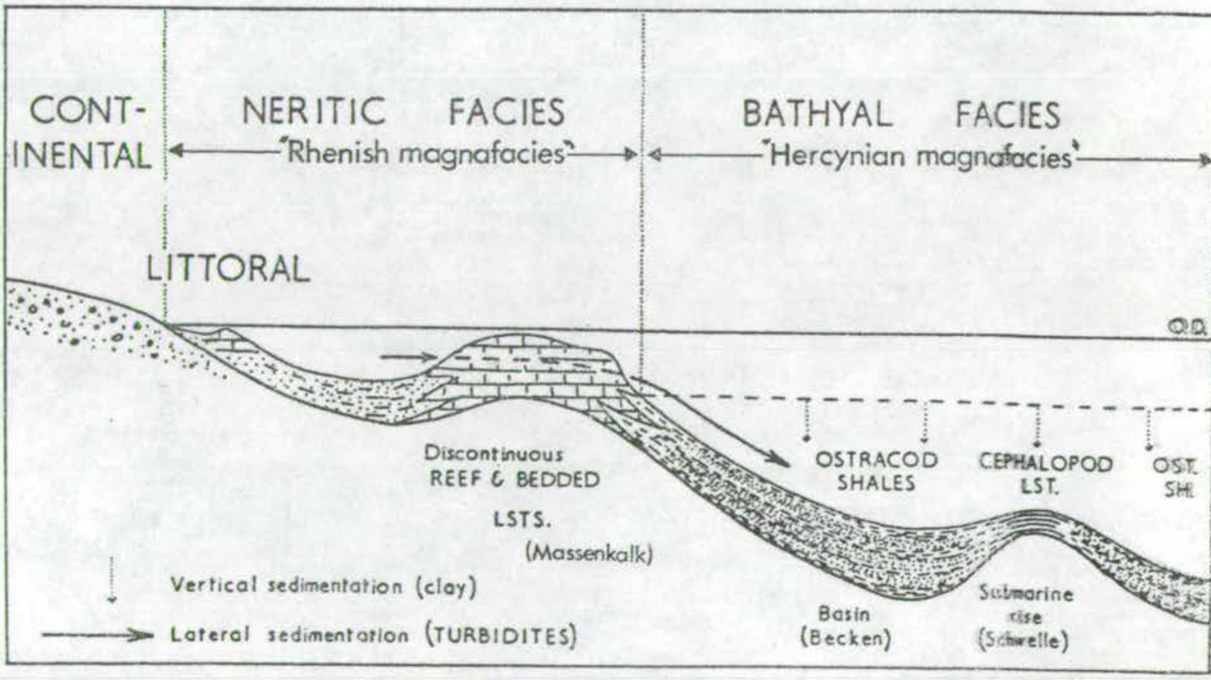


Fig. 1·3 Diagrammatic cross section illustrating the distribution of major facies in the Variscan Geosyncline and their major characteristics. (modified after Goldring,1962).

evidence that farther south in the Taunus region (Figs 1.1 and 1.2), where the rocks have been metamorphosed during the Variscan Orogeny, shallow water sediments are again located (Krebs, 1970) ("Internal Shelf" of Krebs, 1971; "Mitteldeutsch Schwelle" of Brinkmann, 1948 and Henningson, 1970). Indeed the massive limestones south of Giessen (fig 1.4) are considered by Krebs (op. cit.) to be of shelf type, as distinct from the isolated limestone masses found within the trough area of the Lahn Syncline.

The area dealt with in this thesis lies in the Lahn Syncline, close to the eastern edge of the Rheinisches Schiefergebirge, about ten kilometres northwest of the town of Giessen (Fig 1.4).

Very little work has been done on the Devonian rocks in this area since the region was mapped by Kegel (1933). Prior to this, mention had been made to the area only in a more general stratigraphic context by Ahlburg (1908, 1910), Kegel (1922) and Schwartz (1925), and by Beyer (1896), Maurer (1875, 1889), Parkinson (1903), and Sommer (1909) who published palaeontological lists from several localities within the area. Since 1933, the only detailed work done in the area has been that of P.Bender (1965, 1969) who, at the time of writing, was engaged in remapping the area north of Dünsberg (Fig 1.5). The Lower Carboniferous greywackes in the southeast and east of the area have been studied by Henningson (1961, 1966, 1968).

Objective

The purpose of this study is to describe and give

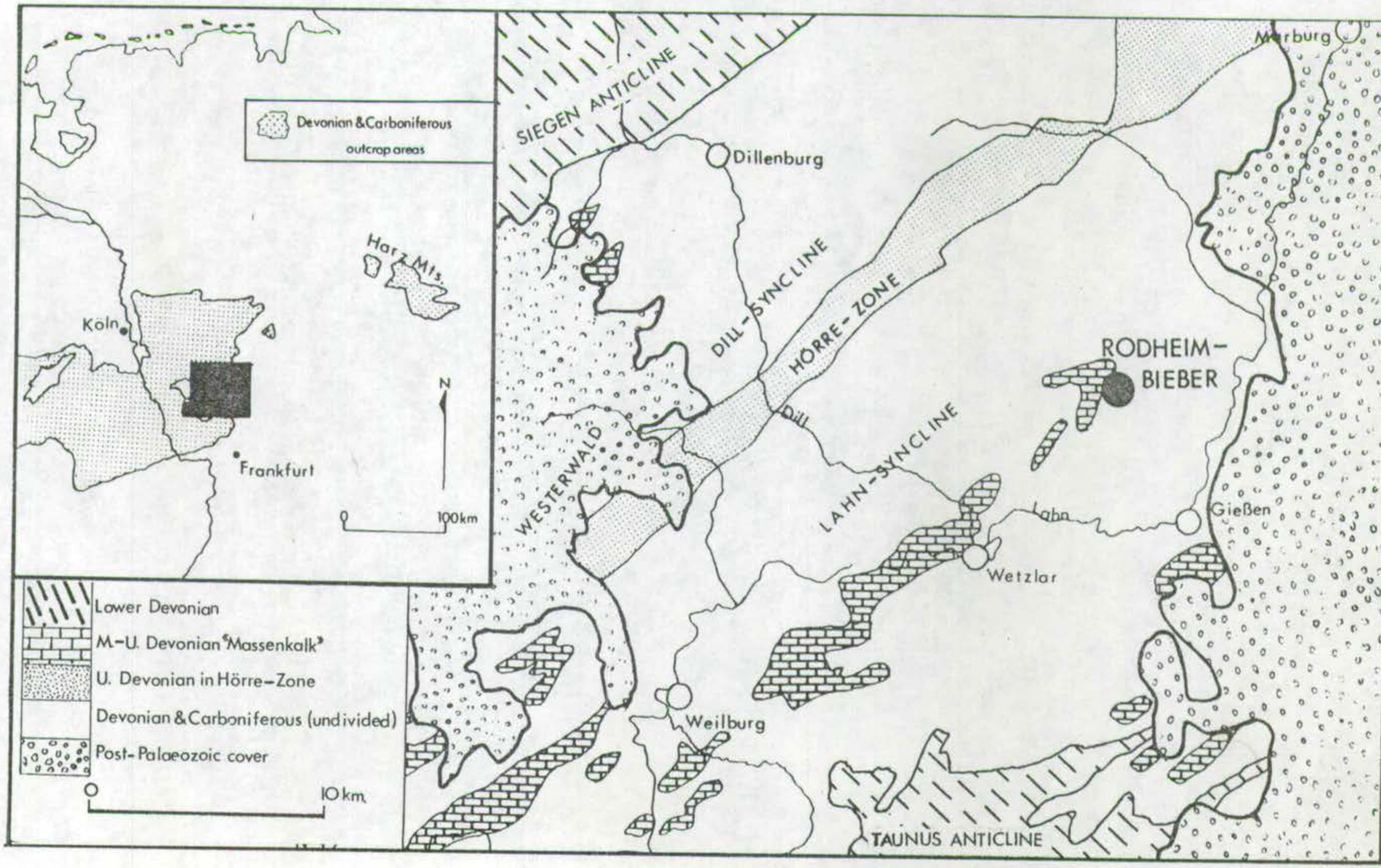


Fig. 1.4 Map showing the location of Rodheim-Bieber in relation to the main geological features of the Lahn Syncline.

an environmental interpretation of the various facies both within and surrounding the Rodheim-Bieber carbonate complex.

Very early in the study, it became obvious that to achieve this aim very precise biostratigraphic work was essential to relate lithologics from different exposures, as the rocks are, in the main, highly tectonized. Detailed study of the rocks themselves also revealed that many of their textural and mineralogical features were the result of post-depositional not syndepositional processes. Careful study of diagenetic and later processes was therefore thought necessary before conclusions could be drawn about the depositional environments in which the rocks were fomed.

Method of Study

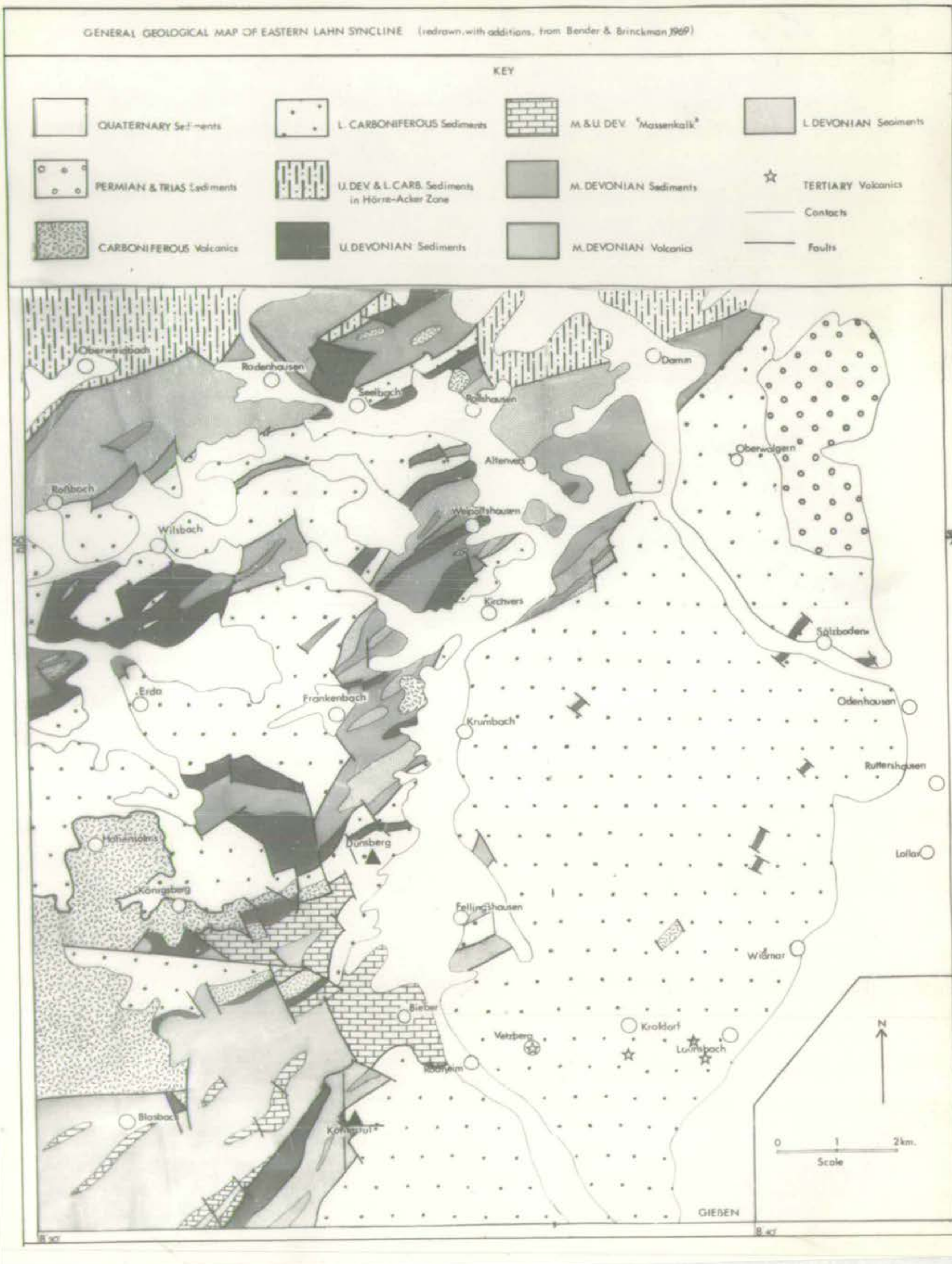
All areas mapped as Middle or Upper Devonian by Kegel (1933) on the lower half of Geologisch Karte, Blatt 3106, Rodheim, were examined for exposures. Localities in the northern area of the map, recently dated by Bender (1965), were also visited and samples collected for comparative purposes, no detailed petrographic work having been done by Bender (op. cit.).

The majority of exposures did not lend themselves to detailed field study however, for the following reasons:

(a) exposures were generally too small for details of relationships between individual lithologies to be shown;

(b) in many places, tectonic deformation had disturbed

Fig. 1.5


GENERAL GEOLOGICAL MAP OF EASTERN LAHN SYNCLINE (redrawn, with additions, from Bender & Brinckman 1969)
KEY
QUATERNARY Sediments
L. CARBONIFEROUS Sediments
M. & U. DEV. 'Massenkalk'
L. DEVONIAN Sediments
PERMIAN & TRIAS Sediments
U. DEV. & L. CARB. Sediments in Hörre-Acker Zone
M. DEVONIAN Sediments
TERTIARY Volcanics
Contacts
CARBONIFEROUS Volcanics
U. DEVONIAN Sediments
M. DEVONIAN Volcanics
Faults
Rodenhausen
Seelbach
Damm
Oberwalgern
Altenvers
Roßbach
Wilsbach
Weipoltshausen
Kirchvers
Sälzbaden
Erda
Frankenbach
Krumbach
Odenhausen
Ruttershausen
Dünsberg
Lollar
Königsberg
Fellingshausen
Wißmar
Bieber
Krofdorf
Vetzberg
Launsbach
Blasbach
N
0
1
2km.
Scale
GIEßEN

the rocks to such a degree that very little of the primary texture could be seen;

(c) the surfaces of many exposures were often weathered or covered by vegetation, flowstone, or slickensides.

The largest areas of exposure were found in quarries at Eberstein (69140, 11780) and Bieber (10360, 70885) (Map 1, Appendix 2) but, as the former is at present being worked, detailed examination of faces was restricted by time and by the shattered and dusty nature of the faces. Only in Bieber quarry could detailed field examination be carried out, but here, faulting and dolomitization restricted the areas of quarry wall suitable for study.

Many specimens, therefore, had to be collected for detailed laboratory examination. All samples were slabbed and ground flat, and either peels or sections (some stained by the method described by Dickson (1964)) made from all specimens. Staining slabs with methylene blue before taking peels was found to enhance textural details in some samples (especially microsparites and micrites). Certain specimens were also selected for microfossil extraction and for estimation of insoluble residue content.

Stratigraphy

Prior to the 1960s, zonation of the Devonian and Carboniferous in the Rheinisches Schiefergebirge was based mainly on the use of ammonoids, the Middle Devonian being divided into two major "stufen", the Upper Devonian into six, and the Lower Carboniferous into three.

Fig 1.6

Correlation of Middle and Upper Devonian, and Lower Carboniferous conodont zones with ammonoid "stufen" in the Rheinisches Schiefergebirge. *

A. - Ancyrognathus.

Gn. - Gnathodus.

I. - Icriodus.

P. - Palmatolepis.

Pg. - Polygnathodus.

Pol.- Polygnathus.

Pro.- Protognathodus.

Ps. - Pseudopolygnathus.

S. - Spathognathodus.

Sc. - Scaliognathus.

Sch.- Schmidtognathus.

Sg. - Scaphignathus.

Si. - Siphonodella.

* Since compilation of this table, a new conodont zone, the lower "rhomboidea" zone, has been established between the "crepida" and "rhomboidea" zones, the latter having been renamed the Upper Crepida-zone. The succeeding quadrantinodosa-zone has also been renamed the marginifera-zone. (Sandberg and Ziegler, 1973).

SERIES	STAGES	"STUFEN"	AMMONOID ZONES				CONODONT ZONES
L. CARBONIFEROUS	VISEAN	GONIATITES	Goniatites	Cu III	γ		Pg. nodosus
					β		Gn. bilineatus bilineatus
					α		
		PERICYCLUS	Pericyclus	Cu II	δ		anchoralis–bilineatus interregnum
					β/γ	upper	Sc. anchoralis
						lower	
	TOURNASIAN				α	upper	Si. crenulata
						lower	
		GATTEN-DORFIA	Gattendorfia	Cu I			Ps. triangula triangula
							Ps. triangula inaequalis
U. DEVONIAN	FAMENNIAN	WOCKLUM	Wocklumeria	to VI			Pro. kockeli
						?	? ?
				to VI		upper	S. costatus
						middle	
		DASBERG	Goniaclymenia	to V/VI		lower	
				to V		upper	Pol. styriacus
				to IV/V		middle	
		HEMBERG	Platyclymenia	to IV		lower	
				to III	β	upper	Sg. velifer
						middle	
					α	lower	
		NEHDEN	Cheiloceras	to II	β	upper	P. quadrantinodosa
						lower	
							P. rhomboidea
					α	upper	P. crepida
						middle	
						lower	
	--?--	ADORF	Manticoceras	post to Iδ		upper	P. triangularis
						middle	
						lower	
	FRASNIAN			to I	δ	uppermost	P. gigas
					γ/δ	upper	
						lower	
					γ		A. triangularis
					(β)γ	upper	Pol. asymmetricus
						middle	
					α	lower	
						lowermost	
					?	upper	Sch. hermanni–cristatus
M. DEVONIAN	GIVETIAN		Maenioceras	tm		lower	
						upper	Pol. varcus
						lower	
						upper	I. obliquimarginatus
						lower	

In the late '50s and early '60s attempts were made to establish a fuller, more reliable and more universally applicable zonation based on more common microfossils. Thus zonal schemes, based on conodonts, were established for the Middle Devonian (Bischoff and Ziegler, 1957; Wittekindt, 1955; Ziegler, 1965, 1971), the Upper Devonian (Ziegler, 1958, 1962, 1969, 1971), and the Lower Carboniferous (Bischoff, 1957; Voges, 1959, 1960; Meischner, 1970) (Fig 1.6).

The sequence of conodont zones recognized in the German Devonian has since been found in North America (Winder, 1966; Clark and Ethington, 1966; Pollock, 1968; Klapper et al, 1971 etc.) and in Australia (Glenister and Klapper, 1966; Druce, 1969; Seddon, 1970). With very minor modifications, this zonal scheme appears to have almost worldwide applicability. The sequence erected by Voges (1969), and modified by Meischner (1970), does not appear to be recognizable on a worldwide scale, or even between different parts of Europe, due to the provincial nature of many Lower Carboniferous conodonts. Since both the Devonian and Carboniferous sequences were established within the Rheinisches Schiefergebirge, however, both can be used with confidence in the studied area.

Conodont zonation in the Rodheim-Bieber area

156 samples from the studied area were treated for conodonts using the methods described by Collinson (1963). Of the 156, 44 samples were either completely barren or

contained only bars and blades of little or no stratigraphic value. The remaining 112 either contained faunas rich enough to enable a precise age to be given, or their age could be estimated using supplimentary information from samples collected from the same locality. A list of the samples with their stratigraphic and geographic locations is given in Appendix III. Sample locations are also given on Map 2, Appendix I, and faunal lists are tabulated in Tables 1-6, Appendix III.

Further dates for certain localities (mainly outside the main study area) were obtained from samples collected and dated by P.Bender (1965, 1969) and D.Henningson (1966).

In addition to the conodont ages above, a few scattered pre-1933 macropalaeontological investigations provided additional evidence of age at certain localities (Beyer, 1896; Maurer, 1875, 1889; Parkinson, 1908; Sommer 1909). Although very common within the massive limestones, however, macrofossils (e.g. stromatoporoids, tabulate corals and brachiopods) are of little stratigraphical value because of the relatively long time ranges of species present.

Stratigraphical problems

Even with as good a means of correlation as conodonts, precise correlation was hampered by the following problems.

(a) The nature of the rock types present.

In the Bieber area, the complex interrelationships between rock types severely limits the use of litho-stratigraphical correlation methods. Furthermore, the lithologies where an accurate means of correlation is most needed are often those poorest in conodonts (e.g. the massive limestones).

(b) Tectonic disturbance.

All of the rocks in the studied area have been affected to a greater or lesser degree by Variscan and later earth movements. Although none of the studied rocks can be described as metamorphic, the development of cleavage in the basinal shales and to a limited extent within some of the shaly limestones, neomorphism and dolomitization within limestone sequences often make assessment of true thickness of sedimentary sequences difficult.

(c) Difficulties in relating microfossil dates to rock ages on published maps.

Due to the relative lack of biostratigraphic control (ammonoids being uncommon or absent in many of the rocks in the area) and the consequent high reliance on litho-stratigraphic correlation methods, the published maps contain many serious inaccuracies especially in the basin areas where similar poorly fossiliferous lithologies occur several times in the Devonian and Carboniferous. As much of the tectonic interpretation of the area depends on an accurate knowledge of the age and distribution of the strata, the structural interpretation of these areas is brought into question and revision of these maps is required. (At the time field-work was being done (1972), remapping of the area north of Dünsberg was being carried out by P.Bender of Marburg University). Fortunately the southern part the area shown in Fig 1.5 contains fewer major errors because macrofossils are common in the majority of sediments of this area. The use of conodonts has however resulted in a refinement of correlation in the area. As a result, the ages of many sediments mapped by Kegel (1933) have been revised.

(d) Limits of exposure.

Apart from the quarries at Eberstein and Bieber, where exposure was more or less continuous, outcrops were normally small. Because of the massive and frequently inhomogeneous nature of the massive limestones, it was often difficult to decide whether two dissimilar but adjacent outcrops were in their correct stratigraphical position or had been tectonically juxtaposed. For this reason, accurate field measurement of complete sections was almost impossible.

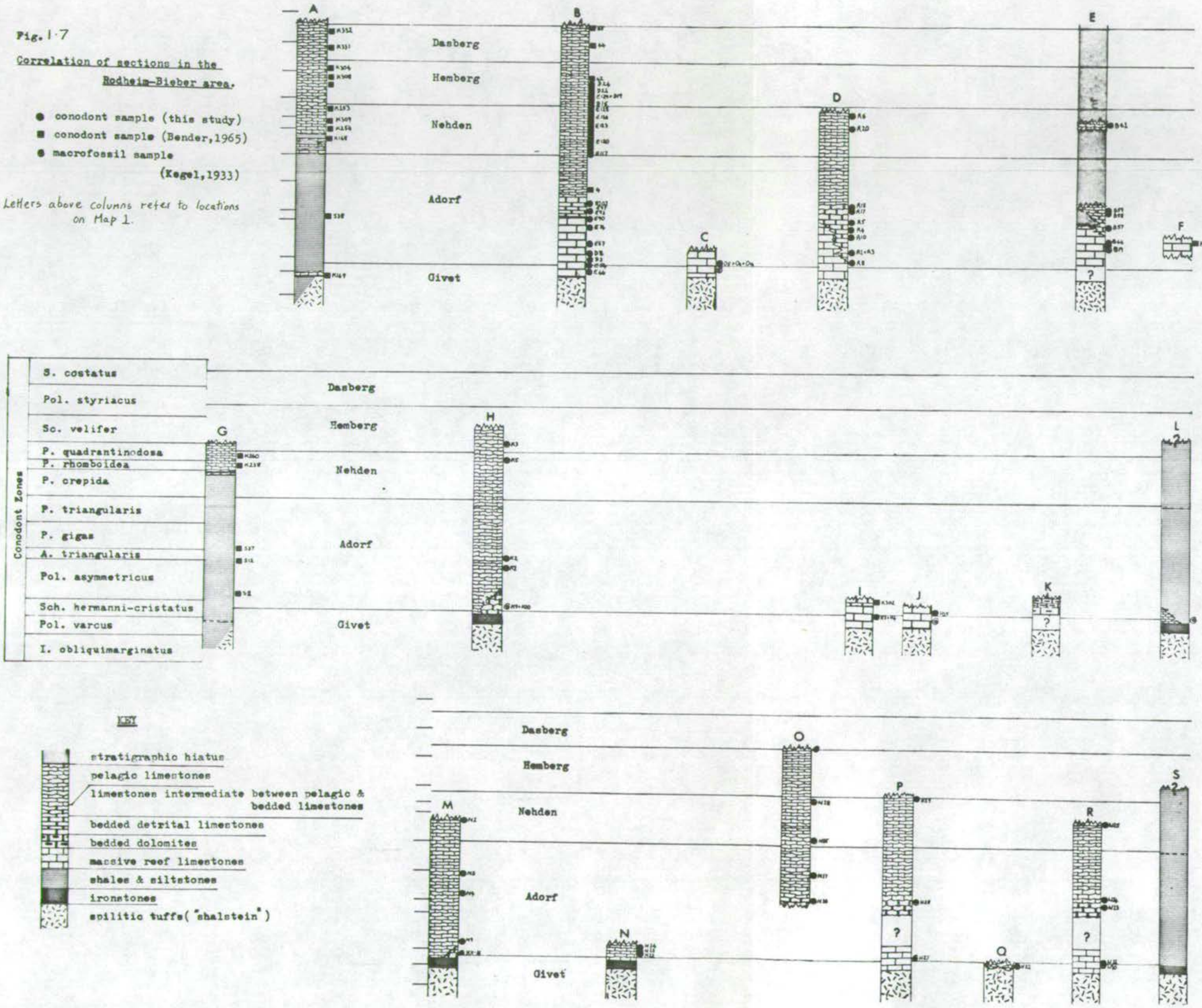
Stratigraphical results

Despite the difficulties in measuring sections in the field, the actual dating of individual specimens from a given locality enables one to place these specimens in a time-space field independant of their field relations, so that a history of sedimentation for that locality emerges. This method of sample plotting forms the basis for the stratigraphic columns in Fig 1.7 many of which have been compounded from the samples collected within a small area rather than in a continuous section. Time lines are here approximately horizontal.

Accurate total thicknesses of sediment cannot be

Fig. 1·7

Correlation of sections in the Rodheim-Bieber area.

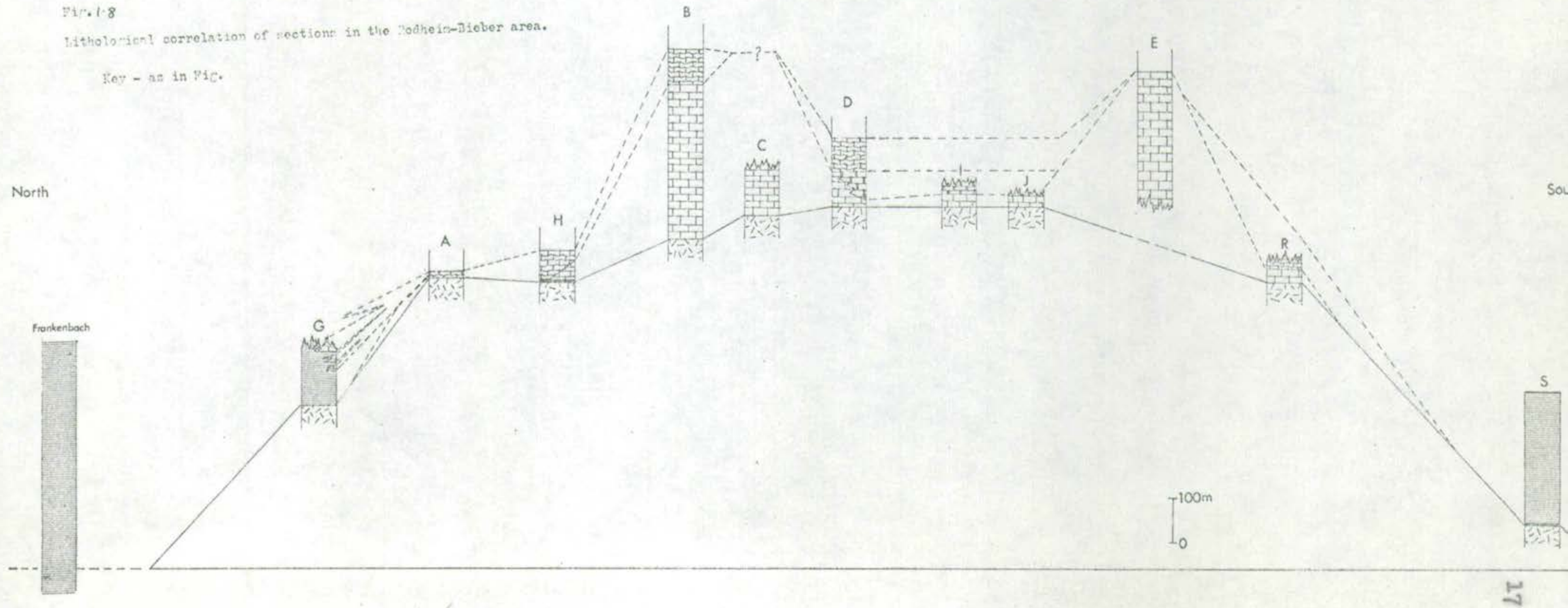


measured in most areas, but rough estimates of relative thickness can be made, both from field observation and from estimation of thicknesses of mapped units. Fig 1.8 shows the relationship between some of the sections in Fig 1.7, with the relative thickness between litholigic units in the various sections represented.

Fig. 1·8

Lithological correlation of sections in the Rodheim-Bieber area.

Key - as in Fig.



CHAPTER 2

Igneous Rocks

Igneous rocks are important constituents of the Devonian-Carboniferous succession in the Lahn-Dill area. Their prime importance, however, is not so much the contribution they make to the overall volume and thickness of the rock sequence, but the way in which igneous activity has controlled facies distribution and development throughout the Upper Devonian and Lower Carboniferous.

Petrographically, the rocks fall into three main groups:

(a) Pyroclastics - defined here as fragmental rocks derived directly or indirectly from explosive volcanic activity. Reworked tuffitic sediments are thus included within this group.

(b) Lavas and sills - mainly vesicular pillow lavas and small, petrographically similar, intrusions.

(c) Volcanochemical deposits - although not strictly "igneous", the origin of these deposits is so intimately linked with volcanic activity, that they are described and discussed here for convenience.

(a) Pyroclastics

As in the rest of the Lahn Syncline, the pyroclastic rocks in the studied area are dominated by vitric tuffs

("Schalstein" of German authors*). They are restricted to the region south of Königsberg and west of Bieber (Map 1, Appendix I), this marking the northeasterly termination of a more or less continuous volcanic ridge running the entire length of the Lahn Syncline. The thickness of the tuff sequence is difficult to assess due to the tectonic complexity of the area, but Kegel (1933) estimates it to be between 0 and 500 metres. Good exposures are found only in small quarries and recent road cuttings, but even here the rocks are chloritized and sheared, and in some cases, thrust over rocks of younger age (Fig 2.1a).

In thin section, the greater part of most of the rocks is made up of "streaked-out" and "welded" apple-green chlorite globules enclosed in a fine grained chlorite-calcite matrix, sometimes clouded with limonite (Fig 2.1b). The flattening of globules has imparted a pronounced lineation to the rocks which is interrupted in places by apparent flowage of the chloritic mass around small ovoid lumps of fine grained olive green to brown vesicular spilite.

Near the top of the Schalstein sequence, southwest of Dicke Eiche (Map 2, Appendix I), which in the Givetian

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* "Schalstein" is an old term used originally for the green and purple chloritized sheared pyroclastics of Givetian age in the Lahn-Dill area. Hentschel (1952) proposed that the term be used only in this precise lithostratigraphic sense, though others (e.g. Middleton, (1960) have extended its use to rocks of similar composition and texture, regardless of age or geographical location.

Fig 2.1

(Scale bar represents 1mm. length unless otherwise indicated)

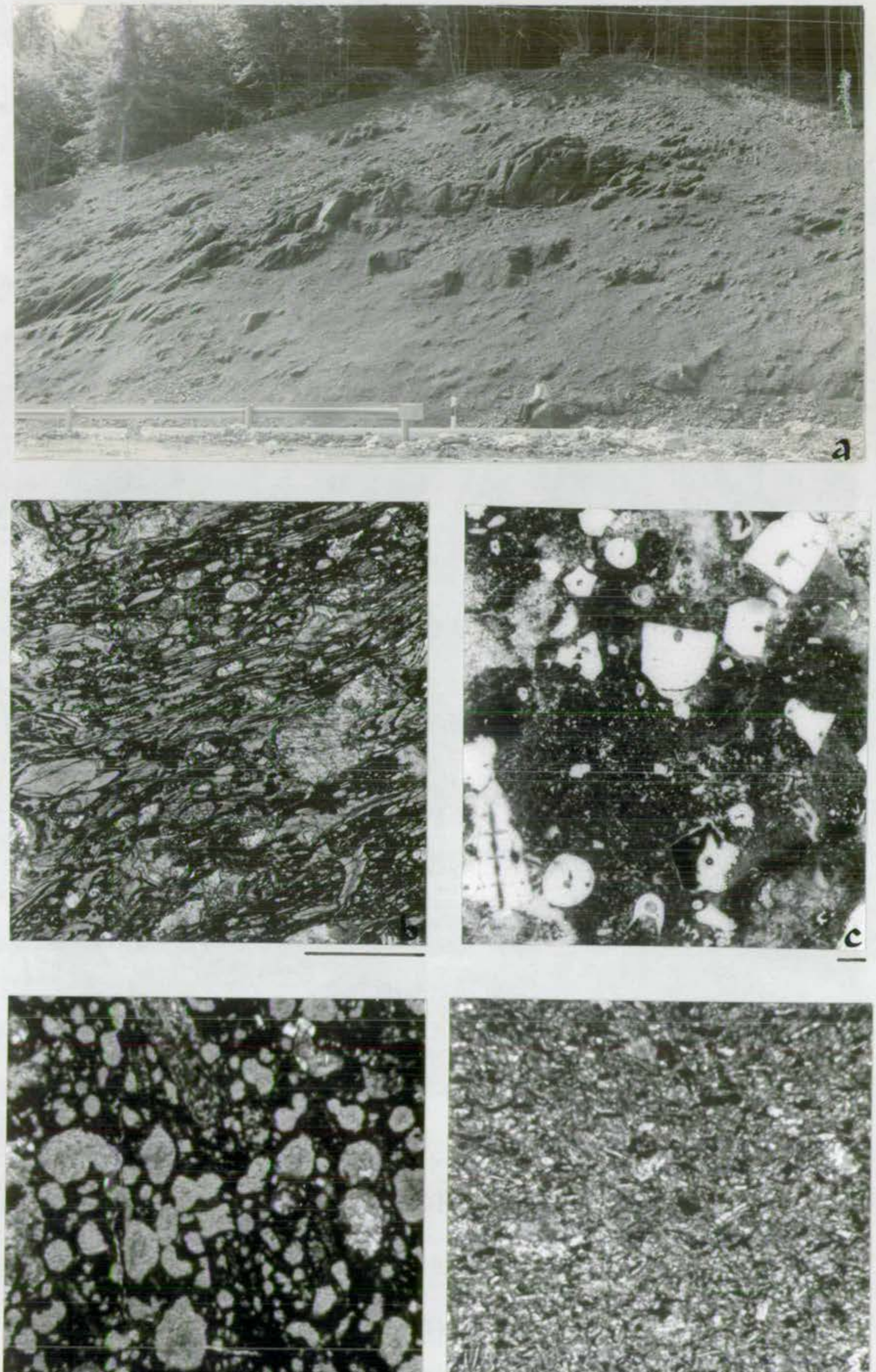
(a) Contact between strongly sheared Schalstein (above left) and massive limestone (below right) in roadside exposure immediately south of Eberstein. The pyroclastics are thrust over the younger limestones at this locality. Scale given by figure, right of centre at base of photograph.

(b) Givetian Schalstein showing typical "streaked-out" and "welded" chloritic globular texture. Thin section, plane polarized light. Sample no. E1.

(c) Crinoid ossicles and plates in shaly tuffitic matrix. Ossicle near bottom of photograph has been partially haematitized. Methylene blue stained peel, plane polarized light. Sample no. H32.

(d) Undeformed chlorite globules in tuff fragment from breccia in upper part of volcanic sequence, south of Dicke Eiche. Thin section, plane polarized light. Sample no. H31.

(e) Bedded tuffitic sediment from basinal area south of Frankenbach. The bulk of the sediment is composed of discrete chloritized shards, in contrast to the schwellen Schalstein in which interconnected globules are more common. Thin section, plane polarized light. Sample no. F8.


a
b
c
d
e

probably lay near the summit of the volcanic rise, the massive unbedded Schalstein pass upwards into bedded shaly tuffs containing blocks of volcanic and sedimentary materials (e.g. Emsian(?) sandstone and Givetian biomicrites) as well as mixed volcanic-sedimentary breccias rich in crinoid material (Fig 2.1c). The tuff fragments found in these breccias differ from the Schalstein, described above in having a less well developed foliation, the globules being unflattened here, and in the presence of extremely small chloritized feldspar lathes in an olive green, partially silicified, chlorite mesostasis (Fig 2.1d).

The lack of foliation and the preservation of the original shape of the globules may be due to differential palagonitization within the tuff sequence. Only the "clear" apple-green chlorite areas show signs of "streaking-out" and flattening, and then only when they are not enclosed in olive-green "turbid" chlorite areas, with or without feldspar microlites. It is possible that the "clear" areas were originally sideromelane (very quickly quenched basaltic glass) which, during diagenesis, was transformed to palagonite and was compacted in the process. The "turbid" areas, on the other hand, may have originally been tachylite (basaltic glass clouded with iron ore) which failed to compact since tachylite resists palagonitization (Peacock, 1926).

The foliation, flowage around fragments, and welding of globules typical of the Schalstein, therefore, may be at least partly a secondary feature, due to early palagonitization, and not produced by any primary volcanic process.

Although the bulk of obvious Givetian magmatic activity was restricted to the area southwest of Dünsberg (Fig 1.5), thin, often graded, tuffitic siltstone horizons of this age are found interbedded with black shales in the area north of Dünsberg (Bender, 1965).

Microscopically these bedded reworked (?) tuffitic sediments most differ from the Schalstein in possessing discrete angular, chloritized shards rather than globules (Fig 2.1e).

In composition and texture, these pyroclastic rocks closely resemble the Triassic "aquagene tuffs" of British Columbia (Carlisle, 1963). The globulation found in most of these rocks is, according to Carlisle, due to fast quenching of pieces of molten lava in contact with sea water, with subsequent hydration of the glass to palagonite and then to crystalline chlorite.

(b) Lavas and sills

The oldest lavas in the Devonian of the Lahn Syncline are of keratophyric and quartz keratophyric composition. These are well developed in the area south and west of Wetzlar (Fig 1.4), but are only found in the southwest part of the studied area interbedded with shales and tuffs of Givetian age (Fig 2.2a).

The majority of lavas, however, are more basic in

Fig 2.2

(Scale bar represents 1mm. length unless otherwise indicated)

(a) Givetian quartz keratophyre with strongly embayed quartz phenocrysts in a greenish-grey glassy groundmass with fluidal structure picket out by limonite staining. Thin section, crossed polarized light. Sample no. W16.

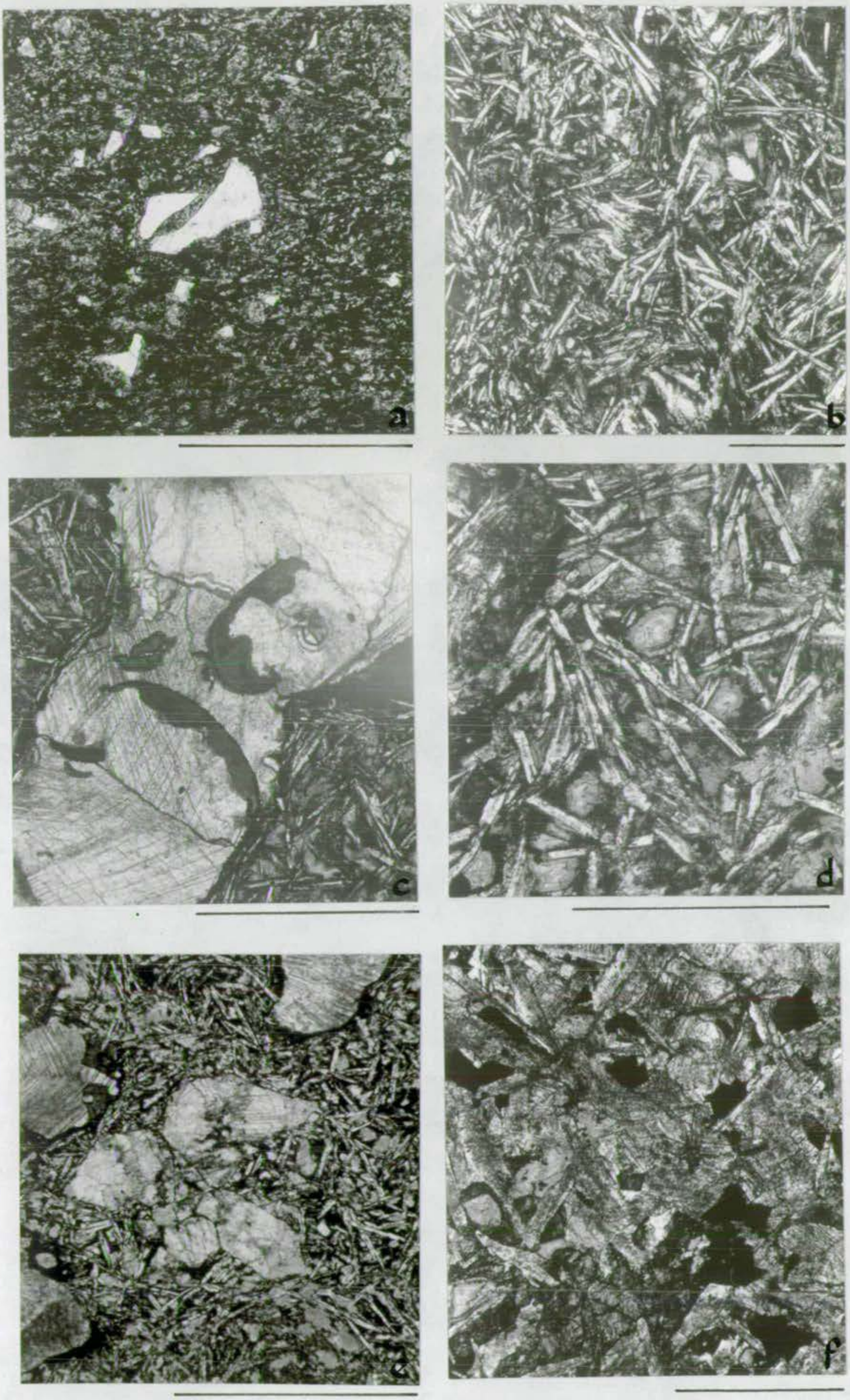
(b) Lower Carboniferous feldsparphyric spilite. Feldspar lathes either randomly orientated or arranged in fan-like clusters in a chloritic groundmass. Thin section, plane polarized light. Sample no. D29.

(c) Lower Carboniferous spilite. Detail of calcite filled vesicle showing dislodged dusty olive green chlorite rims. Thin section, plane polarized light. Sample no. D29.

(d) Lower Carboniferous spilite. Detail of groundmass showing patches of clear apple green chlorite pseudomorphing groundmass mafic minerals (? clino-pyroxene) between feldspar lathes. Thin section, plane polarized light. Sample no. D29.

(e) Lower Carboniferous spilite. Detail showing probable calcite pseudomorphs after(?) pyroxene glomerocrysts. Thin section, plane polarized light. Sample no. D30.

(f) (?) Carboniferous partially spilitized dolerite. Heavily altered euhedral plagioclase feldspars (andesine, in part) and ilmenite (+ leacoxene) subophitically enclosed within subhedral titanaugite crystals. Thin section, crossed polarized light. Sample no. K16.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f

composition. These occur as lenses, flows and fragments within the Schalstein, and as flows of Lower Carboniferous age ("Deckdiabas" of German authors), the latter being by far the best developed in the studied area.

The Deckdiabas reach their maximum thickness west and south of Königsberg (250 metres, according to Kegel, 1933), thinning rapidly to less than 50 metres west of Eberstein and dying out entirely to the south and east (Map 1, Appendix I). It appears, therefore, that the Deckdiabas were originally confined to the northern part of the rise, the volcanic centre probably lying somewhere southwest of Königsberg. (The present northerly extent of the Deckdiabas is due solely to the lavas having been thrust northwards along the Hohensolms Thrust).

Until recently, age estimates of the Deckdiabas were imprecise, the only indications of age being that metamorphosed nodular limestones of Wocklum age occur within the lavas (proving them to be post-Devonian), and that shales of Upper Pericyclus age overlie them in places (Kegel, 1933). More accurate age information has since been provided by Walliser (reported in Bender, 1965), who obtained a CuIIβ conodont fauna from a black fine-grained limestone lense within the Deckdiabas sequence south of Hohensolms (66360, 12450). This age agrees with those obtained for the Deckdiabas elsewhere in the Lahn-Dill region (Krebs, 1966; Goldman, 1968).

In the field the lavas often appear massive and vesicular, though, in several exposures, poorly developed pillows are seen.

In thin section, both the Givetian and Lower Carboniferous lavas are amygdaloidal hyalopilitic feldspar-rich spilites, the feldspars (albitic, in composition) being more or less randomly orientated in a dark "turbid" olive green chlorite groundmass (Fig 2.2b). This chlorite (possibly originally tachylite) also occurs as discontinuous, and often dislodged, vesicle rims (Fig 2.2c). Clear apple-green chlorite and calcite occur as final vesicle-fills, and also in the groundmass, apparently as pseudomorphs, the former of groundmass ferromagnesian mineral (Fig 2.2d), the latter as glomerocryst pseudomorphs (Fig 2.2e).

In places the spilites are heavily enriched in haematite, either dispersed throughout the matrix, or filling vesicles. These are discussed more fully in the next section.

Intrusive spilitized doleritic rocks (probably also of Carboniferous age) are common in the basin as well as rise areas, normally as more or less concordant sill-like bodies. In some samples, spilitization has not gone to completion and relics of the original mineralogy and texture are preserved, indicating that these rocks were originally dolerites with zoned andesine feldspars and ilmenite subophitically enclosed within titanaugite crystals (Fig 2.2f).

It is widely accepted (Kegel, 1933; Krebs, 1966, 1971; Goldman, 1968) that the spilitic pillow lavas in the Lahn-Dill region, are the products of submarine volcanism, and nothing in the association of the spilites in the Rodheim-Bieber could be found to contradict this. The actual process by which the present mineralogy was derived, however, is more controversial, some authors believing the present mineralogy to be primary (Lehmann, 1952, 1974), others that it is due to alteration of basalts, extruded into sea water, either at an early stage or some time after burial (Vallance, 1960, 1969, 1974; Cann, 1969; Juteau and Rocci, 1974). A primary origin for the Lahn-Dill spilites is clearly unacceptable, due to the partially preserved basic mineralogy of the intrusive rocks, the frequent occurrence of pseudomorphs of mafic minerals, and the fact that published analyses of the Devonian-Carboniferous spilites fall within the range of basalts (Hermann and Wedepohl, 1970; Floyd, 1972). Early spilitization caused by reaction with sea water on eruption also seems highly unlikely since spilitization in recent ocean-floor lavas increases with depth and is not advanced in those basalts actually in contact with sea water (Cann, op. cit.). The changes in composition most probably resulted from hydrothermal solutions connected possibly with intrusion of dolerites, the formation of ironstones in the Devonian, Carboniferous or Tertiary (or, perhaps, all three), or from widespread regional metamorphism during the Variscan Orogeny.

(c) Volcanochemical Deposits

In places the junction between the volcanic rocks and the overlying sediments is marked by a zone of haematite enrichment. This ore is present above both the Givetian and Carboniferous volcanics, and in places haematite impregnation of the overlying limestones in the Devonian case has extended the area of these ores. Only the primary ores are discussed here.

The ores associated with the Schalstein are by far the most important. They are found throughout the Lahn-Dill area always at, or near, the boundary between the Givetian volcanics and Adorf sediments (thus the German name "Grenzelager", literally "boundary-layer"). Ammonoids (Pharciceras sp.) have been found within the ores at many localities proving them to be mainly upper Givetian in age (Kegel, 1933). Conodont samples of Adorfian age age have also been obtained from sediments immediately above the ore layers in other areas of the Lahn-Dill Syncline.

The "Grenzelager" ores are of two distinct petrographic types: haematite-calcite ores ("Kalkig roteisenerz") and haematite-quartz ores ("Kieselig roteisenerz") (Bottke, 1965).

Calcareous ores consisting of irregular interconnecting angular flakes of haematite scattered throughout a recrystallized calcium carbonate matrix, are rare in the studied area, only one occurrence having been noted (Fig 2.3). Siliceous ores are by far the most widespread. Texturally

Fig 2.3

(Scale bar represents 1mm, unless otherwise indicated)

(a) "Grenzelager" haematite-calcite ore. Non uniformally distributed angular haematite flakes, often containing angular quartz or chert fragments, in coarse ferroan calcite cement. Thin section, plane polarized light. Sample no. K13.

