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ASPECTS OF PELAGIC SEDIMENTATION IN THE DEVONIAN OF WESTERN EUROPE

by

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Abstract

During the Devonian of the Rhenish geosyncline (sediments exposed in the Rheinisches Schiefergebirge and Harz Mountains, Central Germany) condensed pelagic limestones, locally rich in cephalopods (Cephalopodenkalk), accumulated on submarine rises (termed Schwellen). These limestones pass laterally into shales with nodules and nodular limestones commonly involved in sedimentary slumping or reworking which were deposited on the slopes of the Schwellen. Silty shales, locally with turbidites were deposited in the deeper water areas (Becken) between the rises. The cephalopod limestones occur in three situations, above basement rises (geanticlines) submerged reefs and volcanic ridges. The Schwellen facies is present elsewhere in Europe, particularly during the Upper Devonian and has been examined in S.W. England and the Montagne Noire (S. France).

The depth of deposition of the pelagic limestones probably did not exceed a few hundred metres and in some cases was 50 m or less. There is no evidence of emersion and all early diagenetic events took place subtidally. The basinal sediments probably accumulated at depths in the region of 1000 m.

The Schwellen limestones are fine grained carbonates with a dominantly pelagic fauna (cephalopods, thin-shelled bivalves, conodonts and cricoconarids) which have suffered extensive recrystallization and are now mostly homogeneous microsparites. Laminated carbonates and thin terrigenous units occur locally, the latter indicating deposition from low density suspension currents (or possibly nepheloid layers). Fossil concentrates of thin-shelled bivalves, cricoconarids and ostracods also indicate current activity.

Lack of compaction and certain sedimentary structures suggest early lithification of the cephalopod limestones and hardgrounds indicating synsedimentary cementation are locally developed. Planar corrasional hardground surfaces cut cavity-fill cements and skeletal material. Radiaxial calcite filling the truncated cavities has nucleated from the erosion surface indicating replacement of an earlier cement.

Cryptohardgrounds having surfaces with an irregular relief and showing evidence of subsolution are encrusted by arenaceous foraminifera. Cricoconarid microcoquinas were cemented early through syntaxial overgrowths. The fibrous overgrowth crystals, showing some similarities with radiaxial calcite, are a replacement of an early acicular cement or the host sediment.

Sheet cracks in the Schwellen limestones filled by microsparitic internal sediment and radiaxial calcite are of variable size and shape and are considered to have formed by shear-failure. Neptunian dykes also occur.

Ferromanganese encrustations associated with encrusting foraminifera (from the Montagne Noire) are depleted in iron, manganese and nickel relative to modern manganese nodules. Chemically the Schwellen limestones are comparable with Recent pelagic sediments and are significantly different from limestones of other facies in being low in magnesium, high in iron and manganese. Differences exist between the Devonian Schwellen limestones, slope and basinal nodules; the slope sediments with a more variable chemistry, tend to be enriched in magnesium, iron and manganese. The enrichments are attributed to the effects of the slope and movement of connate waters.

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CHAPTER 1

Introduction

The Upper Devonian sediments of the Rhenish geosyncline * exposed in the Rheinisches Schiefergebirge and Harz Mountains, Central Germany, are essentially pelagic of two contrasting types. These are condensed fine grained limestones, locally rich in cephalopods (Cephalopodenkalk) and much thicker shales often rich in pelagic ostracods (Cypridinenschiefer). These two rock-types, deposited at the same time, may crop out within a few kilometres of each other. In 1925, Hermann Schmidt suggested that the condensed limestones accumulated on submarine rises (termed Schwellen) and that the ostracod shales were deposited in the deeper water areas (Becken) between the Schwellen. Rabien (1956) described two types of submarine rise, high Schwellen supporting reefs and shallow water carbonates and low Schwellen where the deeper water cephalopod limestones accumulated. The Schwellen limestones and ostracod shales representing pelagic sedimentation belong to the Hercynian facies of German geologists (Erben, 1964).

Schwellen limestones, typically with numerous tectonic stylolites and shale streaks, pass laterally into nodular limestones and shales with nodules which were deposited in the slope region. The slope sediments have commonly been involved in sedimentary slumping and reworking. The basinal deposits are mostly silty shales, locally with turbidites.

* The term geosyncline for the trough of deposition in the Rhenohercynian zone of the Variscan fold belt has been applied by many authors (e.g. Aubouin, 1965; Krebs, 1968a; Rutten, 1969) and is used in this thesis.

Rich faunas of the Schwellen limestones, mostly of pelagic organisms, have enabled zonal schemes to be erected for ammonoids and conodonts and to a lesser extent for trilobites. Much detailed stratigraphical work on the Devonian of the Rheinisches Schiefergebirge and Harz Mountains has been published using these faunal elements. Advantage has been taken of the German literature on this which has enabled the sedimentological and geochemical work presented in this thesis to be undertaken and facilitated the placing of samples and sections stratigraphically within the Devonian.

Outside the Rhenish geosyncline Devonian cephalopod limestones occur in S.W. England, Montagne Noire (S. France), the Pyrenees, the Eastern Alps, East Germany, Bohemia, the Urals and North Africa.

Aim of the thesis

This thesis is concerned with the Schwellen, slope and basin sediments and the aim is towards an understanding of their sedimentary and diagenetic environments. These sediments have not previously been examined sedimentologically in detail and their fabrics, sedimentary structures and diagenesis are described. Evidence is presented concerning the conditions and depths of deposition. The chemistry of these sediments is discussed as a result of atomic absorption work and comparisons made with Recent pelagic sediments and ancient carbonates of other facies. Ferromanganese nodules were discovered during the course of this work and these are described and their sedimentological significance discussed. The occurrence of Schwellen sediments within the Rhenish geosyncline in space and time is discussed, with their stratigraphical and structural association.

Field Work

The Devonian pelagic sediments have been examined in Germany, S. France and S.W. England. In Germany, the author was attached to the Geologisches Institut, Göttingen University, during the summer of 1969 and April 1970, and examined sections recommended by Professor D. Meischner in the Harz Mountains and Rheinisches Schiefergebirge. The Schwellen limestones were examined in detail in the N.W. Harz where several sections show the transition to the basin sediments. Details of the localities examined with map and literature references are given in Appendix 1 . Certain areas (Langenaubach-Breitscheid (Dill Syncline), Lahn Syncline and the Attendorn region) and certain horizons (Kellwasser limestones) were not examined where German geologists are working (Professor Krebs and his students and geologists from Marburg University).

A month was spent in the Montagne Noire, S. France, in July 1970 when the Upper Devonian griotte was examined. A note on the presence of ferromanganese nodules was published recently and is appended. A visit was also paid to the Spanish Pyrenees where similar rocks are exposed.

In S.W. England, the Upper Devonian basinal sediments were studied in the Padstow region, N. Cornwall during September and October 1968, and January 1969. A paper on crinoidal turbidites occurring in this area was published as a result of this work (Appendix). In South Devon, the Schwellen facies has been examined at Chudleigh, near Newton Abbot and deeper water shales at Saltern Cove (March 1969; May 1970). Research in these two areas in South Devon with P. van Straaten (Göttingen) (October, 1969; January, 1971) using conodonts for accurate dating has been published and is in press. These papers are also placed in the Appendix.

General Techniques Used

In the field, sections have been measured and described and where the age of the sediment was not known conodonts were obtained from acetic acid treatment. Apart from disused quarries, normally providing continuous outcrop, most exposures are along forest road cuttings which are frequently overgrown or badly weathered. The fine grained nature of the sediments studied, the relative paucity of sedimentary structures at many localities and weathering of the limestone have necessitated the collecting of numerous samples for examination in the laboratory. For geochemical analyses an unweathered sample was collected, charateristic of the lithology for each Upper Devonian stage. Each sample is shown on the section for that locality (eg ⑮).

Petrological studies of samples collected were made in Reading with polished surfaces, acetate peels and thin sections (all made by the author). Some limestones were examined with the electron microscope to look at the finer details of the carbonate fabrics. A selection of samples from the Schwellen, slope and basinal sediments was analysed by atomic absorption spectrophotometry for the elements calcium, magnesium, strontium, iron and manganese. X-Ray diffraction analyses were undertaken to ascertain the minerals present and insoluble residues were also determined. Ferromanganese nodules were found in the pelagic limestones from the Montagne Noire and these were analysed by X-Ray fluorescence. The inter-element relationships were established with the electron microprobe.

Further Work

This thesis represents the first attempt at a detailed sedimentological study of the Schwellen facies. Many aspects require further study. In particular, the geochemical data presented here although showing that the pelagic limestones can be distinguished from other limestones is only a reconnaissance. Further analyses are required with additional elements when it will be possible to make better comparisons between the facies and with other published work. Further more detailed palaeoecological work of certain aspects of the fauna might also be expected to yield valuable information on the environment of deposition. More work on the fine details of carbonate sediments can be made with the scanning electron microscope (the transmission electron microscope was used in the work here) when it may be possible to determine the nature of the original sediment, and perhaps find relict organic structures.

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CHAPTER 2

The Rhenish Geosyncline and the Occurrence of Devonian Pelagic Sediments

The Rhenish geosyncline was a depositional trough from the Lower Devonian until the Lower Carboniferous and represents the northern part of the Variscan geosyncline which covered much of Europe. The Old Red Sandstone Continent and the Brabant Massif formed the northern margin. It was bounded to the south by a geanticline (the Mitteldeutsche Schwelle, Fig. 2.1A, Brinkmann, 1948), a land ridge for much of the Devonian and Carboniferous, which through uplift and erosion provided the synorogenic flysch of the Rhenish geosyncline. The Thuringian basin, where similar sediments were deposited, was situated south of the geanticline and bordered on its southern margin by the German - Bohemian Island, formed by the Erzgebirge and Fichtelgebirge, Fig. 2.1A. To the west deposition in the Rhenish geosyncline was continuous with that in the Ardennes. The Devonian and Carboniferous development in S.W. England is very similar to that in Germany.

The sediments of the Rhenish geosyncline are exposed in the Rheinisches Schiefergebirge and Harz Mountains (Fig. 2.1B). The thickness of sediment varies across the strike and a total of about 6 to 8 km were deposited between the Lower Devonian and Lower Carboniferous (Kegel, 1949; Fig. 2.2). The geosyncline was divided into north-western and south-eastern basins (here termed Sauerland and Taunus basins respectively, Fig. 2.3), by a mid-geosynclinal ridge upon which sediments of various types were deposited. The deposits of this ridge, only 20 km wide, are exposed in the Kellerwald where the ridge has been termed

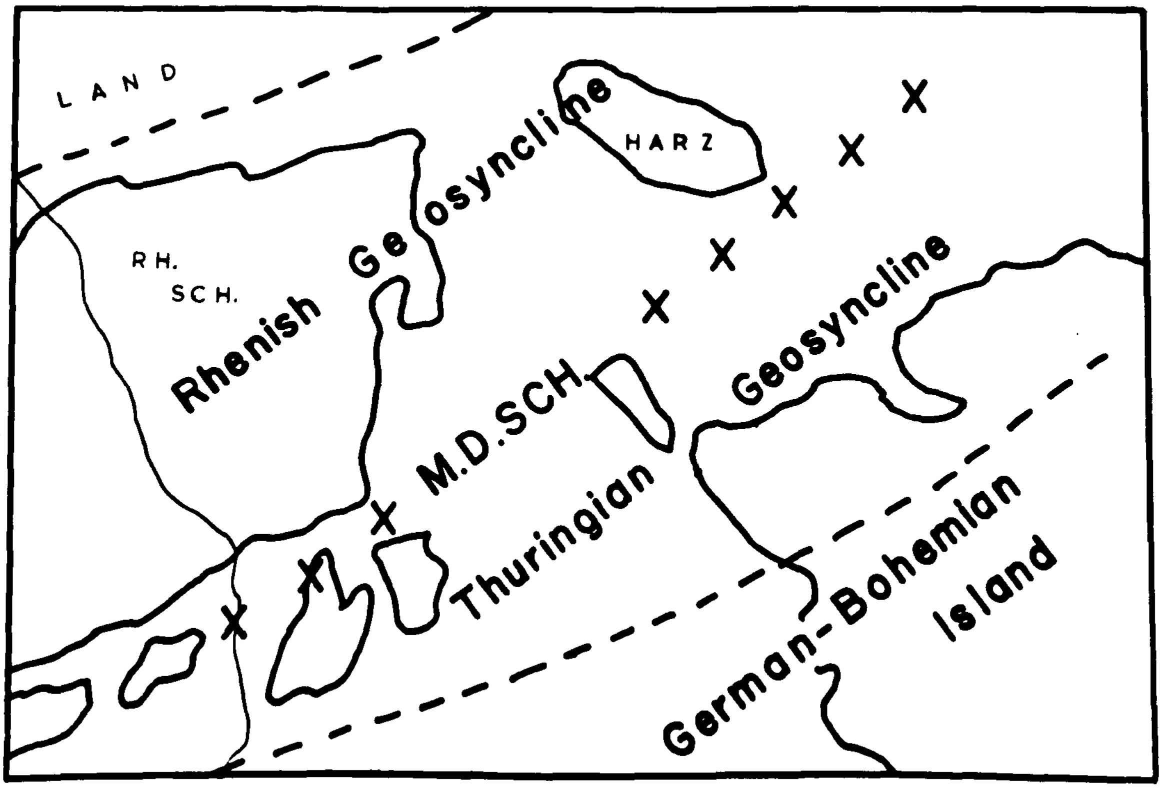


Fig. 2.1A Sketch map showing location of Rhenish geosyncline within Central Europe. The Mitteldeutsche Schwelle (M.D.Sch.) separated the Rhenish from the Thuringian geosyncline. (Rh. Sch. = Rheinisches Schiefergebirge).

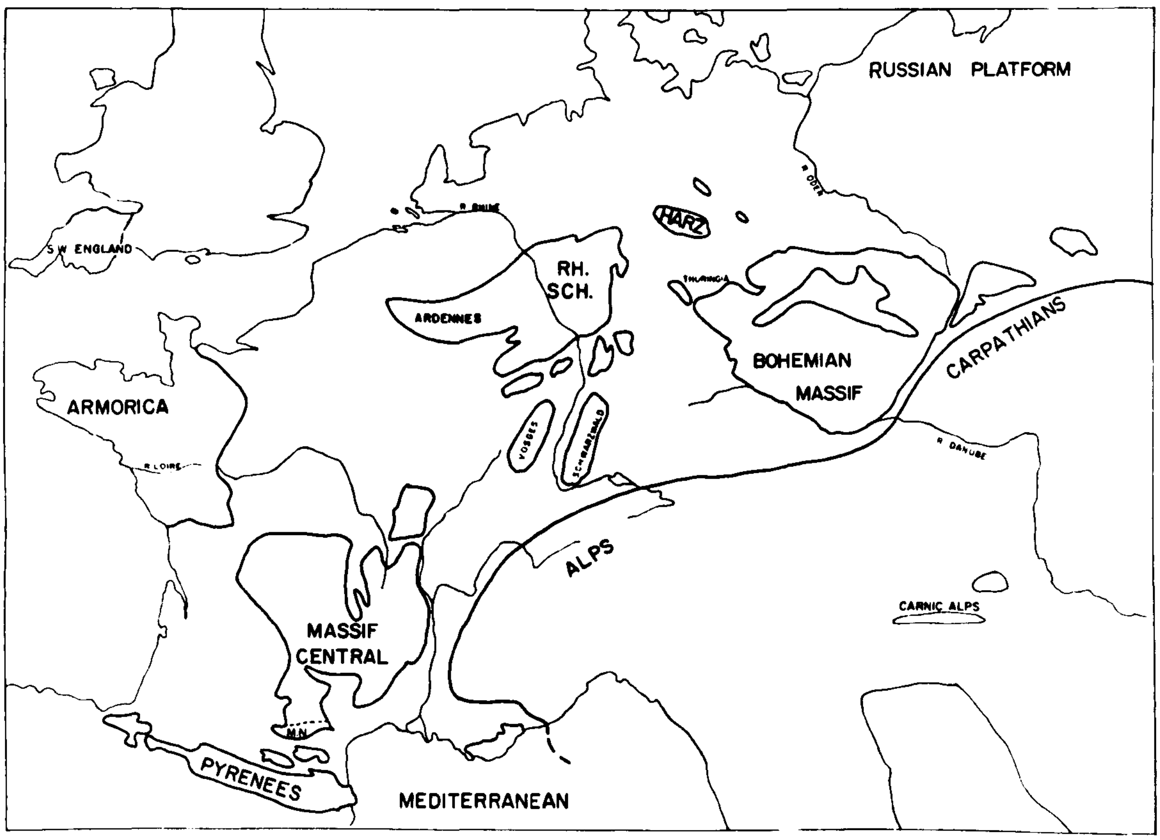


Fig. 2.1B Map of Europe showing location of Hercynian areas mentioned in the text.

(RH.SCH. = Rheinisches Schiefergebirge,
M.N = Montagne Noire)

DEVONIAN OF RHEINISCHES SCHIEFERGEBIRGE

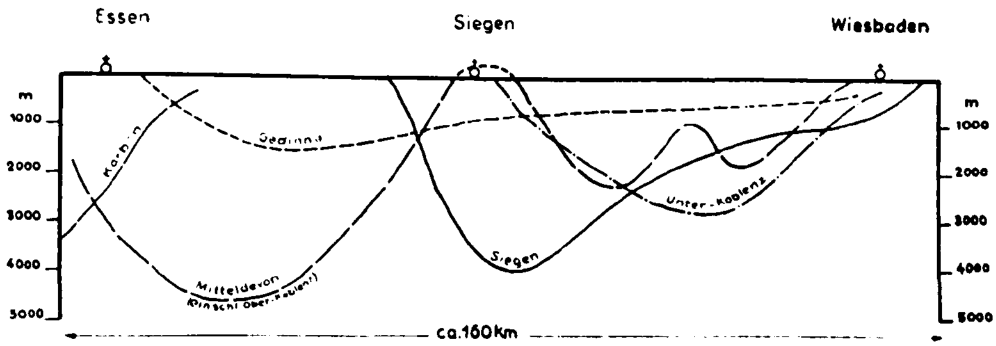


Fig.60. Composite stratigraphic section across the Rheinisches Schiefergebirge, from Essen to Wiesbaden, showing variations in thickness of parts of the Devonian and of the Carboniferous. (After KEGEL, 1948.)

Fig. 2.2 Section across the Rheinisches Schiefergebirge showing variations in thickness of the Lower and Middle Devonian (From Kegel, 1950).

the Kellerwald - Grossschwelle (Meischner, 1968). On this ridge (Fig. 2.7, p.31), smaller ridges (Ense, Haingrube and Steinhorn Schwellen) were active at various times upon which pelagic carbonates accumulated. These sediments form narrow belts of facies along the strike (SW - NE) and although they may only be a few kilometres across they are seen again in the Harz (200 km away).

Section 2.1 Volcanism

Three main phases of volcanism occurred in the Rhenish geosyncline. 1) During the Emsian (upper lower Devonian) vast quantities of keratophyres (lavas and tuffs) were extruded, particularly in the Attendorn region (Fig. 2.4, Sauerland basin). 2) Basaltic spilites (mostly pillow lavas) and intrusives with subordinate andesites, picrite basalts and serpentinites were developed at the end of the Givetian and in the lower Frasnian. Most volcanic activity took place in the Lahn and Dill synclines (Fig. 2.4) and in the north-eastern part of the Sauerland basin. 3) Extrusive and intrusive basaltic spilites were developed during the Lower Carboniferous, before the main flysch phase. Most activity took place in the Lahn and Dill regions again. (A similar volcanic history occurs in S.W. England, Middleton, 1960). The distribution of volcanics is shown in Fig. 2.3.

Section 2.2 Tectonism

The Rhenish geosyncline according to Krebs (1968a) and MacGillavry (1970) shows two main phases, a tensional phase followed by a compressional phase. During the tensional phase vertical crustal movements allowed subsidence and the formation of fault-bounded basins (accompanied by acid and later basic volcanism). The compressional phase, which affected the Taunus basin first, led to uplift and erosion of the Mitteldeutsche Schwelle, and the formation of the synorogenic flysch.

The structure of Hercynian Middle Germany diverges from the Mitteldeutsche Schwelle (Brause, 1970) with overturning of major folds

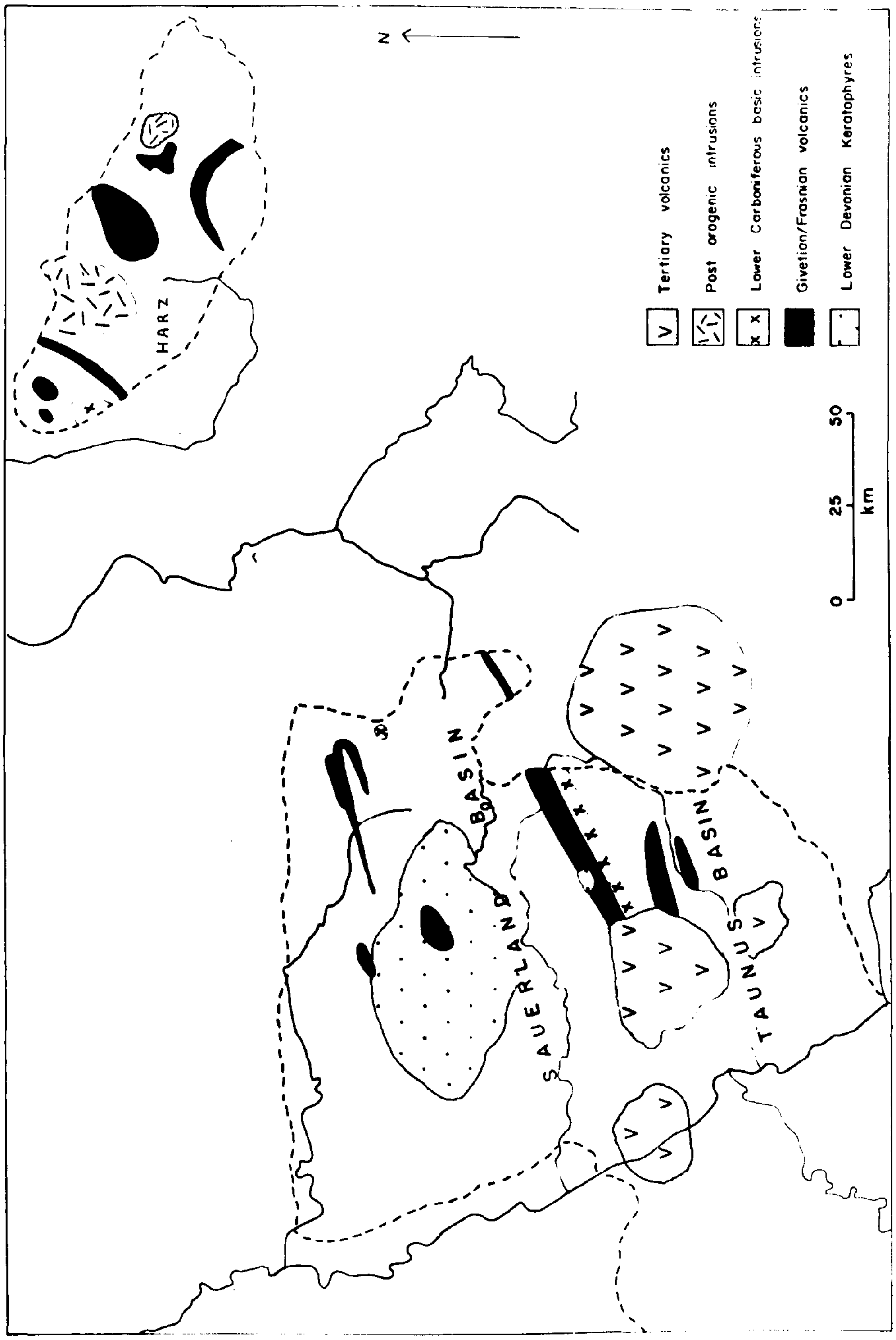


Fig. 2.3 Sketch-map showing the distribution of igneous rocks in the Rheinisches Schiefergebirge and Harz Mountains (after Schmidt and Plessmann (1961), Kockel (1958) and Möbus (1966)).

directed away from this former land ridge. Thus in the Rheinisches Schiefergebirge folding is overturned to the north-west. Southward facing folds occur on the south side of small stable blocks within the Rhenish geosyncline (e.g. on the south side of the Siegener block, Rutten, 1969). The Rhenish geosyncline was folded during the Lower Carboniferous into a number of anticlinoria and synclinoria which pitch towards the north-east. The subsidiary folds have wavelengths of about 10 km. Cleavage dips to the south-east. Schuppen tectonics (high angle thrusts) is usually associated with the former ridge areas; for example in the Kellerwald (Meischner, 1968). Major overthrusts and nappe structures are absent in the Rheinisches Schiefergebirge and Harz.

The sediments in the Rheinisches Schiefergebirge show very little metamorphism and do not exceed chlorite grade. A metamorphic belt is developed just to the north of the Mitteldeutsche Schwelle in the Taunus region, southern margin of the Rheinisches Schiefergebirge (Stenger, 1961) and in the south-eastern part of the Harz (Reichstein, 1964).

Late orogenic and post orogenic volcanism occurred particularly in the vicinity of the Mitteldeutsche Schwelle (Schwab, 1970). Post orogenic granites were intruded in the Harz (e.g. Brocken granite) and are thought to exist below some parts of the Rheinisches Schiefergebirge. The Harzburger gabbro is also post orogenic.

Section 2.3 Depositional history

Sedimentation in the Rhenish geosyncline can be divided into five phases. (Details of the sediments can be found in Bogdanoff et al. (1962), Kockel (1958), Schmidt and Plessmann (1961), Möbus (1966) and Meischner (1968)).

1. Deposition of a thick terrigenous sequence in rapidly subsiding troughs, with detritus derived mostly from the Old Red Sandstone Continent to the north (Lower to Middle Devonian).

2. Formation of shallow water carbonates and reefs along the margin of the shelf bordering the northern continent and on volcanic rises within the basin. Carbonate turbidites, derived from the 'reefs' common in the basinal sediments (Givetian to middle Frasnian).
3. Pelagic sedimentation over much of the geosyncline with ostracod shales in basins and cephalopod limestones on submarine rises (Schwellen and Becken facies) (Upper Devonian and Lowest Carboniferous (Gattendorfia Stufe). More uniform black shales and cherts in Lower Carboniferous.
4. Flysch sedimentation, deposition of greywackes derived from the Mitteldeutsche Schwelle. Mainly Lower Carboniferous, but began in the south in the Middle Devonian.
5. Molasse developed only in the most northern part of the Rhenish geosyncline. Much of the geosyncline uplifted at this time and undergoing erosion (Upper Carboniferous).

On the mid-geosynclinal ridge pelagic limestones, cherts and neritic sandstones were deposited (Lower Devonian to Lower Carboniferous).

It is the deposits of the third phase, the Schwellen limestones, deposited on submarine rises, the slope sediments and basinal shales which have been examined in this study. In terms of geosynclinal sequences these sediments belong to the euxinic phase of Pettijohn (1957), the leptogeosynclinal phase of Trümpy (1960), the bathyal lull of Goldring (1962) and the preflysch of Aubouin (1965). It is considered that the Rhenish geosyncline was intracratonic.

Section 2.4 Pelagic Sedimentation in the Rhenish geosyncline

German geologists have recognised two major facies in the Devonian, the Rhenish and Hercynian facies (Schmidt, 1926; Rabien, 1956; Erben, 1964.). The distinction is based effectively on depth of water and/or agitation of the environment. The Rhenish facies is characterised by 'impure' sediments, conglomerates, sandstones and silty shales with very little $CaCO_3$. Brachiopods are common. This facies is particularly well developed during the Lower Devonian when neritic sediments were deposited over wide areas. The Hercynian facies consists of 'pure' sediments, fine grained limestones and argillaceous shales with a dominantly pelagic fauna of cephalopods, thin-shelled bivalves and styliolinids. Reef carbonates are placed in the Hercynian facies. Schwellen and Becken sediments belong to the Hercynian facies. Erben (1964) distinguishes four intrafacies of the Hercynian magnafacies from the Devonian in Bohemia. Schwellen limestones belong to the Dvorce - Prokop intrafacies, and the Becken shales belong to the Badeholz intrafacies.

The Schwellen facies is developed in three main situations within the Rhenish geosyncline, above reefs or shallow water carbonates, on basement rises (geanticlines) and on volcanic ridges (Fig. 2.10, p. 41). Generally, the geanticlines accumulated pelagic limestones from the Lower or Middle Devonian, the volcanic ridges from the Givetian or Frasnian and the reefs from the middle Frasnian. In most cases, pelagic carbonate deposition continued until the end of the Famennian, locally it persisted into the Gattendorfia Stufe (Lowest Carboniferous). Some of the reefs were also developed above volcanic rises within the Sauerland basin during the Givetian and Frasnian, the hydrographical conditions determining whether reef or pelagic carbonates formed. Locally, during the upper Givetian and Frasnian cephalopod limestones were deposited on the flanks of reefs and then contain thin bands of displaced shallow water material.

The main Schwellen with their important localities are shown in Table 2.1 and the distribution of Schwellen limestones in the Rhenish geosyncline is shown in Fig. 2.4.

There are few Schwellen which did not receive carbonate sedimentation. Two subsidiary ridges of the mid-geosynclinal rise (the Hundsdorfer and Keller Schwellen, Fig. 2.7) supported neritic terrigenous sedimentation during the Upper Devonian, with the sediment probably derived from erosion of the ridge (Meischner, 1968). Rarely shales were deposited on rises (e.g. Hühnertalskopf, West Harz Schwelle, Fig. 2.6), although thin shale horizons are interbedded with the pelagic limestones.

In the following pages, the stratigraphical development of the Schwellen and Becken facies in the Rhenish geosyncline is described, with the situation and context of the pelagic limestones.

Reef Schwellen	Arnsberg Attendorn Brilon Langenaubach Lahn Syncline	Beul, Wocklum Obermarpe Kattensiepen Langenaubach Gaudernbach
Volcanic rises	Dill Syncline Haingrube Hauptgrünsteinzug Oberharzer Diabas-Zug	Eibach Haingrube Adorf am Martenberg Buntenbock, Hutthaler Widerwaage
Basement Rises	Westharz Schwelle Ense Schwelle Steinhorn Schwelle	Aeketal, Hühnertalskopf, Langestal, Riesenbachtal, Romkerhalle. Bicken, Blauer Bruch, Steinbruch Schmidt, Steinbruch Syring. Silberstollen
Reef Flank	Attendorn reef	Bonzel, Grevenbruch

Table 2.1. The main Schwellen in the Rhenish geosyncline with their important localities.

Fig. 2.4 Sketch map showing position of main Schwellen during the Devonian in the Rhenish geosyncline.

Schwellen of the mid-geosynclinal rise with stage from which they affected sedimentation are as follows:-

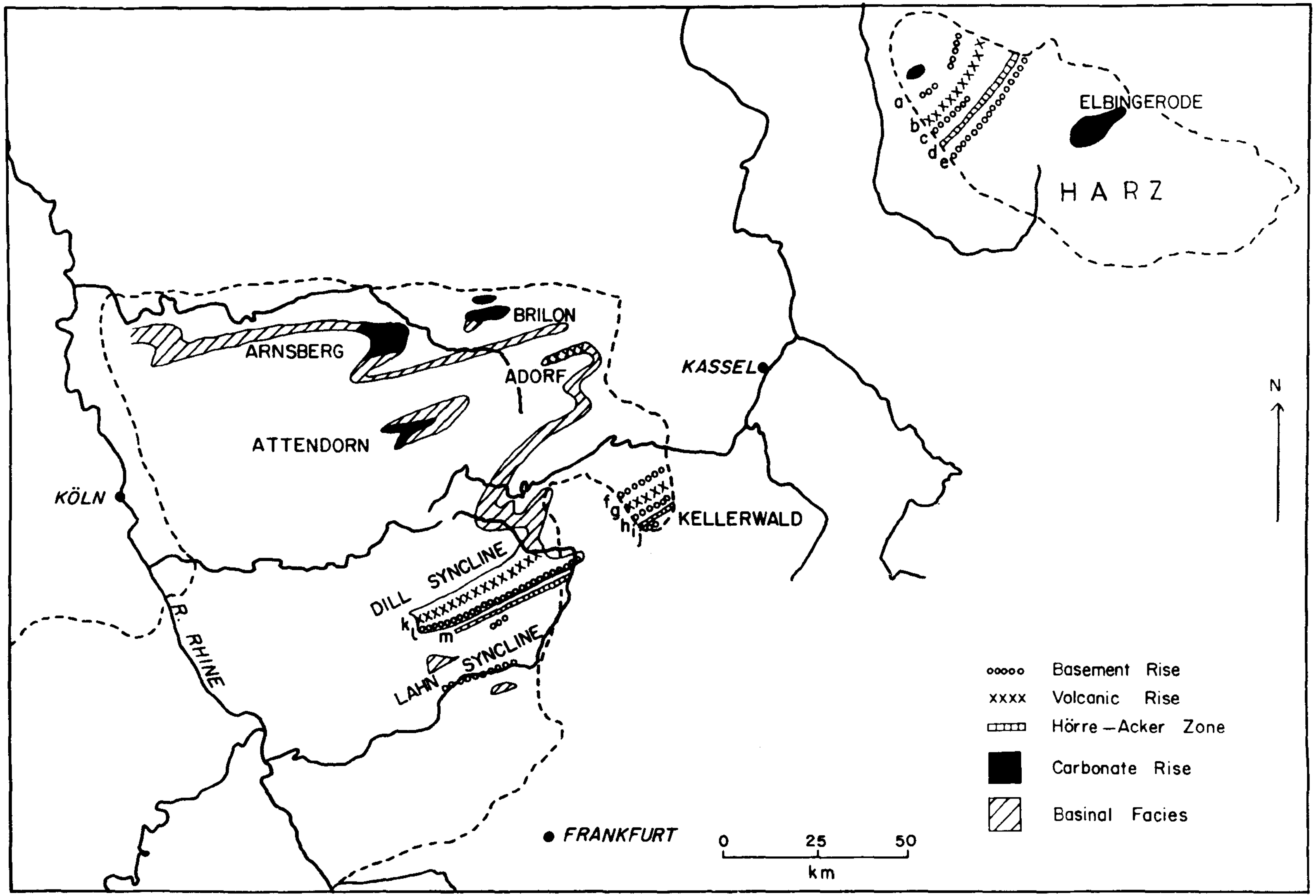
a) West Harz Schwelle (Eifelian)
b) Oberharzer Diabas-Zug (upper Frasnian)
c) Schwelle in the Söse Syncline (Frasnian)
d) Acker-Bruchberg Zone (Frasnian)
e) 'Sautal' Schwelle in the Sieber Syncline (Lower Devonian)
f) Hundsdorfer Schwelle (Frasnian)
g) Haingrube Schwelle (Frasnian)
h) Ense Schwelle (Eifelian)
i) Keller Schwelle (Frasnian)
j) Steinhorn Schwelle (Lower Devonian)
k) Volcanic Schwelle in Dill Syncline (Frasnian)
l) Ense or Bickener Schwelle (Eifelian)
m) Hörre Schwelle in Lahn Syncline (Lower Devonian)

Pelagic carbonates accumulated on Schwellen a,b,c,e,g,h,j,k and l. Neritic sandstones were deposited on d,f,i and m. Pelagic carbonate sedimentation ceased at the end of the Devonian but the mid-geosynclinal rise still affected sedimentation and reduced thicknesses occur in the Lower Carboniferous. The correlation of Schwellen in the Harz, Kellerwald and Rheinisches Schiefergebirge is as follows:-

a = f
b = g = k
c = h = l
d = i = m
e = j

Other Schwellen, Brilon, Arnsberg, Attendorn and Elbingerode developed above 'reef' limestones were sites of pelagic carbonate sedimentation from the middle Frasnian till the end of the Famennian. The Adorf Schwelle (Hauptgrünsteinzug) accumulated pelagic limestones from upper Givetian to upper Frasnian.

(After Schmidt, 1926; Mohr, 1968; Meischner, 1968; Meischner and Schneider, 1970).


ELBINGERODE
HARZ
BRILON
ARNSBERG
ADORF
KASSEL
ATTENDORN
KÖLN
KELLERWALD
DILL SYNCLINE
LAHN SYNCLINE
R. RHINE
FRANKFURT
a
b
c
d
e
f
g
h
i
j
k
l
m
N
0 25 50
km
Basement Rise
Volcanic Rise
Hörre – Acker Zone
Carbonate Rise
Basinal Facies

Pre-Devonian sediments in the Rhenish geosyncline. A strong unconformity exists in the Sauerland between Lower Devonian and Lower Palaeozoic sandstones and shales. In the Kellerwald the Devonian is conformable with the Silurian, which consists of calcareous and fine sandy sediments (Kockel, 1958).

Lower Devonian. Transgressive Gedinnian sediments occur over the whole region and with the exception of the mid-geosynclinal ridge area they are of fairly uniform thickness (The Ardennes and S.W. England also display a similar transgressive sequence at this time). In the Harz, sedimentation began a little later (Siegenian or Emsian). The Lower Devonian in the Rhenish geosyncline was mainly an episode of terrigenous sedimentation (Rhenish facies). Sediment was largely derived from the Old Red Sandstone Continent to the north. Deposition occurred within troughs orientated parallel to the geosynclinal axis and for most of the time sedimentation kept pace with subsidence. The neritic deposits pass laterally (south-east) into deeper water shales. The mid-geosynclinal rise was a positive area and received little sediment. During the Lower Devonian two subsidiary ridges on the mid-geosynclinal rise, the Ense and Steinhorn Schwellen supported pelagic carbonates while shales were deposited in the deeper inter-rise depressions (Meischner, 1968).

The palaeogeography for the end of the Lower Devonian is shown in Fig. 2.5.A.

Middle Devonian. Coarse terrigenous sedimentation decreased markedly during the Middle Devonian and shales (Wissenbacher Schiefer, Hercynian facies) became more widespread during the Eifelian. A distinct shelf area developed in the north of the Sauerland basin after a transgression onto

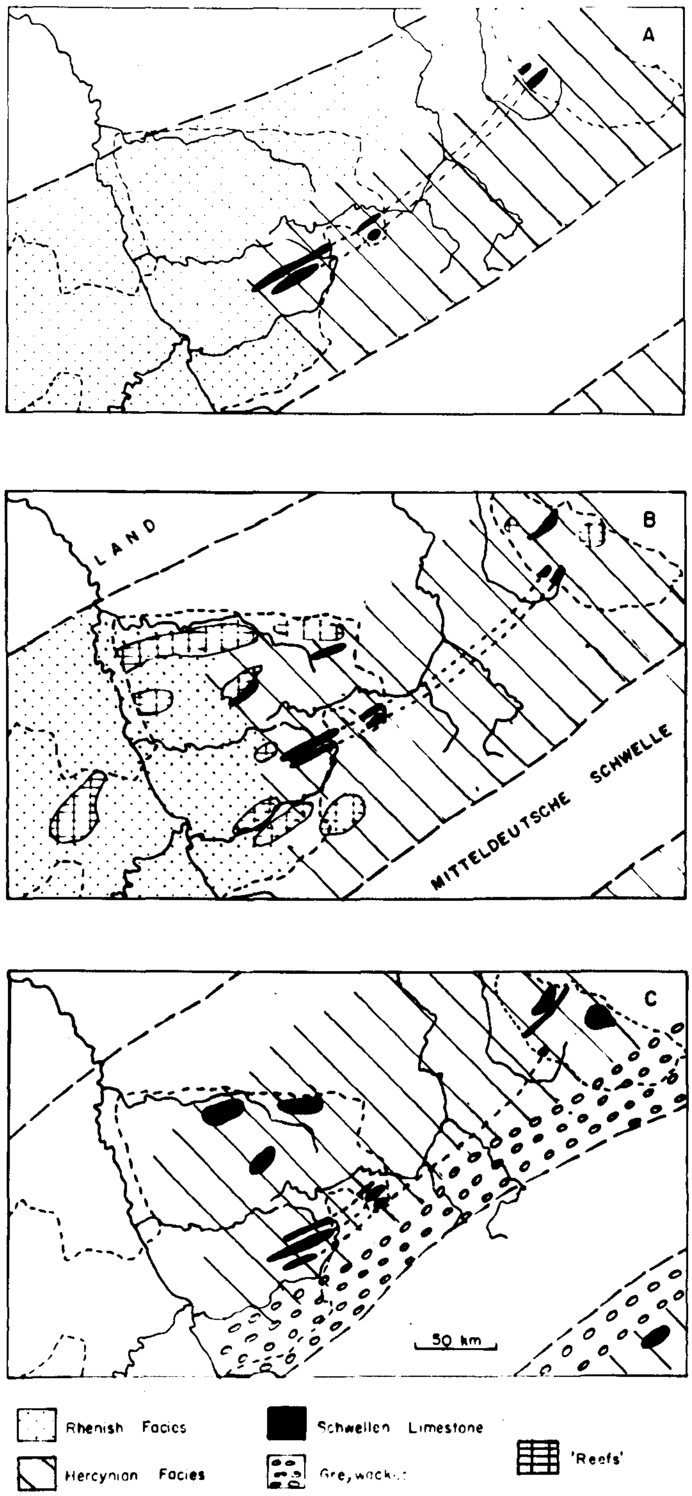
Fig. 2.5 Sketch maps of the generalised palaeogeography of the Rhenish geosyncline. A Lower/Middle Devonian, B Middle/Upper Devonian, C Upper Famennian.

The Rhenish facies representing neritic sedimentation is dominant in the Sauerland basin in the Lower Devonian.

The Hercynian facies (in the maps opposite, shales with a pelagic fauna locally with turbidites) is most widespread during the Upper Devonian (C).

The Schwellen facies and 'reefs' (Hercynian intra-facies, Erben, 1964) are shown separately.

(After Bogdanoff et al, 1962; Schmidt, 1962)


A
B
LAND
MITTELDEUTSCHE SCHWELLE
C
50 km
Rhenish Facies
Hercynian Facies
Schwellen Limestone
'Reefs'

the continent, and reworking of sandstones took place here.

In the northern part of the Harz, the depositional area was divided in two during the Middle Devonian as the West Harz Schwelle began to affect sedimentation. Where subsidence was most rapid 1000 m of Wissenbacher shales were deposited during the Eifelian, while on the Schwelle only 80 m of shales were developed (Mohr, 1968). In the Givetian, pelagic carbonates accumulated on the ridge (10 to 15 m of limestone) whilst in the basin 200 to 300 m of silty shales were deposited.

Volcanic activity towards the end of the Givetian produced large quantities of basic pillow lavas and tuffs and led to the formation of the Hauptgrünsteinzug (a volcanic ridge, the Adorf Schwelle, in the Sauerland basin, Fig. 2.4). Similar ridges were developed during the Frasnian in the Harz (Oberharzer Diabas-Zug) and in the Kellerwald (Haingrube Schwelle) on the mid-geosynclinal ridge (Fig. 2.4). On these volcanic ridges condensed pelagic limestones accumulated, in some cases till the end of the Devonian. The Hauptgrünsteinzug extends for 50 km along the strike. At Adorf am Martenberg 2 m of crinoid-rich limestone occur above the pillow lavas of the Hauptgrünsteinzug, and above this appears the pelagic carbonate facies (5 m representing most of the Frasnian) which itself passes upwards into nodular limestones and ostracod shales (Lower Famennian). Commonly there is an horizon of hematite enrichment (Roteisenstein) immediately over the pillow lavas. Hydrothermal solutions passing through the volcanic material are generally considered to be the origin of this iron ore (Bottke, 1965).

The Givetian is marked by a great development of reefs and other shallow water carbonate sediments. These are particularly well developed along the shelf margin to the north of the Sauerland basin (Krebs, 1968b). (Extensive reef growth occurred in the Ardennes at this time as well). Other reefs developed within the basins where thick piles of tuff and lava had raised the sea floor (e.g. in the Attendorn region, Dill Syncline (Langenaubach, Krebs, 1966) and the Lahn Syncline). The reefs provided detritus which was deposited in the deeper water areas around the reefs as turbidites (allodapic limestones, Meischner, 1968). In the Harz, 'reef' limestones were deposited in two areas, Elbingerode and Iberg.

Cephalopod limestones accumulated on the flanks of some reefs and detrital bands of shallow water material occur within the pelagic facies. The latter passes through shales with nodules into dark grey shales which were deposited in the inter-reef areas.

On the mid-geosynclinal rise, pelagic carbonates continued to accumulate on the Ense and Steinhorn Schwellen.

Uplift and erosion of the Mitteldeutsche Schwelle began in the Middle Devonian and turbidity currents brought greywackes into the Taunus basin. The palaeogeographic situation at the end of the Givetian is shown in Fig. 2.5.B..

Upper Devonian. Volcanic activity continued into the Frasnian in some areas and two volcanic ridges, the Haingrube Schwelle in the Kellerwald (Fig. 2.7) and the Oberharzer Diabas-Zug, were formed at this time. Both ridges (in contrast to the Hauptgrünsteinzug) remained positive areas, maintaining their topographic elevation for the whole of the Upper Devonian. The Haingrube Schwelle supported pelagic carbonates but stromatoporoids, corals and calcareous algae developed at the same time a few kilometres along the ridge (Schneider, 1969). At Hutthaler Widerwaage (on the Oberharzer Diabas-Zug) condensation is extreme. The uppermost Frasnian and Famennian is represented by just 98 cm of pelagic limestone. All conodont zones in this sequence are present though two disconformity surfaces occur (Meischner and Schneider, 1970). The relief between different parts of the volcanic ridges may have been substantial. In depressions (termed Spezial Becken) condensed deposits may form which are transitional between rise and basin sediments (cf. Tucker and Straaten, 1970). The sediments here are usually dark grey shales, locally siliceous and flaser limestones.

Reef growth in the Rheinisches Schiefergebirge, Harz, Ardennes, S.W. England and other areas was terminated during the Frasnian (commonly in the middle Frasnian) by rapid subsidence (Meischner, 1964). Cephalopod limestones accumulated upon these former reef areas but for some, more pronounced subsidence allowed 'basinal' shales to be deposited over the reefs. This occurred in the north-western part of the Sauerland basin and in the central part of the Taunus basin (compare figs.2.5.B and C).

The widespread change from carbonate deposition to deeper water sedimentation is not everywhere synchronous and obviates the suggestion of a catastrophic event. (McLaren, 1970, considered Frasnian reef growth to be terminated by the effect of a large meteorite hitting the earth). Within reef limestones evidence of subsidence before the major facies change occurs. At Chudleigh (S.W. England) Scrutton (1969) interpreted early Frasnian limestones on the coral fauna as showing a gradual deepening environment, which was succeeded by cephalopod limestones (the Dunscombe Farm Goniatite Bed). In the Rhenish geosyncline, reefs gradually decrease in their areal extent with time suggesting reef growth could not keep pace with subsidence (Meischner, 1964). Subsidence may have continued after the cessation of reef growth and the overlying pelagic limestones thus pass upwards into 'basinal' sediments (shales with nodules). This probably occurred in the Chudleigh area where the sediment gradually becomes more argillaceous upwards (Tucker and Straaten, 1970).

In most cases cephalopod limestones accumulated over submerged reefs until the end of the Devonian. Large neptunian dykes penetrate some 'reef' rock and are filled by pelagic carbonates (e.g. Langenaubach, Krebs, 1966).

The West Harz Schwelle was a prominent feature in the Aeketal area (Fig. 2.6) where the Upper Devonian sediments are condensed limestones less than 20 m thick, compared with 400 m of shales in the North-West Harz Basin. In the eastern part of the rise, the Clymenia and Wocklumeria Stufen are absent, and at Langestal (Fig. 2.6) the youngest Upper Devonian sediments are lower Famennian (Mohr, 1962). The Lower Carboniferous is paraconformable over the whole area. At Aeketal (10 km from Langestal) a continuous sequence is developed (Fuhrmann, 1954). Stoppel (1968) invoked subaerial erosion to account for the stratigraphical break. Only one neptunian dyke is known from this region (where styliolinid limestone fills a small pipe 30 cm deep in the Frasnian limestone below, Mohr, 1962) yet Stoppel has suggested karst weathering during Famennian emergence. Upon present day rises, Tertiary sediments are commonly found exposed at the surface (e.g. Milliman, 1966; Cifelli et al, 1966; Ewing et al, 1966). Current activity is considered to sweep seamounts

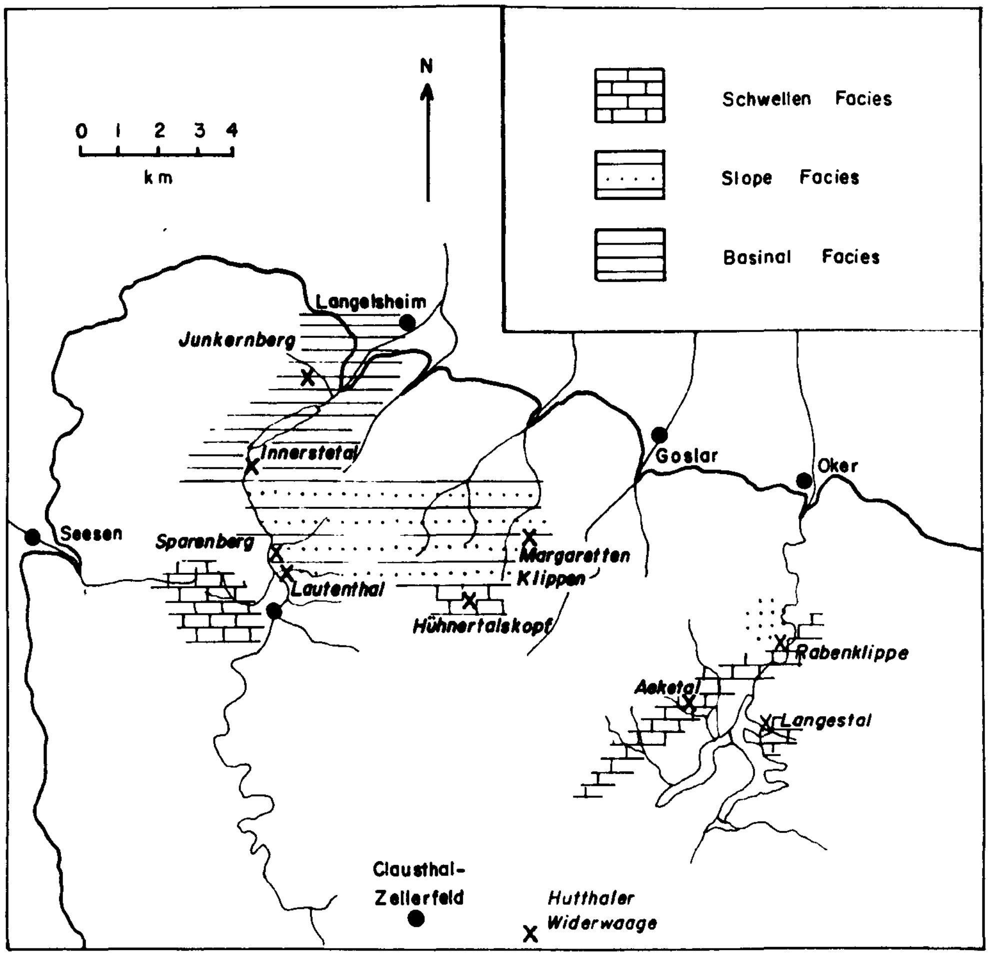


Fig. 2.6 Sketch map showing the distribution of Schwellen, slope and basinal sediments in the N.W. Harz Mountains during the Frasnian. Important localities mentioned in the text are marked. (After Müller-Steffen, 1965).

and ridges and cause erosion. It is therefore unnecessary to invoke emergence for the stratigraphic break in the eastern part of the West Harz Schwelle. There is no indication of the structures characteristic of partial emergence (dessication cracks, birds eyes, stromatolites) or of more extreme emergence (relict soils, karst surfaces, caves etc.). Stratigraphic breaks also occur above reef limestones in the Sauerland basin and although now attributed to submarine erosion, they were thought to indicate tectonic movements between the Devonian and Carboniferous (Krebs, 1968a).

A much condensed Upper Devonian sequence occurs at Hühnertalskopf (Fig. 2.6), which is possibly part of the West Harz Schwelle, though it is separated from the main exposures further east by 5 km of Lower Devonian sandstone outcrop. The Upper Devonian is only 12 m thick, and nearby the Famennian is again absent (Stoppel, 1968). The sediments on this particular rise are unusual since the succession is mostly shale. (Flaser limestones are present in the Frasnian and some nodule bands are developed). Müller-Steffen (1965) and Stoppel (1968) have termed this rise a 'Spezial Schwelle'.

On the mid-geosynclinal rise (Fig. 2.7) pelagic carbonate sedimentation continued on the Ense, Steinhorn and Haingrube Schwellen. Another subsidiary rise became effective during the Upper Devonian (the Hundsdorfer Schwelle) and neritic sandstones and arkoses containing brachiopods were deposited here. Transportation along the ridge is suggested from current directions (Meischner, 1968). These sediments were interpreted by Meischner as shallow water deposits derived by erosion of parts of the ridge, possibly of emergent parts.

The Hörre-Acker-Bruchberg zone, a belt of atypical Upper Devonian sediments which occurs on the mid-geosynclinal rise (Fig. 2.4) is interpreted by Meischner (1968) as a subsidiary ridge (the Keller Schwelle Fig. 2.7). However, Bender and Brinckmann (1969) working in the Lahn Syncline consider this zone to be a deeper water area.

Depressions between subsidiary ridges of the mid-geosynclinal rise received little sediment and siliceous shales and cherts were deposited (Fig. 2.7). Meischner suggested that in these depressions water depths were less than 100 m.

During the Upper Devonian, nodular limestones and shales with nodules were deposited on the flanks of the former reef areas, the mid-geosynclinal

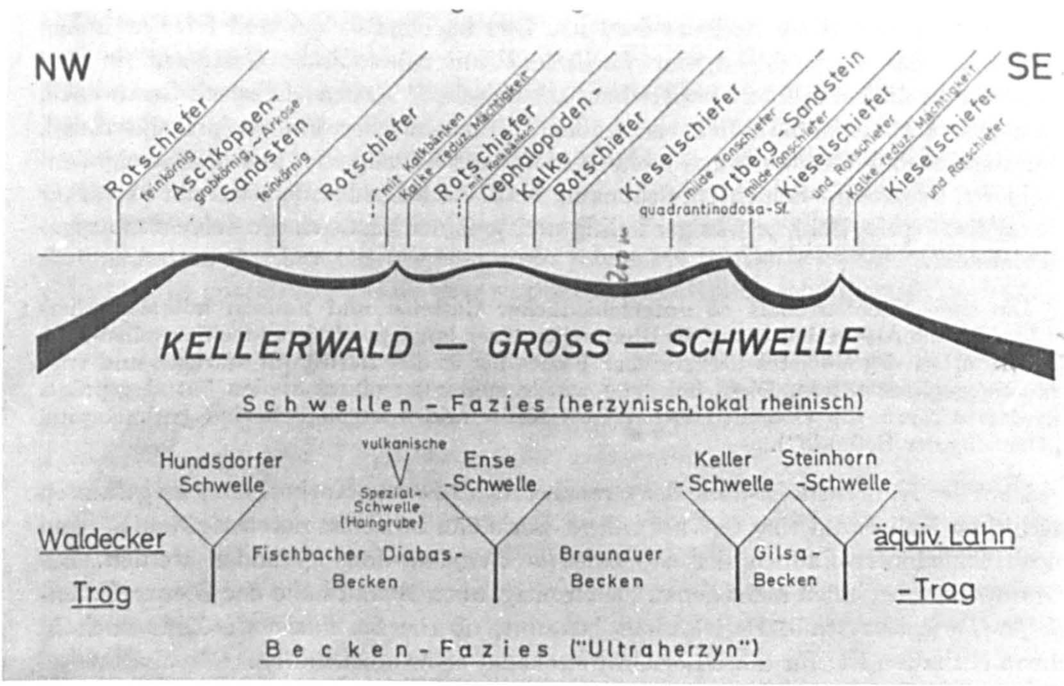


Fig. 2.7 Section across the mid-geosynclinal rise in the Kellerwald during the Upper Devonian showing the subsidiary ridges. Approximate thickness of sediment for the Palmatolepis quadrantinodosa conodont zone (Cheiloceras/Platyclymenia Stufen) indicated by the width of the black band. The lithology at that time is marked above (From Meischner, 1968).

rise and the West Harz Schwelle. Slumping and reworking of sediment commonly took place in this slope environment. In the basins, silty shales rarely with calcareous nodules and locally rich in ostracods were deposited.