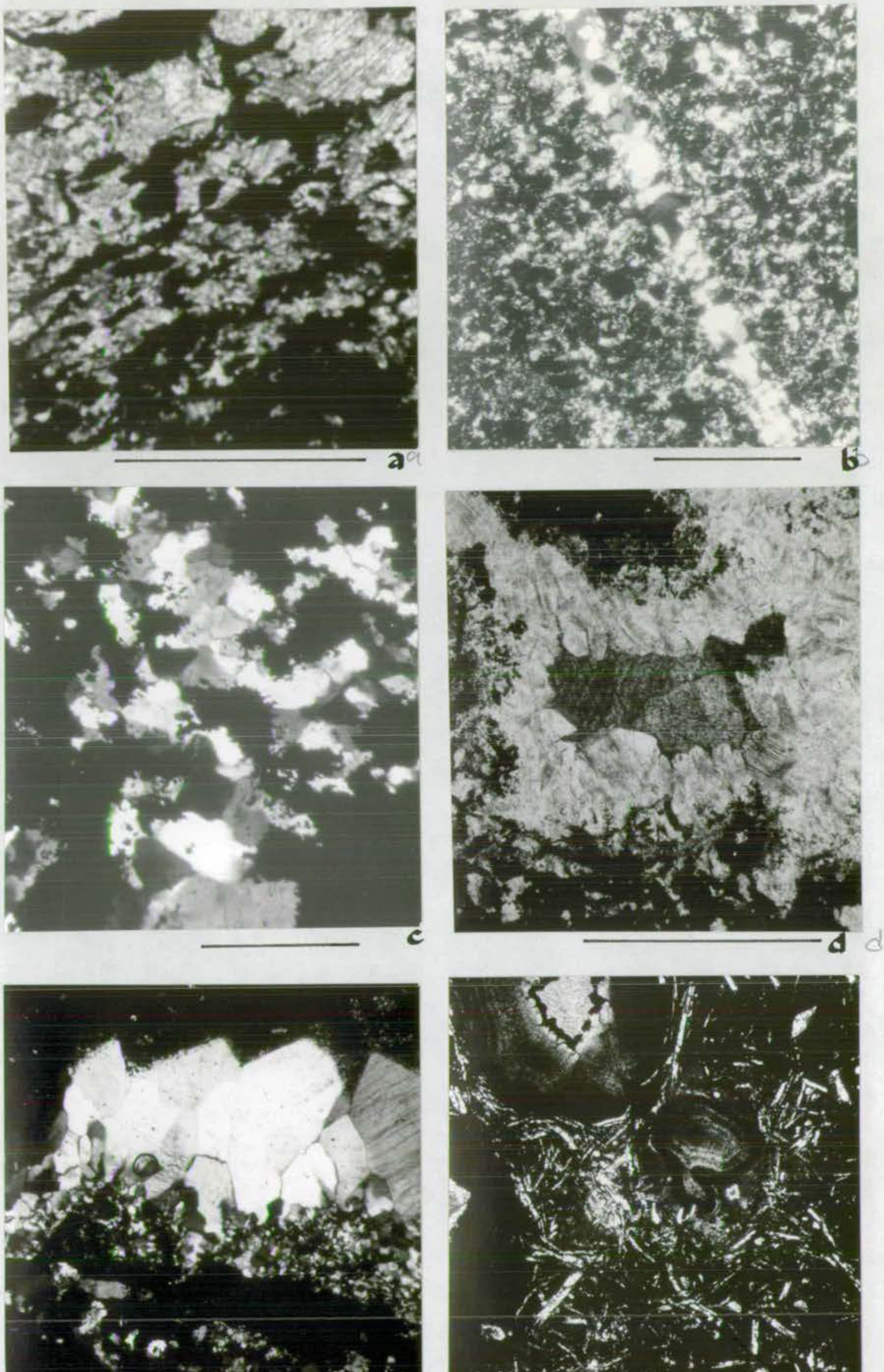
(b) "Grenzelager" haematite-quartz ore. Finely distributed "dusty" haematite in fine quartz or microcrystalline quartz matrix. Drusy quartz filled veins commonly transect ore rocks. Thin section, crossed polarized light. Sample no. W1a.

(c) "Grenzelager" haematite-quartz ore. Large inter-connected haematite flakes, non uniformally distributed in a coarse crystalline quartz matrix. Thin section, crossed polarized light. Sample no. W16.

(d) "Grenzelager" haematite-quartz ore. Detail of vug area with two generations of cavity filling: quartz crystals and ferroan calcite filling cavity in dusty haematite-chert rock. Thin section, plane polarized light. Sample no. K5.

(e) "Grenzelager" haematite-quartz ore. Detail of drusy quartz crystals with well developed crystal terminations overlain by haematite dust, possibly internal sediment deposited on top of crystals. Thin section, crossed polarized light. Sample no. K9.

(f) "Eisenkiesel" ore disseminated throughout spilitic groundmass, and as cherty-haematite vesicle fillings (left and upper left). Thin section, plane polarized light. Sample no. X7.


a
b
c
d
e
f

they vary from "spongey" dispersed haematite-chert rocks with small haematite clots and crystals fairly evenly dispersed throughout a fine cherty matrix (Fig 2.3b) to more massive haematite-quartz rocks, in which large irregularly shaped interconnecting masses of haematite and quartz mosaic form a complex interconnecting network (Fig 2.3c). In addition to these "primary" fabrics, drusy quartz-filled veins and vugs, in which specularite, chalcedony and ferroan calcite also occur, are common (Fig 2.3d). At least some of these cavities and their fillings were formed at a much later date since void filling chalcedony appears to grow from the outer edges of quartz veins.

Some specimens have been intensively brecciated, often with remobilized haematite injected along the cracks, and in places either roof collapse of fine cherty iron sediment into enhedral quartz crystals, or forceful crystallization of quartz into the matrix has taken place (Fig 2.3e).

The Lower Carboniferous "Eisenkiesel" ores are developed on a much smaller scale and are most commonly found in the volcanic rocks themselves or at the junction with the overlying siliceous shales. The ore-bodies are normally thinner (<1 metre maximum thickness) and are poorer in haematite. Frequently the iron is dispersed throughout the groundmass of the volcanic rocks themselves, or is concentrated in vesicles (Fig 2.3f).

Origin

Iron enrichment on the surface of volcanic rock sequences can be caused by surface weathering, with the formation of a lateritic soil, or by volcanic exhalations resulting in the formation of iron-rich precipitates.

Laterization can be excluded as a likely process for formation of the Lahn-Dill ores for the following reasons (Harder, 1954):

i) they are normally coarse grained crystalline quartz-haematite rocks and not soil-like in texture;

ii) they contain marine fossils;

iii) they are poor in Al, Cr, and Ti, elements commonly enriched in lateritic deposits.

An exhalative origin was suggested by Bottke (1962, 1965) in which iron and silica chlorides were released into sea water and precipitated near the vent as mixed hydrosols, passing laterally gradually into deposits richer in $CaCo_3$. A somewhat similar process, involving "extrusion" of "ore-magmas" derived from pre-eruption spilitization, has also been suggested by Lehmann(1972).

Recent drillings in the deep oceans (Bostrom and Peterson, 1966; Bonatti and Joensa, 1966; Bostrom, 1970) and in the Red Sea (Deegens and Ross, 1969) have proved the existence of iron enriched sediments above the volcanic basement. These deposits are thought to have formed through emanation and precipitation at spreading ridges and to have been transported later by sea floor spreading to their present positions. Apparently similar sediments of of Cretaceous age, attributed to hot brine exhalation, have been reported from the Troodos Massif in Cyprus (Elderfeld et al., 1972; Robertson and Hudson, 1972).

The deposits described above differ from the Lahn-Dill ores in several important respects:

i) they are essentially all iron enriched clays, whereas the Lahn ores contain, at most, only a small percentage of clay material;

ii) they are enriched in certain heavy metals (e.g. Cu, Pb, Zn, Ni etc.) missing in the Lahn ores.

Furthermore the preservation of calcareous fossils within the ores cannot be easily reconciled with the high temperature, acidic exhalative origin suggested above.

Low temperature exhalation from CO_2 springs associated with volcanism has been suggested as an alternative (Harder, 1964). In the Mediterranean area, at present, several areas are known where iron and silica are being precipitated at normal temperatures and only slightly lowered pH from such springs (e.g. the Kameni Islands, Santorini; Puchelt, 1973). The iron and silica are thought to be derived in this case from leaching of the volcanics by acidic hydrothermal solutions, a process which has been advocated for the derivation of iron and silica in the Lahn-Dill ores (Hentschel, 1960; Rösler, 1962). The leaching of iron and silica may be associated with the spilitization of the volcanics, as they are noticably depleted in iron-rich, mafic minerals. Heavy metals leached along with iron may have escaped into the atmosphere at the "shallow" depths at which the ironstones probably formed instead of being retained, as they are in deeper "oceanic" areas, within the ore deposits.

The "spongy" texture shown by the ironstones may possibly be a relic of an original colloidal iron-hydroxide/silica mixture (as is found in Santorini), which has since crystallized to haematite and quartz. It is even possible that crystallization was an early event as quartz crystals have been observed forming from iron hydroxide/silica gels within a few weeks of their formation (Harder and Flehming, 1970).

Although the ironstones are stratigraphically equivalent to the massive limestones of the schwellen areas, they are never closely related in the field. The physicochemical conditions in the vicinity of the "exhalative-springs" may have inhibited organic limestone development, though the common occurrence of radiolarian cherts above these deposits suggests that they may have been deposited below the level of active organic carbonate growth, on the schwellen slopes.

Volcanism and Tectonics

Three distinct phases of volcanism are recognized in the Rheinisches Schiefergebirge (Pilger, 1952): an Emsian "Keratophyre" phase, a Givetian "Schalstein" phase, and a Lower Carboniferous "Diabas" phase. In the Rodheim-Bieber area, the first two phases run consecutively with little or no break between phases (Kegel, 1933).

The temporal relationship between volcanic activity and major sedimentary and tectonic features is shown diagramatically in Fig 2.4. Both periods of volcanic activity coincide with major environmental changes, the first marking the beginning, the second the termination, of leptogeosynclinal conditions in the development of the Rhenish Geosyncline, in the studied area.

With the development of the plate-tectonics theory in recent years and the realization that many ancient orogenic belts mark the sites of ancient plate collisions, and are thus the only remaining evidence for the existence of ancient oceans, new interest has been aroused in the tectonic environment of volcanic activity within orogenic belts, in the hope of discovering the relics of ancient subduction zones and eventually reconstructing the relative former positions of plate margins prior to collision.

Recently, several reconstructions of Hercynian Palaeogeography have been attempted, involving lithospheric plate collisions ("Alpinotype" condition of Zwart, 1967) (Nicolas, 1972; Burret, 1972; Burne, 1973; Johnson, 1973; McKerrow and Ziegler, 1973; Floyd, 1973; Riding, 1974; Anderson, 1975).

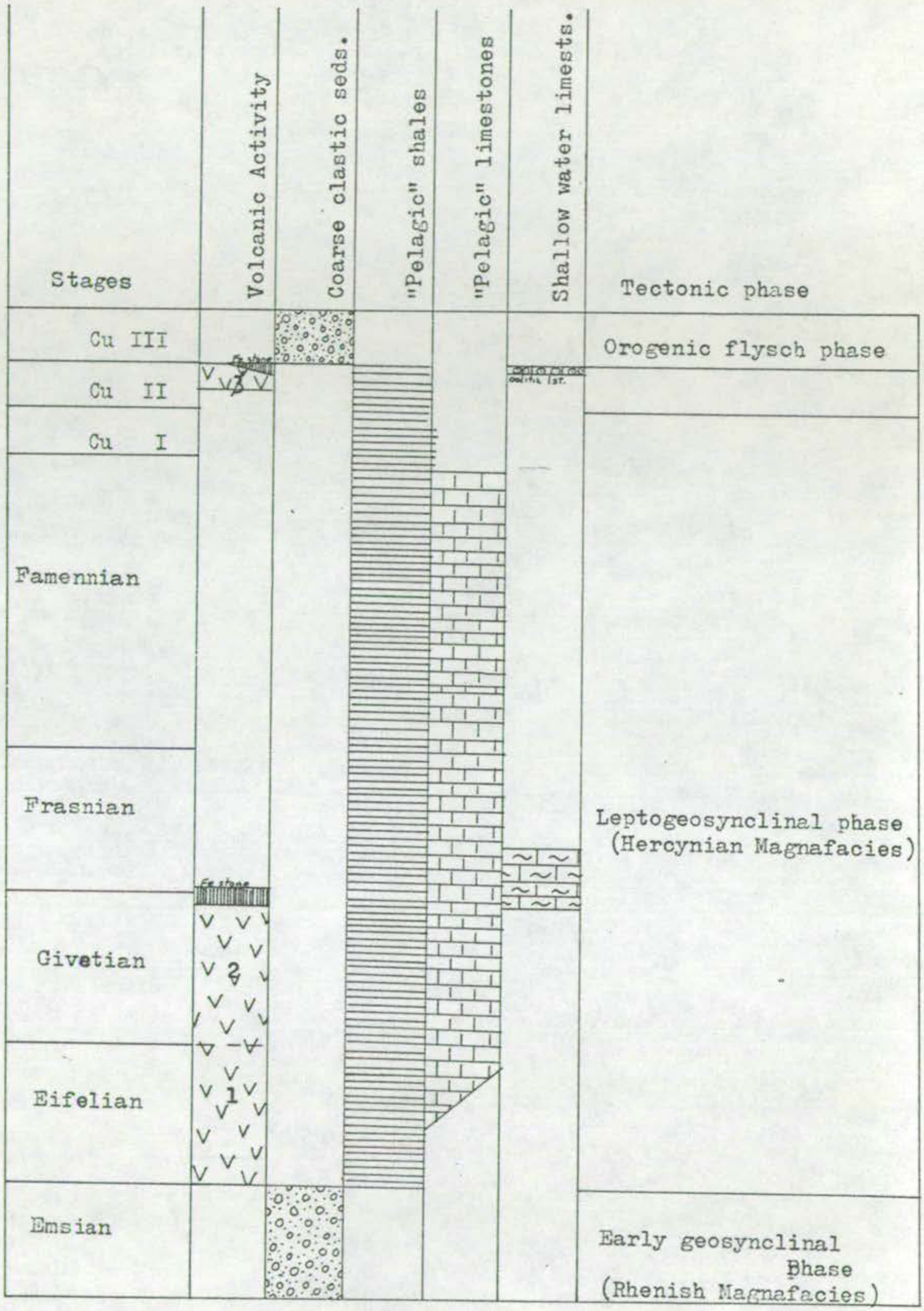


Fig . Time distribution of major lithological groups in the Rodheim-Bieber area. (Numbers refer to Pilger's stages of volcanism).

The difficulties encountered in the above models in locating the positions of oceanic areas and of subduction zones, however, emphasize the problems involved in trying to apply a "simple" plate-tectonics model to the Hercynian fold belt:

i) No evidence (e.g. ophiolites, glaucophane schists, melange deposits) of the former presence of a subduction zone has yet been found anywhere in the Rheinisches Schiefergebirge. Those reconstructions showing a subduction zone immediately to the south of the Lahn area (Burret, Burne, Johnson, op. cit.), therefore, are not based on observable fact but on the <u>assumption</u> that the Lahn-Dill volcanics are of island-arc origin.

ii) Although many of the Middle and Upper Devonian sediments show "oceanic" features, there is no evidence that the underlying crust was oceanic. On the contrary, neritic Emsian sandstones and shales underly the "oceanic" sediments over most of the studied area. Furthermore, if the volcanics, in any way, reflect the nature of the underlying crust, then it is more likely to be of "continental" type (Floyd, 1974), though some of the Lahn-Dill volcanics show characteristics intermediate between "continental" and "oceanic" basalt types (Herrmann and Wedepohl, 1970).

iii) The existence of an oceanic area immediately to the south of the Lahn area is contradicted by the presence of greywackes, in the studied area, thought to have been derived from a crystalline mixed plutonic/volcanic source located in the direction of this ocean area.

Indiscriminate application of plate-tectonics to the Hercynian fold belt, therefore, raises problems which are difficult to solve. There is, however, strong palaeogeographical and faunal evidence for the existence of some kind of "oceanic" area separating North and South Europe in the Devonian (Wittington and Hughes, 1973; McKerrow and Ziegler, 1973). If a true ocean separated these two areas at that time, then it most probably lay farther south, in the sight of the subduction zone lying within the metamorphosed Massif Central region (Riding, 1974). It is also possible, however, that such an "oceanic" area could have been formed by downfaulting of large areas of continental crust to bathyal or abyssal depths, as in the Tyrrhenian Sea at present ("Hercynotype" condition of Zwart, 1967; Krebs and Wachendorf, 1973).

Of the models so far proposed, the one which appears to agree best with most of the sedimentological, palaeogeographical, igneous and geochemical evidence, is that the Rhenish trough developed as a marginal "behind-arc" basin, bordering the Old Red Sandstone Continent to the north, and bordered by an island arc complex, in the region of the Mitteldeutsch Schwelle, to the south (Reading, 1973).

Such a model provides an attractive analogue for the following reasons:

i) Present marginal basins (e.g. the Tonga-Kermadec region) are divided into many fault-bounded rises and basins, with pelagic limestones and reefs on the ridges and terrigenous material in the basins, as in the Rhenish trough (Karig, 1970).

ii) Volcanic activity is almost entirely restricted to fault zones parallelling the basin-ridge system, and is sometimes intermediate in composition between continental and oceanic basalts (Kuno, 1967; Myashiro, 1973) as in the Rhenish trough (Herrmann and Wedepohl, 1970).

iii) In present marginal seas lying between an island arc system and the continent, terrigenous sediment is derived from both areas: mature sands from the continent, immature flysch-type sands from the island arcs. A similar situation is seen in the Rheinisches Schiefergebirge where quartzites from the O.R.S. continent are carried south into the Dill Syncline, being trapped before reaching the Lahn area, whereas greywackes and silts from the Mitteldeutsch Schwelle are carried north and deposited in the Lahn syncline. In many modern examples the island arc complex is entirely volcanic and so the resulting "flysch" is dominated by volcanoclastic material, but in some cases (e.g. Japan) is itself partially made up of continental material and so a more varied flysch composition results (Matsuda and

Uyeda, 1970). The Mitteldeutsch Schwelle probably resembled the latter since the Devonian-Carboniferous greywackes contain high-grade metamorphic and plutonic fragments which could not have been derived from a purely volcanic source.

In summary, no evidence has so far been obtained to indicate that the Rhenish Trough was a true oceanic basin, although it probably reached oceanic depths in places through down faulting producing northeast-southwest trending rises and basins. Whether the trough was part of a larger, purely intracratonic Hercynian complex or developed as a marginal sea bordering an oceanic area to the south cannot be assessed from a study of the Rheinisches Schiefergebirge alone and has not yet been finally resolved (see Riding, 1974 for discussion).

CHAPTER 3

Non Carbonate Sediments

If one excludes the volcanic contribution, sediments within the basin areas can be seen to be composed essentially of verying proportions of pelagic (siliceous) material and detrital clays, silts and occasionally sands. As both the pelagic and detrital processes were continuously active, and thus all sediments contain elements of both processes, attempts to subdivide the sediment types must, of necessity, be somewhat arbitrary. At certain times, however, and in certain areas, one or the other of these processes appears to have been dominant, producing a more or less distinctive lithological group. On this basis the sediments are discussed below under the following headings:

(a) Detrital siltstones and shales - where the influence of fine detrital sedimentation is most marked. With a decrease in detrital supply or an increase in pelagic material these pass into (b).

(b) Siliceous mudstones and cherts - in which evidence of pelagic origin in the form of recognizable siliceous microfossils can be seen.

(c) Coarse detrital deposits, mainly greywackes and breccias, although strictly detrital like (a) are sufficiently distinctive lithologically to be recognized as a separate group and therefore warrant separate discussion.

(a) Detrital siltstones and shales

These are the typical sediments of the basin areas of the Rhenish Geosyncline and are best developed in the studied area in the region north and west of Dünsberg (Fig 1.5).

In the main, the shales show poor fissility, except in the case of black shales of Givetian age. Rock colour varies from black and grey to green and red depending on the degree of oxidation of the iron and the organic content. Black and grey shales are typical of basin areas, especially in the Givetian and Frasnian, with green and red mudstones increasing in importance, the latter, especially, in schwellen areas, in the Famennian.

The shales and siltstones are normally calcium carbonate-poor (usually less than 5% by weight, see Appendix II), but limestone concretions occur locally (Fig 3.1a) and nodular limestones (and, less frequently, limestone turbidites) are interbedded with them in "near-schwellen" areas, the amount of $CaCO_3$ in the sequence as a whole increasing as schwellen areas are approached.

In those areas where the shales are well exposed and have not been intensively deformed by thrusting and development of cleavage, no sedimentary structures, other than fine horizontal lamination, were seen (Figs 3.1b and d).

Petrographically the shales and silts are monotonous, most of the material being in the clay or fine silt range, though scattered silt sized angular quartz grains occur in a few samples. These are never common enough, however, to

Fig 3.1

(Scale bar represents 1mm. unless otherwise indicated)

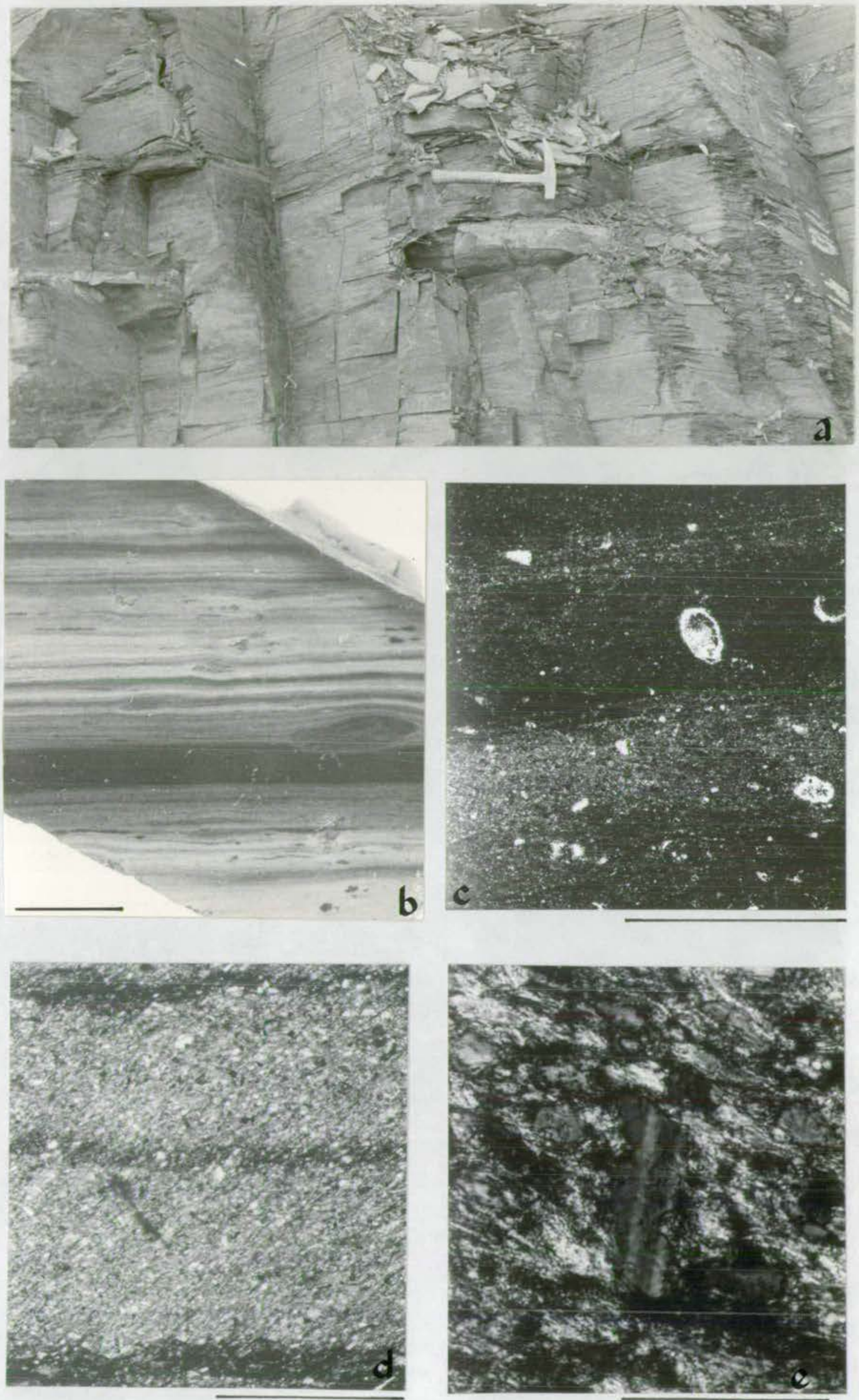
(a) Exposure of Givetian black shales south of Weipoltshausen. Laterally impersistent nodules (Centre, below hammer) commonly developed along the bedding at certain horizons. Early formation of the nodules is indicated by textural features within the rock and by the compaction of shale laminae over the nodules. Scale given by hammer.

(b) Laminated Adorf "Banderschiefer" showing fine compositional lamination, interrupted by small early-formed chert nodules. Dark bands consist of fine black mudstone, light bands of fine siliceous mudstone. Hand specimen. Scale bar = 1cm. Sample no. F11a.

(c) Banded Givetian black shale, showing indistinct lamination and decalcified moulds of cricoconarids. Thin section, plane polarized light. Sample no. Wp22.

(d) Laminated Adorf "Banderschiefer". Atypical silty bands alternate with thin bands of black mudstone. A distinct cleavage, running diagonally across photograph, has deflected laminae slightly in places and has resulted in some reorientation of grains in the silt. Thin section, plane polarized light. Sample no. F11b.

(e) (?)Detrital albite grain in siltstone. Grain edges are ragged. Bedding direction is almost diagonal in photograph. Thin section, crossed polarized light. Sample no. Wp16.



0·1mm

form distinct silt lamellae. It is likely that the importance of fine grained quartz has been underestimated in some cases due to the "obscuring effect" of mica of higher birefringence in the matrix. Rare (?) detrital albite grains occur in some samples (Fig 3.1e).

Many of the sediments were deformed during the Variscan orogeny and a distinct cleavage is developed in them (Fig 3.1d).

The fauna

One of the most distinctive features of the silts and mudstones is the sparseness of the fauna, both in actual numbers and in the range of species represented. Conodonts are rarely seen in thin sections, due to dilution by detritus. Faunal lists compiled by P.Bender (1965), however, indicate that little significant difference exists between the conodont faunas of the basin and rise areas, though these lists contain no mention of Icriodus or Belodella, both genera thought to be restricted or more abundant in shallow water, near-reef environments (Krebs, 1959; Müller, 1962; Seddon, 1970; Seddon and Sweet, 1971; Druce, 1973). (For further discussion, see page 304).

Cricoconarids are represented only by thin shelled Dacryoconarids, which are normally found as rare scattered, decalcified moulds in shales of Givetian and Frasnian age. Their absence in younger sediments is due to the extinction of the group at the end of the Frasnian. Occasionally microcoquinas, several millimetres thick, occur, as in the Givetian shale sequence south of Weipoltshausen, but neither in these nor in the shales themselves do the shells show strong preferred orientation. There is, therefore, no evidence of strong current activity.

The biological affinities of the Dacryoconarids are difficult, if not impossible, to establish with any confidence. Superficially the shells resemble those of some modern pelagic pteropods, but the absence of late Palaeozoic and early Mesozoic pteropods suggests that no phyletic link exists between the two groups. Their facies independence, widespread distribution, and thin tests, however, suggest a pelagic mode of life, and, though possibly unconnected with the pteropods phylogenetically, they may well have occupied a comparable ecological niche in the Devonian Seas (Fisher, 1962; Bouček, 1964).

Ostracods are locally common in the shales, especially those of Famennian age (thus the local lithostratigraphic term "Cypridinenschiefer" used for shales of Fammenian age) and have been used for stratigraphic purposes in some areas (Rabien, 1956). The majority of forms belong to the Entomozoidae (Entomozoe, Entomoprimitia and Richterina being the dominant genera). These are thought to have been planktonic, and are related to the modern pelagic Holocypridae. Occasionally large numbers of shells are found scattered on bedding planes, marking perhaps periods of non-deposition or current winnowing, though normally only scattered specimens are found. Articulated, unbroken specimens, often preserved only as external or internal moulds are common, though loss of $CaCO_3$ through diagenetic solution would tend to result in a preservational bias towards articulated specimens.

Ammonoids are only rarely found in the shales, and only as flattened external moulds, but are more common within shaly limestone nodules within the shales. Their rarity in the shales is, therefore, thought to be due mainly to difficulties in preservation and not to any fundamental ecological control.

Bivalves have occasionally been found in the shales. All belong to the presumed pelagic *Posidonia* or *Buchiola* groups (Bender, 1965).

Trilobites are common only in criconarid microcoquinas, south of Weipoltshausen where thoracic segments, pygidia and cephalons (some with free-cheeks connected to the cranidia) of Phacopids and Proetids, occur in pockets. The occurrence of whole unbroken (though crushed) head-shields indicates that transportation by strong currents is unlikely, and that bottom conditions, on the schwellen slopes at least, were not prohibitive to bottom life.

Corals are extremely rare in these sediments. Only one specimen (? *Syringaxion* sp.) was found buried with the trilobites described above.

Trace fossils have also been reported (Bender, op. cit.). All belong to the *Posichnia* group, typical of the Nereites, deep water, facies (Seilacher, 1953). No burrowing or resting traces have yet been found. The preservation of surface trails on bedding surfaces once again indicates quiet water conditions.

In summary, several generalizations can be made concerning the fauna of the shales and siltstones:

(i) the fauna is sparse;

(ii) all fossils present are exclusively marine forms;

(iii) most fossils are planktonic or nektonic; benthonic forms are rare;

(iv) the faunal diversity and abundance appears to be largely controlled by diagenetic solution of $CaCO_3$ and SiO_2 skeletons, as can be seen by comparing the faunas of the shales with those of limestone nodules within them. The consequences of the dissolution of skeletons in these sediments to estimation of the depth at which they were deposited in the Rhenish trough is discussed more fully below (page 72).

Origin of the clay and silt

The faunal and sedimentological features of the shales suggest a largely pelagic source for much of the clay, with the clay and fine quartz silt being transported through the air from a distant landmass (the O.R.S. continent to the north perhaps). Such a process is known to account for much of the fine quartz silt seen in the present oceans (Rex and Goldberg, 1958). If this were the only, or indeed the main, source of silt and clay, however, one would expect basinal sediments to be much more condensed stratigraphically than their submarine-rise counterparts which would receive just as high a "rain" of clay, but in addition would also receive carbonate sediments. This is the case in the present Central Pacific Ocean where most of the sediment is pelagic in origin and the basinal red or brown mudstones are much reduced in thickness compared to the coccolith-foram oozes on the submarine highs above the Carbonate Compensation Depth. The basinal clays are, in effect, insoluble residues, the calcium carbonate having been completely removed by solution. In the Bieber area, however, the reverse is true: the schwellen limestones are the condensed deposits; the basinal shales being very much thicker in comparison. The pelagic contribution to the clays is therefore of only secondary importance, most of the sediment being swept into the basins by bottom currents as happens around Rodrigues Seamount (Palmer, 1964) and in the Tyrrhenian Sea today (Sartori, 1974).

Two sources for such sediment are possible:

(1) local, i.e. from slumping on the slopes of the submarine rises.

(2) distant, i.e. bottom transportation of silt and clay from a far-off shelf area.

The absence of slump structures and the widespread fine horizontal lamination seen within the shales and silts, suggest that local derivation of material was minimal and that the material was transported and deposited mainly from suspension and not by bottom currents. These deposits may possibly represent distal turbidites (coarse greywackes of middle Adorf age occur in the eastern part of the area, and Givetian greywackes have been found 30km to the south). Alternatively they may have been deposited from suspended sediment-rich layers of water in the water-mass above the basin, similar to those described from the Atlantic Ocean off the Eastern coast of North America (Eittreim et al 1969). These "nepheloid layers" are thought to be responsible for the deposition of significant amounts of hemipelagic sediments at the base of the continental slope and on the basin plains of modern oceans, and may have been equally important in the past (Stanley and Unrug, 1972). At present "nepheloid" deposits are difficult to distinguish from distal turbidites, possibly because the two depositional processes merge at "base of slope".

The sediment could have been derived from the north, as is much of the sediment in the Dill Syncline, but the presence of the Hörre-Acker trough between the Lahn and Dill Synclines would surely have acted as a trap for sediment derived from that direction. Derivation, from a southerly direction, therefore, appears more probable.

(b) <u>Siliceous mudstones and cherts</u>

No clear cut division exists between these rocks and the detrital mudstones and siltstones, into which they grade with increasing clay mineral and silt content, and decrease in the abundance of siliceous fossils. Rocks discussed under the present heading are characterized by fine grain size and the presence of recognizable siliceous fossils.

The siliceous sediments are best developed in the Adorf sequence south and south-west of the Rodheim-Bieber rise, where they have been thrust north-west onto the Schwelle, but occur also in the Adorf sequence west of Dünsberg interbedded with dark grey and light grey shales ("Banderschiefer"), and in the Hemberg/Dasberg sequence at Frankenbach (Fig 1.5). Siliceous sediments also occur in abundance in the Lower Carboniferous throughout the area ("Kieselschiefer").

The siliceous sediments are always conspicuously well bedded, though bedding thickness is often irregular, varying from a few centimetres in the cherts to less than one millimetre in siliceous shales (Fig 3.2a). No sedimentary structures, other than planar lamination, were observed.

No conodonts were obtained from the sequence south-west of Bieber, but as the siliceous sediments directly overlie the "Grenzelager" ores (which contain zonal ammonoids of upper Givetian and lower Adorf age) and, in places, nodular limestones of lower Adorf age (Kegel, 1933),

Fig 3.2

(Scale bar represents 1mm. length unless otherwise indicated)

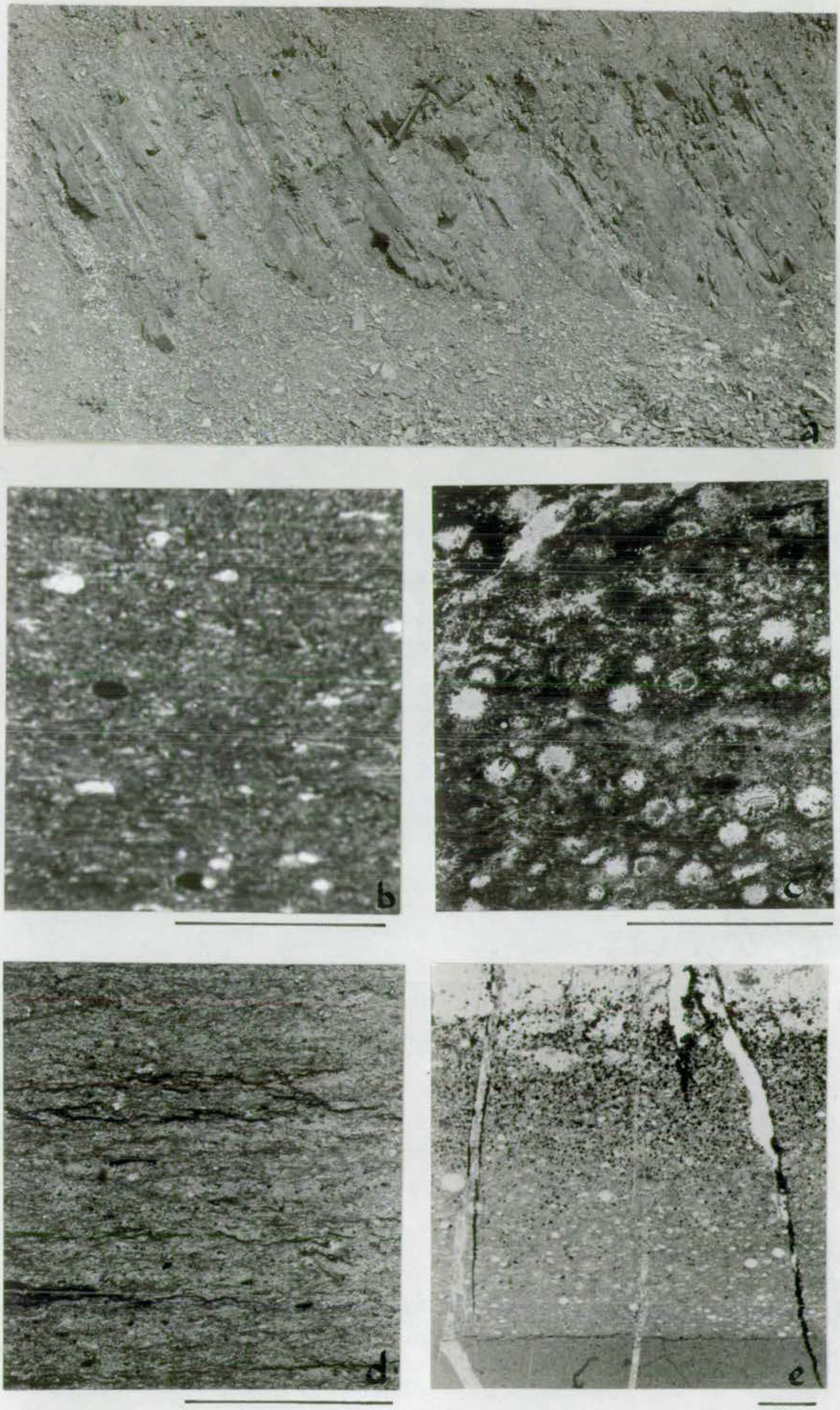
(a) Exposure of Adorf siliceous shales and cherts south east of Rotestrauch. The distinct bedding shown is due to repeated alternation of bands of massive chert and fissile siliceous mudstone. Scale given by hammer.

(b) Siliceous mudstone containing scattered sponge spicules (centre left) and squashed radiolarians (centre right). Thin section, plane polarized light. Sample no. W8.

(c) Radiolarian chert crowded with casts of radiolarians. Matrix is blackstained microcrystalline quartz. In places details of radiolarian wall structure are preserved, but most radiolarians are preserved as chalcedony filled external moulds. Thin section, plane polarized light. Sample no. W18.

(d) Laminated poorly fossiliferous chert. Rare sponge spicules lie parallel to the bedding, but no larger fossils are present. Thin section, plane polarized light. Sample no. W12.

(e) Interlamination of siliceous sediment types. Unfossiliferous chert abruptly overlain by siliceous mudstone with poorly preserved radiolarian ghosts, passing up into recrystallized richly fossiliferous radiolarian chert. Rock is traversed by stringer veins filled with drusy quartz, (?) dolomite and limonite. Thin section, plane polarized light. Sample no. F2.


a
b
c
d
e

the age of the base of the sequence, at least, is known. Ages of other siliceous sediments were obtained by conodont dating of interbedded shales (Bender, 1965).

Like the shales and siltstones described above, the siliceous sediments show a wide range of colours depending on the percentage of organic material and iron contained in them and on the oxidation state of the iron (Grunau, 1965). Grey and black predominate, though as in the shales, red sediments become locally important in the Nehden Stage.

Petrographically the rocks range from mudstones with siliceous fossils (Fig 3.2b) to pure cherts (Fig 3.2c) depending partly on the composition of the matrix. In general, the higher the proportion of siliceous micro-organisms, the "purer" the chert, but this is not always so, as some of the purest cherts contain very few recognizable fossils (Fig 3.2d).

The matrix of the cherts consists of a fine grained mosaic of micro-crystalline quartz and chalcedony, with fine grained almost opaque clots of limonite, pyrite, organic material and clay minerals, intimately mixed in varying proportions, the relative proportions of the various components determining the colour and "muddiness" of the sediment.

<u>Fauna</u>

The most conspicuous difference between the siliceous mudstones and their detrital counterparts can be seen in the marked change in faunal constitution from the dominantly calcareous fauna of the latter to the almost exclusively siliceous fauna of the former, conodonts alone being common to both sediment groups.

Radiolarians are present in many of the sediments but their abundance varies greatly: in many of the purer cherts they are so numerous that individuals are more or less "grain" supported, while in the muddier sediments, they are often more sparsely scattered throughout the matrix. In some cherts radiolarians are almost absent, but this may be due to diagenetic alteration of the chert rather than a primary ecological "exclusion".

Regardless of age and lithology, the type of radiolarians found appear to be the same: spheroidal spumellarians. Specific identification of the individuals is hampered by the poor preservation and the limitations of examination in section, but some, at least, appear to belong to the Heliosphaerinae (perhaps Acanthosphaera and Astrospharina).

Generally in the purer cherts, the shape of the radiolarians tends to be well preserved, but details of test ornament lost. In the mudstones, on the other hand, test details are generally better preserved, but the radiolarians themselves tend to be more flattened by compaction. The range of preservation types encountered is illustrated in Fig 3.3a-h.

Spicules are the only commonly encountered fossils apart from radiolarians. They tend to be more abundant and better preserved in the muddier sediments, partly because of the lack of dilution by radiolarian tests, but also because, like the radiolarians, they probably tend to be much less well preserved and, because of their small size, more difficult to "pick out" in thin section in the purer siliceous sediments.

Some radiolarian spines may have been included with sponge spicules in some samples, but the shape and presence of a distinct central canal in most of the spicules proves them to be sponge spicules (Fig 3.3i-1).

Conodonts are extremely rare, as in the detriral shales and silts, due probably in part to high sedimentation rate, but also to difficulty of extraction. Scattered elements were seen in only one sample, and none were obtained from microfossil samples.

Summary of faunal characteristics

None of the above groups are autochthonous faunal elements: the radiolarians are pelagic, and must have travelled at least vertically downwards to the sea floor, the conodonts are generally considered to be nektonic or planktonic and, therefore, must have again undergone vertical transport and possibly significant lateral transport, if the conodont animal were buoyant after death, before being incorporated into the sediment. The sponge spicules represent the only benthonic component of the fauna, but as no large sponge-like masses were found, and the spicules are always found parallel to the bedding, it is likely that these too have been transported from elsewhere. The presence of small numbers of sponge spicules in practically all recent pelagic sediments was noted by Riedel (1959) who attributed their presence to the fact that they are light and hollow and thus can drift for long distances before being deposited. No calcareous fossils are present, and as they are common both on the rises and in the basins, their absence in the cherts and siliceous mudstones can only be explained by a preservation bias towards non-calcareous faunal components.

Mode of Formation

(1) The source of SiO_2

The origin of silica in siliceous sediments has received a great deal of study in recent years (Grunau, 1965; Krauskopf, 1959; Siever, 1962; Ernst and Calvert, 1966; Pimm et al, 1971; Calvert, 1974; von Rad and Rösch, 1974; Wise and Weaver, 1974). Only two sources, are currently considered important:

(a) that SiO_2 is precipitated inorganically from sea water abnormally saturated with silica;

(b) that SiO_2 is derived mainly from solution of the siliceous tests of planktonic microorganisms.

At present, no substantial amounts of inorganically precipitated silica are known to occur in marine environments (Calvert, 1974). Normal sea water is undersaturated with respect to both amorphous silica and quartz (Krauskopf, 1959) and so inorganic precipitation of silica would be impossible under normal conditions. Local silica saturation of sea water could, however, possibly occur if silica was released in large enough quantities by submarine volcanic activity (Stefansson, 1966). The strong association between cherts and submarine volcanics in the geological record, and the lack of any recognizable biogenous siliceous remains within bedded cherts, have led some authors to favour an inorganic origin (Grunau, 1965; Bonatti, 1965; Heath and Moberly, 1971; Scheidegger, 1973).

In the Rodheim-Bieber a strong connection does exist between siliceous sediments and volcanic activity, the main chert developments immediately following periods of submarine volcanic activity. Siliceous sediments are, however, not restricted to these periods. Some of the siliceous sediments are apparently unfossiliferous or poorly fossiliferous, but neither this nor the connection with volcanic activity is inconsistent with a purely biogenous source of silica (von Rad and Rösch, 1974). Furthermore, the majority of sediments do contain abundant siliceous microfossils. A predominantly biogenous origin is, therefore, suggested for the siliceous sediments described above, although inorganic "volcanogenic" silica is believed to have precipitated locally, immediately after volcanic events, together with iron oxides to form the cherty "Grenzelager" and "Isenkiesel" ores dealt with in Chapter 2.

(2) Conditions of deposition

Assuming a biogenous silica source, the contribution made by silica-secreting organisms to the sediment depends on their population density in the overlying water column, the amount of dilution by non skeletal detrital material, and to a lesser extent on the rate of solution of siliceous tests in the water column or on the sea floor.

Plankton productivity is ultimately governed by the supply of nutrients, of which dissolved silica, in the form of weak silicic acid, is one of the most critical in controlling the growth and development of diatoms (and presumably radiolarians) (Lewin, 1962). At present, rich radiolarian deposits are virtually restricted to regions where the near-surface waters are abnormally enriched in nutrients due to upwelling. Locally, submarine volcanic activity, by releasing SiO_2 along with other nutrients into sea water, can drastically increase radiolarian productivity, resulting in local population bursts or plankton blooms (Gibson and Towe, 1974). The development or siliceous sediments above submarine volcanics is thus more likely due to the establishment of optimum conditions for plankton development than to inorganic precipitation of SiO_2.

Even with optimum conditions for productivity, however, recent investigations of siliceous ooze distribution in the modern oceans suggests that relatively pure siliceous deposits will only be produced if the pelagic "rain" or organisms can be deposited free of dilution from other sediments with higher rates of sedimentation. In the present case dilution could come from two sources:

(i) the basin detrital silts and sands;

(ii) pelagic carbonates.

The absence of any significant terrigenous admixture in many of the siliceous sediments suggests that they were deposited at a faster rate than the detrital mudstones (perhaps during pauses in supply of detrital material, allowing sufficient time for thin purely pelagic layers to develop),or that they were deposited out of reach of the detrital sediments (perhaps on schwellen slopes which could have provided sufficient relief for pelagic sediments to accumulate relatively free of sediment transported by bottom currents). The location of cherts in the studied area precludes the possibility of their having accumulated in "deep ocean" regions far out of reach of bottom carried terrigenous detritus.

Even with elimination of terrigenous material, however, in the oceans today high productivity of calcareous plankton relative to siliceous plankton results in a swamping of the siliceous material by calcareous, resulting in the formation of carbonate oozes. Siliceous oozes only become important when the carbonate material is totally eliminated by solution below the Carbonate Compansation Depth (Peterson, 1966; Berger, 1968). Evidence of apparently pelagic carbonate sedimentation in the area contemporaneous with deposition of siliceous sediments is provided by the presence of fine grained nodular limestones on the schwellen summits (see Chapter 4). The complete absence of primary carbonates in the siliceous sediments suggests that the formation of rich siliceous deposits and the exclusion of carbonate from such deposits are intimately related.

The location of rocks of this type within eugeosynclinal terraines and their resemblance to modern siliceous oozes have led many authors (Steinmann, 1925; Garrison and Fischer, 1968) to regard sediments of this type as direct analogues of present-day deep-sea oozes deposited below the Carbonate Compensation Depth. This would suggest that water depths in the Rhenish geosyncline reached abyssal depths. Concentrations of siliceous microfossils need not in themselves indicate abyssal conditions, however, (Folk, 1973) nor is it possible to estimate with certainty the Compensation Depth in Devonian times (Hudson, 1967; Garrison, 1974). It is possible that CO_2 emission associated with volcanism could account for the absence of carbonate, as this would raise the pH of the sea water and thus the solution rate of $CaCO_3$ (in effect, raising the Compensation Depth) (Granau, 1965).

As sea water is undersaturated with respect to both opal and quartz, deposition of siliceous tests on the sea bottom will not ensure their preservation, and tests will tend to dissolve in sea water, the effectiveness of this solution being controlled to a large extent by water depth and sedimentation rate.(Riedel and Funnel, 1964).

Solution appears to be greatest in shallow water (less than 250m) decreasing regularly to a depth of about 2,000m in modern oceans, below which solution rates are constant and low (Berger, 1968). The optimum depth for silica preservation, therefore, would appear to be below 2,000m, all other factors excluded. Solution, however, is effective only so long as skeletons are in contact with sea water. Increased sedimentation would result in faster burial of skeletons and compaction would expel much of the pore water, thus reducing the rate of solution. Rich siliceous deposits could persist at relatively shallow depths, therefore, if sedimentation rates were high enough to counteract the solution effects.

(3) Diagenesis

The lack of observed compaction features in the radiolarian cherts and the fine structural detail preserved in some radiolarians suggests that the cherty matrix was completely lithified before much solution or crushing of radiolarian tests could take place. The silica cement was derived most probably by solution of opaline skeletal material in interstitial undersaturated pore waters with subsequent precipitation when saturation was reached. In the radiolarian rich sediments, however, the large volume of cement required could not have been derived from indiscriminate local skeletal solution since the majority of skeletons appear to have dissolved after lithification. It appears likely, therefore, that the bulk of the cement in these rocks must have been derived from some external source, probably from solution of silica in sediments more affected by compaction and solution. These richly fossiliferous rocks often grade into poorly fossiliferous or unfossiliferous, often finely laminated cherts (Fig 3.2e). In these rocks, perhaps, the rate of sedimentation failed to keep pace with the rate of solution of skeletons, resulting in an almost featureless "amorphous" siliceous sediment. Increased detrital sedimentation with consequent lowering of solution rate might also explain why microfossils are generally better preserved in these rock types, and why they show conspicuous signs of compaction (e.g. flattening of radiolarians) not seen in the "purer" radiolarian cherts.

The present microcrystalline quartz fabric probably developed through gradual crystallization ("maturation") of the original opaline sediments (Ernst and Calvert, 1969; Calvert, 1974; von Rad and Rösch, 1974).

Many of the sediments are traversed by thin veins filled with chalcedony, quartz, (?)dolomite and limonite (Fig 3.2c). Such veining is common in subsurface cherts in the modern oceans (Pimm et al., 1971) and appear here to be caused by compaction pressure. Whether the veining in the present case is "early diagenetic" as in the Pacific today, or occurred during deformation in the mid-Carboniferous is, however, uncertain.

Fig 3.3

(Scale bar represents 0.1mm, unless otherwise indicated)

All figures - thin sections (plane polarized light- a, f, g, h: others crossed polarized light).

(a) Poorly preserved chert filled radiolarian ghost. Sample no. N11.

(b) External mould of radiolarian filled by chalcedony fans. Sample no. W18.

(c) Squashed radiolarian filled with red siliceous mud. Wall preserved as aggregates of radially orientated quartz crystals. Sample no. W8.

(d) Black siliceous mud filling radiolarian test. Spinose nature of wall well preserved. Sample no. W18.

(e) Detail of above. Spines and associated spine bases consist of single quartz crystals, laterally linked to form skeletal wall. Sample no. W18.

(f) Detail showing honeycomb nature of radiolarian test. Sample no. X3.

(g) Radiolarian showing possible internal sphere. Sample no. O22.

(h) Pillar structure in radiolarian test, preserved by sediment infill. Sample no. X3.

(i) Triaxion sponge spicule. Sample no. W8.

(j) Circular cross section of sponge spicule, showing central canal. Sample no. X13.

(k) Needle like spicules. Sample no. O7.