Turbidites occur in the basinal shales at some horizons in the Rheinisches Schiefergebirge. During the Frasnian, allodapic limestones were derived from the reef areas. In the Famennian, sandstones deposited by turbidity currents (which flowed around former reef areas) were derived from the shelf area to the north (Plessmann, 1962; Einsele, 1963).

Continued uplift and erosion of the Mitteldeutsche Schwelle, with the formation of greywackes had nearly filled the Taunus basin by the end of the Devonian. (Palaeogeography of the Upper Devonian, Fig. 2.5 C and Fig. 2.8).

Lower Carboniferous. Pelagic sedimentation in the Sauerland basin of Schwellen limestones and light grey ostracod shales persisted locally until the *Gattendorfia* Stufe. Black shales (Liegend Alaunschiefer) and cherts (Kulm Kieselschiefer) yielding only conodonts, goniatites and posidonids were deposited over wide areas with fairly uniform thickness during the *Gattendorfia* and *Pericyclus* Stufen. The mid-geosynclinal rise only caused a small reduction in thickness.

Shelf carbonates were extensively developed in the Ardennes during the Lower Carboniferous and extended across the shelf area to the north of the Sauerland basin. Turbidity currents brought carbonate detritus into the basin forming allodapic limestones (Kulm Plattenkalk) (Fig. 2.8).

The main period of flysch deposition occurred during the Lower Carboniferous and a vast amount of detritus was brought from the south by turbidity currents and slump flows (Kuenen and Sanders, 1956; Plessmann, 1962). They reached the mid-geosynclinal ridge during the *Pericyclus* Stufe, and some of this sediment was reworked on the mid-geosynclinal rise to form the Keller Quartzite (Fig. 2.8). Greywackes filled the Sauerland basin during the *Goniatites* Stufe, and detritus for this probably came from the uplifted Taunus basin to the south Fig. 2.8. The turbidites are interbedded with dark grey shales (Kulmtonschiefer) containing a pelagic fauna of posidonids and goniatites.

Upper Carboniferous. The Taunus and Sauerland basins were both uplifted above sea level by the end of the Lower Carboniferous, and pelagic sedimentation ceased. Deposition only persisted in the most northern

Fig. 2.8 Sections across the Sauerland basin from Middle Devonian to Upper Carboniferous (From Wunderlich, 1965). The south-eastern Schwelle (labelled KGS) is the mid-geosynclinal rise of Meischner (1968).

Section 1. Upper Givetian, with reef developed along the margin of the shelf area, and pelagic limestone deposited on the mid-geosynclinal rise.

Section 2. Upper Frasnian. Marine transgression on to continental area with spread of Hercynian facies. Pelagic limestones deposited above reefs.

Section 3. Upper Famennian.

Section 4. Lower Carboniferous (Gattendorfia/Pericyclus Stufen). Taunus basin filled by flysch, which re-worked on mid-geosynclinal rise into the Kellerwald Quartzite. Extrusions of basalt (Diabas) along flanks of rise. Pelagic carbonate sedimentation absent and black shales and cherts accumulated in the Sauerland basin. Development of shallow water carbonates (Kohlenkalk) on northern shelf.

Sections 5, 6 and 7. Lower Carboniferous (Goniatites Stufe α, β and γ). Sauerland basin filled by greywackes derived from uplift and erosion of the Taunus basin. Allodapic turbidites in northern part of Sauerland basin derived from shelf.

Sections 8,9 and 10. Upper Carboniferous (Namurian, Westphalian and Stefanian). Uplift and erosion of Sauerland basin with formation of molasse.

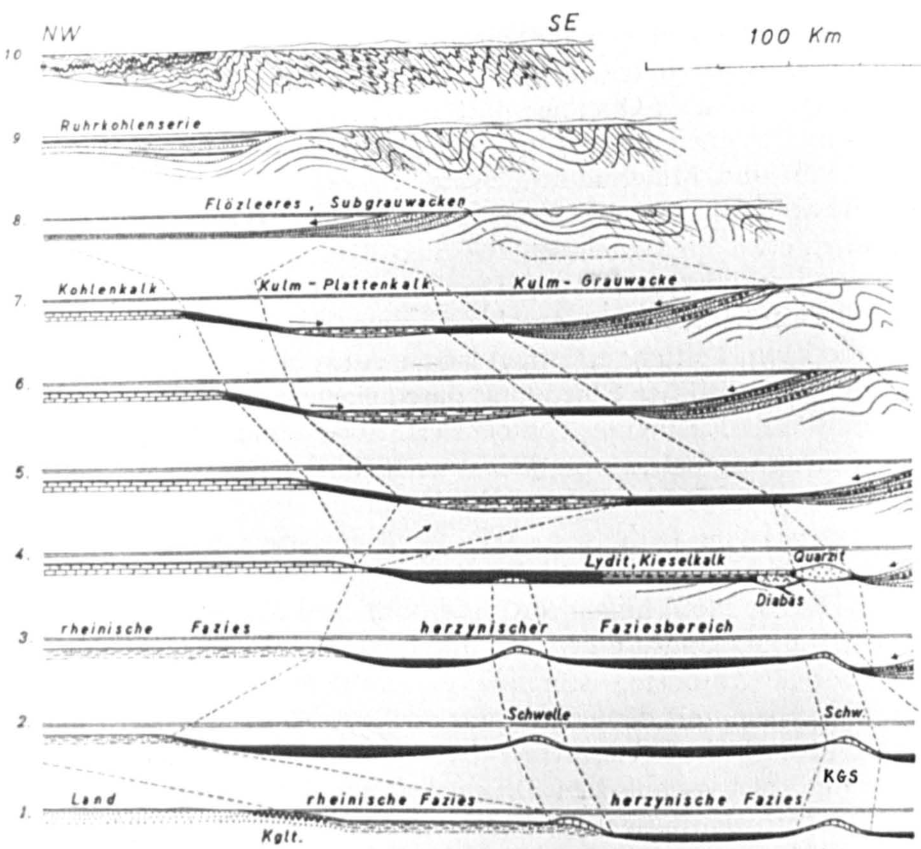


Abb. 5. Schematische Serienprofile der faziellen und tektonischen Entwicklung des Ostrheinischen Schiefergebirges vom Mittel-Givet bis Stefan.

part where subgreywackes, sandstones (Flözleeres) and paralic coal measures were deposited. (Fig. 2.8).

Section 2.5 Oceanic crust in the Rhenish geosyncline?

The association of basaltic spilites, serpentinites, picrites and pelagic sediments (which characterize the second phase of volcanicity) is an ophiolite suite (Aubouin, 1965; Krebs, 1968a; Dewey and Bird, 1970). This might suggest the geosyncline was underlain by oceanic crust and Mitchell and Reading (1969) suggested that "Schwelle subfacies (of preflysch) will occur on rises and Becken subfacies in small basins on the mid-ocean ridge". However, the pillow lavas have a widespread distribution (Fig. 2.3) and are not concentrated along one particular zone, as might be expected if a suture zone existed in the Rheinisches Schiefergebirge.

Hermann and Wedepohl (1970) analysed spilitic lavas and intrusives from the Hauptgrünsteinzug (Sauerland basin), Kellerwald, Lahn-Dill region and the Harz. Their analyses show that the basaltic spilites, andesites and picribasalts have major element and mineral compositions comparable to Recent spilites from oceanic ridges, but that the spilites have rare earth abundances typical of continental tholeiites and oceanic islands.

Tectonism and metamorphism in the Rheinisches Schiefergebirge are less intense and quite different from the Alps, west coast of North America and island arc situations. The Rhenish geosyncline was unlike many present-day areas which are considered to represent geosynclines. These are marginal to continental areas (e.g. Dewey and Bird, 1970; Mitchell and Reading, 1969). Crustal tension existed during the early part of the Rhenish geosyncline (Krebs, 1968a; MacGillavry, 1970) but the work of Hermann and Wedepohl (1970) suggests that the geosyncline did not open sufficiently for oceanic ridges and sea floor spreading to develop.

Section 2.6 Other Occurrences of Devonian Pelagic Sediments

Erben (1964) summarized the development of the Hercynian and Rhenish facies in Germany and noted the occurrence of pelagic limestones (Hercynian facies) elsewhere in Europe. In central Europe, cephalopod limestones are found in Thuringia, Bohemia and Austria (Carnic Alps) developed on geanticlines and above reef limestones. They also occur in the Urals geosyncline and in North Africa.

In southern Europe, pelagic limestones occur in the Montagne Noire (S. France) (Fig. 2.18, p. 13) and the Pyrenees, where they are known as griotte. Locally the Devonian is conformable on the Silurian in the Montagne Noire and the Lower Devonian consists of sandstones and conglomerates. During the Middle Devonian, shallow water carbonates and in some areas reefs developed until the end of the Givetian or the early Frasnian, when pelagic carbonate sedimentation spread over much of the Montage Noire (Bayer et al, 1968). The Upper Devonian griotte is exactly the same facies as the Schwellen limestone, although the griotte is generally red through the presence of hematite. In the Pyrenees an identical facies developed on geanticlines during the Middle and Upper Devonian. Griotte occurs in the Cantabrians in the Upper Devonian, but is very widespread during the Lower Carboniferous.

In S.W. England, pelagic limestones occur at Chudleigh near Newton Abbot and are described in Tucker and Straaten (1970) which is appended. The pelagic limestones occur above shallow water carbonates during the Frasnian in a situation similar to that in Germany. Deeper water shales are developed extensively in S. Devon and N. Cornwall during the Upper Devonian and contain turbidites at some horizons.

The Hercynian facies of condensed pelagic limestones and ostracod shales is only well developed in the Variscan geosyncline. In other areas (North and South America and Australia) these sediments do not occur (Oswald, 1968) and neritic sediments of the Rhenish facies are developed.

Section 2.7 Comparison with Recent and Ancient Geological Situations

Environments can be found today which are comparable with the conditions of deposition envisaged for the Schwellen limestones but exact analogies including the tectonic situation are difficult to find. Many Recent examples can be cited of pelagic sediments developing on submerged 'reefs' or volcanic rises in an oceanic environment (e.g. Hamilton, 1956). The Blake Plateau, off the east coast of North America is a continental area which has undergone rapid subsidence since the Cretaceous (Sheridan et al, 1969). Lower Cretaceous algal limestones have been recovered from the plateau indicating that it was once at sea level. Rapid subsidence prevented the continued formation of shallow water carbonates, and pelagic foraminiferal oozes have been accumulating since the Tertiary. A similar situation exists off the Iberian coast where Black et al (1964) described non-magnetic seamounts which are part of the continental basement. Recent pelagic sediments are accumulating over some shallow shelf areas where there is little terrigenous influence (e.g. the Yucatan shelf in the Gulf of Mexico, Logan, 1969). This type of situation must have existed during the Upper Devonian in the Rhenish geosyncline when pelagic shales and limestones were deposited over wide areas. Submarine rises of continental crust are present off the N.W. coast of New Zealand (Brodie, 1964) and are receiving pelagic carbonate sedimentation. These may be similar to the geanticlines or basement rises occurring in the Rhenish geosyncline (West Harz Schwelle, mid-geosynclinal ridge).

The pelagic limestones of the Tethyan Jurassic are comparable to the Devonian cephalopod limestones. In the Alps and Sicily, pelagic carbonate was deposited over Triassic 'reef' limestones when the rate of subsidence became too great for continued 'reef' growth. A break-up of the carbonate platform into seamounts is suggested by Jenkyns and Torrens (1971). The Briançonnais zone in the Western Alps had a similar effect on sedimentation as the mid-geosynclinal ridge of the Rhenish geosyncline. It was a stable ridge with subsidiary ridges where calcareous sediments considerably reduced in thickness were deposited during the Mesozoic.

In the Lower Palaeozoic, sediments equivalent to the Schwellen limestones are condensed graptolitic shales, which are considered to have been deposited on submarine rises (Williams, 1962). The basinal shales (Becken facies) of the Rhenish geosyncline are similar to the Schistes lustrés of the Alps and to mudstones and siltstones occurring in Lower Palaeozoic basins before the development of flysch.

Section 2.8 Summary

In the Rhenish geosyncline condensed pelagic carbonates occur on three types of topographic high (Fig. 2.10) (Schwellen) (i), a submerged reef or shallow water carbonate area, (ii), a volcanic ridge, (iii), a basement rise (geanticline). Outside the Rhenish geosyncline, Devonian pelagic carbonates were developed mainly above reef limestones. The relationship between reef, basement rise and basin (after Krebs, 1968b) is shown in Fig. 2.9.

The carbonate sediment is similar on the different types of rise. However, basalt fragments and tuff may occur within pelagic sediments on volcanic rises, and neptunian dykes containing pelagic carbonate may penetrate reefs.

Pelagic sediments occurring above reef limestones are normally middle Frasnian to Upper Famennian in age, those on volcanic ridges are upper Givetian or Frasnian to the end of the Devonian. The mid-geosynclinal rise was a persistent feature of the Rhenish geosyncline and supported pelagic carbonates on two subsidiary ridges (Ense and Steinhorn Schwellen) from the Lower Devonian till the upper Famennian. The West Harz Schwelle affected sedimentation from the Eifelian and pelagic limestones accumulated from the Givetian till the end of the Devonian.

The slope facies, (Fig. 2.11) transitional sediments between the Schwellen and Becken facies, is characterised by the presence of slumping (possibly due to faulting along the flanks of the Schwellen) and normally consists of shales with nodules and nodular limestones. It is difficult to ascertain the width of the slope facies because of folding and non-exposure but in the Harz, the distance today from the West Harz Schwelle (Aeketal region) to the basinal sediments (Innerstetal) is about 12 km. (Fig. 2.6).

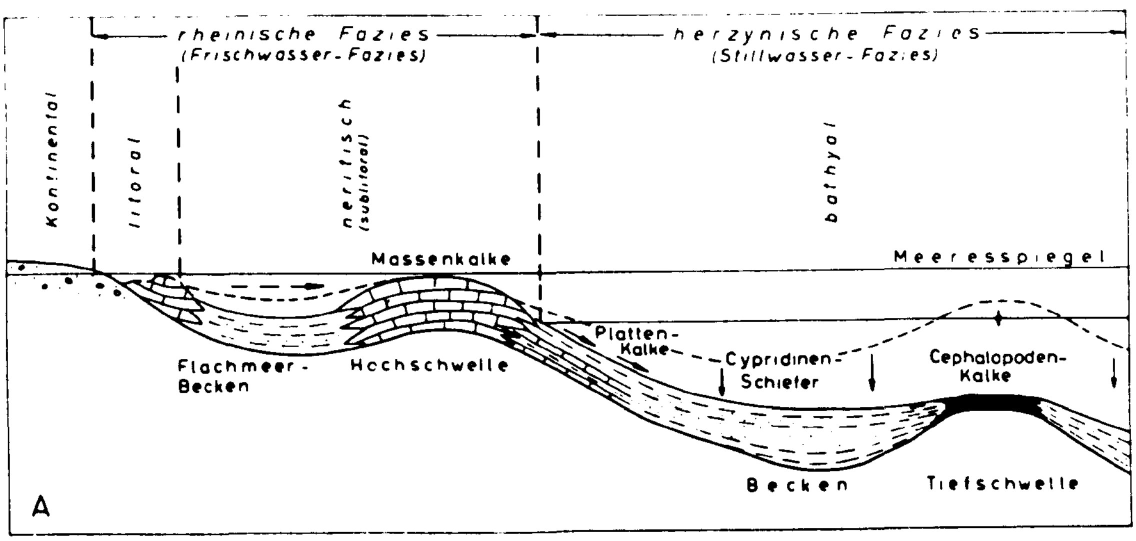


Fig. 2.9 Diagrammatic section showing relationship between reefs, basins and Schwellen. (From Krebs, 1968b, based on Rabien, 1956). Rabien considered reefs as forming on Hochschwelle (high rise) and cephalopod limestone as forming on Tiefschwelle (deep rise). Reefs occur along the shelf margin or on volcanic rises in the basins. Cephalopod limestones form above basement rises (geanticlines) (e.g. mid-geosynclinal rise) volcanic rises, submerged reefs and on the flanks of reefs.

Fig. 2.10 Diagrams showing the situations of the Schwellen limestones (shaded black).

A. Pelagic limestones occurring above a submerged reef and within neptunian dykes penetrating the reef. The pelagic carbonate facies is also shown on the flanks of the reef where it contains horizons of displaced shallow water material. The reef itself is shown occurring above volcanics (e.g. Attendorn reef) but reefs also developed along the shelf margin of the continent to the north.

B. Pelagic limestones occurring on a volcanic ridge. Depressions occur on the volcanic ridges (also on submerged reefs) where Flaser limestones and siliceous shales commonly formed.

C. Schwellen sediments occurring on a basement rise. Pelagic limestones were deposited on some subsidiary ridges and neritic sandstones (stippling) on others. Siliceous shales accumulated in the inter-rise depressions.

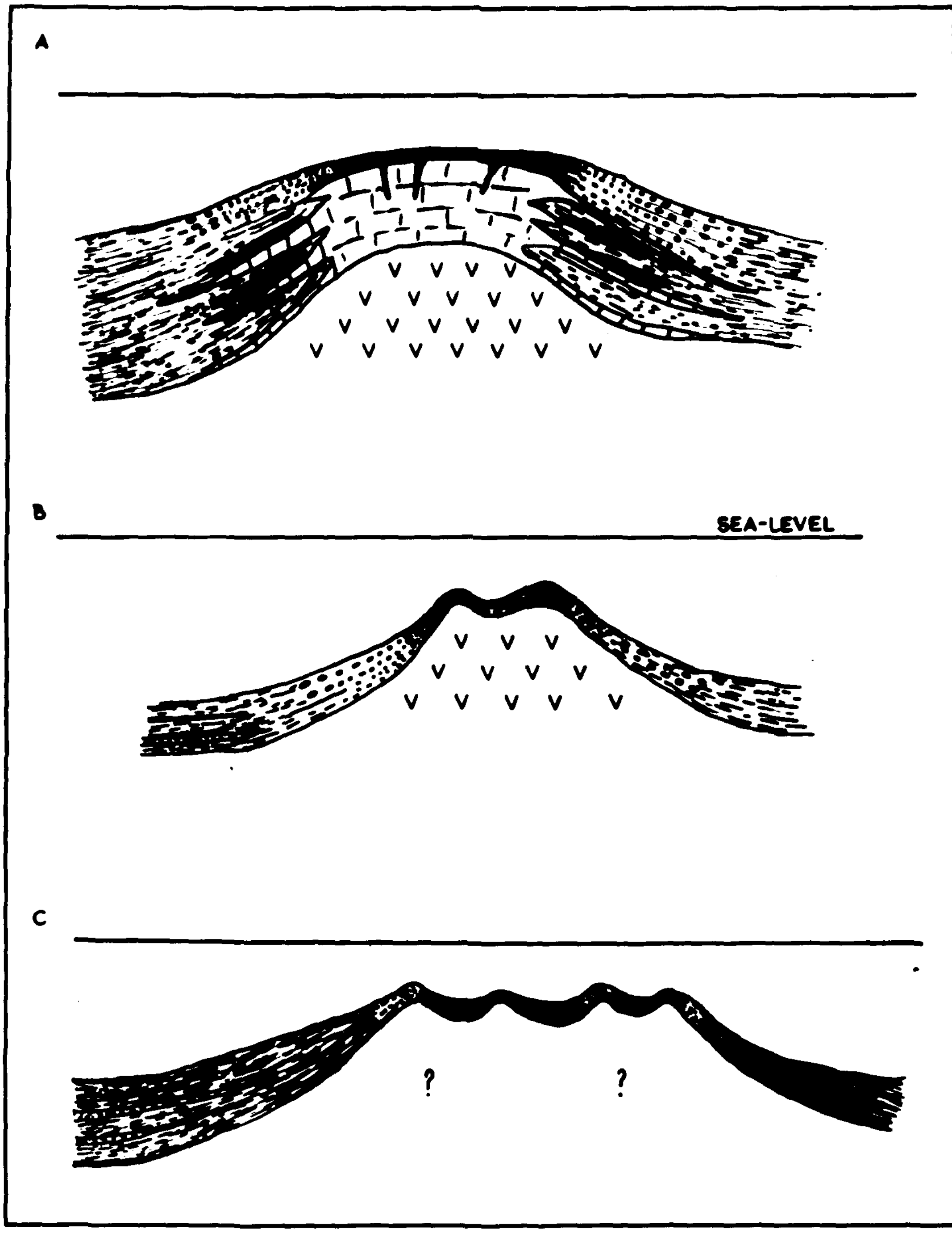
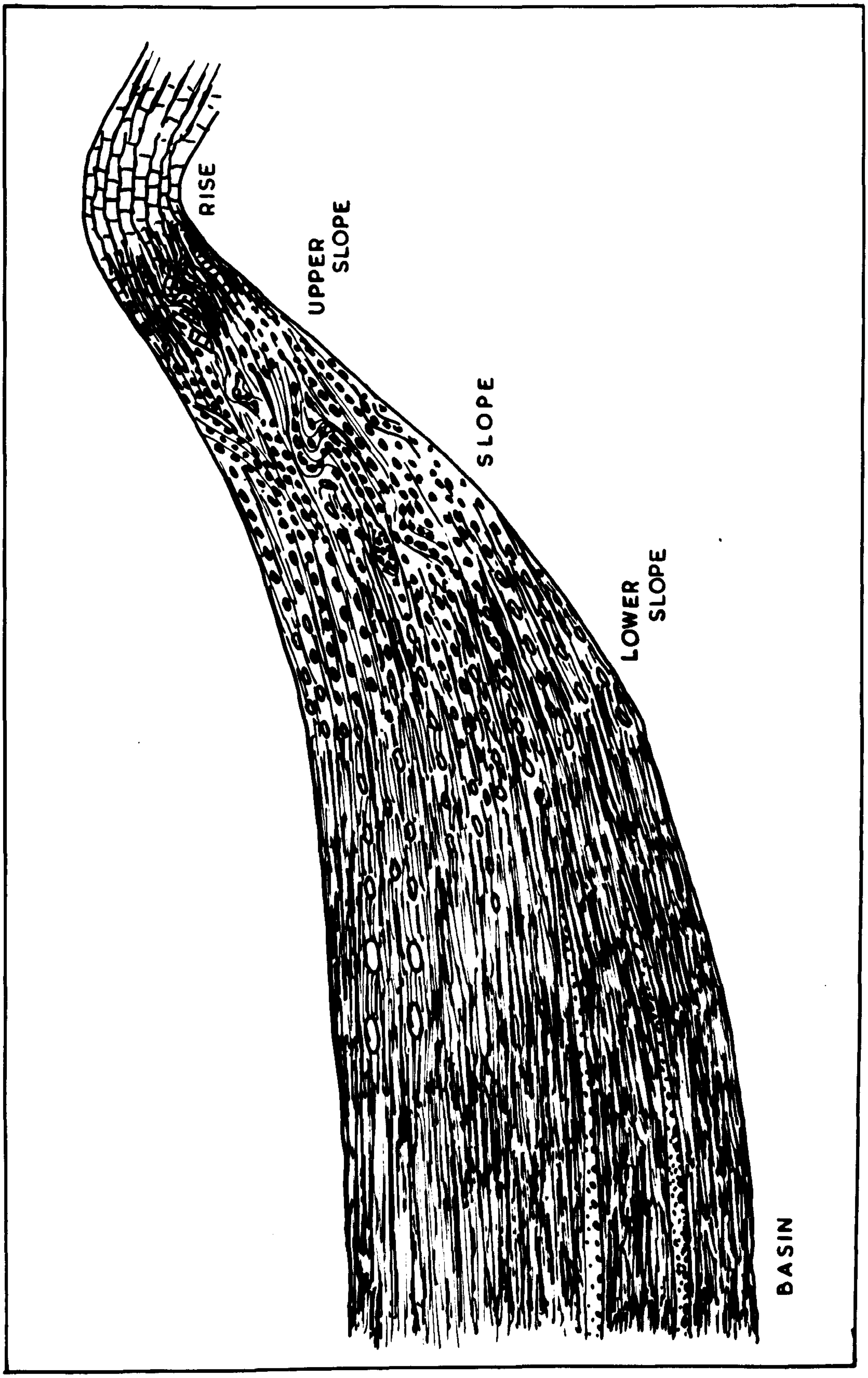

A
B
SEA-LEVEL
C
?
?

Fig. 2.11 Diagrammatic Schwellen to Becken section showing main features of the sediments. Limestones, formed on top of the rise and in the upper slope region may occur as slumped blocks lower down the slope. The slope sediments, shales with nodules, nodular limestones and shales, were commonly involved in slumping or reworking. In the lower part of the slope, shales with nodules are unaffected by slumping and pass laterally into silty shales of the basin, locally with clastic or carbonate turbidites.


RISE
UPPER SLOPE
SLOPE
LOWER SLOPE
BASIN

Deep water shales with a pelagic fauna (Hercynian facies) are developed from the Lower Devonian onwards and their distribution, then limited to the south-eastern part of the geosyncline, became more widespread in the Middle Devonian and in the Famennian covered most of the depositional area, apart from the topographic highs.

In the next three chapters, the sedimentology of the Schwellen, basin and slope sediments is described with their lithological features, sedimentary structures, fauna and diagenesis. The limestones deposited on the Schwellen are described in detail in chapter 3, but the neritic sandstones which occur on subsidiary ridges of the mid-geosynclinal rise have not been examined. The sandstones have been studied by geologists from Göttingen (Meischner, 1968 and Habilitation (in press) and students unpublished dissertation). Similarly the Kellwasser limestone, a feature of Frasnian Schwellen sediments has not been examined (Professor Krebs). The division of the Upper Devonian pelagic sediments into Schwellen, slope and basin is based primarily on lithology, although there were times when limestones were developed over a slightly larger area. For example, during the Frasnian and Platyclymenia Stufe carbonates were generally more widespread and Flaser limestones occurred lower down the slope, whereas in the Cheiloceras Stufe shales were deposited over larger areas and commonly limestones only developed on the top of the rise (Müller-Steffen, 1965). The relationship between the three facies and the lithologies is shown in Fig. 2.11, p.43. The Schwellen sediments described are Flaser limestones which would also have been deposited in the upper slope region and occur in small depressions on the rises (e.g. Adorf am Martenberg, Fig. 3.37, p.95). The slope facies (chapter 5) is distinguished on the presence of slumped and reworked sediments, probably resulting from deposition on the steeper parts of the slope. The basinal sediments (mostly silty shales, locally with turbidites) pass laterally into shales with calcareous nodules showing no signs of sediment movement, which accumulated in the lower slope region (hence chapter 4 deals with 'basin and lower slope facies', deeper water sediments with no indication of slumping).

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CHAPTER 3

Sedimentology of the Schwellen Facies

Schwellen limestones, representing sedimentation on submarine rises in the Variscan geosyncline, are fine grained limestones, typically with many shale streaks or Flasers (hence term Flaser limestone) formed through diagenesis and tectonics. They are reduced in thickness compared with the sediments deposited on the slope and in the basins. The limestones are typically homogeneous microsparites or biomicrosparites, rich in pelagic microfossils, and locally thin-shelled bivalves and ammonoids. The presence of ammonoids has led to the German stratigraphical term for this limestone, Cephalopodenkalk.

The main factors controlling deposition are a) rate of sedimentation, determined by carbonate production and terrigeneous influence, b) depth, in effect, whether the rise is near or within the photic zone (<150 m), c) current activity and d) amount of organic matter (a factor important in diagenesis). Other factors affect the facies locally.

Evidence for deposition on submarine topographic highs is provided by the presence of sedimentary slumps and reworked sediments in the slope facies (p. 235), the greater thickness of basinal sediments compared with Schwellen limestones and the different faunas of the rise and basin environments. The Schwellen sediments pass downslope into nodular limestones, shales with nodules and shales, locally rich in ostracods. The depth of deposition of this pelagic carbonate facies probably did not exceed a few hundred metres and in many cases was 50 m or less (p.196). No evidence for emergence has been found in the Schwellen limestones. The approximate rates of sedimentation (Fig. 3.1) are comparable with modern pelagic sediments (normally less than 2 mm per 1000 years). In the basins the rate was much higher (20 to 30 mm/1000 yrs, not allowing for compaction). Typical successions in Schwellen facies are shown in Figs. 3.2 and 3.3. Localities referred to in this chapter are shown in Fig. 3.4.

Section 3.1 Schwellen sediments

The Schwellen sediments are typically well-bedded Flaser limestones (Figs. 3.5,3.6 and 3.7) varying in thickness from a few centimetres to about a metre. More nodular horizons may be present and these are particularly well developed in the Montagne Noire and Pyrenees, where these fine grained limestones are termed griotte (Figs. 3.8, 3.9 and 3.10). Irregular bedding surfaces are commonly developed through pressure solution (Fig. 3.9). Thin shale horizons a few millimetres thick occur between limestones and may also be the product of pressure solution. Thicker shale beds (up to 5 cm) rarely occur. They are usually dark grey in colour and contain little carbonate.

The limestones are composed of 80 to 95% $CaCO_3$, The matrix is mostly micro-sparite (5 to 20μ; Folk, 1959), but patches of coarser calcite (pseudosparite) are commonly present. Micrite (<4μ) is not very common and occurs as vague clots resembling structure grumeleuse (Cayeux, 1935). Pelletoids up to 200μ in diameter and consisting of micrite grains rarely occur (Fig. 3.11). About 90% of the limestones examined in thin section are homogeneous with no lamination. This could be through extensive bioturbation and neomorphism (recrystallization).

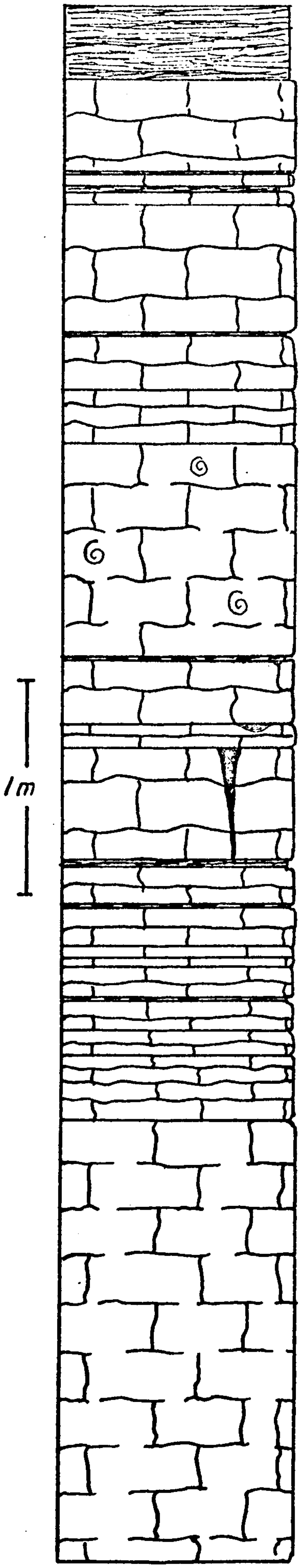
The limestones are generally rich in skeletal debris (biomicro-sparites). The cricoconarids Styliolina sp. and Nowakia sp. are common microfossils in Givetian and Frasnian sediments and thin-shelled bivalves and goniatites are important macrofossils during the Upper Devonian.

The origin of the carbonate in Schwellen limestones is not

Locality	Stage/Stufe	Thickness	Time M/yr	Rate of sedimentation mm/1000 yrs.
Aeketal	Frasnian	6m	3.5	1.7
Riesenbachtal	Frasnian	7m	3.5	2.0
Hühnertalskopf	Frasnian	6m	3.5	1.7
Aeketal	Cheiloceras	1.5m	2.0	0.75
Riesenbachtal	Cheiloceras	4m	2.0	2.0
Huhnertalskopf	Cheiloceras	3m	2.0	1.5
Aeketal	Platyclymenia	3m	3.0	1.0
Riesenbachtal	Platyclymenia	7m	3.0	2.3
Lautenthal	Frasnian	5m	3.5	1.4
(Slope)	Cheiloceras	12m	2.0	6.0
	Platyclymenia	15m	3.0	5.0
	Clymenia/Wocklum.	20m	4.0	5.0
Innerstetal	Frasnian	60m	3.5	17.0
(Basin)	Cheiloceras	50m	2.0	25.0
	Platyclymenia	100m	3.0	33.0
	Clymenia/Wocklum.	90m	4.0	22.5

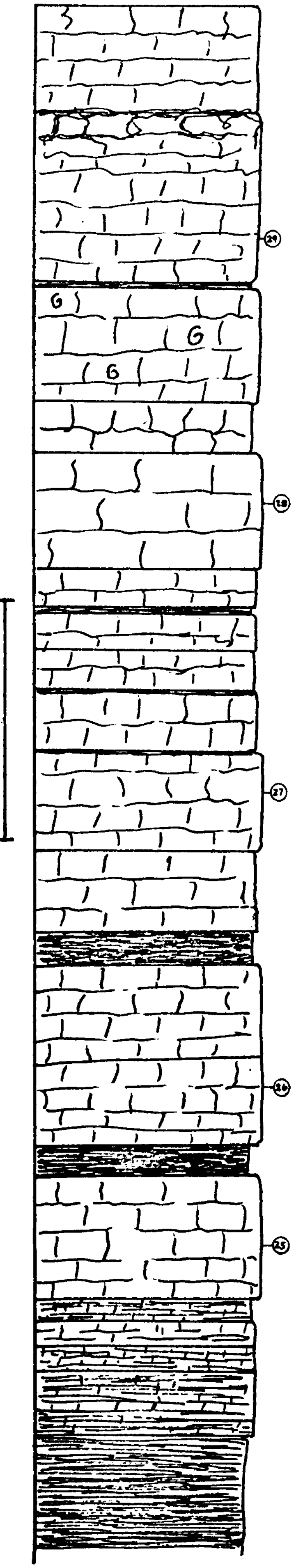
Fig.3.1. Approximate rates of sedimentation for the Schwellen and Becken facies of the West Harz Schwelle. Length of Upper Devonian stages after Friend and House, 1964.

Fig.3.2. Succession at Adorf am Martenberg (Sauerland)



+50 cm	Grey shale with ostracods (*Cheiloceras* Stufe)
40 cm	Light grey Flaser limestone, partly dolomitic.
20 cm	Thin dolomitic limestones.
60 cm	Light grey limestone, dolomitic in the upper part.
25 cm	Red Flaser limestone.
25 cm	Thin bedded pink Flaser limestone.
1 m	Red Flaser limestone, irregular bedding and many goniatites. Thin cavity structures in lower part.
35 cm	Grey limestone band.
12 cm	Thin platey Flaser limestone and cavity structure.
50 cm	Reddish grey Flaser limestone and sedimentary dyke.
60 cm	Thin red and grey limestones containing foraminiferal /algal nodules..
60 cm	Red thin-bedded platey limestone and styliolinids.
2 m	Hematitic limestone with crinoids and cricoconarids, indistinctly bedded.

Fig. 3.3. Upper Devonian Succession at Acketal (Harz)



1 m	Grey Flaser limestone (*Clymenia* Stufe)
1 m 50 cm	Grey Flaser limestone (*Platyclymenia* Stufe)
3 cm	Thin shale horizon.
1 m	Grey Flaser limestone + goniatites and bivalves (*Platyclymenia* Stufe)
30 cm	Flaser limestone.
1 m	Badly exposed weathered limestone (*Cheiloceras* Stufe)
3 m	Thin bedded Flaser limestone + shale partings (Frasnian)
25 cm	Upper Kellwasser limestone rich in styliolinids.
1 m 50 cm	Grey Flaser limestone.
20 cm	Lower Kellwasser limestone.
1 m	Flaser limestone (Frasnian)
+2 m	Calcareous shales with thin limestone bands.

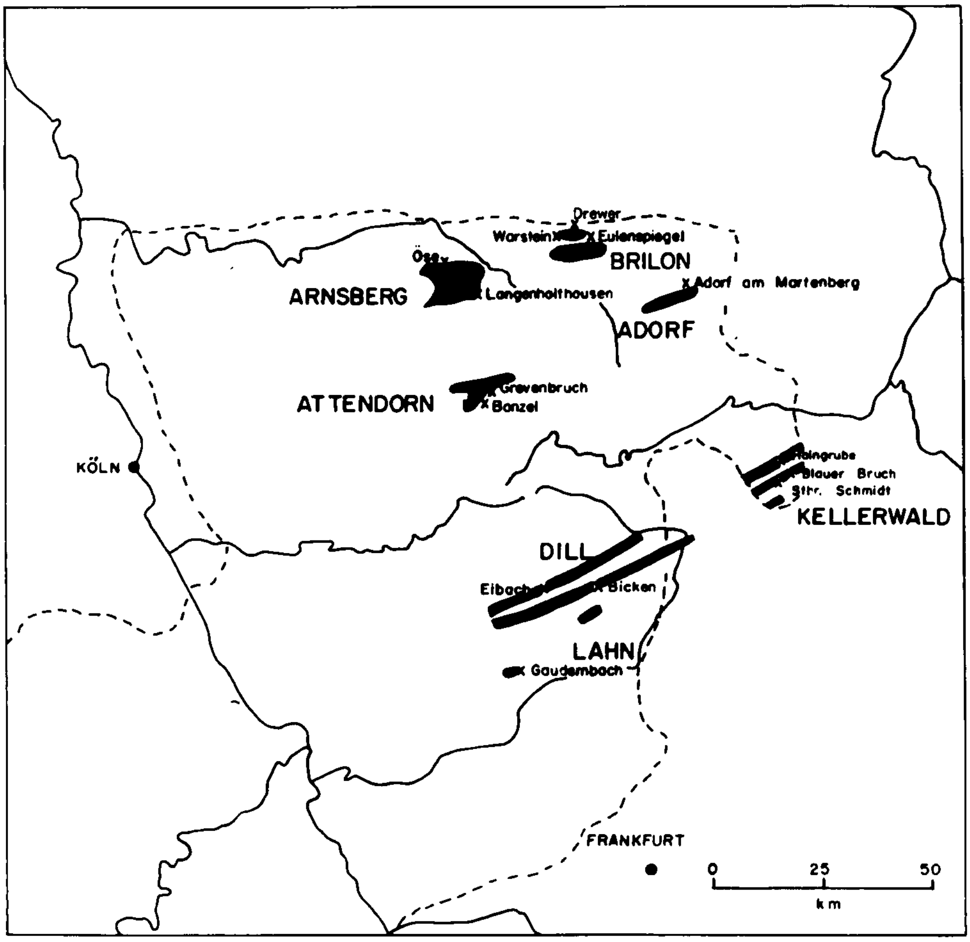
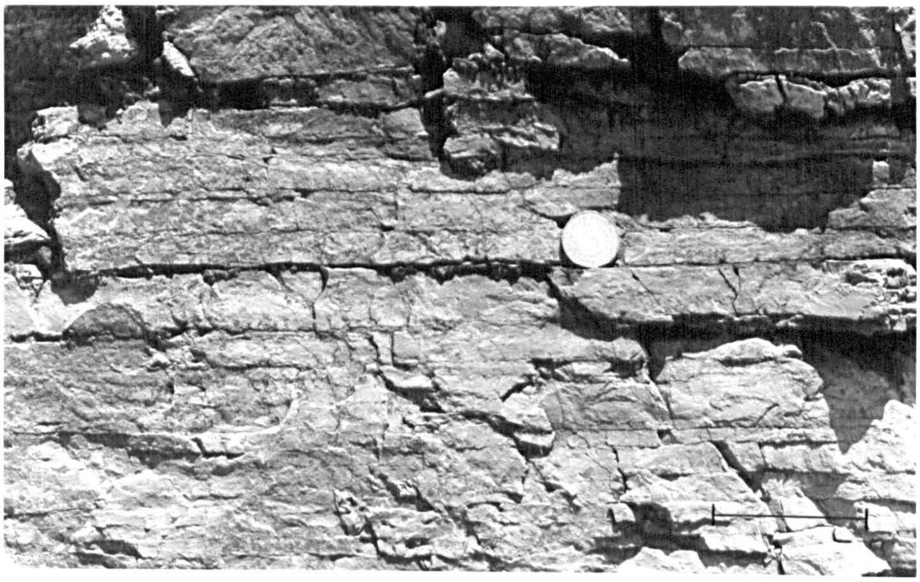


Fig. 3.4 Sketch map of the Rheinisches Schiefergebirge showing location of main outcrops mentioned in the text. Important localities in the Harz Mountains are shown in Fig. 2.6 (p.29). Full details and references of localities are given in the appendix (p. 342).

Fig. 3.5 Bedded Flaser limestones separated by millimetre thick shale partings. Upper Frasnian. Steinbruch Schmidt (Ense Schwelle) Kellerwald.
Scale bar = 10 cm.

Fig. 3.6 Typical cephalopod limestone, grey fine grained limestone cut by horizontal and vertical stylolites (Flasers). Frasnian. Quarry south of Syring (Ense Schwelle) Kellerwald.
Scale bar = 10 cm.



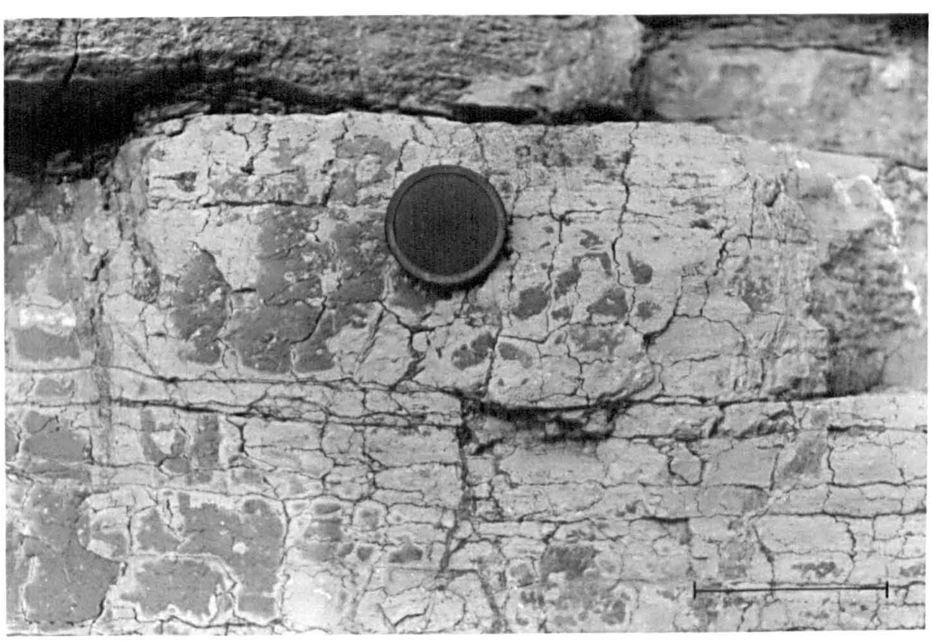


Fig. 3.7 Grey Famennian limestones from Aeketal (Harz) cut by shale streaks (Flasers) parallel to the cleavage direction. Horizontal shale partings also present. Scale bar = 5 cm.

Fig. 3.8 Red hematitic limestone (griotte) cut by numerous shale steaks, giving the rock a brecciated appearance. Scale bar = 10 cm. Lower Famennian. Mont Peyroux, Montagne Noire.

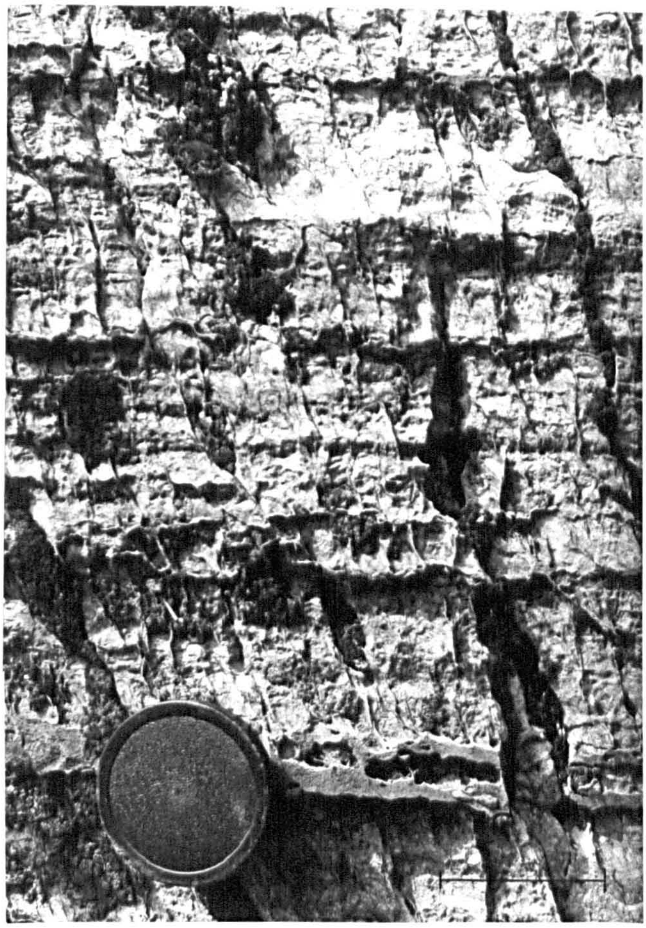
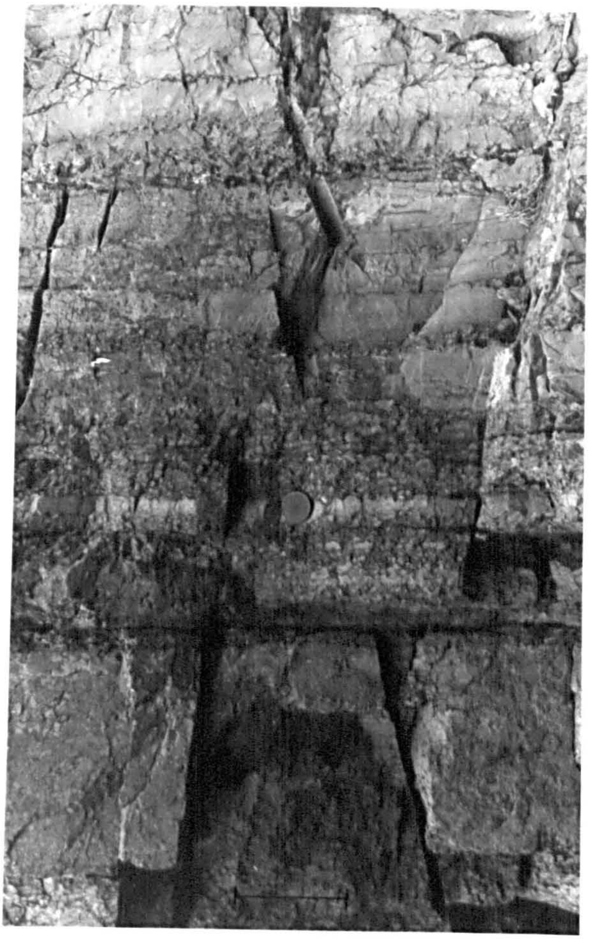


Fig. 3.9 Famennian griotte with prominent horizontal stylolites between beds and nodular horizons. Mont Peyroux, Montagne Noire. Scale bar = 20 cm.

Fig. 3.10 Irregular, nodular bedding surface of griotte, rich in goniatites. Lower Famennian. Mont Peyroux, Montagne Noire. Scale bar = 10 cm.



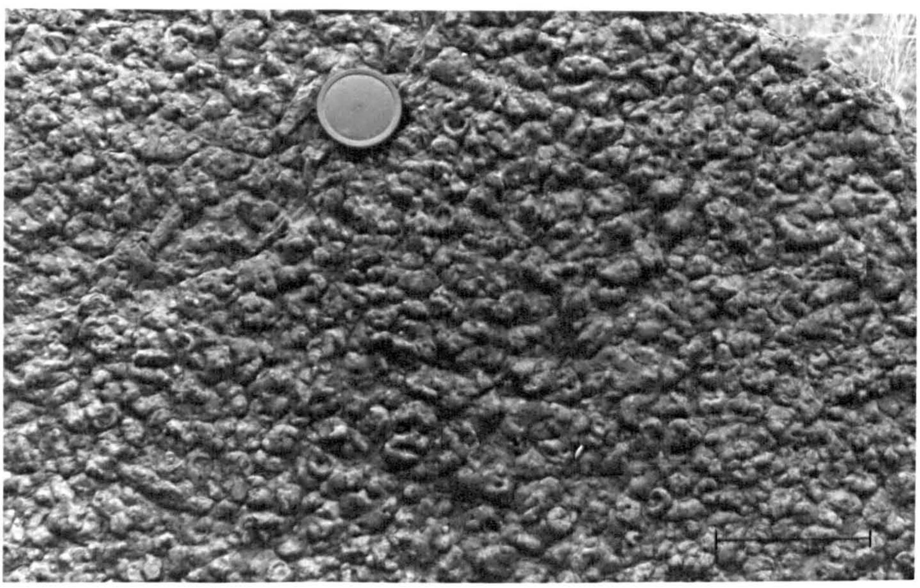
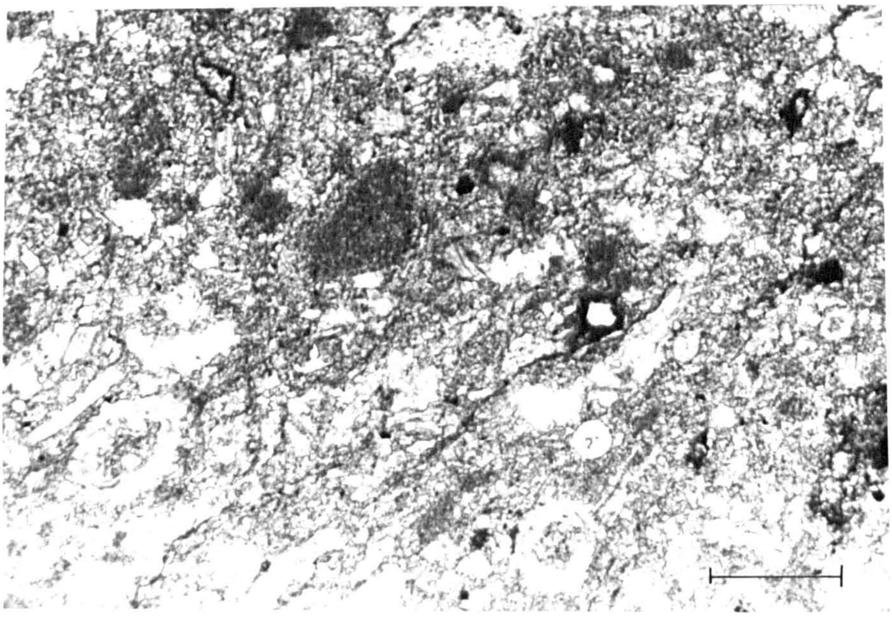


Fig. 3.11 Photomicrograph of microsparitic limestone containing numerous skeletal fragments and pelletoids of micrite grains. Lower Frasnian. Bicken, Dill Syncline. Thin section S 22950b. Scale bar = 200μ.

Fig. 3.12 Electron micrograph of Jurassic pelagic limestone with coccoliths. Lower Jurassic. Maurach, Sonnwendgebirge, Austria. Replica of sample S 23076. Scale bar = 5μ.



known. The facies is very similar to the Jurassic pelagic limestones of the Alps which are locally rich in coccoliths (Fischer et al, 1967). Electron microscopy was used to search the Schwellen limestones for comparable organic structures and a Jurassic limestone was examined for comparison (Fig. 3./2) and to check preparatory techniques (Appendix II). Although coccoliths are recorded from the Devonian Nöel (1961) structures resembling them were not found. Possible organic structures in the Schwellen limestones (Figs. 3./3, 3./4 and 3./5) were rarely encountered and were mostly circular structures with central hollows. Neomorphism has largely obliterated the original texture of the Schwellen limestones. In some cases, the limestones appear to be composed entirely of broken skeletal material which has been altered to microsparite (Fig. 3.//). Sediments such as these with a median grain size of 10μ are coarse compared with Recent and fossil coccolith oozes which are micritic (less than 4μ) (Bramlette, 1958). Coarser pelagic oozes (75% carbonate silt) consisting of comminuted pelagic organisms occur on the Yucatan shelf today (Logan, 1969) and the Schwellen limestones may have been more like these, than finer pelagic deposits.

Other components: Limestone clasts are occasionally found, particularly in neptunian dyke fillings. Ferromanganese encrustations (p.287) occur around shell fragments and limestone clasts at some localities in the griotte of the Montagne Noire (Mont Peyroux, Combe D'Izarne and Coumiac) and at Bicken in Germany. Phosphatic nodules are rare and were only found at Eibach (Dill Syncline).

Quartz grains (up to medium silt) are a minor constituent of the limestones (about 1%) but may locally be 5%. Volcanic material, generally chloritized tuff and basaltic clasts occur in the limestones at Adorf am Kartenberg (Fig. 3./6), Haingrube and Hutthaler Widerwaage.

Kellwasser Limestone. Limestones containing a planktonic and nektonic fauna of bivalves, goniatites, orthocones, ostracods and fish are commonly developed at two horizons during the Frasnian. The limestones are black, bituminous and pyritic, and the absence of trace fossils suggests euxinic conditions during deposition. This facies has been described from Germany, but occurs at about the

Fig. 3.13 Electron micrograph of Schwellen limestone with organic structure. The recrystallised appearance of the matrix with few discrete grains is typical of most Schwellen limestones examined. Upper Frasnian. Dunscombe Farm, Chudleigh, S.W. England. Replica of sample S 22986. Scale bar = 5μ.

Fig. 3.14 Electron micrograph of possible organic structure in Devonian pelagic limestone. Frasnian. Acketal, N.W. Harz. Replica of sample S23038. Scale bar = 2μ.

Fig. 3.15 Possible organic structure in Schwellen limestone. Frasnian. Aeketal, N.W. Harz. Electron micrograph of sample S 23038. Scale bar = 5μ.

Fig. 3.16 Chloritized volcanic fragments cemented by calcite. Frasnian. Adorf am Martenberg, Sauerland. Thin section S 23069. Scale bar = 1 mm.

same time in Belgium, S. France, Pyrenees and North Africa. Schmidt (1935) suggested a situation similar to the Sargasso Sea.