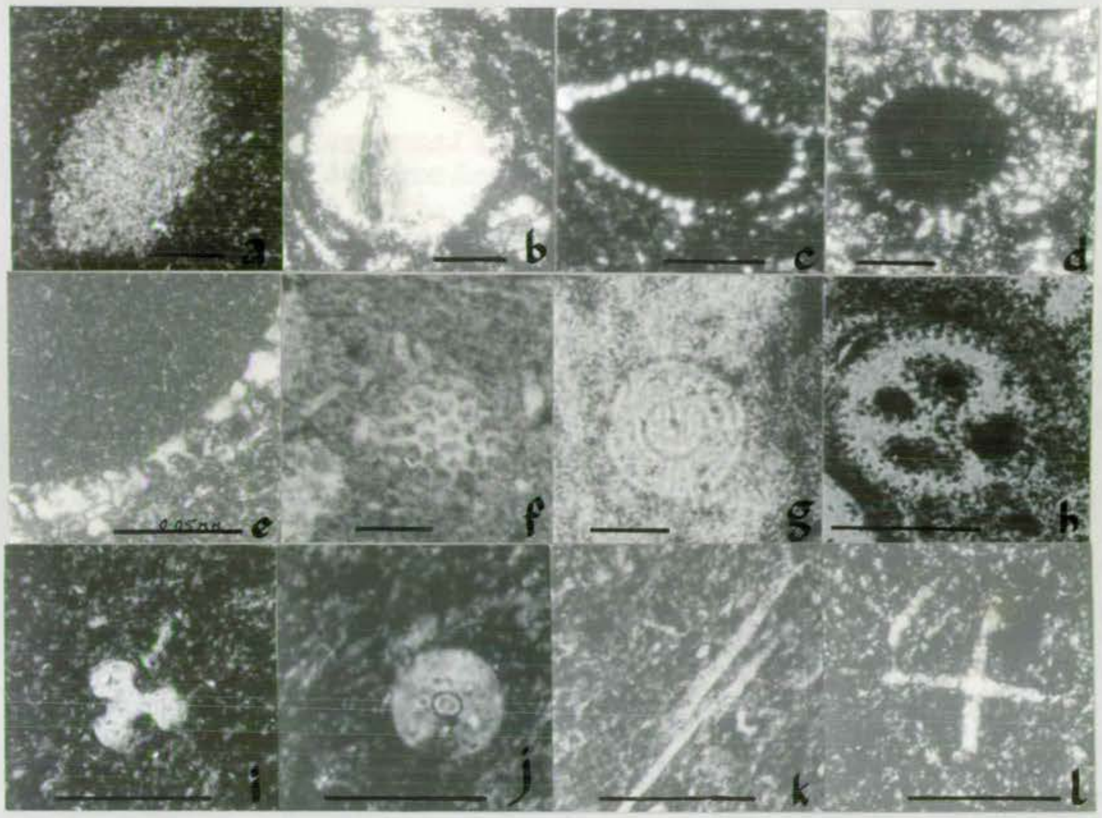
(l) Spicule cross. Sample no. O7.

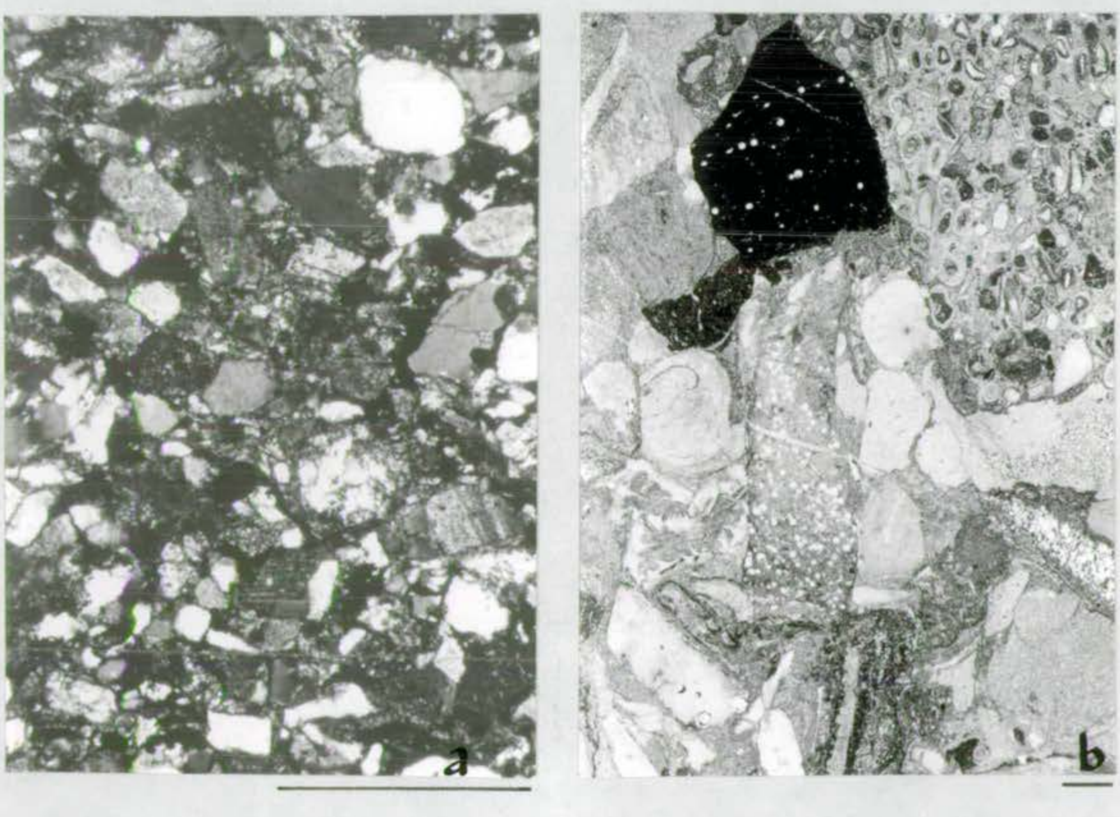
Fig 3.4

(Scale bar represents 1mm, unless otherwise indicated)

(a) Poorly sorted and rounded grains of quartz, feldspar, chert and igneous and metamorphic rock fragments in Adorf greywacke. Thin section, crossed polarized light. Sample no. H1.

(b) Angular chert, siliceous shale and oosparite fragments and fossils in silty matrix. Lower Carboniferous breccia. Thin section, plane polarized light. Sample no. X3.





(c) Arenaceous and Rudaceous detrital deposits

Coarse clastics of arenaceous or rudaceous grade are rare in the Middle and Upper Devonian sequences in the Rodheim-Bieber area. The only known occurrences (all definitely of Middle Adorf age, Henningson, 1966) lie in inliers within the main mass of Kulm Greywackes ("Giessen Greywackes" of Henningson 1966) east of Bieber, between Sälzboden and Wissmar. Here thin beds of dark grey to brown sandstones are interbedded with the "normal" siliceous mudstones and siltstones of the basin facies. Graded bedding is poorly developed and directional sedimentary structures, such as loadcasts, toolmarks etc. were not found, because of restrictions of exposure and the fact that bedding planes were never exposed. Little could be deduced about the lateral persistance of these beds but thickness and evenness of bedding never varied over the lengths of exposure examined.

Petrographically the rocks resemble the Carboniferous greywackes, both in general texture and in composition (Fig 3.4a). The mineralogy consists in the main of fragments of quartz (with a significant proportion of "metamorphic-platy" polycrystalline grains), plagioclase (essentially all albitic in composition), microcline, perthitic feldspars and various rock types (mainly metamorphic-gneissose, schistose and phyllitic). Rock types found locally in abundance (such as limestone, ironstone, schalstein and shale) tend to be absent. The Devonian greywackes appear to be more feldspathic than those at the Kulm, but the number of samples studied was low and comparison between different mineral estimates in greywackes are notoriously difficult (Welsh, 1976).

Transport Direction

No directional sedimentary structures were found in the greywackes, but several factors indicate that the material was derived from the East or South East.

(i) Coarse greywackes of Adorf age are restricted to the East and South East of the Bieber Schwelle; contemporaneous sediments further to the North West are fine grained.

(ii) The presence of a deeper water trough to the North West in the Hörre Acker Zone would act as a trap for any sediment derived from the direction of the O.R.S. continent in the North.

(iii) The immature nature of the Adorf greywackes contrast with the more mature nature of the recycled, presorted turbidites derived from the shelf areas bordering the continent to the North (Einsele, 1963).

(iv) The compositional and textural similarities between the Adorf and Kulm greywackes, strongly suggest a similar source area (or areas) for both groups, Grain orientation studies of the Kulm greywackes around Giessen (Henningson, 1963) indicated transport from the South east and east (Henningson, 1963).*

Source Area

The almost complete absence of rock fragments typical of the Rhenish Geosynclinal trough (cherts, shales, nodular and massive limestones, ironstones and spilites), together with the common occurrence of plutonic and gneissose fragments led Henningson (1963) to conclude that the Kulm greywackes were not locally derived but came from a largely crystalline source area or series of source areas somewhere to the Southeast of Giessen (the Mitteldeutsch Schwelle, Brinckmann, 1948).

No such "Devonian-Carboniferous" crystalline area is at present exposed at the surface, but several deep borings made during the last 15 years have located plutonic igneous and high grade metamorphic rocks beneath Permian sediments, about 80km East southeast of Giessen (Henningson 1966, 1970). Because of the lithological similarities between the Adorf and Kulm greywackes, the

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* The use of grain orientation as a reliable index of transport direction is only valid if the beds have been deposited mainly by grain flow as direct deposition from suspension would result in near random grain distributions. Recent work (Middleton 1967), however, suggests that the basal parts at least of turbidite sequences are deposited by bed flow, and, as Henningson's measurements were made in the coarser basal members of the greywackes sequences and show a high degree of orientation, they probably are a fairly reliable index of transport direction.

former are also thought to be derived from the Mitteldeutsch Schwelle.

Sandstones of Famennian age are an important constituent of the sedimentary sequence in the Hörre-Acker Zone and neighbouring regions. The resemblance of many of these sandstones to the well sorted sandstones of the Dill Syncline and the absence of coarse sediments of this age in the area south of the Hörre-Acker Zone suggests derivation, at least in part, from the North with the possibility of "along trough" transport from the North east, and local derivation from the trough slopes, especially for the impure limestone turbidites ("Allodapischer Kalk", Meischner, 1964) which occur locally especially in the uppermost Upper Devonian in this area.

Breccias in the basin sequences

No terrigenous rudites appear to be present in the Devonian sediment sequence in the studied area, but in the Lower Carboniferous, two breccia occurrences are worthy of mention as they provide evidence of tectonic activity at the time of their formation, and contain fragments of rock types no longer seen in the "normal" sedimentary sequences. They are, therefore, the only remaining evidence of some environments once present in the areas from which the breccias were derived.

One such case is a breccia 1.5m thick interbedded with black siliceous shales, north of Königsberg (67780, 12240). No conodonts were obtained from the breccia (Sample no. X3); but a rich macrofauna of $Cu III \alpha$ age was obtained from the matrix by Parkinson (1903) and Somer (1909). Most of the cherts are platy fragments of chert or siliceous shale, but more rounded fragments of biomicrite and oosparite, as well as isolated broken brachiopod and gastropod shells, bryozoans and crinoids, often only weakly disarticulated, occur in a black shaly matrix. (Fig 3.4b).

The compositional and textural disorder of this sediment, with its haphazard mixture of shallow water benthos and rock types mixed with deep water radiolarites and shales, together with its location on the Northern slope of the Bieber Schwelle and its post-volcanic, pre-flysch age indicate that transport was initiated suddenly on a slope steep enough to allow mixing of sediments, many already lithified, of widely varying bathymetric facies.

Breccias of this type are by no means uncommon in siliceous mudstone sequences (Garrison and Fischer 1966; Schlager and Schlager 1973; Cox and Prath 1973). The sediment described above shows a remarkable resemblance to that described by Cox and Prath (op. cit.) who attributed the admixture of chert, shale and oolitic limestones to submarine sliding initiated by reverse faulting on the slope of a submarine volcanic rise. Bearing in mind that the direction of tectonic transport in the Rheinishes Schiefergebirge is towards the northwest, and that the occurrence of this breccia coincides with the date of orogenic movements in the area, one is tempted to conclude that the Königsberg Breccia, also, was caused by reverse faulting on the Schwelle.

In the area south west of Weipoltshausen, a breccia, mainly composed of Upper Devonian nodular limestone fragments in a shaly crinoidal matrix of CuIIβ age was discovered by P.Bender (1965). Although not investigated in the present study, the stratigraphic and geographic position (close to the Weipoltshausen Schwelle) suggests that this two may have formed by submarine slumping on a fault-prone slope.

Summary

For much of the Middle and Upper Devonian, sedimentation within the basinal areas around the Bieber Schwelle consisted of deposition of detrital mudstones and silts from the slow moving turbidity or bottom "nephroid" currents. During pauses in sedimentation and in sites not easily reached by these sediments, however, the pelagic contribution to sedimentation became more important, with siliceous oozes being deposited on Schwellen slopes, especially immediately after periods of volcanic activity when unusually favourable conditions for plankton productivity prevailed. For a short time during the Middle Adorf, the supply of coarse material from the Mitteldeutsch Schwelle intensified leading to the brief introduction of coarse sediment in the eastern part of the basin, foreshadowing the more intensive deluge of greywackes of the Lower Carboniferous Kulm series which were eventually to fill the basins. Apart from these local incursions from the South, sedimentation of coarse clastic material appears to have been short lived and restricted to local tectonically unstable situations.

Estimation of the absolute water depth in which any of the basin sediments were deposited is difficult, but was almost certainly in excess of 300m. Several lines of evidence suggest this figure as an upper depth limit:

(1) Above this depth SiO_2 is extremely soluble, so the abundance of siliceous sediment suggests deeper water.

(2) Modern siliceous sponges are found today at depths between 200 and 300m.; above this depth calcareous sponges predominate.(A similar relationship seems to hold true for Palaeozoic sponge faunas (Finks 1960)). The frequency of occurrence of sponge spicules in many of the basin sediments indicate that they accumulated at or below the zone of sponge growth, the spicules being swept downslope after death of the sponge.

(3) Neritic faunas typical of shelf conditions are missing from the sediments. The fauna is, rather, dominated by pelagic forms with only a sparse vagile benthos (mainly blind trilobites). Blindness in arthropods is generally linked to life in relatively deep dark waters (though not necessarily abyssal (Clarkson, 1967).

A lower depth limit is almost impossible to estimate. Much of the faunal and lithological evidence is suggestive of abyssal depths (Rabien, 1956; Goldring, 1962). The rarity of $CaCO_3$, especially, in basin sediments is difficult to explain except by solution of $CaCO_3$, either at the sediment-water interface or within the sediment, in the environments in which these sediments were deposited. It therefore appears likely that the bulk of non carbonate sediments accumulated below the Carbonate Compensation Depth (C.C.D.) in that part of the Rhenish trough. The actual depth below which solution became effective in removal of $CaCO_3$ cannot, however, be determined. At present the C.C.D. in the oceans varies from one area to another over a depth range of over 2 kilometres (Berger and Winterer, 1974), and there are indications that even in the Mesozoic, solution of $CaCO_3$ became effective at shallower depths than at present, allowing pelagic non-carbonates to accumulate, undiluted by calcareous material, in environments (e.g. continental shelves and oceanic spreading ridges) where pelagic carbonates are at present abundant (Garrison, 1974; Robertson and Hudson, 1974).

CHAPTER 4

Nodular limestones

For much of the Devonian and Lower Carboniferous in the Rheinisches Schiefergebirge, carbonate deposition gave rise to limestones of two distinct types:

(i) Thick massive limestones, produced directly or indirectly by the build-up of benthonic organisms. These are largely restricted to the "shelf" area bordering the O.R.S. continent to the north, though scattered occurrences are found over volcanic submarine elevations in the trough region of the Variscan geosyncline where waters were apparently shallow enough for prolific benthonic carbonate development. These limestones form the subject of Chapter 5.

(ii) Thin nodular limestones, in which the benthonic contribution is small or completely lacking, the bulk of the sediment being apparently of pelagic origin (Rabien, 1956). Like the massive limestones, they are restricted, in the trough area, to submarine highs since it was only on such submarine highs that relatively pure limestones could develop free of dilution from terrigenous detritus (Fig 4.1).

On the lower slopes of these submarine rises, where terrigenous dilution is high, no truly primary carbonate rocks are found. Limestones in these basin sediments

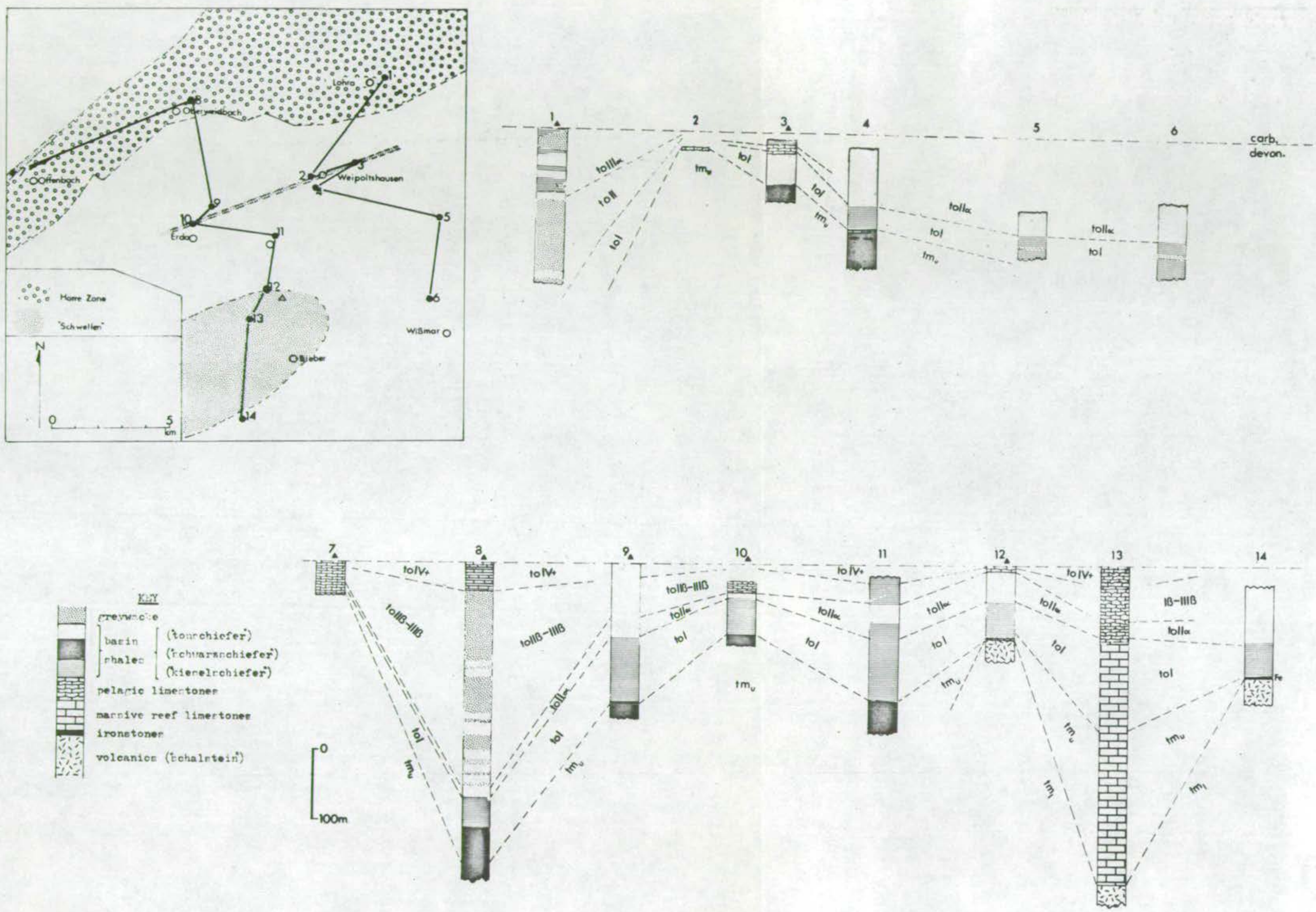

Fig. 4·1 Lithological variation in the North east Lahn Syncline
(▲ = column taken from Bender, 1969)

tend rather to occur as isolated impure elongate nodules or nodule bands along bedding planes in shales (Fig 3.1a). As the overall calcium carbonate content of the sequence increases further up the schwellen slope, the frequency of limestone nodules within the shales increases, and the limestones become purer (Fig 4.2a), until on the schwellen summits where terrigenous detritus is almost completely lacking, the nodules coalesce leaving only thin shale seams ("flasers" of German authors) between them (Fig 4.2b).

When discussing the nodular limestones it is important to distinguish between those nodules which were formed diagenetically within the lower slope sediments (and are essentially "calcified shales" from a sedimentalogical point of view), and those sediments which, although they may have undergone some $CaCO_3$ remobilization and enrichment during lithification, were nevertheless originally deposited as calcium carbonate rich sediments, relatively free of detrital non-carbonate material.

The main features of these two limestone types are listed below:

	Basin limestones (Fig 4.2c)	Schwellen limestones (Fig 4.2d)
Nodule shape and size	Large, elongate- up to 1 metre or more in length and greater than 10 cm. thick	Small, "isometric"- rarely more than a few cm. in size.

	Basin limestones Fig 4.2c)	Schwellen limestones (Fig 4.2d)
Colour	Always black, like surrounding shales	Grey on schwellen slopes; red and buff-yellow also found on schwellen tops. *
Insoluble residues (see Appendix II)	35-60% (by weight) Mainly clay material evenly disseminated throughout.	10-15% on schwellen tops; up to 30% on slopes. Mainly clay material (+ rarely, quartz silt), largely restricted to clay seams or bands between nodules.
Carbonate matrix	Micrite	Microsparite (rarely pseudosparite).
Sedimentary structures	Fine lamination marked by fossil alignment. (Continues into surrounding shales.	Mottled texture and bioturbation structures. Skeletal material haphazardly distributed.
Fauna	Rich - dominated by pelagic and nektonic animals.	

In the studied area nodular limestones are restricted to the area between Königsberg, Bieber, and Waldgirmes in the south (Fig 4.1; 3) and to a thin linear belt between

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* Only in the case of the Kellwasser Limestones, have black schwellen limestones been noted. These are bituminous, pyritic biomicrites, rich in nektonic and planktonic fossils and certain heavy metals. They are thought to have been deposited in local reducing conditions caused by either (a) an overabundance of vegetable matter (floating seaweed) in the environment of deposition, as in the Sargasso Sea today (Schmidt, 1925), or (b) temporary restriction of water circulation in the geosynclinal trough (Rabien, 1956); or, (c) periodic volcanic gaseous emissions along north north west - south south east trending geofractures, in this case the "Unna-Giessen Fraktur" (Krebs, 1969).

Fig 4.2

(Scale bar represents 1cm. in a+b and 1mm. in c and d)

(a) Nodular limestone from schwellen slope environment. The mottled irregular appearance is typical of the nodular limestones. The junction between nodular bands and shaly horizons is always marked by a sharp decrease in nodule size giving the rock a brecciated look. The nodular bands are broken up by high angled solution and movement planes. Sample no. X12.

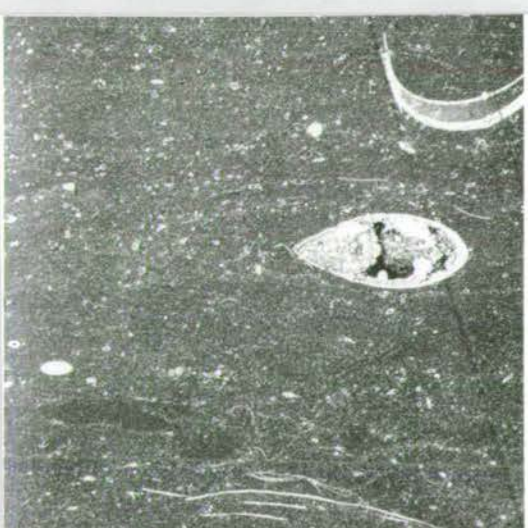
(b) Nodular limestone from schwellen summit. The limestone is again strongly mottled but clay material is virtually restricted to irregular clay seams in the rock. Sample no. E119.

(c) Nodule in basin shales. Fossiliferous calcareous mudstone matrix containing ammonoids (centre and upper left), bivalves (lower centre) and scattered cricoconarids (in matrix - light dots). Note the tendency for strong horizontal alignment and the crushing of the bivalve shells through compaction. Methylene blue stained peel, plane polarized light. Sample no. Wp15.

(d) Nodular schwellen limestone. In contrast to (c), typical schwellen limestones are intensively bioturbated with shell fragments, lacking any preferred orientation. The matrix is microsparite. These sediments are frequently transected by solution stringers ("Flasers" of German authors), (right of centre). Thin section, plane polarized light. Sample no. D21.


a


b


c

Erda, Weipoltshausen and Altenvers, in the north (Fig 4.1; 2,3,10).

These two areas mark the positions of former schwellen areas which, in the Upper Devonian were flanked by basin areas in which limestones are never developed (Fig 4.1; 1,8,9; 4,11; 5,6,14.).

Origin of the carbonate matrix

In the shaly nodules in the basin shale facies, most of the calcium carbonate is either present as fossil fragments or as micrite-sized cement between terrigenous silt and clay-sized grains, with minor amounts of sparry calcite filling skeletal voids. The nature of these nodules suggests that all of the calcite cement has migrated into the environment during diagenesis from solution of skeletal fragments elsewhere in the shale sequence.

The matrix of the majority of the limestones dealt with in this section, however, could have been formed by any of the following processes:

(i) by deposition of an original silt-sized matrix between skeletal grains;

(ii) by degrading neomorphism of an original sparite cement between skeletal grains;

(iii) by aggrading neomorphism of an original micrite matrix between skeletal grains.

Several objections can be raised against a primary calcisiltite matrix. Modern carbonates of silt grade are relatively uncommon, the only known "primary" carbonate silt being the "vadose silt" of Dunham (1969) which could not have developed in the Devonian schwellen limestones as evidence suggests that they were never subaerially exposed in the Devonian. Alternatively, a silt could have developed from comminution of skeletal fragments, as is the case on areas of the Campeche Bank today (Logan 1969). Larger microspar grains are posibly recrystallized skeletal fragments (Tucker 1973), but it is difficult to imagine where such a finely broken down matrix could have been derived from, especially in the case of the post-Adorf limestones when few if any shallow high energy areas existed in the studied areas to supply the detritus. In situ breakdown is largely ruled out by the fine preservation of thin shells within the limestones. Furthermore, breakdown of most skeletons tends to result in micrite-sized grains (Bathurst 1969).

The present nature of the microspar within the limestones is clearly secondary. Grain size distributions are patchy, grain contacts are curved and wavy, and in several cases microspar crystals "nibble" into earlier features (see page 118). It is doubtful whether such a matrix could have developed by degrading neomorphism of a sparite, however, as this type of replacement is rare and normally very localized (Bathurst, 1969). It is improbable that such a process could produce the microspar matrix, usually found in all schwellen limestones.

The most likely explanation for the microspar, therefore, is that it resulted from aggrading neomorphism of an originally finer grained (micritic?) matrix (Folk, 1965).

Origin of micrite

The origin of calcium carbonate mud has been dealt with in detail elsewhere (see Bathurst, 1969), and only a few points relevant to the present limestones will be dealt with hers.

Of the various methods by which lime-mud can be produced, only three can be seriously considered as possible sources for the matrix of the schwellen limestones.

(i) Mechanical or biological breakdown of shells.

(ii) Disintegration of calcareous algae.

(iii) Inorganic precipitation of $CaCO_3$.

(i) Mechanical or biological breakdown of shells.

The abundance of thin, delicate, apparently unabraided skeletons showing no signs of algal degradation, would seem to indicate that insitu comminution was not an important process in the formation of matrix in the studied limestones. Nor could such a matrix have been derived from local shallow wave-washed areas, since no such possible source areas are known for most of the time the schwellen limestones were being deposited.

(ii) Disintegration of calcareous algae.

Benthonic algae, for the reasons given above, can also be excluded as a likely source for lime mud. Such mud could only be derived from the disintegration of planktonic algal skeletons, as in Recent pelagic foram-coccolith oozes. No plated planktonic algal skeletons have been found in Devonian limestones, though a few problematical forms have been described from the

Carboniferous (Gartner and Gentile, 1972). Even if present, however, no trace of such skeletons would be expected since they would not have survived neomorphism. Recent altered coccolith-foram oozes from the Pacific Ocean already show features similar to those of Devonian nodular limestones:

> "The anhedral calcite crystals lie mostly in the 2-20 μ range and typically have very irregular shapes.....the boundaries between the coccoliths platelets often are indistinct suggesting the onset of recrystallization in which the multicrystal coccoliths are reformed into a few crystals or a single crystal" (Pimm et al., 1971).

Plated planktonic algae are generally thought to have developed since the Triassic (Black et al., 1967), but planktonic algae must have been important constituents of the pelagic ecosystem throughout the Phanerozoic, since without these "primary producers", pelagic animal life could not have been sustained. In view of the strong lithological resemblance between the Devonian nodular limestones and those of the Alpine Jurassic (Hallam, 1967 ; Garrison, 1967; Jenkyns, 1970; Garrison and Fischer, 1969), which do contain coccoliths, it is possible that the micritic matrix could have been derived from the plates of similar, though not necessarily closely related, pelagic calcareous algae.

(iii) Inorganic precipitation of $CaCO_3$.

Until recently inorganically precipitated $CaCO_3$ was considered only to occur in very shallow-water environments, where supersaturation of sea water with $CaCO_3$ could take place. Recent investigations in deep marine environments, however, have revealed the presence of calcium carbonate of micrite grade in carbonate sediments on the sea floor, which is definitely of inorganic origin:

> "..... the finding of inorganically precipitated lutites in such environments may offer a possible clue to the mode of deep-sea carbonate sedimentation prior to the evolution of coccolithophoroids and planktonic foraminifera in the early Mesozoic" (Milliman and Müller, 1973).

At the present time, however, such cases of inorganic micrite precipitation in open marine environments, appear to be restricted to areas in which the salinity and water temperature are unusually high, due to submarine exhalation of hot saline brines (Gewirtz and Friedman, 1966; Milliman, Ross and Ku, 1969). In addition to $CaCO_3$, such brines are often responsible for the precipitation of heavy metals within the sediment (Deegens and Ross, 1969). Heavy metal concentrates occur in the Givetian basin sediments in parts of the Rhenish trough (Krebs, 1971), but are absent in basin areas in the studied region. Exhalation has also been suggested as an explanation for heavy metal enrichment in the Kellwasser limestones (Krebs, 1969). There is, in addition a long history of volcanism in the area. Such periods were episodic, however, and it seems unlikely that thermal activity could account for the matrix of <u>all</u> of the nodular limestones, irrespective of age.

Palagonitization of volcanics, by increasing the pH and the Ca^{++} concentration of sea water, can result in precipitation of $CaCO_3$ (Garrison, Hein and Anderson, 1974). Such a process could, at best, only provide $CaCO_3$ over a restricted period (the duration of palagonitization), and area (locally, above the volcanic pile), It is thus felt inadequate as a source for the matrix of the widespread limestones.

The fauna

The fossil types encountered in both the shaly nodular limestones and schwellen limestones are alike. They are therefore discussed together in this section. Both limestone types are extremely fossiliferous, due in large part to their condensed nature and to the increased possibilities of fossilization offered by the limestones, though the range of fossil-types encountered commonly in them is restricted.

Conodonts are common in insoluble residues of the nodular limestones though their frequency of occurrence relative to other components is difficult to assess as they are only rarely seen in thin sections. Their contribution is, therefore, not as great as one might imagine from faunal numbers obtained per unit volume of rock, and, in sheer numbers, both cricoconarids and ostracods, for instance, often heavily outnumber conodont elements. The high concentrations of conodonts in schwellen limestones, relative to other sediments would seem to be due mainly to the condensed nature of the limestones, the relative insolubility of conodont elements, and their high density which could be an important factor in areas where fine calcareous mud was periodically winnowed by currents.

The world-wide reputation of conodonts as"facies-breakers", and their common association with pelagic fossils, especially in areas where indigenous benthos are absent, has led to the conclusion that conodonts formed part of some nektonic or planktonic animal (Müller, 1962), though the exact nature of this animal is still in doubt (Seddon and Sweet, 1969; Melton and Scott, 1973; Lindström, 1973).

Cricoconarids, with the exception of the genus Styliolina, which appears in rocks of late Adorf age, are restricted to rocks of Givetian and early Adorf age, due to the almost complete extinction of the group in the middle Adorf. Styliolina is by far the most common genus present, though Nowakia is almost as widespread but not so common. Tentaculitids are virtually absent and have only been seen in the Kellwasser limestones (toI(β)γ) at Altenvers (Fig 1.5).

Cricoconarid tests seldom show any preferred orientation in the limestones, a feature which cannot easily be reconciled with passive pelagic settling, which would tend to produce a pronounced horizontal shell orientation. Bioturbation is thought to be the most likely cause for the haphazard orientation of the shells.

In places cricoconarid concentrations occur in which the shells are overgrown by radiaxial bladed fibrous calcite (Fig 4.3A). Such overgrowths are considered to be replacements of an original acicular cement (Kendall and Tucker, 1973).

The presence of cement rims on these shells, which are always found in areas where the sequence is very condensed (e.g. Weipoltshausen and the north slopes of Dünsberg) indicates that these shells were not mud

Fig 4.3

(Scale bar represents 1mm. unless otherwise indicated)

(a) Cricoconarid-rich schwellen limestone. Styliolinid shells with syntaxial fibrous calcite overgrowths, randomly orientated in shaly micrite. Methylene blue stained peel, plane polarized light. Sample no. F20.

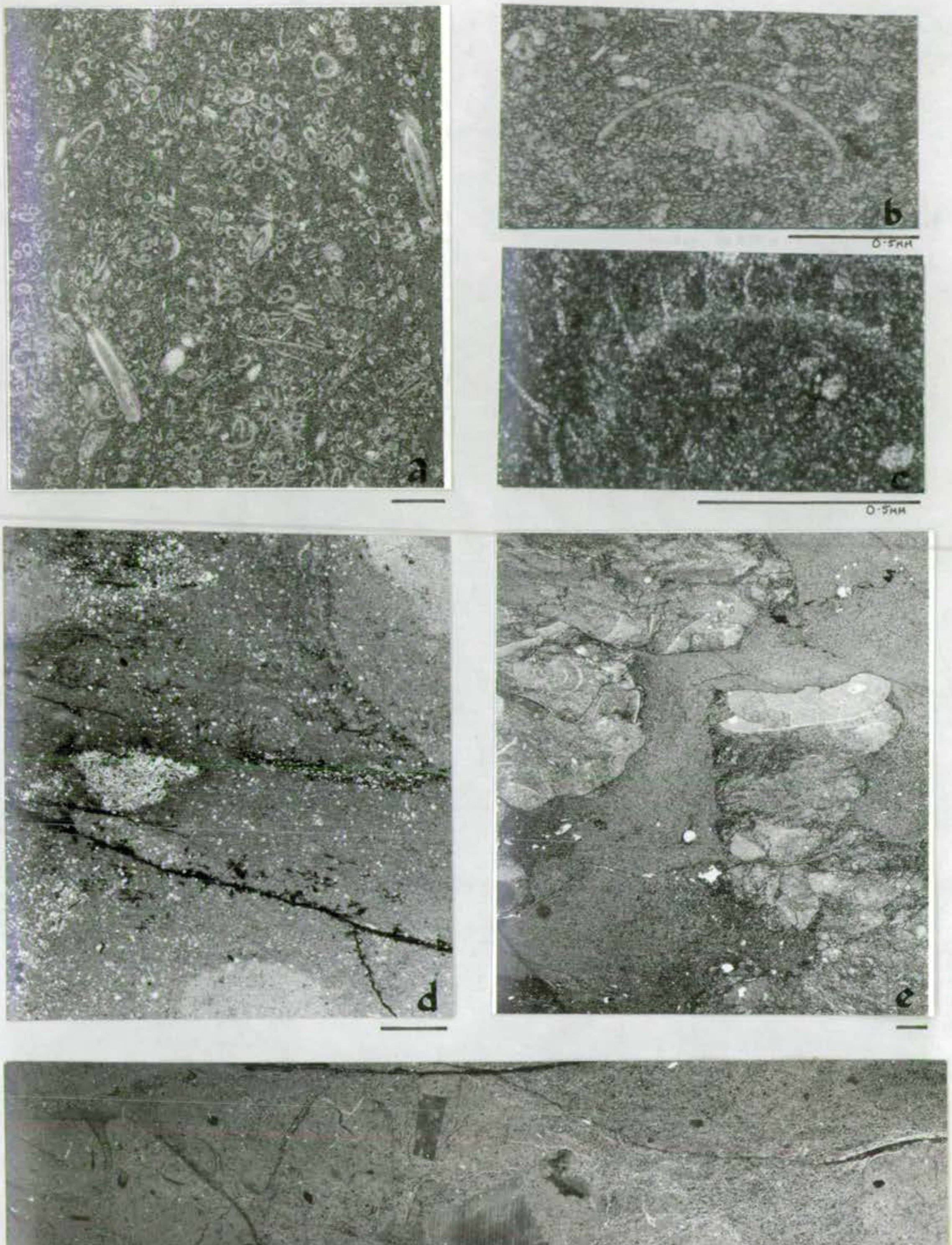
(b) Smooth-shelled ostracod (possibly Entomozoe, Ungarella, Entomoprimitia or Waldeckella sp.) found in most of the nodular limestones examined but never abundant. Thin section, plane polarized light. Sample no. D18.

(c) Ridged-shelled ostracod (possibly Richterina, Fossirichterina, Maternella or Volkina sp.) never seen in Givetian and Lower Adorf nodular limestones, but important constituents of those of later age. Thin section, plane polarized light. Sample no. E124.

(d) Burrows in nodular limestone (lighter mottled patches) rich in quartz silt (white specks). The patchiness of the quartz distribution is thought to be due to bioturbation. Thin section, plane polarized light. Sample no. F23.

(e) Irregular (?)submarine karst in nodular limestone. The ammonoid capping the pillar (right of centre) shows no signs of differential solution of shell relative to sediment infilling. It is thought, therefore, that solution took place after shell replacement. The fact that the ammonoid is filled with sediment, indicates that it must have been partially broken or dissolved prior to burial. Methylene blue stained peel, plane polarized light. Sample no. K3.

(f) Possible hardground in nodular limestone. Coarse material from above penetrates down into microsparite. The sharpness of the junctions suggests that such "funnels" may be borings rather than burrows in soft sediment. Thin section, plane polarized light. Sample no. D18. Negative print.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supported, a situation which could only arise if sedimentation of fine material had virtually stopped over the period the shells were being deposited, or if the shells themselves were winnowed from the sediment by bottom currents.

Ostracods. The importance of ostracods in the basin sediments has already been emphasized (page 47), but they are no less abundant in many of the schwellen limestones.

Rabien (1956) concluded that the ostracods present were pelagic, but in view of the identification difficulties encountered in the present study, the possibility of some benthonic forms being present in the limestones cannot be excluded. In thin section, only two ostracod types could be identified (most of the specific characters being observable only in completely exposed specimens).

(i) A smooth unornamented or weakly ornamented elliptical form, present but never abundant, in rocks of all ages and all environments of deposition (Fig 4.3b). These are generally articulated, either with shells preserved or as calcite filled moulds.

(ii) A ribbed form (or series of forms), absent in limestones older than Lower Adorf age, but an important constituent of younger rocks, occurring both as articulated and disarticulated valves, lying at all angles to the bedding in the majority of sediments, but parallel to the bedding, "convex-up" in rocks with high clay mineral content (Fig 4.3c). In rocks of post-Frasnian age, these ribbed ostracods are often the dominant fossil components, frequently rivalling the cricoconarids of the Givetian and Lower Frasnian in their rock-building potential. Their importance in Upper Devonian nodular limestones and their appearance on the scene just at the time when circoconarids were on the point of extinction suggests that they may have taken over the pelagic niche left by the then fast declining cricoconarids.

Arenaceous foraminifera, though present in a few of the limestones, do not appear to have been important faunal constituents.

The commonest types encountered were meandering tube-like forms (Tolypammina sp), though a few planispirally coiled (Ammodiscus sp) and globular (Saccamina + Saccaminopsis sp.) forms were encountered in some samples (Appendix III, plate 4, Figs 37-44). A complex siliceous labyrinthine structure observed in a few specimens, is tentatively classed here as a foram though its exact taxonomic position is unknown (Fig 4.3d).

The ecological significance of the foraminifera in these sediments is difficult to assess due to their rarity and the fact that most were recovered from insoluble residues, No encrusting forms were seen, however, and it is possible that the few specimens seen had undergone bottom transport or that they were originally epiplanktonic, attaching themselves to floating algae, no trace of which has been preserved.

Ammonoids are sporadically distributed, but are locally so abundant as to suggest concentration by condensation or current action. With the exception of the larger shells (i.e. those greater than 1-2cm.), skeletons seldom show any preferred orientation. This is not surprising, however, since many of the ammonoids are globular in shape and little preferred orientation would be attained under any current conditions.

In many cases, the ammonoid shells were severely damaged, often with inner whorls completely missing, suggesting predator breakage or partial solution, prior to burial. No strong evidence for current or compaction breakage could be found.

It is assumed that Devonian ammonoids, like the modern nautiloids, were nektonic carnivores or scavengers.

Bivalves. Definite bivalve shells (Buchiola sp.), which could be identified in hand specimen, were seen only in the Kellwasser limestones at Altenvers, but many of the limestones contain small, thin, slightly curved, recrystallized shells or spar-filled moulds thought to be bivalve shells.

Apart from Buchiola, the only bivalves reported from the Middle and Upper Devonian schwellen sediments in the Rheinisches Schiefergebirge are thin shelled Pectinaceans (Posidonia sp.) (Rabien, 1956). The characteristics of the thin shells seen in thin sections are wholly compatible with those of the above genus. No large Megalodus bivalves, found in Adorf massive limestones elsewhere in the Rhenish trough, were found (Clarke, 1885;

Lotz, 1900). It is thought unlikely that these thin bivalves could have been transported from a shallower water source for the following reasons:

(i) they show no increase in frequency of occurrence adjacent to presumed shallow-water sediments;

(ii) they lack micrite envelopes, a common feature seen in molluscs in shallow water limestones;

(iii) the shells are so thin that they would stand little chance of preservation if derived from a shallow high-energy environment.

The occurrence of these bivalves in both the schwellen limestones and basin shales in association with a dominantly pelagic fossil assemblage, suggests that they are "native" to the basin realm and that their distribution in life was independent of both water depth and substrate type. For these reasons the Posidoniids are thought to have been planktonic in the Devonian as they apparently were in the Jurassic (Jeffries and Minton, 1965). An epiplanktonic mode of life has been proposed for Buchiola (Schmidt, 1935).

Most of the valves seen in the present study were disarticulated with "convex-upwards" orientation along bedding planes, though some valves occur at high angles to the bedding, perhaps due to bioturbation. Severe shell breakage is rare, and that which is seen is more likely to have been caused by compaction or biomechanical processes than by strong current activity.

Crinoids are represented in many of the nodular limestones by ossicles and plate fragments, but only in those sediments stratigraphically above or adjacent to detrital crinoidal biosparites in the Eberstein area, are they at all common. In cases such as this crinoidal derivation from neighbouring bioclastic sediments seems the most likely source for the crinoidal material, but isolated ossicles also occur in situations where no immediate source can be found (e.g. Weipoltshausen). Sparse crinoidal development may, therefore, have been possible in some areas in which schwellen limestones were deposited. It is equally likely, however, that these scattered ossicles underwent considerable transportation from a distant shallow-water source, as crinoidal material is easily transported due to its buoyancy after death (Cain, 1968).

Articulated brachiopods, though common in some of the massive limestones, are only occasionally found in the nodular limestones, and then normally only as disarticulated and broken shells.

Brachiopods are most common in those nodular limestones directly overlying or laterally adjacent to massive bioclastic schwellen limestones, in which brachiopod fragments are common, suggesting that most, if not all of the brachiopod material was derived from a shallower higher-energy environment. A similar degree of disarticulation and breakage could, of course, have been accomplished by scavengers or predators, but the possibility of most of the brachiopods being indigenous to the"nodular limestone" environment is rejected as little evidence of widespread hardground formation has been found, the brachiopods requiring a hard substrate for attachment. At Weipoltshausen, however, no shallow-water provenance from which the brachiopods found there (Herrmann, 1909) could have been derived has been located. Perhaps, here, and locally in the Bieber area, small brachiopod faunas could establish themselves by attachment to shell fragments, or to local submarine hardgrounds. (The entire Givetian sequence is, at Weipoltshausen, less than 5 metres thick and indications of disconformities can be seen in the limestones.) Some modern brachiopods (e.g. Chlidonophora chuni, Blockmann) are capable of attaching themselves directly to "soft" substrates by the use of a root-like pedicle system (Rudwick, 1961), but it is unlikely that this type of attachment had developed in the brachiopods in the rocks examined.

Inarticulate brachiopods

Several small Acrotethaceans were recovered with conodonts in insoluble residues of many of the nodular limestones (Appendix III, Plate 4, Figs 45 and 46). They are not, however, restricted to these rocks and have been found also in several samples of massive limestones from the Bieber area.

In view of the statements above on the ecological needs of brachiopods, and the facies independence of these forms, it is thought most likely that they were epiplanktonic, perhaps attaching themselves by their pedicles to floating algae (Ager, 1962). Furthermore, their small size would make life on a soft substrate virtually impossible (Rowell and Krause, 1973). Trilobites are ubiquitous though numerically unimportant members of the fauna of these limestones. Disarticulated fragments are normally all that are seen in sections, although nearly complete specimens have been recovered from the Givetian Weipoltshausen limestones. (Hermmann, 1909).

Trilobites as a group appear to have inhabited a wide range of environments in the Rhenish geosyncline, and there are no reasons for believing the trilobite fragments found in the limestones to be anything but autochthonous, though the fragments probably have undergone a certain amount of post-mortem movement, if only through bioturbation.

Rabien (1956) reported 16 species (all Phacopids) from the shales and nodular limestones of the Waldeck Syncline. (Proetids were also found in rocks of Wocklum age, but no limestones of this age were examined in the present study). No other trilobite genera were noted. The trilobite fauna listed for the Weipoltshausen limestones, however, is more varied containing Cheirurids, Harpids and Cyaphaspids as well as Phacopids and the nature of this fauna resembles those of massive limestone sequences of equivalent age elsewhere in the Rheinisches Schiefergebirge (Clark, 1885; Maurer, 1889; Beyer, 1896) rather than those described by Rabien (op. cit.).

Corals are normally absent in the nodular limestones, but in the Weipoltshausen sequence one small solitary rugose coral (tentatively identified as "aff. Barrandeophyllum") was found. As no evidence exists for a shallower water area in the neighbourhood from which it may have been derived, it is assumed that sparse coral development was possible on the Weipoltshausen Schwelle, at least for part of the Givetian. As has been noted above, many aspects of the fauna of the Weipoltshausen limestones suggest that these were deposited in water depths shallower than those in which the majority of the nodular limestones were apparently deposited, yet not shallow enough to allow reef development.

Apart from the material listed above the only recognizable fossil fragments found were isolated sponge spicules in a few of the shalier limestones.

Skeletal preservation

The degree to which fossil particles can be recognized depends partly on their mode of fossilization and partly on the extent to which original characteristics have been destroyed by subsequent neomorphism. Neomorphism is dealt with in detail later (page 116) and only fossilization processes are dealt with below.

In the studied rocks, skeletal material is generally preserved in one of the following ways:

(1) Unaltered hard parts, where no apparent mineralogical or textural changes appear to have

Fig 4.4

(Scale bar represents 1mm. unless otherwise indicated)

(a) Preservation of original fibrous microstructure in brachiopod. Methylene blue stained peel, plane polarized light. Sample no. K1.

(b) Concentration of broken ammonoid outer whorls (Preservation type aIII). Methylene blue stained peel, plane polarized light. Sample no. E21.

(c) External mould of articulated ostracod. Void filled with sparry calcite. (Preservation type cII). Thin section, plane polarized light. Sample no. E118.

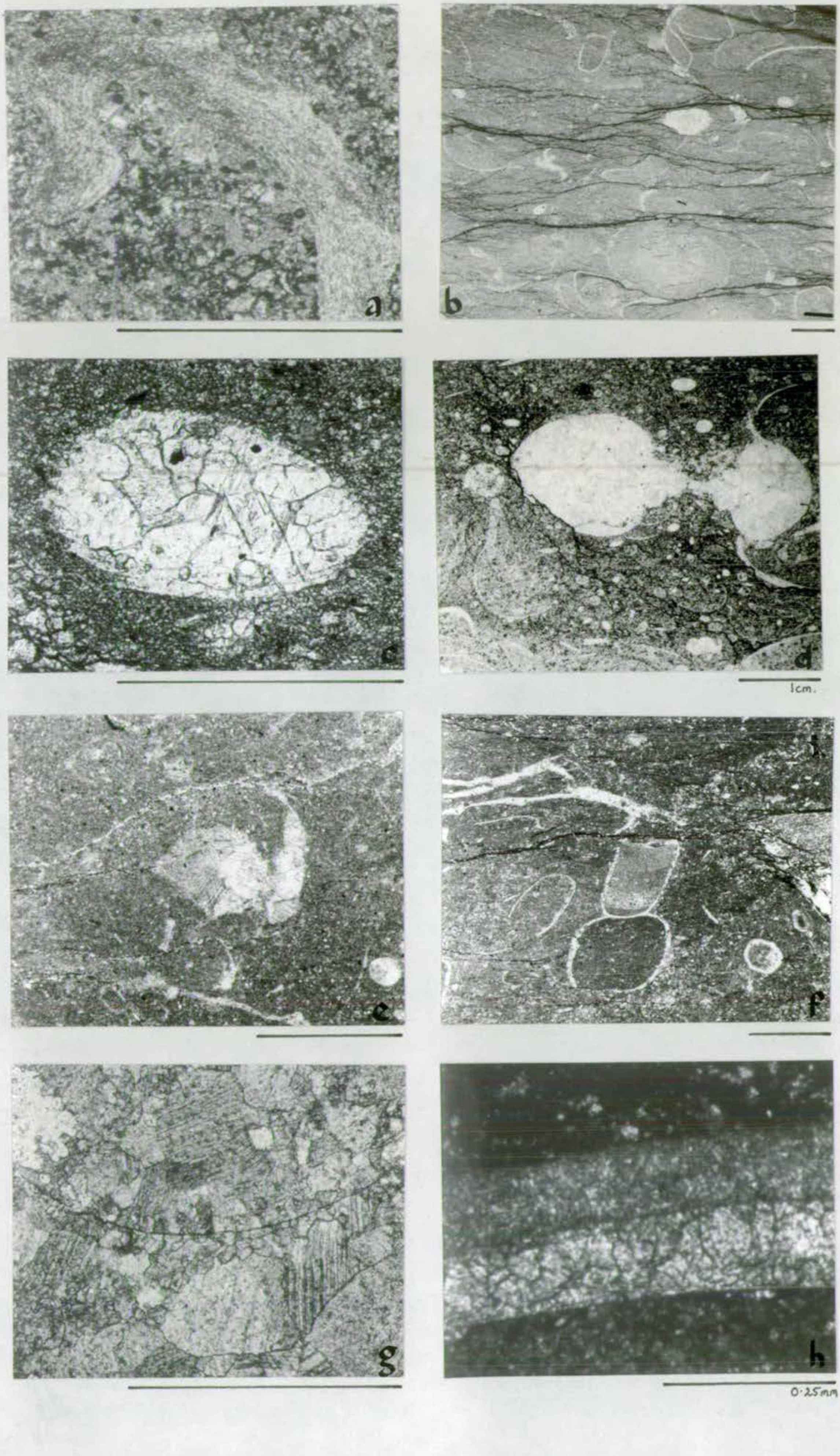
(d) Apertural filling of ammonoid. Wall shape retained in outer but not inner whorls. (Preservation type bII). Methylene blue stained peel, plane polarized light. Sample no. A2.

(e) Partial apertural filling has resulted in formation of a cavity above geopetal sediment in outer whorl (vertical junction). Lower part of ammonoid is no longer preserved. (Preservation type bII/cIII). Thin section, plane polarized light. Sample no.D17.

(f) Draft filled ammonoid. Wall shape preserved throughout. Sediment laminae prove shell filled in vertical position. (Preservation type aIII). Thin section, plane polarized light. Sample no. E132.

(g) Arcuate dust line in sparry calcite marking former wall edge of inner whorl of ammonoid. The high angled indistinct dustline (upper left) is a septal trace within the spar. (Preservation type cIV(?)). Thin section, plane polarized light. Sample no. N18.

(h) Shell layering preserved in ammonoid. Upper (originally outer) layer is composed of small dusty crystals orientated roughly parallel to shell edge, those in lower (inner) layer are coarser grained and resemble void-filling spar. (Preservation type aII). Methylene blue stained peel, plane polarized light. Sample no. Wp15.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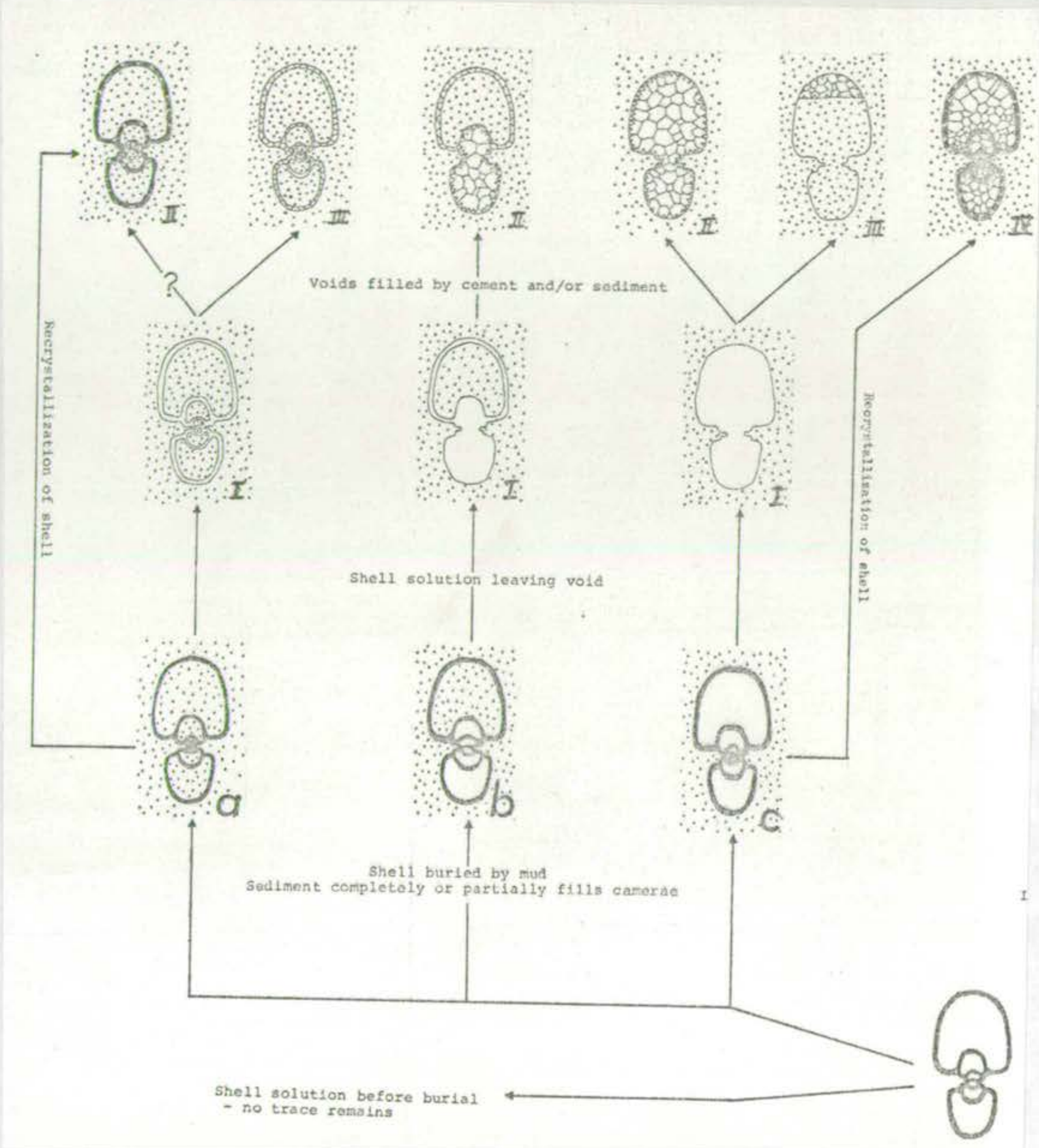


Fig. 4.5. Diagram of fossilisation possibilities for ammonoid shells.

black - original shell material

black & white banded - recrystallized shell material

stipple - sediment matrix

mosaic - void-filling calcite

blank - void

taken place.

(2) Replaced hard parts, where skeletal replacement by calcite and, more rarely, haematite and silica, has taken place, through filling of voids resulting from solution of skeletons.

(3) Recrystallized hard parts, generally involving textural but not compositional change without an intervening void stage.

(1) Unaltered hard parts

Skeletons and skeletal fragments retaining details of original microstructure are common in those elements of the fauna possessing mineralogically stable calcite or calcium phosphate shells (e.g. brachiopods, crinoids, corals, foraminiferids, conodonts) (Fig 4.4a). In some cases, however, later neomorphic crystal growth has affected some skeletons with a consequent loss of detail.

(2) Replaced hard parts

Replacement of carbonate skeletons by non-carbonates is rare, and only occurs in local "special" environments. (Page 398).

Replacement by void-filling calcite spar is common, however, especially in molluscs. Possibly because of their complex morphology, ammonoids show a greater range of preservation types than other molluscs, and so the discussion below is largely concentrated on them.

How well the form of an ammonoid is preserved in the studied limestones is dependant on the effectiveness of burial, the best preserved ammonoid shapes being both surrounded by and completely filled with fine sediment. Non-burial would have resulted, in extreme cases, in complete solution and non-preservation of ammonoid shells. As sediment deposition rates were probably very low over most of the period of deposition of these limestones, complete solution of ammonoids at the sediment surface may have occurred. In some limestones, evidence for post-burial (and post-lithification) solution is common (Fig 4.3e), and some limestones contain concentrations of "incomplete" ammonoids, in which outer whorls are common, but inner whorls rare (Fig 4.4b). In the absence of any evidence of strong current abrasion, the present state of the ammonoids can only be adequately explained by breakage by predators or scavengers, or preferential solution of the inner (thinner) whorls, prior to burial.

Ammonoids in which the shell material has been dissolved and infilled by calcite are found in one of the following forms in the studied limestones:

(i) As spar or sediment-filled moulds. All traces of the shell have been removed by solution, so that only the external shape is preserved (Fig 4.5c II + III). Preservation of the shell will not be accomplished, of course, in unlithified sediments, due to collapse of sediment into the solution void. Bivalves and articulated ostracods are often preserved in this way (Fig 4.4c), though ammonoids seldom are due to the ease with which sediment can enter the living chambers of ammonoid shells.

(ii) As part sediment-, part spar-filled moulds. Many ammonoids are preserved in this way, the walls of the interior whorls having completely dissolved as in (i), but the internal and external wall positions are retained in part of the outer whorl by infilling of the last chamber with sediment ("apertural-fill"; Seilacher, 1973). (Fig 4.5bII, Fig 4.4d). In those cases where the living chamber has only been partially filled by sediment, the internal-shape of the wall is preserved only adjacent to the sediment infill, which is overlain by a spar filled geopetal cavity in which no signs of wall structure can be seen (Fig 4.4e).

(iii) <u>As completely sediment-filled moulds</u>. Internal and external wall boundaries are preserved throughout, if filling is complete (Fig 4.5aIII, Fig 4.4f). Effective filling of all the chambers could not have been accomplished through the siphuncular opening alone; secondary openings (caused by abrasions, corrosion, or predator damage) are essential if sediment is to gain access to inner chambers ("draft-dill"; Seilacher, 1973). The common occurrence of this type of preservation indicates that most of the ammonoids in the studied limestones must have been severely damaged prior to burial.

In one specimen the mode of preservation is problematical. The camerae have been partially filled with internal sediment yet "ghost" traces of internal walls and septa are seen, not only beneath internal sediment lenses, but also within the spar areas above (Fig 4.4g). The original textural relations have been modified by recrystallization of spar and sediment alike. The retention of traces of wall within the spar can be explained only if recrystallization of the shell took place with subsequent filling of the camerae with cement, or if the shell was dissolved and later filled with spary calcite after all of the camerae had themselves been filled by calcite cement (Fig 4.5cIV). Recrystallization is thought more likely, though no microstructural details have been preserved within the "shell areas".

(3) Recrystallized hard parts

Only in those cases where details of shell microstructure are retained can recrystallization be proved (Fig 4.5 II). The only ammonoids retaining a distinct double layering within the shell occur in the nodules in the basin shales (Fig 4.4h). The retention of shell structure indicates that no complete shell void stage was present at any time. The special diagenetic conditions offered by this clay rich environment are thought to be largely responsible for this mode of fossilization.

Diagenesis

Bioturbation

The earliest post-depositional event recorded in the schwellen limestones is bioturbation. It is believed that burrowers were responsible for wholesale reworking of the sediment during and after deposition for the following reasons:

(i) No signs of original sedimentary lamination are present in the limestones. The absence of such laminations could only be due to unchanging sedimentation over a long period of time, or to the destruction of any record of such sediment changes by thorough mixing. In the case of the limestones considered here both processes probably played some part, as sedimentation is likely to have been constant over large periods of time. The lack of any record of any vertical change however would seem to suggest that such minor changes were obliterated by biogenic homogenization of the sediments. In some limestones, for instance, "pockets" of quartz silt are found (Fig 4.3d). Such localized patchy concentrations are difficult to reconcile with even, pelagic sedimentation of the carbonate <u>unless</u> one is prepared to accept that the localization of the terrigenous silt occurred after and not during deposition.

(ii) Fossils within the limestones show no preferential orientation parallel to the horizontal, but are normally found at all angles to the "bedding".

(iii) Intensive mixing of sediment is indicated by colour and grain size mottling (Fig 4.2d).

(iv) Direct evidence, in the form of sediment filled burrow structures, is found in many limestones.

No signs of bioturbation were seen in the nodules in the basin shales. In common with the surrounding shales, they show a well developed lamination marked by fossil orientation along the bedding (Fig 4.2c).

Lithification

Lithification (i.e. the establishment of a rigid framework irrespective of porosity) took place at a very early stage in the diagenetic history of the pelagic limestones studied:

(1) fragile curved bivalve shells retain their curvature and remain unbroken, even at high angles to the bedding in the schwellen limestones, indicating that the rocks were already cemented prior to compaction. In the shaly nodules in the basin sediments, however, thin shells are often slightly crushed. Some compaction had obviously already taken place before lithification of these rocks (Fig 4.2c).

(2) Burrow structures do not appear squashed.

(3) Shell moulds of originally aragonite fossils are preserved proving that the matrix was self supporting prior to solution of aragonite.

(4) Sediment filled fissures are present in a few limestones, indicating that the limestones were cohesive enough to fracture while still in the surface or near surface environment.

No evidence can be found in either the schwellen or basin limestones to indicate that they were ever subaerially exposed prior to orogenic uplift in the Carboniferous. On the contrary, the lithological features of the rocks indicate that the sea bottom underwent almost continuous deepening from the Middle Adorf stage to the Carboniferous. One must conclude therefore that the cementation of these limestones occurred soon after deposition at or near the sediment surface in the submarine environment.

In recent years many examples of early submarine lithification of calcareous muds have been reported (Friedman, 1964; Gevirtz and Friedman, 1966; Milliman, 1966, 1971, 1973; Cifelli et al., 1966; Fischer and Garrison, 1967; Thompson et al., 1968; Friedman et al., 1971; MacIntyre et al., 1971; Marlowe, 1971; Müller and Fabricius, 1974). In most cases the cement is magnesium calcite, though aragonite cements are found in areas of unusually high salinity and temperature. (Gevirtz and Friedman, 1966). Both magnesium calcite and aragonite tend to develop as acicular or micritic cement crusts on and within pelagic limestone components.

Due to subsequent recrystallization, the nature of the cement in the nodular limestones studied remains uncertain, although it appears probable, if one accepts that most of the schwellen limestones were once "micrites", that much of the cement was itself micritic. Perhaps much of the micritic "matrix" was indeed cement. If this were so then the origin of the micrite (page 84) no longer presents such a problem. In a few rocks, mainly cricoconarid shell beds, the micritic "matrix" is absent. Instead the styliolinids are "cemented" by overgrowths of fibrous calcite, most probably replacing an earlier acicular cement (Kendall and Tucker, 1973). Whether this acicular cement was aragonite or magnesium calcite however is unknown.

Whether lithification was effected at the sediment/ water interface, (Müller and Fahricius, 1974), or diagenetically below the surface, is open to controversy. Both submarine solution, with formation of hardgrounds (Hollmann, 1964; Garrison and Fischer, 1969), and diagenetic nodule formation (Hallam, 1967; Jenkyns, 1974) have been proposed as the cause of the nodular structure of the lithologically similar Ammonitico Rosso and Adnet Limestones of the Alpine Jurassic.

As with the above mentioned limestones, evidence can be found in the Devonian nodular limestones to support both processes. Foram-encrusted hardgrounds have been found in the Rheinisches Schiefergebirge in extremely condensed schwellen limestone sequences (Tucker, 1973). Encrusted hardgrounds have not been observed in the present study, but a few limestones show irregular lithological junctions, not modified by subsequent stylolitization, with structures which could be interpreted as borings penetrating into the limestone (Fig 4.3f). In some cases, the irregular relief is demonstrably the result of subsolution as the limestone surface is deeply embayed and sediment filled fossil shells are truncated (Fig 4.3e). Subsolution probably played some part in the lithification of limestones on the schwellen tops, but no evidence could be found that it was primarily responsible for the nodular structure of the limestones. A diagenetic origin through early sub surface migration of $CaCO_3$ is thought more likely for the following reasons:

(1) With increasing $CaCO_3$ content, shales with scattered nodules pass gradually into almost pure limestones composed of aggregates of nodules.

(2) The frequent alternations of limestone nodule bands and thin shale bands are difficult to explain in terms of primary sedimentalogical controls, since there is no indication in the area of cyclical changes in depositional conditions which could give rise to such marked sediment changes. Such a lithological alternation, however, would be expected if $CaCO_3$ was dissolved at certain levels in the sediment and redistributed in others. Such a process would result in exaggeration of original lithological variations.

No new explanation for nodule origin could be found from a study of the rocks in the Bieber area. Perhaps solution and redistribution of $CaCO_3$ was brought about in the subsurface environment by organic decomposition and heterogeneous nucleation within the sediment (Berner 1968, 1971), by solution in the reducing zone with migration of dissolved carbonate upwards into the oxidation zone and subsequent precipitation (Gründel and Rösler, 1967), or by some process whereby nodule layer spacing was controlled by rate of sedimentation (Jenkyns, 1974).

Whatever the process of nodule formation, the rate at which nodules formed was obviously dependent on the total $CaCO_3$ content of the sediment. Thus nodules in the lime-rich sediments on the schwellen tops and upper slopes were formed before early compaction could have any effect on their constituents, whereas those nodules developed in the lower slope and basin environment, where the calcium carbonate is heavily diluted by detrital clay and silt, show signs of compaction prior to the establishment of a rigid carbonate nodule structure.

At some stage during compaction inversion of the presumed high Mg calcite constituents to the present low Mg mineralogy, probably took place, although no textural evidence of this transformation exists, or would indeed be expected. It is certain, however, that after lithification, some calcitization of aragonitic cements and allochems occurred with the production of calcite pseudomorphs (e.g. the fibrous calcite replacements of original aragonitic (?) acicular cements and some calcitized mollusc shell fragments), although actual solution of aragonitic shells with subsequent infilling of the voids by low Mg sparry calcite cement was of more widespread importance.

Sparry calcite cement is, as yet, unknown in the submarine surface environment, but is a common cement in the magnesium-depleted fresh water environment. It is also believed to be the major cement type in deep burial environments (Folk, 1973). It follows, therefore, that precipitation of sparry calcite either occurred when the rocks were exposed to meteoric waters during the Mid-Carboniferous or during burial while still in the submarine realm. As the spar-pore fillings are both cut by early stylolites (Fig 4.4d) and are affected by microspar growth, the cement appears to be of fairly early age, certainly pre-tectonic, and must therefore have formed in the subsurface environment while the limestones were still submerged.

Stylolitization

As shown above, the process of lithification of the limestones tends to result in nodule formation and the segregation of the original sediment into alternating carbonate-rich and carbonate-poor zones. During compaction these lithological contrasts were further accentuated by the development of horizontal stylolites between the limestones and shale seams and between the nodules themselves. Solution along these stylolites may have been more or less continuous or sporadic, but was certainly long lived as can be seen by the displacement of calcite veins by solution (Fig 4.6a).

In many rocks a later set of stylolites are developed,

Fig 4.6

(Scale bar represents 1mm. unless otherwise indicated)

(a) Calcite filled vein cut by solution seams. No trace of vein found above or below shale seam. Methylene blue stained peel, plane polarized light. Sample no. H29.

(b) Sigmoidal nodules produced by interference of horizontal "compaction" stylolites and high angled "tectonic" stylolites. Hand specimen. Scale bar = 1cm. Sample no. D27.

(c) Bladed fibrous calcite on cricoconarid test, partially removed by pressure solution (longitudinal section). Methylene blue stained peel, plane polarized light. Sample no. F20.

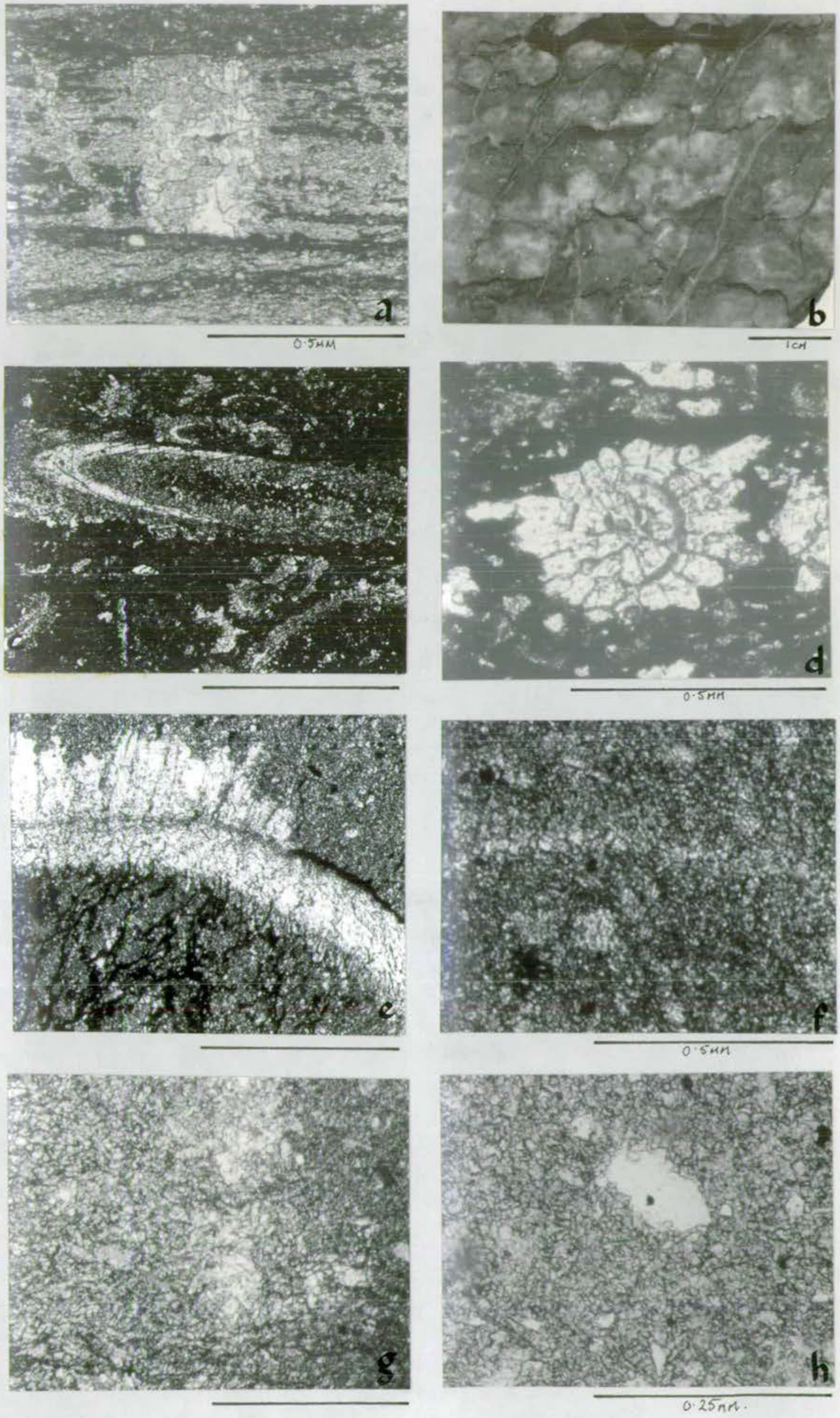
(d) Concentrically arranged inclusion zones in fibrous calcite overgrowths on cricoconarid shell (retained as dark ring) (cross section). Methylene blue stained peel, plane polarized light. Sample no. K1.

(e) Bladed fibrous calcite growth partially developed on brachiopod shell. Methylene blue stained peel, plane polarized light. Sample no. K10.

(f) Thin shell preserved only as clear zone within more turbid microspar surroundings. Shell margins obliterated by microspar growth. Thin section, plane polarized light. Sample no. E128.

(g) Microspar (right) replaced by loafish grains of pseudospar. No primary textural details recognizable in pseudospar. Thin section, plane polarized light. Sample no. N18..

(h) Microspar "nibbling" of detrital quartz grain. Thin section, plane polarized light. Sample no. E118.


a
b
0·5MM
1CM
c
d
0·5MM
e
f
0·5MM
g
h
0·25MM.

often nearly perpendicular to the earlier set. These are the result of the change from vertical (compactional) to lateral (orogenic) linear stress during the Mid Carboniferous. The interference of these two sets of stylolites tends to break the rock up into a series of more or less sigmoidal nodules, each separated from its neighbours by a thin clay seam (Fig 4.6b). How well these structures are developed, however, depends on the nature of the original rock (purer limestones often tend to be more irregular in their reaction to stylolitization) and its orientation relative to the axis of linear stress at the time of "tectonic" stylolitization (if vertical, for example, the second set of stylolites are developed in roughly the same direction as the first set).

The shaly nodules in the basin shales are always free from signs of stylolitization. All of the compaction effects are probably absorbed by the surrounding shales (which are heavily compacted around the limestone nodules) so that none are transmitted to the limestones themselves. Similarly, signs of tectonic deformation are common in the shales, but none were apparent in the limestone nodules examined.

<u>Neomorphism</u>

Very few if any of the textures shown by the Devonian nodular limestones, studied are primary. Most are the result of neomorphic changes which have taken place since the sediments were deposited.

Three types of neomorphic change are important, all three being "aggrading" in character (Folk, 1965):

(1) Replacement of original aragonitic shell material by calcite with retention of details of the shell structure;

(2) Replacement of originally acicular cements by fibrous calcite pseudomorphs;

(3) Replacement of matrix, allochems and sometimes diagenetic fabrics by coarser grained calcite (usually microspar, but sometimes pseudospar).

(1) Replacement of aragonitic shells.

The only definite signs of calcite replacement of shell material are to be found in the shaly nodules of the basin facies, where ammonoid and bivalve shells are preserved with the original shell layering intact (page 106). The calcite within the shell shows none of the characteristics of void-filling spar (Bathurst, 1971) but appears to have crudely imitated the original crystal orientation trends within the shell. This mode of preservation contrasts sharply with that of the ammonoids in the schwellen limestones which are always preserved as spar-filled moulds. Most probably the low permeability and high magnesium content of the shales, together with the protection of organics (e.g. amino-acids) were responsible for the preservation of aragonite in this environment, when it was being dissolved elsewhere (Füchtbauer and Goldschmidt, 1964; Kennedy and Hall, 1967). The time of replacement by low magnesium calcite cannot be determined due to the lack of other diagenetic features within these nodules.

(2) Replacement of acicular cements.

Apart from one or two shells with bladed fibrous overgrowths projecting into the matrix (which may not have had any acicular "base") (Fig 4.6e), this type of neomorphism is found only in cricoconerid rich layers of Givetian and Lower Frasnian age. These cement-replacement fibrous calcites have recently been intensively studied by M.Tucker (1973) who convincingly demonstrated how such a replacement could occur. The original acicular cement is thought by Tucker to have been aragonitic, but no evidence exists for this assumption in the limestones in the present study, and it may equally well have been high or low Mg calcite. The time of replacement is thought by Tucker to coincide with subaerial exposure of the limestones in the Mid-Carboniferous, mainly because a source of Mg depleted pore water is needed to bring about the neomorphic change. As was shown above, however, such magnesium deficient water could be present in a deep-burial environment and so replacement may have occurred at a much earlier stage, especially so since these fibrous calcite overgrowths are affected by both pressure solution and microspar replacement (Fig 4.6c).

(3) Aggrading neomorphism.

The most important neomorphic change the limestones have undergone is wholesale aggradation neomorphism to microspar, or, in a few cases, to pseudospar (Fig 4.6g).

The increase in grain size has mainly affected the matrix, but recrystallization has been so widespread that thin ostracod and bivalve shells are barely discernable against the matrix (Fig 4.6f).

Aggrading neomorphism beyond the micrite size limit (4μ) can take place only when magnesium contained between the grains, inhibiting their lateral growth, has been removed (Folk, 1973). This can easily be accomplished during subaerial exposure by flushing with fresh (magnesium-deficient) waters, but could equally well occur during burial, by flow of magnesium deficient connate waters through the rocks.

Textural evidence in the rocks studied suggests that growth of much of the microspar occurred at a late stage in diagenesis since quartz grains are etched by, and both void filling and fibrous calcite replacements, are affected by microspar formation (Fig 4.6h). Calcite filling tectonic voids, however, is unaffected by neomorphism indicating that the microspar growth was either pre- or syntectonic.

There is no evidence, however, that all of the microspar is of the same age, and it is likely that microspar formation has occurred more than once, whenever conditions existed in which magnesium could be removed effectively from the rocks. During dedolomitization of the partially dolomitized limestones at Rotenberg, for example (page), magnesium was removed on a large scale. This may explain why these rocks are the most affected by microspar formation of those examined in the present study. Aggrading neomorphism has been reported accompanying dedolomitization and surface weathering elsewhere (Chafetz, 1972).

No trace of aggrading neomorphism was observed in the shaly nodules. This is not surprising however when one considers that the grain size of the non-skeletal calcite is largely determined by the size of the pore space between clay grains. Thus, it is doubly difficult for these calcite grains to increase in size since they are not only bounded by non-carbonate material, which would have to be forcefully displaced during growth, but this material contains magnesium which itself inhibits the growth of the calcite.

A summary of the diagenetic history of the schwellen and basin limestones is given in Fig 4.7.

Fig 4·7 Comparison between diagenotic histories of schwellen and basin "pelagic" limestones

	Schwellen Limestones	Shaly Nodules
Early Near Surface Diagenesis	Deposition Bioturbation Cementation and nodule formation	Deposition Slight compaction Cementation and nodule formation
Burial Diagenesis	Shell solution and inversion of unstable carbonates Precipitation of sparry calcite in voids Microspar growth Veining Development of horizontal stylolites (polyphasal)	Shell recrystallisation (?) inversion of unstable carbonates Precipitation of sparry calcite in voids
Tectonism and Metamorphism	Development of neat vertical stylolites, deformation, non-$CaCO_3$ mineral growth and veining	
Metasomatism	Non $CaCO_3$ mineral growth +(?) microspar growth	
Exposure	(?)Microspar growth Present rock texture	Present rock texture

Summary

In its general nature the fauna of the pelagic limestone facies differs little from that of the contemporaneous basin facies. In both, nekton and plankton predominate, benthos being rather rare. The present differences between the faunas are, it is thought, due mainly to differing sedimentation rates and selectivity of preservation, so that siliceous organisms are abundant only in siliceous sediments, and calcareous fossils only in carbonate sediments. The major faunal features of the "pelagic" facies * are compared in Fig 4.8 in which the probable ecological roles of the faunal constituents are also listed.

Certain faunal differences do exist, however, between the basin and schwellen, sediments: in the basin sediments, benthonic animals are virtually absent, but are found in the schwellen limestones, especially those which developed on schwellen tops or adjacent to massive limestone developments.

The preserved fauna is made up of the following elements:

(a) a pelagic nektonic death assemblage dominated by ostracods, cricoconarids, ammonoids and conodonts, but

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* The terms "pelagic facies" and "pelagic environment" refer to those sediments and that general environment where little or no apparent contribution was made by shallow-water deposits or by benthonic life-forms, generally considered to have lived in shallow marine environments. In the studied area such environmental conditions were present in areas away from the Bieber Schwelle in the Upper Givetian and Adorf, and throughout the region from then until the Lower Carboniferous.

	PAUNA	Ammonoids	Conodonts	Ostracods	Cricoconarids	Radiolarians	Bivalves	Inarticulate Brachiopods	Foraminifera	Articulate Brachiopods	Crinoids	Rugose Corals	Tabulate Corals	Bryozoans	Stromatoporoids	Siliceous sponges	Algae	Trilobites	Gastropods	"Surface-feeding Trails"	"Burrows"
	LIFE HABIT	NEKTON	PLANKTON					EPIPLANKTON	FIXED EPIFAUNAL BENTHOS									VAGILE EPIFAUNAL BENTHOS			VAGILE INFAUNAL BENTHOS
CONSUMER TYPES	Primary Producers																X				
	Grazing herbivores																		X	X	
	Detritus feeders																	X		X	X
	Suspension feeders		X	X	X	X	X	X	X	X	?X	X	X	X	X	X					
	Carnivores and Scavengers	X	X	X																	
PREDOMINANT ORIGINAL SKELETAL COMPOSITION	Calcite	-		X	X					X	X	X	X		X		X	X			
	Aragonite	X																			
	Calcite and Aragonite						X												X		
	Calcium phosphate		X					X						X							
	Silica					X			X							X					
MODE OF FOSSILISATION	Fossil preserved entirely	-	C	C	-	-	-	C	C	C	C	C	C	C	C	C	C	C	-		
	Fossil replacement	R	-	C	C	C	R	-	-	-	-	-	-	-	-	-	-	-	-		
	Fossil disolutiant fill	C	-	C	-	C	C	-	-	-	-	-	-	-	-	-	-	-	C		
	Replacement by new mineral	-	-	R	-	-	R	-	R	-	-	-	-	-	-	R	-	-	-		
LITHOLOGICAL DISTRIBUTION	Greywackes	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	R	-
	Cherts	-	P	-	-	C	-	-	-	-	-	-	-	-	-	C	-	-	-	-	-
	Siliceous shales	R	P	-	-	P	-	-	-	-	-	-	-	-	-	C	-	-	-	-	-
	Detrital shales and silts	P	P	C	P	-	P	-	-	R	-	R	-	-	-	-	-	P	-	R	-
	Nodules in basin shales	C	C	C	C	-	P	P	-	R	-	-	-	-	-	-	-	P	-	-	-
	Pelagic limestones	C	C	C	C	-	P	P	R	P	P	R	-	-	-	R	-	P	-	-	P
	'Reef' limestones	-	P	P	R	-	R	R	R	C	C	P	C	R	C	-	P	R	R	-	P

Fig. 4.8. Summary of the faunal & floral features of the 'off-reef' areas.
(C= common; P= present; R= rare; -= not seen)

containing minor pseudoplanktonic or epiplanktonic animals.

(b) a sparse indigenous death assemblage containing a "mixture" of basin forms (e.g. trilobites and rarely traces of siliceous sponges) and schwellen forms (e.g. crinoids, brachiopods and corals), the latter being especially noticeable in those limestones lying above or immediately adjacent to the massive limestones of the Bieber area.

(c) an exotic death assemblage, rich in crinoids and brachiopods found as intercalations within those nodular limestones lying downslope of massive fossiliferous limestones at the time of deposition.

Faunal relationships

The inferred relationships between different elements in the fauna of the "pelagic environment"* are shown diagramatically in Fig 4.9. Construction of a food-web such as this, can, at best give only an approximate model of the real system, since information on feeding habits of Devonian animals is unobtainable and recourse

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* The terms "pelagic facies" and "pelagic environment" refer to those sediments and that general environment where little or no apparent contribution was made by shallow-water deposits or by benthonic life-forms, generally considered to have lived in shallow marine environments. In the studied area such environmental conditions were present in areas away from the Bieber Schwelle in the Upper Givetian and Adorf, and throughout the region from then until the Lower Carboniferous.

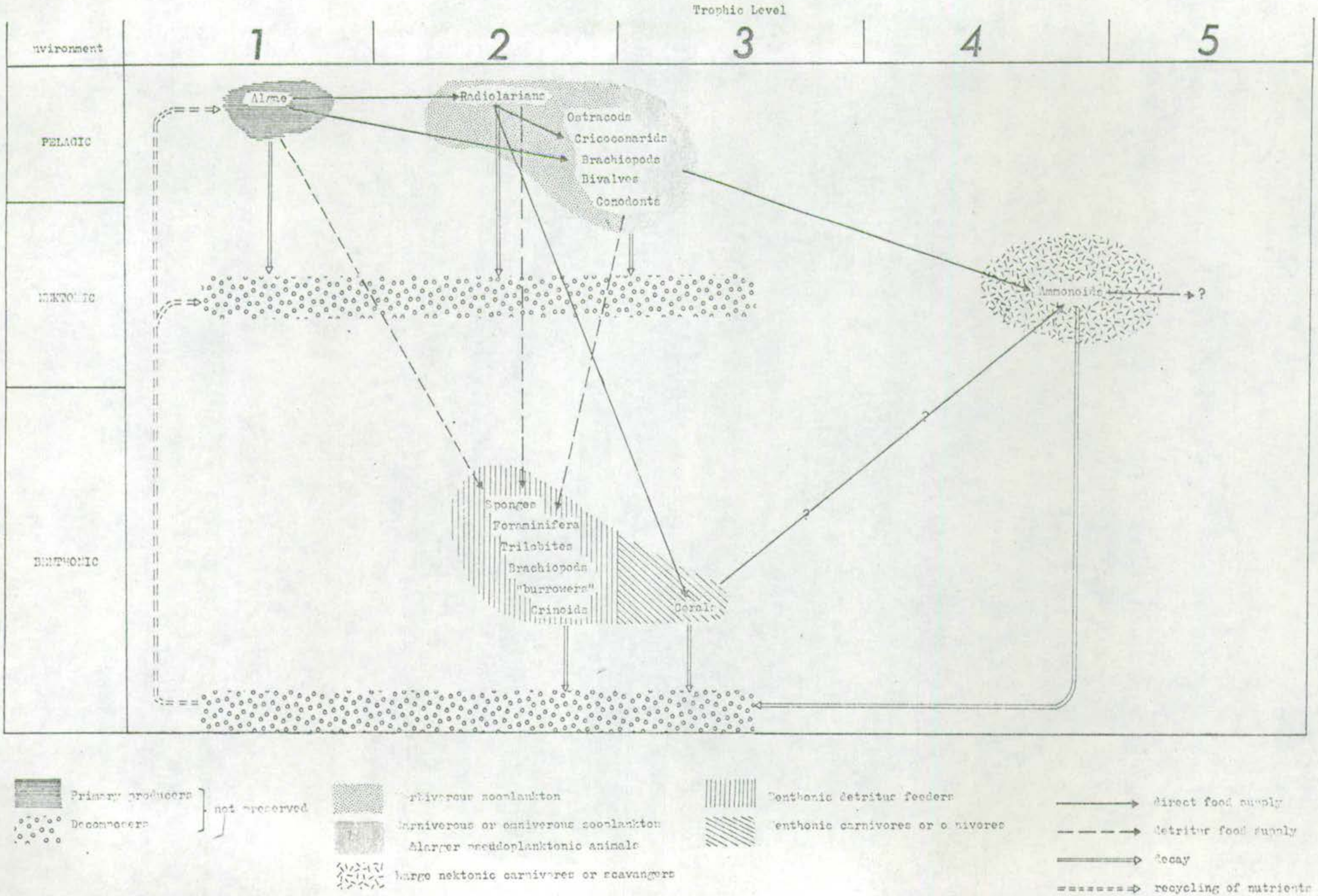


Fig 4.9. Hypothetical faunal reconstruction of "soft-reefs" food web.

must be made to Recent related forms, or, in the case of extinct groups, to generally accepted life-habits. Also, the preserved fauna almost certainly gives a very biased picture of the original biota. The presence of burrows in nearly all of the schwellen carbonates, for example, indicates the former importance of an infauna of which there is now no other record. Likewise there is no preserved record of "primary producers", though they must have existed to support the preserved fauna.

The food-web model presented in Fig 4.9 is therefore based on only that fraction of the living date, which has been preserved. It is thought unlikely, however, that the addition of "missing-links" in the food-chain would severely upset this model.

Post-mortem transport

With the exception of the exotic elements in the fauna, the lack of abrasion of fossils in general suggests relatively weak current transport on the sea floor, though winnowing of microfossils from the sediment may occasionally have taken place. All of the pelagic and nektonic fossils must have undergone vertical transportation to the sea floor prior to burial and it is probable that many of the larger forms (e.g. ammonoids and conodonts) underwent some lateral transport by surface drifting. By analogy with modern nautiloids and marine vertebrates, such lateral drifting may have amounted to several hundreds of miles. Even allowing for such drifting of individuals, it is not thought likely that this has severely upset the ecological distribution of animal groups in "pelagic" environments.

Environmental conditions

(1) Temperature + salinity: The presence of radiolarians in the basin areas and the prolific development of skeletal carbonates in the Bieber area suggests a tropical climate. All elements of the fauna are open-marine in nature.

(2) Bottom aeration: The importance of bioturbation in all but the Kellwasser limestones indicates that not only the bottom waters, but also the upper layers of the sediment, were generally fully oxygenated.

(3) Nature of substrate: Bioturbation within the sediments indicates that, for much of the time, they were relatively unconsolidated and probably much like the cocolith-foram oozes in the Oceans today. Substratum type had no effect on most of the preserved fauna, being pelagic or nektonic, and the preserved benthos indicates a soft bottom rather than hard. Hard-bottom encrusters have not been found in any of the studied limestones, though the increasing frequency of crinoids and brachiopods in shallow schwellen top limestones could possibly indicate an increase in formation of temporary hardgrounds in these areas.

(4) Water Depth: Little direct control is exerted over the distribution of marine life by water depth alone, but

such depth-controlled parameters as light intensity, salinity, temperature, dissolved oxygen concentration and food supply can have an important effect.

The majority of the calcareous mudstones are thought to have been formed at depths in excess of 50m for the following reasons:

(a) nodular limestones of Givetian and Adorf age exist as lateral equivalents of actively growing reefs, and those of later age lie on top of "dead" reefs. The calcareous mudstones, therefore, were formed at depths below the zone of reef growth, probably at depths in excess of 50 or 70m.

(b) benthonic algae have not been found in any of the limestones examined, but are common in the Adorf limestones around Bieber. Their absence is taken to indicate deposition below the photic zone, in depths greater than 50-100m, for most of the nodular limestones. Isolated occurrences of red algae have been reported from schwellen limestones elsewhere in the Rheinisches Schiefergebirge, indicating that locally perhaps these limestones may have been deposited at even shallower depths (Tucker, 1973). The presence of red-algae, however, is, in itself, no guarantee of shallow-water origin (Playford and Cockbain, 1969).

A lower depth limit is more difficult to assess. Estimates of maximum depth of deposition of the nodular limestones in the lithologically and faunally similar Ammonitico Rosso range from a few hundred metres (Hallam, 1967; Wendt, 1969; Jenkyns, 1970) to several thousand metres (Trümpy, 1960; Garrison and Fischer, 1969). All depth estimates have hinged on interpretation of the "oceanic" nature of the limestone fauna, with its sparse benthos and rich plankton and nekton. It is, therefore, necessary to establish the reasons for the virtual exclusion of benthos, before one can give any depth values to the sediments.

In the present marine environment a benthonic density (though not diversity) gradient exists between the shallow continental shelf areas, where benthos are plentiful, and the deep ocean areas, where benthos are scarce (Hessler and Sanders, 1967). The main factor influencing the amount of benthonic life is not depth alone, however, but food availability. Benthonic animals are most abundant in shallow continental-shelf conditions for the following reasons:

(a) the concentration of nutrients supplied by rivers from the land is highest in shallow waters bordering the landmass.

(b) primary production by benthonic algae on which much of the zoobenthos feed is possible only in water depths shallow enough to allow light penetration to the sediment surface for photosynthesis.

(c) the zoobenthos live in the same environment as phytobenthos, thus little energy wastage is met with to consumer from producer, either through decomposition or passage through several food-chain links.

The sparse nature of the benthonic fauna in the Devonian schwellen limestones can therefore be explained by the following:

(a) they were deposited in an area distant from the continent. Land derived nutrient supply was, therefore, negligible.

(b) they were deposited, for the main part, below the photic zone. Primary production was, therefore, limited to phytoplankton.

(c) the benthonic animals did not live in the environment where primary production was taking place. Consequently with increasing depth, the consumers became increasingly separated both spatially and temporally from the primary food supply. With increasing depth the feeding characteristics of the benthos possibly also changed. As most of the food drops from the water column above, detritus feeders would predominate, and deposit feeders would be at an advantage over suspension feeders, in being able to obtain concentrated food supplies from the sediments or sediment surface.

Great depths do not have to be resorted to to explain the nature of the fauna of the pelagic limestones, therefore, since food supply is the overall control. Certainly selective dissolution of aragonite does not appear important (as it is in the Jurassic nodular limestones) as ammonoid shells are found in nodules in the basin shales and are not restricted to schwellen sediments.

The presence of arenaceous foraminifera (mainly Tolypammina sp.) in both the Jurassic Alpine nodular limestones and Devonian limestones of Germany has been taken as indicative of depths of less than 200m (Wendt, 1969; Tucker, 1973). Modern Astrorhizidae and Ammodiscidae have their maximum development in deep oceanic sediments, largely due to the exclusion of calcareous foraminifera below the Compensation Depth (Murray, 1972). The apparent contradiction in these results emphasizes the difficulties (and dangers) in extrapolating modern ecological distributions into the past. It is probable that in the Devonian, these agglutinated foraminifera occupied a much wider range of environments, being subsequently displaced by the development of calcareous foraminifera in the late Palaeozoic and Mesozoic.

CHAPTER 5

Massive limestones

In this chapter, a range of limestone types are considered which, though showing a wide range of lithological variation, are nevertheless sufficiently distinctive both in their lithological, faunal and floral characteristics and in their mode of origin, to warrant independent consideration. The major differences between the nodular schwellen limestones already discussed and those dealt with in this chapter are summarized below.

	Nodular limestones	Massive limestones
Grain size	Microsparites	Mainly sparites or sparudites
Insoluble residues	high levels (5-40% by weight)	low levels (0-8% by weight)
Texture	mud supported allochems	generally grain supported allochems
Fauna+flora	dominantly pelagic	dominantly benthonic
Gross characteristics	poorly bedded-nodular	thinly-bedded to massive
Colour	generally grey, yellow or red	generally grey to black.

The richness and diversity of the mainly benthonic fauna and flora, and the absence of fine detritus in grain-supported sediments, suggest that limestones of this type were deposited in environments where current energy was much greater than was the case in those Schwellen areas where nodular limestones developed. It is thus assumed that the limestones dealt with here were deposited on submarine rises whose summits were shallow enough to allow a prolific benthonic development and on which current strengths were strong enough to keep the majority of the sediments relatively free from fine detritus.

Geographical distribution

In the studied area, these massive carbonates are restricted entirely to the volcanic rise area between Königsberg, Bieber and Waldgirmes (Fig 1.5). To the north and west of Königsberg and to the south of Waldgirmes, they are replaced by time equivalent pelagic nodular limestones and siliceous shales, and to the east of Bieber by shales and greywackes (Fig.4.1). The massive limestones, therefore, appear to be virtually surrounded by sediments considered to have been deposited in deeper waters.

No massive limestones occur in the Erda-Weipoltshausen-Altenvera schwellen area. (Fig 1.5).

Stratigraphical distribution

Over most of the area, complete sequences have not been preserved, but where the base of the limestone is exposed against the Givetian volcanics (though generally a tectonic boundary), conodont faunas of Upper Givetian age have been obtained. Limestone fragments within the volcanic tuffs on the summit of the Schwelle also yielded conodonts of this age (Appendix III).

The top of the sequence is rarely exposed. Erosion has removed the overlying sediments over much of the area, cutting well down into the massive limestones. Only at Eberstein is a complete sequence exposed, and here the change from massive to nodular pelagic limestones occurs in the P. gigas zone.

The majority of the massive limestones are, therefore, of upper Givetian or lower Adorf age. Younger sediments are, however, included within fissures and caverns within the limestones and these are dealt with in this chapter, for convenience, regardless of rock type. Fragments of limestone of Carboniferous age are also dealt with in this section, since they shed considerable light on the late evolution of the schwellen area.

Local correlation

Correlation between sections and hand specimens within these limestones presents more problems than are presented by nodular limestones for the following reasons:

(i) rapid facies variations are more pronounced in the massive limestones and lithological correlation cannot be applied successfully over most of the sequence.

(ii) conodonts are rare in these limestones, and those found are generally conservative forms of limited stratigraphical value.

(iii) in contrast to the nodular limestones, great thicknesses of massive limestones were deposited during one conodont zone, and so even with perfect conodont chronology, correlation within zones is impossible.

Macrofossils are common, but generally of little stratigraphical value, either because they are long ranging forms, or because discrimination of accurate ages depends on recognition of specific characters difficult to distinguish because of poor preservation.

I Facies types in the massive limestones

Massive limestones of Middle and Upper Devonian age in the eastern Rheinisches Schiefergebirge are collectively known as "Massenkalk" (Paeckelmann, 1922). Over much of the Rheinisches Schiefergebirge this "Massenkalk" sequence can be subdivided into three major lithostratigraphic units: the Schwelm, Dorp and Iberg Facies (Krebs, 1968b).* These three facies units are recognized also in the Rodheim-Bieber area, with the addition of one other facies - herein referred to as the Oolitic Facies, which although temporally divorced from the Devonian "Massenkalk" is dealt with in this chapter for convenience as it too was developed on the Rodheim-Bieber Schwelle, but at a later date (CuIIβ/γ). The main features of these Facies are tabulated below:

Facies type	Age	Lithological character in the Rodheim-Bieber area
Oolitic	Lower Carboniferous	Light grey oosparites-oomicrites. Not seen in situ but only as fragments in schwellen-slope breccia.
Iberg	Middle Adorf	Light grey bedded calcarenites, developed locally over Dorp limestones and grading into nodular pelagic limestones.

* The names Schwelm, Eskesberg, Dorp and Iberg were originally applied to biostratigraphic divisions of the "Massenkalk" in the Bergish Land by Paeckelmann (1922). More recent work, however, has shown that these divisions are not biostratigraphic but lithostratigraphic units, and that the Eskesberg and Dorp limestones are lithologically indistinguishable. The term "Eskesberg" therefore has been abandoned in reapplying these terms as names of major lithofacies units by Krebs (op. cit.).

Facies type	Age	Lithological character in the Rodheim-Bieber area
Dorp	Lower to Middle Adorf	Mainly light grey massive to bedded calcirudites, calcarenites and calcilutites. Rapid lateral and vertical variation in lithology.
Schwelm	Upper Givetian	Dark grey to black generally massive calcilutites, calcarenites and calciridites with scattered algal-stromatoporoid boundstones.

It must be stressed, however, that many of the characteristics of one Facies may be, at least partially, replicated in subfacies of another (e.g. many features of the Schwelm Facies are found in certain lithologies in the Dorp Facies, as are many of the limestones of the Iberg Facies).

Schwelm Facies

Rocks belonging to this Facies are all of Givetian age and are best exposed in the southern part of the quarry at Eberstein (Fig X, Appendix I) where their dark colour, generally fine grain size and "splintery" nature distinguish them from the light to dark grey bioclastic limestones of the overlying Dorp Facies.

The base of the sequence is exposed only immediately south of Eberstein (Fig 2.1a), but here it is a thrust contact, the limestones adjacent to the overthrust volcanics being deformed, and little of the primary depositional fabric can be distinguished. Because no limestone is exposed between this locality and outcrops bordering the quarry at Eberstein, the account given here is based mainly on lithologies developed in the upper half of the Schwelm sequence though, in hand specimen at least, the rocks in the upper part of the sequence appear to be similar to those seen near the base.

Three major lithological groups are recognized within the sequence, though many individual specimens do, of course, show features transitional from one group to another:

(a) Stromatoporoid-Amphipora intrasparites and micrites.

(b) Dense poorly fossiliferous calcilutites.

(c) Poorly fossiliferous pelsparites and pelmicrites.