Section 3.2 Sedimentary structures

Generally sedimentary structures are not common in the Schwellen limestones, and some have only been found at a few localities. Structures described are 1) those produced by current activity, 2) hardgrounds, 3) sheet cracks and 4) neptunian dykes.

3.2.1 Sedimentary structures produced by current activity

Current activity is shown by four types of structure, 1) thin graded units of terrigenous material, 2) laminated carbonates, 3) fossil concentrates and 4) corrasional hardgrounds. The last structure is described separately under the section on hardgrounds.

3.2.1.1 Thin graded units of terrigenous material (Figs. 3.17 to 3.19)

Limestones from Aeketal (West Harz Schwelle) and from quarries in the Kellerwald in Ense Schwellen limestone (Schmidt, Syring and Blauer Bruch) commonly contain thin graded units of terrigenous silt and clay. The units are usually 1 to 5 mm thick and contain up to 20% fine and medium silt grade quartz grains. The base of the units may be gradational, but it is usually sharp, locally erosive or irregular with small scour structures (Fig. 3.17). The top of the units is always gradational into the background carbonate. The terrigenous material may be piped down into the carbonate below (Fig. 3.18). In some cases the units are not continuous and this is probably through bioturbation. Pyrite as cubes and aggregates is concentrated in the graded units but there is normally very little skeletal debris compared with the fine grained carbonate between (Fig. 3.19).

The features of these graded units suggest that they were produced by deposition from suspension currents. However, clastic turbidites do not occur in the basinal facies of the N.W. Harz Mountains and are only locally developed in the Rheinisches Schiefergebirge in the Upper Devonian. Thin silty laminae are common in the basinal shales.

Another process which may have been operative

Fig. 3.17 Thin unit of terrigenous silt and clay with irregular base. Skeletal material concentrated in the background carbonate. Frasnian. Aeketal, N.W. Harz. Thin section S 23041. Scale bar= 1 mm.

Fig. 3.18 Lamina of terrigenous material with irregular base and burrow structure filled by terrigenous material and sparite. Upper Frasnian. Steinbruch Schmidt (Ense Schwelle) Kellerwald. Thin section S 22992. Scale bar = 1 mm.

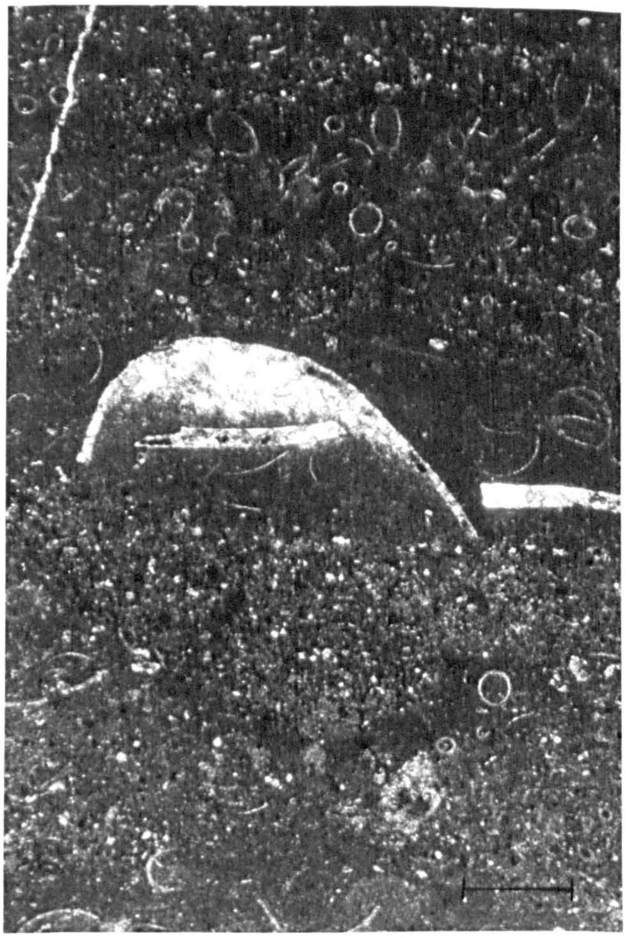
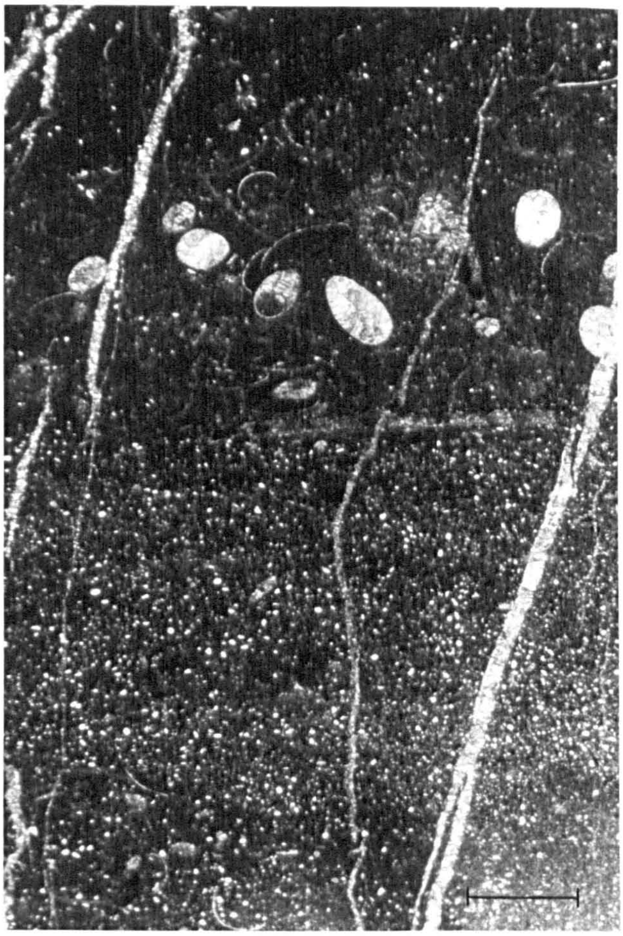
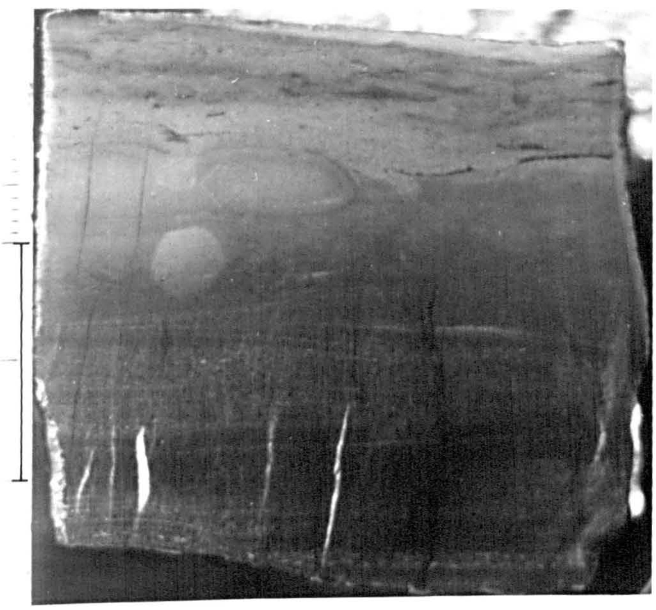


Fig. 3.19 Graded unit of terrigenous silt and clay. Microfossils (ostracods and thin shelled bivalves) common in background carbonate sediment. Upper Frasnian. Aeketal, N.W. Harz. Thin section S 23042. Scale bar = 1 mm .

Fig. 3. 20 Limestone with horizontal and cross lamination. Silty carbonate bioturbated in upper part. Lower Frasnian. Blauer Bruch (Ense Schwelle) Kellerwald. Polished surface S 22973a. Scale bar = 2 cm.





concerns the so-called nepheloid layers. Suspension-rich layers have recently been described from the waters above the continental slope and rise off the Atlantic coast of the United States (Ewing and Thorndike, 1965). These layers range up to 950 m in thickness and are considered to be a medium of transporting fine grained material from shelf areas into deeper waters. The sediment is thought to go into suspension through the action of storms and currents in the shelf. Deposition from such a sediment layer would give graded units rich in silt and clay. Little shell debris would be expected since this would be too coarse to go into suspension. However, as pointed out by Stanley (1969) it is very difficult to distinguish between the deposit from a nepheloid layer, and that from a low density, slow moving turbidity current and maybe the two mechanisms are related. Organic matter too would be brought along with the sediment and could account for the concentration of pyrite in the graded units. Although these suspension-rich water masses occur today at depths of 1000 m or more, depths greater than those envisaged for the Schwellen limestones, the process should still operate at shallower depths.

Quartz grains of fine to medium silt occur scattered throughout the pelagic carbonate, normally constituting about 1%, and some of this silt, along with finer terrigenous material could have been transported by wind. Rex and Goldberg (1958) showed that much of the clastic material less than 4μ in diameter, and some of the quartz grains up to 20μ, occurring in the Recent pelagic sediments of the Atlantic Ocean are aeolian in origin, derived from the Sahara. Wind transportation of several thousand kilometres was reported. The situation must have been very similar in the Devonian with the Old Red Continent subjected to tropical weathering only a few hundred kilometres to the north. Some clastic material in the Schwellen limestones could be derived from the neritic Schwellen.

3.2.1.2 Laminated carbonates

Limestones from Blauer Bruch and Haingrube display horizontal and cross lamination. At Blauer Bruch, Frasnian Flaser limestones are bioturbated microsparites rich in cricoconarids (occurring at all angles to the bedding). Laminated carbonate 1 to 3 cm thick containing less clay material occur within and in the lower parts of limestone beds (Figs. 3.20 and 3.21). The contact of the laminated part with the microsparite below is often irregular, and generally erosive. Cricoconarids are only present in some laminae where they are all parallel to the bedding. The carbonate grains in the laminated part are mostly 5 to 15μ in diameter but silt-size micrite clots occur which may be pelletoids or the products of neomorphism. Cross-laminae cutting horizontal laminae at a few degrees are present in small lenses. Microstylolites have formed between some laminae.

Laminated carbonates from Heingrube are composed of slightly coarser grains (medium to coarse silt). Cross lamination at 5° to 8° to the horizontal is continuous as a thin bed 5 mm thick in one limestone.

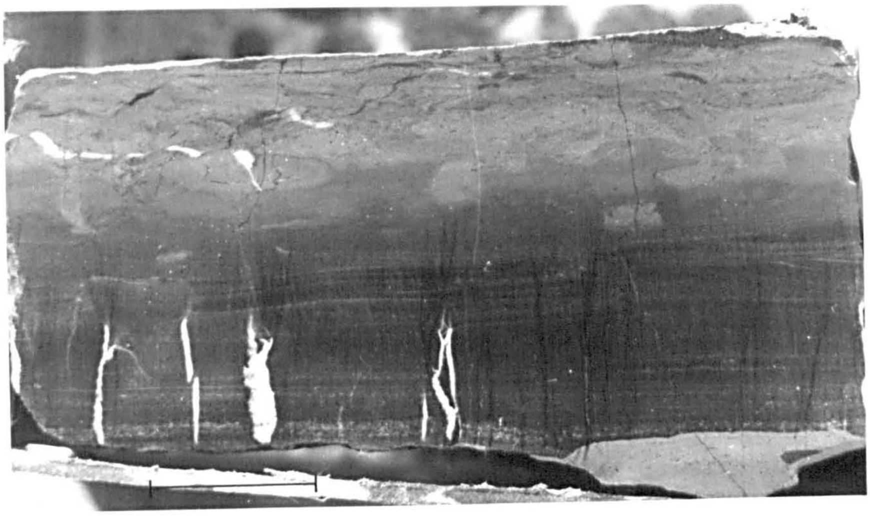
These structures indicate that at certain times current activity worked the pelagic sediment into horizontal lamination and small scale ripples. These structures were probably more frequently developed but have been destroyed by bioturbation and diagenesis.

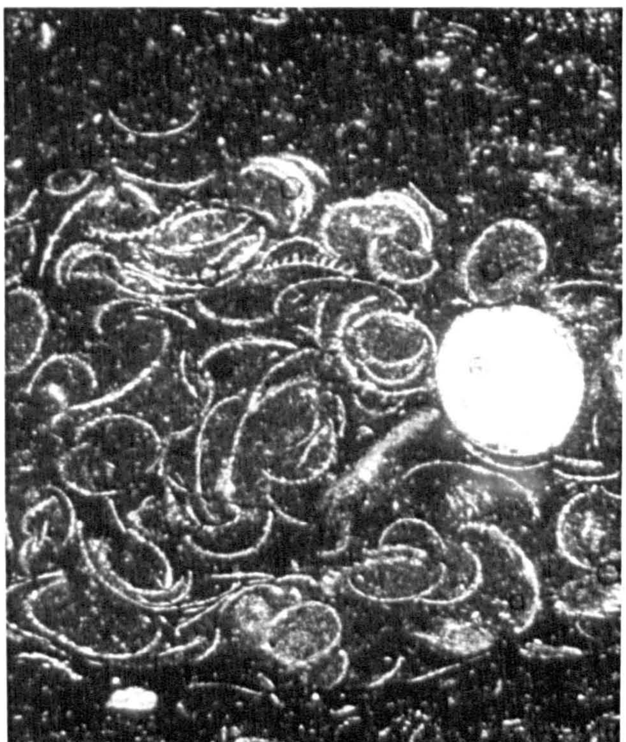
3.2.1.3 Fossil concentrates

Fossils may be concentrated with and without preferred orientation. In most limestones, the fossils are randomly orientated and cricoconarids and goniatites fo example occur normal to the bedding. Microcoquinas consisting of 90% cricoconarids with a preferred orientation commonly occur in Givetian and Frasnian limestones. Such cricoconarid micro-coquinas commonly exhibit a type of graded bedding: at the base of each unit the limestone is largely composed of cricoconarids in contact with each other but with the appearance and upward increase of microspar matrix they grade into cricoconarid free microsparites (thickness of units 5 to 10 cm). Thinner cricoconarid microcoquinas are commonly associated with sheet cracks (p.97). Ostracods and thin-shelled bivalves locally form shell bands less than a centimetre in thickness (Fig. 3.22). In some cases there is very little matrix present and the shells are now in a sparry calcite cement (biosparites).

Fig. 3.21 Horizontal and cross laminated limestone, bioturbated in upper part. Lower Frasnian. Blauer Bruch (Ense Schwelle) Kellerwald. Polished surface S 22973b. Scale bar = 2 cm.

Fig. 3.22 Fossil concentrate of ostracods and thin shelled bivalves in calcareous shale. From a count of 150 shells 33% were concave upwards, 46% convex upwards and 21% vertical or subvertical. Upper Givetian. Hühnertalskopf, N.W. Harz. Thin section S 23072. Scale bar = 1 mm.





Pelagic limestones which were deposited on the flanks of 'reefs' (e.g. Bonzel and Grevenbruch) have thin bands of derived shallow water material. These are usually 2 cm or less in thickness and contain fragments of corals, brachiopods, calcareous algae, stromatoporoids and bryozoans, and silt-size carbonate clasts.

Faunal remains in the Schwellen sediments generally indicate that sedimentation was quiet but that intermittent current activity reworked the sediment and produced shell concentrates. Fine grained carbonate was probably carried off the Schwellen into the slope regions. Current activity can be considerable on modern submarine rises (seamounts, guyots and ridges) compared with the flank and basinal regions, and currents up to 1 m per sec have been recorded (Heezen and Hollister, 1964).

3.2.2 Hardgrounds and disconformity surfaces

Hardgrounds provide evidence of early lithification. A slow net rate of sedimentation, or a pause in sedimentation is one of the main requirements for their formation. Hardgrounds are produced by lithification on the sea-bottom and, once formed, the surface may be modified by later processes (e.g. mechanical erosion), which may facilitate the recognition of these surfaces in the field. Two types of hardground surface (or discontinuity surface) can be recognised in the Devonian pelagic limestones, 1) a planar surface which cuts shells and cavity cements, and 2) an irregular surface with a relief of up to a centimetre which is encrusted by fixosesile foraminifera. The first type is an erosional surface and its nature suggests that the planation was caused by corrasion. Disconformity surfaces of this type occur in the Ordovician and Upper Devonian of Russia (Hecker, 1970) where it is suggested that the abrasion was caused by the movement of shell debris across the surface. They are also recorded from the Middle Jurassic of Britain (A. Kendall, pers. comm., 1971) and the movement of oolith sheets is here invoked as the cause of planation. Purser (1969) and Kazmierczak and Pszczolkowski (1968) describe similar hardgrounds from the Jurassic of France and Poland respectively. One of the features of the erosional type of hardground is their lateral extent, which can be considerable (e.g. up to 50 km in the case of the Polish hardgrounds). Newell (1967) raised some objections to the existence of lithified sediments before planation by erosion, and suggested that a firm substrate only is needed to produce the features of hardgrounds. However, the cutting of shells and cavity cements to the same level as the surrounding sediment, shows that they must have been of the same hardness, i.e. the sediment must have been cemented. Newell's interpretations were made before the occurrence of Recent lithified subtidal carbonate muds was known. In the last few years, there have been many references to this (e.g. Fischer and Garrison, 1967).

Hardground surfaces of the second type have not been previously described. They are characterised by an irregular surface which is encrusted by fixosessile foraminifera. The surfaces were detected in thin section but could not be found in the field, although their presence was known at a definite horizon. They are therefore termed cryptohardgrounds. In some cases, there is evidence that subsolution has occurred. These hardgrounds probably represent local lithification of the sediment as a result of carbonate solution/precipitation in a manner similar to that described by Fischer and Garrison (1967) for Recent lithified subtidal carbonate muds.

3.2.2.1 Hardgrounds with corrasional surfaces (Figs. 3.23 to 3.28, p.76, 79 and 81).

Hardgrounds with evidence of erosion were found at Bicken, Ense Schwelle (a basement rise in the Rheinisches Schiefergebirge) and Mont Peyroux, Montagne Noire (where pelagic sediments follow 'reef' limestones). Both hardgrounds are present in microsparitic limestones of lower Frasnian age. The hardground surfaces are overlain by a few millimetres of calcareous shale before the next Flaser limestone. In the case of the Bicken disconformity surface, the presence of bivalves, orthocones and crinoidal debris within the shale (Fig. 3.23) indicates that there has been no bedding - plane slip at this horizon. The surface at Bicken can be traced for 3 m across the outcrop, and is irregularly exposed over one square metre. The Mont Peyroux surface can be followed for 5 m, and terminates a red hematitic limestone (Fig. 3.24) which is rich in microfossils. Both

Fig. 3.23 Bivalve resting on corrasional hardground surface. Lower Frasnian. Bicken, Dill Syncline. Hand specimen S 23063. Scale bar = 5 cm.

Fig. 3.24 Bedding plane is a corrasional hardground surface. Sediment below enriched in hematite. Lower Frasnian. Mont Peyroux, Montagne Noire. Scale bar = 10 cm.

hardground surfaces cut at several points a thin wavy calcite filled cavity structure or sheet crack (Fig. 3.25 and 3.26). This clearly shows that erosion has taken place, and that the calcite filling the sheet crack is synsedimentary (Purser, 1969). The Mont Peyroux surface also cuts shells with a ferromanganese coating. Ferromanganese nodules occur in the top few centimetres of limestone below the corrasion surface, and in the sheet crack at Bicken (Fig. 3.26). These ferromanganese encrustations are probably associated with a slowing down of the sedimentation rate, which led to the formation of the hardground.

Two types of borings penetrate the Bicken hardground surface; very fine borings, 10μ across within the top 1mm and larger ones, 200-300μ in diameter penetrating down a few millimetres. Borings were not observed in the Mont Peyroux surface.

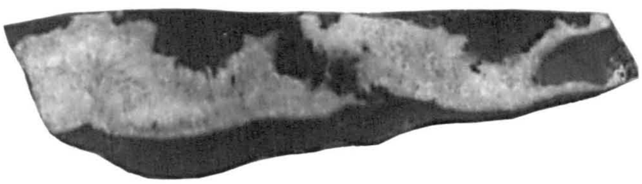
Hematization associated with Mont Peyroux hardground. The sediment below the Mont Peyroux hardground surface is particularly enriched in hematite. Indeed in some places hematite has replaced much of the fauna and in other places, the actual sediment has been replaced. Crinoids and cricoconarids that project into the cavity (Fig. 3.27) are hematized, indicating that this occurred before the cavity-fill stage. Cricoconarids appear to be the material most easily replaced. Crinoids seem to be replaced with difficulty, only pores of the stereom being filled by hematite. On the other hand, where the fine grained matrix is hematized (Fig. 3.28) skeletal material is unaffected. This, however, may be a later diagenetic event, since cricoconarids within a hematized matrix alone possess syntaxial overgrowths. Also hematized matrices occur within more shaley parts of the limestone associated with pressure solution. Pressure solution has also affected the upper portions of some cavities where sutured stylolites outlined by hematite occur.

For the formation of the hardground surfaces the sediment must have been lithified and the smooth nature of the surface indicates corrasion. Other prominent bedding planes occur at Bicken (and other localities) and it is possible that some of these are

Fig. 3.25 Corrasional hardground. Above: view of surface showing truncated ferromanganese encrusted shells, and cavity fill cement. Below: side view showing truncated <u>Stromatactis</u>-like cavity. Lower Frasnian. Mont Peyroux, Montagne Noire. Hand specimen S 22951. Scale as shown.

Fig. 3.26 Section below hardground surface. Corrasional hardground at top has cut a calcite filled sheet crack. A bivalve is resting on the hardground surface. Ferromanganese encrusted shells and clasts occur below the surface in the sediment and sheet crack fill. Lower Frasnian. Bicken, Dill Syncline. Photograph of thin section S 22950a. Scale bar = 0.5 cm.


31
10 20 30 40 50
50 MILLIMETRES



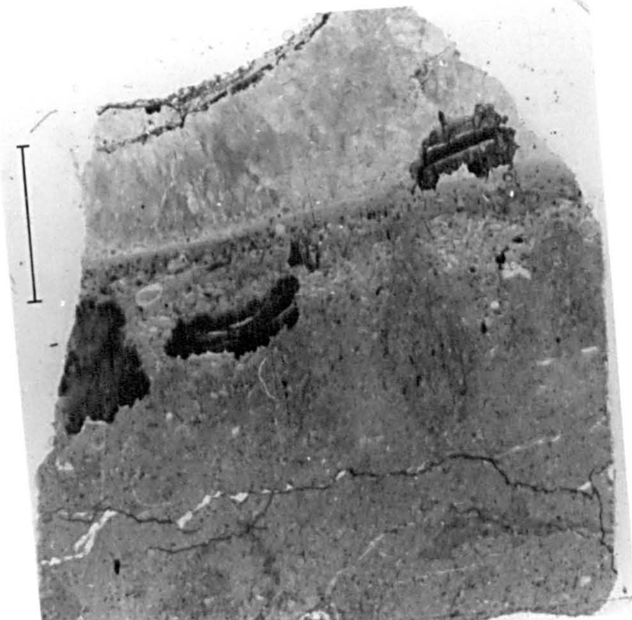
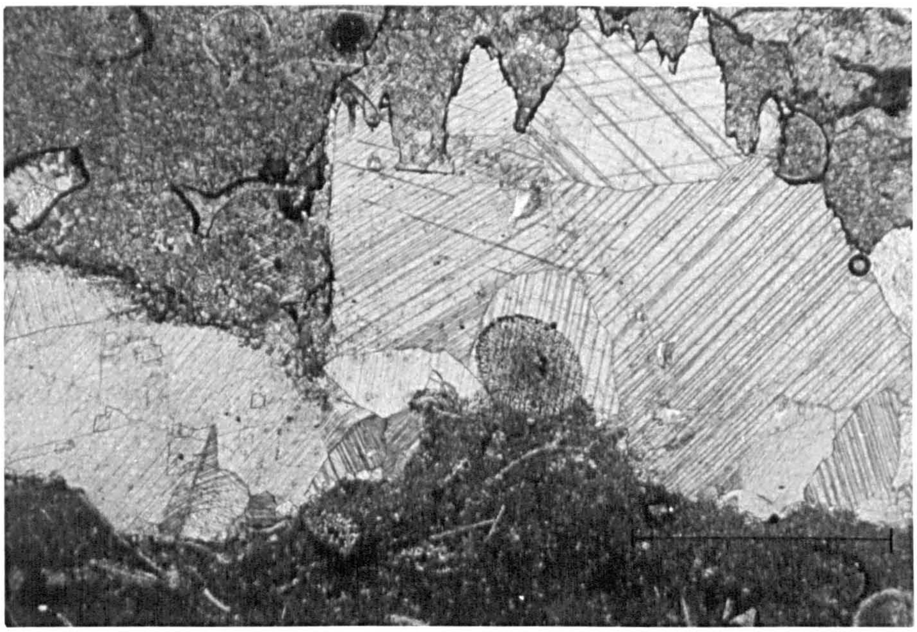
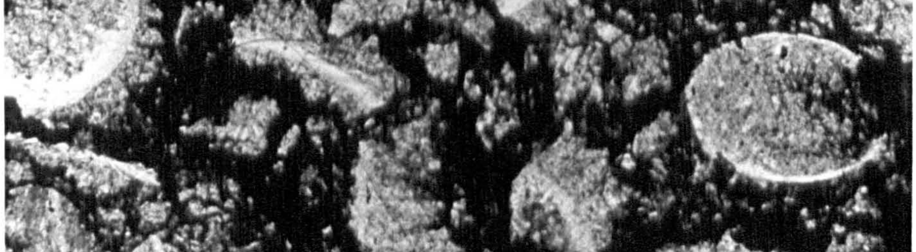


Fig. 3.27 Sheet cavity below corrasional surface. Crinoid fragment in cavity partly replaced by hematite. Hematite also present along stylolite at top of cavity. Lower Frasnian. Mont Peyroux, Montagne Noire. Thin section S 22957b. Scale bar = 1 mm.

Fig. 3.28 Hematite replacement of sediment below corrasional surface. Lower Frasnian. Mont Peyroux, Montagne Noire. Thin section S 22957a. Scale bar = 200μ.





hardground surfaces. However, hardground surfaces can only be definitely recognised by the presence of truncated structures, and tectonic movement and pressure solution have commonly affected bedding planes.

To summarize, the order of events for the corrasional hardgrounds is:-

1) curtailment of sedimentation and formation of ferromanganese nodules and encrustations.

2) lithification of the carbonate mud.

3) formation of sheet cavity before or after initial lithification.

4) replacement of skeletal material by hematite (Montagne Noire surface).

5) cavity filling by internal sediment and cement.

6) corrasion of lithified sediment, cutting shells and the cavity cement.

7) grain growth around cricoconarids followed by hematization of matrix. Exact time of this not known, but probably after 6.

3.2.2.2 Cryptohardgrounds (Figs. 3.29 to 3.31, p.83 and 86).

Hardgrounds with irregular relief were found at three localities,
1) Haingrube (upper Famennian) a volcanic Schwelle in the Kellerwald.
2) Adorf am Martenberg (Frasnian).
3) Riesenbachtal (lower Famennian) West Harz Schwelle.

The disconformity surfaces cut microsparite carbonate and in the first centimetre of sediment above there is usually an enrichment of terrigenous silt and clay indicating that there was a lower rate of carbonate sedimentation. The maximum distance that one of these surfaces has been traced is 14 cm (limited by size of specimen). ~~However~~ In one case a surface was noted to fade into the background sediment. The relief of the surface reaches up to two centimetres, though mostly it is a few millimetres.

The cryptohardground surfaces are encrusted with arenaceous foraminifera, of the genus Tolypammina (identified by Dr. G. Eickhoff). In many cases, the tests have only three sides - the fourth being the firm substrate (Fig. 3.29). Tolypammina is not resticted to the hardgrounds and also occurs encrusting shell fragments and

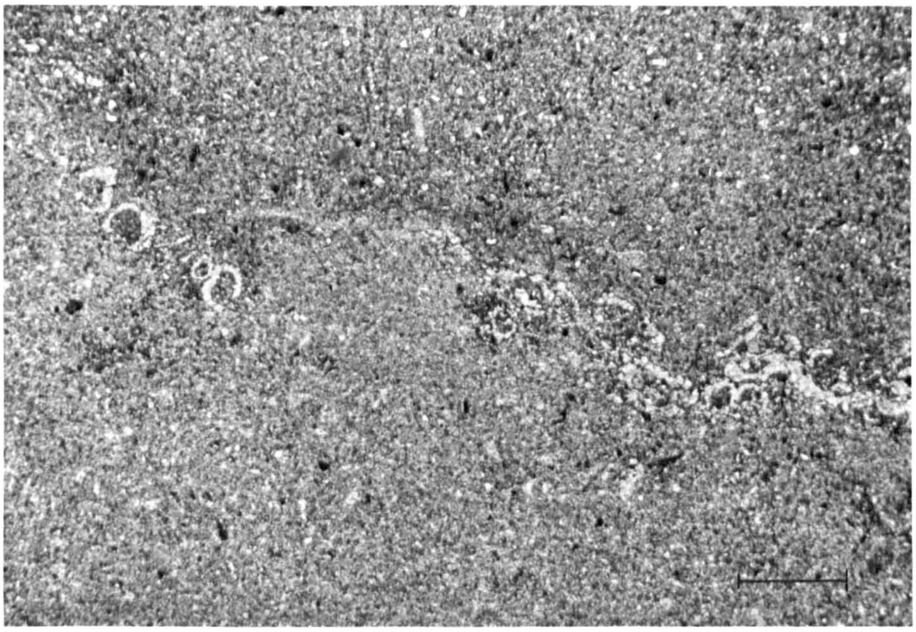


Fig. 3.29 Cryptohardground surface encrusted by arenaceous foraminifera. Upper Framennian. Haingrube, Kellerwald. Peel S 22970. Scale bar = 1 mm.

goniatites and is associated with ferromanganese encrustations (p.287). That the surface was lithified rather than firm is shown by a layer of foraminifera that can be followed over a goniatite with its upper part truncated (Fig. 3.30; c.f. Lindström, 1963). The same situation is shown with a crinoid fragment (Fig. 3.31).

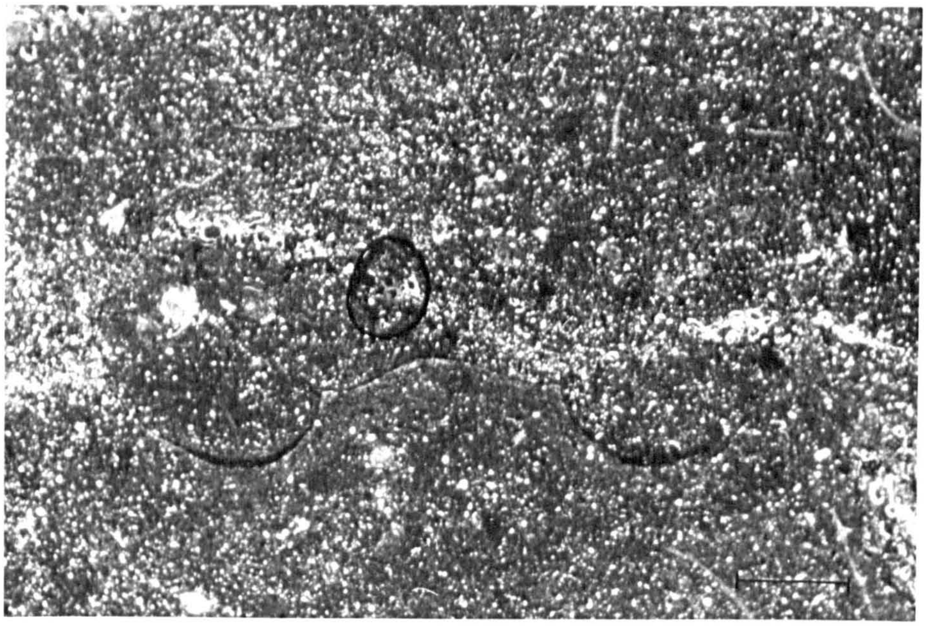
Some of these hardground surfaces have a superficial resemblance to pressure solution planes, suggesting that the foraminifera were encrusting a shell or occurring in the sediment, and after pressure solution were left as part of the insoluble residue. Pressure solution could be invoked for the cryptohardground cutting a crinoid ossicle (Fig. 3.31) but for the truncated goniatite (Fig. 3.30) and the Haingrube surfaces (Fig. 3.29) there is no evidence of a stylolite. Foraminifera of this type are rare as individuals in the sediment , and normally occur on substrates that were obviously hard (e.g. shell fragments). The foraminifera are not deformed as one might expect if pressure solution had taken place at this junction. The persistence of the foraminiferal bands for at least 10 cm or more also precludes a pressure solution origin.

Origin: The presence of a truncated goniatite and crinoid ossicle with the irregular relief of the hardground surface, suggest that subsolution of the lithified sea-bottom occurred, rather than submarine corrasion. The latter may have been operative before subsolution took place. Identical truncation of ammonites occurs in the red pelagic limestones of the Tethyan Jurassic (Hollmann, 1962) (a very similar facies to the Devonian rise sediments) and subsolution surfaces are commonly coated with a ferromanganese crust (Fig. 7. 14,p.305) (Jenkyns, 1970b).

The cryptohardgrounds represent local lithification of the sea floor. This is suggested by the occasional merging of a disconformity surface into the microsparitic limestone. Lithification could be caused by solution and precipitation of calcium carbonate in a similar way to that proposed by Fischer and Garrison (1967) for Recent lithified oozes. Cementation is commonly through high magnesium calcite (Milliman, 1966) which is precipitated as a micrite from sea water and later reverts to low magnesium calcite

Fig. 3.30 Cryptohardground surface with truncated goniatite encrusted by foraminifera. Frasnian, Adorf am Martenberg, Sauerland. Thin section S 22974. Scale bar = 1 mm.

Fig. 3.31 Cryptohardground with truncated and bored crinoid, encrusted with foraminifera. Frasnian. Adorf am Martenberg, Sauerland. Peel S 22975. Scale bar = 1 mm.



with the loss of magnesium. However, aragonite micritic cements (Friedman, 1968), acicular high magnesium calcite cements (Ginsburg and Schroeder, 1969) and low magnesium calcite cements (Garrison et al, 1969) are also reported from subtidal cemented carbonate muds. It is unlikely that the exact process of cementation for ancient carbonate muds can be deduced, since it is impossible to distinguish between grains and the intergranular cement.

3.2.3 Sheet cracks and cavities

Sheet cracks and cavity structures are the most common type of sedimentary structure in the Schwellen limestones and two main types can be distinguished on their size and cements, a) sheet cavities and cracks ranging in height from 1 to 5 cm, some showing many stages of internal sedimentation and more than one phase of cement formation. These have probably formed by shear failure on a slight slope. b) Very thin sheet cracks, 1 cm or less in thickness, persistent over at least 2 m and characterized by kite-shaped calcite crystals filling the void. The second type are probably unique to the Devonian and have not been described before. They are restricted to cricoconarid-rich horizons, and were found at Bicken and Bonzel in the Rheinisches Schiefergebirge.

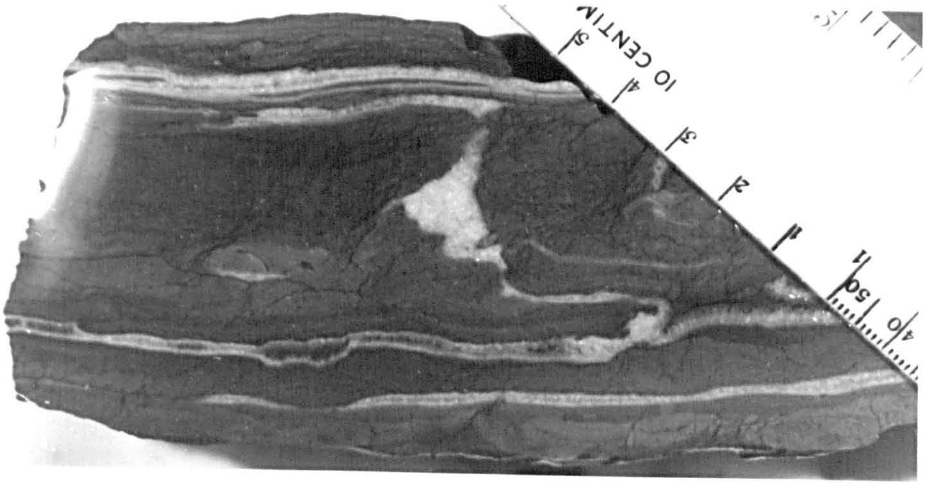
3.2.3.1 Sheet cavities produced by shear failure (Figs. 3.32 to 3.37).

Sheet cavities of this type were found in the Cephalopodenkalk at Adorf, Bicken, Blauer Bruch, Eibach (Rheinisches Schiefergebirge), Langestal (Harz), Mont Peyroux (S. France) and Chudleigh (S.W. England). Slight movement of partially lithified sediment on a slope is considered the most likely mechanism of their formation.

The cavities are always associated with microsparitic limestones with very little clay material present. The cavities vary in size from a few millimetres to 5 cm in height and the lateral extent, commonly 10 to 20 cm, may be up to 2 m. The cracks are mostly parallel to the bedding but occasionally cavities at different levels are connected (Fig. 3.32). In most cases the roof parallels the floor of the cavity both of which are rather flat. Other cracks (e.g. that cut by the Mont Peyroux corrasion surface, p. 79) have a smooth

Fig. 3.32 Thin sheet cracks filled by internal sediment, fibrous calcite and red calcisiltite. Frasnian. Adorf am Martenberg, Sauerland. Polished surface S 22978. Scale as shown.

Fig. 3.33 Deep cavity structure with skeletal fragments and limestone clasts in lower part, and fine graded bands higher up. Burrow structures present in the internal sediment. Small displacement of internal sediment upper left of cavity. Frasnian. Adorf am Martenberg, Sauerland. Polished surface S 22979. Scale as shown.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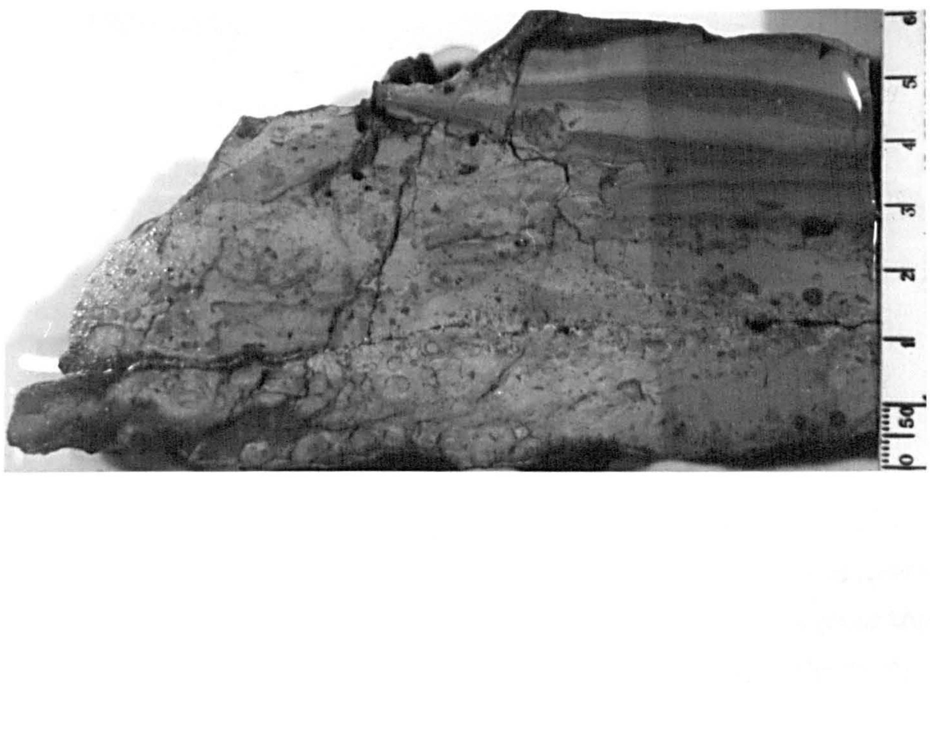


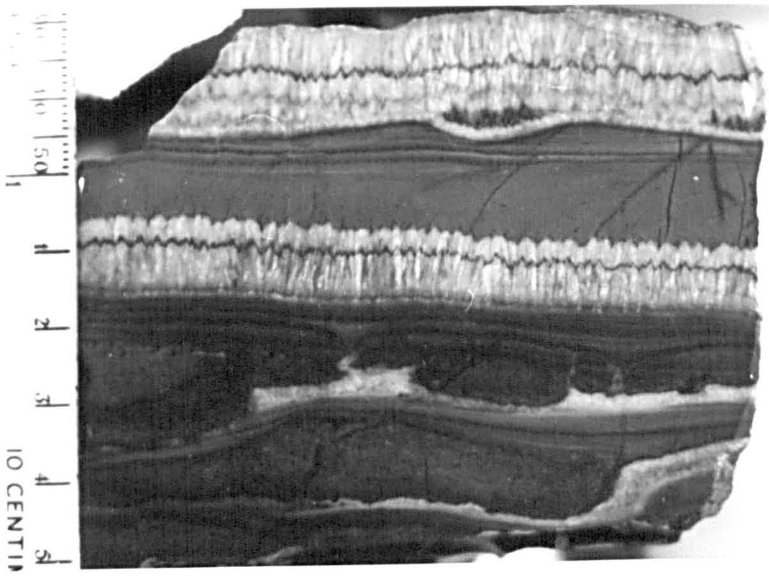
undulating floor, but a more irregular roof. This type can be 2 cm in height, and resemble some of the sheet varieties of '<u>Stromatactis</u>' (Lees, 1964). Some cavities though, cut the bedding quite considerably (Fig. 3.33) and have skeletal fragments and sediment intraclasts in the lowest part of the sediment fill.

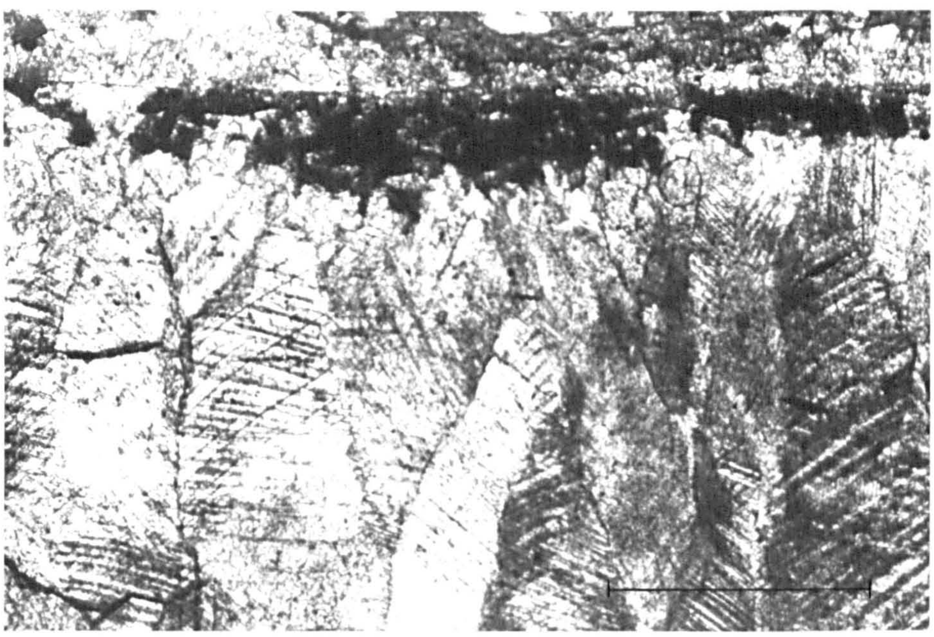
The sediment filling the cavities is thin, normally graded bands of microsparite. However bands may coarsen upwards through a crack, a feature also noted by Schwarzacher (1961). Rarely cross lamination or small channel structures are developed. Red pelite (Krebs, 1969) also forms thin bands in some fills. Small burrows (1 mm in diameter) occur in some internal sediments, filled by a slightly coarser sediment than the surrounding matrix. The final internal sediment is often a coarse hematitic calcisiltite which is later than the main fibrous calcite cement phase. Very thin bands composed of single fibrous calcite crystals may occur between bands of internal sediment. These indicate pauses in the internal sedimentation when cement was precipitated from the ambient water.

Most cavities show an extensive development of fibrous calcite as the first cement, and some crystals may be as much as 1 cm long (Fig. 3.34). The fibrous calcite usually has curved twin planes (concave away from the substrate from which it grew) and subcrystals are developed. Convergent optic axes are present in the crystals. The fabric is the radiaxial fibrous type described by Bathurst (1959a, 1969). The sheet cracks cut by corrasion surfaces at Bicken and Mont Peyroux (p.79) show that the radiaxial calcite is replacive. The calcite filling the sheet cracks is not cut by the corrasion surface, but has seeded from the erosion surface itself (Fig. 3.35). This fabric must therefore have developed after the corrasion surface was cut, by replacing an earlier cement. The possibility that the first cement dissolved.and that the fibrous calcite was precipitated in a void is precluded by the presence of borings into the cavity-fill cement of the Bicken hardground surface. The borings would have fallen to the bottom had there been a void stage. These facts then clearly show that the radiaxial fibrous calcite is replacing an earlier cement. A note on this in press (Kendall and Tucker, 1971) is appended (p.403 to 405).

Fig. 3.34 Large sheet cavity filled by numerous graded units and fibrous calcite. Red calcisiltite is final internal sediment after fibrous calcite. Frasnian. Loose block, Adorf am Martenberg, Sauerland. Polished surface S 22977. Scale as shown.

Fig. 3.35 Sheet crack cement cut by corrasional surface. Fibrous radiaxial calcite crystals have nucleated from the erosion surface, and contain borings. Lower Frasnian. Bicken, Dill Syncline. Thin section S 22950a. Scale bar = 1 mm.





Where a layer of fibrous calcite is overlain by internal red hematitic sediment the crystal terminations may be broken (Fig. 3.36) and fragments of the scalenohedral terminations may be seen in the red sediment above. If this red sediment is similar to the vadose silts of Dunham (1968), then it was introduced at a late stage, above the water table, presumably towards the end of the Lower Carboniferous. The replacement of the early cement by radiaxial fibrous calcite must thus have taken place between the Frasnian and Lower Carboniferous.

The final occlusion of the cavities is by equant sparry calcite. In some cavities, equant calcite has replaced the fibrous variety and large sparite crystals stretch across the whole cavity. However, ghosts of former fibrous crystals can be seen in some of these equant crystals which have straight cleavage planes, but undulose extinction. This type of cavity-fill cement, fibrous calcite and then drusy equant sparite is well documented in the literature (e.g. Bathurst, 1959a; Orme and Brown, 1963; Zankl, 1969; Krebs, 1969).

Rarely, the sediment and cement of a cavity is slightly displaced (Fig. 3.33) which could be due to a settling of the sediment, and shows that it was firm or lithified. Small displacements of this kind in cavity-fill sediments were figured by Orme and Brown (1963).

To summarize, these horizontal sheet cavities show that the carbonate mud was either a cohesive, partly compacted sediment, or lithified to some extent, so that a fairly rigid framework could provide cavities which remained open for some period of time.

Generally the paragenesis recorded here is as follows:-

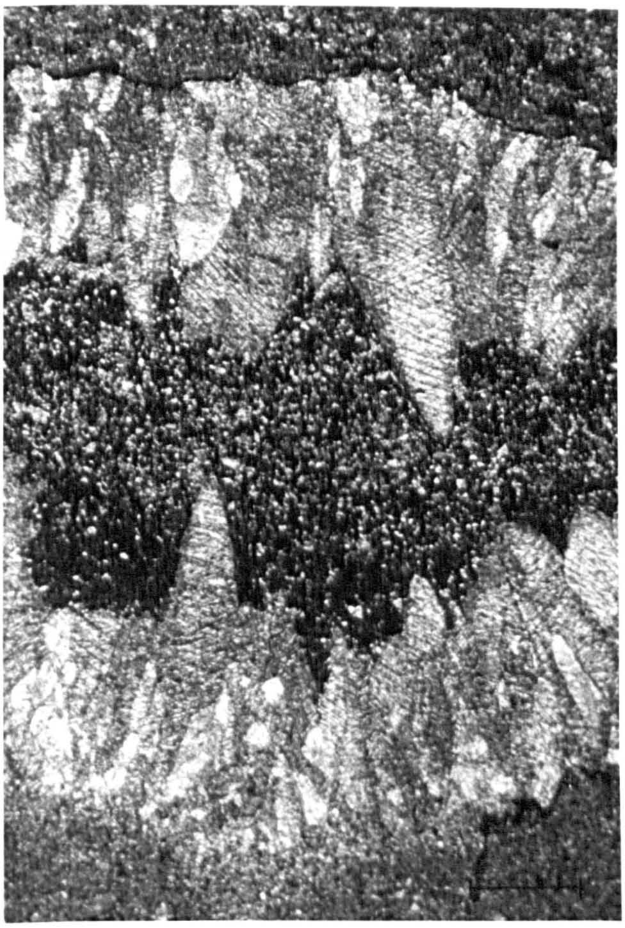
a) compaction or partial lithification of carbonate mud,

b) formation of the sheet cavities,

c) filling of the cavities by the mechanical introduction of internal sediment, and precipitation of an early cement, growing up from the floor, and down from the roof. Cement may also develop during breaks in internal sedimentation. Replacement of the acicular cement by fibrous calcite. Hematitic calcisiltite is commonly brought in after the fibrous calcite has formed. Final occlusion of the cavity is by drusy sparite, which may replace earlier fibrous cements.

Fig. 3.36 Sheet cavity filled by fibrous calcite and red calcisiltite. Some of the crystal terminations have been broken. Frasnian. Adorf am Martenberg, Sauerland. Peel S 22978. Scale bar = 1 mm.

Fig. 3.37 Exposure of cephalopod limestone at Adorf am Martenberg, Sauerland (looking west). The cliff face is of Frasnian limestones, capped by ostracod shales, of lower Famennian age. Rocks exposed in front of the cliff are crinoidal limestones. These sediments were deposited in a small depression on the volcanic ridge, Hauptgrünsteinzug.



Origins of the sheet cracks: Numerous origins of sheet-cracks and cavities have been proposed:- a) an organic origin, caused by the decay of a soft bodied animal, or through burrowing, b) cracks and cavities formed through dessication, c) openings caused by slight movement downslope, d) cavities produced by dewatering of sediment (syneresis cracks), e) cavities formed by internal erosion of sediment and f) tectonic openings.

An organic origin, either due to burrowing, or the former presence of a soft bodied animal, as has been invoked for some 'Stromatactis' structures (Bathurst, 1959) or the 'Stromatactis' spars of Lees (1964) is not applicable to the shhet cavities in the Devonian. The lateral extent, and frequent confinement to the bedding implies a mechanical rather than an organic origin. Fischer (1964) figured similar sheet cracks from the Lofer cyclothems (Dachkalk) of the Alpine Triassic. He interpreted them as dessication cracks since the presence of stromatolites, birds-eyes and mud cracks suggests subaerial exposure. These associated features are completely absent from the Devonian sediments, indicating an exlusively subtidal environment.

Internal erosion alone is unlikely to have formed the cavities but may have occurred once the cavities were formed by some other process. Mechanical erosion has probably occurred in the case of Fig. 3.33 for example, and cut down into carbonate mud. Tectonic movements have created sheet-like cavities in the Alpine Jurassic (Wendt, 1971) but they exhibit evidence for many periods of infill, indicating that they were reopened a number of times. The cavities in the Devonian pelagic limestones were opened and filled during a single event. There is no need to invoke tectonic movements.

Schwarzacher (1961) and Lees (1964) described sheet cracks as another type of 'Stromatactis' structure (sheet spars of Lees) from Carboniferous 'reef-flank' deposits in Ireland. The orientation of the cavities was related to primary 'reef' slopes. Schwarzacher suggested that they formed along planes of shear failure developed during sediment creep, slumping, or compaction at an early stage. For some Devonian sediments with sheet cracks, there is no evidence of an appreciable depositional slope, however, slopes of only a few degrees are needed to cause sediment movement (Lewis, 1971). At Adorf, where the best developed cavities were found, there is evidence of a depositional slope. The limestones at this locality appear to have been deposited in a small basin, at least 10 m across (Fig. 3.37). The present depositional dip is 10^{o}, and could easily have caused some slight movement of sediment. There is no need to invoke lithification before or during the formation of these cavities, as long as the sediment is coherent, cracks can form. However, some of the cracks are quite high (5 cm) and in these cases, at least partial lithification is needed to keep them open. If the sediment were impermeable, cavities will not close by hydrostatic pressure alone (Schwarzacker, 1961). The cavities appear to be confined to relatively pure carbonate muds (suggesting some lithification) for within the more argillaceous sediments downslope, they are absent. It is considered then that small movement of sediment on a slope is the most likely cause of the sheet cavities.

3.2.3.2. Horizontal sheet cracks in cricoconarid rich sediments (Figs. 3.38 to 3.60).

Thin calcite filled sheet-cracks occur in lower Frasnian limestones at Bicken. They are also present at Bonzel, near Grevenbrück (Rheinisches Schiefergebirge) where the pelagic facies accumulated on the flanks of the Attendorn reef complex. Thin detrital bands consisting of displaced shallow-water material occur within cricoconarid-rich sediments. These sheet cracks appear to be confined to this type of sediment and are connected with early cementation of cricoconarid microcoquinas. At Bicken, the sheet cracks occur some 60 cm above the hardground described above (Fig. 3.38) and can be traced for up to 2 m across the outcrop (Fig. 3.39). The cracks have a spacing of 1 to 3 cm and about 10 occur. The calcite cement fill is only 1 or 2 mm thick and is a distinctive kite-shaped fibrous calcite. Below this, internal sediment between 1 and 10 mm thick occurs as a number of thin graded units, some containing cricoconarids. There is some evidence for pulling apart in the cracks from Bicken. If the internal sediment and cement were removed from the cavities (particularly the central part of Fig. 3.40 and Fig. 3.41), it appears as if the cracks could be closed up.

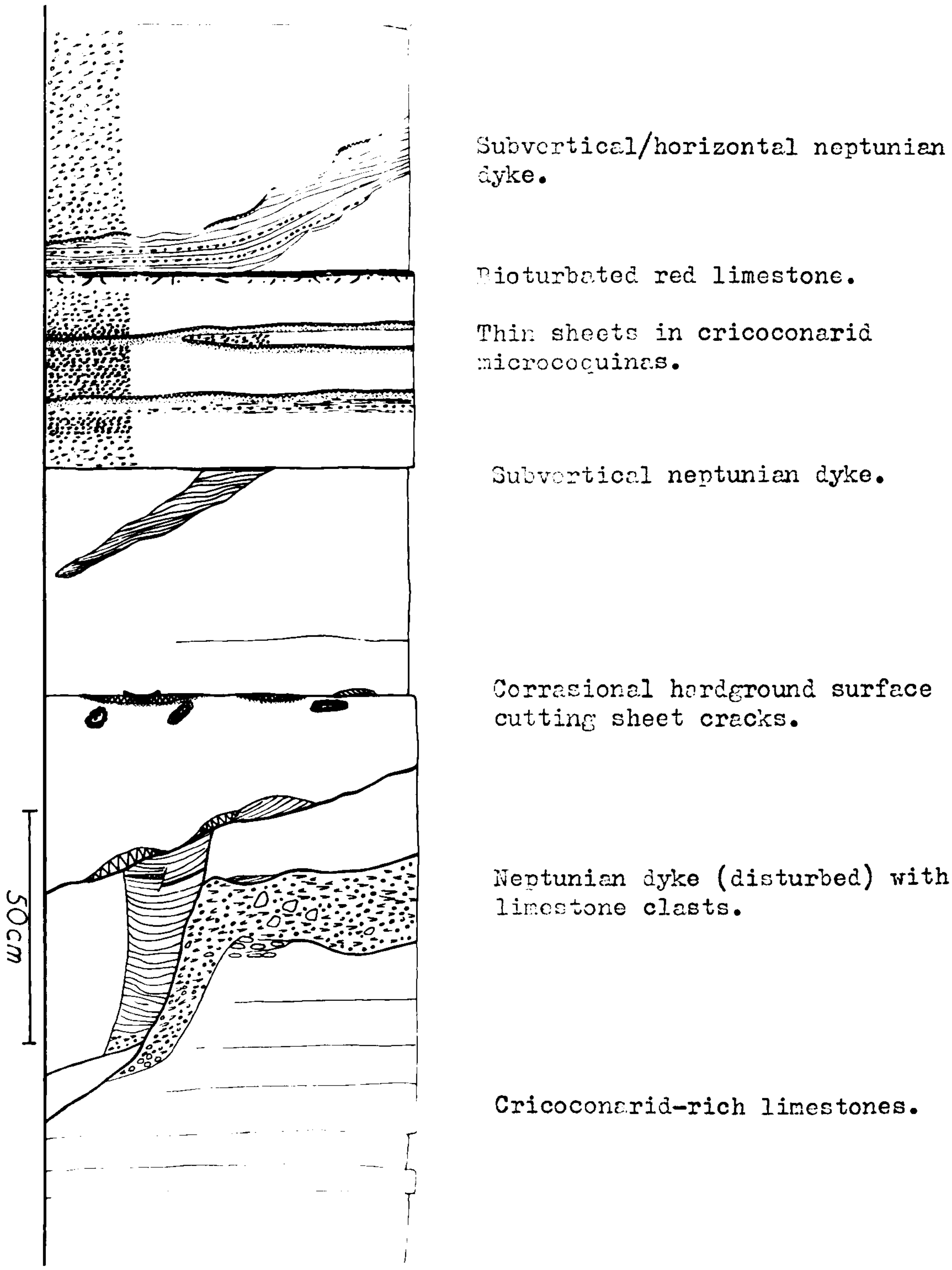


Fig. 3.38 3 m of the lower Frasnian succession at Benner Quarry, Bicken (Dill Syncline). The top of this section is situated 3.7 m below the lower Kellwasser Horizon.

Fig. 3.39 Thin sheet cracks in cricoconarid-rich sediments. Bicken, Dill Syncline. Scale bar = 5 cm.

Fig. 3.40 Sheet cracks in cricoconarid-rich sediments, filled by internal sediment and fibrous calcite. Lower Frasnian. Bicken, Dill Syncline. Photograph of thin section S 23065. Scale bar = 2 cm.

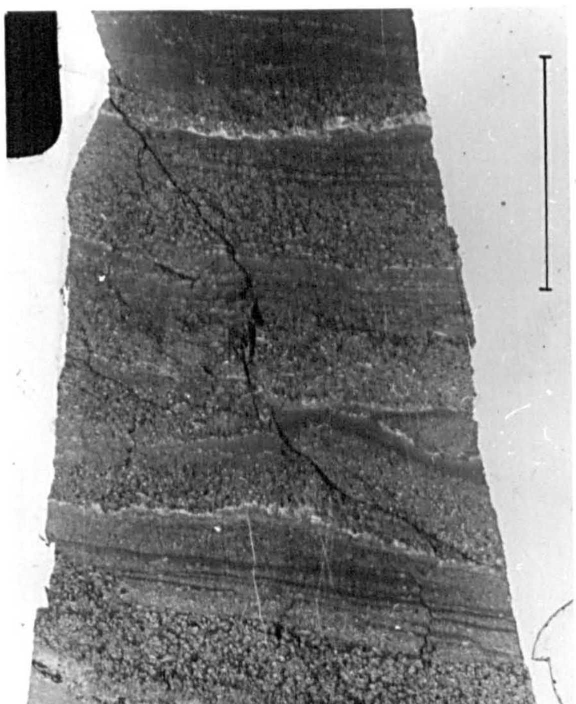
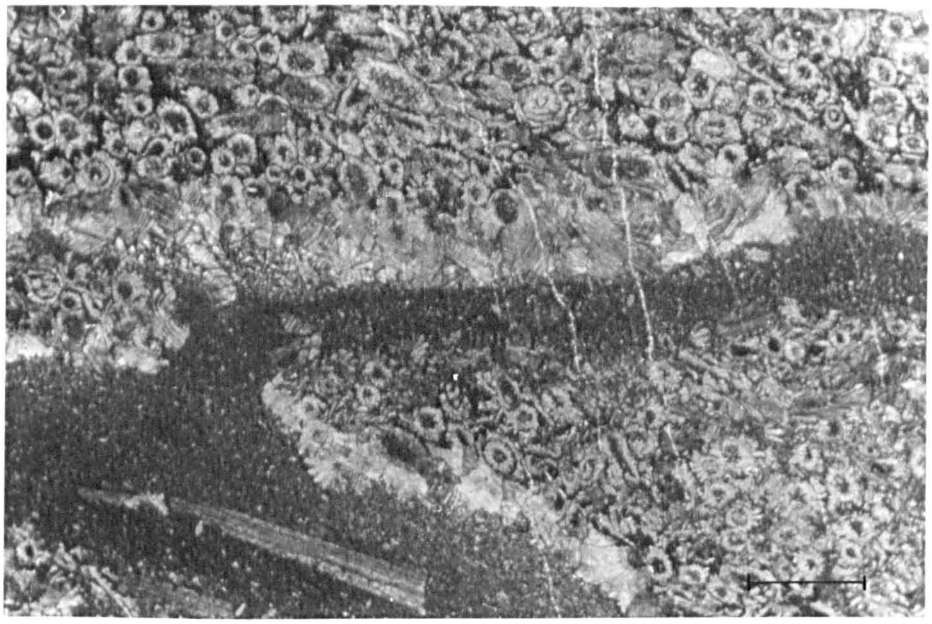
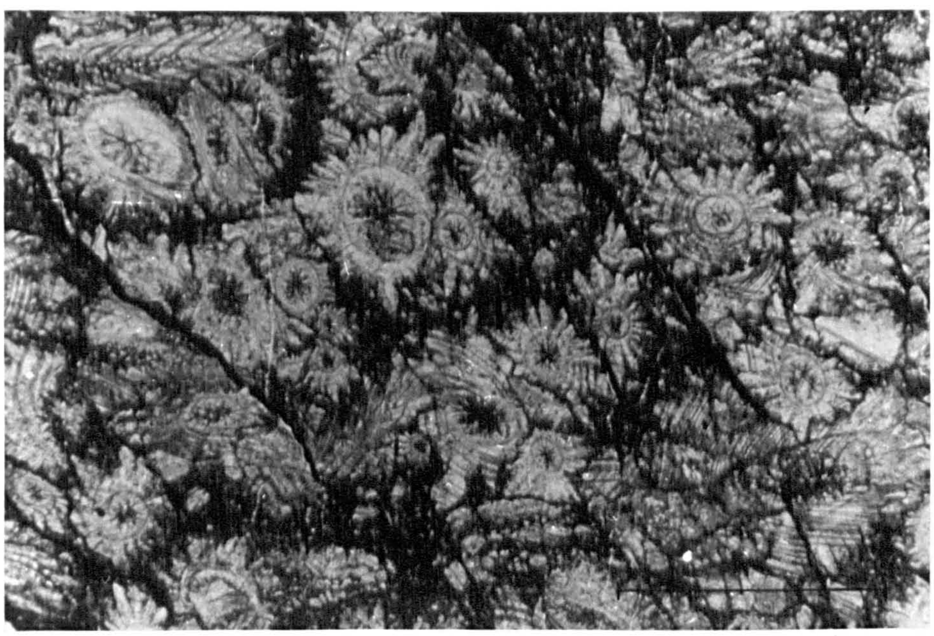


Fig. 3.41 Detail of Fig. 3.40. Sheet crack with matching sides suggesting a pull - apart mechanism. Lower Frasnian. Bicken, Dill Syncline. Thin section S 23065. Scale bar = 1 mm.

Fig. 3.42 Cricoconarids with fibrous calcite overgrowths, some partly removed by pressure solution. Lower Frasnian. Bicken, Dill Syncline. Thin section S 23065. Scale bar = 1 mm.



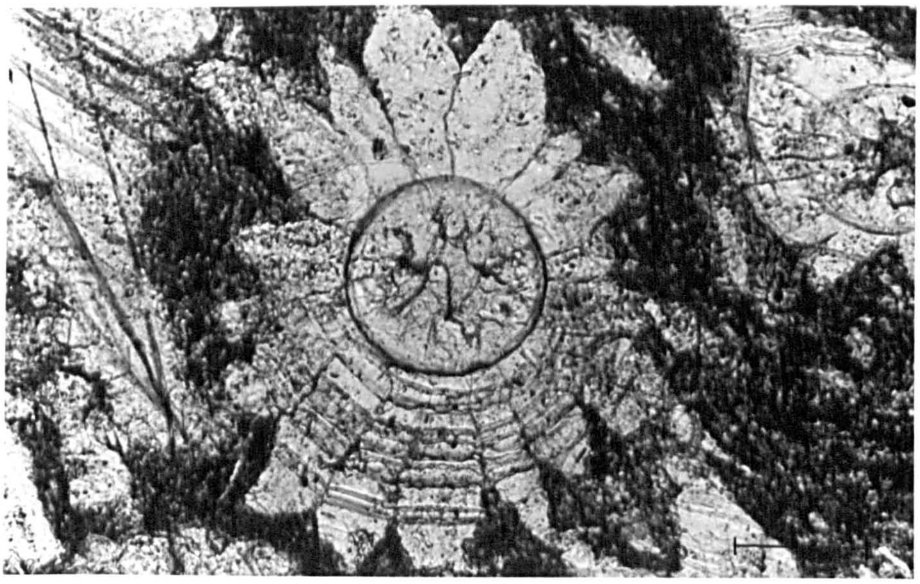


The sediment associated with the cracks is very rich in cricoconarids, and in some parts there is very little matrix at all. There is a preferred orientation of the cricoconarids in thin section, and often two or three cricoconarids occur inside each other. These features suggest current activity (Fisher, 1962). Thin clay seams, less than a millimetre thick, have been injected into the sediment through pressure solution at the same time as the formation of cleavage. The cricoconarids show extensive syntaxial overgrowth development, with the formation of fibrous calcite crystals up to 500μ x 100μ (Fig. 3.42). These have grown outwards and inwards from the walls of the cricoconarids. The centre of the cricoconarid may be filled by sediment or equant calcite. In transverse section, the fibrous calcite overgrowths are directed with their c-axes normal to the criconarid wall. The crystals increase in size outwards, and usually have scalenohedral terminations (Fig. 3.43). Ghosts of former growth stages may be indicated by dusty areas (Fig. 3.44), but ghosts are best seen in the calcite filling the sheet crack. A pseudouniaxial cross is present under crossed nicols (Fig. 3.45). The twin planes $\{01\bar{1}2\}$ are usually very prominent and may be straight (Fig. 3.43). However, they are frequently curved (Fig. 3.44) and the twin planes are always concave towards the styliolinid. The twin planes from one crystal are often coincident with those of adjacent crystals. If this is perfectly developed, then a series of concentric rings is obtained; the whole resembling an oolith (Fig. 3.46). In longitudinal section, the twin planes are always directed towards the open end of the cricoconarid. Fibrous crystals developed along the shell in longitudinal section in the same way as in the transverse section, but commonly, one large overgrowth is developed (Fig. 3.47) which has undulose extinction.

There is substantial evidence to show that the fibrous calcite overgrowths are replacing a former acicular overgrowth, or replacing the host sediment. In transverse section, within the fibrous overgrowths, lines of inclusions can be discerned normal to the cricoconarid wall (Fig. 3.48) (observation of Dr. A.C. Kendall). These could

Fig. 3.43 Fibrous overgrowth around a cricoconarid. Calcite crystals have straight twin planes, but match up with those of adjacent crystals. Lower Frasnian. Bicken, Dill Syncline. Thin section S 23065. Scale bar = 200μ.

Fig. 3.44 Fibrous overgrowth around a cricoconarid. Ghosts of earlier growth stages can be seen, and twin lamellae are curved and continuous. Lower Frasnian. Bicken, Dill Syncline. Thin section S 23065. Scale bar = 100μ.



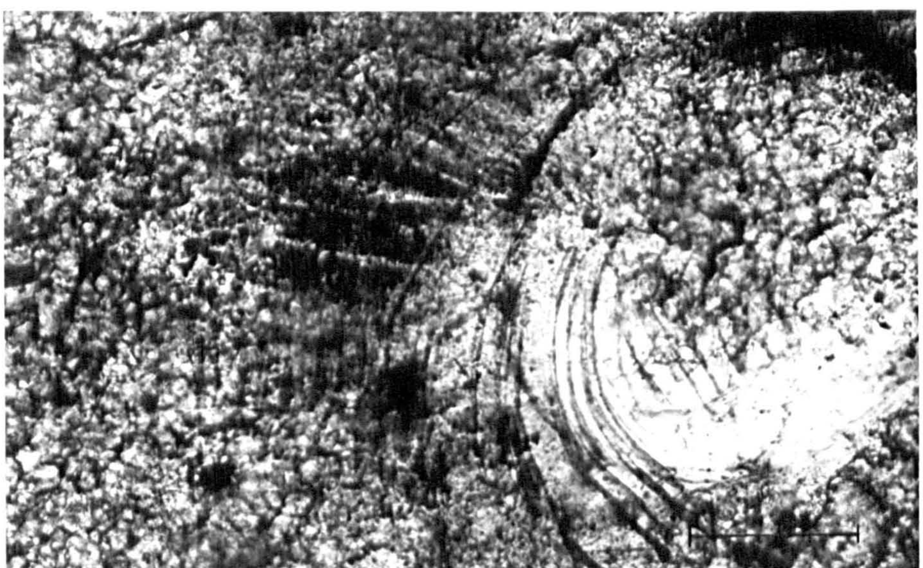


Fig. 3.45 Pseudouniaxial cross obtained with fibrous overgrowths around cricoconarids, under crossed nicols. Lower Frasnian, Bicken, Dill Syncline. Thin section S 23065. Scale bar = 1 mm.

Fig. 3.46 Pseudo-oolitic structures produced by continuous twin planes around cricoconarids. Lower Frasnian. Bicken, Dill Syncline. Thin section S 23064. Scale bar = 200μ.

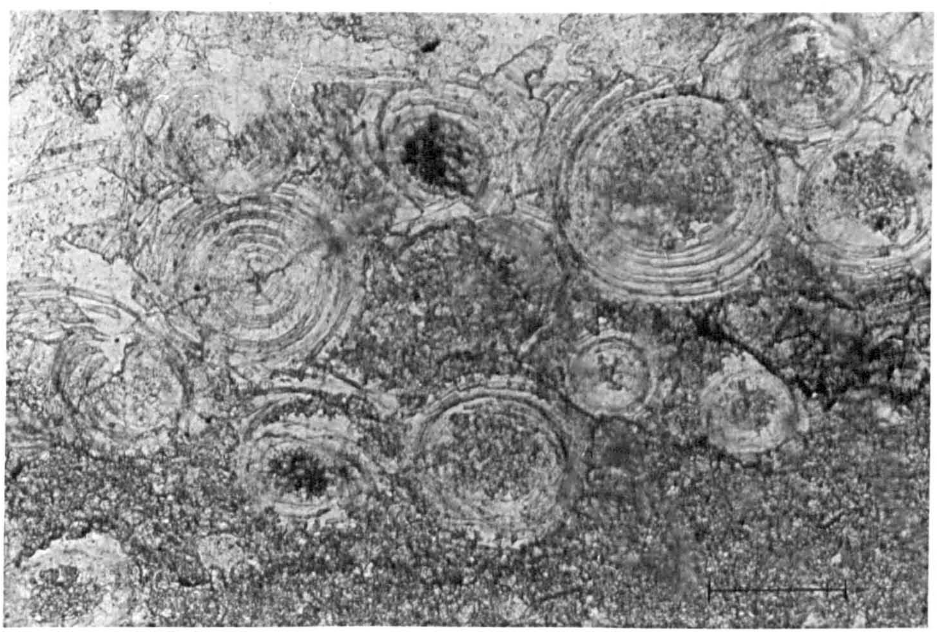
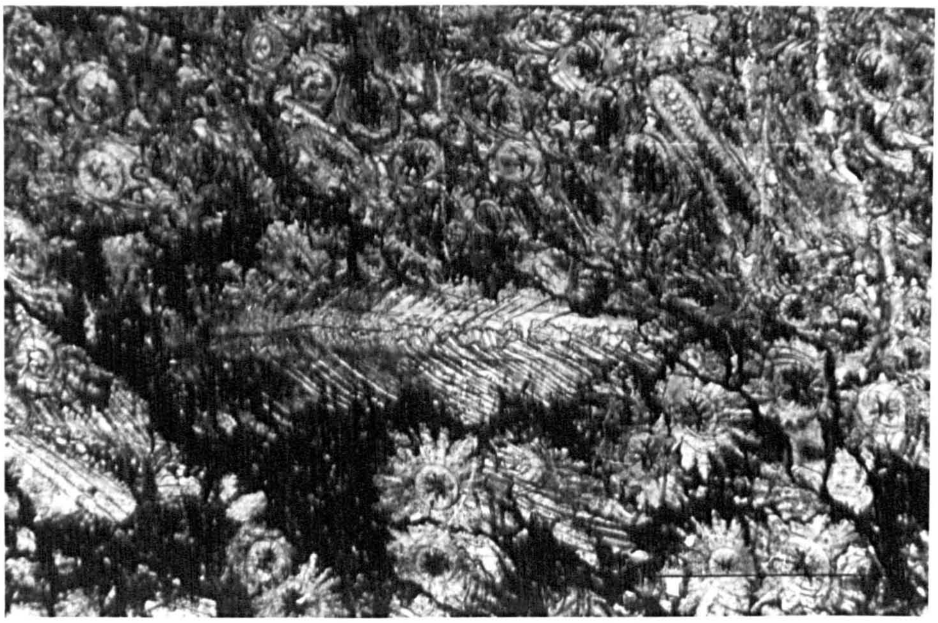
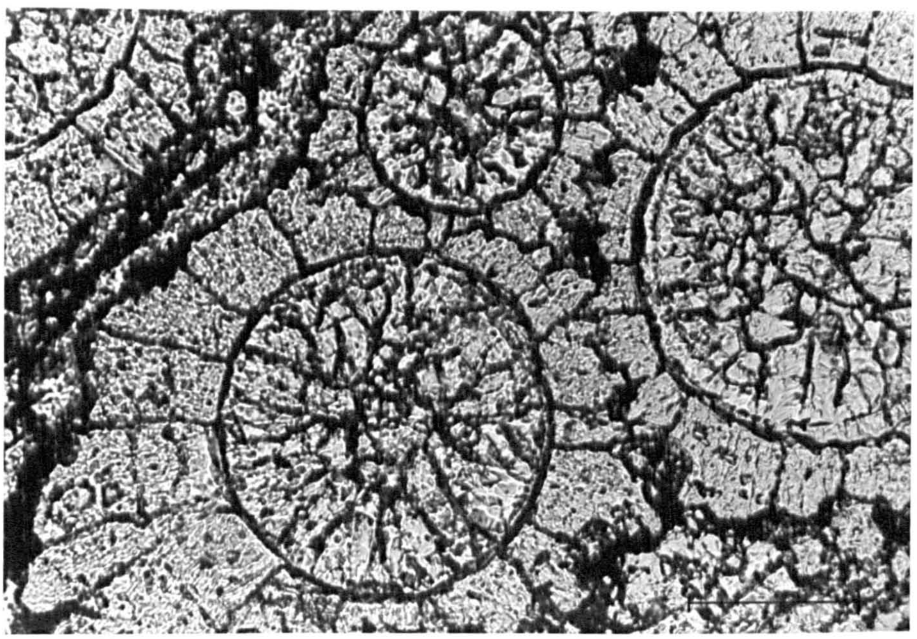


Fig. 3.47 Longitudinal section of cricoconarid with fibrous overgrowth. Lower Frasnian. Dicken, Dill Syncline. Thin section S 23065. Scale bar = 1 mm.

Fig. 3.48 Fibrous calcite overgrowths around cricoconarids showing lines of inclusions radial to the shell wall. Lower Frasnian. Langestal, N.W. Harz. Peel S 23045. Scale bar = 100μ.



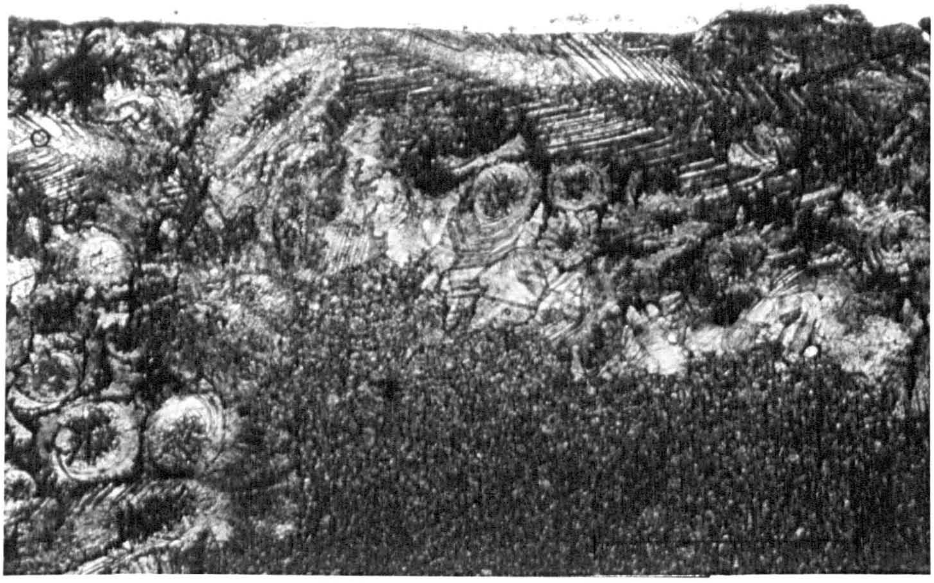


only occur if the fibrous calcite crystals had replaced an earlier acicular carbonate, which grew radially from the styliolinids. On theoretical grounds, it is easier to explain the fibrous calcite overgrowths as replacive. Since cricoconarids have a prismatic shell structure (Fisher, 1962), one would expect a greater number of seeding points for crystal overgrowth development, than are indicated by the number of fibrous calcite crystals seen now as the overgrowth. In a number of places, particularly where a geopetal cavity has formed inside a cricoconarid, the overgrowth can be seen replacing the host sediment (Fig. 3.49). The fibrous calcite within the cavity is clear, but that outside is full of inclusions and has obviously replaced the sediment. This situation is well seen in a section from the Montagne Noire (Fig. 3.50) where the twin planes of the overgrowth clearly pass into the surrounding hematitic sediment. Other evidence that this type of calcite has replaced an earlier carbonate is given above (p. 90).

Sheet crack cement fill. The sheet cracks at Bicken are filled by large kite-shaped crystals of fibrous calcite, which may be up to 2 mm x 150μ in size. These are also overgrowths, developed radially from cricoconarids that either project into the cavity or are just within the sediment above the crack (Fig. 3.51). Smaller fibrous crystals of the type described above, may be present on the upper parts of cricoconarids which have large fibrous crystals growing down into the cavity. It seems reasonable to infer from this that the fibrous crystals in the sheet crack have grown in a similar way to those around the cricoconarids in the host sediment, i.e. by replacing an earlier acicular cement. The fibrous calcite crystals filling the cavity also have well-developed twin planes which are mostly straight, but nearer the cricoconarid they tend to be curved. The twin planes of adjacent crystals frequently coincide in the same way as those overgrowths developed in the host sediment. The extinction may be undulose, dividing large grains up into subcrystals as in radiaxial calcite (Bathurst, 1959a, 1969), but more commonly each crystal extinguishes at the same point (Fig. 3.52). Within many of the kite-shaped crystals, ghosts of earlier growth

Fig. 3.49 Longitudinal section of a cricoconarid showing replacement of host sediment by fibrous overgrowth. Lower Frasnian. Bicken, Dill Syncline. Thin section S 23065a. Scale bar = 1 mm.

Fig. 3.50 Longitudinal section of cricoconarid projecting into sheet crack. Fibrous overgrowth has replaced host sediment, and grown in continuity with adjacent calcite crystals in the cavity. Lower Frasnian. Mont Peyroux, Montagne Noire. Thin section S 22957a. Scale bar = 1 mm.



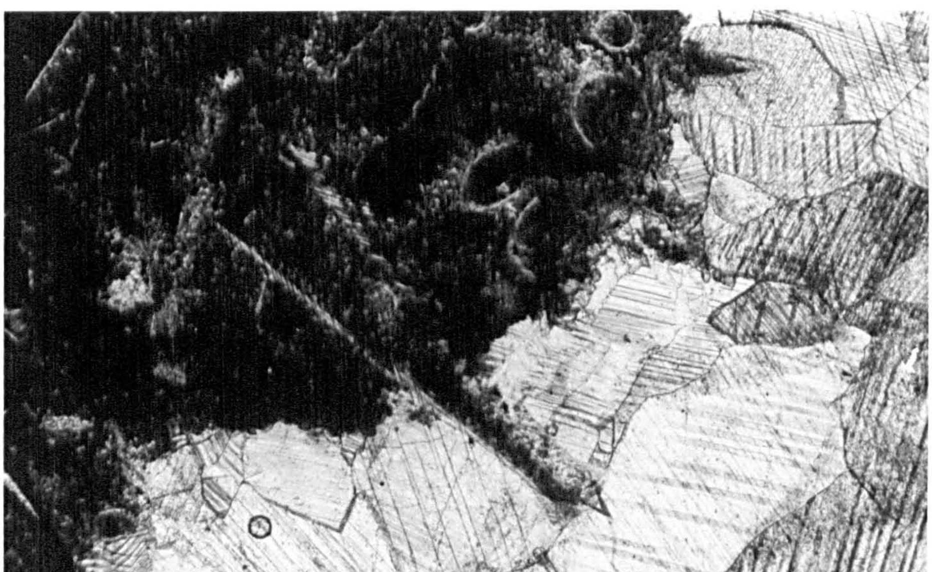
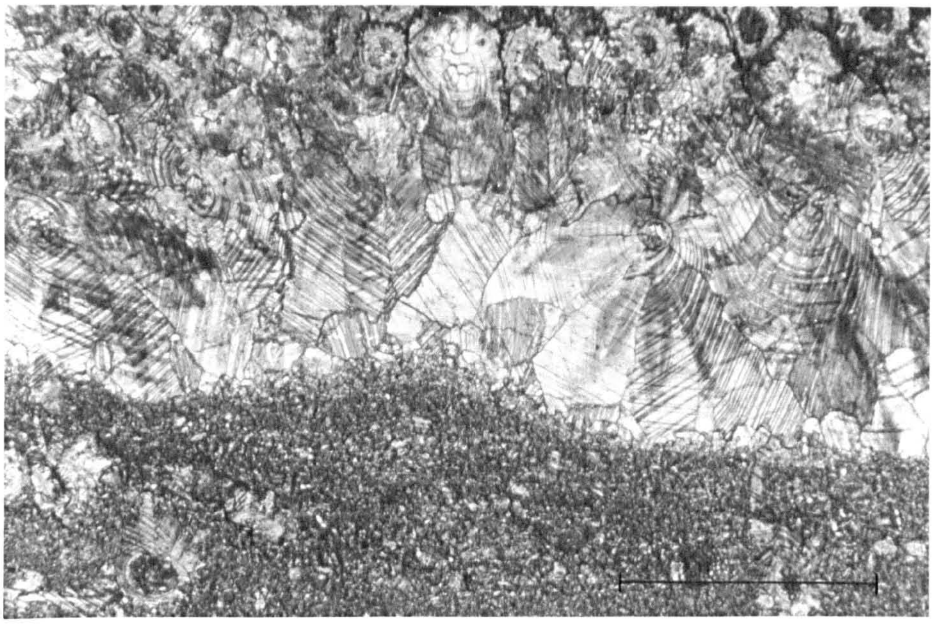
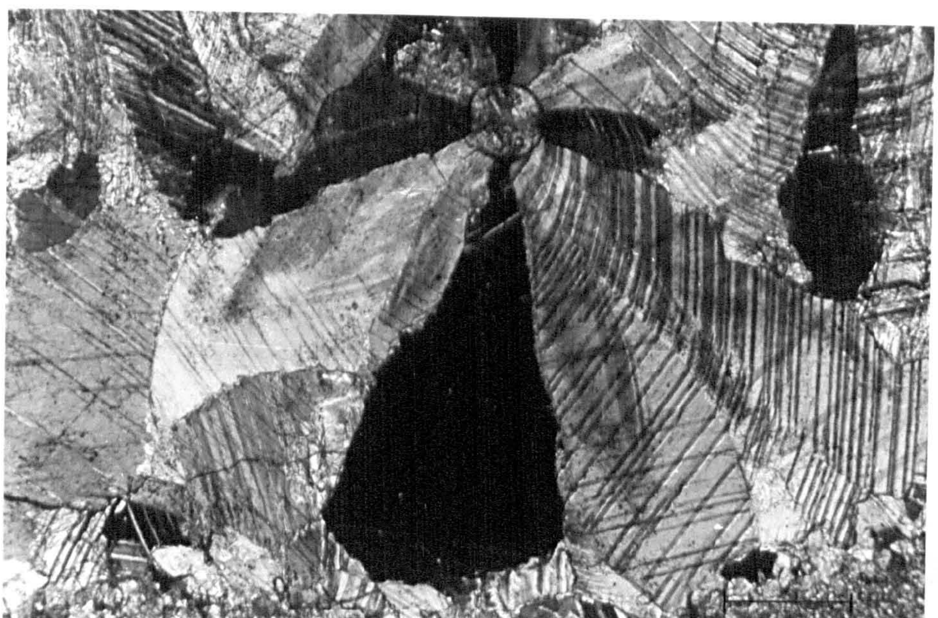


Fig. 3.51 Kite-shaped fibrous calcite crystals, which have grown from cricoconarids, filling sheet crack. Lower Frasnian. Bicken, Dill Syncline. Thin section S 23065. Scale bar = 1 mm.

Fig. 3.52 Fibrous calcite crystals with complete extinction under crossed nicols. Ghosts of scalenohedral terminations of earlier growth stages present within fibrous crystals. Lower Frasnian. Bicken, Dill Syncline. Thin section S 23065. Scale bar = 200μ.





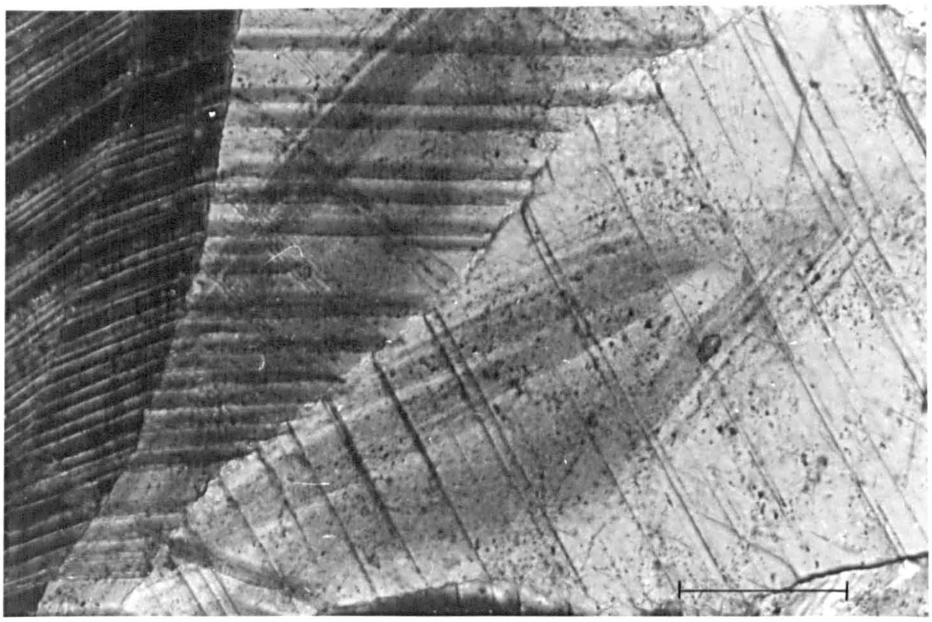
stages can be seen, with fine scalenohedral terminations (Figs. 3.52 and 3.53). The presence of these ghosts is shown by a light brown/ colourless banding. Inclusions are present in the darker bands making the crystals pseudopleochroic. In longitudinal section, growth stages of the same type are normal to the shell wall, but all now included in one large overgrowth crystal (Fig. 3.54). Where the ghosts are not so well preserved, their former presence is indicated by a dusty area full of inclusions. The light brown pseudopleochroic bands contain many inclusions which could be organic matter or clay particles. Hudson (1965) described brown pseudopleochroic calcite from Jurassic bivalves and attributed the colour to organic matter. Similar zonation patterns due to varying amounts of iron in the calcite lattice, were described by Evamy and Shearman (1965; 1969) and Dickson (1967). Staining of the samples from Bicken with potassium ferricyanide and Alizarin Red showed that ferroan calcite is not present, but only calcite. Adjacent crystals have slightly irregular consertal contacts, and it can be seen that material has been lost here. The outer parts of the earlier growth stages have been lost along the junction with adjacent crystals. This probably occurred through competition during growth of adjacent crystals as they replaced the original acicular crystals. The fibrous calcite in the sheet cracks is replacing a carbonate which was clearly a void-filling cement.

The sheet cracks at Bonzel are slightly different in that secondary cracks are also present. The main cracks are about a centimetre thick and filled by grey and red internal sediment, and then fibrous calcite. The latter is of two types, fibrous calcite that is an overgrowth from cricoconarids, and radiaxial fibrous calcite (Bathurst, 1959a; 1969) which has mainly grown up from the cavity floor (Fig. 3.55). Secondary cavities formed where partly lithified host sediment cracked and slipped down slightly towards the main cavity (Fig. 3.56). The smaller cracks are filled by fibrous calcite which has mainly grown from cricoconarids.

Internal sediment in sheet cracks. The internal sediment now filling the sheet cracks at Bicken is a microsparite, with grains up to

Fig. 3.53 Ghosts of earlier growth stages within fibrous calcite overgrowths filling a sheet crack. Lower Frasnian. Bicken, Dill Syncline. Thin section (crossed nicols) S 23065. Scale bar = 100μ.

Fig. 3.54 Ghosts of earlier growth stages seen in longitudinal section of cricoconarid. Lower Frasnian. Bicken, Dill Syncline. Thin section S 23065. Scale bar = 1 mm.



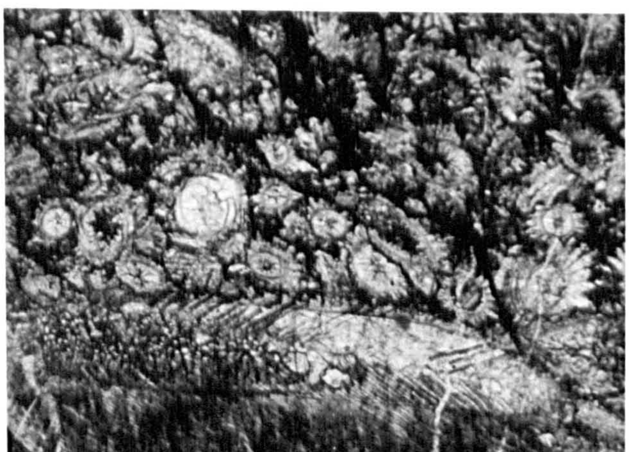


Fig. 3.55 Sheet crack filled by radiaxial calcite and overgrowths from cricoconarids. Givetian/ Frasnian. Bonzel, Attendorn region. Thin section S 23067. Scale bar = 1 mm.

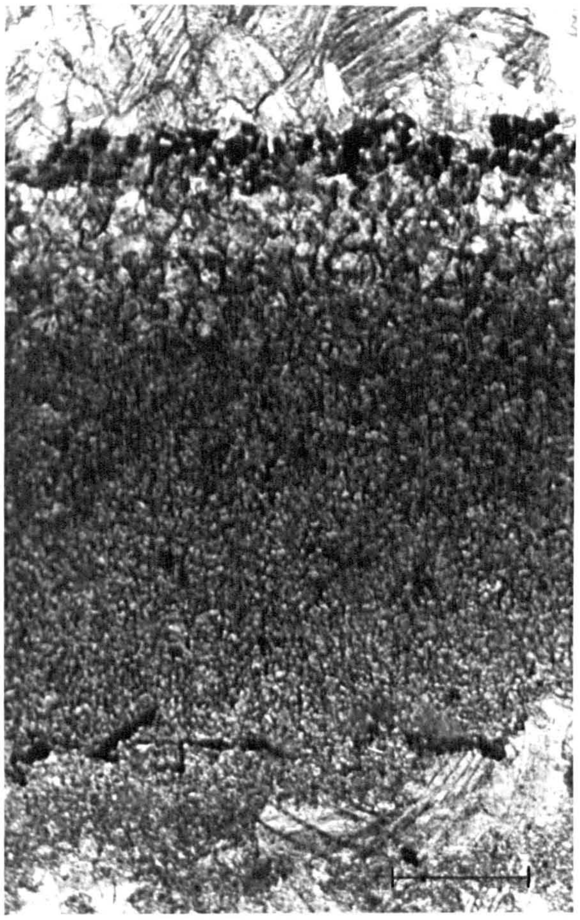
Fig. 3.56 Secondary cracks formed above larger sheet crack, filled by radiaxial calcite and overgrowths from cricoconarids. Givetian/ Frasnian. Bonzel, Attendorn region. Thin section S 23067. Scale bar = 1 mm.

20μ. Although neomorphism has occurred to give a coarser fabric, grading of the sediment is still preserved. Some cracks have several graded bands of grey microsparite, which contain cricoconarids with syntaxial overgrowths. The red internal sediment partly filling the sheet cracks from Bonzel is reversely graded at the top- a feature noted by Schwarzacher (1961) in Carboniferous '<u>Stromatactis</u>' cavities. The red sediment consists of a tight mosaic of interlocking grains (Fig. 3.57) with a border 0.5 to 1μ thick between the grains, where clays and iron minerals are concentrated. A later neomorphism of the internal sediment has partly replaced the fibrous overgrowths. Some of the fibrous crystals radiating outwards from cricoconarids have been replaced by microsparite (Fig. 3.58), and occasionally, an odd patch of fibrous calcite can be seen cut off from its growth nucleus by microsparite. The twinning in the overgrowth fragment however corresponds to that of the cricoconarid from which it originally extended (Fig. 3.59).
<u>Interpretations, origins and lithification</u>. The cricoconarid-rich sediments were cemented before the formation of the sheet cracks. The cracks would not remain open over such distances as 2 metres, unless the host sediment was lithified. In modern lithified pelagic oozes, lithification often occurs in bands, with softer uncemented material between (Milliman et al, 1969).

Cementation of the cricoconarids probably took place by the formation of overgrowths. The fibrous calcite overgrowths replaced an earlier acicular cement which was probably aragonitic (brachiopods with calcitic skeletons originally and crinoids which had high-magnesium calcitic skeletons are perfectly well preserved without overgrowths of this type). Modern acicular aragonitic cements are well documented and a Recent example that is very similar to the Devonian cemented cricoconarid microcoquinas occurs at the bottom of the Red Sea (Gevirtz and Friedman, 1966; Milliman et al, 1969). Here lithified pteropod oozes are cemented by syntaxial overgrowths of acicular aragonite (pteropods are composed of aragonite) (Fig. 3.60). It is suggested that cementation of the cricoconarids occurred through the formation of similar aragonitic overgrowths which also grew

Fig. 3.57 Internal sediment partly filling sheet cracks. Sediment coarsens upwards. Givetian/Frasnian. Bonzel, Attendorn region. Thin section S 23067. Scale bar = 200μ.

Fig. 3.58 Internal sediment has recrystallized to a microsparite and replaced the lower part of the fibrous overgrowths. Lower Frasnian. Bicken, Dill Syncline. Thin section S 23065. Scale bar = 200μ.



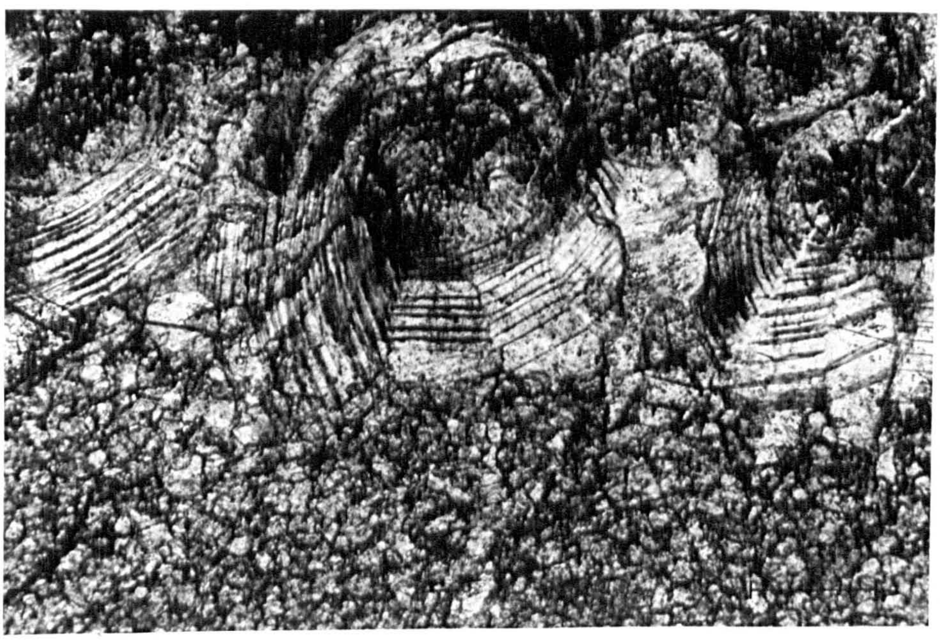
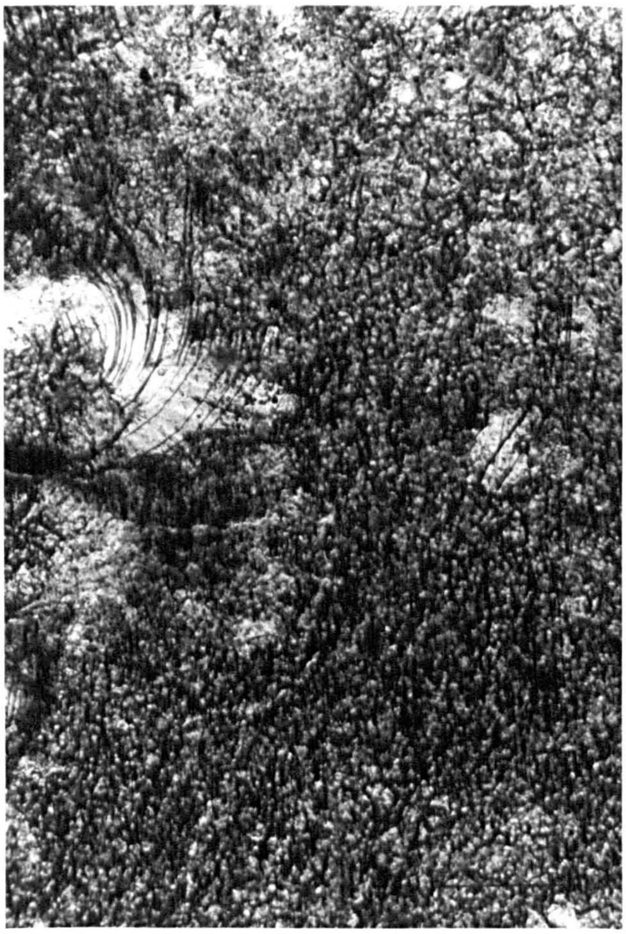


Fig. 3.59 Recrystallization of the internal sediment to microsparite has partly removed the fibrous overgrowths around cricoconarids. Remnants of the overgrowth are separated from the cricoconarid, but can be recognised by the twin plane directions. Lower Frasnian. Bicken, Dill Syncline. Thin section S 23065a. Scale bar = 200μ.

Fig. 3.60 Recent acicular aragonite which has formed syntaxially around pteropods (From Gevirtz and Friedman, 1966).



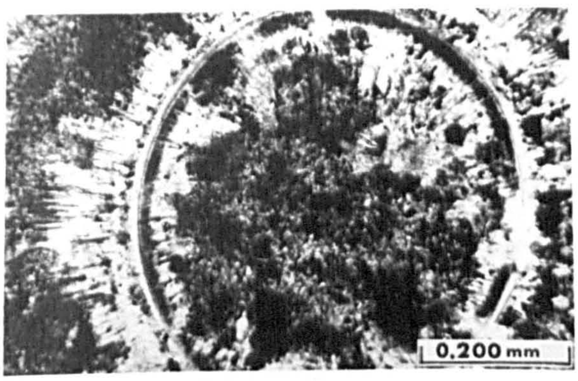


FIG. 5.—Cross section of Pteropod in hard layer showing syntaxial and drusy fibrous aragonite.

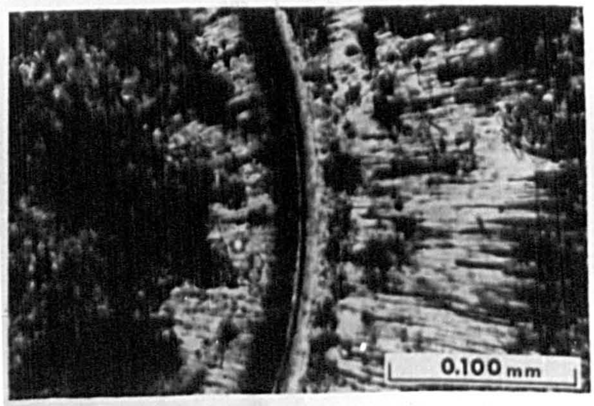


FIG. 6.—Enlargement of fig. 5.

into the sheet cracks. Later, the aragonite fibres were replaced by the growth of fibrous calcite, seeded from the cricoconarid wall. These replacements grew outwards and the ghosts of the growth stages reveal the shape these replacing crystals took.

The presence of continuous twin planes from one crystal to the next, at first a very difficult thing to explain, can be understood knowing that the fibrous calcite crystals are replacing acicular aragonite. The original aragonite overgrowths would have been very fine crystals, (modern ones are 100μ x 10μ) and adjacent fibres will differ little in optic orientation. Hence the aragonite fibres radiating round the shell can be regarded as one large overgrowth. As fibrous replacement crystals grow, they would include a large number of aragonite fibres. Since the aragonite fibres do not differ much in orientation, they can be accomodated by the developing replacement crystal, but the net result, will be a strained fibrous crystal. This strain is released by the formation of twin planes and since effectively, there is one large overgrowth crystal, release of strain will cause concentric twin planes to form. The continuation of twin planes from one crystal to the next, could only occur if the crystals have the same a-orientation. They have the same c-orientation, this can be seen in thin section. For the overgrowth crystals to have exactly the same orientation, can only mean that the shell structure of the cricoconarid has determined how the overgrowth crystals were seeded.

The paragenesis of the sediments with the sheet cracks from Bicken, is as follows;

1. Deposition of cricoconarid microcoquinas, which are cemented by syntaxial acicular aragonite overgrowths around the cricoconarids.

2. Formation of sheet cracks, which are filled by internal sediment, locally containing cricoconarids. Growth of acicular aragonite overgrowths into the cavity.

3. Replacement of aragonite fibres by fibrous calcite seeded on the cricoconarid wall. Growth outwards enveloping aragonite producing strained crystals.

4. Release of strain producing curved twin planes, frequently

continuous forming bodies resembling ooliths.
5. Neomorphism of internal sediment and matrix between cricoconarid shells producing a coarser mosaic. Replacement of some fibrous calcite overgrowths by microsparite.
6. Production of thin diagenetic veins.
7. Intense pressure solution, coincident with the development of cleavage, producing numerous solution stringers, injecting clay, removing parts of the cricoconarids and/or their overgrowths, and displacing thin veins.

Two mechanisms operate today forming cavities below lithified sediments, a) lifting and erosion and b) burrowing. Cavities can form from a slight lifting up of a lithified layer through force of crystallization. Shinn (1969) described low amplitude folds from the Persian Gulf which may have formed in this way. Lindström (1963) figured similar fold structures from the Ordovician Orthoceras limestones of Scandinavia and invoked movement or gliding of cemented bands over the marls. After lifting, scouring out of soft sediment may occur. Soft sediment below a lithified band may be excavated by crustaceans or fish and produce small cavities (Shinn, 1969).

The lateral extent and spacing of the Bicken sheet cracks is the main problem and the mechanisms outlined above would probably not give rise to cracks over 2 m long. The origin of cracks in cricoconarid microcoquinas is not known but early cementation of the sediment is probably one of the main factors since this type of crack does not occur in other lithologies.

3.2.4 Neptunian dykes (Figs. 3.61 to 3.69).

Neptunian dykes occur locally in the Devonian pelagic sediments at Adorf, Bicken, Calvarienberg, Eulenspiegel and Gaudernbach (Rheinisches Schiefergebirge), Langestal (Harz) and Combe D'Izarne (Montagne Noire). The dyke from Langestal has been interpreted previously as due to karstic weathering during emergence (Stoppel, 1968). Dykes formed by tectonic fracturing occur in Jurassic pelagic sediments of the Alps (Wendt, 1969).

Two types of dyke can be distinguished on scale and age,

a) dykes where the fill is of a similar lithology and age to the host sediment and b) dykes penetrating down into appreciably older sediments where the host lithology and fills are quite different.

Dykes of the first type commonly occur in the same sediments as sheet cracks, Flaser limestones with very little clay. The dykes are mostly vertical or subvertical and range in size from 1 to 30 cm in width (Figs. 3.61 and 3.62). They penetrate down to at least one metre. One dyke at Bicken cuts across the bedding and then follows the bedding, possibly through coinciding with a sheet crack. Dykes at Eulenspiegel are commonly shallow structures, 20 cm wide and cutting down 10 to 15 cm (Fig. 3.63).

All the dykes are filled by numerous graded units of microsparite, each 5 mm or less in thickness (Fig. 3.64). Irregularities in the internal bedding occur in subvertical dykes near the roof. Dykes in Givetian/Frasnian limestones commonly have bands of internal sediment composed soley of cricoconarids. In some cases, the internal sediment is very rich in pyrite, as minute grains 1-2μ across or aggregates (Fig. 3.65). Limestone intraclasts are commonly present in the dykes, and are mostly 1 cm across. Evidence of lithification of the host sediment before the dyke sediment was introduced (apart from intraclasts) is shown in one instance where a shell filled with the host sediment projects well in the dyke sediment (Fig. 3.66). Much of the shell had dissolved before the dyke sediment came in but the shell-fill retained its shape. Thin dykes, 1 cm wide, commonly cut limestone beds and are filled by numerous thin laminae of microsparite.

Thin fibrous calcite bands may be present within the dyke sediments. The cement at the top of the dyke is mostly a drusy equant calcite, but in places, a fibrous mosaic is developed. The latter was probably the original cavity fill, much of it having been replaced by sparry calcite.

Microstylolites commonly 'underline' or slightly displace the laminae of the internal sediment. The sides of dykes have sometimes been affected by pressure solution or small tectonic movements.

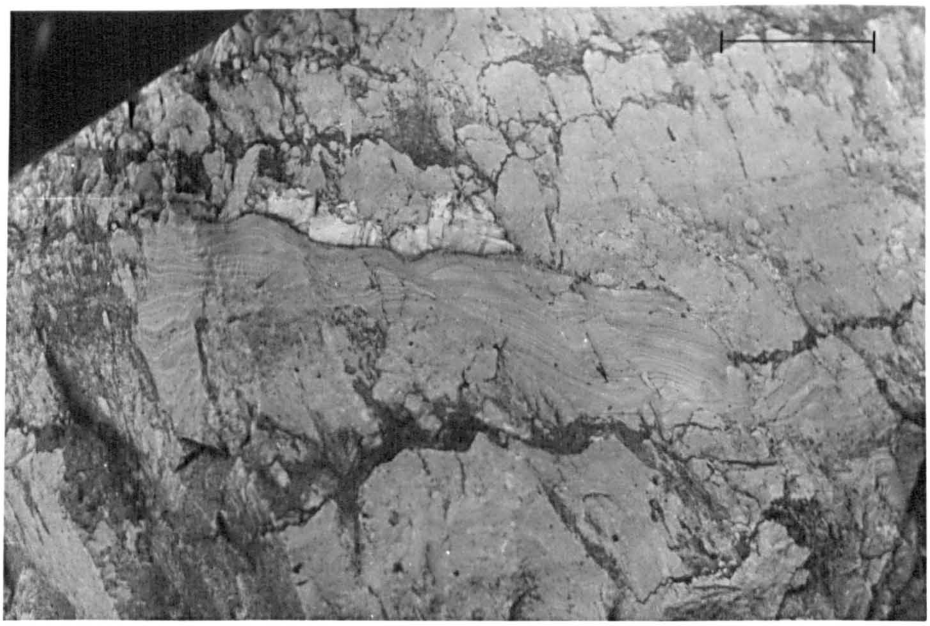
Larger more complicated dykes of Upper Devonian pelagic sediments

Fig. 3.61 Vertical neptunian dyke filled with numerous graded bands of microsparite. Thin bands of fibrous calcite occur within the sediment. Dyke on right filled with cricoconarids, limestone and shale clasts. Lower Frasnian. Bicken, Dill Syncline. Scale bar = 5 cm.

Fig. 3.62 Vertical neptunian dyke penetrating cricoconarid rich sediments. Lower Frasnian. Bicken, Dill Syncline. Scale bar = 5 cm.

Fig. 3.63 Shallow neptunian dyke in grey Flaser limestone. Calcite has filled final void. Frasnian. Eulenspiegel, near Warstein, Sauerland. Scale bar = 10 cm.

Fig. 3.64 Horizontal neptunian dyke cutting cricoconarid-rich limestone. Some internal sediment bands composed entirely of cricoconarids. Lower Frasnian. Bicken, Dill Syncline. Polished surface S 23064. Scale bar as shown in cm.