(a) Stromatoporoid-Amphipora intrasparites and micrites

These limestones are easily distinguishable from those of the other two groups by the rich fauna and flora contained within them. In the exposed sequence of Schwelm limestones at Eberstein, limestones of this type are most important in the lower part of the sequence being replaced in the upper part by limestones of types (b) and (c).

The faunal composition varies considerably from place to place in the sequence. In some areas the sediments are composed entirely of bulbous and elongate massive in-situ stromatoporoids (Fig 5.1a) commonly reaching heights in excess of 15cm. Stromatolitic micrite coatings are common on the surfaces of these stromatoporoids. The sediment between these growth colonies is dominated by algal degraded fragments of Amphipora, Stachyodes and by abundant gastropods ("Turbinopsis" and "Murchisonia" types) (Fig 5.1c).

Microscopically the matrix of these rocks ranges in composition from "dark bituminous" intrasparites to dismicrites, the intraclasts frequently showing an elongate "whispy" structure. These intraclasts are in most cases probably algal grains, as Girvanella threads have been noted in some of them. The most common faunal constituents of the matrix are ostracods (normally articulated), foraminifera (Bisphaera, Parathurammina, Earlandia sp.) fragments of dasycladacean algae (? Vermiporella sp.) and calcispheres (Fig 5.1b).

Fig 5.1

(Scale bar represents 1 cm. unless otherwise indicated)

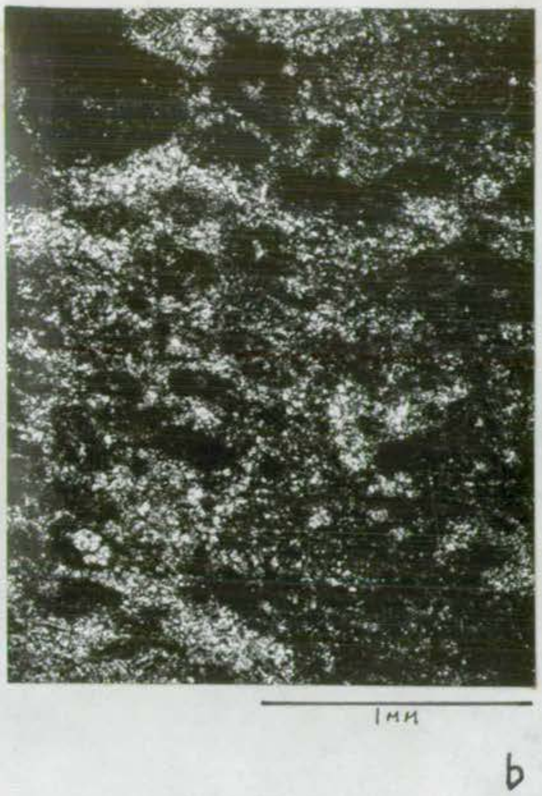
(a) Stromatoporoid-Amphipora subfacies. Recrystallized massive stromatoporoid, with well developed latilaminae, underlying (?) algal intrasparites similar to (c) below. Limestone cut by near-vertical fibrous calcite filled fissure. Hand specimen. Sample no. E15.

(b) Stromatoporoid-Amphipora subfacies. Dark pelleted matrix consists mainly of forams, ostracods, and poorly preserved algal traces. Recognizable macroskeletal debris is rarely present. Thin section. Plane polarized light. Sample no. E18. ~~Negative print.~~

(c) Stromatoporoid-Amphipora subfacies. Large calcite filled gastropod moulds (right), algal coated and degraded Amphipora and Stachyodes stems, and smaller gastropods in a partially recrystallised micrite/intrasparite matrix. Hand specimen. Sample no. E15.

(d) Dismicrite with scattered Amphipora stems and thin-shelled (?)brachiopods(left). Note blackened (algal degraded?) edges of Amphipora, and elongate encrusting(?) colony at base of specimen. Hand specimen. Sample no. E18.


a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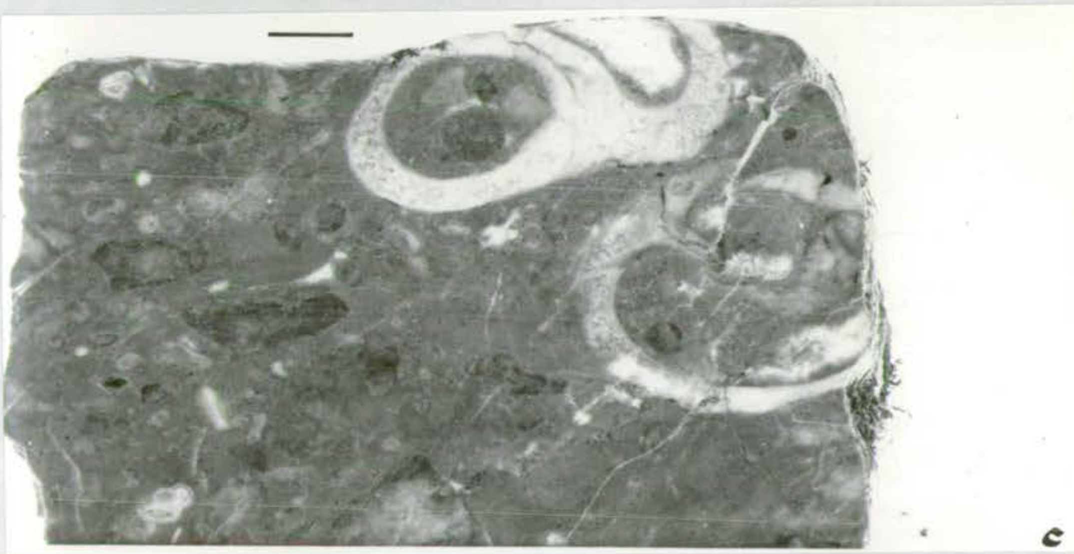
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d

In the upper part of this subfacies massive stromatoporoids decrease in importance and are replaced by dense *Amphipora* colonies in a fine dismicrite matrix lacking the coarse intraclasts and algal degraded colonies seen lower in the sequence (Fig 5.1d).

Discussion

Several features shown by the limestones discussed above are considered particularly important in discussing their origin.

(i) The abundance of algae and algal products suggests that they were formed in water depths less than 50 metres and probably less than 20 metres (Swinchatt, 1969).

(ii) The common occurrence of massive bulbous and elongate stromatoporoids, provides strong evidence for formation in the turbulent water zone above wave base.*

(iii) The lack of orientation and poor sorting shown by allochems, and the presence of a fine grained matrix on the other hand, suggests deposition in quiet-water areas where fine sediment would not be winnowed by current action. Because of the abundance of algal traces in the matrix of these poorly-sorted rocks, however, it is likely that allochems were stabilized by algal mats, preventing their reworking and removal, even in high energy environments.

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* See page 190 for full reference list.

(iv) The restricted nature of the fauna, with organisms normally only found in back-reef environments in the overlying Dorp Facies (e.g. Amphipora, Parathurammina) apart from massive stromatoporoids being important constituents. Skeletons typical of fore-reef environments in the Dorp Facies are not found in limestones of the Schwelm Facies, suggesting that environmental conditions were, in many respects, similar to those in the back-reef areas of the Dorp complex.

(b) Dense poorly fossiliferous calcilutites

Rocks of this type are extensively developed in the upper part of the Schwelm sequence between the fossiliferous rocks discussed above and the basal skeletal calcarenites of the overlying Dorp Facies.

They are unbedded and in hand specimen in places have a conspicuous mottled appearance (Fig 5.2a) which may suggest that these are massive "organic" structures of some kind, especially as subspherical "lumps" of similar massive featureless calcilulite are seen in the type (a) limestones below. Microscopically, however, no trace of any organic structure can be seen, all traces of primary texture having been lost through neomorphism. In the upper parts of this sequence a few scattered Thamnopora fragments and rare traces of gastropods were noted.

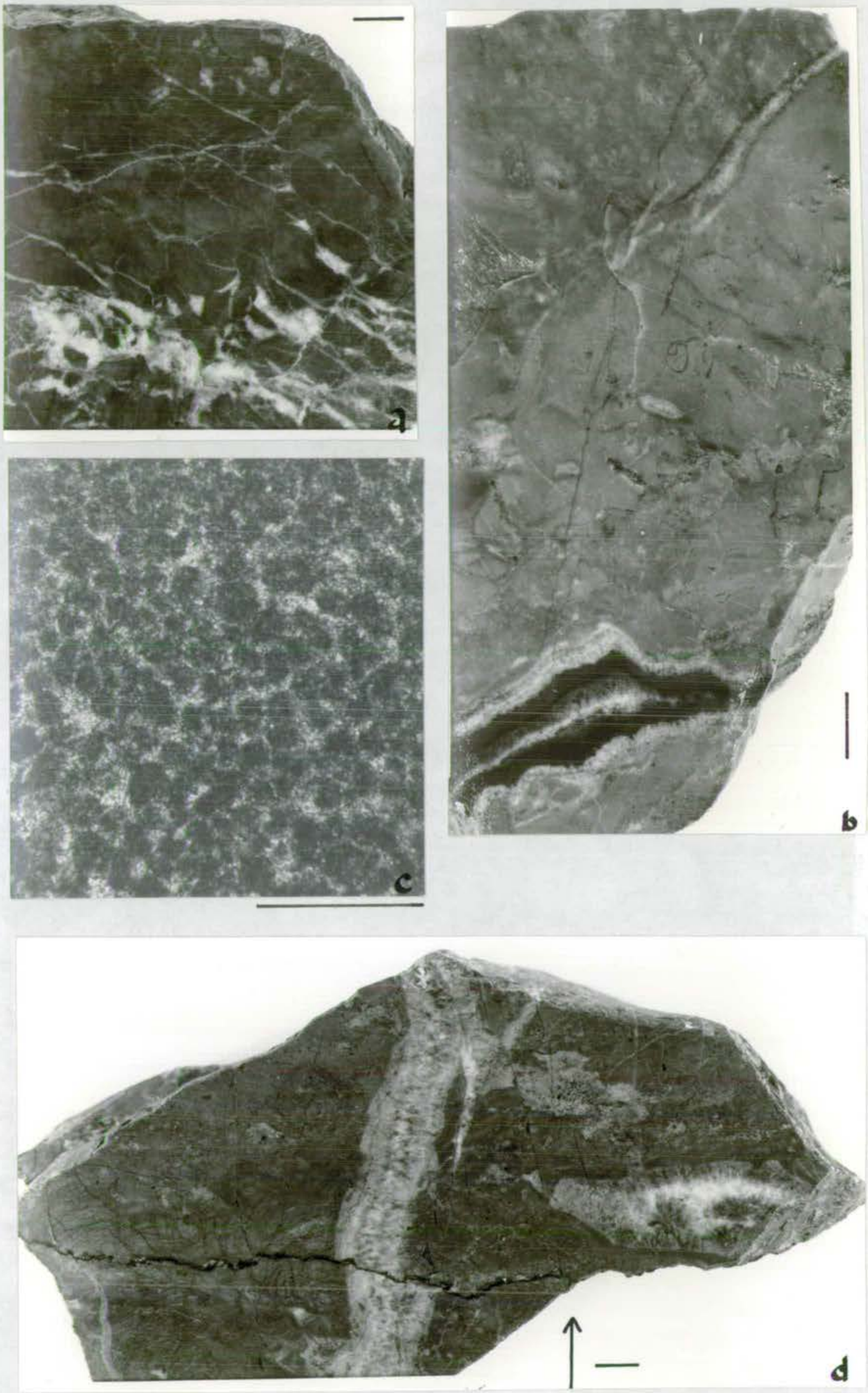
Fig 5.2

(a) Dense calcilutite subfacies. Light and dark mottled unfossiliferous microsparites are typical of this group. Hand specimen. Scale bar = 1cm. Sample no. E22.

(b) Pelsparite subfacies passing up into dense calcilutite subfacies. Large sheet cracks and small horizontal elongate fenestral voids (lower right). Spar filled mould of gastropod (upper centre) is the only fossil present. Hand specimen. Scale bar = 1cm. Sample no. E23.

(c) Pelsparite, (detail of above). Dense well sorted ovoid pellets in sparite cement. Thin section, plane polarized light. Scale bar = 1mm.

(d) Massive multidomed colony of Sphaerocodium and Girvanella overlain by pelsparite. Colony cut by vertical fibrous calcite filled fissure. Large "bedding-parallel" cavity on right appears to be a primary feature. Hand specimen. Scale bar = 1cm. Sample no. E34.


a
b
c
d

Discussion

Because of the widespread recrystallization which has affected these rocks, little information remains which might shed some light on the environment of deposition of these rocks. Their massive character in the field suggests that they are not primarily organic growth colonies since the rocks are a completely homogenous thick mass. It is thought most probable that these rocks were deposited as lime mud sediments in areas where environmental conditions excluded the faunal elements seen in the underlying fossiliferous limestones.

(c) Poorly fossiliferous pelsparites-pelmicrites

Rocks of this type are indistinguishable texturally in hand specimen from those of type (b) above and occur as lenses within them near the top of the Schwelm sequences at Eberstein. They are always lighter in colour, however, than the dense calcilutites.

Macroscopic bedding is not developed though microscopically aligned sheet cracks and fenestral voids occur, and bands of pelsparite and pelmicrite, resulting from differences in pellet grain size and degree of compaction, alternate (Fig 5.2b). Gastropods are virtually the only fossils present, though rare Amphipora fragments have been noted. In the upper part of this sequence bulbous growths of Sphaerocodium and Girvanella occur on a pelleted sediment base (Fig 5.2d).

Most of the limestones are composed of well sorted pellets (Fig 5.2c), few other allochems being present.

Discussion

Because of their regular ovoid shape, the absence of any internal structure and the association with gastropods, the pellets are interpreted as being of faecal origin. The occurrence of domed algal growths within this sequence suggests that these sediments were formed in shallow waters, and the common development of parallel sheet cracks may indicate that at times these sediments were exposed subaerially, allowing the sediment to dry out, as do lithologically similar sediments in the Bahamas today (Kornicker and Purdy, 1957).

In addition to the exposures noted at Eberstein, rocks of the Schwelm Facies crop out in places on the Bieber Schwelle. Nowhere else, however, does the thickness of the sequence approach that seen at Eberstein. In the central schwellen region (around Dicke Eiche and Haina), the Schwelm Facies, when seen, tends to be lithologically very like the overlying back-reef limestones of the Dorp Facies and appears to grade upwards into them.

Summary of Schwelm Facies deposition

Because rocks of this Facies are well exposed only at Eberstein, the spatial relationships between individual lithological groups could not be assessed.

The faunal, floral and sedimentological characteristics of the sediments discussed above suggest

(i) that the sediments were deposited in shallow water conditions, probably above normal wave base and possibly, at times, above low water.

(ii) that during their deposition, many of the sediments may have been stabilized by algae. Stromatolitic crusts are certainly developed on massive stromatoporoids.

(iii) that the sediments were deposited over a wide range of environments, ranging from areas where skeletal organisms were apparently excluded (Dense calcilutites) through areas able to support a meagre faunal (pelsparites) to areas with a rich fauna and flora (stromatoporoid-*Amphipora* intrasparites).

The severely restricted nature of the fauna of the dense calcilutes and pelsparites may be partially due to the fine grain size of these sediments, but is thought to be largely caused by fluctuations from subtidal to intertidal conditions within the areas in which these sediments were deposited. The fauna and flora of the stromatoporoid-Amphipora intrasparites, though rich, is restricted in its diversity. Once again, many organisms commonly found in the overlying Dorp limestones are not found in this subfacies presumably because their development is inhibited by some environmental limiting factor. This factor cannot have been food availability due to the abundance of algae present. Nor could it have been depth since these sediments formed in very shallow waters. It is concluded therefore that either increased temperature or salinity, or temporary sub-aerial exposure were the most probable factors controlling the biotic composition of these sediments.

It is therefore suggested that the Schwelm limestones were deposited on intertidal to subtidal "flats" developed on the Givetian volcanic summit and that varying hydrographic conditions on various parts of these flats gave rise to coarse fossiliferous carbonates with scattered stromatoporoid growths in the more exposed areas, while fine calcilutites and pelleted muds were formed in quieter more restricted embayments.

Limestones showing somewhat similar characteristics are developed in the overlying Dorp Facies, where the field associations and petrographic characteristics suggest that they were formed in similar environments to those in which it is suggested the Schwelm limestones were deposited.

Dorp Facies

Rocks belonging to this Facies group are best developed in a crescentic zone between Königsberg, Eberstein and Bieber. Scattered outcrops occur also in the area between Rotenberg, Dicke Eiche and Rotestrauch, but here the Dorp limestones are thinner than those in the main outcrop area. (Maps 1+2, Appendix I).

All the limestones in this group are of Lower Adorf age (mainly P. asymmetricus zone). The base of the sequence is exposed only at Eberstein where it appears to be transitional to the underlying Schwelm limestones. The base has been taken at that point in the sequence marked by the influx of light grey crinoidal-brachiopod calcarenites (page 153). The top of the sequence is well marked at Eberstein, where the upper few metres of limestone are riddled with sediemnt filled caverns, but is not exposed elsewhere.

A wide range of laterally impersistent lithological groups can be recognized in the Dorp Facies suggesting that at this time on the Bieber Schwelle, environmental conditions varied considerably from one area to another. In view of the fact that the majority of the limestones in the main outcrop area (Königsberg-Bieber) consist mainly of skeletons and skeletal debris of large colonial organisms, and that those of the central schwellen area (Rotenberg, Dicke Eiche, Haina) contain few if any colonial organisms, it is thought likely that the diversification of environmental conditions was caused by the establishment of an organically constructed "barrier" or reef on the northern and eastern edges of the schwellen area.

The major subfacies recognized in this study are listed in Fig 5.3 below. Their skeletal and non skeletal characteristics are compared in Fig 5.21.*

The subfacies have, as far as is possible been shown, more or less, in correct sequence relative to the "reef-margin" zone. In this study a distinction has been made between sediments of the fore reef area and sediments of the back-reef area. These two groups are almost symmetrically arranged about the reef margin area (characterized by large massive in-situ colonial growth complexes), both showing a progressive decrease in grain size away from this marginal area (Fig 5.3), the former grading into deeper water off-reef pelagic limestones, the latter into fine non-skeletal sands and muds of extreme lagoonal type, as the influence of the reef wanes with distance from it, and local environmental conditions increase in importance.

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* Subfacies distinctions are based mainly on large scale faunal, floral and sedimentary features, since the matrix, especially in the fore reef area, is commonly similar in limestones of different subfacies. For this reason, polished rock slabs have been used to illustrate subfacies types in preference to thin sections in many cases, since the essential differences between subfacies types cannot be shown on the small area of a thin section.

(a) Subfacies type	Interpreted Environment	(b) Krebs 1966	Krebs 1969	Krebs 1971	Franke 1973
Skeletal biosparites	Fore reef	K?		6/7?	C
Stachyodes-Thamnopora biosparudites (type1)		I		3	G
" " (type2)					
Alveolites-stromatoporoid biosparudites (type 1)		H/I		4	F
" " (type 2)		I			D
" " (type 3)		H		2	E?
Stromatoporoid biolithite	Reef margin	F			H,I,K
Skeletal biosparudites (type 1)	Back reef	E	VI		L,M,N
" " (type 2)			VII		O
" " (type 3)		A	XI		Rin part
Laminites (type 1)		C	I		P
" (type 2)			II		
Non skeletal limestones (type 1)		B	IX		Q?
" " (type 2)			VIII		Rin part
Bedded dolostones			?		

Fig 5.3 (a) Subfacies types in the Dorp Facies, their inferred environments of deposition. (Vertical bars I, II and III indicate respectively (a) facies range over which massive or laminar colonial skeletons are common, (b) facies range over which dendroid colonial skeletons are common, and (c) facies range dominated by fine skeletal or non skeletal debris.

(b) Comparison between Dorp subfacies described here and those described from Devonian reefs elsewhere in the Rheinisches Schiefergebirge.

Sediments of the fore reef area

(1) Skeletal biosparites

Limestones belonging to this group were found at Strohmuhle and Eberstein. In the latter locality, where field relations could be examined, they lie directly on Schwelm limestones and are, to some extent, transitional to them. They are overlain by Alveolites-stromatoporoid biosparudites (Fig X, Appendix I).

The most distinctive feature of sediments of this group is the general lack of large colonial skeletons. In most cases brachiopods (both whole shells and broken fragments) are the only commonly recognizable macrofossils apart from scattered *Thamnopora* stems (often algal coated) (Fig 5.5a). As the contact with the overlying sediment group is approached, however, small flat lying *Chaetetes* and encrusting *Alveolites* colonies and *Thamnopora* stems become common (Fig 5.5b).

The matrix of these rocks is a moderately well sorted crinoidal biosparite containing fragments of fenestellid bryozoans, ostracods, gastropods, foraminiferids (normally *Bisphaera*) and small micritic intraclasts common to the matrix of the majority of the detrital massive limestones (Fig 5.5c).

The main features of this group are summarized in Fig 5.4.

	Skeletal	Stachyodes - Thamnopora biosparudite		Alveolites - stromatoporoid biosparudite			
	1) biosparudite	2)Type 1	3)Type 2	4)Type 1	5)Type 2	6)Type 3	7)Biolithite
Colour	← light-dark grey →						
Bedding Type	← massive						
Skeletons: essential / subordinate / accessory	Crinoids + brachiopods / tabulate corals / bryozoans + algal coatings	Stachyodes / Thamnopora + Alveolites / Solitary rugose corals		Lamellar stromats+/Alveolites (normally in growth position) / Stachyodes + Thamnepora / Rugose corals		Stroms + Alveolites (never in growth position). Crinoids + Stachyodes + Thamnopora	Massive stroms. / Sphaerocodium(?) + brachiopods
Carbonate matrix	Finely comminuted skeletal + intraclasts	← finely comminuted skeletons-brachs. crinoids, bryozoans, intraclasts etc. →				Crinoid ossicles + (rarer) intraclasts + fine skeletal debris.	Rare matrix consists of finely comminuted skeletal debris + Renalcis fragments.
Insoluble residues	← 1-3% →						
Voids (Interskeletal)	—	voids occupy all interskeletal areas.	Voids - rare. Only shelter cavities beneath skeletons.	Voids - common to rare - all shelter cavities beneath skeletons.	Large geopetal floored voids between large colonial skeletons	—	rare
Cement	Sparry/fibrous calcite	fibrous calcite abundant	fibrous calcite, (sparry calcite in parts).	fibrous calcite (sparry calcite in parts).	fibrous calcite abundant	fibrous/sparry calcite. syntaxial overgrowths on crinoids	fibrous
Mean size range	+1 to -1Ø	-2 to -5Ø	-2 to 5Ø	-2 to -6Ø	-2 to -7Ø	-2 to -5Ø	-2 to -8Ø
Sorting	high - low	high - low	low	v. low*	v. low*	v. low	v. low*
Degree of grain orientation	generally low - large skeletons//bedding	very low		v. low in ground mass, but large colonial organisms show pronounced near horizontal alignment.		v. low	low - high
Internal fabric	Random. Fine skeletal debris	Random . Concentrations of coral-strom. stem fragments or stems.		Strong alignment shown by skeletons - suggests growth fabric.		Random - coarse debris in crinoidal matrix.	Growth complexes of stromatoporoids +/- algae, either insitu or disturbed.
Relations with other subfacies.	Forms transition from Schwelm Facies to 3	Occurs as elongate pockets in 3.	Interfingers with 4, 5 + . 6.	Interfingers with 2, 3, 5 + 7	Forms lenses within 3 + 4	Lense-like concentrations within 3 + 4	Massive lenses within 4 normally.

(* Sorting low in many cases because of insitu frame growth).

Fig. 5.4 Summary of main features of "fore-reef" Dorp limestones.

Fig 5.5

(Scale bar represents 1cm. unless otherwise indicated)

(a) Skeletal biosparite. Scattered broken brachiopod shells (and rare larger articulated shells with geopetal sediment infillings) and crinoid ossicles are typically the only recognizable macrofossils seen in the fine skeletal calcarenite matrix. Hand specimen. Sample no. E39.

(b) Skeletal biosparite. Large brachiopod shells (left), algal coated _Thamnopora_ stems (lower right), and in-situ _Chaetetes_ colony (overgrown by _Alveolites_) (centre right) in fine crinoidal biosparite matrix. Hand specimen. Sample no. E40.

(c) Skeletal biosparite. Transition to Stachyodes-Thamnopora biosparudites shown by the increased importance of solitary and dendroid rugose and tabulate corals (_Thamnopora_) and by the appearance of stromatoporoid fragments in a fine, partially dolomitised matrix. Hand specimen. Sample No. B46.

a

b

c

Discussion

The characteristic lack of "reef" organisms in these limestones suggests either that they were deposited in an area so remote from the reef-margin that coarse skeletal material never reached them or that no reef existed at the time of their deposition and thus no source of coarse skeletal material was available for coarse skeletal debris.

It is not considered likely that these limestones were deposited in the distal areas of the fore reef slopes (as were lithologically similar fore reef limestones described by Krebs (Microfacies 6 and 7, 1969) and Franke (facies C, 1973)) as they lie directly over Schwelm limestones showing "back-reef" shallow-water characteristics.

The restriction of limestones of this type to the junction between the Schwelm and Dorp Facies, is considered critical to any interpretation of their origin because at this time the reef-margin rim, which was to control sedimentation throughout the lower and middle Adorf, had not yet established itself. No source for coarse skeletal debris, therefore, existed at this time.

This subfacies, therefore, is best interpreted as a transition between the Schwelm and Dorp Facies, during which skeletal crinoidal sands were deposited over the Schwelm sediments prior to the establishment of a reef-rim, It is thought significant that crinoid (+/- brachiopod) rich sediments are generally developed immediately after major facies-breaks (see page 295) and may represent "pioneer" sediments developed during the initial periods of marine transgressions.

The degree of breakage of brachiopod shells indicates that these sediments underwent considerable movement, as does the high degree of large grain orientation and sorting. The depth of water in which these sediments were deposited cannot be accurately assessed, though the presence of algal crusts on some Thamnopora stems suggests that water depths were probably less than 20 metres or so.

The gradual appearance in higher rocks in the sequence of low lying encrusters (Chaetetes and Alveolites) and the upward transition to the Stachyodes-Thamnopora biosparites is thus interpreted as the record of gradual establishment of reef-conditions, rather than decrease in water depth.

Because of their field associations, rocks of this subfacies are probably more closely related to the "skeletal calcarenites" of Klovan (1964) than to the reef-distant fore reef deposits of Krebs (1971). Klovan also noted that crinoidal calcarenites poor in "reef-skeletons" tended to occur intimately associated with fore-reef and even back-reef sediments, an association he found difficult to explain but thought may have been the result of localized quiet-water regions within the reef and near-reef zones.

(2) Stachyodes-Thamnopora biosparudites

Limestones included within this class are exposed at Eberstein, Rehmuhle and Bieber (Maps 1 and 2, Appendix I). In addition a few limestones similar to Thamnopora-rich varieties seen at Bieber were noted southwest of Rotestrauch (Sample no. W10).

Two varieties of this group can be distinguished:

(a) Limestones consisting entirely of coarse skeletal debris, lacking any interskeletal detrital matrix (Type 1).

(b) Limestones containing coarse skeletal fragments in a finer skeletal calcareous matrix (Type 2)

The main features of these limestone-types are summarized in Fig 5.4.

(a) Type 1

A large variety of lithologies are represented within this group, ranging from limestones containing only Stachyodes, Thamnopora, Alveolites or solitary rugose corals, to limestones with all four components, mixed in varying proportions (Fig 5.6a-d). In all cases interskeletal voids are filled by fibrous calcite, any sediment present being restricted to localized interskeletal "traps" in the rock (Fig 5.6d).

In some cases, the size of the interskeletal voids is so great that it is difficult to imagine such rocks being composed of grain supported skeletal fragments (Fig 5.6a and d). Such sediments possibly represent cemented dendroid branching growth colonies, while others

Fig 5.6

(Scale bar represents 1cm. unless otherwise indicated).

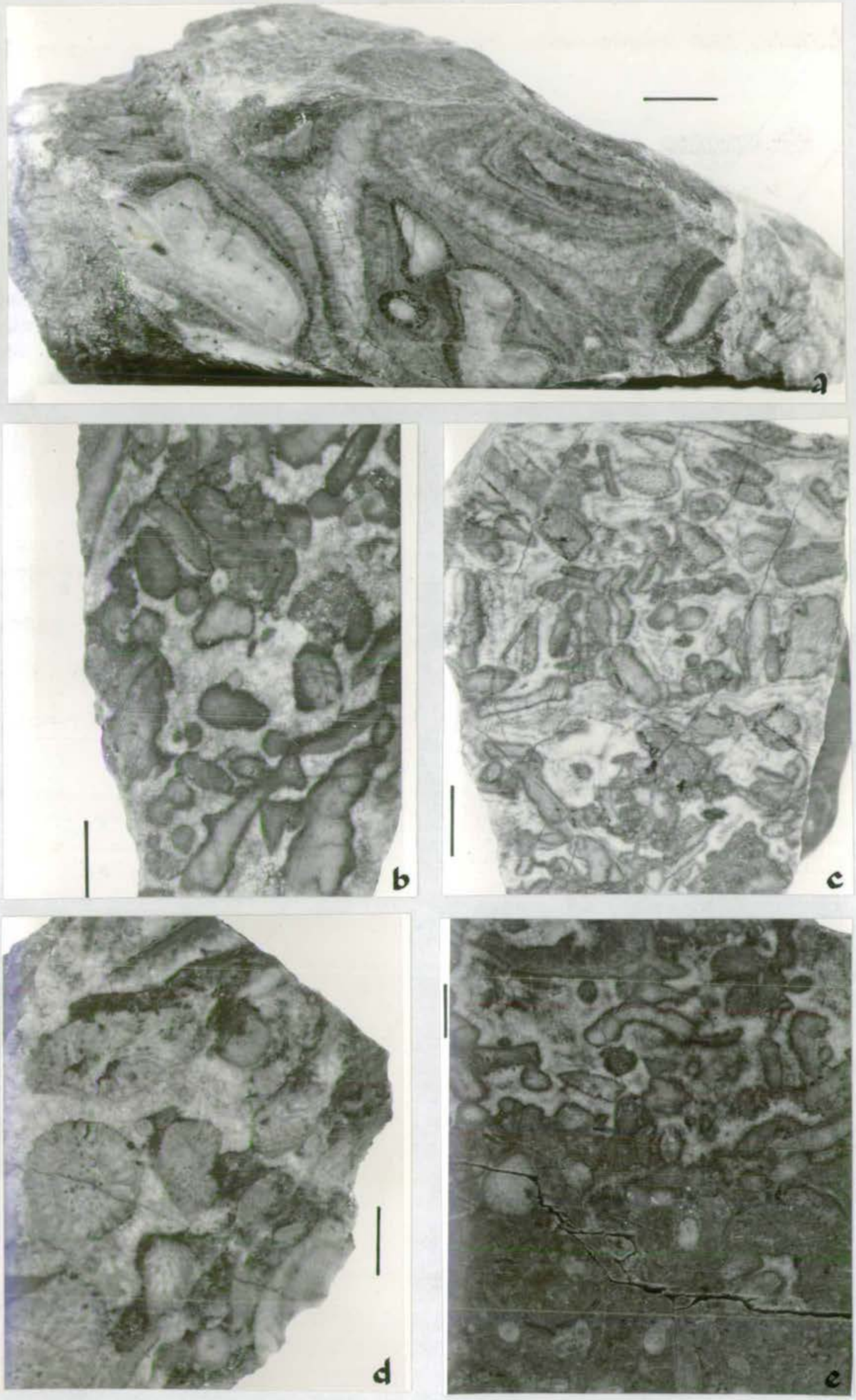
(a) Stachyodes stems in fibrous calcite cement. Stems do not appear grain supported but "floating" in spar, suggesting that they are connected in three-dimensions. Hand specimen. Sample no. B62.

(b) As above, but Stachyodes stems here are smaller, probably grain supported fragments. Edges of stems are blackened by algal(?) degradation or by solution. Small patches of sediment are trapped between stem fragments (upper centre). Hand specimen. Sample no. B36.

(c) Small Alveolites and Thamnopora fragments in fibrous calcite cement containing dark inclusion zones parallel to cavity walls. Hand specimen. Sample no. B4.

(d) Thamnopora stems (possibly growth colony) and solitary rugose coral (Disphyllum sp.) in fibrous calcite cement. Patches of fine skeletal detritus trapped between and above skeletons. Some subsequent solution is indicated by sharp vertical edge on sediment (arrowed). Note internal sediment deposited after fibrous calcite (upper centre). Hand specimen. Sample no. B3.

(e) Junction between Stachyodes-thamnopora biosparudite type 1 (above) and type (2) below. Only small shelter cavities are found in the latter sediment group. Grains in upper sparry zone are mainly Alveolites and Thamnopora, though Stachyodes and broken fragments of solitary rugose corals also occur. Hand specimen. Sample no. E56.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with more closely packed stems may represent concentrated gravel-like deposits of stromatoporoid and coral sticks (Fig 5.6b and c).

A conspicuous blackening of stem edges (marginal micritization) is seen in many limestones of this group (Fig 5.6b). Such blackening may have been caused by boring endolithic algae or by solution alteration (Ginsburg and Schroeder, 1973).

(b) Type 2

The macroskeletal components of these rocks are similar to those of type 1 but here interparticle voids are restricted to small geopetal shelter cavities beneath skeletons as rocks of this type, unlike those of type 1, contain appreciable amounts of detrital sediment between coarse skeletons (Fig 5.7a). The junction with the above described lithological group is normally sharp (Fig 5.6e).

Crinoid ossicles and brachiopod fragments are the only macrofossils, apart from those already mentioned, which can be readily seen in the matrix. They are normally little abraded and are associated with fragments of fenestellid bryozoans, trilobites and foraminifera. The greater part of the matrix, however, is not skeletal but consists of well sorted micritic (now largely microspar) intraclasts. Many of these appear to be badly preserved algal remains, some are stromatolitic and detrital crusts on microfossils ("woolly overcoats" of Toomey et al, 1970), and some may be sediment clots,

Fig 5.7

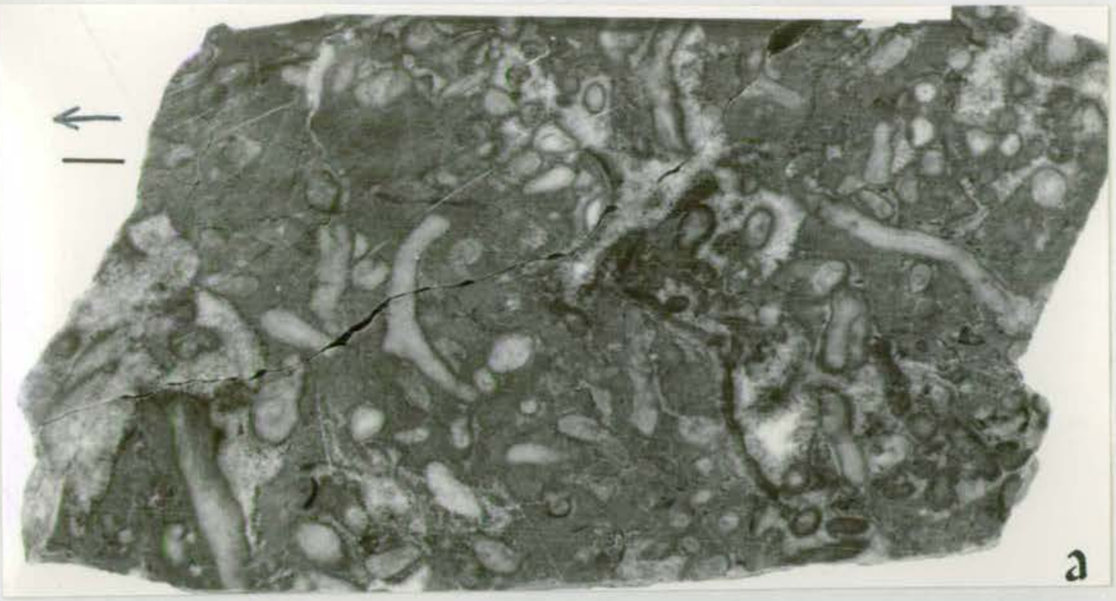
(Scale bar represents lcm. unless otherwise indicated)

(a) Stachyodes-Thamnopora biosparudite (type 2). Thin randomly orientated stems of Stachyodes and Thamnopora formingpoor poorly sorted skeletal gravels. Shelter cavities, now filled with fibrous calcite, present beneath many of the larger fragments. Matrix composed of broken skeletal debris and intraclast-like grains. Hand specimen. Sample no. E56.

(b) Stachyodes-Thamnopora biosparudite (type 2). Transition to Alveolites-stromatoporoid biosparudites(type 2) shown in upper left where large solution fragments , still retaining their original orientation, are surrounded by fibrous calcite cement.

Zone A below, characterized by the abundance of small bulbous and laminar stromatoporoids and Alveolites, transitional to type 1 Alveolites-stromatoporoid biosparudites.

The bulk of the limestone below the above mentioned zones consists of broken Stachyodes (and, more rarely, rugose and Thamnopora) stem fragments. Zone B consists of small closely packed fragments in a fine biodetrital matrix; Zone C of larger, less well sorted, fragments, which, due to their greater size, are more loosely packed and commonly roof shelter cavities filled with fibrous calcite. Hand specimen. Sample no. B33.



A

B

C


transported into their present depositional environment, or degraded pieces of skeletal debris (Wolf, 1965b; Braithwaite, 1966).

These limestones are texturally immature or submature well to poorly washed biosparites with little preferred skeletal orientation. Fragmentation of large skeletons is generally uncommon, as is abrasion.

Discussion

Some of the limestones described above and included in the type 1 variety, may be growth frames of dendroid stromatoporoids and tabulate or rugose corals, but the majority most likely were formed as concentrates of dislodged skeletal branches. The absence of sediment between branches suggests that these limestones were deposited in environments in which current strengths were strong enough to keep the skeletonal voids free of sediment, while those of type 2 were deposited in lower energy environments. The transition from the depositional conditions of type 1 to type 2 limestones could have been due to increased water depth, type 1 being deposited above, and type 2 below wave-base for example, or the environmental changes may have been more complex involving local variations in current strength along the reef-front, due to the presence of surge-channels or other such features.

Alternatively the absence of matrix in the type 1 limestones may have been exaggerated by post-depositional solution of the matrix, which appears to have been responsible for enlargement of void areas elsewhere in the fore-reef area (see below). The general closer packing of skeletal grains in type 1 limestones, and the common occurrence of monotypic skeletal rocks, which are never found in type 2 varieties, suggests that the essential differences between the two groups are due to primary depositional effects and not to post-depositional processes.

It is therefore concluded that the majority of these limestones were deposited as skeletal debris, at times in areas where currents were strong enough to wash-out the finer skeletal debris. In a few cases, however, small dendroid growth colonies, (not necessarily in situ) may be present. The marked variation in fossil-type between samples suggest that the skeletal fragments were derived from growth colonies close at hand, though other limestones, showing a high degree of skeletal mixing, often display a marked upwards decrease in skeletal grain-size, These are interpreted as skeletal talus fans deposited on the fore-reef slopes (Fig 5.7b).

At Eberstein, rocks of this type are separated from stromatoporoid biolithites (interpreted as reef-margin deposits) by the Alveolites-stromatoporoid biosparudites discussed below. It is therefore concluded that dendroid colonies either grew at a greater depth, or were deposited at greater depths farther from the reef-margin, dendroid skeletons being more easily dislodged and transported than low lying encrusters (Fig 5.14). The presence of high interskeletal porosity, however, suggests that the facies relationships may have been more complex and that Stachyodes-Thamnopora biosparudites could form in waters shallow enough for sediment winnowing to be accomplished.

(3) Alveolites-stromatoporoid biosparudites

Limestones of this subfacies are easily distinguished from those of other groups by the increased importance assumed by massive and lamellar colonial corals and stromatoporoids.

Three varieties are here distinguished:

(a) Limestones containing abundant apparently in-situ lamellar (and less frequently massive) corals and stromatoporoids, showing pronounced preferred orientation. Voids are restricted to local shelter cavities beneath skeletal sheets (type 1).

(b) Limestones in which the lamellar skeletons are partially unsupported by matrix, and appear to "float" in cement in places. The tendency for preferred skeletal orientation is still retained (type 2).

(c) Limestones containing disorientated apparently dislodged and perhaps reworked colonial skeletons (type 3).

The main features of each of these groups are summarized in Fig 5.4 and only brief descriptions are given below:

(a) Type 1

Lithological varients of this type are exposed at Eberstein, where they succeed Stachyodes-Thamnopora biosparudites in vertical sequence (Fig x, Appendix I). The transition from one subfacies to the other is marked by the appearance in the sequence of flat-lying lamellar stromatoporoid colonies, normally only a few centimetres thick encrusting a biodetrital matrix almost identical to that of the underlying Stachyodes-Thamnopora biosparudites. Complex intergrowths of Keega, algal (?) micrite and Renalcis are also found (Fig 5.8a).

Higher in the sequence the importance of stromatoporoids as growth masses is reduced as Alveolites increases in abundance and size (Fig 5.8b). This coral occurs either as flat-lying encrusting sheets or as hemispherical "heads" sometimes greater than 15cm. in diameter, the latter growth habit being more common higher in the sequence.

The matrix of these rocks is virtually identical to that of the underlying Stachyodes-Thamnopora biosparudites with fenestellid bryozoan fragments, crinoid ossicles, Renalcis fragments and non-skeletal (?) micritic intraclasts being the most important constituents.

At Bieber too, similar rocks are found. Encrusting stromatoporoids predominate in some rocks (Fig 5.8c) while, in others, Alveolites colonies are more important (Fig 5.9a). In both cases shelter cavities are found below branches of growth colonies, indicating that they were not wholly encrusting the sediment surface during growth.

Unlike Eberstein, where Alveolites dominated limestones overlie lamellar stromatoporoid rocks which, in turn, succeed Stachyodes-Thamnoporoid biosparudites, all rock types at Bieber are found in close proximity

Fig 5.8

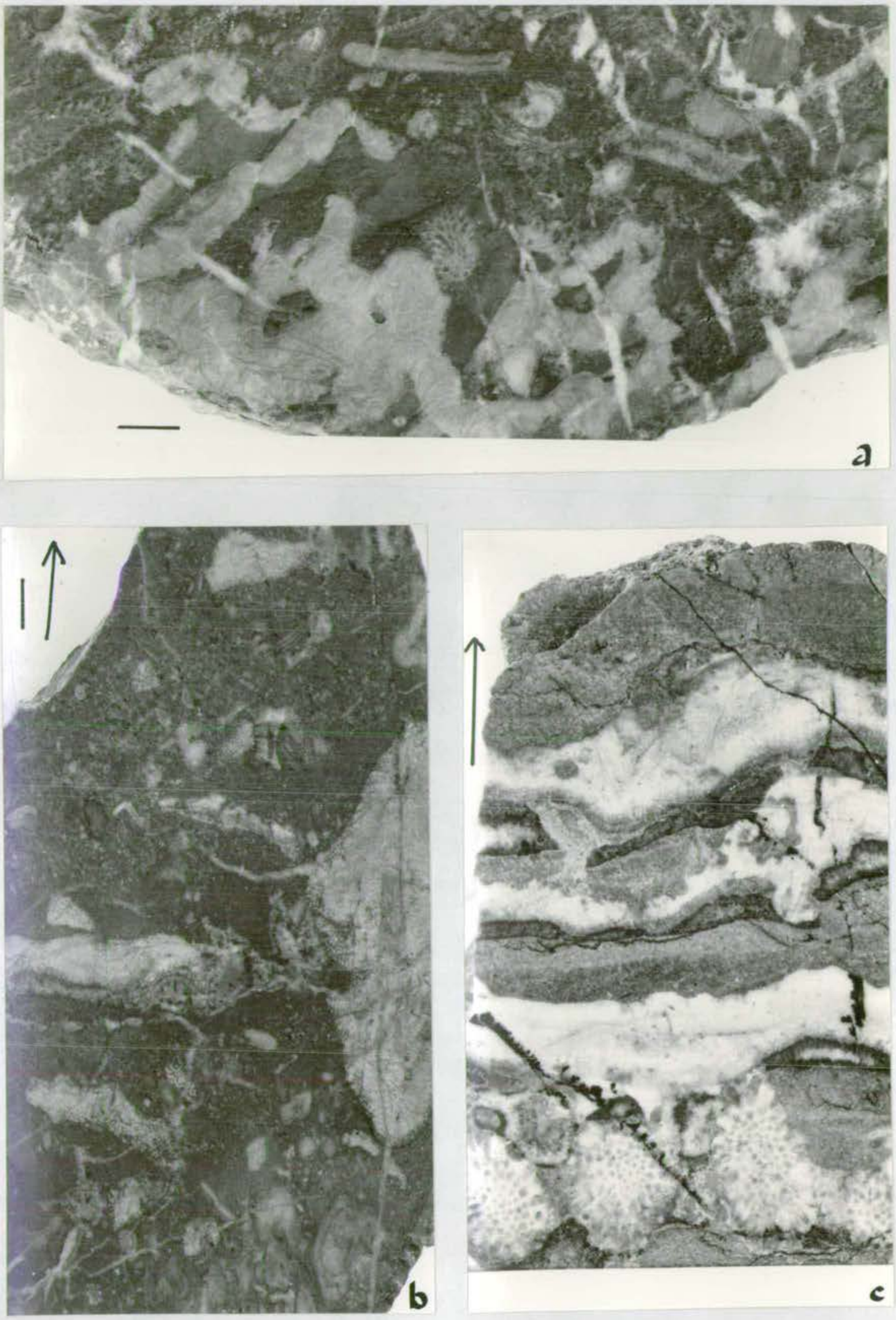
(Scale bar represents 1cm)

All specimens are of Alveolites-stromatoporoid biosparudites (type 1).

(a) Tangled growth colony of Keega encrusted and overlain by dense algal micrite (with Renalcis) in which scattered Thamnopora stems are trapped. Coarse skeletal matrix above contains fragments of Stachyodes and crinoids. Hand specimen. Sample no. E72.

(b) Massive (right centre) and tabular Alveolites colonies (various locations), apparently in-situ in coarse skeletal matrix similar to (a) above. Small upward facing colonies (top centre) apparently grew directly on sediment, but downward facing laminar colonies underlain by shelter cavities (centre left) filled by fibrous calcite must have grown freely projecting into the water column. Hand specimen. Sample no. E74.

(c) Horizontal branches of Keega and other stromatoporoid(?) branching at various levels from central stem (just out of picture to right). Shelter cavities present beneath all branches indicate that colonies did not hug sediment surface during growth. Upper surfaces of some branches intensively bored (centre). Note interconnected Thamnopora stems below Keega (bottom) lying on bedding. Lack of large skeletal fragments in matrix between branches is probably due to close sheeting of stromatoporoid colony restricting access of sediment. Hand specimen. Sample no. B42,


a
b
c

and interfinger with one another.

This group is also seen at Strohmuhle, between the two localities mentioned above.

Discussion

The high degree of orientation of large skeletons, none of which appear to be disturbed, indicates that these rocks represent in-situ growths of coral and stromatoporoid within a detrital matrix, which has either been trapped in a skeletal frame or encrusted by low lying growths (Fig 5.14).

In those sediments directly above the Stachyodes-Thamnopora biosparudites at Eberstein, the large skeletons are all low-lying encrusters dominated by stromatoporoids and upward growing crusts of Alveolites, but farther up the sequence, between these rocks and the Stromatoporoid biolithites (described below), the appearance of shelter cavities beneath colonial skeletons indicates that sediment trapping may not have been solely due to sediment encrusting, but that sediment may have been trapped in low lying dendroid, laminar or massive growth colonies capable of growth above the sediment surface. In some sediments the presence of lateral and downward skeletal growths (Fig 5.9a) and of vertically interconnected lamellar stromatoporoids (Fig 5.8c) suggests that many of the skeletal sections seen were interconnected in three dimensions.

Fig 5.9

(a) Alveolites-stromatoporoid biosparudite (type 1). Large in-situ(?) Alveolites colony (left) underlain by shelter cavities, as are near horizontally aligned Alveolites and Keega skeletons (centre and left). Matrix consists of fine skeletal debris with fragments of Stachyodes. The delicacy of some colonial skeletons (e.g. Alveolites (lower centre) and their lateral, and often downwards orientation indicates that they are in-situ sections of a larger framework. Hand specimen, Scale bar = 1cm. Sample no. B10.

(b) Alveolites-stromatoporoid biosparudite (type 1) (type 2 transition). Lower zone crowded with low angled thin in-situ Alveolites and Keega branched in fine detrital matrix in upper part of rock. Hand specimen, scale bar = 1cm. Sample no. B11.

(c) Alveolites-stromatoporoid biosparudite (type 2). Lamellar Keega colony roofing cavity. Skeletal debris matrix trapped on cavity bottom formed by algal encrusted Stachyodes and in "traps" on upward directed stromatoporoid branches (left). Matrix consists of algal degraded grains Renalcis fragments, forams and broken shell material typical of matrix in all rocks of this facies. Thin section, plane polarized light. Scale bar = 1mm. Sample no. B32.


a


b


c

(b) Type 2

Rocks of this type were observed only at Bieber quarry. They appear to be missing from the sequence at Eberstein and elsewhere.

The coarsest varieties of this type occur as elongate lenses (normally about 1 metre thick) within Alveolites-stromatoporoid biosparudites of type 1 (Fig 5.10a). The composition of the rock varies between localities within the quarry depending on the size, type and abundance of the various skeletal components and the amount of fibrous-calcite filled cavity area. In areas where skeletal components are small (Fig 5.9b) the rocks are less obvious in the field, as the spar filled interparticle space is reduced.

The most abundant fossil constituents are elongate sheets of lamellar stromatoporoids. These are generally unbroken but are commonly severely pitted, usually, but not invariably on their upper surface (Fig 5.10b). The orientation of the vast majority of these sheets is subhorizontal, though high angled sheets (normally small) do occur. Smaller, possibly detrital, fragments of stromatoporoid occur within the matrix (Fig 5.10c).

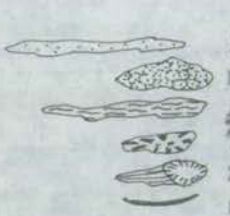
Alveolites occurs both as an encruster on the upper and lower sides of stromatoporoids and as discrete coral heads. Detrital fragments are rarely seen. Syringoporella, commensally intergrown with stromatoporoid, is also seen, but is rare.