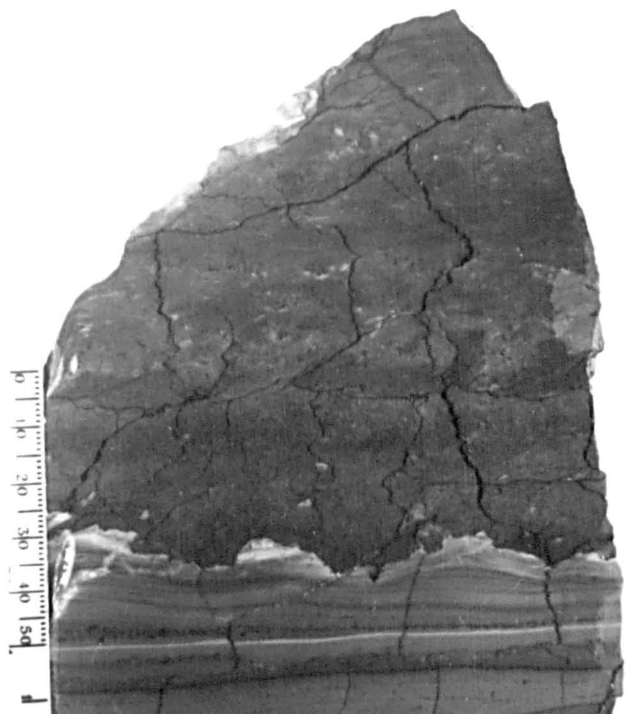
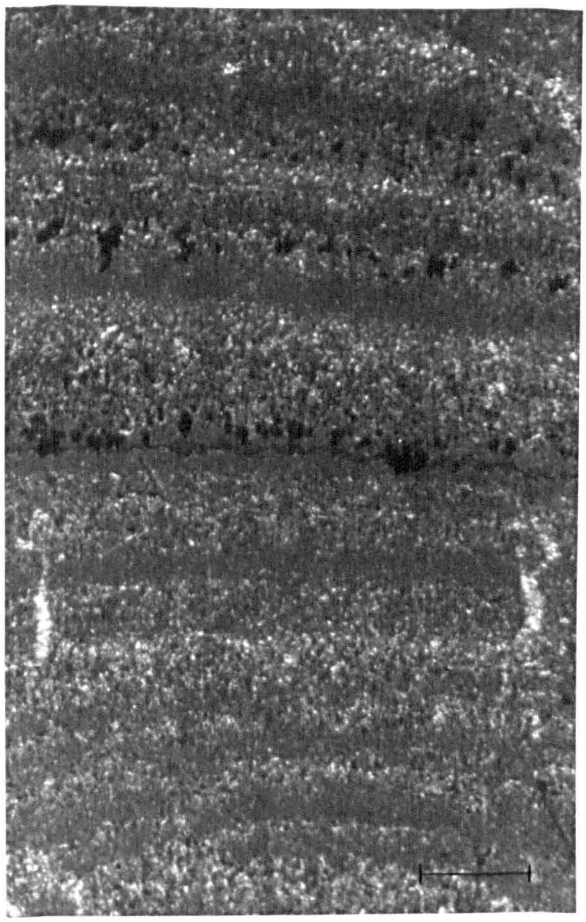
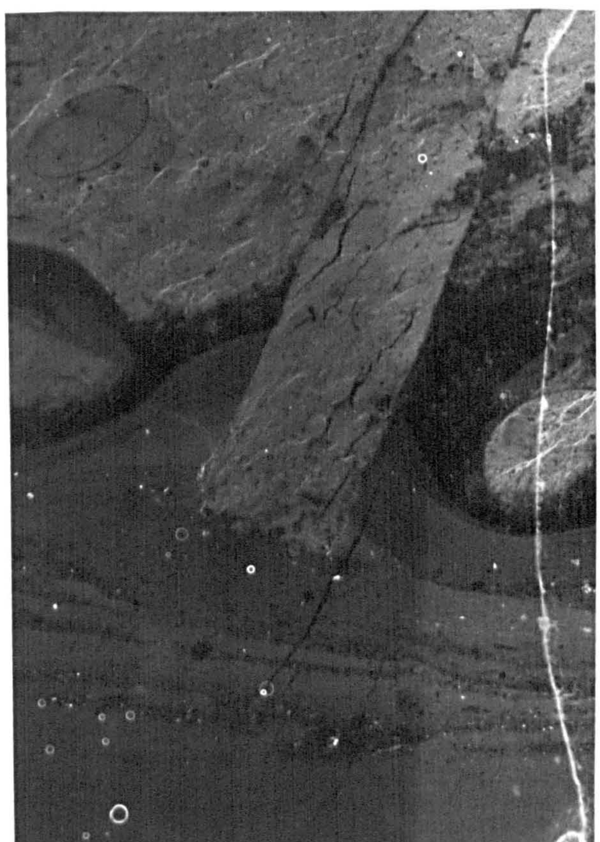

0
10
20
30
40
50

Fig. 3.65 Bands of micrite and microsparite in a neptunian dyke. Pyrite is present as minute grains 2μ in diameter, and as aggregates. Frasnian. Eulenspiegel, near Warstein, Sauerland. Thin section S 23051 Scale bar = 1 mm.

Fig. 3.66 Projection of host sediment which had filled a skeletal fragment into neptunian dyke. The distinct shape and undisturbed nature of the projection suggests lithification of the host sediment before internal sedimentation in the dyke. Lower Frasnian. Bicken, Dill Syncline. Negative print of thin section S 23064. Scale bar = 5 mm.





in fissures in the lower Frasnian/Givetian 'reef' limestones, occur at Gaudernbach and are recorded from other localities that were not examined (Krebs, 1968a). These dykes extend down at least 30 m into the 'reef' limestone and similar dykes eleswhere give a variety of Upper Devonian ages, based on conodonts (Krebs, 1968a). Large clasts of reef limestone are included in some of the dykes, and also pieces of fibrous calcite cement which filled voids in the 'reef' limestone (Figs. 3.67 and 3.68). These are normally embedded in a coarse red calcisiltite. Other dykes are filled by numerous graded bands of calcisiltite and micrite of various shades of yellow, red and brown (Fig. 3.69). At Calvarienberg, black Clymenia shales fill a large fissure in Platyclymenia Flaser limestone.

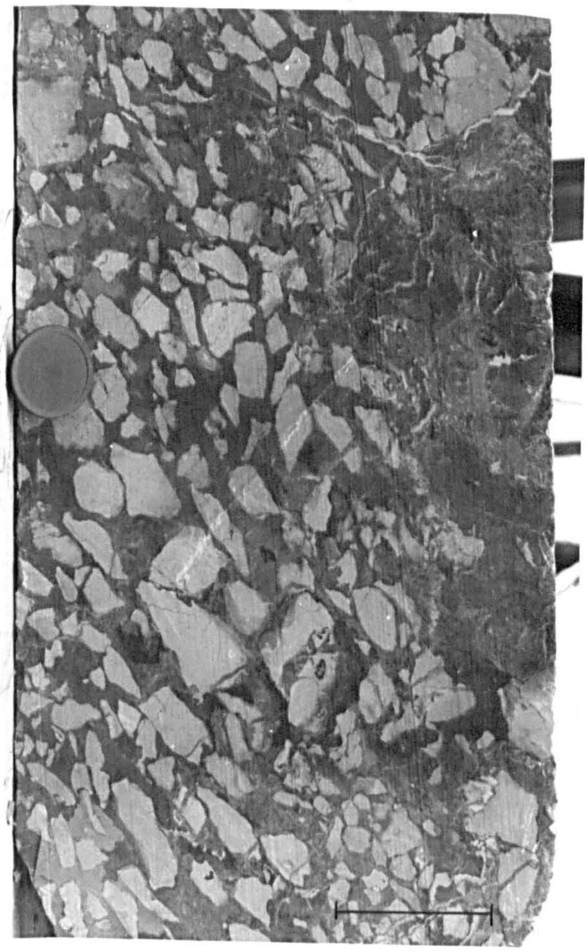
Origin: For vertical or subvertical sedimentary dykes there are four possible mechanisms for their formation (Wendt, 1971), a) Karst weathering, b) submarine erosion, c) slight movements of the lithified sediment, and d) tectonic fracturing contemporaneous with sedimentation.

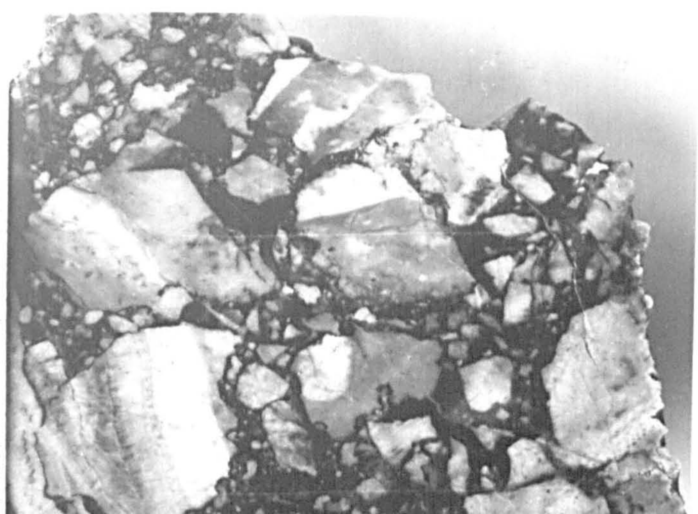
There is no evidence for emergence although Stoppel (1968) envisaged such a mechanism for a small sedimentary dyke in similar sediments at Langestal (Harz). Submarine erosion is more likely to be a contributary factor, modifying the dyke form rather than an initiating process. For example, erosion has probably occurred in the case of the shallow dykes at Eulenspiegel (Fig. 3.63). Settling of lithified carbonate, mass transport of sediment downslope, or shear failure on a slope could give rise to sedimentary fissures and this type of process probably occurred with the smaller dykes in the Schwellen limestones.

Dykes frequently are formed by small tectonic movements. Wendt (1971) has shown this to be the main factor in the formation of the neptunian dykes which pervade much of the Jurassic of Sicily and the Alps. Pelagic sediments, locally with specially adapted faunas, penetrate down to several hundred metres, into the Triassic reef limestone. Also in the Northern Calcareous Alps, Fischer (1965) describing similar structures in the Loferites considered tectonic fracturing a likely cause. Large scale mudcracking as well as karst weathering are also suggested as possible mechanisms. The absence

Fig. 3.67 A fissure filling in a Givetian/Frasnian 'reef' consisting of angular clasts of 'reef' limestone embedded in red calcisiltite. Gaudernbach, Lahn Syncline. Scale bar = 10 cm.

Fig. 3.68 Angular fragments of reef material and fibrous calcite (formed in voids in the reef) embedded in a red calcisiltite and occurring as a fissure filling in a Givetian/Frasnian 'reef'. Gaudernbach, Lahn Syncline. Polished surface S 23047. Scale as shown in cm.





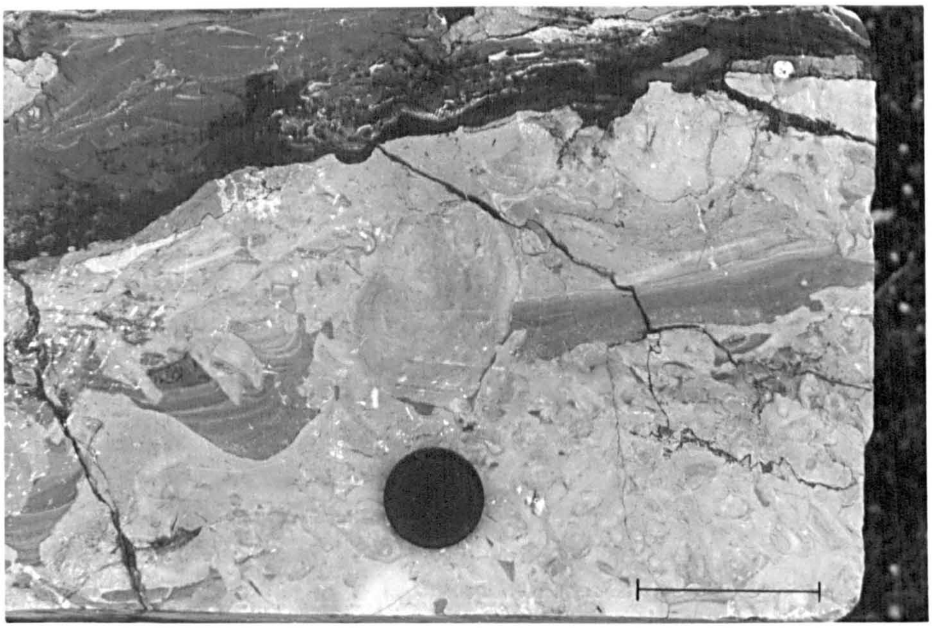


Fig. 3.69 Laminated red and brown calcisiltite of Upper Devonian age occurring in channels and dykes in Givetian/Frasnian 'reef' limestone. Gaudernbach, Lahn Syncline. Scale bar = 10 cm.

of any evidence for emergence in the Devonian sediments, precludes the latter two. Unfortunately, not enough dykes were encountered in Germany to make measurements of their orientation for comparisons with tectonic directions. However, tectonic fracturing seems likely for the larger dykes at Gaudernbach and Calvarienberg and could apply to some of the smaller ones too. In some cases, lithification of the host sediment before dyke formation has clearly taken place and so useful information is gained if the dykes fills are not much younger than the host sediment.

Section 3.3. Lithification

The Devonian pelagic limestones are fine grained and have suffered extensive recrystallization. Within the greater part of the sediments therefore cements cannot be distinguished. However, evidence from sedimentary structures described above (sheet cracks and neptunian dykes) and the presence of limestone clasts, shows that at least where these occur, lithification occurred during early diagenesis when the sediments were still in a completely submarine environment. The presence of hardgrounds (p.73) shows that in some instances lithified sediments formed the sea floor. Generally, sedimentary structures are not very common but lack of compaction before lithification is a feature of the limestones and also suggests early lithification.

In modern carbonate muds, most compaction takes place in the first 50 cm below the sediment/water interface, when a great loss of water occurs, reducing an original soupy sediment to a mud with a porosity of 50 to 70% (Ginsburg, 1957). The porosity is gradually reduced by compaction, and in Recent deep water sediments the transition from dominant compaction to solution welding of grains and cementation is between 250 and 500 m (Scholle, 1971). There is very little evidence for compaction due to overburden pressure in the Schwellen limestones and the sediments must have been lithified before much burial. Only one example was encountered where a geopetal cavity had collapsed (Fig. 3.70) and this was in a nodular limestone (griotte) rich in goniatites which are filled with carbonate and have more argillaceous sediment between them. The nature of the fracture suggests that it occurred after lithification of the sediment and before the filling of the geopetal cavity by calcite. One fractured shell was found (Fig. 3.71) which may have been broken through compaction. Otherwise all other fossils are full bodied and breakages have not occurred. Trace fossils are circular in section and show no evidence of deformation. The pelagic carbonate must have been sufficicently cemented before the overburden pressure was high enough to cause appreciable compaction. Observations of this kind on ancient shallow water carbonate muds have been made by many authors e.g. Pray (1960), Beales (1965), Bathurst (1970). As Bathurst (1970) concluded, a vast quantity of carbonate is needed for the cementation of these carbonate muds - more than half the volume of the limestone. The problems of the origin of this material, how it was transported and precipitated are still unsolved. Compaction after cementation has occurred in many Schwellen limestones with the formation of microstylolites and later diagenetic pressure solution planes.

The features of the Schwellen limestones then suggest early diagenetic lithification in the submarine environment. A few years ago it was considered that carbonate sediments could only be lithified in subaerial environments (Friedman, 1964) but there have recently been many records of submarine lithification from various environments at various depths (Fischer and Garrison, 1967). One of the main factors in early lithification of Recent submarine sediments is slow sedimentation, and it is probably the slow sedimentation of the Schwellen sediments which led to their early cementation. Their early diagenetic history is in marked contrast to carbonate turbidites which accumulated at a much faster rate and show evidence of early compaction before cementation (Scholle, 1971).

Fig. 3.70 Geopetal cavity formed by a goniatite which was fractured before being filled by calcite, but after lithification of the host sediment. Lower Famennian. Coumiac, Montagne Noire. Peel S 22966. Scale bar = 1 mm.

Fig. 3.71 Shell with a ferromanganese coating that has been fractured. Fragments of the ferromanganese coating occur in the drusy calcite and inside the shell indicating that a void stage existed. The shell is now a sparite pseudomorph. Lower Famennian. Combe D'Izarne, Montagne Noire. Thin section S 229546. Scale bar = 1 mm.

Section 3.4 Neomorphism (late diagenesis)

Folk (1965) defined recrystallization as applied to carbonate rocks as replacement of one calcite type by calcite of another grain size, morphology or orientation. For the change aragonite to calcite the term inversion is used. Folk introduced the word neomorphism when it is not possible to decide whether recrystallization or inversion has occurred. Folk also distinguishes two types of neomorphism, aggrading and degrading, and divides aggrading into porphyroid and coalescive depending on how the increase in grain size takes place. Porphyroid aggrading neomorphism, proceeding from a number of isolated nuclei is considered by Folk (1965) to be the initial process which is rarely preserved. Coalescive neomorphism is general coarsening of the mosaic by the growth of some grains, in preference to others, and is considered to take over from porphyroid neomorphism once a certain grain size is reached. In the Schwellen limestones, most types of recrystallization described by Folk can be recognised.

3.4.1 Aggrading neomorphism (Figs. 3.72 to 3.83).

The matrix of many of the Devonian pelagic limestones has a mottled appearance where micritic areas are surrounded by areas of slightly coarser microsparite. These features may represent original inhomogeneities in the sediment, through burrowing or the presence of compound grains, but irregular neomorphism could also give rise to this fabric. Coarser grains, 20 to 50μ in size, are commonly scattered throughout the sediment and probably represent fragments of skeletal debris which have undergone neomorphism. Relicts of shells which were nearly obliterated through aggrading neomorphism of the sediment can frequently be discerned. A number of limestones (e.g. those from Bicken) consist entirely of microsparite or pseudospar with numerous skeletal fragments, and the original sediment must have been a calcisiltite rather than a calcilutite that has undergone extensive neomorphism.

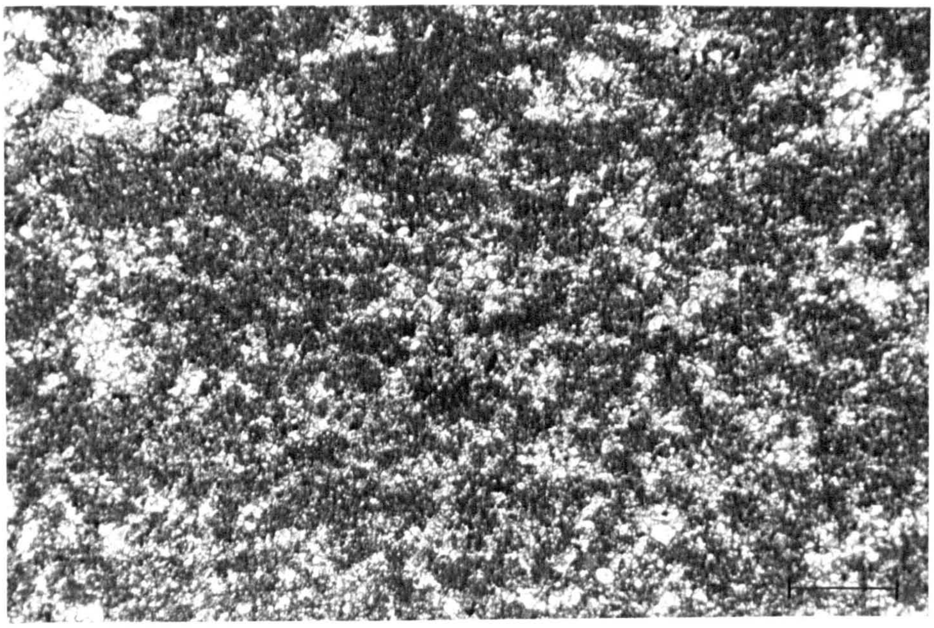
A type of fabric which resembles the porphyroid aggrading neomorphism of Folk (1965), the grain growth of Bathurst (1958, 1959) and Orme and Brown (1963) is present in the Frasnian limestones ~~from the~~ (Ense Schwellen limestones) from Steinbruch Schmidt in the Kellerwald. These are limestones with very little clay material. Small patches, a few centimetres across, consisting mostly of pseudosparite grains 20μ to 30μ in size are developed in some examples (Fig. 3.72). These patches grade into the background carbonate which here consists mostly of micrite and microsparite. The fabric resembles the structure Grumeleuse of Cayeux (1935) and the diagenetic fabrics described by Beales for pelleted limestones (1965). These specimens show that recrystallization began from a number of isolated centres and could eventually lead to a complete coarsening of the original mosaic. This type of neomorphic fabric may develop from original variations in the sediment, but since these patches of pseudospar grade into 'normal' micrite/microsparite, this fabric is probably just a diagenetic feature, and may be an example of Folk's porphyroid neomorphism.

Samples from Grevenbrück where pelagic carbonates deposited on the flanks of the Attendorn reef are exposed, show an advanced stage of porphyroid neomorphism where the original fine grained sediment is left only as clots in a pseudosparitic matrix consisting of crystals up to 70μ across (Fig. 3.73). The remains of skeletal material can be discerned in the centres of some pseudosparitic crystals (p.161).

Bedding-parallel bands with aggrading neomorphism in some of the limestones from Steinbruch Schmidt (Fig. 3.74) contain some patches of very coarse mosaic (Fig. 3.75) which on first examination would be identified as '<u>Stromatactis</u>'- type cavity structures. The calcite grains in the coarse patches are mostly up to 60μ to 70μ across but some are 2 mm or more in diameter. Cricoconarids and shell fragments occur within this horizon (Fig. 3.76) showing that it is a neomorphic fabric, and not a complex cavity system filled by drusy calcite. Grading of sparite grains may occur into the coarser parts but the shape of the grains is normally irregular. Enfacial junctions (Bathurst, 1964) are absent, indicating

Fig. 3.72 Irregular patches of microsparite and pseudosparite formed by neomorphism of the host sediment. Frasnian. Steinbruch Schmidt (Ense Schwelle) Kellerwald. Peel S 22990. Scale bar = 1 mm.

Fig. 3.73 Micrite clots in pseudosparitic limestone. Givetian/Frasnian. Grevenbrück, Attendorn region. Peel S 22998. Scale bar = 200μ.



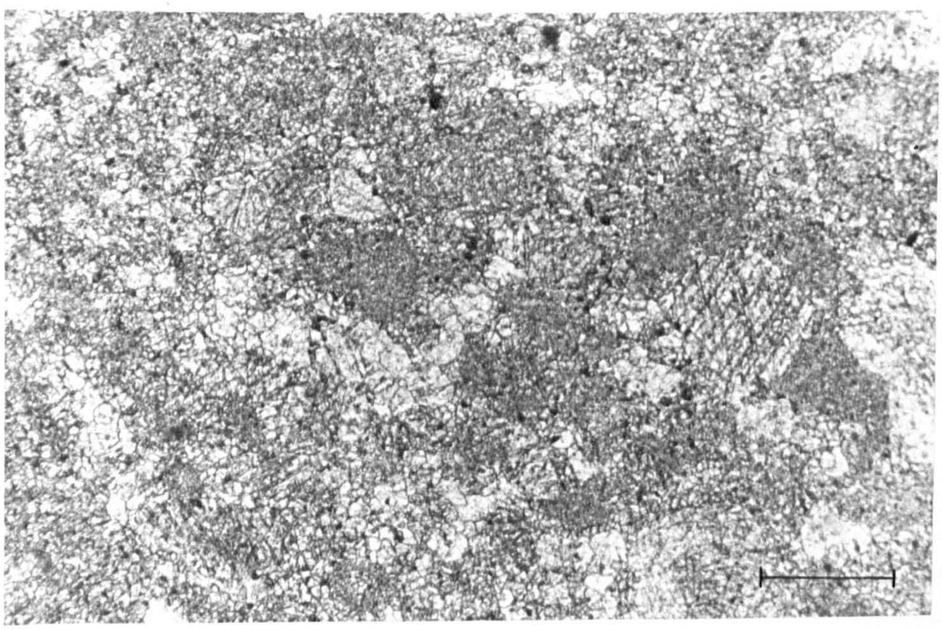
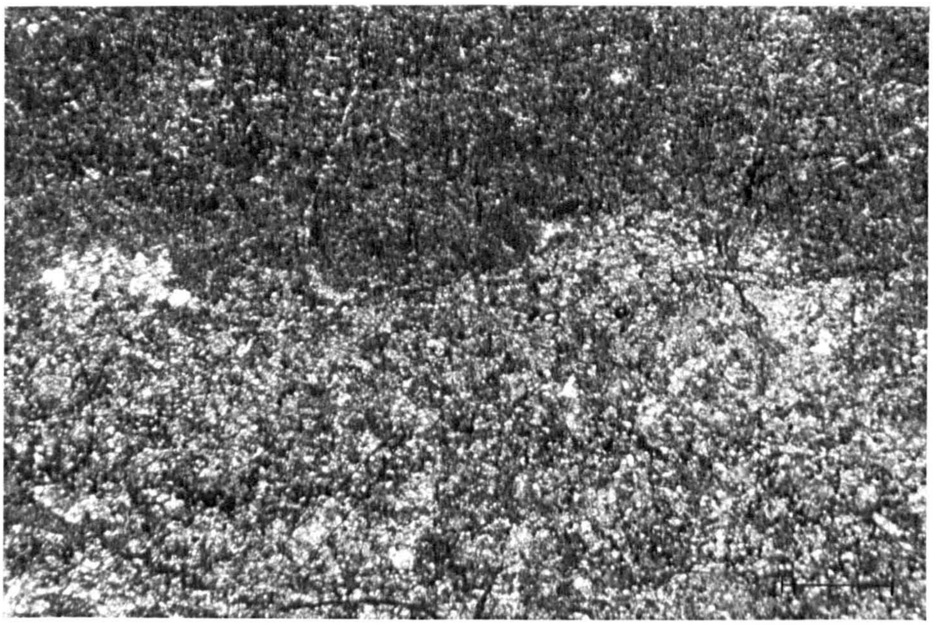


Fig. 3.74 Patches of pseudosparite and coarse microsparite forming a band parallel to the bedding. Skeletal material occurs within the band of coarser mosaic. Frasnian. Steinbruch Schmidt (Ense Schwelle) Kellerwald. Peel S 22991. Scale bar = 1 mm.

Fig. 3.75 Area of coarse pseudosparite grading into the background micrite/microsparite. Frasnian. Steinbruch Schmidt (Ense Schwelle) Kellerwald. Peel S 22991. Scale bar = 1 mm.



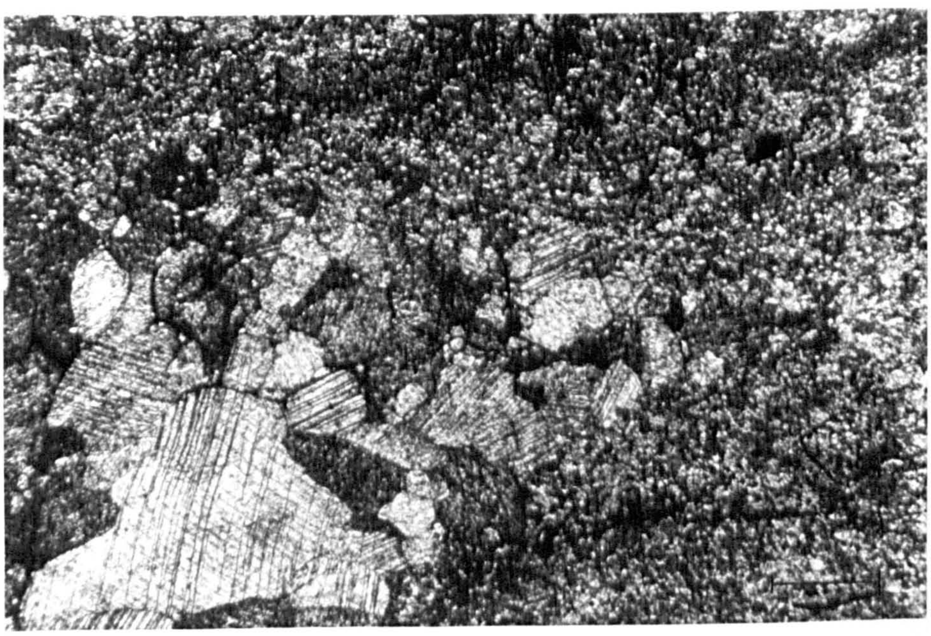
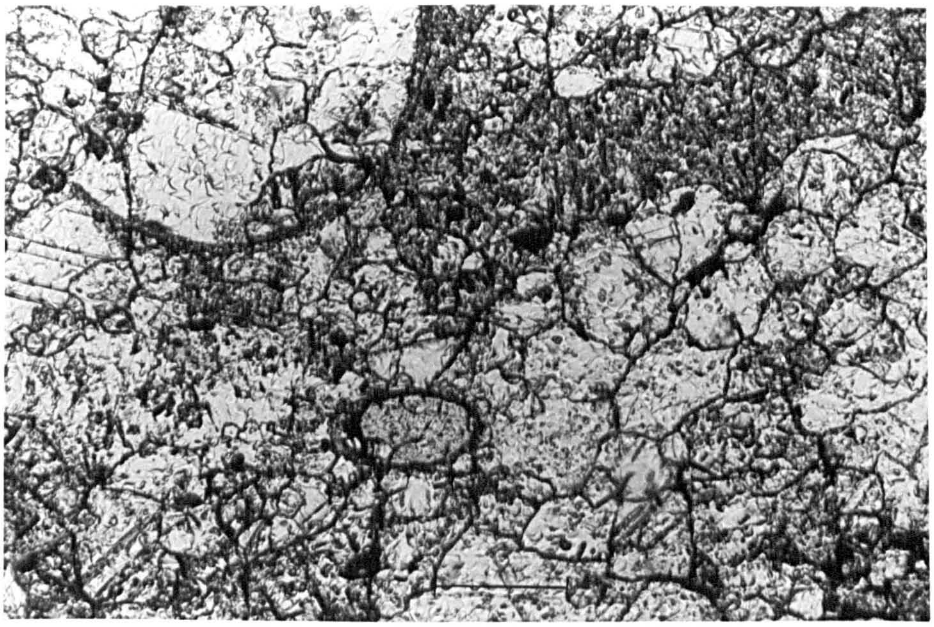
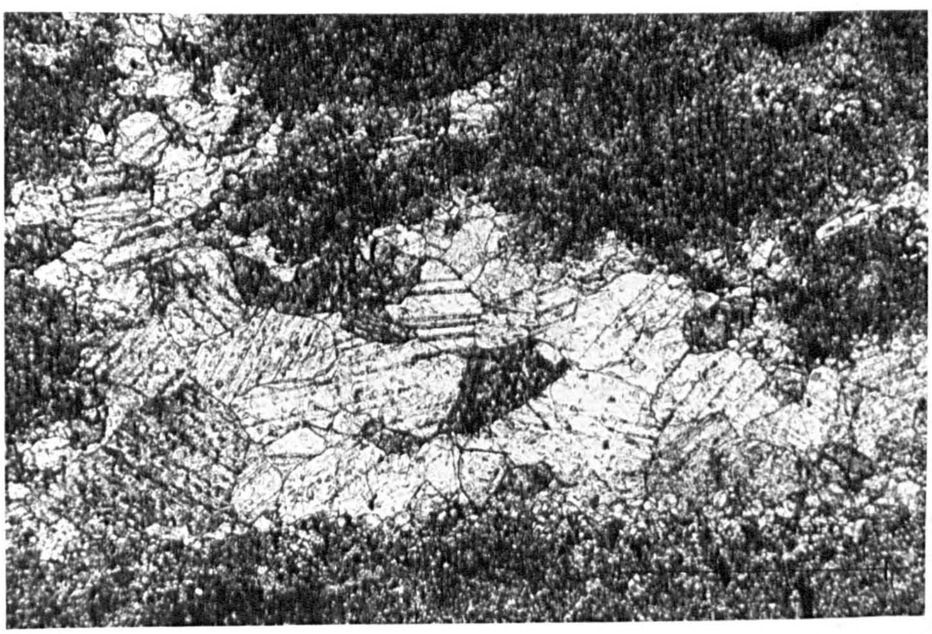


Fig. 3.76 Coarse microsparite and pseudosparite with skeletal fragments. Triple junctions are present but enfacial junctions rare. Frasnian. Steinbruch Schmidt (Ense Schwelle) Kellerwald. Peel S 22991. Scale bar = 200μ.

Fig. 3.77 Area of coarse pseudosparite with flat base. This was probably a cavity structure originally and later modified by neomorphism. Frasnian. Steinbruch Schmidt, (Ense Schwelle) Kellerwald. Peel S 22991. Scale bar = 1 mm.





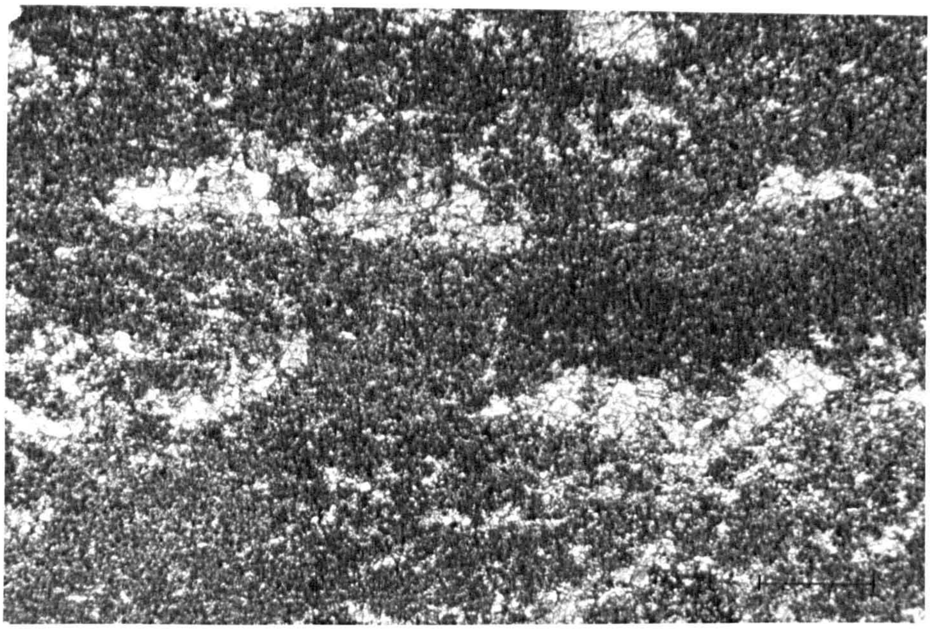
the presence of pseudosparite. These neomorphic fabrics tend to occur in the central parts of limestone bands.

There are patches of coarse calcite which have flat bottoms and a line of small equant crystals at the base (Fig. 3.77). There is also some suggestion of an internal sediment. These were probably cavities originally, although they grade into the recrystallized areas. Gradations can be found between definite cavities, and obvious recrystallization patches (Fig. 3.78). In the latter, coarse patches with micrite 'islands' occur but with quite flat bases. It is difficult to ascertain the importance of bioturbation in the neomorphism here. Some of the '<u>Stromatactis</u>' structures could have been caused by burrowing and then later modified by neomorphism. Structures which are clearly burrows usually have a coarser fabric than the host microsparite and fine downwards (Fig. 3.79). The borders of the burrows are commonly diffuse.

At Acketal (Harz), limestones containing graded bands rich in terrigenous material, may have patches of coarser pseudosparite with grains up to 50μ (Fig. 3. 80) in the background sediment between the silty bands. These are simply irregular coarse patches gradational into the surrounding micrite. With the electron microscope, the mosaic of limestones from Aeketal, Hühnertalskopf, Hutthaler Widerwaage (Harz) and Chudleigh (Devon) is very irregular and discrete grains are badly developed. Coarse grains 10μ to 20μ with diffuse boundaries are interspersed with smaller grains. Cleavage planes of calcite crystals are commonly present suggesting that much of the mosaic has undergone recrystallization. Limestones from Steinbruch Schmidt (Ense Schwelle) show a tight mosaic of grains, mostly 10μ or less in diameter, but larger grains up to 20μ are also present (Fig. 3. 81 ,?organic structure in lower right of photograph). In detail, the grains are interlocked and the contacts are very irregular (Fig. 3. 82). Fischer et al (1967) termed this mosaic amoeboid and suggested solution welding as the process causing the interpenetration of grains. In some replicas very large crystals up to 100μ across showing cleavage planes appear to be replacing the granular mosaic (Fig. 3. 83).

Fig. 3.78 Patches of pseudosparite and coarse microsparite. The upper patch has a flat base and may have been a primary void. Other patches are irregular and grade into the background micrite/microsparite. Frasnian. Steinbruch Schmidt (Ense Schwelle) Kellerwald. Peel S 22990. Scale bar = 1 mm.

Fig. 3.79 Burrows filled by a coarser carbonate sediment and drusy sparite. Frasnian. Aeketal, N.W. Harz. Thin section S 23038. Scale bar = 1 mm.



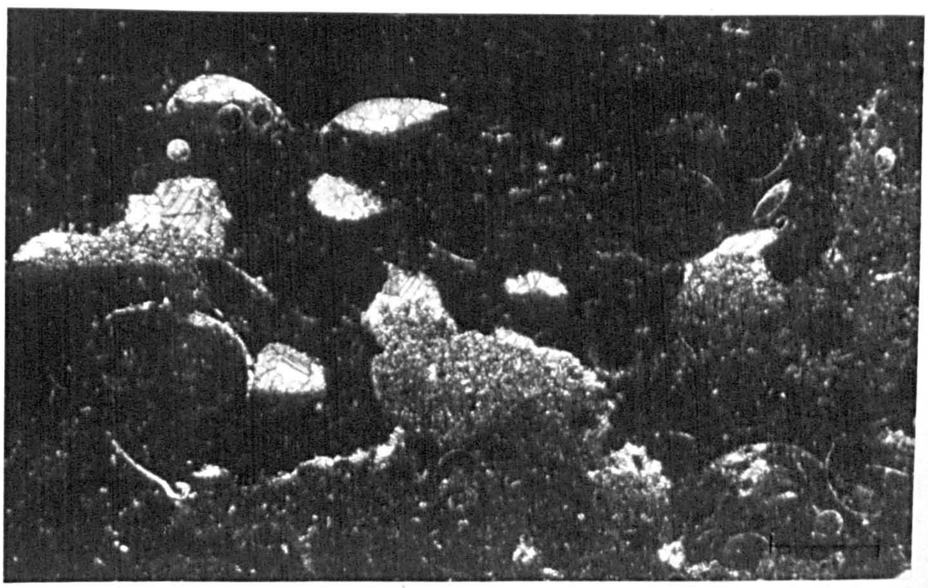


Fig. 3.80 Patches of coarse microsparite and pseudosparite which occur in the carbonate sediment between terrigenous bands. Frasnian. Aeketal, N.W. Harz. Thin section S 23038.
Scale bar = 1 mm.

Fig. 3.81 Tight mosaic of microsparite with some much larger 'grains'. Possible organic structure lower right. Frasnian. Steinbruch Schmidt (Ense Schwelle) Kellerwald. Electron micrograph of sample S 22990. Scale bar = 10μ.

Fig. 3.82 Electron micrograph showing sutured contacts of grains. Frasnian. Steinbruch Schmidt (Ense Schwelle) Kellerwald. Replica from sample S 22990. Scale bar = 5μ.

Fig. 3.83 Granular mosaic of microsparite with large crystal showing cleavage planes that appears to be relacing the surrounding mosaic. Frasnian. Steinbruch Schmidt (Ense Schwelle) Kellerwald. Electron micrograph from sample S 22990. Scale bar = 10μ.

To summarize, recrystallization of the carbonate sediment has been extensive and little remains of the original texture. Recent suggestions that recrystallization of limestones is controlled by the clay content and occurs if the insoluble residue is less than 2% (e.g. Bausch, 1968) are borne out to a certain extent in the Schwellen limestones. In the limestones from Aeketal, recrystallization is confined to the carbonate rich parts between thin terrigenous units. Also in the Ense Schwelle samples, each limestone band is underlain and overlain by a thin shale band (3 mm or less in thickness) and recrystallization is most prominent in the central portion of the bed. Many Schwellen limestones now have a higher clay content through tectonic pressure solution (p.166).

Folk (1965) considered that the neomorphic process leading to the formation of microsparite involves a replacement in the solid state with the aid of interstitial solutions, leading to the formation of coarser grains. Other mechanisms suggested include the formation of overgrowths (rim cementation), the solution of minute supersoluble grains and pressure solution (Bathurst, 1958, 1959; Folk, 1965).

3.4.2 Syntaxial overgrowths (Figs. 3.84 to 3.87).

A common type of neomorphism in the Schwellen limestones, particularly those containing cricoconarids is the formation of syntaxial overgrowths, where calcite had grown around suitable hosts in optical continuity, either by replacing the sediment or as a cement (i.e. filling a void). The overgrowths around cricoconarids are described above in section 3.2.3.2.

Syntaxial overgrowths occasionally occur around crinoid ossicles. This is well documented and can either be a cement, as in crinoidal biosparites (Evamy and Shearman, 1969) or through neomorphism of the surrounding matrix (both types described by Bathurst, 1958, and Orme and Brown, 1963). In the rise limestones, crinoidal overgrowths are of the neomorphic type (Fig. 3.84).

Overgrowth formation has been suggested as a possible mechanism of aggrading neomorphism (Bathurst, 1959; Folk, 1965) and some evidence for extensive overgrowth development is seen in samples

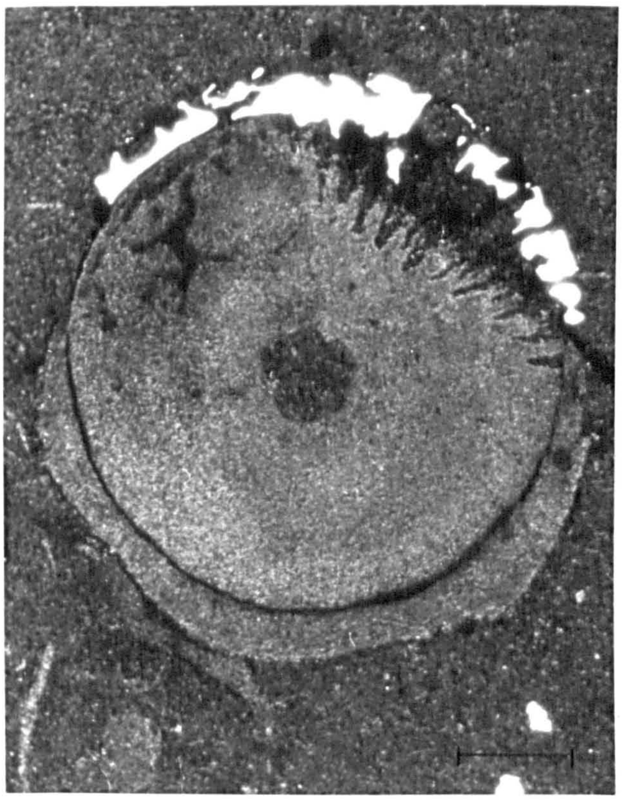


Fig. 3.84 Bored crinoid with syntaxial overgrowth. Frasnian. Adorf am Martenberg, Sauerland. Peel S 23075. Scale bar = 1 mm.

from the Haingrube volcanic Schwelle (Kellerwald) and from Grevenbrück (Rheinisches Schiefergebirge). Coarse limestones at these localities, a few centimetre thick, are very similar to the pseudosparites of Folk (1965), composed of 'loaf' shaped grains (rounded to subequant grains) with a diameter up to 0.4 mm (Fig. 3.85). However, on closer examination the remains of crinoid and shell fragments can be seen, enclosed within the coarse grains (Figs. 3.86 and 3.87).

3.4.3 Degrading neomorphism (Figs. 3.88 to 3.91).

Few examples of degrading neomorphism were encountered although this could be the result of extensive degradation, leaving few cases of partly micritized skeletal fragments. A number of micritized crinoid ossicles were found (e.g. Fig. 3.88), but of more interest is a bivalve from Langenholthausen (Rheinisches Schiefergebirge) which is preserved as a sparite pseudomorph in one part, and then the grain size gradually decreases along the shell until it is indistinguishable from the host sediment (Figs. p 165). This bivalve shows that degrading recrystallization has occurred and in other cases it may have completely removed all evidence of the former presence of skeletal material. This could suggest that the original sediment was generally much coarser, or contained more skeletal debris and that neomorphism has led to an homogenization of the mosaic.

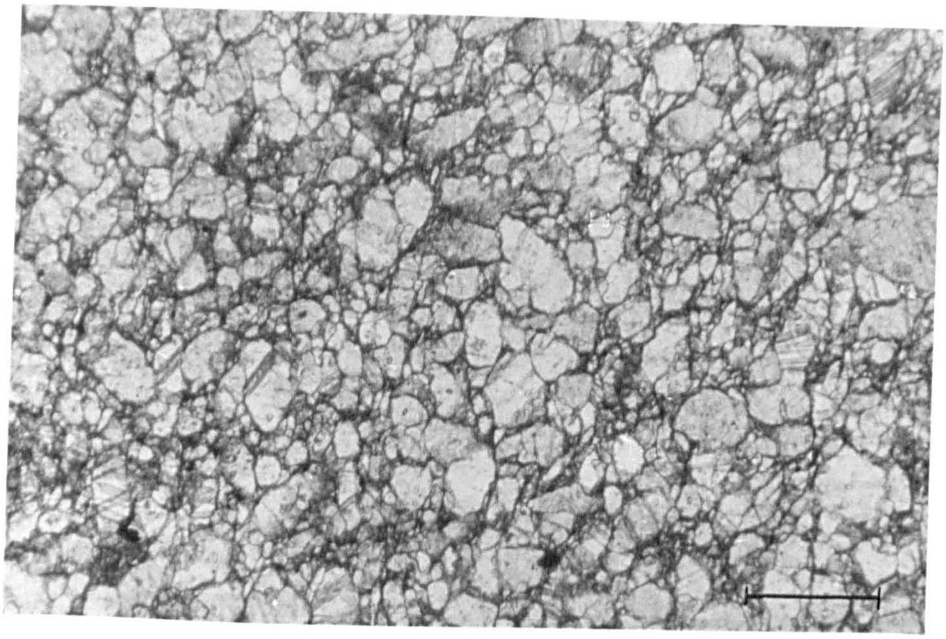
Section 3.5 Pressure solution (Figs. 3.92 and 3.93).

The effects of pressure solution in the formation of stylolites (Flasers) are often very marked in these pelagic limestones. Three periods of stylolite formation can be recognised, early diagenetic, late diagenetic, and tectonic.

Early diagenetic pressure solution has caused microstylolites to form parallel to the bedding. These are not always laterally persistent. They are particularly well seen where the limestone consists of coarse and fine bands where they occur at the base of the coarse band. Thin diagenetic calcite veins displace these

Fig. 3.85 Pseudosparitic limestone consisting of rounded to subequant 'grains'. Upper Famennian. Haingrube, Kellerwald. Peel S 22969. Scale bar = 0.5 mm.

Fig. 3.86 Pseudosparitic limestone (as Fig. 3.85 above) but with syntaxial overgrowth around echinoderm fragment. Upper Famennian. Haingrube, Kellerwald. Pell S 22969. Scale bar = 200μ.



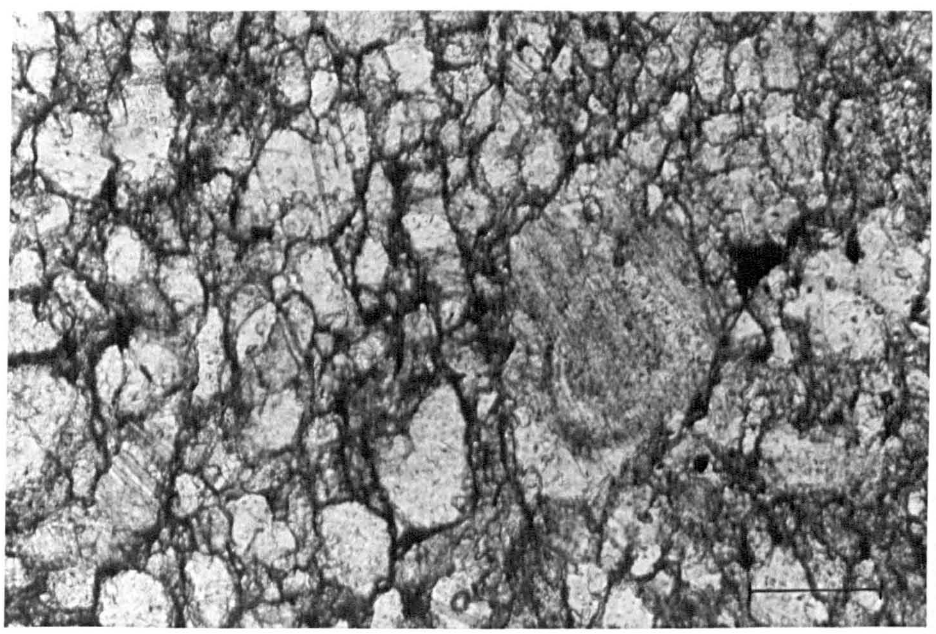
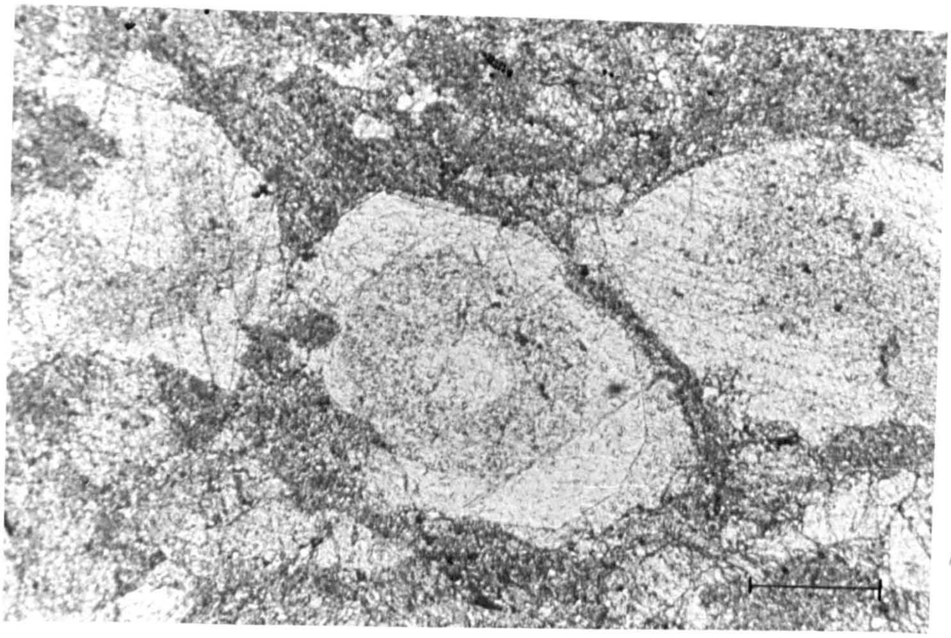


Fig. 3.87 Syntaxial overgrowths around echinoderm fragments. Central example shows a later stage of overgrowth. Givetian/Frasnian. Grevenbruch, Attendorn region. Peel S 22998. Scale bar = 200μ.

Fig. 3.88 Partially micritized crinoid fragment. Conodont present upper left, and foraminiferal/algal nodule lower left. Frasnian. Adorf am Martenberg, Sauerland. Thin section S 22974. Scale bar = 1 mm.



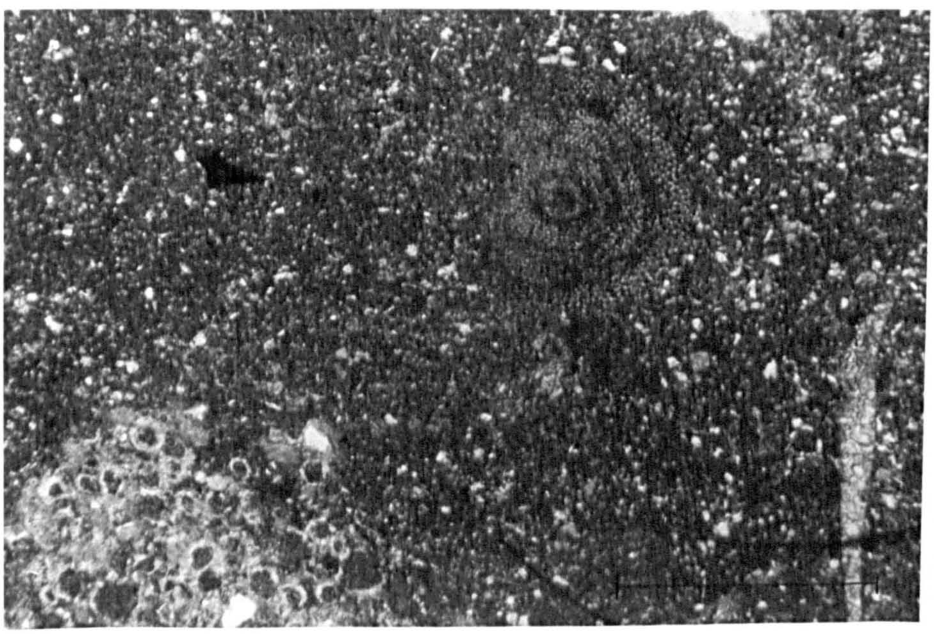
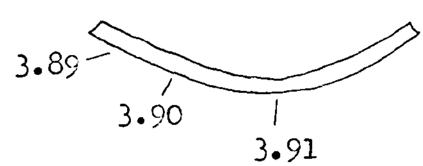


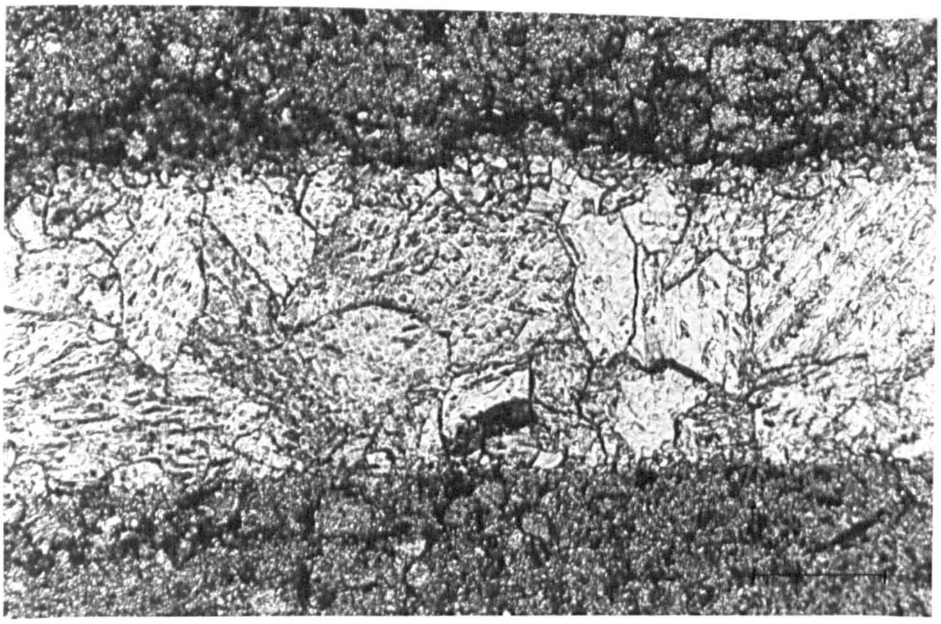
Fig. 3.89, 3.90 and 3.91.