Fig 5.10

(a) Alveolites-stromatoporoid biosparudite/biolithite. Field photograph of fibrous calcite filled voids between large (mostly tabular) skeletons and fine skeletal sediment. The large cavity system is roofed by laminar <u>Alveolites</u> (left) and downward projecting lamellar stromatoporoids (right). Large angular interconnected "lumps" within cavity system retain their origin orientation (and are therefore not dislodged boulders). Irregular nature of cavity complex with vertical contacts with host rock sediment suggests solution enlargement of primary cavities. Scale bar = 5cm. Bieber quarry.

(b) Large colonial skeletons (mainly stromatoporoids) roofing fibrous calcite filled cavities. Presence of detrital matrix in patches on undersides of some skeletons (lower right) indicates the former presence of sediment, now largely removed by solution. Scale bar = 1cm. Cut Slab. Sample no. B32.

(c) <u>As above</u>. Tracing of skeletal distribution.



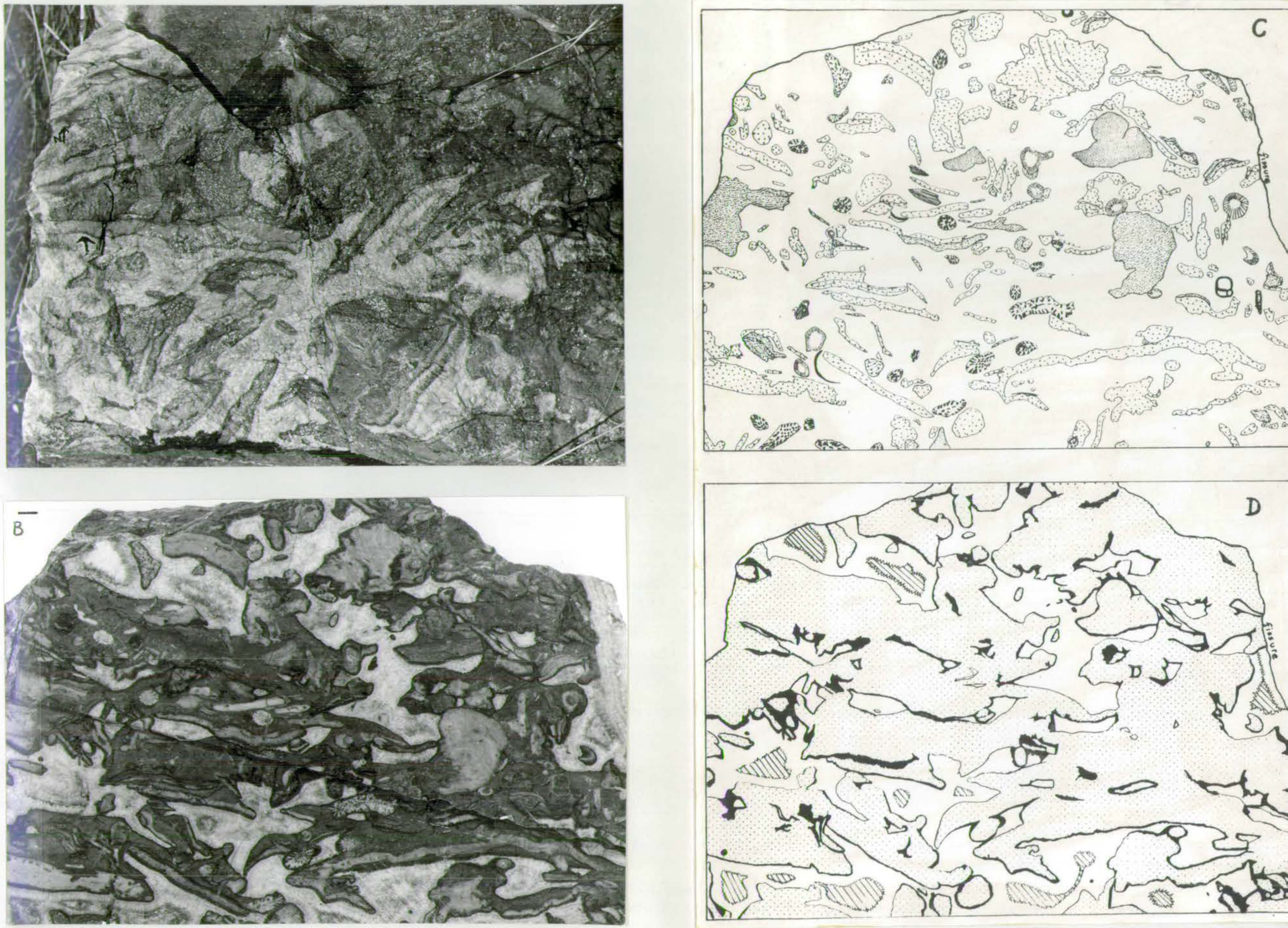
lamellar stromatoporoids
massive stromatoporoid + Syringoporella
<u>Alveolites</u>
Thamnopora
rugose corals
shell fragments.

Large skeletal organisms show a strong horizontal alignment and are not overturned. Frequent encrustation by <u>Alveolites</u> (which also occurs as massive colonies) are seen. Dendroid skeletons (<u>Stachyodes</u>, <u>Thamnopora</u>, rugose corals) are of secondary importance are randomly orientated. Some fragmentsof brachiopods and gastropods also occur.

(d) <u>As above</u>. Tracing of cavity and cement distribution.

Stipple:- matrix and skeletons
Black shading:- initial cement rim (dark in hand specimen)
Unshaded:- fibrous calcite "cement"
Cross-hatched shading:- sparry calcite cement.

Note the strong control exerted on cavity distribution by large skeletons, and also the irregular anastomizing cavity shape, suggestive of solution. First generation cement (black) is more common on cavity walls and roofs than on floors.


B
C
fissure
D
fissure

Dendroid skeletons (<u>Thamnopora</u>, rugose corals, and more rarely <u>Stachyodes</u>) are normally unbroken, sometimes branched and, in contrast to the larger colonial organisms, do not show any preferred orientation. The occurrence of branching colonies apparently "floating" in fibrous calcite (Fig 5.10b) may indicate, however, that many of these skeletons are in growth position anchored to other skeletons or to the detrital matrix in three dimensions.

Broken brachiopod valves and gastropods are rarely found in the matrix, which resembles that of the type 1 Alveolites-stromatoporoid biosparudites (Fig 5.21).

<u>Discussion</u>

Limestones showing the characteristics of this group could have been formed by either of the following processes:

(a) Large skeletons could have been dislodged and deposited as a coarse-skeletal debris pile, with finer detrital material filtering into the interstices producing geopetal sediment layers.

(b) The large skeletons could represent an open in-situ growth framework, the detrital matrix partially infilling cavities between framework branches.

In some places a detrital origin for these limestones seems reasonable, but in the majority of cases where these limestones occurred, a purely detrital origin is rejected for the following reasons:

(1) colonial skeletons show a high degree of preferred orientation, with large lamellar sheets of stromatoporoid (Keega) lying nearly horizontally and sometimes growing down into voids (Fig 5.10a). Massive colonial corals always grow vertically upwards, though solitary and fasiculate colonies tend to be poorly orientated. In the latter case, however, lack of orientation would be expected if these organisms were in situ, a pronounced horizontal alignment indicating transportation prior to burial.

(2) The majority of the skeletal components are undamaged. Some breakage would be expected if they had undergone transportation. Thin plate-like stromatoporoids especially, some of which have very complex shapes (Fig 5.10b, lower right) and are in places extensively pitted and corroded, show no signs of abrasion or breakage.

(3) Skeletal packing is extremely loose (Fig 5.10c). It is highly improbable that such a loose skeletal network could be produced merely by grain support of skeletal fragments. The common occurrence of skeletons "floating" in fibrous calcite cement also indicates a non detrital origin.

(4) Alveolites overgrowths are found both on the upper and lower sides of lamellar stromatoporoids, but in the latter case are always underlain by fibrous-calcite filled voids, suggesting that the corals grew on the stromatoporoids in their present position.

The majority of limestones of this group are, therefore, interpreted here as in-situ growth complexes of lamellar stromatoporoids and tabulate corals, between whose branches, fine skeletal detritus has been partially trapped (Fig 5.14).

Limestones of this group resemble the "type 4" fore-reef limestones of Krebs, 1969 [= Microfacies I; Krebs, 1965] interpreted by Krebs (op. cit.) as deposits of coarse skeletal rubble deposited in the upper part of the fore-reef zone, the size and nature of the skeletal material being controlled by variations of stromatoporoid growth in the immediate vicinity. For the reasons given above, a purely detrital origin is rejected for the limestones discussed here.

The nature of the complex cavity system common to all rocks of this type requires further comment. Three modes of origin are possible:

(a) The cavities were primary open interframe voids which have since been partially filled by sediment washed into the framework.

(b) The cavities are secondary in origin, having developed by leaching or winnowing of matrix from between the skeletons of colonial organisms.

(c) The cavities originated in part as primary interframe voids, were partially filled with sediment and were subsequently enlarged by leaching or winnowing.

The common occurrence of downward growing encrustations of Alveolites into voids, and of geopetal tops to sediment flooring many voids, provide evidence for a primary origin, but areas of matrix sediment below lamellar growth colonies and above cavities (Fig 5.10b, lower right), laminar colonies apparently floating in calcite cement, and high angled junctions between cement and matrix are all suggestive of a secondary origin. In places irregular blocks of sediment appear to float in calcite (Fig 5.8b upper), but these blocks are interconnected in three dimensions and the orientation of skeletons within them is like that seen elsewhere in the rock, indicating that these are not dislodged boulders but in situ interconnected solution remnants. A combined primary/secondary origin is thus thought to best explain the void characteristics noted.

The millimetre-thick black seams present at the contact between skeletal components and void filling fibrous calcite described by Krebs (1969), and attributed by him to encrusting blue green algae, are also especially conspicuous in limestones of this group (Fig 5.10d). No trace of any algal structure could be found in these seams and they appear to be part of the "cement". As the fibrous calcite is secondary, however (page 343), any possible algal structures may have recrystallized with the original cement. It is interesting to note that the anastomising secondary voids in recent Bermuda reefs are marked by a distinct blackening of the cavity edges (Ginsburg and Schroeder, 1973). Marginal micritization of skeletons, presumably due to algal boring, is commonly developed, especially on lamellar stromatoporoids.

(c) Type 3

Limestones of this type are exposed at Strohmuhle and Eberstein (Fig x, Appendix I). In both cases they are intercalated with Stachyodes-Thamnopora biosparudites.

The most apparent difference between these limestones and those described above is the complete lack of skeletal alignment of any kind.

Skeletons are normally large (commonly > 1cm, some >5cm), angular, and show little or no signs of abrasion. Massive Alveolites heads (a few centimetres in size) are the most common faunal constituents, though lamellar stromatoporoids, Thamnopora and Stachyodes are also common (Fig 5.11a). The matrix, consisting predominantly of crinoid ossicles, is, as a rule, much better sorted and coarser grained than that of the other fore-reef subfacies (Fig 5.12c) (Fig 5.21).

Discussion

Rocks included within this group are essentially bioconglomerates composed of disturbed colonial coral and stromatoporoid colonies as well as smaller fossils. The replacement of the fine detrital matrix seen in other fore-reef limestones, by a coarser well-washed crinoidal calcarenite is considered as possible evidence of deposition in the turbulant zone, these sediments representing local concentrations of skeletal rubble within the reef margin area.

Fig 5.11

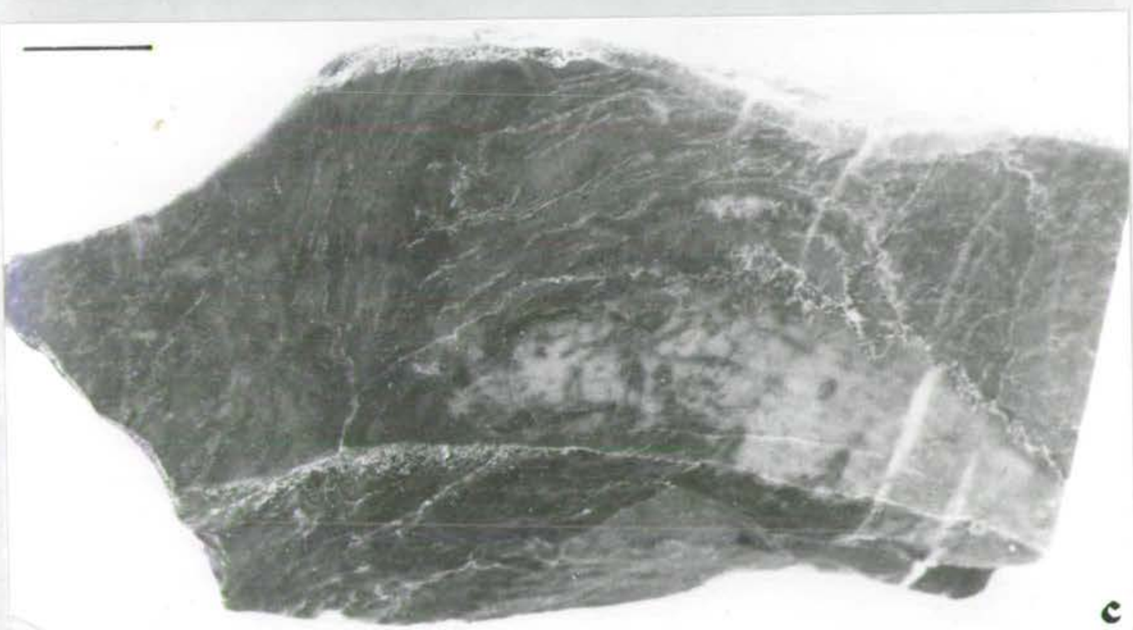
(Scale bar represents 1cm. unless otherwise indicated)

(a) Alveolites-stromatoporoid biosparudite, type 3. Randomly orientated <u>Alveolites</u> colonies (mainly massive Bulbous forms), <u>Stachyodes</u>, and lamellar stromatoporoids in well washed coarse crinoidal matrix. Hand specimen. Sample no. E46.

(b) Stromatoporoid biolithite. In situ massive hemispherical stromatoporoid mound developed on larger stromatoporoid base. Field photograph. Bieber quarry.

(c) Biolithite. Stromatolitic(?) (+ Sphaerocodium?) algal overgrowth on stromatoporoid (lower right). Hand specimen. Sample no. E78.

a



c

(4) Stromatoporoid biolithites

As was mentioned above, some of the rocks included here within the Alveolites-stromatoporoid biosparudite subfacies, composed largely of in situ stromatoporoid and coral growths forming a porous framework complex, are considered to be biolithites. Discussion in this section is limited to rocks, with a very low primary interskeletal porosity, composed mainly of massive in situ stromatoporoid growths.

Rocks of this type are developed at Bieber quarry (where they are the oldest rocks exposed in the quarry entrance), at Eberstein (immediately below the cavernous zone preceding deposition of the Iberg limestones, and at Strohmuhle (Figs x and y, Appendix I).

They are distinguished from other rocks of the Dorp Facies by the abundance of massive, subspherical, bulbous or hemispherical stromatoporoids and, in places, tabulate corals, normally so tightly packed that other skeletons and detrital matrix are virtually excluded (Fig 5.11b) (Fig 5.21). At both Eberstein and Bieber large areas of limestone are composed almost entirely of stromatoporoid colonies some larger than 30cm in diameter. They are mainly in growth position but, where transitions to fore-reef limestones occur, fragmented colonies form a stromatoporoid rubble with shelter cavities partially filled with calcareous sand similar in composition to the matrix of other fore-reef limestones.

At Eberstein many of the stromatoporoids are

Fig 5.12

(Scale bar represents 1mm. unless otherwise indicated)

(a) Stachyodes - Thamnopora biosparudite. Detail of matrix. The majority of the allochems appear to be "algal grains", with skeletal fragments (mainly brachiopods and fenestellid bryozoans) of secondary importance. Thin section, plane polarized light. Sample no. E56. [illegible].

(b) Algal(?) rimmed Stachyodes (centre right) in fine biodetrital matrix consisting almost entirely of fragments of Renalcis, crinoids, and unilocular forams. Thin section, plane polarized light. Sample no. E80.

(c) Crinoid - brachiopod coquina from reef-margin area. Note the high proportion of articulated valves of widely varying size and the lack of broken and abraded skeletal debris. Inter-skeletal voids occupied partly by unfossiliferous microsparite, similar to that found in nearby cavern systems in the reef-margin limestones, and by fibrous calcite cement filling geopetal voids within, and below brachiopod shells. Peel, plane polarized light. Sample no. E76. Negative print.

(d) Stromatoporoid biolithite subfacies. Massive stromatoporoid (below) overlain by thick crust of (?)Sphaerocodium and/or stromatolitic deposits. Detail of Fig 11c above. Thin section, plane polarized light. Sample no. E78. Negative print.

(e) Stromatoporoid biolithite subfacies. (?)Algal encrusted Stachyodes underlying brachiopod biosparites similar to (c) above. Renalcis fragments are particularly common in the upper layers of these dense (?) algal masses and in the base of the overlying bioclastic limestones. Thin section, plane polarized light. Sample no. E78. Negative print.

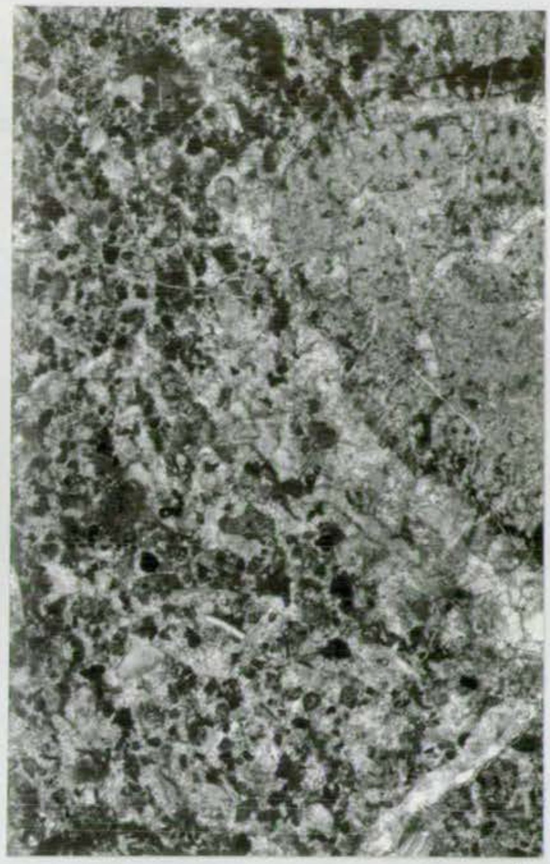
a

c

1cm.

e

0·5cm.



b

d

0·5cm.

intergrown and overgrown by dark grey to black stromatolitic crusts and mounds, developed initially as thin crusts on skeletal fragments or massive stromatoporoids but developing into substantial hemi-spherical colonies as layer upon layer was build up (Fig 5.11c, Fig 5.12d). No algal tubes have so far been identified in these stromatolites but this may be due partly to recrystallization. Girvanella is definitely not present but Sphaerocodium or some finer, perhaps non-calcareous filamentous alga may have been responsible for their formation. Renalcis is normally found on or near the surface of these growths. These structures appear to be superficially similar to the stromatolitic crusts described by Land and Goreau (1970) and to structures referred to the algal genus Spongiostroma (Hadding) by Scoffin (1971). They seem to have fully encrusted and bound skeletons (Fig 5.12e) and thus probably played an important role in this subfacies as a "cementing" agent, adding considerably to the strength and wave-resistance of the structure.

Discussion

Evidence obtained in this study indicates that only a small proportion of the reef-complex is actually composed of rocks of this type, the vast majority of the limestones being biodetrital. The rarity of exposures of in-situ reef margins can be explained by the following:

(i) In modern atoll reef-complexes the reef crest area (composed of massive algal boundstones in the Pacific) is thin (less than 100 metres wide) compared to the total width of the reef complex. In Pacific atolls, for example, in-situ reef-builders form only about 5-10% of the total volume of carbonate in the reef complex (Ladd and Tracey, 1950). Even with 100% preservation of the reef-complex, the number of exposures expected of this facies would be low.

(ii) It is likely that this rock type did not occur as a continuous "wall" around the rim of the reef-complex, but as discontinuous patchy mounds, perhaps often short lived, established only on isolated parts of the rim where bathymetric or hydrographic conditions were suitable for the development of massive stromatoporoid growth.

The actual lateral thickness of the reef margin zone is impossible to assess due to lack of adequate exposure, but at Eberstein, limestones rich in stromatoporoids (with or without stromatolites) occupy a vertical zone less than 30 metres thick. The thickness of this zone in other German Devonian reefs is thought to have been less than a few hundred metres, when developed (Krebs, 1974). The Devonian reefs are not, however, peculiar in this respect as similar reef-margin boundstones are only rarely developed in reef-complexes in the Carboniferous (Wolfenden, 1958; Broadhurst and Simpson, 1973), Permian (Smith and Francis,

1967) and Triassic (Zankl, 1969).

Of all the fore-reef rock-types discussed above, the stromatoporoid biolithites are thought to have formed nearest to sea level for the following reasons:

(i) they contain abundant massive stromatoporoids, generally thought to have been adapted to conditions above wave base, by most authors *

(ii) at Eberstein they occur immediately below a zone of extensive karst erosion, indicating proximity to intertidal or supratidal conditions.

Stromatolitic algae also occur in this subfacies, but the presence of such structures is no guarantee of very shallow waters as similar structures have been found in limestones deposited in water depths of at least 45m in Devonian reefs in Australia (Playford and Cockbain, 1969).

Summary of reef margin and fore-reef lithological groups

Although the limestones described above have been discussed in terms of distinct groups, transitions obviously exist between them. With the exception of the Skeletal Calcarenites, which are in some respects best regarded as transitional between the Schwelm and Dorp Facies, typical Dorp Facies sediment types can be

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* (Lecompte, 1954; Dinely, 1960; Edie, 1961; Perkins, 1963; Klovan, 1964; Leavitt, 1968; Jenik and Lerbekmo, 1968; Fischbuch, 1968; Embry and Klovan, 1971; Gwodsz, 1972; Krebs, 1966, 1971, 1974).

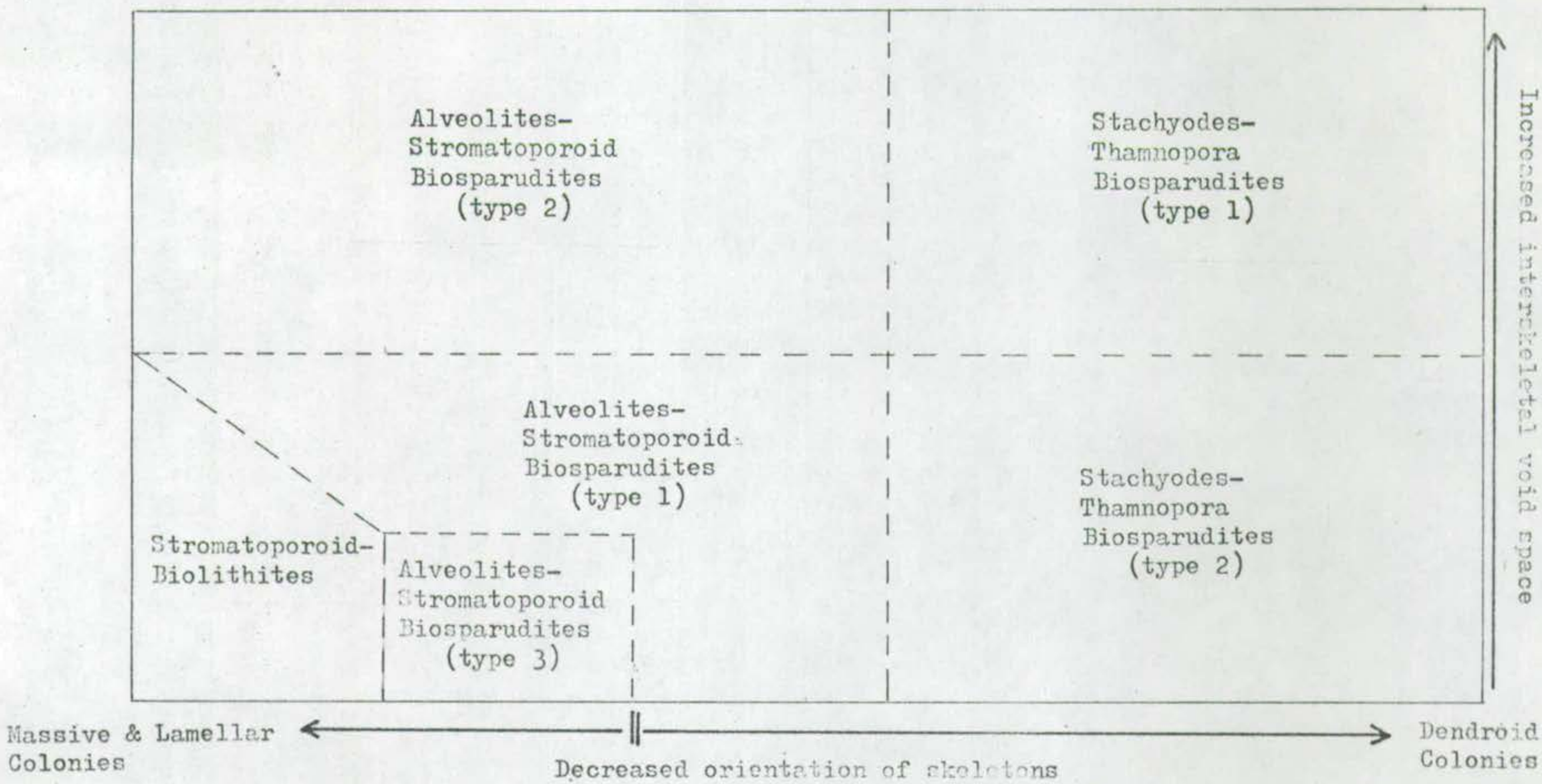


Fig. 5.13 Compositional relationships between fore-reef & reef-margin limestones.

described in terms of variations in interskeletal porosity and the size and nature of the constituent skeletal components. (Fig 5.13 and Fig 5.21).

Reconstruction of the spatial relationships between the lithological groups discussed above is difficult however, for the following reasons:

(1) Only "proximal" fore-reef subfacies are exposed in the studied area, and so no sediments deposited in the transitional zone between fore-reef and pelagic limestones were seen in the present study.

(2) The most common lithologies (i.e. Alveolites-stromatoporoid and Stachyodes-Thamnopora biosparudites), are closely associated in the field, especially at Bieber where transitions from one group to the other occur again and again within a few metres. At Eberstein, on the other hand, a clear vertical transition from the latter to the former occurs below stromatoporoid biolithites, indicating that, in general, Alveolites-stromatoporoid biosparudites were formed in areas between the reef-core and the environment of deposition of the Stachyodes-Thamnopora biosparudites. In Fig 5.14, therefore, the Stachyodes-Thamnopora biosparudites have been interpreted as having formed seawards of the Alveolites-stromatoporoid biosparudites, though the factors controlling the distribution of these subfacies may not have been primarily depth-dependent, but controlled by local hydrodynamic and topographic conditions along the reef-front.

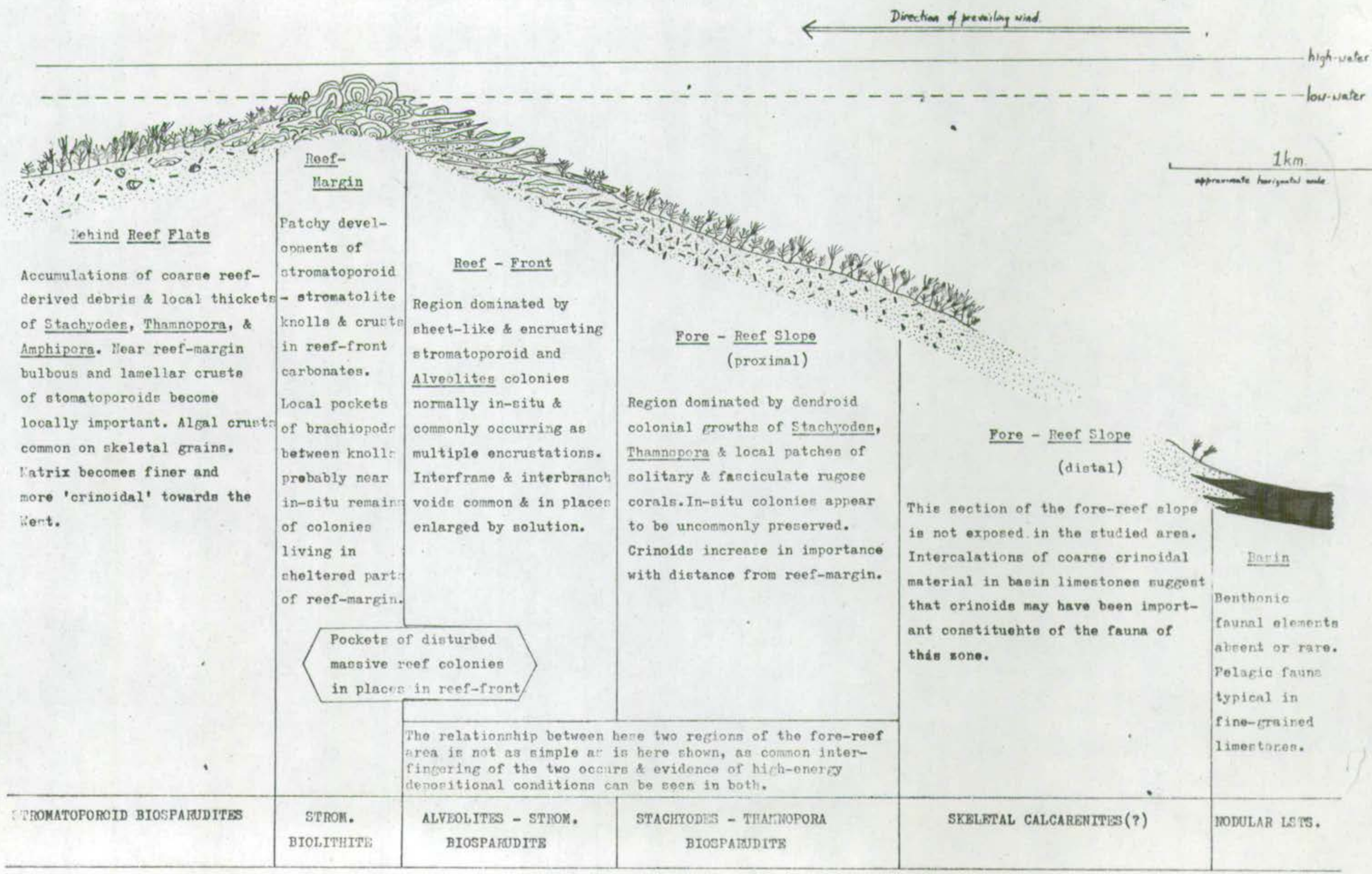


Fig. 5.14. Interpretive reconstruction of fore-reef and reef-margin environments in the Rodheim-Bieber carbonate complex.

Sediments of the back-reef area

(1) Skeletal biosparudites

Rocks of this type are exposed in roadside outcrops between Eberstein and Bieber, and in forest path outcrops on Rotenberg.

Three subfacies can be recognized, though transitions between subfacies are common because of the heterogeneous nature of the rocks:

(a) Stromatoporoid biosparudites-containing abundant "reef-skeletons" in a crinoidal pelsparite or intrasparite matrix.

(b) Stachyodes biosparudites-containing well-sorted "reef-skeletons" in a fibrous calcite cement.

(c) Crinoidal biosparudites and biosparites-in which "reef-skeletons" are of only secondary importance.

The important features of these subfacies are summarized in Fig 5.15.

(a) Stromatoporoid biosparudites

The limestones included within this subfacies are unbedded, and characterized by an abundance of poorly sorted, coarse skeletal fragments (mainly of colonial organisms) showing little or no preferred orientation (Fig 5.15).

The composition, size, orientation and proportion of skeletal material present varies from sample to

	SKELETAL BIASPARUDITES				LAMINITES	NON-SKELETAL LIMESTONES		
	1)Stromatoporoid biosparudites	2)Stachyodes biosparudites	3)Crinoidal biosparudites	4)Fenestral micrites	5) 'Algal flake' limestones	6)Micro-intrasparites	7)'Micrites'	8)Bedded dolostones
Colour	← light grey			→	← dark-grey/black		→	red
Bedding type	← massive			→	← thin			→
Skeleton: essential	Stachyodes, Amphipora + Thamnopora	Stachyodes	crinoids		—	—	—	—
Skeleton: subordinate	Bulbous stromats. Sphaerocodium(?)	Amphipora + Thamnopora	Stachyodes, Amphipora Thamnopora	Amphipora	—	forams	—	—
Skeleton: accessory	crinoids	—	Sphaerocodium(?)	gastropods calcispheres	ostracods	dasyclad. algae	—	—
Carbonate matrix	finely comminuted debris, intraclasts + microfossils	—	finely comminuted debris, intraclasts + microfossils	Micrite (structureless)	micrite flakes	intraclasts + microfossils	microsparite	recryst.dolomi
Isoluble residues	← 2-4%		→	← 0-5%	→	← 5-8%		
Voids	Geopetal cavities in places between skeletal fragments	Cavities present between all fragments	rare	fenestral voids //bedding	rare	—	—	vugs(//bedding in places).
Cement	fibrous calcite	fibrous calcite	fibrous calcite/ syntaxial overgrowths	Micrite	fibrous + sparry calcite	fibrous + sparry calcite/syntaxial overgrowths	Micrite	—
Mean size range	-2 to -5ø	-2 to -4ø	-3 to +1ø	+7ø	+3 to +5ø	+2 to +4ø	+4 to +6ø	+2 to +4ø
Sorting	v. low	high	low-high	high	v. high	v. high	v. high	v. high
Degree of grain orientation.	low	high	low	high	high	← low		
Internal fabric	Random - debris in spar/matrix	Aligned - debris in spar	Random - debris in matrix	Fine micrite slight lamination	Strong grain alignment	← Random non oriented grains		
Relations to other subfacies.	Interfingers with 2, 3, 4 + 5	Interfingers with 1, 3 + 4	Interfingers with 1, 5 + 6	Interbedded with 1, 2 + 3	Interbedded with 1 + 3	Interbedded with 3 + 7	Interbedded with 6	Lateral equiv. 6 + 7.

Fig. 5·15 Summary of main features of "back-reef" Dorp limestones.

sample. In the eastern exposures, in-situ massive stromatoporoids become common near the reef margin area. These massive to nodular stromatoporoids occur as small isolated colonies, commonly overgrowing rugose coral, Stachyodes and Thamnopora fragments, the latter, in places, themselves coated by stromatolitic micrite. In general, however, only fragmented colonies are found, especially farther west where both the size and nature of the large colony fragments change. Stachyodes, Amphipora and Thamnopora are the most widespread colonial skeletons (Fig 5.21).

In many cases, little matrix is present, the broken and abraded fragments being wholly cemented by fibrous calcite. More commonly, however, a crinoidal pelsparite matrix is present, especially west of Rotenberg where crinoidal material increases and "reef-skeletons" decrease in importance (Fig 5.16b).

Discussion

Lithologically these limestones are almost indistinguishable especially in the eastern part of their outcrop area from detrital reef material within the reef and fore-reef areas. The more common occurrence of Amphipora and algal (Sphaerocodium?) rims on skeletal grains, together with their association in the field with limestones of undoubted "back-reef" origin, suggest however that these sediments represent piles of dislodged and transported framework grains deposited immediately leeward of the "reef margin" zone (Fig 5.19).

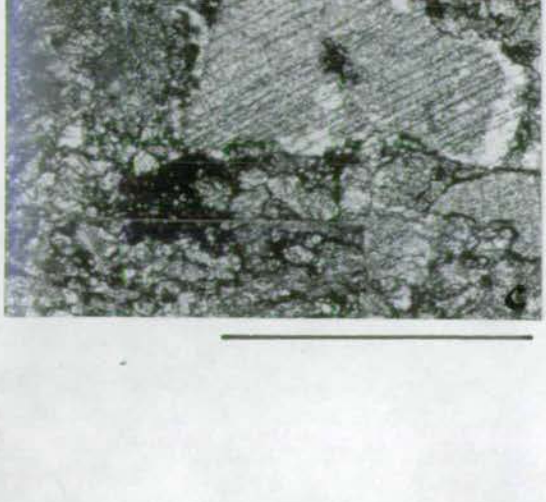
Fig 5.16

(a) Stachyodes biosparudite composed mainly of horizontally aligned, algal coated Stachyodes, Amphipora (and rare rugose corals, lower right) in fibrous calcite cement. Hand specimen. Scale bar = 1cm. Sample no. S5.

(b) Fragments of bulbous stromatoporoid in crinoidal Stachyodes intrasparite overlying (with stylolitized contact) finer crinoidal biosparite similar to (d). Hand specimen, scale bar = 1cm. Sample no. R2.

(c) Crinoidal biosparite. Poorly sorted crinoid ossicles with syntaxial rims in pelleted sparite matrix. Peel, plane polarized light. Scale bar = 1mm. Sample no. S3.

(d) Graded crinoidal biosparite. Basal part of unit contains unabraded Thamnopora and Amphipora stems and algal degraded crinoid ossicles in an intrapelsparite matrix. Upper part of unit lacks "reef" skeletons and is essentially a crinoidal pelsparite. (Transitional lithology between crinoidal biosparite and microintrasparite subfacies). Thin section, plane polarized light. Scale bar = 1mm. Sample no. H14.



(b) Stachyodes biosparudites

These are massively bedded or unbedded, well washed grain supported rocks composed largely of broken, and often rounded, unually algal-coated dendroid framework constituents in a partially recrystallized fibrous calcite cement (Fig 5.15).

The skeletal components are almost entirely Stachyodes and Amphipora fragments, showing a pronounced near-horizontal alignment in hand specimen (Fig 5.16a). Less frequently, algal coated solitary rugose corals, Thamnopora stems, Chaetetes fragments, and brachiopod valves occur, and in one case a small Manticoceras shell was noted (Fig 5.21). The size of individual skeletal fragments varies between samples but is generally fairly uniform within any one sample. In areas where the fragment size is smaller, alignment is less marked and grains are more tightly packed.

Rocks of this type are less common than the skeletal biosparudites described above, and were seen only at a few localities southeast and southwest of Rehmuhle.

Discussion

At least two modes of origin could account for the features seen in the above described rocks.

(a) They may be lag deposits of coarse skeletal material from which the matrix has been removed by current action.

(b) They may represent strand line deposits.

The latter explanation is favoured here since the former does not explain the pronounced current-orientation of skeletal fragments, the abrasion of the fragments, or the admixture of reef and off-reef elements (e.g. ammonoids) in the sediment. Furthermore, the rocks bear a striking resemblence to recent Acropora cervicornis gravels in Isla Perez (Kornicker and Boyd, 1962) and to similar shingle ramparts of the Low Isles of the Great Barrier Reef (Fairbridge and Teichart, 1948).

In the Rodheim-Bieber complex, such rocks are restricted to the area between the massive skeletal reef-like limestones of the Bieber area and the bedded "non-skeletal" back-reef limestones of the Haina area (Fig 5.20). No evidence exists, however, for the former presence of any exposed volcanic "land area" between these two localities and it is thought more probable that no such land area existed, but that material torn from the reef area by wave action and storms was piled up leeward of the reef area as low island ridges, as is the case with many present day atolls (Fairbridge and Teichart, 1948; McKee, 1959) (Fig 5.19).

(c) Crinoidal biosparudites and biosparites

These are generally located farther from the reef margin than either of the two limestone-types described above and differ from them in being essentially crinoidal sparites, in which "reef" organisms (e.g. Stachyodes, Thamnopora etc.) are of secondary importance (Fig 5.21).

As the proportion of large skeletal debris increases eastwards, these rocks pass laterally into stromatoporoid biosparudites, and into microintrasparites as the amount of skeletal debris decreases westwards.

In places, pure encrinites, with few if any allochems apart from crinoids, occur (Fig 5.16c), but transitional rock-types containing transported (often stromatolitic coated) "reef" organisms (Fig 5.16b), and graded units containing Amphipora, Thamnopora, forams, intraclasts and peloids (some algal) are more common (Fig 5.16d).

Discussion

The location of crinoidal sediments in areas intermediate between the reef-margin (stretching from Eberstein to Bieber) and the central back-reef area (around Haina and Dicke Eiche) suggests that in life they probably lived in this general environment.

Crinoids are common in back-reef areas in the Rheinisches Schiefergebirge (Krebs, 1966, 1969, 1974; Franke, 1971, 1973) but are absent in the back-reef areas of reef-complexes of the same age in Canada (Klovan, 1964; Jamieson, 1967; Leavitt, 1968). As modern crinoids (and presumably ancient crinoids too) are stenohaline, the occurrence of crinoids in their present situation can only be explained by one of the following alternatives:

(i) Crinoid material may have been swept into the back-reef area from the fore-reef region through channels in the reef margin by storms.

(ii) Crinoids may have been able to grow in the back-reef areas in German but not Canadian reefs. This would suggest that the salinity of German back-reef areas was normal marine, whereas that of Canadian lagoons was higher than normal.

The first of the above alternatives is not considered likely in the present case for the following reasons:

(i) Crinoidal sediments are purer and more abundant in back-reef than in fore-reef sediments.

(ii) Crinoidal material is commonly very poorly sorted

(iii) Crinoids are rare or absent in Stachyodes biosparudites, interpreted as being (at least in part) accumulations of skeletal material swept into the back-reef area from the reef margin and fore-reef areas.

The absence of "reef" limestones in the western part of the Schwelle and the occurrence of conodonts in sediments of the "lagoonal" region suggests that the back-reef and open marine environments were interconnected. It is, therefore, likely that normal marine conditions prevailed throughout the greater part of the back-reef area, and that local thickets of crinoids grew wherever the substrate and hydrographic conditions were suitable (Fig 5.19).

(2) Laminites

As used here, the term "laminite" includes those limestones found which are both fine grained and show distinct banding in hand specimen. All of the rocks of this group are unfossiliferous or very sparsly fossiliferous.

Two main rock-types are included within this category:

(a) Fenestral micrites

(b) "Algal-flake" limestones.

(a) Fenestral micrites

Limestones of this type occur in roadside exposures south of Rehmuhle interbedded with the "skeletal biointrasparudites" described above.

Macrofossils are extremely rare, only Amphipora stems have been noted (Fig 5.17a). Microfossils appear to be restricted to calcispheres (Fig 5.17b), no algal filaments or conodonts were found (Fig 5.21).

Sheet cracks and smaller fenestral voids occur parallel to the bedding, the former commonly weakly interconnected by sporadic vertical cracks.

(b) "Algal-flake" limestones

Limestones of this type are exposed at Obermuhle and Rotenberg where they are interbedded with "skeletal biointrasparudites".

No macrofossils occur in these limestones, only

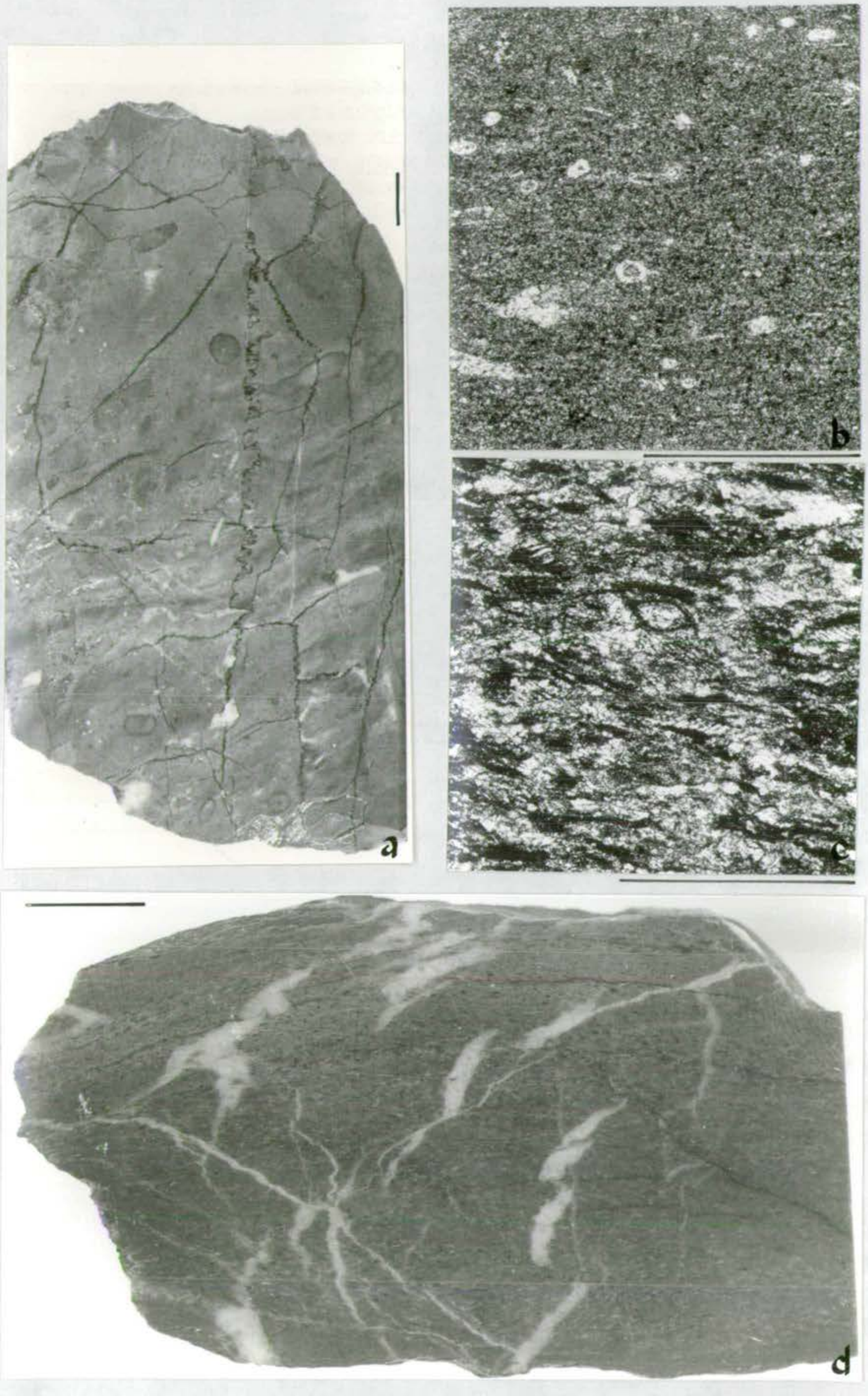
Fig 5.17

(a) Dense, massive to poorly laminated micrite cut by horizontal and vertical stylolites. Laminated zone (lower centre) rich in bedding-parallel fenestral voids. Amphipora stems are the only macrofossils present. Hand specimen. Scale bar = 1cm. Sample no. S10.

(b) Detail of above. Finely laminated micrite with fenestral voids (lower left) and scattered calcispheres. Thin section, plane polarized light. Scale bar = 1mm. Sample no. S10.

(c) Laminated "algal-flake" limestone composed of dark micritic (algal?) flakes aligned horizontally in a recrystallized spar matrix. Possible ostracod(?) upper centre. Some flakes resemble Girvanella(?) filaments. Thin section, plane polarized light. Scale bar = 1mm. Sample no. R8.

(d) Dark recrystallized laminated limestone with bedding-parallel fenestral voids(?). Laminae at base are horizontal but those in upper part of rock are updomed, (stromatolitic?). Hand specimen. Scale bar = 1cm. Sample no. R8.



rare (?)ostracod shells or (?)forams and crinoid fragments being noted (Fig 5.21). Conodonts were recovered from rocks of this group. In contrast to the limestones described above, traces of algae (?Girvanella) are common; the entire matrix consisting of horizontally aligned thin "flakes" of dense micrite, thought to be algal in origin (Fig 5.17c). In hand specimen, the lamination produced by the elongation of these flakes is very pronounced and convex laminated structures (resembling stromatolitic domes) commonly occur (Fig 5.17d).

Interpretation

Several features of these limestones are thought to be very relevant to the interpretation of their mode of origin.

(1) The low fossil content and restricted range of fossil types.

(2) The presence of sheet cracks and fenestral voids.

(3) The pronounced lamination (shown in hand specimen, at least).

(4) The association with well-washed and sorted skeletal calcirudites showing a marked alignment of skeletal fragments.

(1) The absence or extreme rarity of many fossil groups normally seen in the massive limestones is interpreted as evidence of environmental conditions, in the areas of deposition of these sediments, extreme enough to exclude

the vast majority of the animals found elsewhere in the Dorp Facies. Since no stenohaline animals are represented within these sediments, but only forms (e.g. Amphipora) apparently capable of withstanding increased salinity (Krebs, 1966, 1969; Klovan, 1964; Leavitt, 1968), they are thought to have formed in hypersaline, shallow water or possibly intertidal conditions.

(2) Deposition at times above low tide level is indicated also by the many bedding-parallel sheet cracks and fenestral voids, resulting most probably from dessication of the sediment in this zone (Fischer, 1964; Shinn, 1968).

(3) The preservation of lamination within these sediments indicates that sedimentation was at times discontinuous and that burrowing activity within the environment of deposition was low or absent. In present-day environments burrowers effectively turn over sediments in all subtidal environments (Rhoads, 1967), their activity only decreasing when the subsurface environment is reducing in nature or when salinity is increased or reduced, normally by deposition above low tide level.

(4) The field occurrence of these sediments also favours an intertidal or supratidal environment of formation as they are intimately associated with coarse skeletal gravels, some of which resemble beach like concentrates seen on modern atoll islands. The fine grained nature of the limestones discussed here however suggests deposition in a low energy environment. This paradoxical situation can only be resolved if during deposition these sediments were stabilized by algal mats.

It is, therefore, concluded that these sediments were deposited in intertidal areas on the leeward sides of shoals (possibly, emergent at times), of skeletal debris, resembling the shingle-ramparts seen in modern atolls (Fig 5.19). Such islands could have provided the necessary intertidal areas required, as well as providing, on their leeward side, a sheltered, possibly at times hypersaline, environment in which algal mud flats could develop, as is the case on the leeward side of Andros on the Bahama Bank at the present time (Monty, 1967; Shinn et al., 1969).

(3) <u>Non-skeletal dark calcarenites and calcirudites</u>

Limestone types grouped together in this category all share the following gross characteristics (see also Fig 5.15 for comparison of major features).

(i) they are dark coloured, normally black.

(ii) they are well bedded, and traversed by bedding-parallel stylolites.