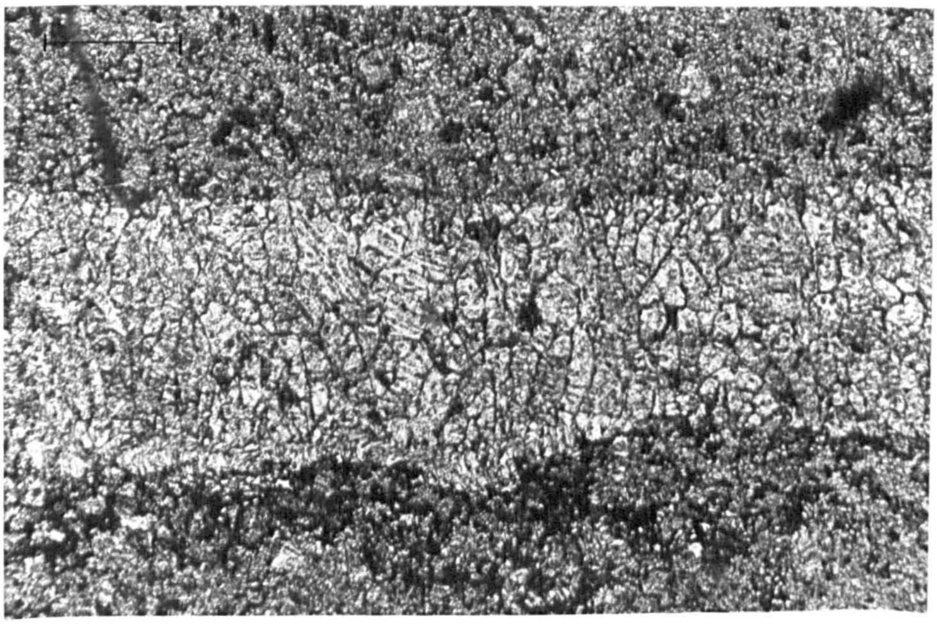
Photomicrographs taken at three points along a bivalve.

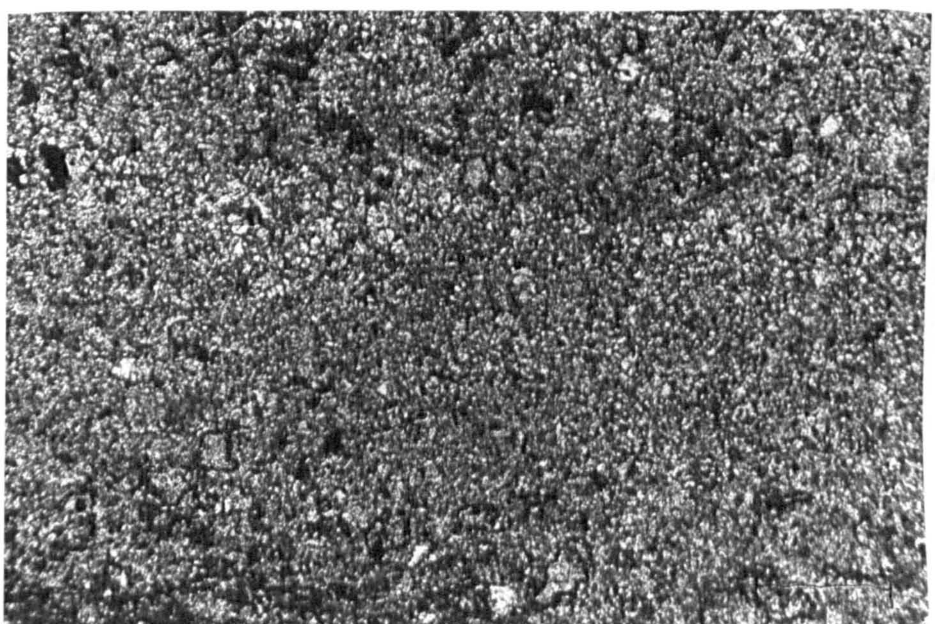


A gradual decrease in grain size of the shell mosaic occurs until at the lowest point (3.91) the shell is indistinguishable from the host sediment. Lower Famennian. Langenholthausen, Arnsberg region. Peel S 22997.

Scale bar = 200μ.







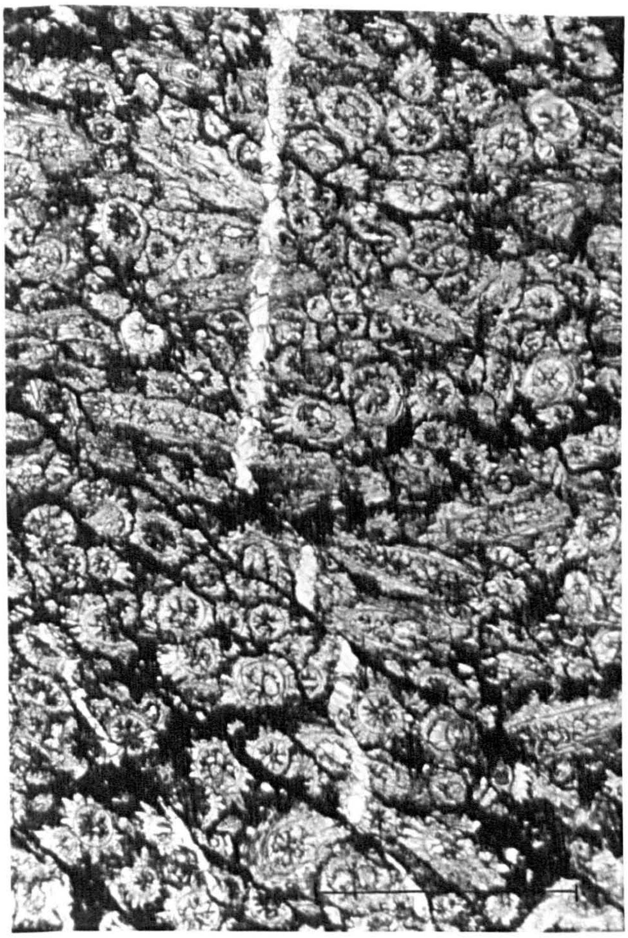
early stylolites. Limonite and clay minerals are frequently developed along these planes and fragments of skeletal material may have been removed a little or displaced slightly.

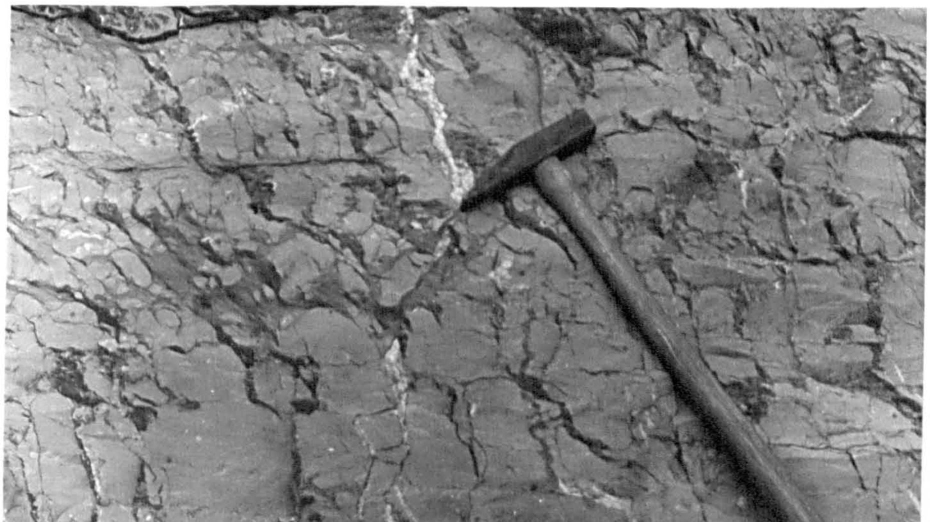
Stronger stylolites are present in many limestones and it is often difficult to distinguish between late diagenetic and tectonic ones. The latter are best recognised by their consistent orientation, parallel to the cleavage direction and cutting earlier structures. Late diagenetic stylolites cut geopetal and other cavity structures. They are therefore post cementation, and are themselves cut by tectonic veins and stylolites. The effect of the two later types is very similar though. With horizontal stylolites, the maximum amount of material lost, as seen from goniatites and shells, does not appear to exceed a few millimetres. Pressure solution is more prevalent in clay-rich parts of the the limestone and for this reason, many of the limestones have irregular bedding surfaces. The millimetre thick shale partings between beds are accentuated and distorted through pressure solution and much of the shale itself could have originated through late stage compactional pressure. Shale streaks (Flasers) frequently penetrate the limestones from the shale partings and commonly the Flasers are concentrated along the outer parts of the bed, suggesting an injection of clay into the limestone. Upper and lower parts of Flaser limestones may have a rather nodular appearance due to this same effect. In some cases Flaser limestones break laterally into nodules (Fig. 3.93) where the injection of clay has been more extreme.

Solution stringers (Schmidt, 1965) a few millimetres long and filled by clay are common in cricoconarid limestones (Fig. 3.92) and were formed by pressure solution coincident with the development of cleavage. One of the effects of extensive pressure solution has been to increase the clay content of the limestones. With the cricoconarid microcoquinas, for example, the clay is concentrated along the solution stringers.

Fig. 3.92 Solution stringers filled with clay cutting a diagenetic vein in a cricoconarid-rich limestone. Lower Frasnian. Bicken, Dill Syncline. Thin section S 23065. Scale bar = 1 mm.

Fig. 3.93 Flaser limestone with nodules developed tectonically within the bed. Frasnian. Bicken, Dill Syncline. Scale bar = 20 cm.





Section 3.6 Faunal preservation and ecology of the Schwellen limestones (Figs. 3.94 to 3.113).

The fossil abundance of the Schwellen limestones varies considerably from outcrop to outcrop and within sections, but generally the limestones are quite rich. Ammonoids are perhaps the best known fossils from these rocks (hence the term Cephalopodenkalk), but often the limestone is full of conodonts or in the Givetian and Frasnian with cricoconarids. Thin shelled bivalves and foraminifera are locally common. Faunal lists of Schwellen limestones are given by Beushausen (1900), Schmidt (1921), Fuhrmann (1954) and Rabien (1956).

Skeletal material in the Schwellen facies is preserved in several ways 1) with little alteration of the skeletal carbonate (or with aragonitic shells, paramorphic replacement) preserving the original structures, 2) solution of the skeletal carbonate (particularly if aragonite) forming a void which is later filled by calcite, producing a sparite pseudomorph, 3) micritization of the skeletal carbonate, 4) formation of overgrowths, and 5) replacement of skeletal carbonate by other minerals, particularly hematite and pyrite.

Cephalopods (Goniatites/Clymeniids/Orthocones)

In some Schwellen limestones, ammonoids are common (i.e. about 10 may occur in a 2 kg sample of limestone, e.g. Adorf am Martenberg). Horizons of the griotte in the Montagne Noire (particularly the Cheiloceras Stufe) are also very rich (Fig. 3.10, p. 56) and the goniatites give the rock a nodular appearance with calcareous shale concentrated between the 'nodules'. In many limestones from Germany, however, ammonoids are rare and are only occasionally seen in thin section (e.g. Bicken, and sections in the Harz Mountains). Goniatites and clymeniids generally have no preferred orientation and occur at all angles to the bedding, although large shells tend to be parallel to the bedding (cf. Reyment, 1970). Goniatites, clymeniids and orthocones, like the Mesozoic ammonoids, may be considered to have been nektonic or nektobenthonic.

The preservation of goniatites can be quite complicated and it is often difficult to decide how replacement of the original aragonitic shell occurred. Rarely, a two-layered shell structure is seen (Fig. 3.94), a thin outer layer and a thicker inner layer. Limonite may be deposited between the two layers, suggesting that one of them (the inner layer from its mosaic) passed through a void stage, during which time the limonite was precipitated. Where geopetal cavities were formed by goniatites, the shell wall and septa are commonly missing in the cavity. The void, filled by equant sparite, may have small crystals around the edge (Fig. 3.95) as most drusy cavities of this type (Bathurst, 1964) or be filled by several large crystals which are in direct contact with the sediment (Fig. 3.96). This situation in a geopetal cavity is considered by Bathurst (1964) to indicate that the shell experienced a void stage.

In other instances, the shell is still distinguishable although surrounded by drusy calcite (Fig. 3.97), suggesting that the wall did not go through a void stage.

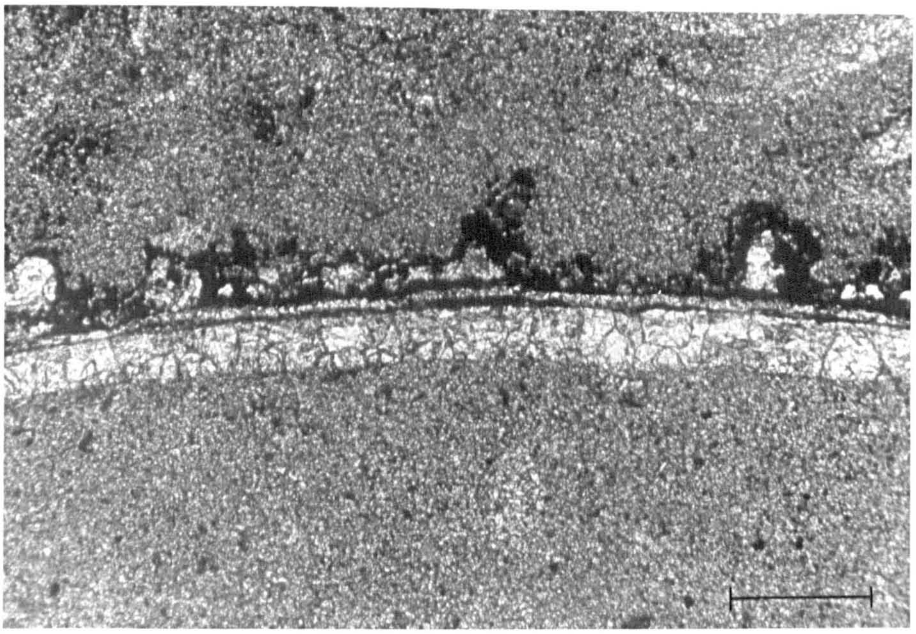
Some of the goniatites from the Montagne Noire have camerae filled by coarse calcite. One such goniatite (Figs. 3.98 and 3.99) has a complicated history. Internal sediment with ostracods was deposited on the septa but there is now no trace of the septa or the shell wall and the septal fills appear unsupported. The void was filled by equant calcite, showing many triple junctions, but few enfacial junctions (Bathurst, 1964). The calcite crystals themselves have strongly developed twin planes and undulose extinction, characters not typical of sparite druses. Around the edge of the cavity, calcite with undulose extinction is full of inclusions. The simplest explanation for the unsupported internal sediment is that the cavity was filled by an early cement which was later replaced by equant calcite. All trace of the septa was removed during the replacement (cf. Kendall, 1969). However, the same effect could arise from lithification of the internal sediment (so that it could support itself) followed by a phase of solution affecting the septa and shell wall. The void, containing cemented internal sediment, was then filled with calcite.

Bivalves

The characteristic types of lamellibranch in Schwellen limestone ~~[illegible]~~ mostly belong~~ing~~ to the genera Posidonia and

Fig. 3.94 Goniatite shell with a two-layered wall. Limonite is present between the layers and the mosaic of the inner (lower) layer suggests that a void stage existed. Lower Famennian. Combe D'Izarne, Montagne Noire. Peel S 22955. Scale bar = 200μ.

Fig. 3.95 Geopetal cavity formed by a goniatite shell. Cavity is filled by sparite druse with small peripheral calcite crystals. All evidence of the shell wall in the cavity has disappeared suggesting that a void stage existed. Frasnian. Bicken, Dill Syncline. Thin section S 23064. Scale bar = 200μ.



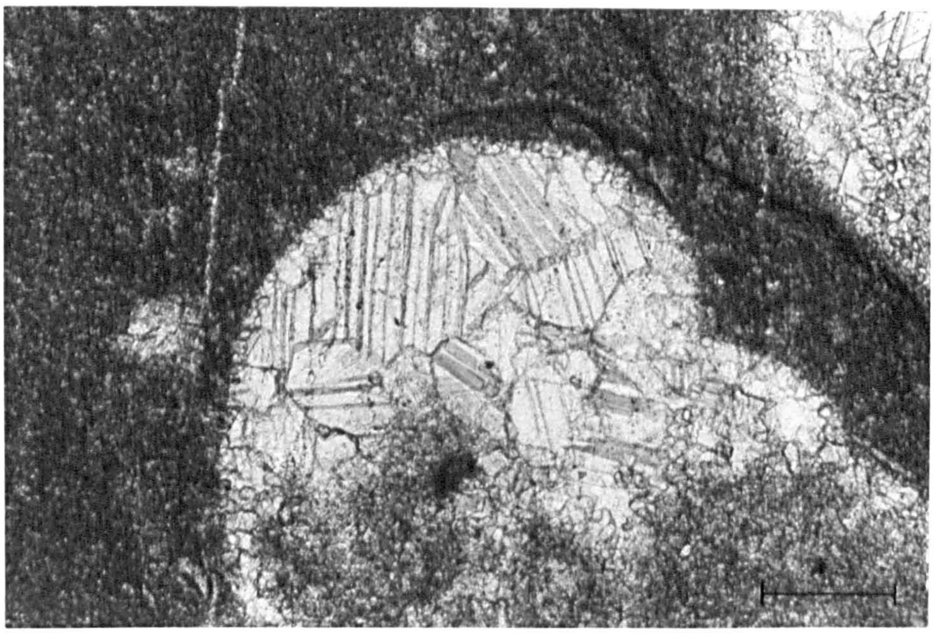
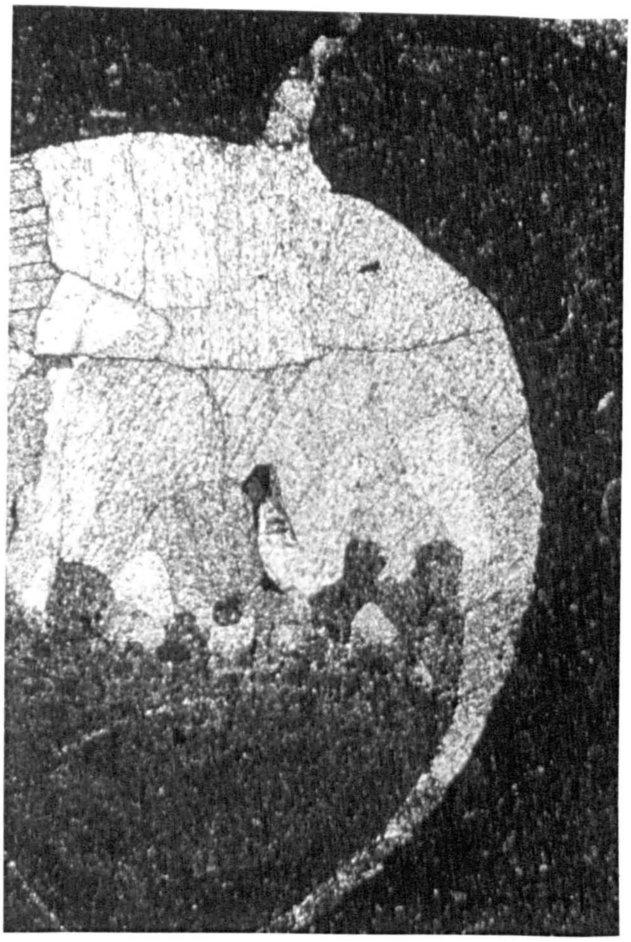


Fig. 3.96 Geopetal cavity formed by goniatite shell, filled by large calcite crystals. Frasnian. Adorf am Martenberg, Sauerland. Peel S 22983. Scale bar = 1 mm.

Fig. 3.97 Geopetal cavity formed by goniatite where the shell is still distinguishable within the drusy sparite. Platyclymenia Stufe, Upper Devonian. Langenholthausen, Sauerland. Peel S 22997. Scale bar = 1 mm.



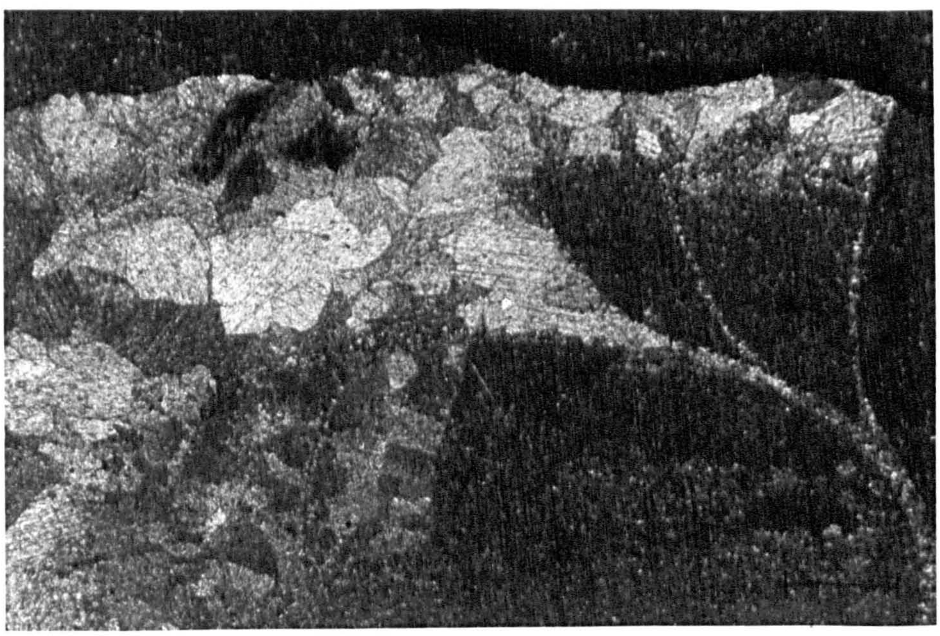
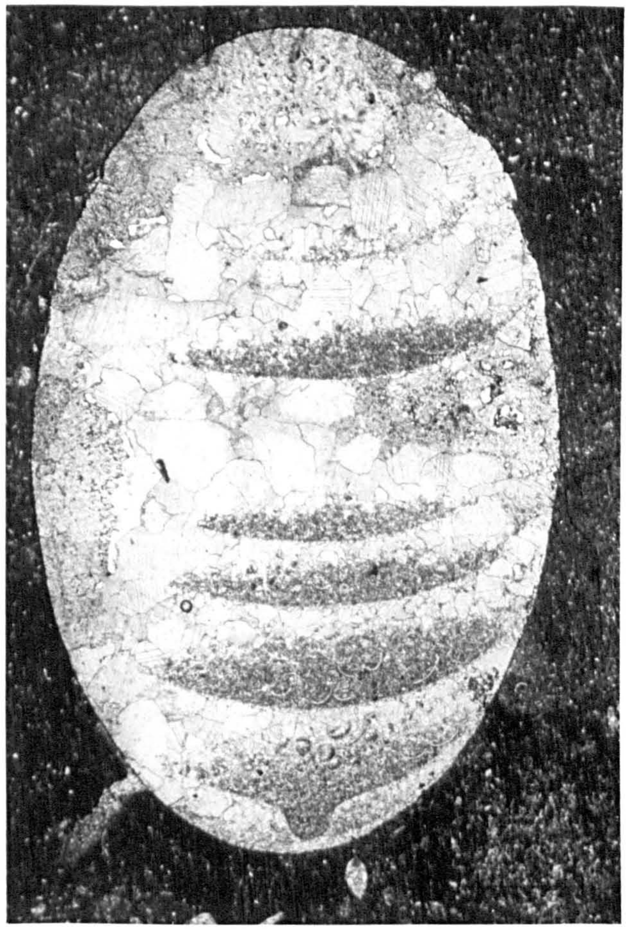
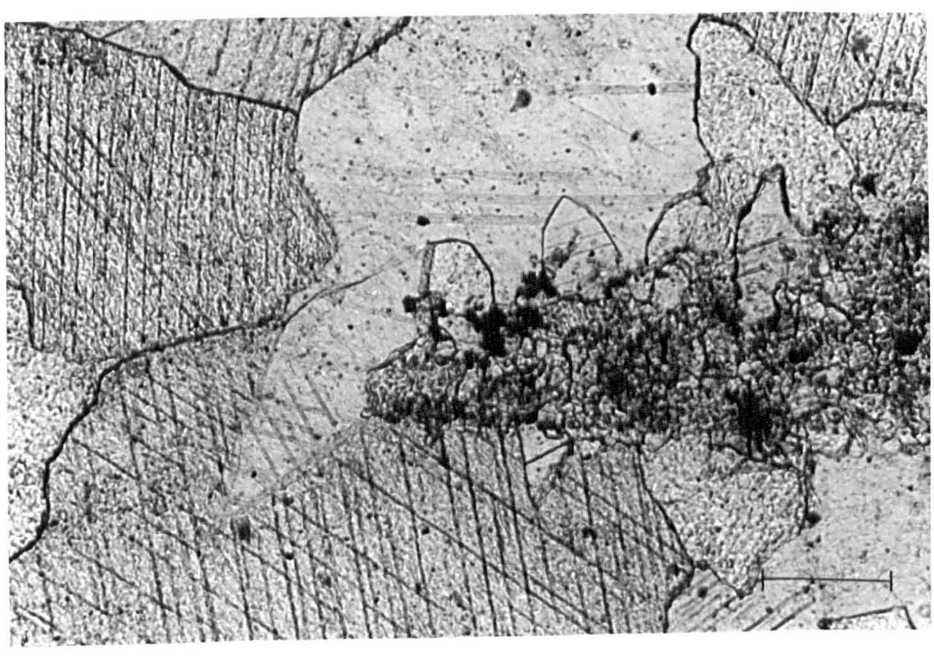


Fig. 3.98 Goniatite camerae with apparently unsupported internal sediment. Cavity filled by equant calcite with triple junctions. Lower Famennian. Mont Peyroux, Montagne Noire (See p. 170).
Thin section S 22960. Scale bar = 1 mm.

Fig. 3.99 Detail of goniatite fill above. No evidence of septum in the calcite filling cavity. Lower Famennian. Mont Peyroux, Montagne Noire. Thin section S 22960. Scale bar = 200μ.





Buchiola. These have a thin shell (less than 200μ thick) and are commonly about a centimetre across. Schmidt (1935) suggested a pseudoplanktonic mode of life for Buchiola, attached to floating weed. Jefferies and Minton (1965) recognised that adult Palaeozoic posidonids were nektoplanktonic. Very small bivalves, 2 to 3 mm across were obtained in some insoluble residues and may also have been pseudoplanktonic. Thicker-shelled bivalves, up to 5 cm across are not so common and mostly belong to the genera Cardiola and Kochia.

The thin shelled bivalves are common locally in Schwellen sediments. They may form coquinas a centimetre or less in thickness consisting soley of shells in a sparite matrix, which can be traced for up to a metre in the field. In most cases, however, they are randomly orientated in the limestones and mainly occur as single valves.

Very thin bivalves (shell wall <30μ thick) commonly have fibrous calcite overgrowths developed with their c-axes normal to the shell wall (Fig. 3.100) replacing or growing into the host sediment. Others have lost their outer walls and occur as a band of inclusions within large overgrowth crystals (Fig. 3.101).

Few bivalves in the Schwellen limestones show evidence of in-situ replacement, preserving the internal structure with inclusion rich bands (Fig. 3.102). In most cases the shell dissolved and the shell mosaic is now a granular calcite (Fig. 3.104 and 3.113). Micritic envelopes are present in a few cases with fine borings penetrating the shell (Fig. 3.103). Where these are broken, fragments of the envelope may occur within the shell (Fig. 3.104) indicating the existence of a void stage (Bathurst, 1966). Bivalves from the Montagne Noire commonly have a thin coating of limonite (Fig. 3.71, p.141) which behaved in a similar way to the micritic envelope and may also occur within the shell.

Partial micritization has occurred with some bivalves which may be due to the activity of boring organisms (Fig. 3.105) or through degrading neomorphism of the shell (Fig. 3.91, p.165)

Fig. 3.100 Thin-shelled bivalve with fibrous calcite overgrowth. Frasnian. Blauer Bruch (Ense Schwelle) Kellerwald. Peel S 23077. Scale bar = 1 mm.

Fig. 3. 101 Thin-shelled bivalve preserved as a band of inclusions with sparite overgrowth crystals. Lower Famennian. Bicken, Dill Syncline. Thin section S 23035. Scale bar = 200μ.

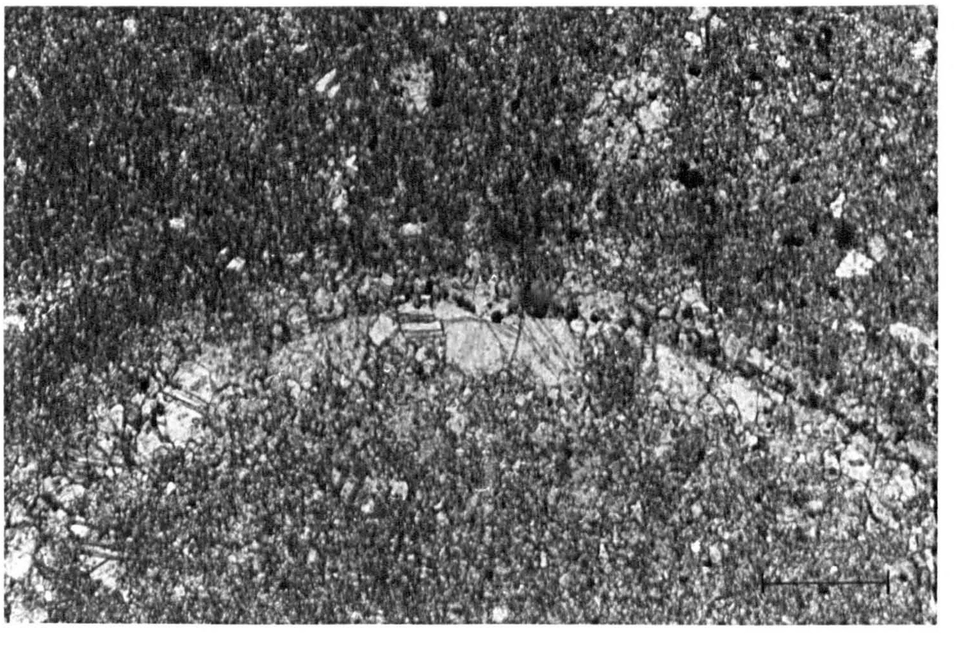
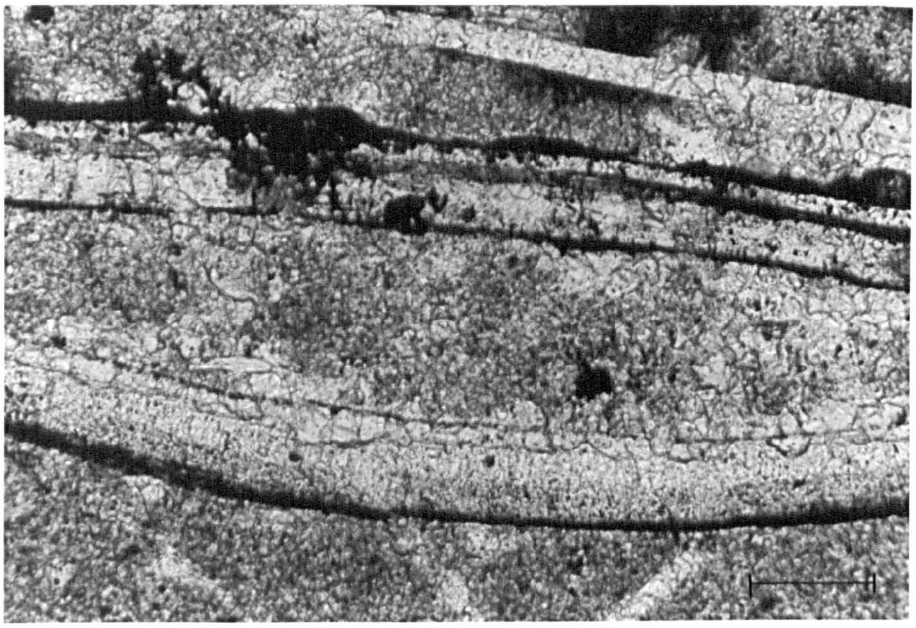


Fig. 3.102 Two bivalves showing original shell structures and inclusion-rich bands. A hematite coating is present on the outside of the shells. Lower Frasnian. Mont Peyroux, Montagne Noire. Thin section S 22957. Scale bar = 200μ.

Fig. 3.103 Bivalve with a limonitic coating on one side and borings penetrating the shell from this surface. Lower Famennian. Combe D'Izarne, Montagne Noire. Thin section S 22965. Scale bar = 100μ.



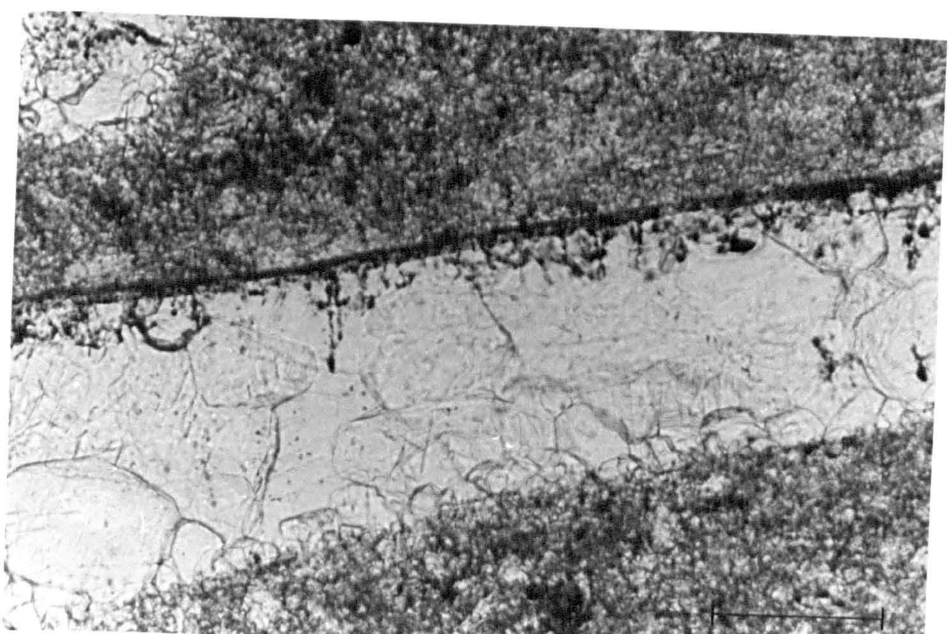
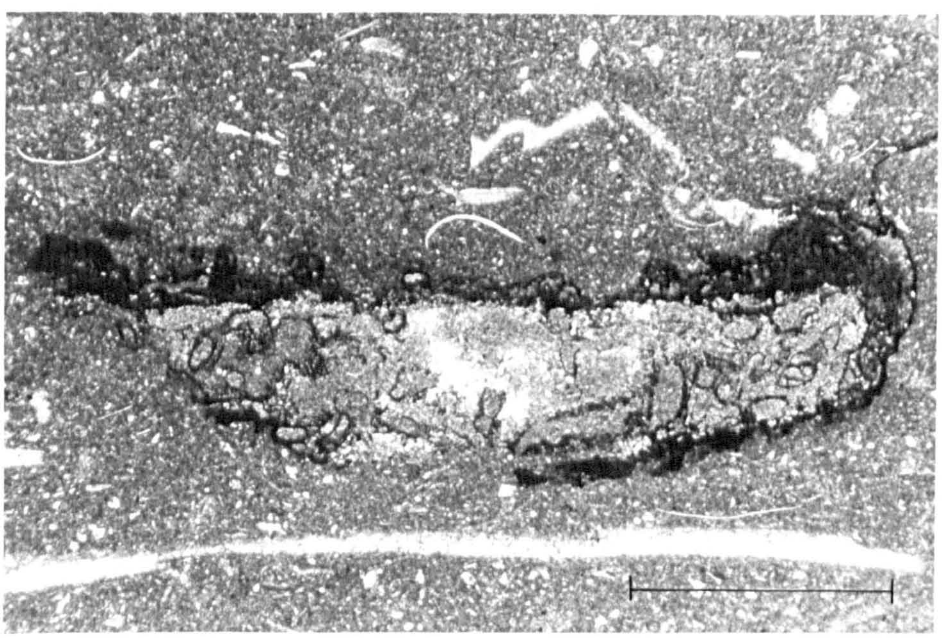


Fig. 3.104 Bivalve with a micritic envelope and fine borings on one side. The envelope is broken and a fragment occurs within the shell showing that a void stage existed. The shell is filled by sparite which has grown across the fracture. Lower Famennian. Combe D'Izarne, Montagne Noire. Thin section S 229546. Scale bar = 1 mm.

Fig. 3.105 Bivalve extensively bored and nearly completely micritized. Borings are coated in limonite, and ferromanganese occurs around the outside of the shell. Lower Famennian. Combe D'Izarne, Montagne Noire. Thin section S 22954a. Scale bar = 1 mm.



Gastropods.

Gastropods are rare in the pelagic carbonates. They vary in size from globular forms a few millimetres across, to turretted types up to 3 cm long. They have usually been preserved through a void stage, and are now sparite pseudomorphs with no trace of the original shell structure (Fig. 3.106). Pleurotomaria sp. and Naticopsis sp. appear in faunal lists but ecologically, these gastropods are not restricted to particular depths and the smaller ones may be planktonic.

Other macrofaunal elements.

Brachiopods are uncommon but orthids, rhynchonellids and athyrids have been recorded (Rabien, 1956). Insoluble residues occasionally yield single valves, less than 3 mm in length, of inarticulate brachiopods of the orders Lingulida and Acrotretida. These brachiopods, were probably pelagic and Rudwick (1965) suggested they may have been pseudoplanktonic attached to floating weed.

Crinoids are not important in the Schwellen limestones and mostly occur as single ossicles. Two hematized calyxes (2 mm in diameter) of microcrinoids were obtained from insoluble residues from the Montagne Noire. At Adorf am Martenberg (Sauerland) crinoidal limestones occur below the cephalopod limestones suggesting a greater depth of deposition for the latter (p. 26). Borings with a diameter of about 100μ commonly occur in crinoid fragments. Micritization of crinoids (Fig. 3.88, p.163) has rarely occurred and probably took place by degrading neomorphism rather than through the activity of boring algae.

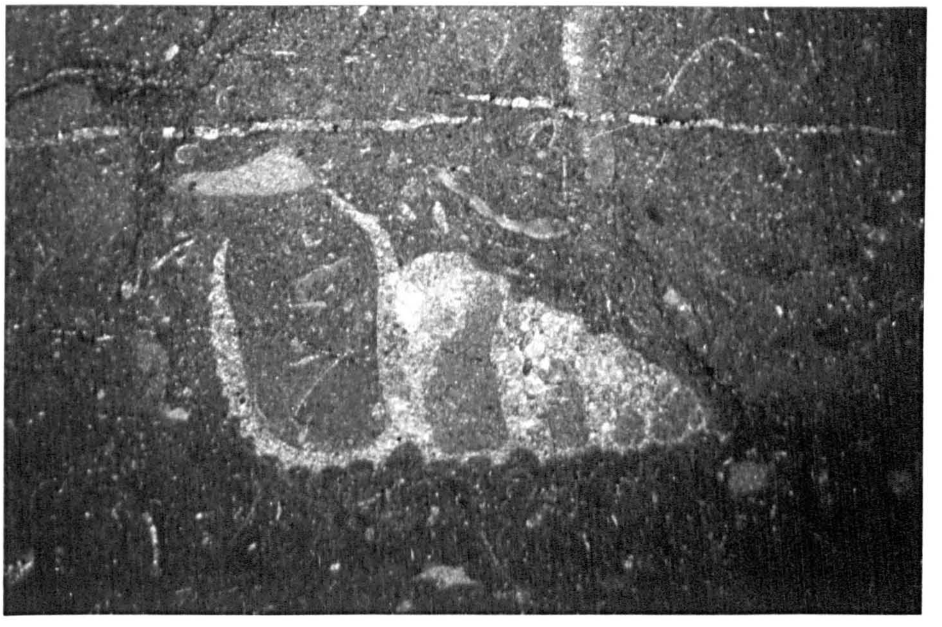
Corals are very rare and only single corals of the Syringaxon type (identified by Dr. C.T. Scrutton) have been found. Trilobites are occasionally seen in thin section where they are usually disarticulated. During the Upper Devonian most trilobites belonged to the blind phacopid and proetid groups (Clarkson, 1967).

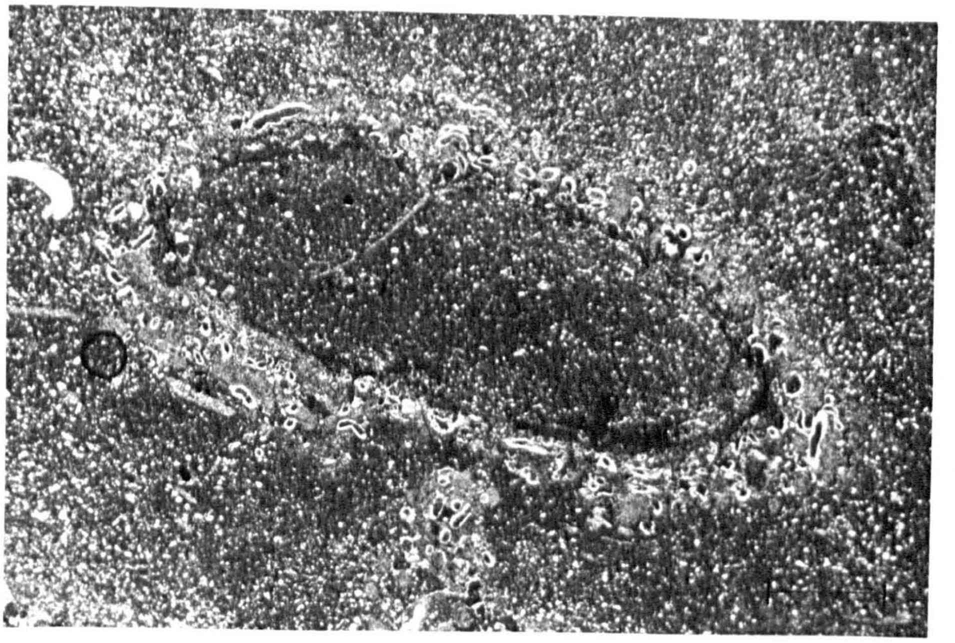
Cricoconarids.

Givetian and Frasnian Schwellen limestones are usually very rich in cricoconarids which may form microcoquinas with about 10% sediment (p. 70). These microfossils were easily affected by current

Fig. 3.106 Gastropod preserved as a sparite pseudomorph with no trace of the original shell structure. Lower Famennian. Combe D'Izarne, Montagne Noire. Peel S 22952. Scale bar = 1 mm.

Fig. 3.107 Limestone clast encrusted by arenaceous foraminifera and a red alga (light grey areas around clast). Frasnian. Adorf am Martenberg, Sauerland. Thin section S 22974. Scale bar = 1 mm.





activity and can be found with a preferred orientation. Their shells may also be packed inside each other. Three types occur, Styliolina, Tentaculites and Nowakia. The latter two are distinguished by a corrugated test and Tentaculites has septa. They are normally preserved with a fibrous calcite overgrowth which may be a cement or a replacement of the host sediment (p.103). Locally, hematization or pyritization of these microfossils has occurred. Most authors regard the styliolinids and Nowakia as pelagic (Rabien, 1956; Fisher, 1962; Boucek, 1964) in view of their thin tests. Their world wide distribution would also suggest this. Tentaculitids are not so common and with a thicker septate shell are considered to have been nektobenthonic.

Ostracods.

These do occur in the Schwellen limestones but are more common in shales of the slope and basin regions. Rabien (1956) considered the dominant group, the Entomozoacea, to be pelagic.

Conodonts.

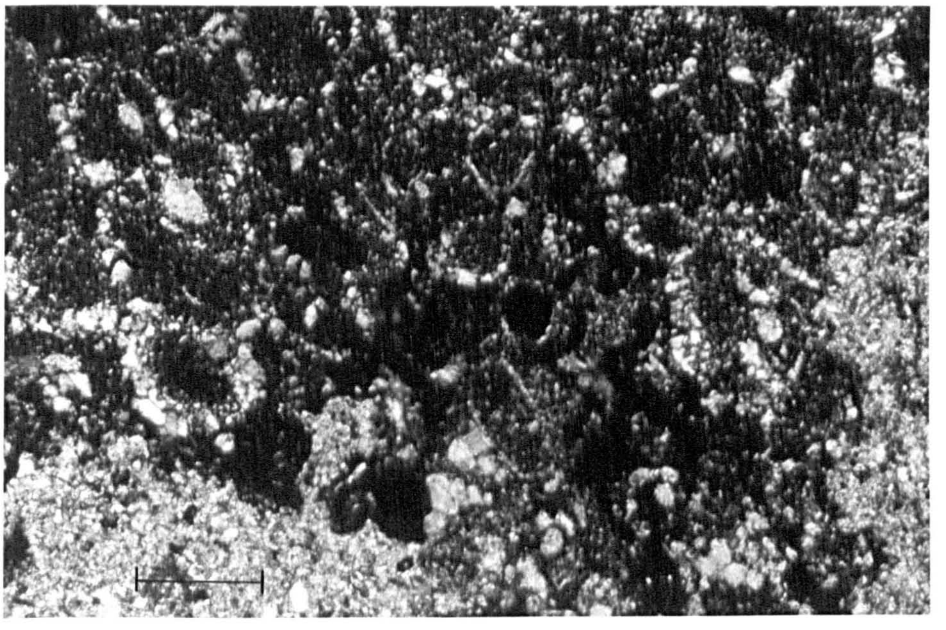
Conodonts are very abundant in Schwellen limestones and a few hundred gm of sediment can yield more than a 1000 individuals. Their abundance is probably due to slow rates of sedimentation and current activity removing fine grained carbonate but leaving the relatively dense conodonts. Even in the most condensed successions (e.g. Upper Devonian is 98 cm, Hutthaler Widerwaage, Meischner and Schneider, 1970) all conodont zones are represented and mixed faunas only occur at disconformity surfaces. Current ideas on the affinities of conodonts refer them to early nektonic fish (Halstead, 1969).

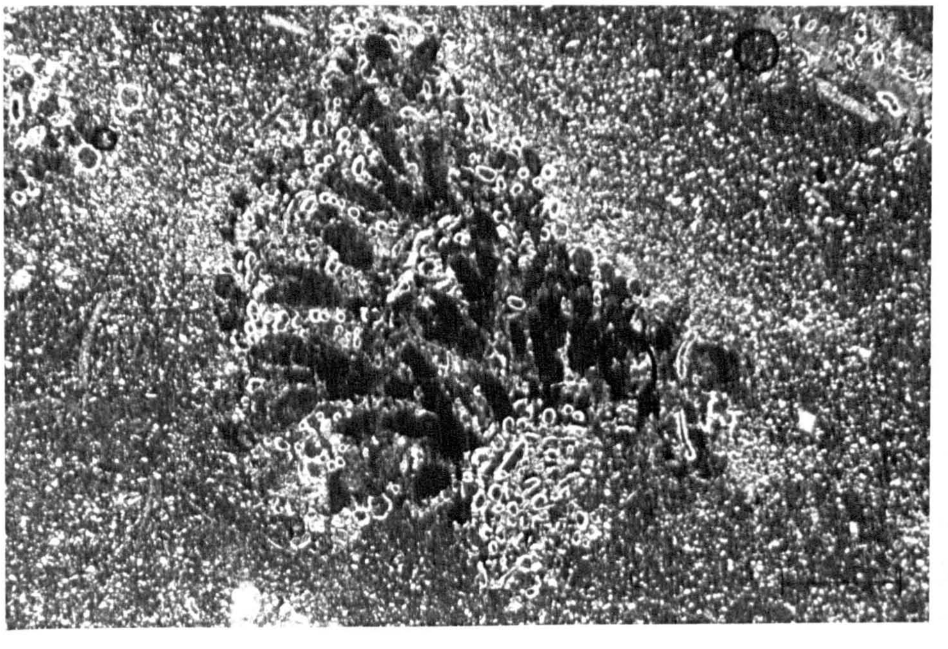
Foraminifera.

Arenaceous foraminifera are common at some horizons and can be found encrusting bivalves, goniatites and limestone clasts (Fig. 3.107). They also encrust cryptohardgrounds (p. 83) and occur within ferromanganese encrustations (p.300) (Tucker, 1971). Foraminifera also occur associated with algae in nodules. The foraminiferal tests, 100 to 200µ across, are composed of medium silt size quartz grains (Fig. 3.108). Some forms are simple meandering tubes while others are planispirally coiled. In many cases, the test is not a complete tube, but is absent where the organism was in contact with the

Fig. 3.108 Foraminifera of the genus *Tolypammina* with tests composed of numerous quartz grains. Frasnian. Adorf am Martenberg, Sauerland. Thin section S 22974. Scale bar = 200μ.

Fig. 3.109 Foraminiferal/algal association. Frasnian. Adorf am Martenberg, Sauerland. Thin section S 22974. Scale bar = 1 mm.





substrate. Most of these encrusting foraminifera belong to the genus Tdypammina (Eickhoff, pers. comm. 1970). The foraminifera are important since they are definitely benthonic. Wendt (1969) described encrusting foraminifera of the same genus from the Alpine Jurassic and comparing them with modern examples, considered a depth of deposition not exceeding 200 m (sublittoral zone).

Foraminiferal/algal nodules were found at Adorf am Martenberg where they occur as white clasts in the limestone. These two organisms have grown together (Fig. 3.109). Some nodules are rounded and have probably been rolled on the sea floor (Fig. 3.110). The alga has a growth form similar to that of Solenopora sp., a red alga. Schneider (1970) described foraminifera encrusting Upper Devonian conodonts and noted a possible association with algae. Wolf (1965) also records arenaceous foraminifera growing with algae. Foraminiferal/algal nodules occur in pelagic sediments on the Yucatan Shelf (Logan, 1969). They occur predominantly between 33 and 66 m, but locally extend up to 20 m. These nodules are not attached but are occasionally moved by strong currents.

Calcispheres.

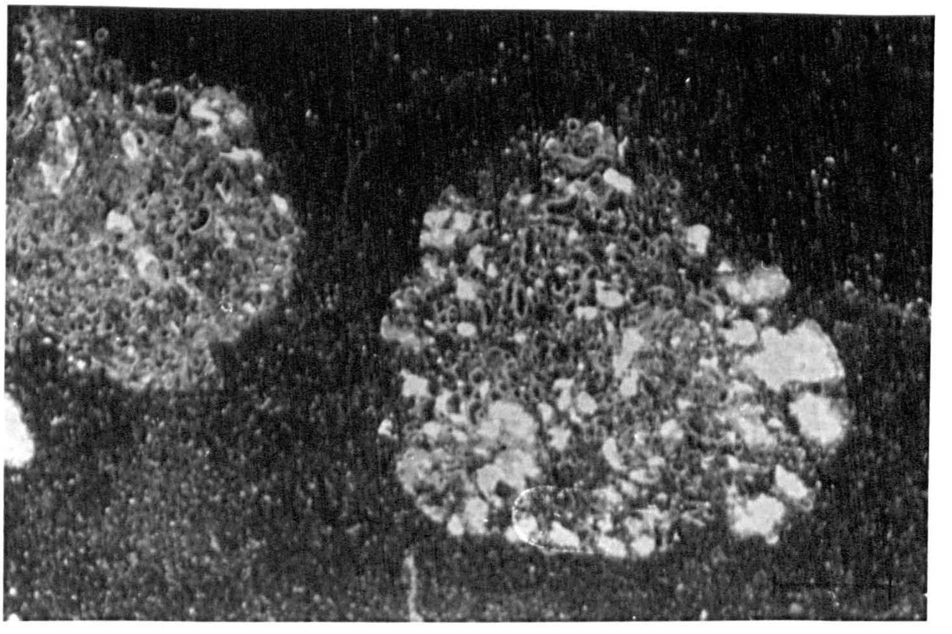
Spherical calcitic structures have recently been described from the Givetian and lower Frasnian 'reef' limestones in Germany (Flügel et al, 1971) but they are rare in the pelagic facies. Similar bodies, with a diameter of 100μ were found in the Frasnian limestones from Bicken. They have a thin outer wall composed of single calcite crystals, 2μ in diameter, and are filled by coarse calcite crystals. These structures may be algal in origin and are probably planktonic (Flügel et al, 1971).

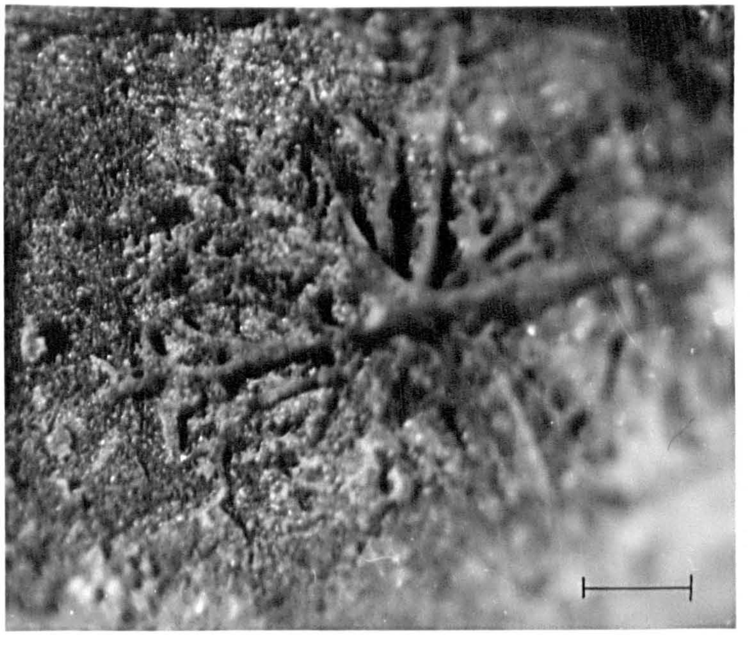
Bioturbation.

Burrow structures are present in many limestones and vary in diameter from 0.5 mm to 5 mm. They are commonly filled by a slightly coarser sediment, or calcite druse and have diffuse boundaries. Some have been modified during diagenesis and now resemble 'Stromatactis' structures. They are rarely visible on bedding surfaces and are then simple meandering types, occasionally branching. In one case where trace fossils could be seen on a surface, there was no indication

Fig. 3.110 Foraminiferal/algal nodules. Frasnian.
Adorf am Martenberg, Sauerland.
Thin section S 22974. Scale bar = 1 mm.

Fig. 3.111 Radiating borings in a brachiopod shell.
Lower Frasnian. Aeketal, N.W. Harz.
Hand specimen S 23036. Scale bar = 1 mm.





of their presence in thin section. Bioturbation may have been more extensive than is now apparent, and been destroyed during diagenesis by recrystallization.

Larger borings

Brachiopods from a calcareous shale at Aeketal (Fig. 3.111), and goniatites from Adorf contain borings 100 to 300μ across. Those at Aeketal radiate from a central point, with main branches giving off smaller ones. Those from Adorf are more irregular in arrangement, and meander in the shell, crossing each other but rarely branching. Pits 0.2 mm across also occur in the shells. Echinoderm fragments in thin section commonly have borings with a similar diameter (Fig. 3.112). Schindewolf (1962) described borings (with a diameter of 15 to 20μ) in ammonite shells of the lower Jurassic and recorded similar borings in goniatite shells from Adorf. A fungal origin was suggested and this interpretation could well apply here, although the borings are larger. Boring polychaete worms (e.g. Polydora sp.), bryozoans and sponges also produce borings of this size.

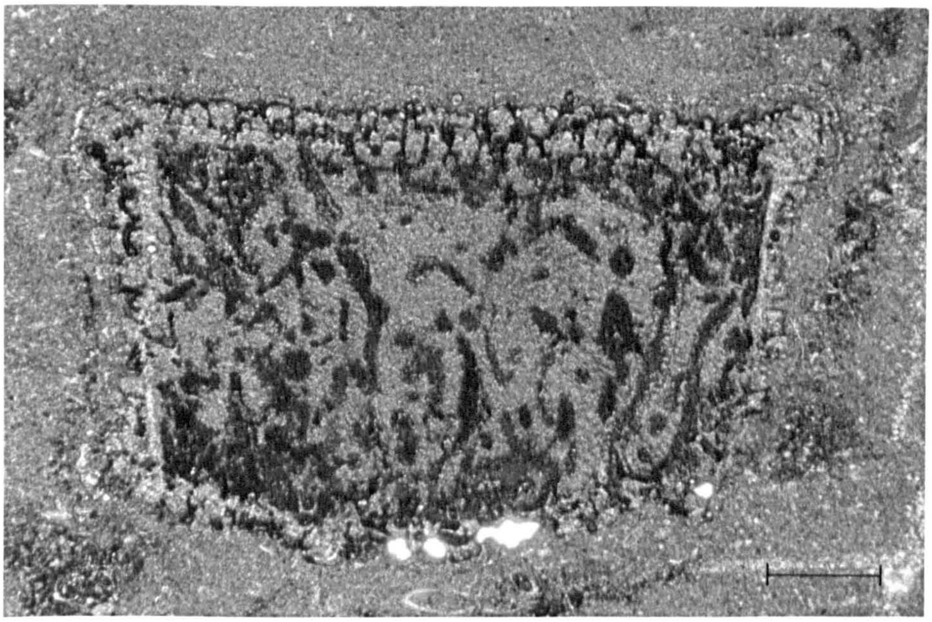
Borings in bivalves from the Montagne Noire have similar diameters (100 to 200μ) to the German examples and were probably made by the same organism. Three types of preservation are noted, a) filled with micrite with a vague border, gradational into the sparite of the shell, b) filled with micrite, but with a definite border and c) borings filled with equant calcite (Fig. 3.113). Most are filled with brown micrite and in places much of the shell is micritized (Fig. 3.105). The bored surface in the shell is commonly coated in brown limonite, indicating that the borings were open for some time before they were filled by micrite.

Smaller borings

Some lamellibranchs from the Montagne Noire have a hematitic or limonitic envelope formed by boring organisms. The borings are 2 to 4μ in diameter (Fig. 3.103) and are similar to the micritic envelopes of algal origin described by Bathurst (1966) and Wolf (1965). Borings of this size enriched in iron also occur in the Jurassic pelagic limestones (Jenkyns, 1970). If the borings are algal in origin then this would indicate a depth of deposition within the photic zone

Fig. 3.112 Bored crinoid which is partly micritised. Ferromanganese encrustations and 'limonitic cauliflowers' are also developed. Lower Frasnian. Mont Peyroux, Montagne Noire. Peel S 22959. Scale bar = 1 mm.

Fig. 3.113 Bivalve with borings, some filled with micrite others with sparite. A micritic envelope occurs around the central boring. The shell has an irregular surface and is encrusted with foraminifera and ferromanganese. Lower Famennian. Combe D'Izarne, Montagne Noire. Thin section S 22954a. Scale bar = 200μ.



(less than 150 m, Hudson, 1967). However, algal borings are similar in size to fungal borings which are not restricted in depth (Friedman et al, 1971). The presence of a micritic envelope around a large boring within a shell (Fig. 3.113) suggests that these borings may indeed be fungal in origin.

Summary and depth of deposition

The fauna is dominated by pelagic organisms, with goniatites, conodonts, cricoconarids and posidonid bivalves being the most common. This aspect of the fauna suggests 'relatively deep water', where conditions were not suitable for benthonic organisms. Encrusting foraminifera were found at a number of localities and enable one to be more precise about the depth of deposition. Following Wendt (1969) for the Jurassic, the foraminifera encrusting hardgrounds and skeletal fragments indicate a depth of 200 m or less. The presence of foraminiferal/algal clasts by analogy with modern examples suggest even shallower depths, in the region of 50 m.

The stratigraphic position of some Devonian pelagic sediments, commonly occurring above shallow water carbonates (reefs), suggests depths of deposition greater than that suitable for stromatoporoids, rugose and tabulate corals (e.g. Schwellen sediments in S. Devon, Scrutton, 1969). During the upper Givetian and Frasnian, cephalopod limestones accumulated on the flanks of reefs, contemporaneous with reef growth, and contain thin detrital bands of reef debris (e.g. Attendorn region, localities Bonzel and Grevenbrück). The Haingrube volcanic Schwelle in the Kellerwald supported calcareous algae, corals, stromatoporoids and brachiopods at one point, while a few kilometres along the ridge, pelagic carbonates accumulated (Schneider, 1969). Recent reef corals grow at depths of less than 25 m, and maximum growth takes place at 10 m or less (Stoddart, 1969). Illumination is considered to be the most critical factor.

In conclusion, certain faunal elements and stratigraphical relationships suggest a depth of deposition in the lower sublittoral zone to the upper bathyal zone, from some tens of metres to a few hundred metres.

Section 3.7. The formation of Devonian condensed limestones

Jenkyns (1971) recently reviewed the genesis of Jurassic condensed sequences and concluded that two processes are involved. These are stratigraphic condensation, where little sediment is entering, or being formed in, the depositional area (starved basin situation) and reworking, where reduction in thickness occurs through current activity. With the Schwellen limestones both processes were operative. Sedimentary structures within the Devonian limestones (laminations, fossil concentrates, hardgrounds and limestone intraclasts) attest to current activity which could also have removed fine carbonate particles off the Schwellen onto the adjoining slope regions. The hematite enrichment of the pelagic limestones in the Montagne Noire and the presence of ferromanganese nodules suggests a low organic content which may result from current activity (p.316). Stratigraphic condensation is suggested by the general reduction in thickness of the Upper Devonian over the whole Rhenish geosyncline. The shales deposited in the deeper water areas between the Schwellen are about 300 m thick for the Upper Devonian in the N.W. Harz (Müller-Steffen, 1962). The thickness for a comparable time span (10 million years) in the Lower Devonian is several kilometres (Fig. 2.2, p.14). Also in the inter-rise depressions on the mid-geosynclinal ridge, condensed cherts and shales were developed suggesting a 'starved basin' situation in this case caused by the remoteness of land. A slow net rate of sedimentation on the Schwellen caused by stratigraphic condensation and current activity are considered to be responsible for the reduced thickness of the cephalopod limestones.

The carbonate of the Schwellen limestones was probably biogenic in origin, similar to Jurassic (Fischer et al, 1967; Jenkyns, 1971) and Recent (Bramlette, 1957) pelagic calcareous sediments. Whether the carbonate was generated uniformly over the whole depositional area (giving a pelagic 'rain') or developed only above the Schwellen is not known. The absence or paucity of carbonate in the basinal shales might suggest the latter, but solution during diagenesis could have removed $CaCO_3$ from the shales.

Section 3.8

History of Sedimentation of the Devonian pelagic limestones

Sedimentation

Sedimentation of the Devonian pelagic carbonates was generally rather quiet giving rise to fine grained limestones with randomly orientated microfossils. Depths of deposition probably never exceeded a few hundred metres and in many cases was 50 m or less. The micritic carbonate is considered to be biogenic in origin - mainly by analogy with Tethyan Jurassic limestones of the Alps and Recent pelagic carbonates. With coarser limestones the original sediment was composed of a high proportion of finely broken skeletal fragments. During periods of current activity microcoquinas were deposited and occasionally the pelagic sediment itself was reworked into horizontal and cross lamination. Deposition from low density suspension currents (or nepheloid layers) gave rise to graded units with erosive bases rich in terrigenous material.

During periods of reduced sedimentation, or pauses in sedimentation, local cementation of the carbonate sediment occurred on the sea floor, and gave rise to hardgrounds. The cement causing lithification was probably a micritic carbonate, but in view of the fine grained nature of the sediment and later neomorphism, it is not possible to detect the cement. Once formed, the hardground surfaces in the Schwellen limestones were modified by two processes, corrasion and subsolution. Corrasion hardgrounds are characterised by a smooth surface cutting cavity cements and skeletal material. Cryptohardgrounds, showing evidence of solution, have a more irregular relief and may also truncate shells.

Ferromanganese nodules are locally developed (particularly in the Montagne Noire) and formed on the sea bottom. This is shown by the presence of encrusting foraminifera actually within the ferromanganese crust (p.294).

Evidence of bioturbation is found in many of the limestones but in most cases it appears to be simple burrowing of soft bodied organisms of various sizes. The homogeneous nature of most Schwellen limestones may be due to extensive bioturbation.

Early Diagenesis

During early diagenesis - i.e. after some burial, sheet cracks and cavities and neptunian dykes formed. The lack of compaction suggests that lithification occurred during early diagenesis. Some sheet cracks formed through the cementation of the host sediment others through shear failure on a slope. Sedimentary dykes also formed after lithification of the sediment. The internal sediments in these structures are graded calcisiltites and calcilutites, occasionally with microfossils.

Acicular cements were precipitated in the cavities which were later replaced by fibrous calcite. The exact timing of this replacement is not known, and so this event is placed in late diagenesis. Also with goniatites and some bivalves the timing of aragonite solution producing a void which was later filled by sparite, is not known. In general early diagenetic events proceeded as follows:-

a) cementation of sediment either by a micritic cement or acicular overgrowths around cricoconarids,

b) formation of sheet cracks and cavities, and sedimentary dykes. Some of the smaller sheet cavities could have formed in unlithified, but coherent sediment,

c) partial filling of the open space structures by internal sediment,

d) precipitation of acicular druses in the voids and at times within the internal sediments during pauses in sedimentation,

e) solution of skeletal aragonite producing a void which was later filled by sparite.

Later Diagenesis

Acicular cements, in cavities and as overgrowths around cricoconarids were replaced by radiaxial fibrous calcite. This calcite also locally replaced the host sediment. The replacement may have been accomplished by inversion, if the original fabric was aragonitic. Generally neomorphism of the sediment, leading to a coarser mosaic, is probably late diagenetic since in places fibrous replacement crystals have themselves been replaced by microsparite. However, other evidence shows that degrading neomorphism has occurred and this may have reduced a large amount of coarse skeletal debris to microsparite. Coarse silts were intoduced as the final internal sediment of cavity structures, and remaining voids were filled by equant drusy sparite. Pressure solution gave rise to horizontal stylolites and microstylolites, which cut geopetal cements.

Generally the order of late diagenetic events is as follows:-

1. Replacement of acicular druse by radiaxial fibrous calcite, and local replacement of sediment by fibrous calcite.
2. Neomorphism of host sediment to microsparite and local replacement of fibrous calcite by microsparite.
3. Degrading neomorphism of skeletal debris to microsparite.
4. Coarse hematitic silts introduced into cavities as final cavity fill.
5. Final occlusion of voids by equant drusy sparite.
6. Pressure solution leading to stylolites and microstylolites, mostly parallel to the bedding, cutting earlier cements.

Tectonic Effects

Deformation of the Schwellen limestones has produced numerous solution stringers and clay streaks. This was caused by tectonic pressure solution, coincident with the development of cleavage and led to the production of Flaser limestones.

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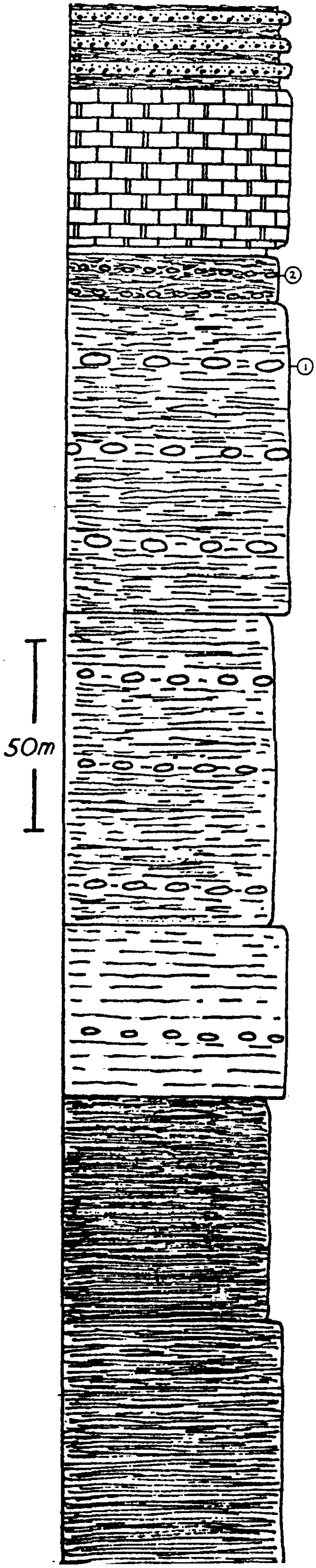
CHAPTER 4

Basin and Lower Slope Facies

The basinal facies of the Givetian and Upper Devonian in Germany has been referred to as the Büdesheimerschiefer and Cypridinen-schiefer by German geologists. This facies, normally a relatively thick succession of shales, locally with turbidites, was deposited in the deeper water areas between topographic highs (Schwellen), where much thinner successions, mainly pelagic carbonates accumulated. For instance, in the Upper Devonian of the north-west Harz Basin (Fig. 2.6, p. 29) more than 300 m of this facies is present, equivalent in age to about 12 m of pelagic limestone, deposited on the West Harz Schwelle, to the south. In some areas (e.g. Innerstetal, N.W.Harz) silty shales comprise some 90% of the succession (a typical basinal succession is shown in Fig. 4.1) and shales with calcareous nodules form the remainder. As one moves towards the Schwellen area the succession decreases in thickness and shales with diagenetic nodules become more important. This lithology may make up 60% to 70% of the succession in the lower slope region (Fig. 4.2). The slope sediments (chapter 5) are distinguished on the presence of slumped or re-worked sediments. Turbidites are important in the deeper water successions of the Rheinisches Schiefergebirge during the Givetian and at certain times during the Upper Devonian. Similar basinal shales with turbidites are developed in the Middle and Upper Devonian of S.W. England. Volcanic horizons, mostly thin tuff bands may be intercalated with the shales.

Sections in the basinal facies have been examined in the North West Harz at Junkernberg, a quarry south of Langelsheim,

Fig. 4.1. Basin succession in the North West Harz (Junkernberg region) after Muller-Steffen (1962).



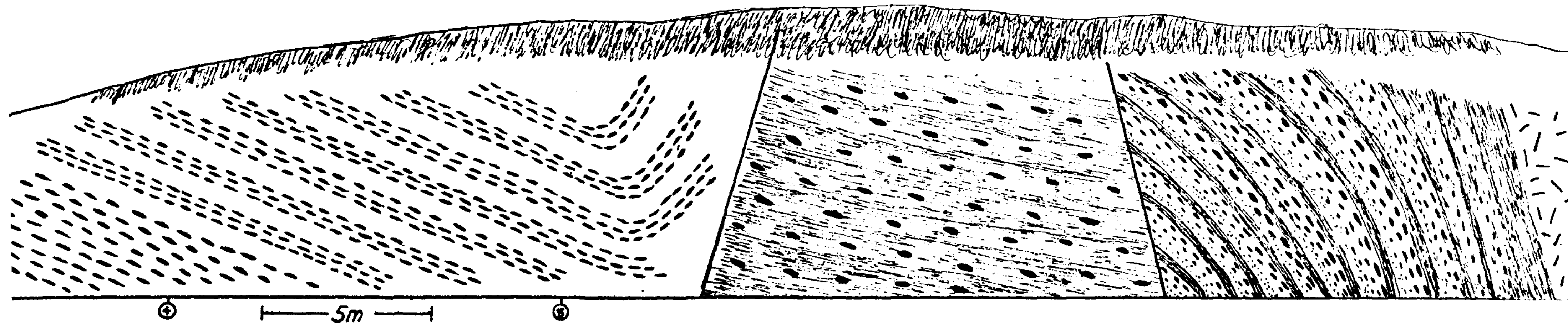


Fig.4.2. Upper Devonian/Lower Carboniferous Succession, railway cutting, south end of Innerstetalsperre, N.W.Harz Mountains.

and the railway cutting along the Innerstetal Resevoir (Fig.2.6,p.29). Smaller outcrops were examined in the Brilon and Arnsberg areas (Sauerland), Kellerwald and in the Dill and Lahn Synclines, in the Rheinisches Schiefergebirge. The distribution of this facies for the Upper Devonian is shown in Fig.2.8, p.29. The sediments are described under the headings 4.1. Shales, 4.2 Fauna, 4.3 Turbidites, 4.4 Shales with nodules and 4.5 Origin of calcareous nodules.

Section 4.1 Shales

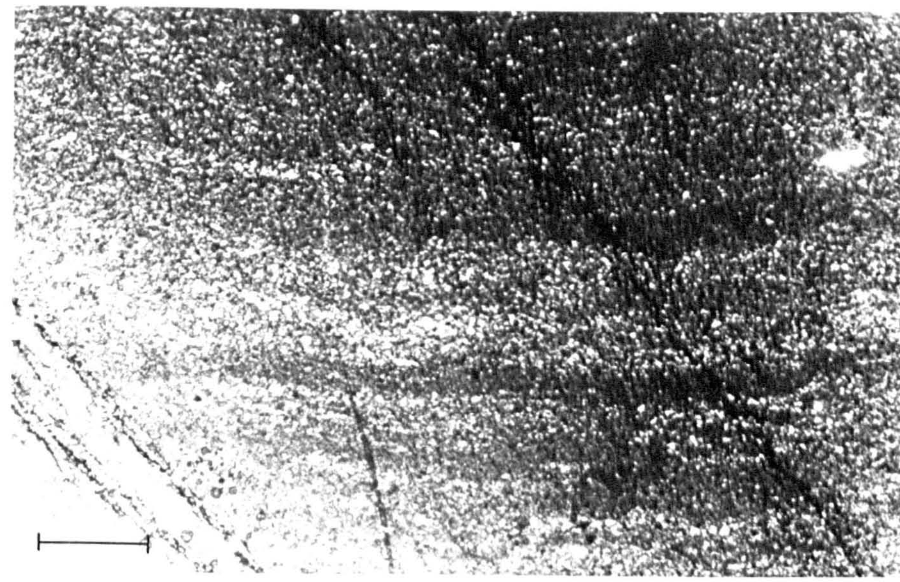
Shales, the background sediment of the Rhenish geosyncline during the Givetian and Upper Devonian, show a variety of colours from bright red to jet black. Generally, grey shales characterize the Middle Devonian and Frasnian in Germany and S.W. England, and lighter colours predominate in the Famennian. Locally, the colours are sufficiently restricted in time for use in geological mapping. Red (or purple) and green shales are developed in Germany and S.W. England during the lower part of the Famennian. The colouration is not confined to the bedding and patches of green shale occur within the red. These may be isolated green areas, or developed along tectonic fracture planes. Commonly the green shales are coarser grained and show better silty lamination. There appears to be a primary colour alternation, which is later modified by diagenesis. Similar colour effects are common in fluviatile sequences and have been related to contemporaneous water table fluctuations (Friend, 1966). Movement of connate and meteoric waters during burial and after uplift has probably affected the colours of these marine Upper Devonian shales.

Apart from colour, three types of shale can be recognised, a) shales with fine silty laminae, b) calcareous shales and c) carbonaceous shales.

Shales with fine silty laminae are most commonly developed (Fig. 4.3). The laminae, due to medium and coarse silt (<60μ), vary in thickness from a few grains to 3 or 4 mm. Some laminae

Fig. 4.3 Shale with silty laminae up to 1 mm thick. Platyclymenia Stufe, Upper Devonian. Junkernberg, Innerstetal, N.W. Harz. Thin section S23070b. Scale bar = 1 mm.

Fig. 4.4 Black laminated carbonaceous shales with scattered silt grains. Frasnian. Innerstetal Reservoir, N.W. Harz. Thin section S 23054. Scale bar = 1 mm.



have sharp bases but most are gradational into the background sediment. The silty laminae must represent deposition from low density suspension currents or nepheloid layers (the latter process is described in chapter 3, p. 69). In some cases, bioturbation has disrupted or destroyed the lamination. Burrows, 2-3 mm across, are usually more quartz-rich than the surrounding sediment. Calcareous shales have the same silty lamination. The carbonate can comprise up to 30% of the sediment and is ferroan dolomite (identified by staining technique). The carbonate grains are now microsparite (<20μ in size) but larger grains probably representing former skeletal fragments do occur. Dark grey or black carbonaceous shales have a fine banding (Fig. 4.4), which is not due to the presence of silt, but to varying compositions of the clay fraction. The banding may be the result of a variable carbon content. Pyrite occurs finely disseminated in the sediment and as nodules up to 2 cm across. This also indicates an original high organic content. These carbonaceous shales are commonly developed during the Frasnian.

Section 4.2 Fauna of the Shales

The fauna of the basinal shales is impoverished and dominated by pelagic forms. Styliolinids and tentaculinids are most common in the Givetian and Frasnian shales, and ostracods predominate in the Famennian. The exact affinities of the styliolinids and tentaculinids are not known and they are placed in their own group, the cricoconarids (Fischer, 1962). The styliolinids are considered to be pelagic whereas the tentaculitids with thicker walls, often septate, are probably nekto-benthonic. These microfossils are rarely found with a preferred orientation (also noted by Rabien, 1956) and minimal current activity is suggested. The ostracods are usually 1 to 2 mm across and had very thin shells composed of chitinous and calcareous material. Most belong to the Entomozoacea group and Rabien (1956) compares them with the modern Halocypriden which are active swimmers. Conodonts are ubiquitous and current ideas on their affinities relate them to early nektonic fish. The world wide distribution of these ostracods, conodonts and cricoconarids indicate that they too were pelagic. Thin-shelled bivalves (posidonids) occur sporadically in the basinal shales and were probably nektoplanktonic (p. 177). Coniatites are rare and occur as compressed imprints with no trace of the shell.

Rabien (1956) described eight types of trace fossil from the Rheinisches Schiefergebirge all of which he ascribed to soft bodied animals, probably annelids. All of these were found on the soles of clastic turbidites in Famennian shales. Trilobites are occasionally found in the basinal shales and most of these belong to the blind Phacopid group. Clarkson (1967) showed that blindness in modern arthropods begins at about 600 m depth, and suggested that for the basinal sediments, there is no need to infer a depth greater than 1000 m.

Section 4.3 Turbidites in the basinal shales

In the Rheinisches Schiefergebirge and the Harz Mountains two main types of turbidites occur in the basinal facies between the Middle Devonian and Lower Carboniferous, a) carbonate turbidites and b) clastic turbidites.

Carbonate turbidites. Carbonate turbidites (Flinz limestones, or allodapic limestones, Meischner, 1964) consisting of allochthonous reef material are well developed during the Middle and lower Upper Devonian in the Rheinisches Schiefergebirge and were mainly derived from the shelf area to the north. Reef growth on the shelf during the Lower Carboniferous produced more carbonate turbidites (Kulm-Plattenkalk, Rhena Kalk). In the Harz Mountains Flinz limestones occur in the vicinity of the Elbingerode reef during Givetian and lower Frasnian times, but are absent in the N.W. Harz basin.

In S.W. England, carbonate turbidites are developed in the Middle and Upper Devonian shale facies of N. Cornwall (Marble Cliff Beds) and South Devon, in the Torquay district. Those in the latter region (exposed at Saltern Cove, Galmpton Point and Elberry Cove) are 'reef' detrital beds of Frasnian age showing some of the features of the allodapic limestones of Meischner (1964). Crinoidal turbidites are exposed in the Padstow area, N. Cornwall and a description of these is appended.

Clastic turbidites. Clastic turbidites are either clean sandstones derived from the shelf area to the north, or immature greywackes derived from the south. The former were prominent during the Upper Devonian in the Rheinisches Schiefergebirge (Wülstsandsteine), and are composed of presorted shelf material, mostly of coarse silt and fine sand (Einsele, 1963). The density currents flowed around the former reef areas which still had topographic expression (Flessmann, 1962). Greywackes gradually became an important feature of basinal deposition from the Middle Devonian onwards, culminating in the Lower Carboniferous. They were derived from the Mittel Deutsche Schwelle, the synsedimentary mountain chain to the south, separating the Rhenish geosyncline from the Thuringian geosyncline (Kuenen and Sanders, 1956). In the Rheinisches Schiefergebirge and the Harz Mountains, the age of the greywackes becomes progressively younger as one goes north, so that in the N.W.Harz basin their age is Goniatites Stufe, Lower Carboniferous.

Section 4.4 Shales with nodules

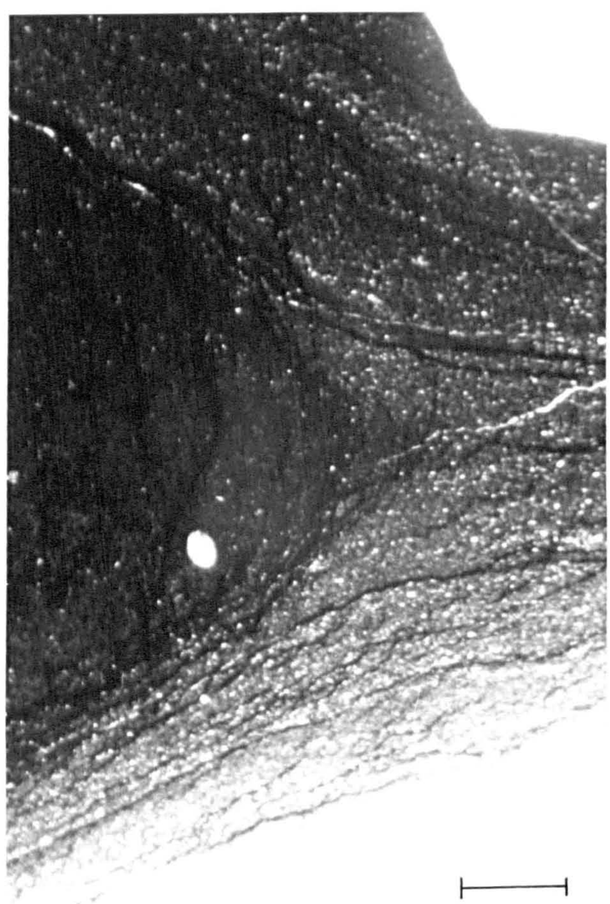
Calcareous nodules occurring in basinal shales are not common but where present range from 10 to 30 cm in diameter and from 3 to 7 cm in thickness. Nodule bands are between 10 and 50 cm apart. Generally, the nodules are larger and nodule bands are further apart than those occurring up the slope. This is a reflection of the carbonate content; when it is low, nodules are larger and bands are further apart, but with a higher content, nodules are thinner and tend to coalesce and bands are closer together. Thick, isolated nodules (Figs.4.5 and 4.6) usually have rounded ends suggesting that their shape has not been affected by compaction or tectonics. Thinner nodules (Fig.4.7) commonly have pointed ends, suggesting some degree of compaction. The nodule/shale boundary is gradational unless pressure solution has occurred along the junction. The nodules themselves are made of calcite; only one exception was encountered, that of ferroan dolomite nodules of Lowest Carboniferous age, from

Fig. 4.5 Grey shales with bands of calcareous nodules. Bands have unusual rhythmic arrangement, 3 or 4 bands occur together within 10 cm of shale, separated by 15 cm of shale without nodules. Nodules are mostly isolated, though some form nearly continuous bands. Trace fossils were found in these nodules. *Clymenia* Stufe, Upper Devonian. Railway cutting, south end of Innerstetal Reservoir, N.W. Harz. (Geochemical sample S 23005 collected here). Length of hammer = 35 cm.

Fig. 4.6 Red shales with isolated calcareous nodules which have rounded ends. *Platyclymenia* Stufe, Upper Devonian. Railway cutting, south end of Innerstetal Reservoir, N.W. Harz. (Geochemical sample S 23004 collected here).
Scale bar = 20 cm.

Fig. 4.7 Thin nodules in grey green shales, some have coalesced to form nearly continuous bands. *Cheiloceras* Stufe, Lower Famennian. Railway cutting, Innerstetal Reservoir, N.W. Harz. (Geochemical sample S 23003 collected here) Length of hammer = 35 cm.

Fig. 4.8 Silty laminae in shales compacted around a calcareous nodule. Difference in lamina thickness across nodule and in shale gives a compaction of 60% in the shale. *Cheiloceras* Stufe, Lower Famennian. Railway cutting, Innerstetal Reservoir, N.W. Harz. Thin section S 23073. Scale bar = 1 mm.



Innerstetal, N.W.Harz. The nodules generally have 70-80% $CaCO_3$ which is present as micrite and microsparite (grains of diameter 15μ and less). Patches of coarser carbonate grains may occur within the nodules and less commonly there is a general coarsening around the periphery of the nodules. There is much more quartz in the shale (up to 30% silt) than in the nodules themselves (1-5%) but this must simply be a reflection of differential compaction. The latter can be measured in some instances where good lamination is present (Fig. 4.8). A compaction of about 60% is indicated.

Trace fossils were found in the nodules of Clymenia age at Innerstetal, N.W.Harz, showing that nodule formation is early diagenetic. The traces are simple burrows occurring at all angles to the bedding but mostly vertical. Some burrows branch and meander. On size, two types can be distinguished. The majority are 2-3 mm in diameter and can be followed for 3 or 4 cm. A few smaller burrows occur with a diameter of less than 1 mm. These burrows can also be discerned in the shales, but very badly deformed by compaction and cleavage. In the nodules, the trace fossils are a lighter colour than the surrounding material due to a smaller amount of clay material in the burrow fills. The carbonate of the burrows is often coarser grained, 15-20μ. Drusy calcite occurs in some of the traces. The nodules often have a boudinaged appearance, but the presence of undeformed trace fossils in spite of the irregular shape of the nodules, shows that the nodules are purely sedimentary and have not been modified by tectonic effects. The trace fossils are of a very simple type, and probably represent the burrows of an annelid or some such worm-like organism.

Shells and trilobites occurring in the nodules are always uncompressed, whereas those in the adjacent shales are flattened in the bedding and distorted by cleavage. This too indicates early lithification of the nodules. Ostracods and bivalves in the shales are usually only imprints. The carbonate that formed their shells may have been dissolved at an early stage to contribute towards nodule formation. Shells in the nodules may have geopetal structures filled by drusy calcite if they provided a cavity. One such structure at the top of a nodule (Fig.4.9) shows a collapsed cavity which is filled with drusy calcite. This suggests that the cavity-fill stage of lithification came after compaction of the surrounding sediment. Bivalves in the nodules are generally well preserved. These are mostly disarticulated and occur at all angles to the bedding. Ostracods too are mostly single valves and have commonly been silicified.

Section 4.5 Origin of Devonian Calcareous Nodules

Gründel and Rösler (1963) divide Upper Devonian nodular beds into the following:-

Knollenschiefer (English translation: shales with nodules)
Carbonate content less than 50% . Normally isolated or partly continuous nodules, parallel to bedding.

Knollenkalk (nodular limestone)
Carbonate content greater than 50%. Nodules parallel to bedding and often forming nodular bands.

Flaserkalk (Flaser limestone)
Carbonate content about 85-95%. Clay material forms thin streaks (Flasers) between the nodules. Nodules not arranged parallel to bedding, often good limestones.

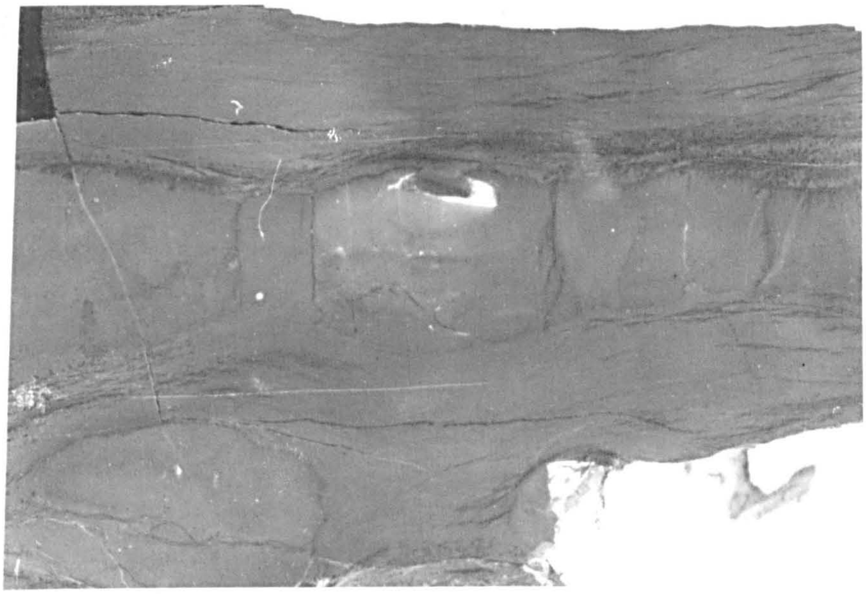
Transitions occur between all three types and the general name applied to this lithology is Kalkknollengesteine. The term Kramenzelkalk is often used but this is really a field term and refers to the weathered appearance of these nodular rocks. The nodules weather more quickly than the shales and the rock appears to be full of holes (Fig. 4.10).

There are six main theories for the origin of calcareous nodules, a) tectonic b) sediment movement c) compaction d) sub-solution e) syngenetic/concretionary f) diagenetic migration of $CaCO_3$.

a) Tectonic origin. Born (1921), Falke (1949) and Weber (1965) considered that continuous beds of limestone were broken up by tectonics to give the nodules. Born (1921) accepted a sedimentary contribution

Fig. 4.9 Fractured geopetal cavity in calcareous nodule. Fracture took place before the formation of calcite. _Clymenia_ Stufe, Upper Famennian. Railway cutting, Innerstetal Reservoir, N.W. Harz. Polished surface S 23052. Scale bar in cm.

Fig. 4.10 Badly weathered nodular or Flaser limestone with typical 'honeycomb~~e~~' appearance known as Kramenzelkalk. Famennian. Kattensiepen, near Warstein, Sauerland. Scale bar = 20 cm.



towards the nodularity but Falke and Weber believed that tectonics alone is the main factor. Shales with nodules can easily be modified by tectonics and indeed one can find nodules elongated or rotated in the cleavage direction. However, the occurrence of nodules in sediments in an area of only minor structural deformation (N.W.Harz) shows that a tectonic origin is untenable.

b) Nodules formed by sediment movement. Greiling (1967) for the Upper Devonian of Frankenwald and Jurgan (1969) for the Alpine Jurassic, suggested that the nodules formed by an unmixing of sediment as a result of catastrophic downslope sliding and movement caused by earth-tremors. Although movement of slope sediments has taken place in the Harz Mountains (chapter 5), the nodules had clearly already formed.

c) Compaction origin. McCrossan (1957) and Wobber (1967) described nodular limestones and isolated nodules and suggested that these were formed by compaction of originally continuous limestone bands. McCrossan envisaged a lateral spreading of the calcareous sediment due to a lower plasticity of this compared with the surrounding clay-rich sediment. A slightly irregular depositional surface or slight lateral variations in the density of the mud are assumed to have initiated the process. Such a hypothesis is clearly not applicable to the calcareous nodules of the Upper Devonian in Germany. The presence of undisturbed trace fossils shows that no movement of sediment was involved.

d) Subsolution. Hollmann (1962) suggested that calcareous nodules were the remnants of extensive subsolution, and that these clasts were transported and embedded in clays to give the typical shales with nodules lithology. Subsolution is of minor importance in the Devonian sediments and cannot be considered to have formed parallel bands of isolated nodules.

e) Syngenetic/concretionary origin. Schindewolf (1921 and 1923) suggested that the bacterial decay of soft bodied organisms would liberate ammonia which would react with seawater to precipitate $CaCO_3$ and form the nodules. Weeks (1953) and Zangerl (1968) invoked a similar mechanism for concretions and Berner (1968) has shown experimentally how this occurs. This process has certainly operated in the formation of some concretions, but is unlikely to have been important in the case of the Upper Devonian nodules.

f) Diagenetic migration of $CaCO_3$. Various authors have suggested that calcareous nodules form by diagenetic migration of $CaCO_3$ (Illies, 1949; Hallam, 1964; Lucas, 1955), but the only full explanation of how this might occur was given by Gründel and Rösler (1963) from a study of Upper Devonian sediments in Thuringia. They noted that the vertical distance between nodule bands, as well as the size of the nodules, is dependent on the carbonate content of the sediment; the higher the carbonate content, the smaller is the vertical distance between the bands and the smaller are the nodules. These relationships have also been observed in the Harz Mountains and the Rheinisches Schiefergebirge.

Hallam (1964) examining the limestone/shale rhythm of the Blue Lias showed that there was a primary alternation which was modified by diagenetic migration of carbonate the nodular appearance of the limestones and the formation of isolated calcareous nodules.

Gründel and Rösler related nodule formation to the oxidation/reduction zone which is found in most sediment profiles today. Both zones contain bacteria which decompose organic matter. The oxidation zone is normally a light brown colour due to the presence of ferric iron, in contrast to the darker colour of the reduction zone where iron is in the form of disseminated pyrite. H_2S is formed in the reduction zone through bacterial activity, and the lowering of pH as a result of this, causes finely disseminated $CaCO_3$ to dissolve in the pore water. This solution moves upwards through the sediment (possibly through the overburden or by compaction) into the oxidation zone. The pH of the oxidation zone is higher (>8) and oxidizes the S^{2-} ion to SO_4^{2-}. The change in pH lowers the solubility of $CaCO_3$ producing a more concentrated solution and eventually precipitation of $CaCO_3$. Growth of nodules takes place in the oxidation zone until sedimentation, and with it the slow upward movement of the oxidation/reduction zone boundary, causes the nodules to pass into the reduction zone. The size of the nodules and distance apart of the bands is easily explained with this hypothesis. Pore solutions low in $CaCO_3$ will start tp precipitate higher up in the oxidation zone, and will have a longer time to develop before passing into the reduction zone. Hence larger nodules will form, and the distance between nodule bands will be greater. With pore solutions rich in $CaCO_3$ the nodules will begin to form earlier, lower down in the oxidation zone, but they will have less time to develop. In this way small nodules at closely spaced intervals will develop.

The process envisaged by Gründel and Rösler outlined above is certainly the most attractive hypothesis for explaining the formation of the Upper Devonian shales with nodules lithology, and accounts for many of the observed facts. Two criticisms can be made in connection with their hypothesis: 1) why are the nodules not dissolved when they pass into the reduction zone and 2) how do nodules form in red sediments.

With the first criticism, it could be a question of size. The $CaCO_3$ which goes into solution is probably the supersoluble micro-crystalline calcite which is disseminated through the sediment. Perhaps the nodules are too large to be affected by the slightly acidic pore fluids which are present in the reduction zone. No evidence of partial solution of nodules has been observed. Nodules may have gradational sharp contacts with the surrounding shale, depending on the degree of compaction or deformation. Shell material is normally well preserved in the nodules but is missing if it extended out of the nodule into the shale. Solution of skeletal carbonate in the shale has taken place and would also contribute $CaCO_3$ for the formation of the nodules.

The second criticism concerns the reduction zone which is essential to obtain the initial dissolution of $CaCO_3$. Nodules commonly occur in red shales (e.g. <u>Platyclymenia</u> Stufe of the Upper Devonian, griotte of Montagne Noire, Ammonitico Rosso of the Alps) and if one interprets these hematite-rich sediments in the traditional way, that is deposited under oxidising conditions, then presumably there was no reduction zone present in the sediment. This would suggest that the reduction zone is not so important in the formation of the nodules, and a process must be found to explain nodules occurring in shales of different colours reflecting deposition under different redox potentials.

There are very few reports of nodules forming in sediments today. Cores have been collected from most environments but none have shown a sediment unmixing of the type under discussion here. Pantin (1958) described a large diagenetic concretion dredged off New Zealand, which has formed in the last 20,000 years. Its formation is attributed to diagenetic migration of organic or detrital carbonate that was originally distributed through the enclosing sediment. Possibly these diagenetic nodules are more widespread today than is thought, and it is a question of time before these are found *in situ*.

The problem of how the calcareous nodules were formed is far from being solved. However, the new observations on nodules made during this work, namely that of trace fossils, slumped and reworked nodules (described in chapter 5) accord with an early diagenetic origin for the nodules. The exact process of nodule formation is not known, but it must be along the lines proposed by Gründel and Rösler and related to solution and precipitation of $CaCO_3$ within the sediment.

Section 4.6 History of sedimentation

The dominant pelagic fauna in the basinal shales indicates that either the sea was very deep, or the bottom was practically inimical to life. A fauna, typical of agitated well oxygenated shallow seas, as is found in Lower Devonian and Eifelian shales (Rhenish facies) is completely absent. The absence of strong bottom currents is shown by cricoconarids which are rarely orientated.

Most of the fine grained clastic sediment in the deeper water areas between the Schwellen must be derived from the shelf area and continent to the north. The sediments are dominantly shales with a silty lamination, the latter representing deposition from low density suspension currents (or nepheloid layers), superimposed on the constant rain of clay and organic matter. Horizons of dark grey or black laminated shales with pyrite indicate a lack of current activity. Turbidites are locally developed in certain areas and may be carbonate turbidites derived from 'reefs' on the shelf or within the basin, sandstone turbidites derived from the shelf to the north, or immature greywackes (the synorogenic flysch) which are derived from the land ridge to the south.

During early diagenesis, calcareous nodules formed in the shales if the carbonate content was sufficiently high. Compaction led to a reduction in thickness of the shales, of at least 60%. Pressure solution and compaction accentuated nodule/shale boundaries, and locally rotation of nodules and fracturing occurred through tectonic stresses.

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CHAPTER 5

Sedimentology of the Slope Facies

The slope facies is variable lithologically and transitional between that of the Schwellen and basins. Shales with isolated or locally continuous nodules and nodular limestones are the most common lithology but these have commonly been affected by mass-transport downslope and contain slumps, or are reworked as intraformational breccias. It is on the basis of slumped or reworked sediments that the slope facies is differentiated from those in the Schwellen and Becken areas. Sediments of the Schwellen and Becken facies may at times be developed in the slope environment. For example, during the Frasnian, Flaser limestones were more widespread and occur interbedded with shales and nodules.

Slope deposits are well developed on the northwestern side of the West Harz Schwelle, where a number of complete Upper Devonian sections show sediments from various parts of the transition from Schwellen to Becken facies. Localities in the slope facies are Sparenberg and Lautenthal in Innerstetal, south of Langelsheim, and Margaretten Klippen and Hessenweg in Granetal, south of Goslar (locations shown in Fig. 2.6, p.29). In the Rheinisches Schiefergebirge, outcrops of these deposits were examined at Drewer Provincial Steinbruch, near Warstein, where the sediments accumulated on the flanks of the Schwelle formed by the Brilon Massive limestone, and at Öse where the sediments accumulated on the slope of the Arnsberg Schwelle (also underlain by Massive limestone), (locations shown in Fig. 3.4, p.50). Other outcrops, where slumped or reworked sediments are exposed, were examined eleswhere in the Harz and in the Lahn Syncline (central part of the Rheinisches Schiefergebirge).

Section 5.1 Slope sediments

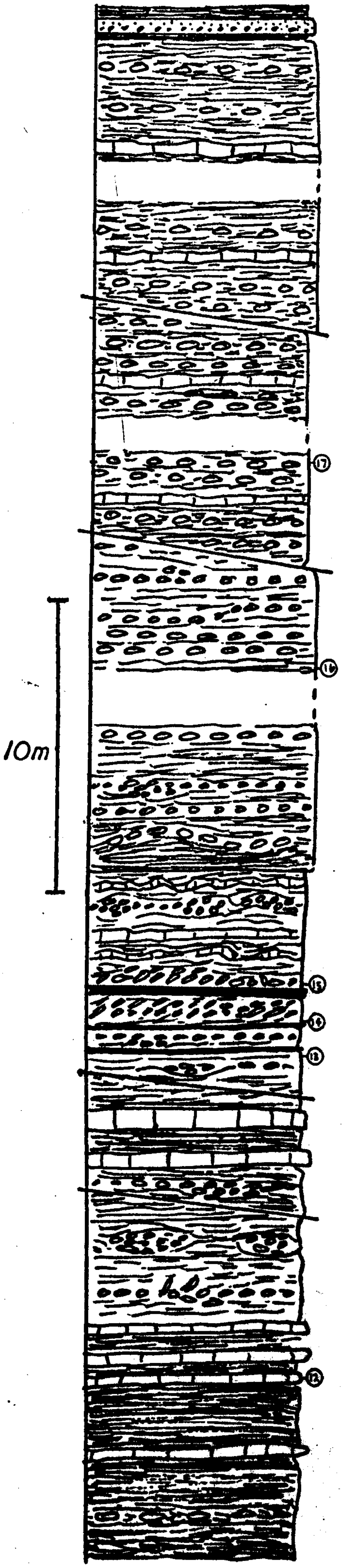
The dominant lithology in slope successions, shales with nodules and nodular limestones generally constitute 50 to 80% of the succession. Flaser limestones are less common and may comprise about 10%. Typical successions in slope sediments are shown in Figs. 5.1 and 5.2. The following sub-facies are described, 5.1.1 Flaser limestones, 5.1.2 nodular limestones, and 5.1.3 shales, and shales with nodules. The sub-facies are effectively based on the carbonate content, and all of these may be involved in slumping and reworking (described in section 5.2).

5.1.1 Flaser limestones

Flaser limestones (i.e. fine grained limestones with diagenetic and tectonic shale streaks) are uncommon in slope sequences. They are similar to Flaser limestone from the Schwellen (chapter 3). Once composed of fine grained calcite with randomly orientated microfossils, the sediments are now microsparites, containing coarser calcite crystals (originally skeletal fragments ?). Some patches of coarser neomorphic calcite occur, with crystals up to 100μ in size, but these mosaics are less common and smaller than those in the Schwellen limestones. Pyrite is commonly present, either as cubes, finely disseminated particles or aggregates. Bioturbation is common and burrows are filled by equant calcite. Hardgrounds, sheet cavities and sedimentary dykes have not been found.

Some Flaser limestones contain bedding types similar to those seen in Schwellen limestones (graded bands rich in silt and clay, carbonate laminations and thin shell bands). Graded carbonate laminae are present in some limestones and these could be derived from the Schwellen.

Fig. 5.1 Upper Devonian succession in slope facies at Lautenthal (Harz)



Carboniferous shale

20 m Grey shale + nodule bands (*Clymenia* and *Wocklumeria* Stufen)

Fault

15 m Brownish-grey shales with many calcareous nodule bands (*Platyclymenia* Stufe)

Fault

20 m Grey silty shales with few nodule bands (*Cheiloceras* Stufe)

7 m Slumped nodule bands and reworked nodules in grey shale. Three Kellwasser horizons (Frasnian)

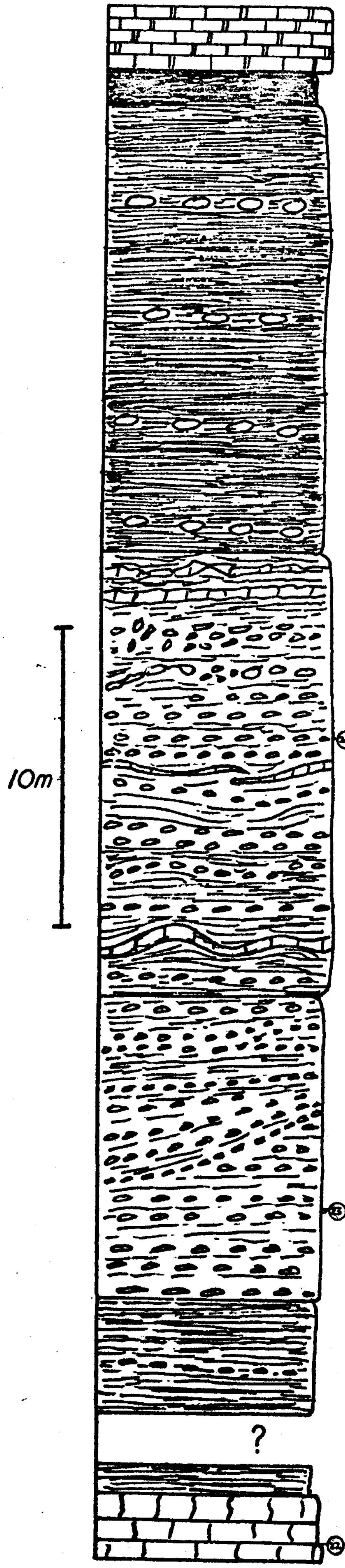
Fault

5 m Grey shales with limestone bands.

Fault

15 m Dark grey shales with limestone beds and pockets of reworked nodules (Givetian)

Fig. 5.2 Upper Devonian in slope facies at Margaretten Klippen (Harz)



	Carboniferous Cherts
40 cm	Alaunschiefer
15 m	Black or dark grey shales with a few nodule bands (*Clymenia* and *Wocklumeria* Stufen)
15 m	Red and green shales with nodule bands and nodular limestones. Slumping common. (*Platyclymenia* Stufe)
10 m	Bands of calcareous nodules in grey shale. Slumping present. (*Cheiloceras* Stufe)
5 m	Grey shales + few nodules.
	No exposure
+1 m	Flaser limestone (Frasnian)

5.1.2 Nodular limestones

The term nodular limestone is used here to refer to continuous limestone bands with irregular, hummocky surfaces, which are interbedded with shales. The origin of the nodular limestones is considered to be a combination of primary changes in sedimentation and diagenetic migration of carbonate (p.219 cf. Hallam, 1964). At Drewer and Öse in the Rheinisches Schiefergebirge nodular limestones with interbedded black shales constitute most of the Famennian. At slope localities in the Harz Mountains, nodular limestones are developed at some horizons.

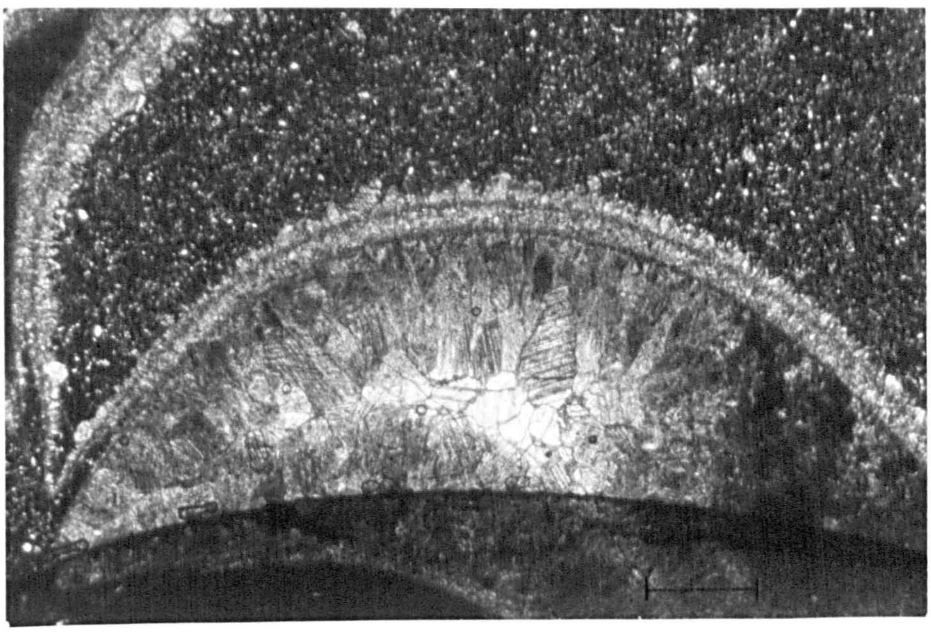
The limestone matrix is a combination of microsparite and micrite, with larger crystals (probably skeletal debris) (c.f. Flaser limestones). Silt is present in variable quantities from 1 to 10%, but in the interbedded shale the quartz content may reach 25%. The shales between the limestones are black and rich in fine grained pyrite. The limestones commonly have a mottled appearance, through varying amounts of carbonate, some of which may be due to bioturbation. Definite burrow structures are common, 2 or 3 mm across, and are filled by slightly coarser calcite. Rarely burrows can be seen extending upwards into the overlying shales. The transition from limestone to shale is gradational where not affected by pressure solution. Where pressure solution has occurred at the junction, skeletal material within limestone is truncated. Fossils are noticeably absent in the shales but are common in limestones. Goniatites present between limestone bands are rare, but where they do occur, calcareous nodules have formed around them. The goniatite shell has been dissolved where it extended outside the nodule.

The limestones are composed of non-ferroan calcite whereas the carbonate in shales is ferroan dolomite.

The thickness of individual nodular limestones and their intervening shales at Drewer is remarkably constant (limestones 5 to 8 cm; shales 1 to 3 cm). The lateral passage of nodular limestones into discrete nodules and the development of nodules around goniatites between nodular limestones indicate that there has been much diagenetic mobilization of carbonate.

Fig. 5.3 Nodular limestones interbedded with black shales. Clymenia Stufe, Upper Famennian. Drewer, near Warstein, Sauerland. Length of hammer = 35 cm. (Text, p. 227).

Fig. 5.4 Cavity within a clymeniid shell filled first with radiaxial calcite full of inclusions and later with clear sparite. Fibrous calcite has also grown from the shell wall into the sediment. Clymenia Stufe, Upper Devonian. Drewer, near Warstein, Sauerland. Thin section S 23061. Scale bar = 1 mm.



Late diagenetic and tectonic pressure solution has affected these sediments by enhancing the limestone shale contact and 'wrapping' the shale around the nodular parts. Flasers may cut the nodular limestones, but stylolites are not so common as in the Schwellen limestones.

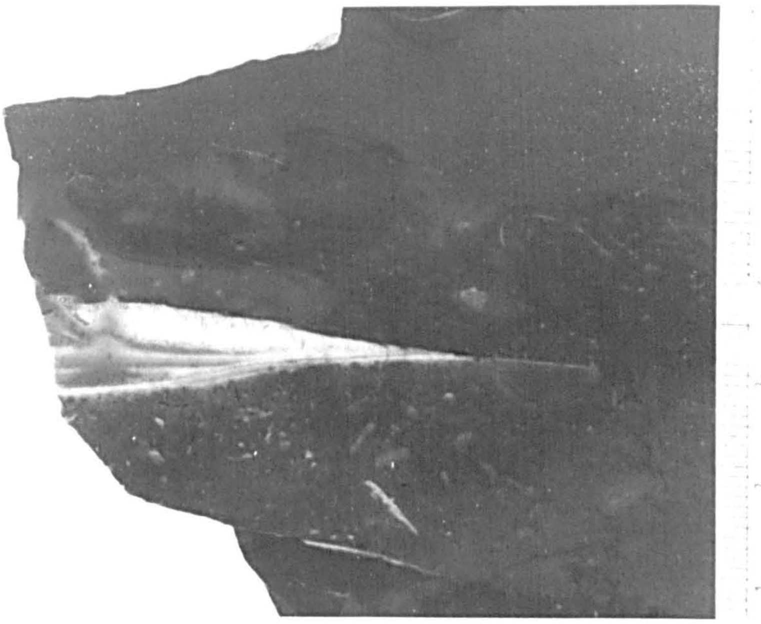
Sheet cracks and sedimentary dykes were not observed in the nodular limestones. Neomorphism of the matrix to very coarse mosaics is also absent. However, clymeniids are present and provided cavities in which calcite was precipitated in a similar way to the sheet cracks of the Schwellen limestones (p. 87).