(iii) they are restricted to the area between Dicke Eiche and Haina (Fig 5.20).

Two main lithological groups can be recognized:

(a) non-skeletal microintrasparites,

(b) microintrasparites rich in macrofossils.

<u>(a) Non skeletal microintrasparites</u> *

Microscopically these sediments range from well sorted intrasparites (Fig 5.18a) to fine laminated "micrites" (now microsparites) in which no allochem structure can be seen (Fig 5.18b). Small dense, black micrite "lumps" are the most common allochems sometimes enclosing tests of forams (forams with "furry overcoats" of Toomey et al, 1967), though scattered ostracods, calcispheres, dasycladacean algae (<u>Vermiporella</u> sp.), filamentous algae (<u>Girvanella</u> sp.) and sparse crinoid ossicles also occur (Fig 5.21). The cement is cloudy sparry calcity (possibly recrystallized).

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* The term "microintraclast" is used here for peloid grains of irregular shape common within these limestones. Similar grains in rocks elsewhere in the Rheinisches Schiefergebirge have been described as "pellets" (Krebs, 1969, 1974).

Fig 5.18

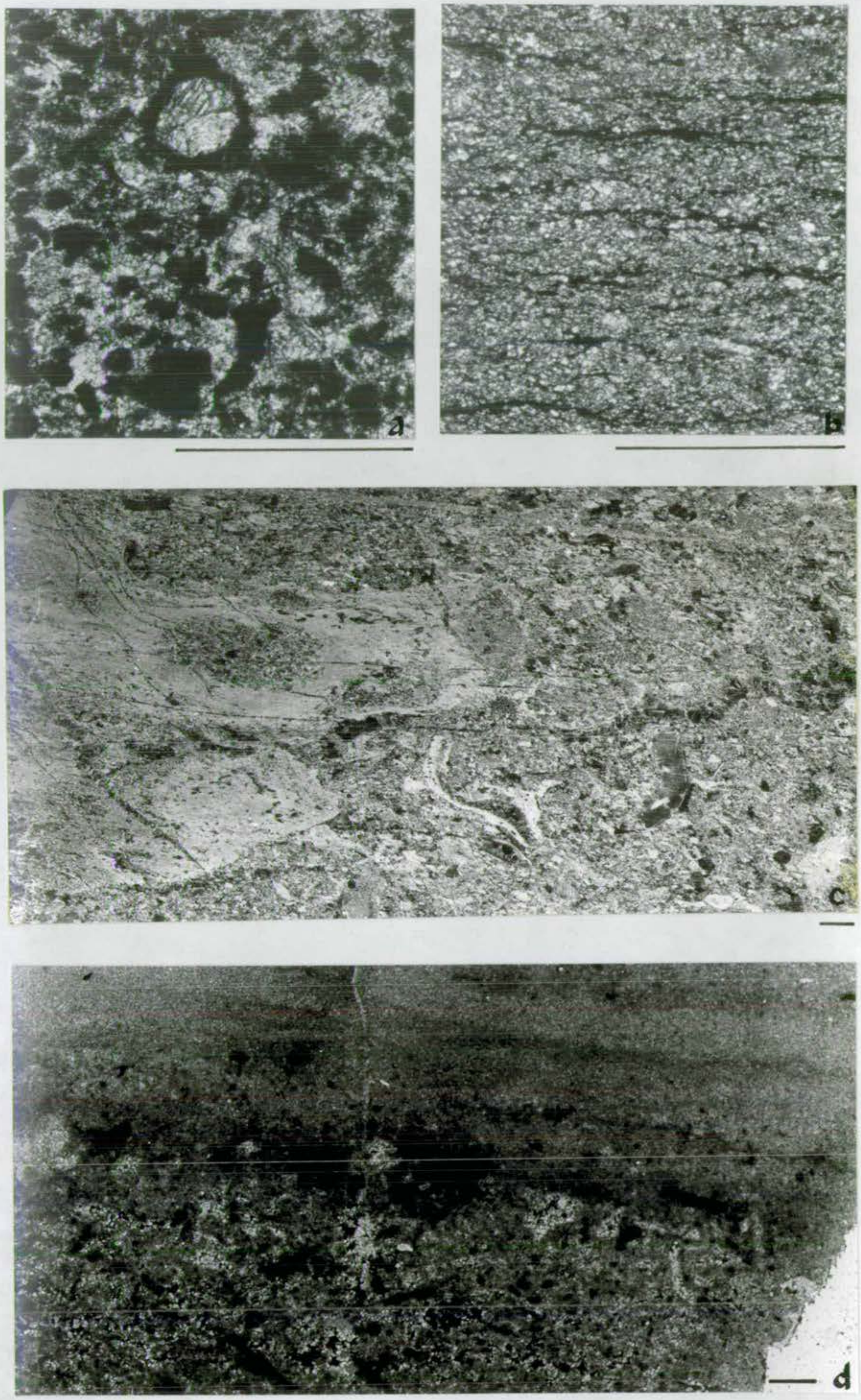
(Scale bar represents 1mm unless otherwise indicated)

(a) Microintrasparite. Dasycladacean algal stem (centre top) and structureless dense black intraclasts and peloids in turbid sparite (possibly recrystallized, in part). Thin section, plane polarized light. Sample no. H8.

(b) Laminated black microsparite (possibly recrystallized micrite). No fossils are present. Peel, plane polarized light. Sample no. H28.

(c) Stachyodes stems (with stromatolitic or Sphaerocodium coating), algal degraded crinoid ossicles and articulated ostracods in fine microintrasparite matrix. Many of the intraclasts, like the fragments in matrix may be pieces of "algal" micrite. Thin section, plane polarized light. Sample no. H6.

(d) Bedded dolostone. Lower part of photograph shows comparatively coarse grained xenotopic to hypidiotopic dolomite (possibly dolomitized equivalent of (b) above). Upper part of photograph shows fine grained xenotopic domomite, with fine sedimentary lamination preserved (possibly dolomitized equivalent of (c) above). Thin section, plane polarized light. Sample no. H17.


a
b
c
d

(b) Fossiliferous microintrasparites

In many respects, rocks of this type are transitional between the non-skeletal limestones above and the skeletal limestones and laminites of the back-reef area. They are poorly sorted due to the varied skeletal content, the composition of which varies from sample to sample, probably reflecting the heterogeneous nature of skeletal growth areas from which the debris was derived. Crinoid ossicles, Thamnopora, Amphipora and Stachyodes (often encrusted by (?)Sphaerocodium) (Fig 5.18c) are the most common skeletons present (Fig 5.21).

Graded units rich in crinoid material form transitions to the crinoidal biosparites described above (Fig 5.16d), and limestones containing algal coated dendroid stromatoporoids (Stachyodes), in which the intraclasts show a tendency to elongation (resembling "algal flakes"), are probably transitional to the laminites found further to the east in the area south of Rotenberg.

Discussion

The black colour, unique to limestones of this subfacies in the massive limestones of the Rodheim-Bieber area, is caused by high concentrations of unoxidized organic material preserved in the intraclasts.* The

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* When dissolved in acetic acid, large amounts of H_2S were given off and a black oily film was developed on the surface of the acid and on the sides of the beaker in which the limestone was dissolved.

restriction of black colouration to this subfacies alone, suggests that the organic material is a primary constituent of the rocks, suggesting that conditions below the sediment surface, and possibly even above the sediment-water interface, at the time of deposition were reducing in character, due perhaps to sluggish water circulation in some parts of the central regions of the back-reef area. Similar dark organic-rich sediments are found in central lagoonal areas of other reef-complexes of the same age in other areas of the Rheinisches Schiefergebirge (Gwosdz, 1971).

The dominantly non-skeletal autochthonous allochem content is consistent with the remoteness of the depositional environment from the "reef-margin" area in the Bieber region to the east. The origin of the non-skeletal intraclasts is, in many cases, difficult to assess, though three possible modes of origin are thought likely:

(i) formation of intraclasts by "clotting" of micrite, or by adhesion of micrite to the spinose tests of unilocular forams and calcispheres (Illing, 1954).

(ii) formation by micritization of skeletal grains due to intensive algal and/or fungal boring (Purdy, 1963; Bathurst, 1966). Grains so formed would be more easily rounded and thus loose any trace of their original skeletal shape (Bathurst, 1967).

(iii) formation by trapping of micrite by algal threads to give algal grains (Wolf, 1965a).

All three processes played some part in the genesis of these grains, but the occurrence of definite Girvanella threads within some of the intraclasts, and the dense black nature of the micrite of the intraclasts suggests that algae were commonly involved in their formation either as "borers" or as "trappers".

These sediments bear a striking resemblance to the "grapestones" of the Bahamas platform, both petrographically (compare Fig 5. with Fig 152, page 124; Bathurst, 1971) and in their site of formation:

(i) both are dominated by non skeletal microintraclasts.

(ii) both are dark coloured and rich in organic material.

(iii) both lack benthonic "reef" organisms but contain a comparatively rich foram fauna. (Forams are even more abundant, however, in Pacific atoll lagoons where the dominant forms are spinose globular species (e.g. Calcarina), superficially similar to the abundant Parathurammina species found in rocks of this subfacies (Cushman et al, 1953)).

(iv) both occur in central lagoonal environments, passing laterally into micritic sediments in deeper or quieter waters.

For these reasons, the sediments of this subfacies are considered as ancient analogues, though not necessarily homologues, of recent "grapestones", and are interpreted as having formed under similar conditions (Fig 5.19).

In those microintrasparites where large skeletal material is abundant, the size grading and orientation of the skeletons within the sediment indicates that they have been brought into the environment from elsewhere. The lack of abrasion, however, suggests that, once in their present environment, little if any current disturbance occurred, as no indications of current sorting or of winnowing were noted.

The increase in grain size and importance of large skeletal organisms (e.g. Amphipora, Thamnipora, Stachyodes and crinoids) from west to east and northeast in the back-reef area indicate that the source of this material lay to the east and north east of Haina. Some of the material may have been derived from the reef area, but it is probable that much of the debris came from areas of stromatoporoid, coral and crinoid growth in the back-reef area itself, closer to the reef rim.

Bedded dolostones

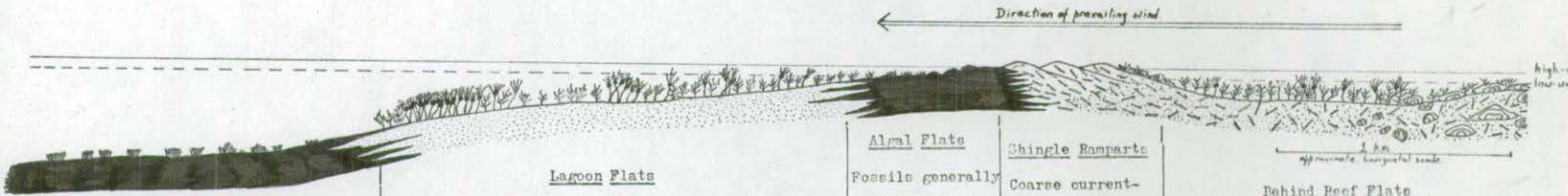
Thinly to thickly bedded dolostones occur in one locality south west of Haina. The sediments are completely replaced by a xenotopic dolomitic fabric, and no fossils are present. Considerable control has been exerted over dolomitization by original grain size differences, however, and fine bedding structures are preserved (Fig 5.18d). In their bedding characteristics, these dolostones resemble the microintrasparites described above, and are most probably dolomitized equivalents (Fig 5.15). Dolomitization in these sediments is considered in more detail in Chapter 6, page 379.

Summary of back-reef lithological groups

The present distribution of the various lithological groups in the back reef area of the Rodheim-Bieber reef-complex indicates that, in general, transitions from one group to another occur laterally rather than vertically, with coarse-grained, skeletal, reef-derived biosparudites developed in the eastern part of the back-reef area, passing westwards into non-skeletal carbonates as the influence of the reef-margin decreased (Fig 5.19 and 5.21). Between the two extremes of the reef margin and the remote back reef areas, local variations in topography, substrate type and salinity are thought to have given rise to the wide variety of sediment types seen in these areas.

Because of the small size of the Rodheim-Bieber reef complex, the range of environments in the back-reef area were not so extensively developed as in other areas of the Rheinisches Schiefergebirge (Krebs, 1967; 1974). Consequently rapid transitions from one lithological type to another occur, and hybridization of sediment characteristics of adjacent environments is common, due to the increased likelihood of sediment mixing over such a small area. Because of the small size, also, skeletal derivation from the reef margin has exerted a strong influence on sedimentation over large areas of the back-reef region.

The absence of a reef "rim" in the western part of the complex, moreover, appears to have resulted in



Lagoon Centre

'Grapestone'-like sediments form in quieter, deeper(?) areas. Unfossiliferous calcilutites form in deepest &/or quietest areas of lagoon centre. Intermittent influxes of coarse skeletal debris from shallower(?) areas nearer to reef-margin, produce skeletal lenses and layers in otherwise nonskeletal calcareous sediments. Dasycladacean algae & forams only fossils commonly found.

NON-SKELETAL LIMESTONES

Lagoon Flats

Predominantly sandy substrate with scattered thickets of crinoids, Amphipora, Thamnopora, & Stachyodes.
Only dendroid colonies developed in this area due perhaps to soft substrate.

CRINOIDAL BIOSPARITES & BIOSPARUDITES

Algal Flats

Fossils generally excluded by extreme environmental conditions. In places, low stromatolite domes, in others sheet-cracks & fenestral voids. Amphipora may have been able to grow in the less exposed areas.

LAMINITES

Shingle Ramparts

Coarse current-washed accumulations of Stachyodes, Thamnopora, etc. piled up by waves & storms as cobble & boulder beach deposits.

STACHYODES BIOSPARUDITES

Behind Reef Flats

Scattered areas of prolific Stachyodes, Thamnopora, Amphipora, & crinoid growth on a sandy substrate. Small massive stromatoporoid colonies locally important in eastern regions close to reef-margin.
Coarse skeletal detritus washed into this region from reef-margin zone to the east.

STROMATOPOROID-BIOSPARUDITES

Fig. 5.19. Interpretive reconstruction of back-reef Dorp environments in the Rodheim-Bieber carbonate complex.

near-normal marine conditions over most of the back-reef area. The main effects of this are seen in the importance of crinoids and in the rarity of Amphipora in the majority of back-reef sediments, the latter occurring abundantly in more restricted lagoonal areas elsewhere in the Rheinisches Schiefergebirge (Krebs, op. cit.) and in Canada (Klovan, 1964; Jamieson, 1967; Leavitt, 1968). Only in local areas on shingle ridges, in the eastern part of the lagoon, where salinity conditions fluctuated is Amphipora an important faunal element.

Summary of Dorp Facies lithologic groups

In the Adorf stage the greatest diversification of lithologies occurred within the Rodheim-Bieber carbonate complex. The relationship between individual lithologies is commonly complex with lenses of one rock-type occurring within large masses of another. In broad terms, however, these major lithological divisions show a fairly distinct geographical separation.

The spatial relationship between the major rock types is summarized in Fig 5.20 below. Such a diagram can at best of course only present a generalized picture of the depositional pattern, since it attempts to illustrate geographical relations over a period of geological time (in this case, the P, asymmetricus zone) during which the position of facies boundaries did not remain constant. In some areas, for example, conditions appear to have fluctuated between "back-reef" and "fore-reef" as at Obermuhle and Rehmuhle. The actual facies groups moreover have not retained their true spatial relationships due to tectonic shortening.

From the evidence still remaining in the Rodheim-Bieber complex, however, the following conclusions can be drawn:

(i) The massive Dorp limestones are virtually surrounded by off-reef sediments deposited in deeper water.

(ii) The limestones are limited in area. Even allowing for tectonic compression, the total width of the complex could not have exceeded a few kilometres.

Fig. 5.20

Areal distribution of major facies units in L. Adorf Stage.

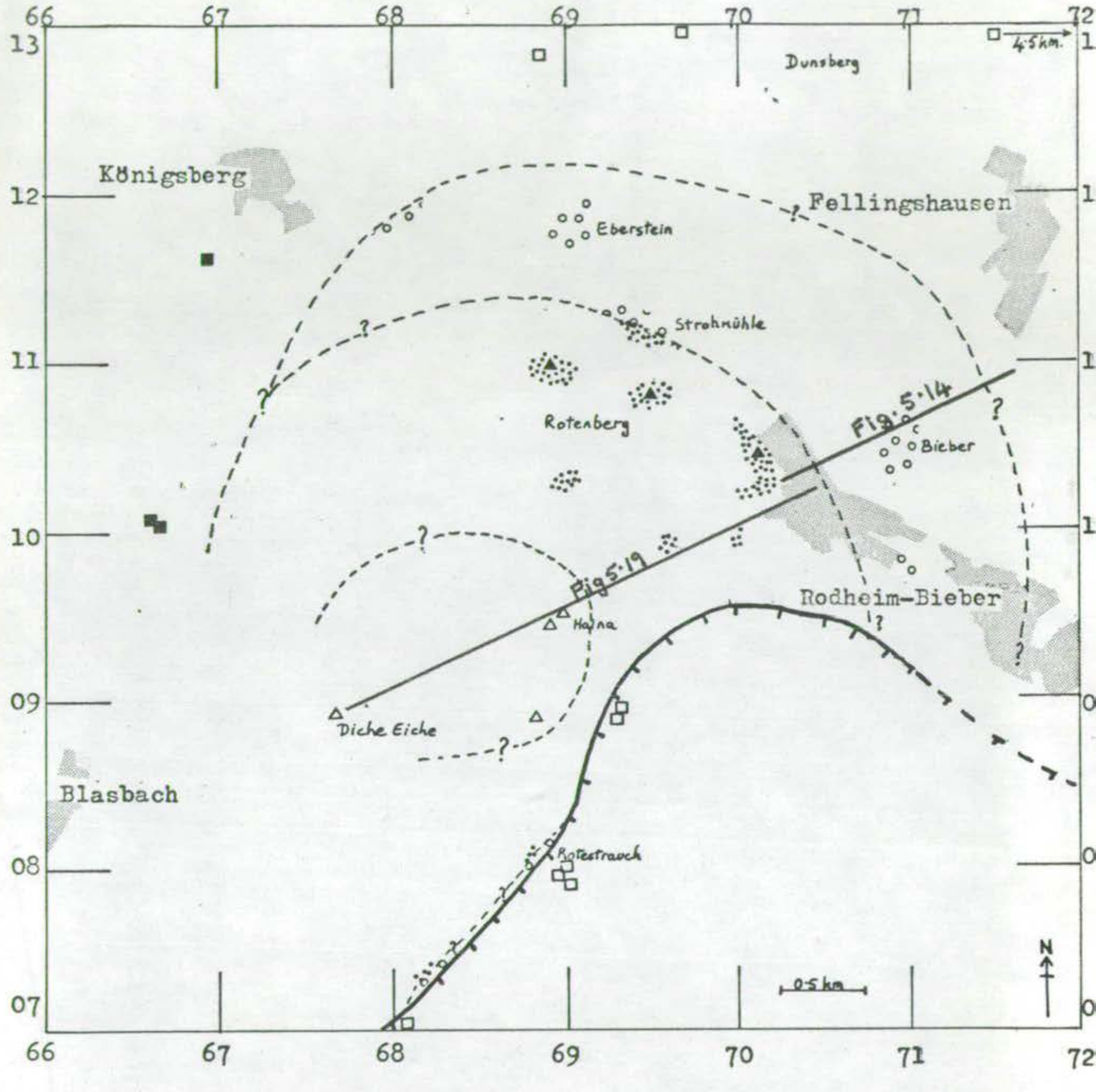
Symbols on map represent outcrops or groups of outcrops. (Compare with Maps 1 and 2, Appendix I).

No subdivision of fore-reef lithologic groups has been attempted since little spatial separation between them exists in the field, and most of the fore-reef limestones are of a proximal nature, limestones deposited distant from the reef margin in the fore-reef/basin transition zone being hidden beneath younger sediments.

The back-reef sediment types show a much greater degree of spatial separation, and therefore are shown as separate groups on the map.

No evidence was found for the existence (or former e existence)of a reef-rim on the western edge of the Schwelle and as pelagic nodular limestones of L. Adorf age occur in areas not far from exposures of limestone displaying "extreme" back-reef characteristics, it is thought highly unliekly that any reef-rim was ever present on this side of the Schwelle.

Exposures to the west, north and east of the schwellen area, containing rocks of L. Adorf age never show any signs of Massive limestones. In their place nodular limestones are developed close to the schwellen area, being replaced basinwards by siliceous and detrital mudstones and siltstones. In the southern area, these latter basin sediments have been thrust onto the schwellen top.



Fore-reef lithologies

Back-reef skeletal lithologies

Back-reef laminites

Back-reef non-skeletal lithologies

Off-reef pelagic limestones

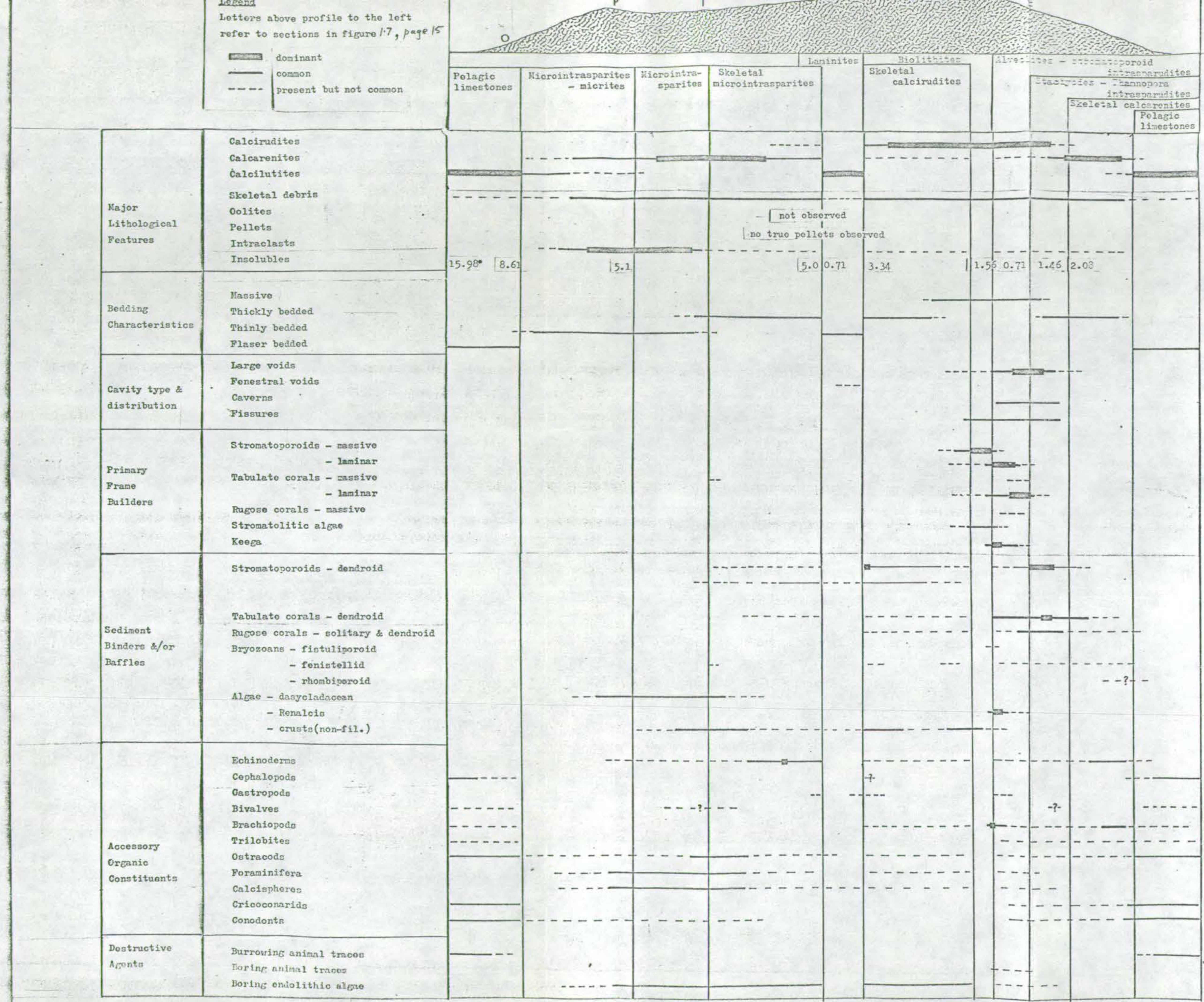
Off-reef non-carbonate lithologies

Thrust zone

(iii) A definite lithological zonation is apparent within the complex. The coarsest and thickest sediments occur along the north-eastern and eastern edge of the complex with thinner and finer sediments developed in the central, south-western and western areas (Fig 5.21).

(iv) A definite faunal and floral zonation can also be distinguished, with massive and dendroid colonial organisms, some capable of frame construction, concentrated in the arcuate eastern zone between Königsberg and Bieber. To the west and south west of this zone the frequency of large colonial skeletons rapidly decreases, and the degree of reworking rises (Fig 5.21).

The distribution of lithological groups in the Dorp Facies, therefore, indicates that at this time the carbonate complex was atoll-like with an arcuate reef-zone enclosing, or partially enclosing, a quieter lagoonal area in which finer grained sediments, lacking the prolific colonial growths seen in the marginal areas to the east, were deposited.



Legend
Letters above profile to the left refer to sections in figure 1·7, page 15
dominant
common
present but not common
O
P
Pelagic limestones
Microintrasparites - micrites
Microintrasparites
Skeletal microintrasparites
Laminites
Biolithites
Skeletal calcirudites
Alveolites - stromatoporoid intrasparudites
Stachyodes - Thamnopora intrasparudites
Skeletal calcarenites
Pelagic limestones
Major Lithological Features
Calcirudites
Calcarenites
Calcilutites
Skeletal debris
Oolites
Pellets
Intraclasts
Insolubles
not observed
no true pellets observed
15.98*
8.61
5.1
5.0
0.71
3.34
1.56
0.71
1.46
2.08
Bedding Characteristics
Massive
Thickly bedded
Thinly bedded
Flaser bedded
Cavity type & distribution
Large voids
Fenestral voids
Caverns
Fissures
Primary Frame Builders
Stromatoporoids - massive
- laminar
Tabulate corals - massive
- laminar
Rugose corals - massive
Stromatolitic algae
Keega
Sediment Binders &/or Baffles
Stromatoporoids - dendroid
Tabulate corals - dendroid
Rugose corals - solitary & dendroid
Bryozoans - fistuliporoid
- fenistellid
- rhombiporoid
Algae - dasycladacean
- Renalcis
- crusts(non-fil.)
Accessory Organic Constituents
Echinoderms
Cephalopods
Gastropods
Bivalves
Brachiopods
Trilobites
Ostracods
Foraminifera
Calcispheres
Cricoconarids
Conodonts
Destructive Agents
Burrowing animal traces
Boring animal traces
Boring endolithic algae
?
?
?
?

Iberg Facies

Location and age

Limestones included within this facies are only exposed in a few areas, and are best developed in the northern extension of Eberstein quarry where a total thickness of 25 metres of these limestones is almost completely exposed. Scattered exposures of lithologically similar rocks were also seen west and south of Dicke Eiche (Sample nos. W14 + N16, Maps 1 and 2, Appendix I), and possibly also at Rotenberg, but there the rocks are tectonically deformed making lithological comparisons difficult.

In all cases rocks of this facies stratigraphically follow massive Dorp limestones of the A. triangularis zone and are succeeded by nodular limestones of Upper P. gigas age.

Macroscopic features

Two distinct limestone groups or subfacies can be recognized within the Iberg limestone unit at Eberstein:

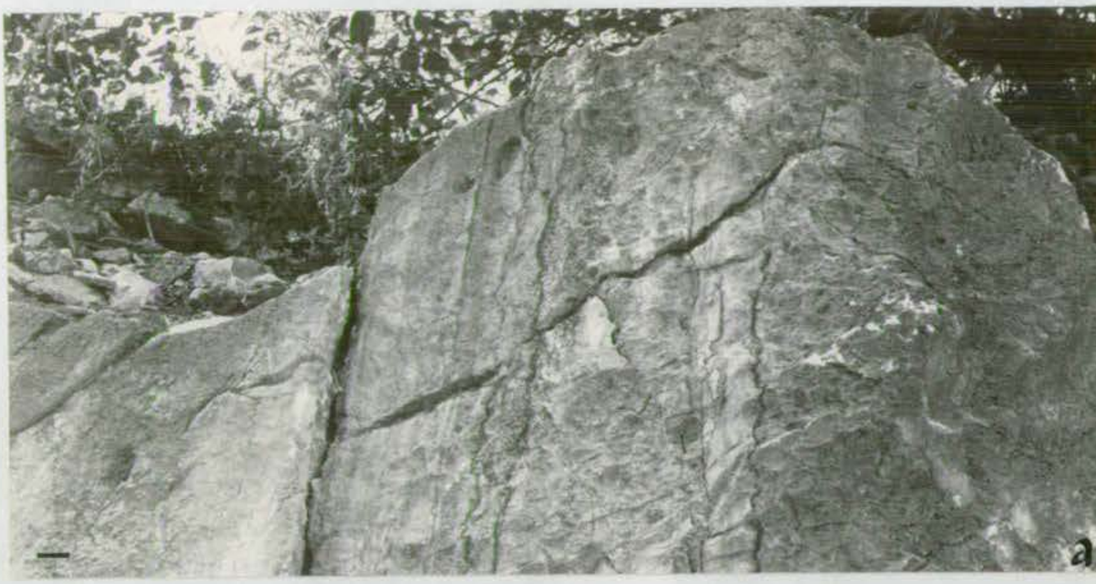
(a) a lower distinctly-bedded group (Fig 5.22a), generally light in colour, coarse grained and often containing lenses of very coarse skeletal debris (e.g. Stachyodes and Thamnopora, normally), more typical of the skeletal Dorp limestones below (Fig 5.22b). Near the base of this group, small flattened mound-like colonies of Alveolites and Philipsastrea are found

Fig 5.22

(a) Vertically bedded Iberg limestones-northern extension of Eberstein quarry. Junctions between major bedded units modified and emphasized by bedding-parallel stylolites. Finer bedding can be seen within these major units. Note tendency for development of incipient cleavage at a high angle to bedding (diagonally from lower left to upper right in photograph). Scale bar = 10cm. . Field photograph.

(b) Poorly sorted crinoidal calcarenite unit, containing disorientated and relatively unabraded Stachyodes (and more rarely, Thamnopora and Amphipora) stem fragments, above calcarenite lacking these macrofossils. (Skeletal subfacies). Scale bar = 1cm. Hand specimen from lower part of Iberg sequence, Eberstein. Sample no. E98.

(c) Mottled and indistinctly bedded fine calcarenites and calcsiltites. Light mottled areas are fine sparry skeletal calcarenite layers within densely packed microskeletal calcisiltites. Macrofossils (e.g. broken solitary rugose coral, lower left) are extremely rare, and always severely abraded. Stylolites common throughout. In places, rocks look superficially like nodular limestones, into which they grade. (Microskeletal subfacies). Scale bar = 1cm. Hand specimen from upper part of Iberg sequence, Eberstein, Sample no. E102.



a

b

c

in-situ in places. These limestones (referred to hereafter as Skeletal calcarenites) appear to be equivalent to the Hahnenfurth subfacies of Krebs (1974).

(b) a finely bedded or indistinctly bedded, normally dark coloured, group of fine grained limestones, in which macrofossils (greater than 0.5cm.) are rare or lacking (Fig 5.22c). This group lies above the skeletal calcarenites and below the nodular pelagic limestones at Eberstein, and is, in many respects, transitional between the two. These limestones (referred to hereafter as Microskeletal calcarenites) appear to be equivalent to the Schlupkothen subfacies of Krebs (1974).

Microscopic features

(a) Skeletal calcarenites

In those parts of the sequence, near the base especially, where in-situ coral growths and dendroid stromatoporoid fragments are common, the limestones tend to resemble the Stachyodes-Thamnopora biosparudites (type2) of the underlying Dorp Facies. Most of the limestones are crinoidal biosparites in which the grains are normally poorly sorted and rounded (Fig 5.23a). Apart from crinoid fragments, which are by far the most important matrix constituents, brachiopod and fenestellid bryozoan pieces, gastropods, ostracods, trilobites, and pieces of Renalcis are found in the matrix. Renalcis also occurs in crusts on skeletons, particularly Amphipora (Fig 5.23b). Allochem size and composition vary from sample to sample

Fig 5.23

(Scale bar represents 1mm unless otherwise indicated)

(a) Thamnopora-rich crinoidal biosparite (skeletal subfacies). Crinoid ossicle (centre) encrusted by Thamnopora. Matrix composed mainly of crinoid fragments and scattered pieces of Renalcis cemented by syntaxial overgrowths on crinoids. Thin section, plane polarized light. Sample no. E89.

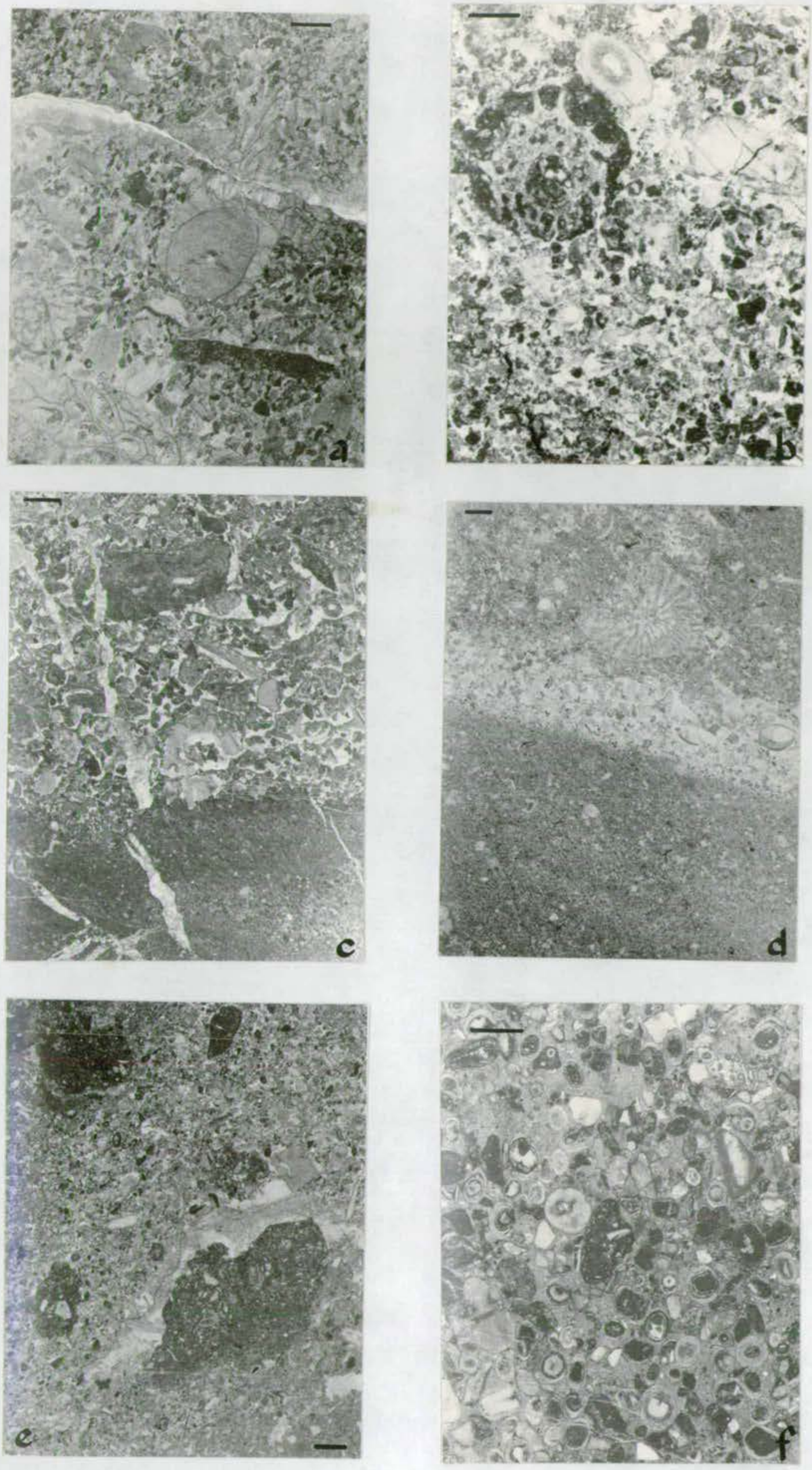
(b) Crinoidal biosparite (skeletal subfacies). Amphipora stem (upper left) wholly encrusted by Renalcis which also occurs fragmented in matrix (black blobs) along with crinoid ossicles. Way up given by geopetal infilling in crinoid ossicle (upper centre). Thin section, plane polarized light. Sample no. E92.

(c) Stylolitized junction, between crinoidal calcarenite (above skeletal subfacies) and fine skeletal calcsiltite (below microskeletal subfacies). Majority of grains in calcarenite are crinoidal, though rarer Stachyodes (upper centre), Thamnopora, Renalcis, brachiopod and ostracod fragments also present. Note fine geopetal internal sediment trapped within interparticle voids. Calcsiltite composed of closely packed microfossils, with Renalcis and finely comminuted skeletal debris. Thin section, plane polarized light. Sample no. E98.

(d) Finely comminuted skeletal and microskeletal calcisiltite (microskeletal subfacies) similar to basal sediment in (c) interrupted by thin crinoid-Renalcis biosparite layer. Way up given by geopetal infilling in ostracod shell (extreme right, centre). Note broken solitary rugose coral (centre) at base of upper microskeletal calcisiltite unit. Thin section, plane polarized light. Sample no. E104.

(e) Finely comminuted skeletal calcarenite (Microskeletal) riddled by sediment filled burrow(?) structures. Fine sediment fillings contain cricoconarids. Thin section, plane polarized light. Sample no. E108.

(f) Oolitic Facies. Crinoid ossicles (frequently marginally micritized) and ooids (mainly with crinoidal nuclei) in recrystallized ferroan-microsparite/pseudosparite matrix. Both concentric and radial structures in oolitic coats preserved despite silicification. Thin section, plane polarized light. Sample no. X3.


a
b
c
d
e
f

but all of the sparites in this group are "well-washed" with fine detritus only occasionally found in "traps" between skeletal grains (Fig 5.23c).

(b) Microskeletal calcarenites

In contrast to the calcarenites described above, limestones in this subfacies rarely contain large skeletal components (such as Stachyodes, Thamnopora, rugose corals etc.), and when these do occur they are always broken (Fig 5.23d).

The matrix is also considerably finer, consisting of finely comminuted crinoid, bryozoan and brachiopod material plus fragments of Renalcis, calcispheres, foraminifera (both uni- and multilocular), ostracods, trilobites, conodonts and, in places, cricoconarids. Allochems tend to be more tightly packed in this upper unit giving the limestones a superficial calcilutite appearance (Fig 5.23c). Near the top of the sequence, burrow(?) structures filled with cricoconarid micro-sparite are found (Fig 5.23e).

One of the most distinctive features of these fine grained limestones is the occurrence of irregular and bedding-parallel sheet crack and stromatactis-like structures (Fig 5.25a). Some of the smaller restricted cavities are obviously shelter cavities, possibly secondarily enlarged, but the larger elongate cracks are postlithification structures, showing no roof-skeleton traces capable of supporting loose sediment above the void. Shear failure, suggested as the cause of similar "sheet-cracking in Carboniferous Walsortian "reefs" of Ireland (Schwarzacher, 1961; Lees, 1964), is considered the most likely explanation for the sheet-cracking in the fine-grained sediments discussed here.

Discussion

The rocks of this facies show the influence of three sources:

(i) indigenous skeletal material, mainly crinoids, Renalcis, ostracods, brachiopods and foraminifera.

(ii) reef-derived material or, at least, material showing more affinities with reef environments (e.g. Stachyodes, Thamnopora, Amphipora, corals) important near the base but decreasing in importance up the sequence.

(iii) pelagic material (i.e. microsparite and pelagic organisms) which is normally swamped by coarser material but can be seen to be present in the form of rare cricoconarids in some samples, particularly in the "transitional" limestones at the top of the sequence.

The limestones therefore probably formed in a transitional environment between the reef and pelagic "off-reef" environments. Such an environment is found in the fore-reef situation of reef-complexes in the Rheinisches Schiefergebirge in the fore-reef facies deposited farthest from the reef source (Krebs, 1966, 1971, 1974; Franke, 1973). These limestones, therefore, could be interpreted as talus material deposited low on the fore reef slope, the shift from "reef dominated" to "pelagic-dominated" sediments reflecting a shift of environment basinwards.

This interpretation is rejected for the following reasons:

(i) rocks of this facies are younger than any known "reef" limestones in the area.

(ii) at the time of deposition of these rocks, material was being deposited in fissures and caves elsewhere (e.g. at Bieber).

(iii) these limestones appear to lie over a karst-eroded area of reef limestone in which pelagic mud was deposited prior to deposition of the limestones here described (see page 261).

(iv) the limestones dealt with here commonly contain Amphipora, normally not found in fore-reef limestones and certainly not seen in fore-reef limestones deposited at a great distance from the reef.

As an alternative to the above interpretation, it is suggested that the limestones of this facies were not deposited on the lower fore-reef slopes, but on the upper slopes or on the reef-margin area itself, and that they represent short-lived "shelf-sands" deposited on or immediately seawards of the reef-margin in the period between reef destruction and the onset of pelagic sedimentation. The upwards decrease in importance of "reef" skeletons in these sediments is thus interpreted as being due not to increasing distance from source but to the progressive elimination of that "source" through reef death, erosion, and subsequent subsidence.

Cap-like deposits of skeletal (mainly crinoidal) sands are commonly found above Devonian reef-limestones in Europe (Krebs, 1966, 1968, 1971, 1974) and Canada (Mountjoy, 1967; Leavitt, 1968). In all cases they are succeeded stratigraphically by nodular calcilutites. Analogous situations are found also in many areas of the Alps, where crinoidal biosparites are commonly found irregularly developed between karst-eroded Triassic and Liassic reef and shelf carbonates and younger Jurassic nodular pelagic limestones (see Jenkyns, 1971 for full references).

The well-washed nature of at least the lower limestones of this facies indicates that these sediments were deposited in a high-energy environment. In this respect they differ from the faunally similar reef-distant limestones which contain a high proportion of fine detrital matrix. The degree of sorting of the crinoid ossicles is not taken to be a reliable measure of environmental conditions, since the post-mortem behaviour of crinoids, when subjected to current movement, makes winnowing and sorting difficult (Cain, 1968). Shallow marine conditions are also indicated by attempts, early in the sequence, by colonial rugose and tabulate corals, to reestablish biolithites similar to the Alveolites-stromatoporoid biosparudites of the underlying Dorp Facies. Failure to reestablish reef-conditions is thought to be due to resumption of subsidence after emergence at too fast a rate for "reef-organisms" to successfully recolonize the old reef surface. Conditions at this time are thought to have been very much like those in parts of the back-reef flats in the lee of the Dorp reef-margin, where prolific crinoidal growth also occurred, as fossils (e.g. Amphipora) typical of back-reef, but never of fore-reef Dorp sediments, are common in the Iberg limestones.

It is therefore concluded that these sediments were formed in shallow water conditions above a dead reef surface (the hard bottom favouring prolific crinoid development (Clark, 1957)) as skeletal sand banks. Continued subsidence throughout the A.triangularis zone resulted in the elimination of these sediments in the lower gigas zone and to their replacement by carbonates lacking a benthonic fauna or flora.

There is considerable evidence, however, (crinoidal deposits in fissures of toII β and CuII β/γ age) that similar conditions to those described above either persisted in a few areas on the dead reef surface throughout the Upper Devonian and Lower Carboniferous or, more likely perhaps, that such conditions were reestablished locally through tectonic uplift associated with fissuring at certain periods in the Upper Devonian and Lower Carboniferous (page 248).

<u>Oolitic Facies</u>

This facies is restricted to the Lower Carboniferous (CuIII α) in the studied area and is represented only by fragments within "deeper-water" breccias in the area north of Königsberg (Chapter 3, page 70).

Though not temporally connected with the main carbonate development of the Rodheim-Bieber area, the presence of oolitic limestones is indicative of shallow water conditions on the Schwelle long after the cessation of reef growth, and therefore provides important information on the post-mortem development of the reef-complex.

The rocks of this facies are all oosparites, in places grading into oomicrites (Fig 5.23f). Crinoids, mainly marginally micritized, are by far the most important faunal constituents and commonly act as nuclei for ooids, though foraminifera (mainly <u>Endothyra</u>, but arenaceous forms also present) are also common both as nuclei, and lying free in the matrix. Fragments of brachiopods and fenestellid bryozoans are seen, but are uncommon. Microsparite intraclasts occur and they too act as ooid nuclei. Some are "riddled" by meandering silicified tubes which resemble <u>Girvanella</u> and are interpreted as such. These intraclasts, therefore, resemble the Girvanella "lumps" seen in many Dorp limestones. Silicified tubes of this type are also present within the cortex of oolitic coatings. Ooids are very common, most fossil fragments having oolitic rims.

All have been silicified (see page 399, figs 6.5, c + d.), though fine concentric and radial structures have been preserved. Broken and recoated ooids are common.

Discussion

Modern ooids occur in a very limited range of environments, often with waters supersaturated with calcium carbonate. Within these environments ooids are normally restricted to water depths of less than 2 metres, though ooids have been noted in depths of 10-15 metres (Newell et al, 1960). Shallow water ooids generally form in high energy environments, strong agitation producing multiple oolitic coating. Ooids found in deeper waters are normally only superficially coated (Bathurst, 1967b). Because of the well developed oolitic rims on nuclei and the presence of stenohaline fossils, the oolitic sediments found here are interpreted as ancient analogues of the marine oolitic sands at present forming in marginal areas of the Bahama Bank (Newell et al. op. cit.). The present location of these oosparites as fragments within deep water breccias suggests that they, like their modern counterparts on the Bahamas platform, formed close to the edge of the schwellen top (Ball, 1967).

The importance of this facies, as mentioned above, is that it provides proof of shallow water conditions on the Bieber Schwelle long after the cessation of reef-growth in the Middle Adorf.

The timing of this renewal of shoal conditions coincides with the resumption of volcanism on the northern edge of the Bieber Schwelle. * It is therefore inferred that the reduction in water depth was due directly to this volcanic activity. Such shallowing could have been brought about in either of the following ways:

(a) Lavas extruded on to the sea bottom, could have accumulated to such a thickness that the top of the volcanic pile either emerged above sea level or came close enough to sea level to allow shallow water carbonate sediments to develop. As the estimated maximum thickness of the Lower Carboniferous volcanics is around 250m. (Kegel, 1933), this would imply that the water depths at which the Upper Devonian pelagic limestones were deposited were of that order of magnitude and were thus certainly not abyssal.

(b) Local shallow water areas could have been formed by block faulting associated with volcanism. A number of large faults are present in the area; many are of Tertiary age but some are Variscan and may have been initiated prior to orogenesis. The fact that these oolitic sediments are now found only in "avalanche" breccias in deeper water off-rise environments strongly suggests that some rapid fault movements did occur about this time.

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* Volcanism is dated as CuIIβ (anchoralis zone); the breccia dated at CuIIIα , but rock fragments in the latter are probably of CuIIβ/γ age (anchoralis-bilineatus interregnum).

The second alternative is considered the more probable explanation, though the evidence now available is not sufficient to draw firm conclusions.

The oolitic sediments are equivalent in age to the youngest fissure and cavern fillings at Eberstein and Bieber (pages 248 + 265). The reopening of these fissure systems too is thought to be a result of volcanic activity. The fissure-filling sediments, however, do not show any definite characteristics of shallow water formation and could have formed at depths below those at which the oolitic limestones formed but probably at depths shallower than those at which the bulk of the pelagic limestones were deposited.

Fissures

High angled neptunean dykes (here termed "fissures") are common in the massive limestones in the quarries at Bieber [70890, 10500] and Eberstein [69150, 11850] and are intermittently seen in roadside exposures of massive-limestone between these two localities (Fig 5.24). Sediments, interpreted as fissure fillings because of their lithological similarity to undoubted fissure fillings, were found also in the mine dumps at Königsberg [68060, 11860].

Orientation and age

The fissures are generally straight sided and maintain a fairly constant orientation at any one locality. Fissures from different localities, however, differ in their orientation (Fig 5.24). The directions followed by fissures at all localities are major tectonic joint directions, but the fissures predate orogenic deformation and many were initiated at a very early stage in the Upper Devonian, as can be proved by conodonts from the sediments filling them. Because of the constancy of direction regardless of age of infilling, fissuring probably occurred in one phase, though fissures were reopened at later times.

The oldest datable fissure sediments at Bieber are almost contemporaneous with reef growth (A.triangularis zone (toIδ)). Fissuring, therefore, must have taken place during the period when the reef was actively growing or very shortly afterwards. Indeed, it is

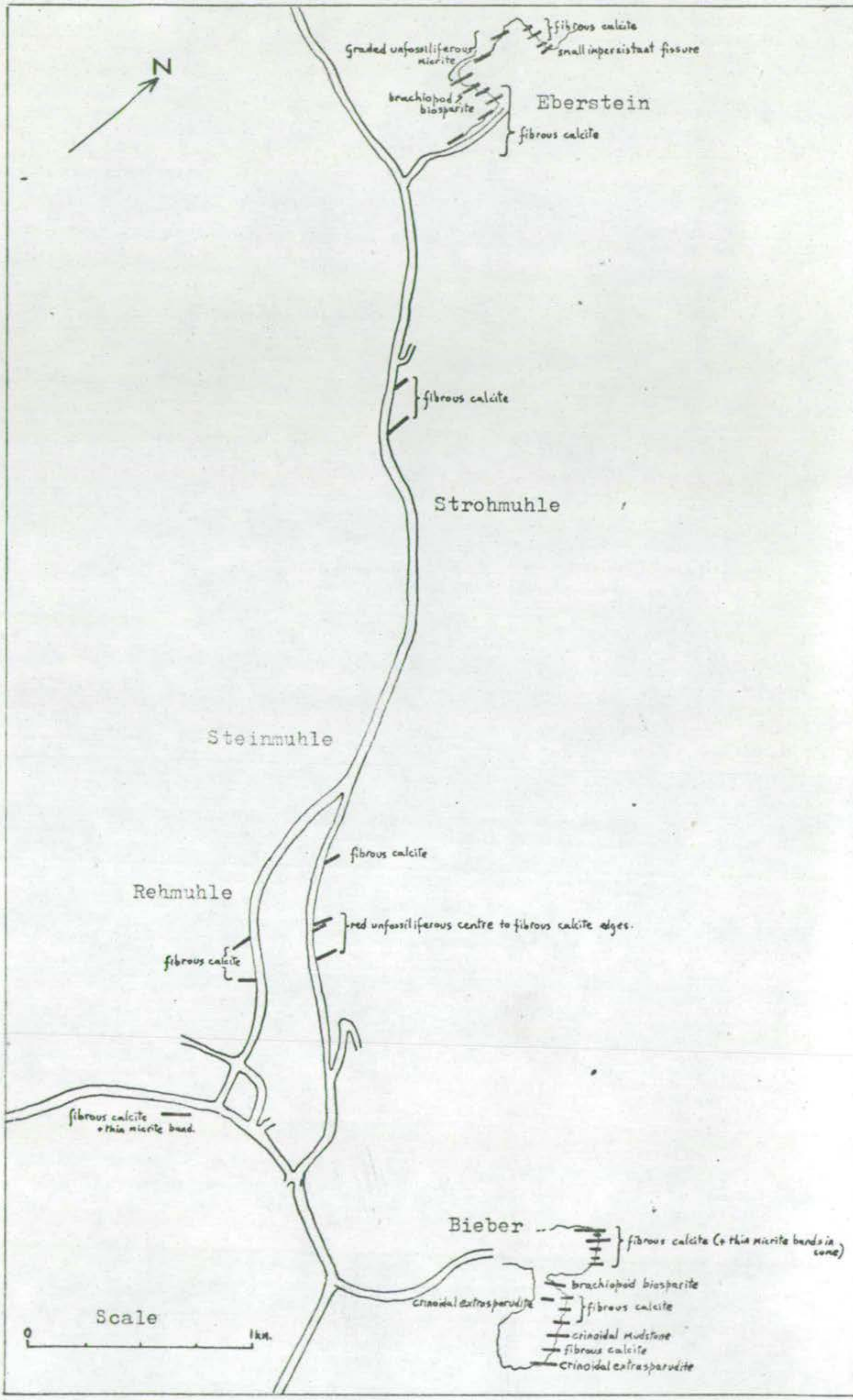


Fig. 5.24 Distribution of fissures in the Eberstein- Bieber massive limestone outcrop area.

possible that fissuring and cessation of reef growth were connected (page 364). Pelagic nodular limestones above bedded massive limestones at Eberstein are not fissured. The youngest rocks in which signs of fissuring were found were of upper gigas-triangularis age, and in these fissuring was on a small scale.

Fissuring described from other areas of the Rheinisches Schiefergebirge appears to follow this pattern (Krebs, 1972).

Size and extent

The extent of fissuring is difficult to assess as nowhere was the top of bottom of a fissure exposed, but most apparently extend vertically up quarry walls for 20 metres or so without interruption. Exceptions to this are found in small localized fissures which are seldom more than a few metres long.

Fissure fillings

No two fissure fillings are exactly the same, and some aspects of sedimentation within a fissure were, no doubt, unique to that fissure. The range of fillings, however, is limited, though many combinations of each may be present in any one fissure, either representing different pulses of sedimentation during one episode of fissure filling or different phases of sedimentation resulting from repeated opening of fissures. In the former case the sediments will normally be more or less conformable with no time gaps. In the latter case, later sediments will transect, and sometimes include fragments of, older fillings and a substantial time gap may exist between filling episodes.

Six main types of fissure filling can be recognized. The main features of these are summarized in Fig 5.32 and they are discussed in detail below.

(i) Localized irregular small fissures, less than a few metres in length, contain only material derived from the immediate surroundings (i.e. wall rock fragments and initial cement). There is no evidence that material was introduced from outside the immediate area (Fig 5.25a). These fissures are not early diagenetic features, since they transect cavities already cemented by fibrous calcite, but are definitely pre-tectonic in age as they are cut and offset by tension gashes (Fig 5.25b). No difference in age between country rock and cavity fill could be detected.

Fissures on a larger scale are more widespread. They are usually straight edged and have sharp boundaries against the country rock (Fig 5.25c). In the simplest cases, these fissures are filled by fibrous calcite crystals growing normally to the fissure walls. Such fissures are small (less than 2cm. wide). With increasing thickness they grade into "compound" fissures, in which sediment of some type is included between these crusts of fibrous calcite (Fig 5.25d).

(ii) In thin fissures (less than 5cm. normally)

Fig 5.25

(a) Near vertical small fissure partially filled with broken carbonate debris, cutting primary cavities and sheet cracks cemented by fibrous calcite, in Iberg limestone. Concordance between sediment levels in fissure and host rock indicate that fissure formation was pretectonic. Sparry calcite cement in fissure suggeste subaerial cementation. Thin section, plane polarized light, Scale bar = 1mm. Sample no. E110.

(b) Small fissure (as above)(centre vertical) offset to left by tectonic veins. Field photograph. Scale given by hammer. Eberstein, Iberg Limestones.

(c) Sharp edged fissure, cutting Stachyodes-Thamnopora biosparudites, composed of many layers of fibrous calcite crystals growing normal to fissure edge. Thin light band in centre is light pink micrite, other bands are zones in fibrous calcite. Field photograph. Scale bar = 5cm. Bieber quarry.

(d) Detail of (c) showing sharp fissure junction cutting fossils (Stachyodes) and matrix alike. Dense band on left edge of specimen is micrite, other bands are clouded and clear fibrous calcite. Hand specimen. Scale bar = 1cm. Sample no. B60.


a
b
c
d

the most common sediment found is grey-pink or red structureless micrite. In wider fissures (up to 20cm. observed width) this micrite is sometimes graded (Fig 5.26a+b), in others fenestral voids and sheet-cracks are developed within the micrite, in places so intensively that the fissure sediment consists of alternating bands of micrite and fibrous calcite (Fig 5.26c).

The general absence of fossils (conodonts were present in only one sample) may be due to introduction of micrite into the fissures through fine openings through which material coarser than silt grade could not pass, or possibly through derivation from an unfossiliferous source. Derivation from "terra rossa" forming on exposed limestone areas has been suggested as a possible source for sediments of this type (Fischer, 1964; Krebs, 1971). Whatever the source, the absence of sedimentary structures other than graded bedding and small scale deformation structures in some layers, suggests that these sediments were deposited in a calm environment.

(iii) Red micrites containing scattered ostracods and rare fragments of fenestellid bryozoa and crinoids are found in many fissures. Sedimentary structures are rare but cross bedding is sometimes present. It is possible that such structures could be due to water movement in the fissure, but may possibly represent structures formed by interaction of several cones of sediment continually building up on the

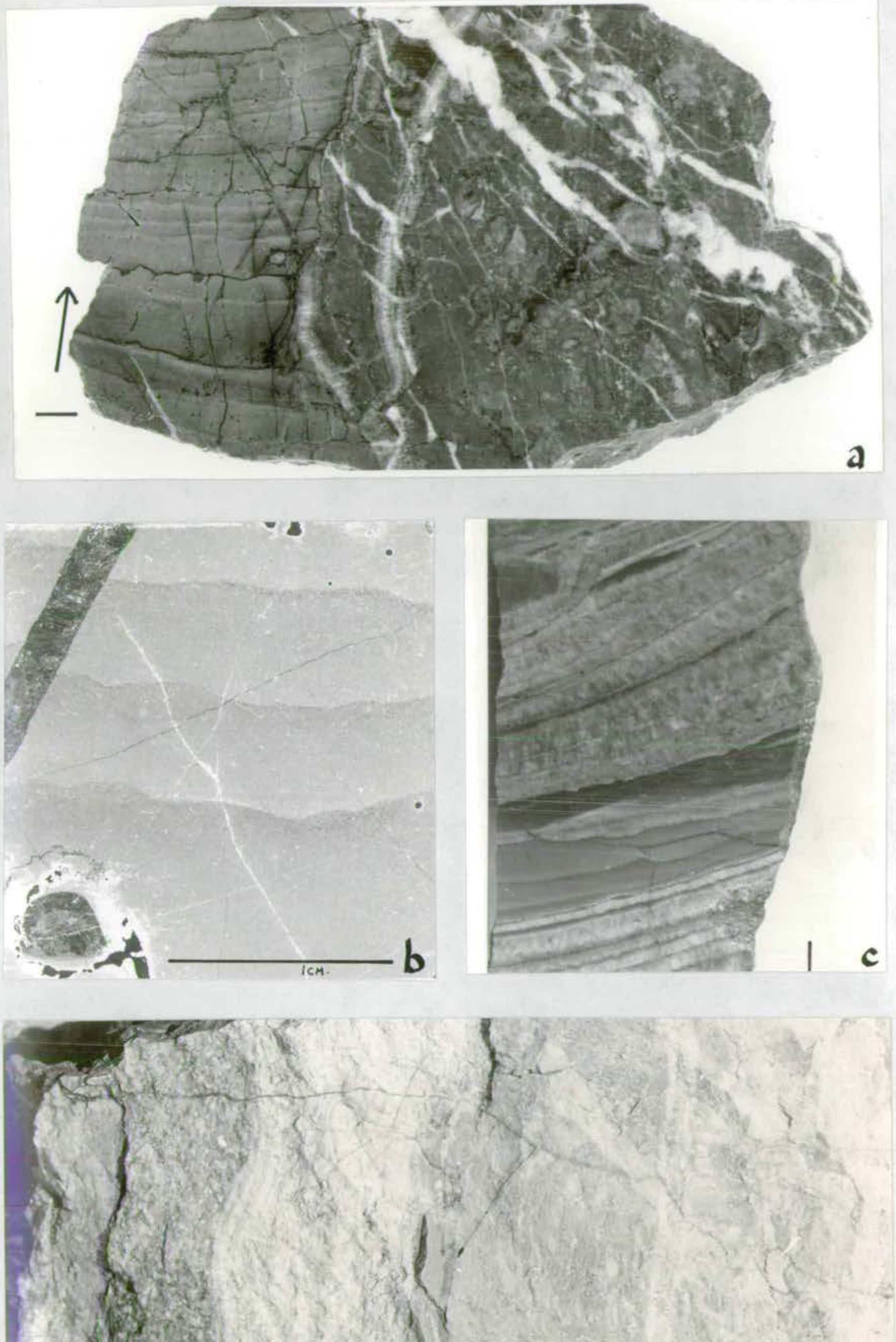
Fig 5.26

(a) Graded nonfossiliferous micrite filling large fissure in Stachyodes-Thamnopora biosparudite. Note graded bedding in fissure sediment. Both fissure and host sediment are cut by irregular tectonic veins. Hand specimen. Scale = 1cm. Sample no. E66a.

(b) Detail of above showing fine graded micritic infill. Methylene-blue stained peel, negative print plane polarized light. Scale bar = 1mm.

(c) Fissure filling consisting of alternating horizontal layers of unfossiliferous micrite (graded in places - lower centre) and fibrous calcite. Washout and small scale deformation structures are also seen in some micrite layers. This rock resembles the "Zebra-rock" fissure fillings of Fischer (1964). Hand specimen. Scale = 1cm. Sample no. E36a.

(d) Large multiple fissure, rimmed by thick fibrous calcite crusts, cutting Alveolites-stromatoporoid biosparudites. Centre of fissure filled with red crinoidal micrite. Note parallel thin fibrous-calcite filled fissure cutting lamellar stromatoporoid to the right of the large fissure. Field photograph. Scale given by pen. Bieber quarry.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fissure floor by sediment "avalanching" down into the fissure.*

(iv) Red micrites, packed with unabraded, non algal-bored, crinoid ossicles are common in younger fissures (Fig 5.26d+Fig 5.27a). The presence of still-articulated ossicles suggests that the crinoids did not undergo much transportation prior to deposition in the fissures. They possibly represent the remains of crinoid "meadows" on the limestone surface above the fissure. The depth at which these crinoids grew is not known, but the association with rare brachiopods suggests depths in excess of those in which "reef" limestones could develop, but shallower than those in which the majority of pelagic nodular limestones were deposited. Alternatively, substrate type rather than depth may have been the primary control on the growth of crinoids. If this were the case, then perhaps rocky exposures of the former reef surface provided an environment more suitable for crinoid development than the surrounding "muddy" pelagic carbonate bottom.

(v) Crinoidal micrites also form the matrix of the youngest fissure fillings (Fig 5.27b). These differ from the other fissures discussed in having very irregular wall edges, no fibrous calcite wall linings,

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* Lithologically these sediments bear a close resemblance to the nodular limestones, but lack most of the fossils commonly found in nodular limestones of equivalent age.

and in containing angular clasts of surrounding country rock, as well as fragments of former fissure-filling sediments of types (ii) and (iv) (Fig 5.27c). The irregular brecciation and rapid filling of these fissures was probably "triggered-off" by contemporaneous volcanism of CuIIβ /γ age. Though no volcanic rocks of this age occur on the southern side of the Bieber Schwelle, the higher insoluble residue content of these sediments (Appendix II) may indicate a contribution from volcanic dust. These sediments resemble, and are equivalent in age to, the Erdbach Limestones I (Krebs, 1966).

(vi) In contrast to the predominantly muddy sediments discussed above, the only other fissure sediment observed is a well-washed biosparite composed entirely of articulated and disarticulated shells of brachiopods(mainly aff. <u>Isopoma</u>, <u>Pugnoides</u>, <u>Donella</u>, in part) (Fig 5.27d). Sediment of this type was found overlying red micrites of type (iii) at Bieber. Conodonts from the biosparite gave an age of toIβ/γ , whereas those from the micrite below were of toIγ/δ age. The underlying apparently younger sediments are however cut by the overlying biosparites, indicating that the latter are in fact the younger, and that the sparse conodont fauna recovered was probably reworked from the host limestone.

Well-washed shell accumulations have been noted previously elsewhere in the Rheinisches Schieferbirge,

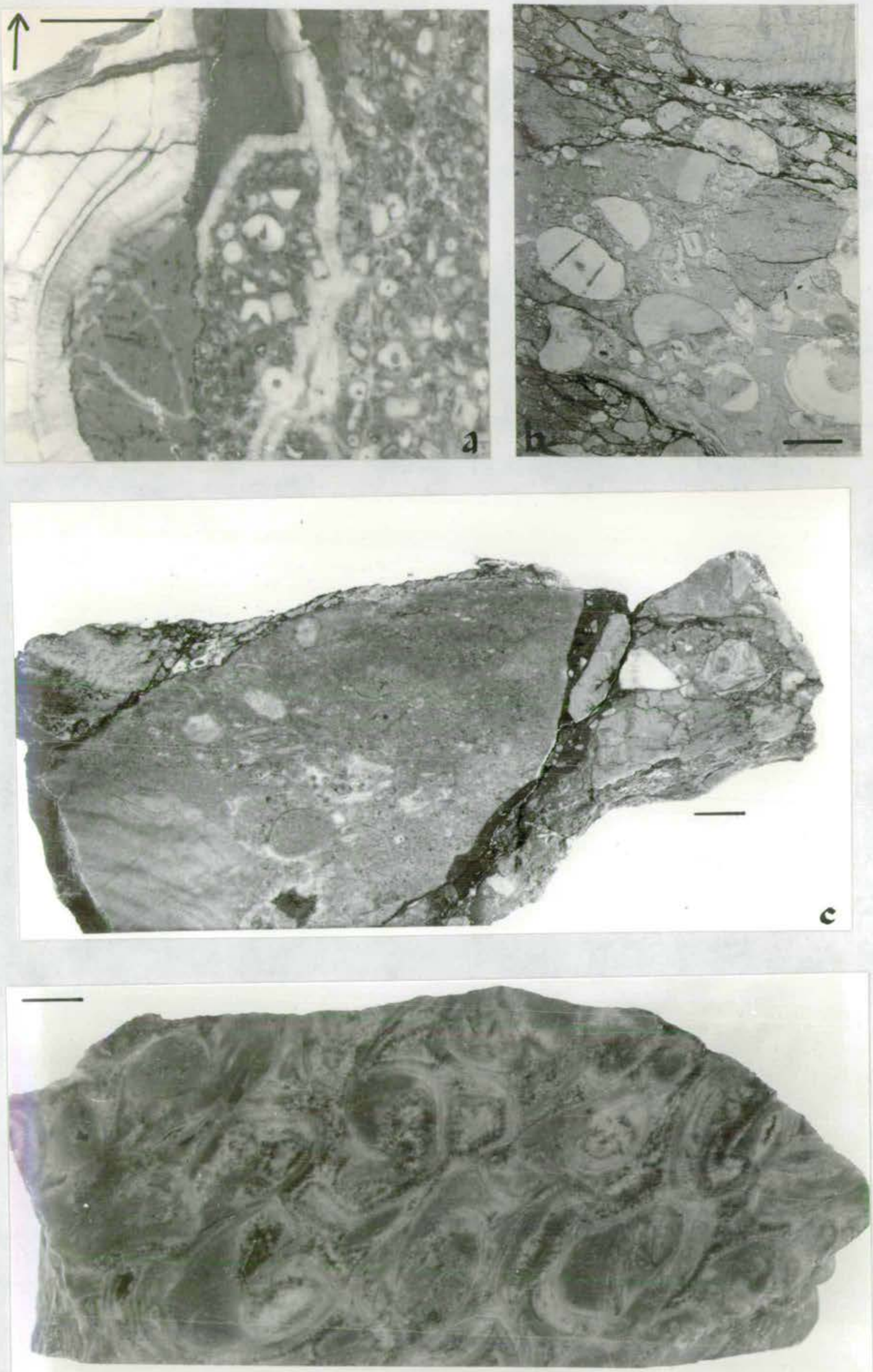
Fig 5.27

(a) Detail of Fig 5.26d, showing (from left to right) (i) multilayered fibrous calcite crust, (ii) red unfossiliferous micrite, cut by (iii) red crinoidal micrite. The latest sediment in filling is itself cut by thin fibrous spar filled fissures. Hand specimen. Scale = 1cm. Sample no. B61.

(b) Red-brown "shaly" crinoidal micrite, cut by clay seams ("flasers"), containing fragments of red unfossiliferous micrite (centre right). Methylene-blue stained peel, plane polarized light. Scale bar = 1mm. Sample taken from matrix of (c) below.

(c) Extrasparudite-type fissure filling. Poorly sorted angular fragments of Dorp limestone (centre), fibrous spar (white fragment, right) and previous fissure fillings in dark red shaly crinoidal micritic matrix (see (b) above). Note fissure sediment infill in cavity in massive limestone fragment (lower centre). Hand specimen. Scale bar = 1cm. Sample no. B63.

(d) Brachiopod biosparite consisting entirely of overpacked, articulated and disarticulated brachiopod shells. Many shells contain geopetal internal sediments (grey micrite). All pore spaces are filled by fibrous calcite. Hand specimen. Scale bar = 1cm. Sample no. B57.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but they do not always contain brachiopods. Cephalopods, bivalves, and corals have been found (Krebs, 1971, 1974).

Brachiopod fissure-fillings are, however, not uncommon in the geological record. Fillings very similar to those described here have been noted in Triassic limestones in the Alps (Fischer, 1964; Scholl and Wendt, 1971). These occurrences have been interpreted as near life-assemblages of fissure-living faunas.

In modern coral-reef environments, brachiopods are restricted to fissures and caves within the reef framework (Rudwick, 1970). It seems possible, therefore, that such environments were also preferred by at least some Devonian brachiopods. Jux and Stauch (1965) concluded that the Middle Devonian brachiopod Martinia inflata, which occurs in shell pockets within the Lower Plattenkalk of the Bergisch-Gladbach-Paffrather area, lived in just such environments within a karst-eroded limestone surface.

In the present case, well preserved articulated brachiopods are rare in most of the massive limestones. Scattered shells occur in fine calcarenites, but in coarse biosparudites only broken disarticulated shells are seen, large concentrations of well preserved brachiopods being found only in fissures (as at Bieber) and in pockets in massive stromatoporoid biolithites of similar age normally "riddled" by solution caverns (see below) (as at Eberstein). In the latter case, however, the brachiopods do not appear to belong to the same genera (and are possibly Atrypids).

Mechanism of fissure filling

Fissures similar in many respects to those described above have been interpreted as having been filled from below (Pray, 1965) and from above (Lewis, 1973). The age of the fissure sediments proves that they were filled from above.

Cause of fissuring

Fissures can be produced in limestones by two main means:

(1) karst erosion.

(2) tectonic fracturing.

Karst erosion is rejected as a primary cause of fissuring for the following reasons:

(a) the edges of most fissures are straight, sharply defined and parallel; not curved or irregular as one would expect if due to solution.

(b) fissures maintain a constant trend at any one locality. In this respect they are more like joints or igneous dykes than solution pipes.

Tectonic fracturing is, therefore, suggested as the cause of fissuring, though the possibility of later karst exploitation of these fissure systems is not ruled out, and did apparently take place in the Tertiary.

The fissures described are very like those described from the Triassic and Jurassic of the Alps (Fischer, 1964;

von Schlager, 1970; Wendt, 1971; Schöll and Wendt, 1971; Krystyn, Schäffer and Schlager, 1971), from the Devonian (Krebs, 1965, 1971; Franke, 1971; Szulczewski, 1973) and from the Silurian (Manten, 1971).

In all of the above examples, tectonic fracturing is thought to have been the primary cause of fissuring. In the case of the Triassic and Jurassic Alpine fissures, differential subsidence between a southerly basin area and northern shelf area is thought to have led to cracking of the shelf edge area with the production of fissures parallel to the shelf edge. Such a mechanism is not thought feasible for production of fissures in the studied area since those described here are radially distributed around the Bieber Schwelle schalstein area and are thus almost normal to the supposed "shelf" edge (Fig 5.24).

Volcanic updoming is thought to have caused the fissuring for the following reasons:

(a) Orientation of fissures in a radial pattern suggests a process of formation in some ways similar to that responsible for radial igneous dyke swarms around volcanic centres.

(b) There is a strong temporal relationship between cessation of reef growth and fissuring which could be due to some kind of volcano-tectonic activity capable of raising the reef above sea level. No extrusive rocks of this age are known in the Rodheim-Bieber area, though they occur elsewhere in the Lahn-Dill area (Krebs, 1966; Quado, 1965; Goldman, 1968).

(c) An equally strong temporal relationship exists between Carboniferous volcanicity and reopening of old fissure systems.

As the upper surface of the massive limestones is rarely exposed and schwellen-basin transition zones are nowhere exposed in the studied area, it is not possible to determine how much surface damage was done to the reef-complex by this fissuring. In the Carboniferous fissuring occurred immediately before deposition of oosparite clasts in the basin area to the north of the schwelle (page 70). Fragments of lithified massive limestone of both Adorf and Lower Carboniferous age are, however, found in basin areas bordering Schwellen elsewhere in the Lahn-Dill area (Krebs 1965; 1971, 1972, 1974).

Caverns

Closely associated with the high angled fissures discussed above, there also occur nearly horizontal cave systems, which, in contrast to the fissures, are not straight sided but branch through the host rock at several levels. These anastomising cave systems are here described as "caverns" though no genetic connotation is necessarily implied by this term.

Location

The caverns are found in two main areas: in Bieber Quarry where they have been found in loose blocks and only rarely in situ; and at Eberstein, where they appear to be restricted to the northern part of the main quarry occupying a zone about 30 metres thick, in which the host limestones are generally more distinctly bedded than the main mass of limestone in the quarry. This cavernous zone is overlain by well-bedded Iberg limestones and then pelagic nodular limestones, neither of which show any signs of caverns. Cavern-like sediments have also been seen in roadside exposures between these two localities at Rehmuhle* and in the area south-west of Rotestrauch, but at these localities the field relations of the sediments were difficult to study due to lack of exposure.

Size

Because of their ramifying nature, the absolute size of these caverns is difficult to assess. At Bieber they are limited in the exposures seen, being less than a few metres thick in total, but "roof pendants" and wall extensions commonly extend well into the cavern systems making them appear smaller (Fig 5.28a). At Eberstein the caverns appear to be larger with many metres occupied by cavern sediments in places. Assessment of the actual shape and size of the caverns at Eberstein is made difficult because later karst erosion (Tertiary)

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* (A sample from this locality showing dissolution features and later infilling by micrite was figured by Krebs, 1971).

has exploited these cavern systems, and many areas are now occupied by slumped masses of Tertiary sediment.

Age

The precise time when cavern formation took place is, unlike that of fissuring, difficult to assess exactly and may have been polyphasal.

At least some of the caverns are thought to have formed at an early stage after limestone lithification for the following reasons:

(i) Sediments in many of the fissures gave conodont ages contemporaneous with those of the host limestone. The conodonts may have been reworked from the limestones but no conodonts of a later age were found in such sediments.

(ii) Some of the sediments contain contemporaneous "reef" organisms (e.g. Amphipora) not rimmed by cement, suggesting that cavern filling took place while such organisms were still actively growing at the surface.

(iii) Likewise styliolinids form important constituents of one type of cavern sediment. As this group became virtually extinct in the late Adorf, an early upper Devonian age of cavern filling is indicated.

(iv) Caverns are restricted to Adorf limestones and are mainly found in the upper parts of the Dorp limestones. Similar caverns in the Alpine Triassic are likewise found near the upper boundary of reef and shelf carbonate sequences (Wendt, 1971).

(v) No caverns, other than primary voids, are found in the bedded calcarenites of lower P. gigas age lying above the Dorp limestones, nor in the nodular limestones above these.

All of the available evidence therefore indicates that, like the fissures dealt with above, initial cavern formation took place in the A. triangularis zone (toI), though these cavern channels may have been exploited, as was the case with the fissures, at several periods thereafter.

Cavern sediment fillings

Unlike the majority of fissures described above, fibrous calcite crusts are not found on the walls and roofs of sediment filled caverns. Only three sediment types are found (see also Fig 5.32):

(i) In the smaller caverns (e.g. at Bieber, Rehmuhle, Rotestrauch), the most common sediment is a pink, red or buff fenestral sparse biomicrite (Fig 5.28b). These are similar to the sparse biomicrites seen in some fissures and may be genetically related to them. In this case, however, the sediment does not lie in sharply defined calcite lined fissures, but in a series of interconnecting dissolution channels transecting host rock allochems, cement and matrix (Fig 5.28a). In some cases clear evidence of cross-cutting of earlier micritic sediments (possibly also cavern sediments) can be seen (Fig 5.28d). These infillings appear to be earlier than

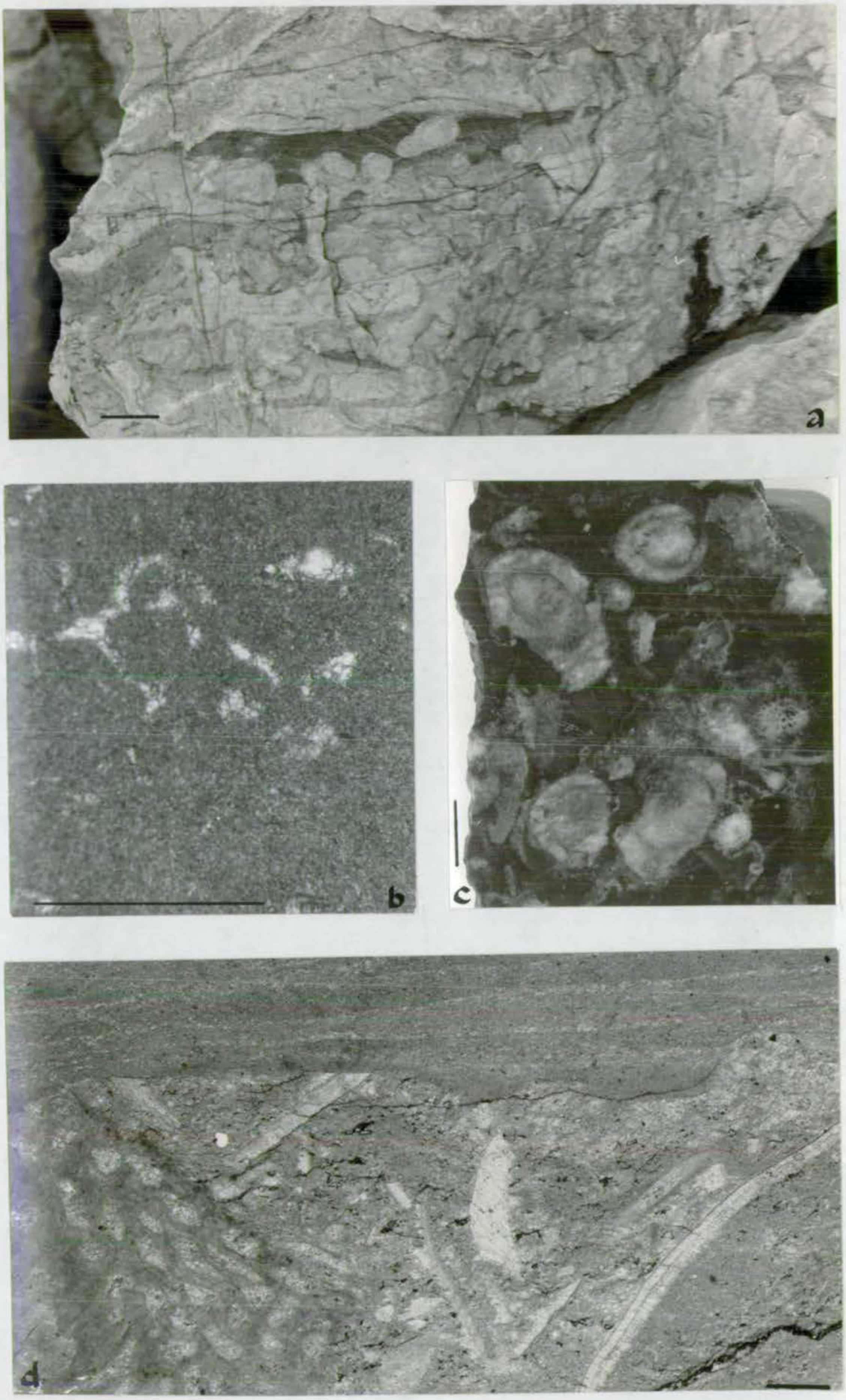
Fig 5.28

(a) Sediment filled caverns in Stachyodes-Thamnopora biosparudite. At least two generations of sediment infilling are present (i) light grey-pink unfossiliferous micrite, filling pockets in lower half of limestone block, and (ii) dark red-brown fossiliferous micrite filling large pocket in centre and smaller pockets lower in block. The shapes of the caverns here are very complex due to the protrusion of bridges and roof pendants of host rock into the cavern areas. Field photograph. Scale bar = 5cm. Bieber quarry.

(b) Detail of light grey-pink micrite from (a) showing small fenestral voids. Such voids normally occur in horizontally aligned zones. Peel, plane polarized light. Sample no. B72. Scale bar = 1mm

(c) Detail of red-brown micrite infilling solution cavities in (a) above. Both rugose corals and fibrous calcite cement rims are transected by the sediment. Hand specimen. Scale bar = 1cm. Sample no. B73.

(d) Pink unfossiliferous micrite "cutting" lithified fossiliferous biomicrite (possibly also a cavern filling). Peel, plane polarized light. Sample no. W10. Scale bar = 0·5cm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the others described below, though they do not contain conodonts.

(ii) Lying above the above-described sediments at Bieber and in caverns of A. triangularis age at Eberstein, one finds red or brown micrites or microsparites rich in cricoconarids (Fig 5.28c). These sediments also contain fragments of trilobites, ostracods, crinoids and bryozoans, and, at Eberstein, commonly pieces of Amphipora (Fig 5.31a).

Conodonts obtained from Eberstein indicate that the filling is not much younger than the host rock, both yielding conodont faunas belonging to the same zone. At Bieber the sediment yielded conodonts of Givetian to lower Adorf age, but as the faunal diversity was small, all conodonts were "conservative" forms, and the filling occurred in host limestones of undoubted lower Adorf age, cavern formation probably occurred at about the same time as at Eberstein.

At Eberstein, the relationship between the cavern-fill sediments and host rock is more complex than at Bieber. The cavern sediments here lie near the top of the Dorp limestone sequence, a few metres below the base of the Iberg limestones (Fig 5.30).

The lowest cave deposits found, like those at Bieber, occur as irregular pockets cutting organisms and cement alike (Fig 5.30A, Fig 5.31c). Above this horizon, however, a distinct irregular widespread junction is found between massive limestones and cricoconarid

Fig 5.29

Serial section through specimen from base of cricoconarid mudstone unit (B in Fig 5.30).

Black areas: "reef" organisms, not in growth position, often overturned.
White areas: fibrous calcite-filled voids.
Coarse stipple: cricoconarid microsparites.
Fine stipple: fine micritic internal sediment.

The following sequence of events can be reconstructed from the serial sections shown:

(1) "Reef" organisms were dislodged from position of growth and deposited either individually (e.g. Alveolites, lower right corner of 1-8), or as interconnected clusters (e.g. Stachyodes, upper right, 1-9) on an irregular solution-relief surface.

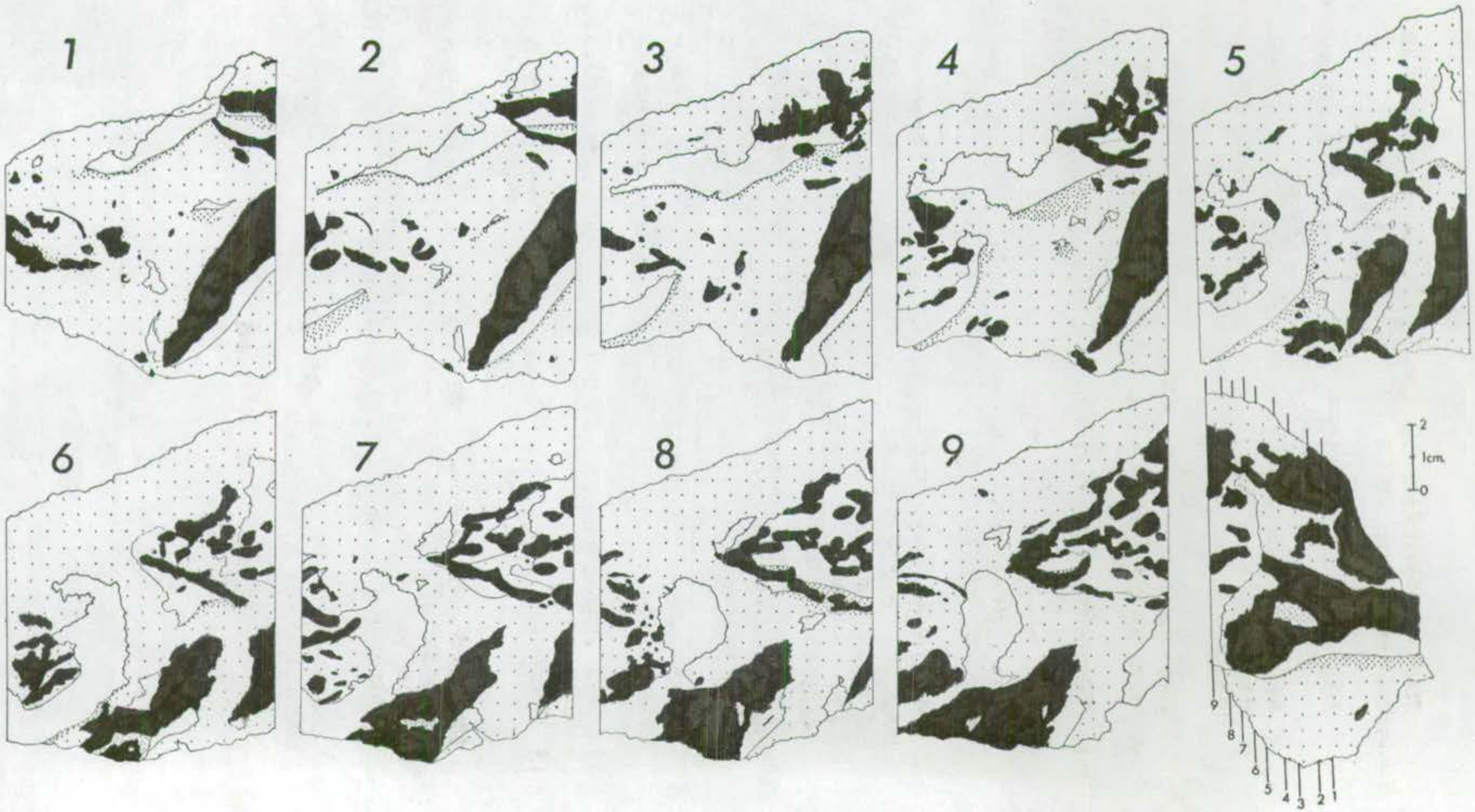
(2) Cricoconarid microsparites with stems of Amphipora, were rapidly deposited over skeletal rubble with the formation of shelter cavities below and between some large skeletons. (The presence of Amphipora in the mudstones suggests that they either passed through areas in which Amphipora grew, or that Amphipora actually grew in places on the dead-reef, and was simply incorporated within the mudstones when they were deposited).

(3) Fine suspended clay sized material settled out forming fine internal sediment in shelter voids.

(4) The cricoconarid mudstones were lithified.

(5) Renewed solution(?) of rock produced Stromatactis-like cavities, with no apparent roof support. These cavities are extremely irregular in shape and interconnect in three-dimensions. They are considered to be post-lithification features as thin almost vertical and therefore presumably cemented pillars of cricoconarid sediment separate adjacent cavities in places (see 5).

(6) Again, fine clay-sized material settled out of suspension onto the floors of these cavities.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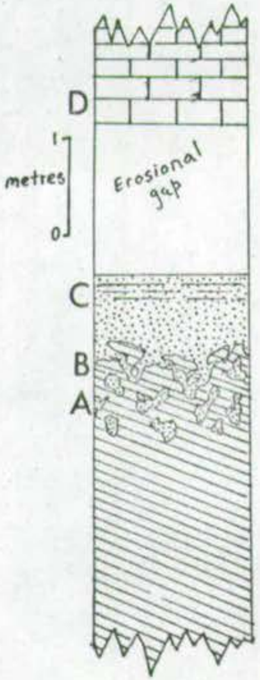


Bedded Iberg limestones.

Pure Cricoconarid microsparite zone (sediment laminated towards top).(Fig.5.31b)

Cricoconarid microsparite & disorientated 'reef' organisms. (Fig.5.29)

Junction between cavern sediment & host limestone.(Fig.5.31a)

Host limestone permeated by solution channels filled with cricoconarid microsparite.(Fig.5.31c)

Host limestone unaffected by cavern systems. (Stachyodes-Thamnopora biosparudites)

Fig. 5.30. Diagram illustrating the relation between cricoconarid microsparites and massive limestones in the northern part of Eberstein quarry.

microsparites, the latter containing many large (greater than 5cm.) dislodged unorientated stromatoporoid and tabulate coral skeletons at the base (Fig 5.30B, Fig 5.31a). The presence of calcite filled shelter cavities below skeletons indicates that the microsparite sediment was deposited after the skeletal material (Fig 5.29, 1-9). The cricoconarid sediments above this zone lack large skeletons, (though stems of Amphipora are common) and within 1 metre of the junction are finely laminated (Fig 5.30C). An erosional gap intervenes at this point below the basal limestones of the Iberg Facies (Fig 5.30D).

The cricoconarid microsparites are interpreted in this case as sediments deposited on and within an exposed cavernous limestone surface. The uppermost sediments are thought to have been deposited on a dead reef surface, the lowermost infillings within eroded cavities within the rock itself. Evidence of renewed solution after lithification of these sediments is provided by the presence of Stromatactis-cavity networks throughout the mudstones (Fig 5.29, 1-9).

(iii) Red-buff crinoidal mudstones occur as cavern fillings only in the upper part of the Dorp sequence at Eberstein, where they are found several metres below the cavern fillings described above.

These sediments are similar in many respects to the matrix of the extrasparudite fissure fillings. Crinoids, unbored by algae but often broken and disarticulated,

Fig 5.31

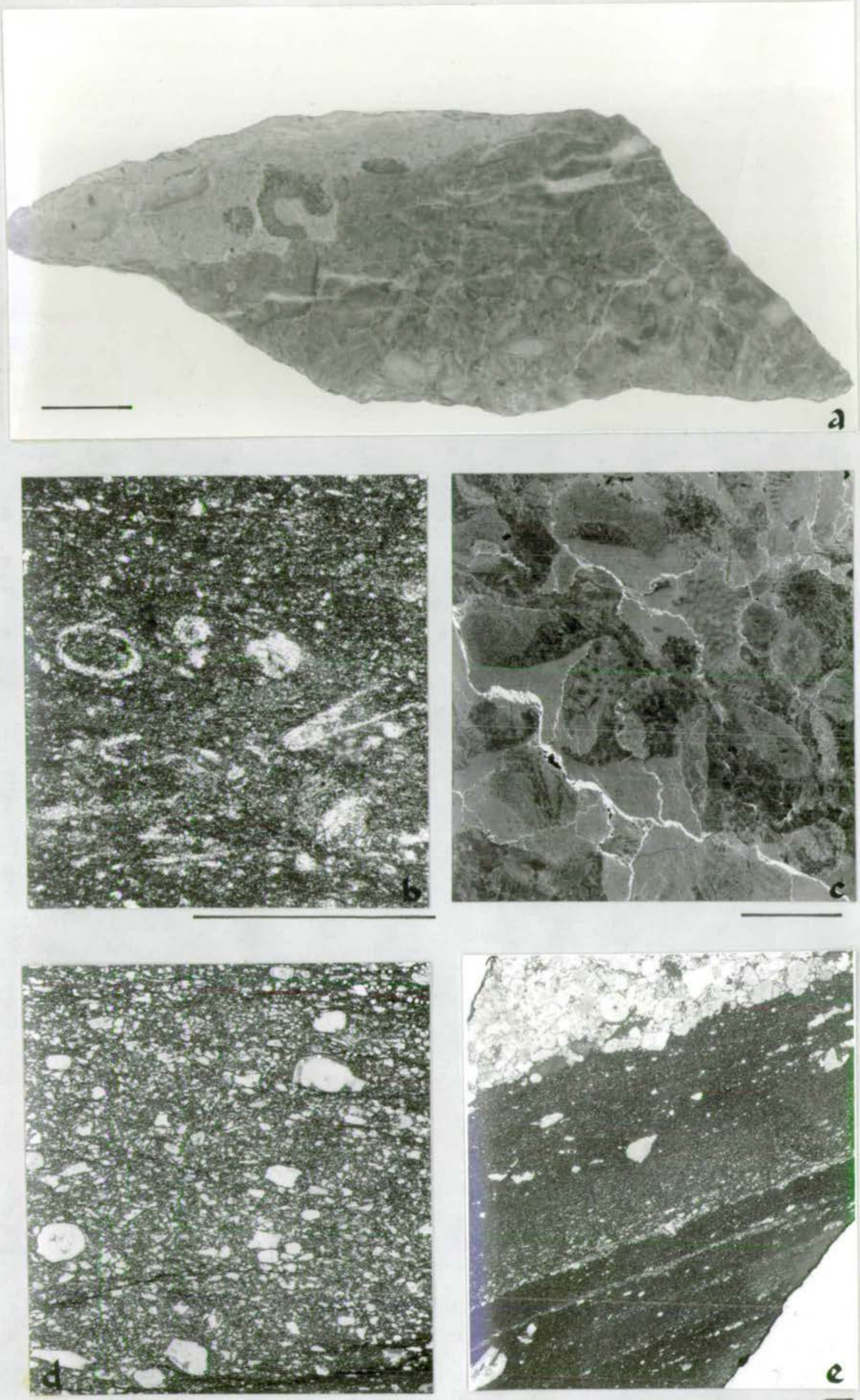
(a) Irregular solution relief between cavern filling cricoconarid mudstones above and Stachyodes-Thamnopora biosparudite below. Note the Amphipora stems at the base of the mudstones. Such stems commonly occur in this sediment but are extremely rare in the underlying massive limestones. Hand specimen. Scale bar = 1cm. Sample no. Eberstein quarry. E86.

(b) Detail of upper sediment in (a), showing cricoconarid shells in fine microsparite matrix. Near the base of this sediment, small foraminifera (Bisphaera sp. and Renalcis) have also been found. Thin section, plane polarized light. Scale bar = 1mm. Sample no. E86.

(c) Cavernous zone in massive Dorp limestones below (a). Internal sediment filled cavities cut both skeletons and fibrous calcite cement. Peel, plane polarized light. Negative print. Scale bar = 1mm. Sample no. E84.

(d) "Massive" red-brown crinoidal mudstone composed almost entirely of broken crinoid fragments in a fine "shaly" calcareous matrix. Methylene-blue stained peel, plane polarized light. Scale bar = 1cm. Sample no. E81.

(e) "Graded" red-brown crinoidal mudstone. The layer of overpacked crinoid ossicles (top) marks the base of a "graded" unit, lying on top of several other graded units. Methylene-blue stained peel, plane polarized light. Scale bar = 1mm. Sample no. E81.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are virtually the sole faunal constituents, though conodonts do occur also. Small quartz grains are found sporadically in the matrix which is of clay size, varying from dominantly calcitic near the base of the cavern complex to shaly (or tuffitic(?)) towards the top. The lower sediments are of rhomboidea age (toIIβ) (the same as lithologically similar crinoidal fissure fillings at Bieber) and are poorly bedded (Fig 5.31d). The uppermost sediments contain conodonts of the bilineatus-anchoralis interregnum (CuIIβ/γ) and are thus equivalent in age to the extrasparudite fissure fillings found at Bieber. These sediments are often conspicuously graded (Fig 5.31e) indicating that filling occurred in short pulses rather than in one filling episode.

Thin cracks in the limestones below these cavern systems are often filled with the fine matrix of the above sediments. Some of this material has even managed, in places, to filter down into unfilled void centres within the Dorp limestones, giving rise in some cases to mixed conodont faunas reflecting both the age of the limestone and of the fillings (Specimen E69a, Appendix III).

Cause of Cavern formation

The origin of the caverns in the Rodheim-Bieber limestones is thought to be intimately linked with the origin of the fissures for the following reasons:

(i) phases of fissure formation and filling are exactly matched in time by phases of cavern filling.

(ii) the sediment types found in fissures are also found in caverns.

The general succession of fissure fillings through time is also seen in cavern fillings.

Either the caverns were formed in the same manner as the fissures (i.e. tectonically) or they were formed by karst solution of the limestones at the time of fissuring.

A purely tectonic origin is here rejected because the caverns show none of the structural characteristics shown by the fissures:

(a) they are not"clean-cut" straight edged features.

(b) they show no parallelism in their development. Rather their irregular ramifying form and the manner in which the caverns cut irregularly across cements, matrix and in some cases fossils, suggests that solution was the prime force in their formation. Furthermore, when field relationships, between cavern fillings and the host rock are seen, they are seen to be solution and not structural features.

The temporal coincidence between fissure and cavern formation therefore is thought not to be due to a single origin for both, but that both resulted from the same primary cause (i.e. "igneous activity"), the fissuring directly, through fracturing of the rock, the cavern formation indirectly, through solution of the limestones

		FISSURES						CAVERNS		
		Fragmental sparite	Non. foss. micrite	Sparse biomicrite	Crinoidal micrite	Extra-sparudite	Biosparite	Sparse biomicrite	Fossiliferous biomicrite	Crinoidal calc. mudstone
FAUNA	Brachiopods	—	—	—	v.rare	—	common	—	v.rare	—
	Bivalves	—	—	—	—	—	—	—	v.rare	—
	Ostracods	—	—	rare	v.rare	—	—	rare	rare	—
	Trilobites	—	—	—	—	—	—	—	rare	—
	Crinoids	—	—	v.rare	v.common	v.common	—	—	rare	v.common
	Bryozoans	—	—	v.rare	—	—	—	—	v.rare	—
	Amphipora	—	—	—	—	—	—	—	common	—
	Renalcis	—	—	—	—	—	—	—	rare	—
	Cricoconarids	—	—	v.rare	—	—	—	—	v.common	—
	Conodonts	v.rare	—	v.rare	rare	rare	v.rare	—	rare	rare/com.
INORGANIC COMPOSITION	Pellets	—	—	—	—	—	—	rare	—	—
	Ooids	—	—	—	—	—	—	—	—	—
	Intraclasts	—	—	—	—	—	—	—	rare	—
	Quartz silt	rare	—	—	—	—	—	—	—	rare
	Clay minerals	—	—	—	—	rare	—	—	—	common
	Fragments of country rock	common	—	—	—	v.common	—	—	—	—
	Fragments of fissure fillings	—	v.rare	—	common	common	rare	—	—	—
	Matrix or cement	spar	micrite	micrite	micrite	micrite	spar/mic.	micrite	micrite	st.micrite
TEXTURE	Graded bedding	—	common	—	—	—	—	—	—	common
	Cross bedding	—	—	common	—	—	—	rare	—	—
	Sheet cracks	—	common	common	—	—	—	common	rare	—
	Sorting	poor	v.good	good	poor/good	v.poor	poor	good	poor/good	v.poor
SIZE		< 2cm.	<1cm. to >10cm.	upto 60cm.	upto 30cm.	upto 50cm.	upto 60cm.	size indeterminate - anastomizing networks		
AGE		same as host rock	to Iγ (?)	to Iγ	to IIβ	Cu IIβ/γ	to Iδ/γ	?	to Iγ	Cu IIβ/γ
Similar sequence seen in normal sequence elsewhere		———	pelagic limestones (?)		———		brachiopod beds at Eberstein etc c.f. page .	pelagic limestones (?)		———

Fig 5.3 Summary of faunal, floral and lithological features of fissure and cavern sediments.

resulting from emergence and exposure of the reef due to updoming of the schwelle.

Origin of cavern sediment

In many respects the sediments found in caverns differ little from those found in fissures of equivalent age and a common source is likely. Two elements are met with in cavern sediments which are not seen in fissure sediments, however, and warrant further discussion.

(i) the tuffitic matrix of the Carboniferous crinoidal mudstones

Deposition of lower Carboniferous fissure and cavern fillings took place immediately after a period of active vulcanicity, and it is likely that the tuffitic matrix may have been derived from erosion of volcanic tuffs. The eroded material was then transported into the cavern system picking up crinoids in its path. The repeated graded bedding seen in these deposits may indicate repeated spasms of sediment influx.

(ii) the cricoconarid microsparites

Faunally and lithologically these sediments most resemble the pelagic limestones of the same age deposited in the Western edge of the schwelle, but differ mainly in the common inclusion of Amphipora stems, Amphipora is generally most common in"back-reef" environments and is rare in "fore-reef" situations. This sediment, therefore, probably marks the breaching of the "barrier" between "off-reef" and "back-reef" areas, and is interpreted as a pelagic sediment swept onto the reef area after cessation of active reef growth.