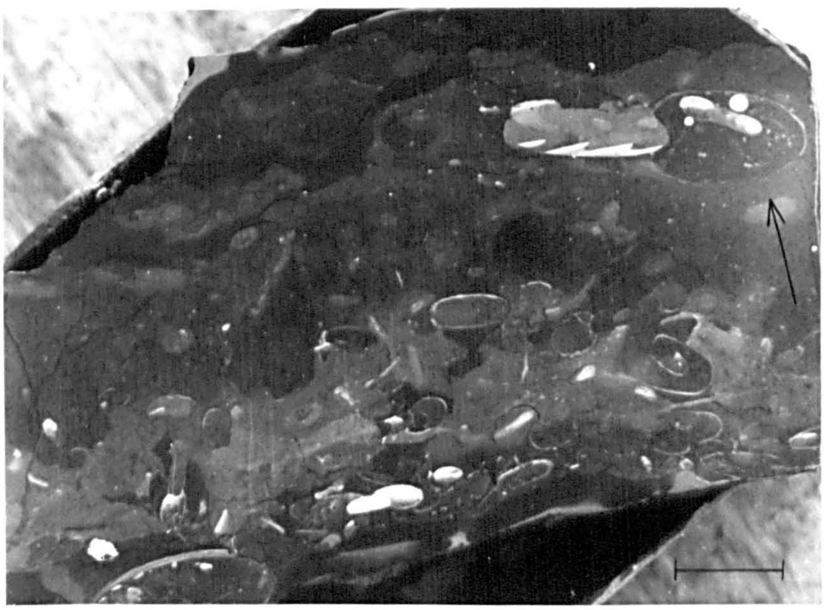
Shells of clymeniids may occur up to 10 cm in diameter. Larger shells are found orientated parallel to the bedding, whilst smaller shells may be at all angles. The clymeniids are filled by internal sediments and coarsely crystalline carbonate. Internal sediments are commonly white to light-grey microsparites occurring as thin graded laminae. The cavity remaining after internal sedimentation, was lined by radiaxial fibrous calcite, and then later filled by equant sparite (c.f. sheet cavities in Schwellen limestones, p. 87). Fibrous calcite is commonly partially or completely replaced by later calcite crystals. Within inner whorls of some clymeniid shells, fibrous calcite crystals are less altered and turbid with inclusions compared with the nearby equant sparite. Radiaxial fibrous calcite has grown from the shell wall of the clymeniid which is still distinguishable (Fig. 5.4). Fibrous crystals have also grown out from the shell into or replacing the host sediment. Where there is no marginal fibrous calcite, the shell wall is no longer visible and sparite is in contact with the host sediment. This suggests that ~~either~~ the shell wall of the clymeniid dissolved before the sparite was precipitated (Bathurst, 1964).

One clymeniid, partly filled with light grey internal sediment, was found with a small patch of this sediment in the host limestone above the shell (Fig. 5.5). The lamination within the shell is deformed. This is interpreted as a fluid - escape structure, similar to those described by Zangerl et al (1969). A sudden release of gas or water carried some of the internal sediment out of the clymeniid.

Fig. 5.5 Section across a clymeniid filled by internal sediment and calcite. A small patch of internal sediment occurs above the shell and bands of internal sediment are deformed within the shell. This is interpreted as a fluid-escape structure. Burrow-fills occur in the nodular limestone which is gradational into the black pyritic shale above. Clymenia Stufe, Upper Devonian. Drewer, near Warstein, Sauerland. Polished surface S 23059. Scale bar in cm. (Text, p. 230).

Fig. 5.6 Mottled nodular limestone rich in ammonoids. Geopetal sediment shows wide variations in angle of deposition. Clymeniid upper right underwent a phase of solution after deposition of white internal sediment on the septa (see text, p. 233). Clymenia Stufe, Upper Devonian. Drewer, near Warstein, Sauerland. Polished surface S 23062. Scale bar = 2 cm.





This implies that neither the host sediment nor the internal sediment was lithified at the time.

Where a clymeniid extended outside the nodule the shell is absent. Shells in the limestone may have been partly dissolved. Clymeniids occur with white internal sediment partly filling the positions once occupied by the septa, forming boat shaped lenses (Fig. 5.6). A thin layer of calcite was precipitated over the internal sediment. A period of solution occurred and the septa were dissolved. The remaining cavity was later filled by a light grey carbonate, and the final void was occluded by fibrous and equant calcites. The first internal sediment must have been cemented before the final filling of the shell since the lenses of this sediment retained their shape after solution of the septal walls, and the lenses have not been affected by the later light grey internal sediment.

Geopetal structures in the nodular limestones can be used to determine the original depositional dip. In the field, these structures are not common and at Drewer some 20 observations were made, varying from parallel to the bedding to a maximum deviation of 10° (Fig. 5.8). However, as can be seen from Fig. 5.6, the angles are not consistent even in one hand specimen. Most measurements made in the Harz (12 readings) do not deviate significantly from the bedding (Fig. 5.9), and a maximum angle of 5° was recorded.

5.1.3 Shales and shales with nodules

Calcareous nodules in silty shales is the most common lithology in the slope facies. Nodules tend to occur in bands commonly regularly spaced less than 10 cm apart. The nodules themselves are small (1 or 2 cm across) but they commonly coalesce forming thin nodular limestones. Nodule boundaries are gradational unless affected by pressure solution. Generally, nodules in the slope region are smaller and bands are closer together than those in the basins. Otherwise they show exactly the same features. The origin of the calcareous nodules is discussed in chapter 4 (p.215).

The shales in the slope region are also very similar to those in the basin (chapter 4). Silty lamination is commonly developed.

Fig. 5.7.

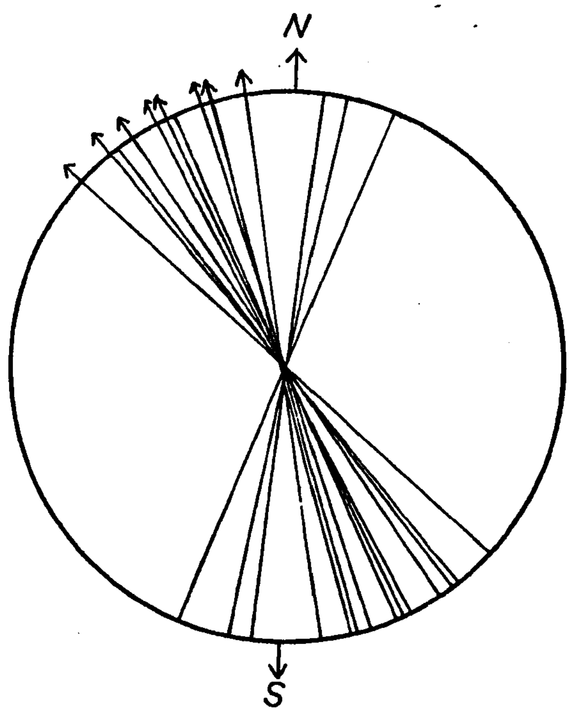


Fig. 5.8.

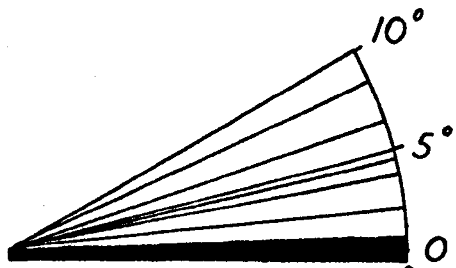


Fig. 5.9

Fig. 5.7. Directions of transport measured from slumped nodule bands (with arrows) and reworked nodules. (Directions are approximate since most sections are two-dimensional). Sparenberg and Lautenthal (Harz).

Figs. 5.8 and 5.9. Angles of slope measured from geopetal structures.
Fig. 5.8 from Drewer (Sauerland), 13 measurements parallel to bedding.
Fig. 5.9 from Sparenberg, Lautenthal and Margaretten Klippen (Harz), 7 measurements parallel to bedding.

Some very calcareous shales occur between nodular limestones and have a fine clay-rich and clay-poor lamination. Muscovite, chlorite and silt are present and the carbonate is ferroan dolomite. Calcite is restricted to large skeletal fragments. Carbonaceous shales are rarely developed apart from the Kellwasser Horizon (a thin organic rich shale or limestone developed over large areas during the Frasnian). Bioturbation is present and has locally disturbed the silty lamination.

Section 5.2 Slumped and Reworked Sediments

5.2.1 Reworked sediments and breccias

Slumped and reworked deposits serve to distinguish the slope sediment from those of the rises and basins. Some Flaser limestones in the slope region contain thin horizons, a few millimetres thick, of angular limestone fragments. The fragments may differ from each other in lithology and fossil abundance. Larger limestone clasts in shaley horizons occur at many localities. At Sparenberg (Innerstetal, Harz) for example, Stoppel and Zscheked (1963) have described blocks of Frasnian Flaser limestone, 3 m across in Frasnian grey shales (Fig. 5.13). Also blocks of reworked limestone occur within the Saltern Cove Goniatite Bed, in S.Devon, an analogous situation (Straaten and Tucker, 1971, in press, manuscript appended).

Horizons of reworked calcareous nodules, shale intraclasts and fragments of limestone beds exhibit imbrication of the type shown in Fig. 5.10. Transport directions were also obtained from slump folds, and all measurements indicate derivation from the south or south-east, the direction of the West Harz Schwelle (Fig. 5.7). Derived nodules are undeformed by compaction, indicating lithification before reworking. Shale clasts and fragments of the Kellwasser Horizon are on the other hand distorted. Isolated pockets of reworked nodules occur at Lautenthal (Fig. 5.10). In the Lahn Syncline, reworked blocks of Middle Devonian limestone are recorded from the Upper Devonian shales (Bender, 1965).

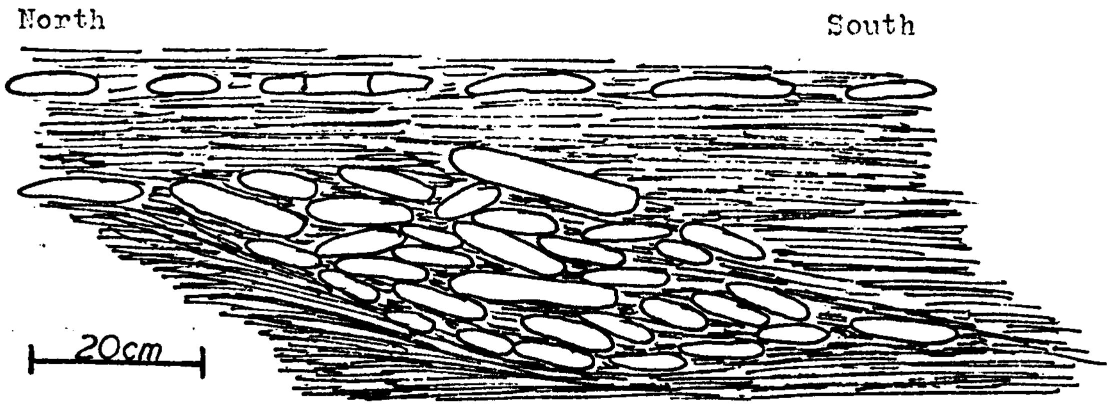


Fig. 5.10.

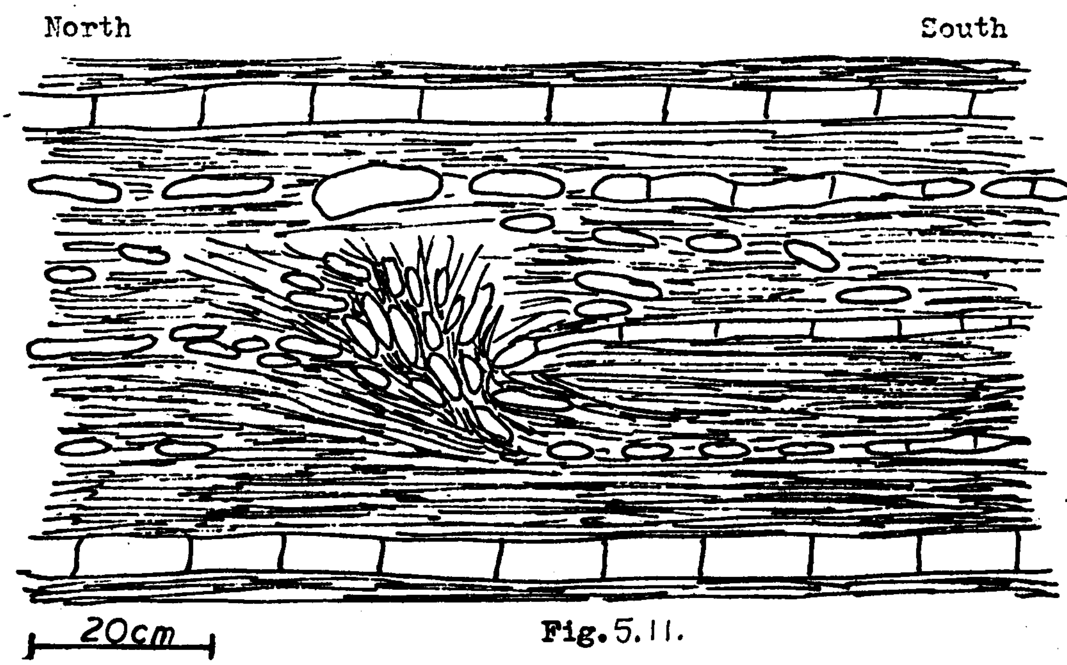


Fig. 5.11.

Fig. 5.10. Pocket of reworked calcareous nodules.
Givetian, Lautenthal (Harz).

Fig. 5.11. Disturbed nodule bands through slumping or fluid escape.
Frasnian, Lautenthal (Harz).

Fig. 5.12.

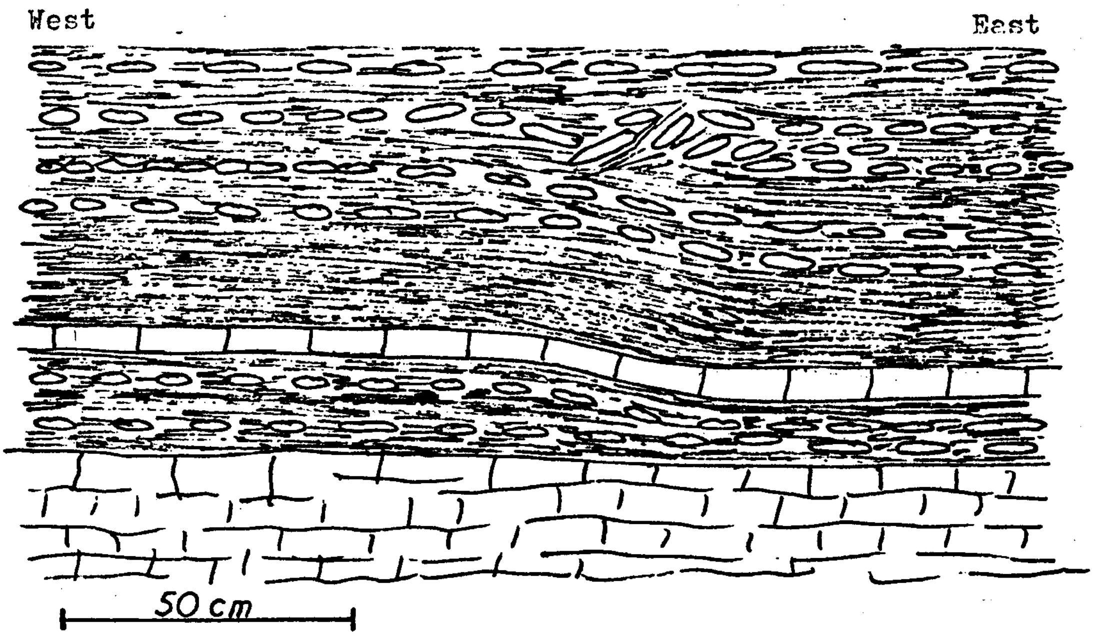


Fig. 5.13.

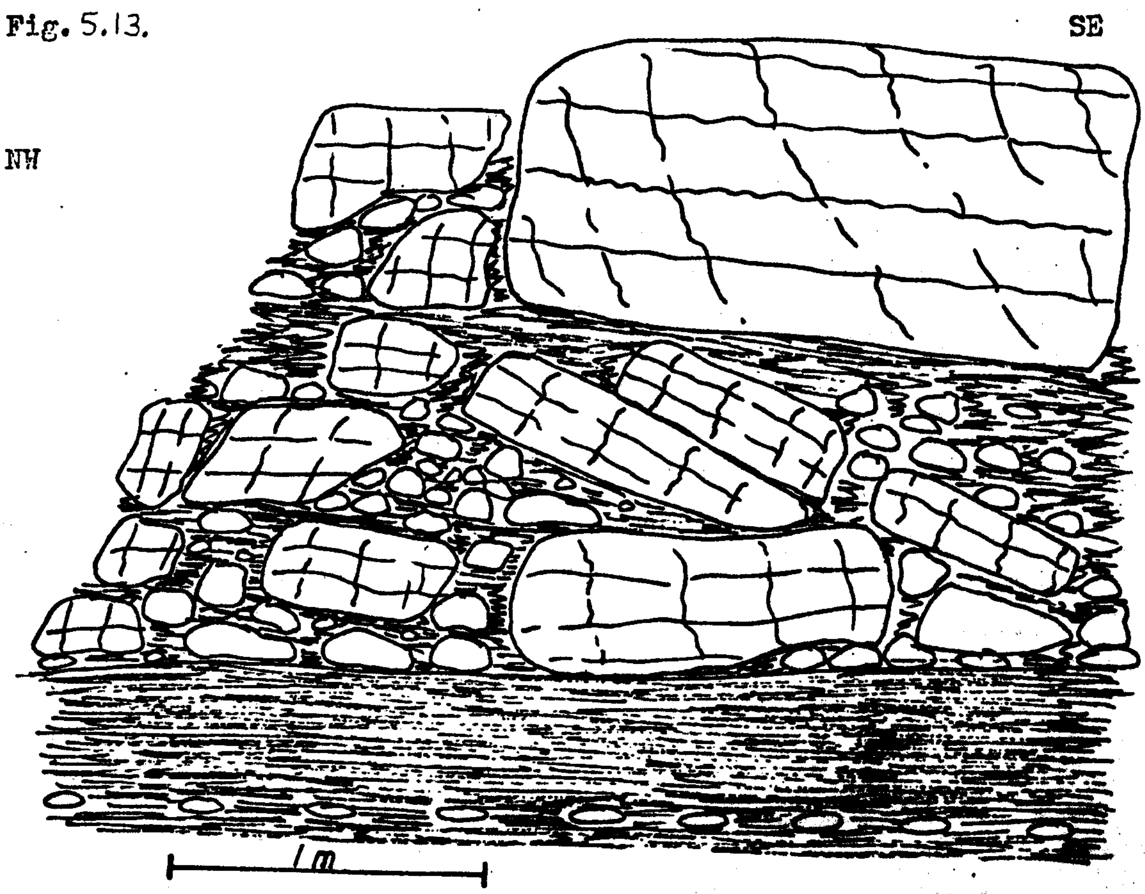


Fig. 5.12. Calcareous nodule bands folded and disrupted by movement of sediment. Platyclymenia Stufe, Margaretten Klippen (Harz).

Fig. 5.13. Slumped and reworked blocks of Givetian/lower Frasnian Flaser limestone in Frasnian shales. Sperenberg (Harz).

Fig. 5.14.

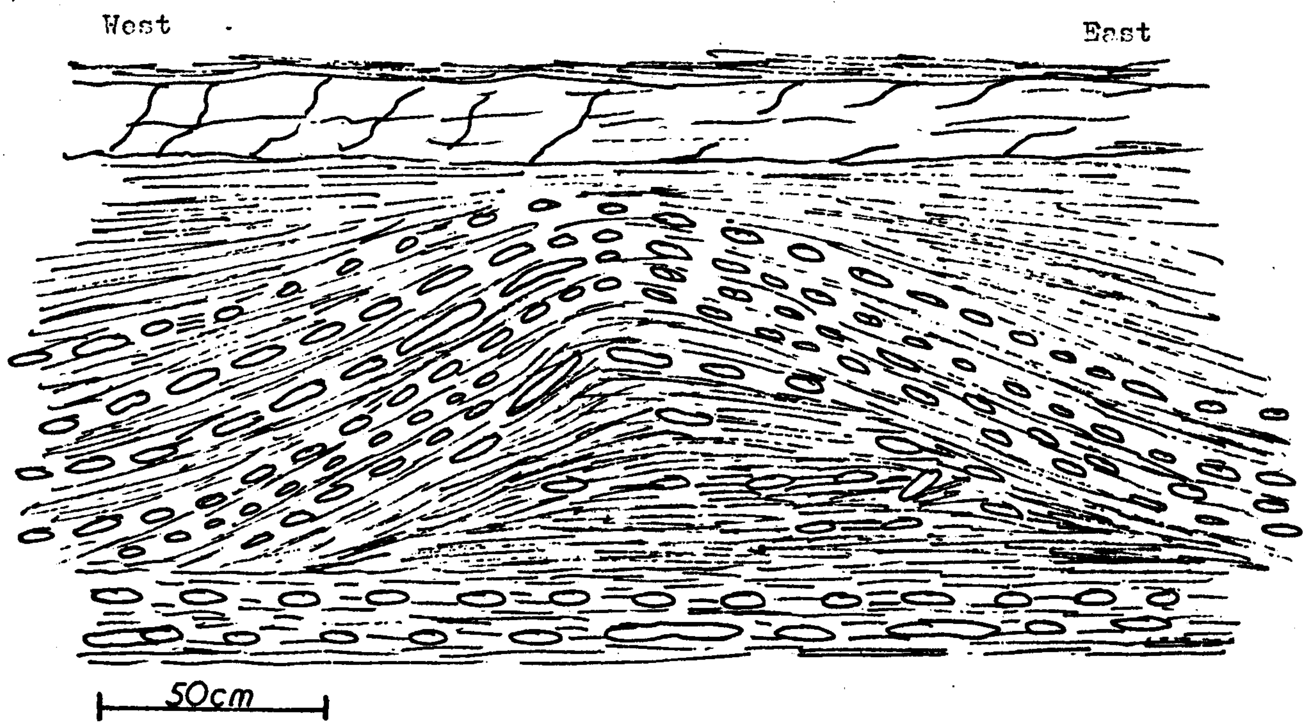


Fig.5.15.

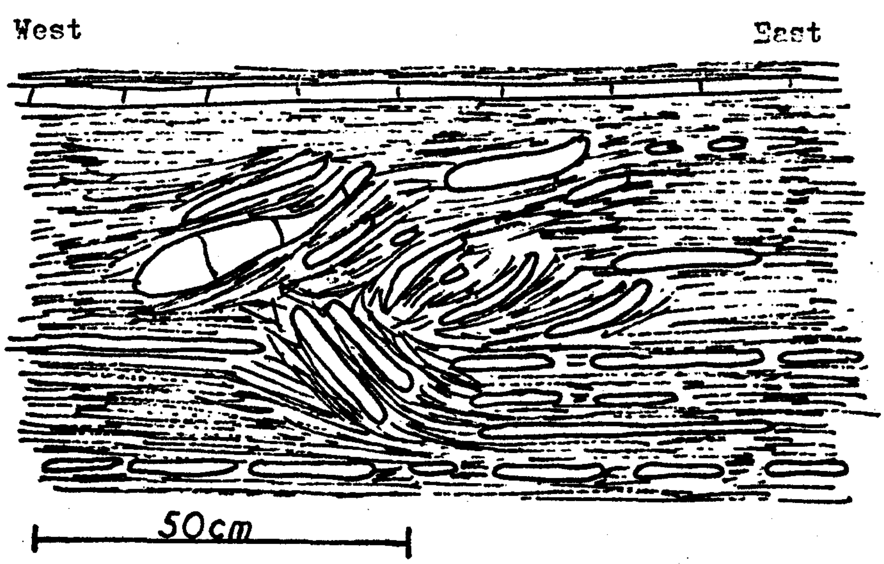


Fig.5.14. Fold produced by slumping. Platyclymenia Stufe, Margaretten Klippen (Harz).

Fig.5.15. Disturbed nodule bands through slumping or fluid escape. Platyclymenia Stufe, Margaretten Klippen (Harz).

In the Montagne Noire and Pyrenees, breccias occur at a number of horizons and form beds up to 10 m thick. Large blocks of Flaser limestone may be reworked along with calcareous nodules and shale fragments. Other breccias from the griotte in the Montagne Noire consist of small fragments of limestone, about a centimetre in diameter, embedded in hematitic shale (Fig. 5.17). The palaeogeography of the Upper Devonian in these regions is unknown thus it is not possible yet to relate the breccias to identified Schwellen.

Fault scarp breccias were developed during the Upper Devonian in the Rheinisches Schiefergebirge. For example, the Schlagwasser-breccia, outcropping near Warstein (Sauerland), is a coarse conglomerate containing a variety of limestone blocks with conodonts of Middle Devonian to upper Famennian age. Schmidt and Plessmann (1961) considered this deposit a tectonic breccia, but it has recently been reinterpreted as having formed through contemporaneous fault movements (Stasche, 1968).

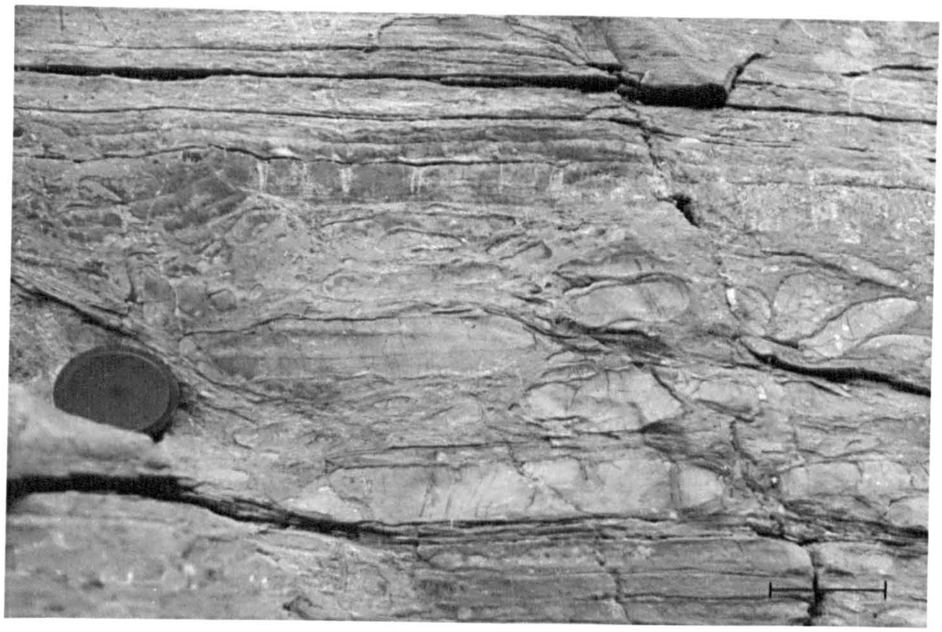
5.2.2 Sedimentary slumping

Small displacements of sediment by downslope slip have produced non-parallel bands of calcareous nodules (Fig. 5.18). The plane of movement (glide plane) cannot usually be seen unless the slide or toreva block has cut down and thus truncates undisturbed nodule bands (Fig. 5.19). Individual bands of nodules may have crumpled and be slightly folded (Fig. 5.20) but elsewhere slump folds with an amplitude of a metre or less have formed (Fig. 5.14) and local underthrusting of nodular limestones has resulted (Fig. 5.21). Alternatively, nodular bands may be locally broken up, leaving the nodules irregularly arranged in the shale (Fig. 5.23). Beds above and below slumped horizons may be planar-bedded, suggesting that only the top few metres of sediment has moved at any one time. However, evidence has not been found showing that erosion of the folds took place, and movement may therefore have occurred under a cover of sediment.

Lamination in slumped shales is distorted or absent altogether The absence of silty laminations itself can often be used as an indication that slumping has taken place. This is used in Straaten

Fig. 5.16 Reworked nodules in grey shale, overlain by a continuous limestone band. Upper Givetian. Lautenthal, N.W. Harz. Scale bar = 5 cm.

Fig. 5.17 Breccia consisting of limestone clasts embedded in hematitic shale. Frasnian. Fontaine de la Santé, Caunes Minervois, Montagne Noire. Peel S 22962. Scale bar = 1 mm.



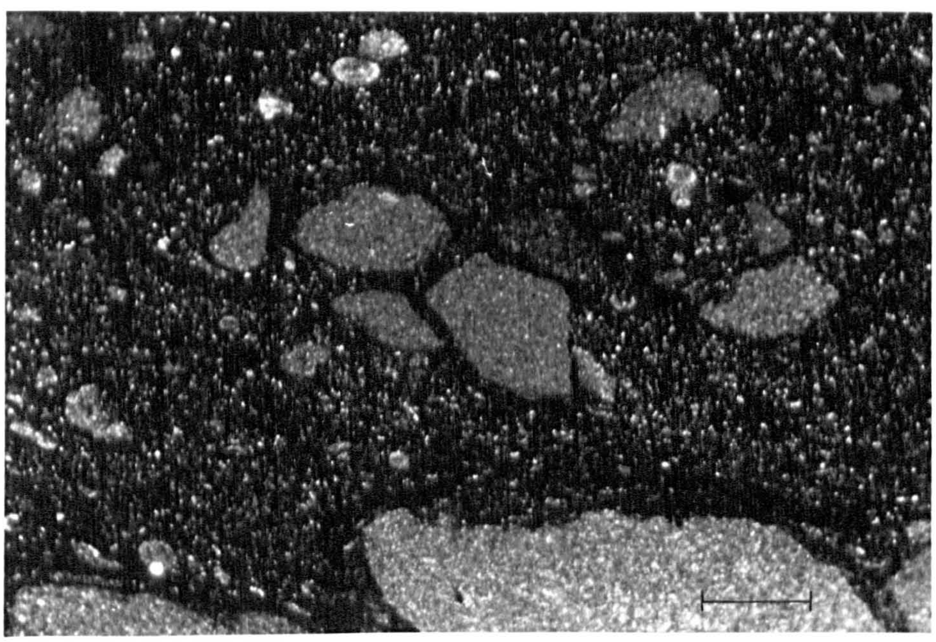


Fig. 5.18 Bands of calcareous nodules with irregularities caused by movement of sediment. Cheiloceras Stufe, Lower Famennian. Margaretten Klippen, N.W. Harz. Length of hammer = 35 cm.

Fig. 5.19 Slumped nodule bands with those in the central part cutting across earlier bands. Upper Givetian. Sparenberg, N.W. Harz. Length of hammer = 35 cm.

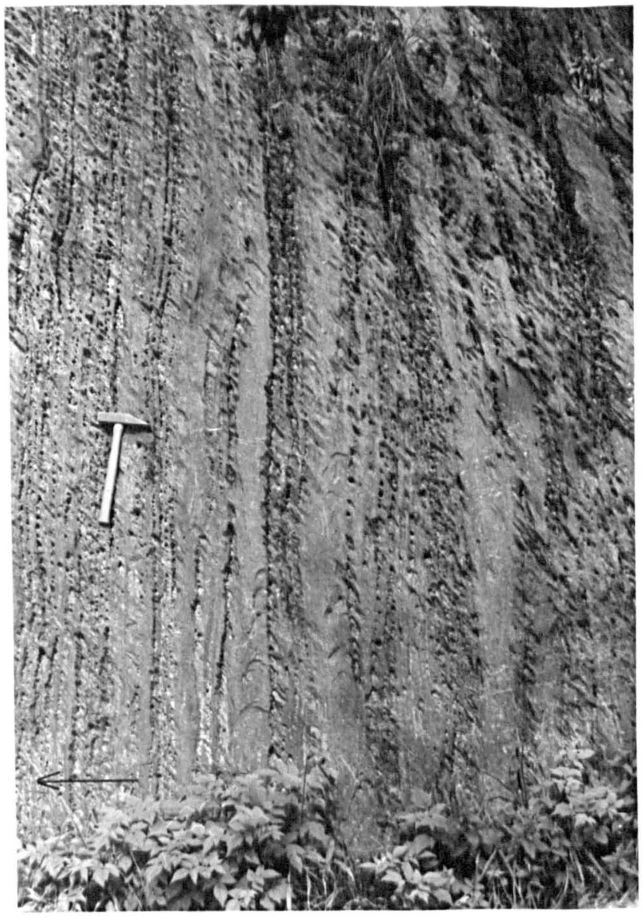
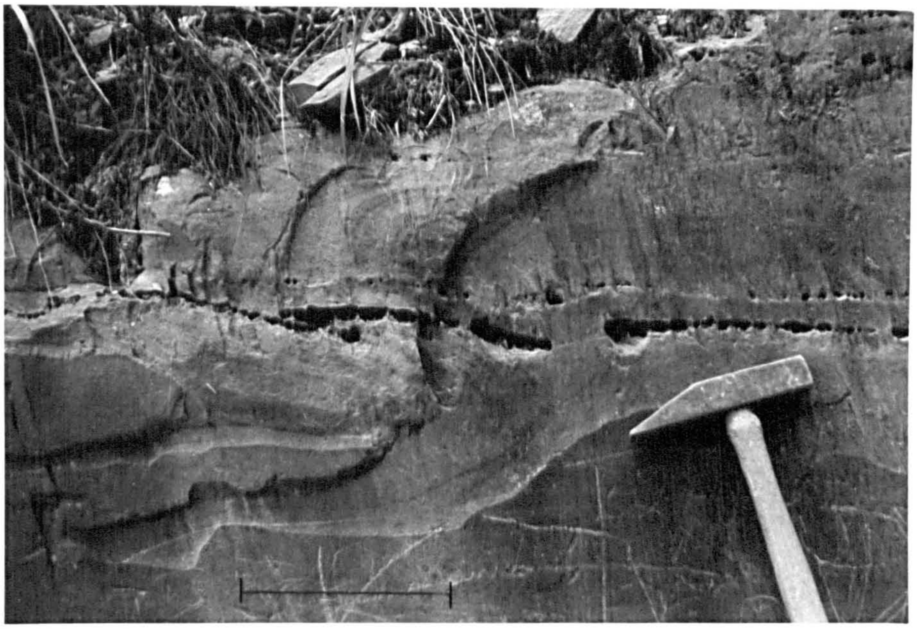


Fig. 5.20 Disturbed nodule bands, some slightly folded, others broken up. _Platyclymenia_ Stufe, Upper Devonian. Weilburg, Lahn Syncline. Scale bar = 10 cm.

Fig. 5.21 Folded and thrusted nodular limestone bands. Broken up nodular limestones also present. Lowest limestone band is planar and undisturbed. _Platyclymenia_ Stufe, Upper Devonian. Margaretten Klippen, N.W. Harz. Scale bar = 20 cm.

Fig. 5.22 Thin band of calcareous nodules which has been thrust over itself during downslope movement of sediment, overlain by continuous undisturbed band of nodules. *Cheiloceras* Stufe, Lower Famennian. Margaretten Klippen, N.W. Harz. Scale bar = 20 cm.

Fig. 5.23 Nodular limestone partly broken up at the crest of the fold. Limestones above and below are planar. *Platyclymenia* Stufe, Upper Devonian. Margaretten Klippen, N.W. Harz. Scale bar = 20 cm.



and Tucker (1971, in press) as additional evidence for the slumped origin of the Saltern Cove Goniatite Bed, S. Devon.

Large scale slumping is present at Drewer where some 50 m of sediment slumped down the flank of the Brilon 'reef' Schwelle during the lowest Carboniferous (Fig. 5.24). Slump folds are developed in the lower part of the moved block in upper Famennian strata (Figs. 5.25 and 5.26). In the upper part two black shale horizons, the Hangenberg-Schiefer and Alaun-Schiefer, have acted as incompetant layers resulting in décollement and wedge out up the quarry face. Undisturbed Carboniferous cherts overlie the slumped beds. The direction of slumping is from the east, where the Schwellen was at the time.

In some nodular limestones (seen at Lautenthal and Rabenklippe in the Harz, and Warstein in the Sauerland) very local disruption of bedding has occurred with fragments of limestone forced upwards (Figs. 5.11 ... 5.27). The broken up part of the sediment in Fig. 5.27 is up to 30 cm high and has the appearance of a sharp fold. It is overlain by a planar nodular limestone. Structures of this type are interpreted as the result of a sudden escape of porewater.

Movement of sediment down submarine slopes has been described from many geological situations, involving various rock types (e.g. Jones, 1940; Kuenen, 1949; Grant-Mackie and Lowry, 1964). Movement occurs when the shear strength of the sediment is exceeded by the component of the shear stress acting downslope. Shear strength of the sediment is determined by many factors including composition, grain size, and rate of sedimentation. Lewis (1971) reports failure of modern sandy-silt shelf sediments off New Zealand at slopes of only 1°. Steep slopes are not a prerequisite of slumping. It is likely that the slope from Schwelle to basin was steeper in some parts than others, and slumping on the steeper parts could cause movement lower down in less steep areas.

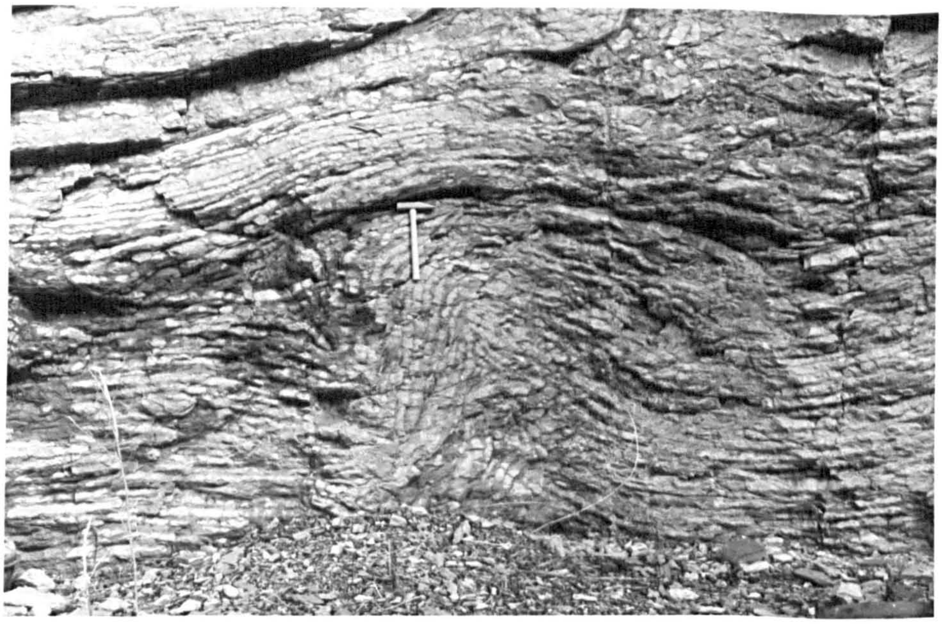
The initiation of slumping is often attributed to earthquakes (Morgenstern, 1967; Lewis, 1971) and this may well have been the cause of slumping in the Upper Devonian slope sediments, perhaps through faulting along the flanks of the Schwellen. Other mechanisms

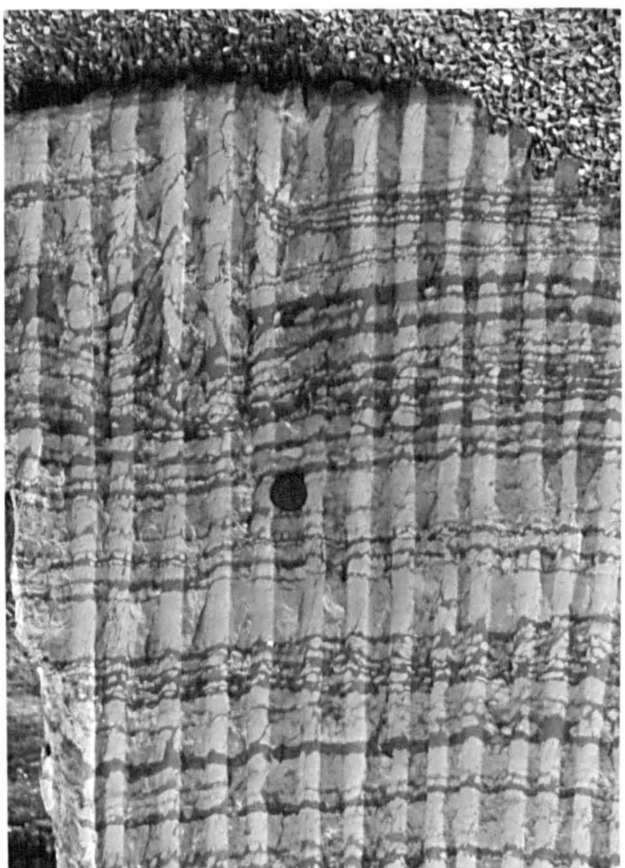
Fig. 5.24 Large scale slumping of sediment. The prominent black shale band (Hangenberg Schiefer) is cut out higher up the quarry face. Slump folds are developed below this in upper Famennain nodular limestones. Carboniferous cherts at the top of the quarry are not involved in the slumping. Drewer, near Warstein, Sauerland.

Fig. 5.25 Slump folds in Upper Devonian nodular limestones Detail of quarry face above; Hangenberg Schiefer in upper part of photograph. Drewer, near Warstein, Sauerland. Scale bar = 1 m.

Fig. 5.26 Slump fold in Upper Devonian nodular limestones. Detail of quarry face in Fig. 5.24. Drewer, near Warstein, Sauerland. Length of hammer = 35 cm.

Fig. 5.27 Nodular limestone with local area of disturbance, probably formed by fluid escape. Upper Famennian. Kattensiepen, near Warstein, Sauerland. Scale bar = 20 cm.





are high sedimentation rates, storms and hurricanes which lead to instability and then slumping.

Section 5.3 Fauna of the Slope Sediments

The fossil abundance of the slope sediments is variable (some localities providing many fossils, others are barren). In the Harz sediments, microfossils are most common, cricoconarids in the upper Givetian and Frasnian, ostracods in the Upper Devonian, and conodonts throughout the Devonian. The ostracods, 1-2 mm long, are well preserved in calcareous sediments and may be articulated or present as single valves. Cricoconarids usually have fibrous overgrowths (p.103).

Nodular limestones at Drewer contain a rich and diverse fauna (goniatites, clymeniids, orthocones, trilobites, brachiopods, thin-shelled bivalves, gastropods, solitary corals and crinoid fragments). Brachiopods, crinoids and corals retain their original microstructure. The molluscan fauna is preserved either as paramorphic replacements (retaining the original microstructure) or as sparite pseudomorphs.

Section 5.4 History of Sedimentation

The slope sediments are very variable in lithology (dependent on the amount of carbonate present) and rock-types characteristic of both Schwellen and Becken facies occur. Generally, shales with nodules and nodular limestones are most commonly developed. The fine grained clastic sediment forming the shales would have been derived from the continent to the north. The carbonate material in Flaser limestones, nodular limestones and nodules, must have the same source as the Schwellen limestones, i.e. probably biogenic in origin.

Slumped and reworked sediments, the distinguishing feature of the slope region, have originated through instability on the slope. Variations in the steepness of the slope must have existed and these may be reflected in the different scales of slumping. Slope angles of less than 5^o are suggested from geopetal structures, and in many cases the slope was probably little more than a degree. Taking 24 km as the distance from the West Harz Schwelle (Aeketal) to the basin (Innerstetal) (distance today is 12 km, shortening factor of 50% used, Wunderlich, 1965), a depth of 400 m is obtained for a slope of 1^o and 2100 m for a slope of 5^o. The Schwellen were probably situated at a maximum depth of 300 m, and so the figures for 1^o and 5^o give basinal depths of 700 and 2400 m. This compares reasonably well with the depth of 1000 m suggested by Clarkson (1967) for the basin.

Current activity in the slope region is indicated by the presence of graded units and shell bands, and since these are less common than in the Schwellen limestones, there was probably less current activity on the slope. Turbulence would be associated with the slumping however, and has given rise to reworked nodules and shales.

Early diagenesis in the slope sediments has been very effective and led to the reorganisation of much of the carbonate material into calcareous nodules. Nodular limestones were also affected by diagenetic migration of carbonate and the interbedded shales are now non-calcareous and unfossiliferous. The reworking and slumping of calcareous nodules shows that their formation took place during early diagenesis.

Late diagenetic and tectonic pressure solution have led to the development of Flasers in some limestones, and accentuated nodule/shale boundaries. Primary structures in the shales are not too badly affected by cleavage.

Section 5.5 Comparison of the Schwellen, slope and basinal facies.

The rise, slope and basinal sediments have now been described and the main features and differences of these facies are summarized in the following table.

COMPARISON BETWEEN SCHWELLEN, SLOPE AND BASIN FACIES

	SCHWELLEN	SLOPE	BASIN
Average thickness for Upper Devonian (from North West Harz)	10 m	100 m	300 m
Dominant lithology	Flaser limestone	Nodular limestone and shales with nodules	Silty shales
Other lithologies	Nodular limestones, thin shale partings between limestones	Flaser limestones, breccias, shales, volcanic horizons	Shales with nodules, turbidites volcanic horizons
Crystal or grain size	Microsparite, less commonly micrite, coarse neomorphic patches.	Micrite/microsparite in calcareous sediments; shales with fine and medium silt-size quartz	Shales with fine and medium silt size quartz
Carbonate type	Calcite	Calcite in limestones locally ferroan dolomite; ferroan dolomite in shales	Ferroan dolomite in shales; calcite in nodules, rarely ferroan dolomite or dolomite

	SCHWELLEN	SLOPE	BASIN
Sedimentary structures connected with lithification	sheet cracks hardgrounds neptunian dykes	absent " "	absent " "
Current structures	thin terrigenous graded units laminated carbonates fossil concentrates	present present present	present absent rare
Slump structures	locally developed	common and diagnostic of this facies	rare
Reworked sediments	occasionally limestone intraclasts	nodules, shale and limestone clasts commonly reworked	rare
Iron minerals	ferromanganese nodules, pyrite (especially in Germany) hematite (especially in Montagne Noire)	pyrite common, as cubes, nodules and disseminated grains	pyrite occurs as nodules and disseminated grains

	SCHWELLEN	SLOPE	BASIN
Diagenesis	Lithification during early diagenesis. Little early diagenetic compaction. Recrystallization during late diagenesis.	Early diagenetic formation of calcareous nodules. Compaction of surrounding shale.	Compaction of shale
Late diagenesis/ tectonism	Formation of Flasers and stylolites through pressure solution, most at the time of cleavage development.	Flasers formed in limestones. Nodule/ shale boundaries accentuated through pressure solution. Cleavage in shale.	Cleavage developed in shale.
Fauna			
Cephalopods	common, locally abundant	present	rare
Bivalves	common, locally abundant	present	present
Brachiopods, corals, gastropods	rare	rare	rare or absent
Trilobites	present	present	present
Crinoids	locally common	present	rare

	SCHWELLEN	SLOPE	BASIN
Conodonts	abundant	common	present
Ostracods	present	locally abundant in shales	locally abundant in shales
Cricoconarids	abundant	locally abundant	common
Foraminifera	locally common	rare	absent
Algae (red)	rare	absent	absent
Calcispheres	rare	absent	absent
Suggested depth of deposition	30 m to 300 m	- transitional -	300m - 1500m?

CHAPTER 6

Geochemical Studies of Devonian Pelagic Sediments

In a rise and basin situation different sedimentological and geochemical processes are operative in different parts of the system. For example, greater current activity occurs on the rise compared with the slope and basinal regions (chapters 3,4 and 5) and, probably as a function of this, clay is concentrated in the basinal sediments compared with those on the rise. In recent pelagic sediments, deposits on topographic highs can be differentiated on their geochemistry from those in the basin. Similarly pelagic carbonates can be distinguished from shallow water carbonates. The object of the geochemical work was to see whether such differences existed in the Devonian and how they compare with Recent sediments.

Several sections in the N.W.Harz Mountains show a transition from basin to rise sediments. Samples for analyses were collected from the following localities (also shown in Fig. 6.1):-

	Sample Nos.
Aeketal (rise sediments)	25-29
Margaretten Klippen (upper slope)	22-24
Lautenthal and Sparenberg (middle-lower slope)	6-17
Innerstetal (lower slope and basin)	1-5
Hutthaler Widerwaage (very condensed volcanic rise)	30,31
Hühnertalskopf (very condensed but mostly shale)	18-21
Bicken (rise in Rheinisches Schiefergebirge)	32-35

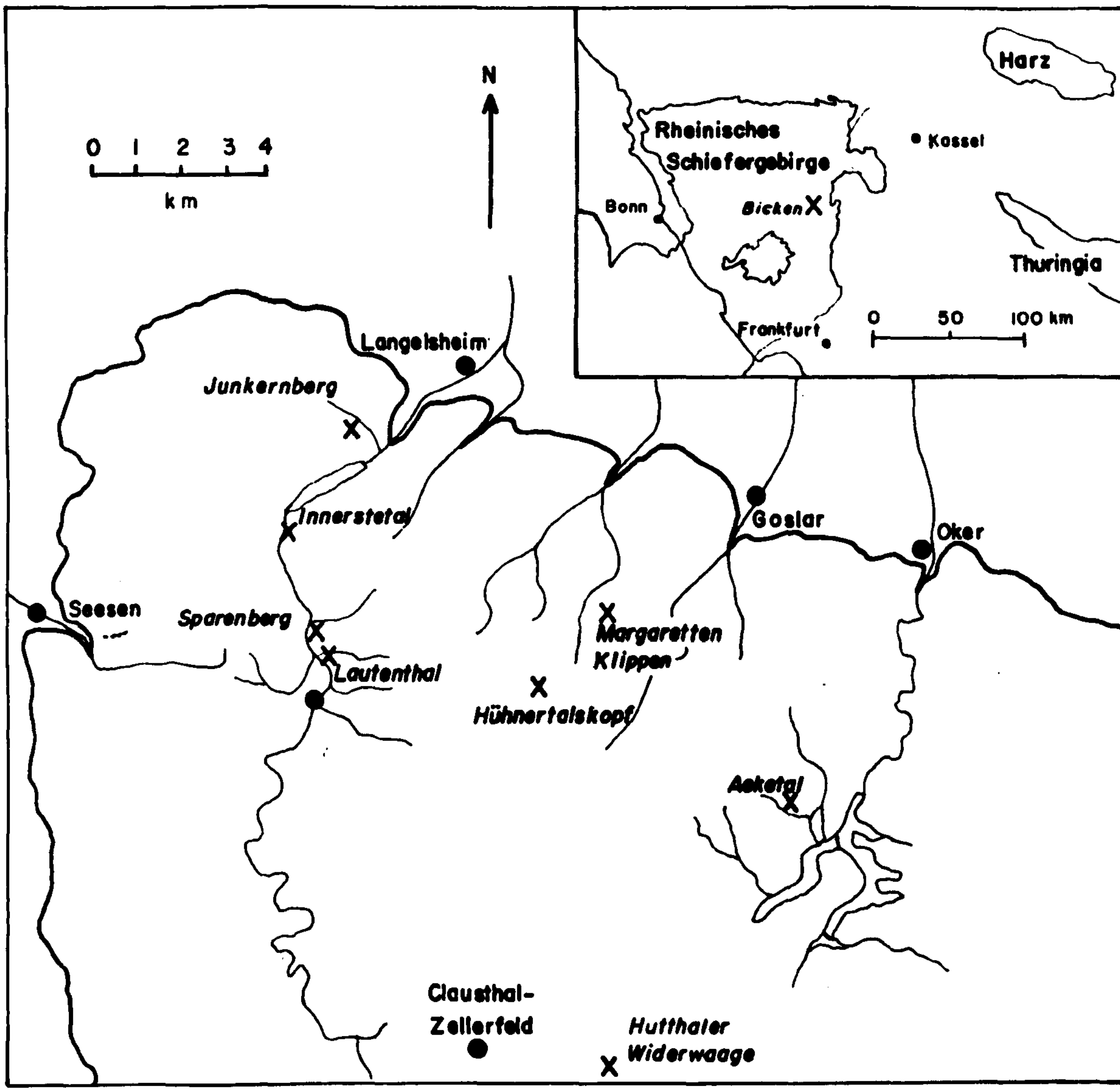


Fig. 6.1 Sketch map of the N.W. Harz showing localities from which samples for chemical analyses were collected.

The age of each sample is based on conodont determinations by Ziegler (1962) for Aeketal, Stoppel (1968) for Margaretten Klippen, Eickhoff (1962) for Lautenthal and Sparenberg, Müller-Steffen (1962, 1965) for Innerstetal and Hühnertalskopf, Meischner and Schneider (1970) for Hutthaler Widerwaage, and Ziegler (1965) for Bicken.

Samples of rise sediments (nos. 18-21, 25-35) are all grey Flaser limestones, except for samples 20 and 21 from Hühnertalskopf, which are calcareous nodules. Samples from the lower slope and basin environments are mostly rather large nodules up to 20 cm across, those from the upper and middle slope are smaller nodules, thin continuous nodular bands, or limestone beds similar to the rise sediments. Thin sections or peels have been made of all samples analysed.

Section 6.1 Analytical Methods

The elements calcium, magnesium, strontium, manganese and iron were determined for 35 samples using a Perkin Elmer 403 Atomic Absorption Spectrophotometer. Details of the preparation of samples are given in the appendix (p.376). Briefly, an exact weight of the powdered sample was digested in a certain volume of acetic acid, and then this solution was diluted to a certain volume depending on the element to be analysed. Standard solutions of the various elements were also prepared, and the absorbance of these solutions was compared directly with that of the sample solutions. Insoluble residues were determined by acetic acid digestion under the same conditions as used in the preparation of solutions for atomic absorption. The mineralogical composition of a selection of samples was determined by X-Ray diffraction and staining of thin sections.

Section 6.2 Results of Chemical Analyses

The results of the atomic absorption work are shown in Tables 1 and 2, and the X-Ray diffraction results in Table 3. The average composition of the three major facies, rise, slope and basin with standard deviations for the elements are shown in Table 4. Sample 2, a Carboniferous basinal nodule, has not been used in the averages since it is quite different from all other samples in being composed of dolomite (Table 3).

Table 4 shows quite clearly that there are major differences between the three facies. Calcium is higher for the rise sediments and lower in the slope and basin. Magnesium content is similar in rise and basin sediments (0.37% and 0.33%), but enriched in the slope deposits by a factor of two (0.67%). Through this the average Ca/Mg ratio is markedly lower for the slope (58) compared with the other two facies (98 and 92). Strontium is rather similar for all three facies though there may be a trend to decrease down-slope. Sr/Ca ratio is appreciably higher for the slope facies. Manganese is rather low for the rise sediments (0.057%) but much richer for the slope (0.33%) and basin facies (0.26%). Iron is markedly enriched in the slope sediments (0.74%) four times that of the rise facies (0.17%) and ten times that of the basinal facies (0.075%). Possible reasons for these variations are discussed later.

Section 6.3 Comparisons with Recent pelagic sediments

The calcium content obviously depends on the clay content and so this element cannot be used in comparison. The Mg content for the rise limestones is similar to that for Recent pelagic sediments which normally have less than 1% (El Wakeel and Riley, 1961). Average values for the Ca/Mg ratio given by Goldberg and Arrhenius (1958), Graf (1960IV), El Wakeel and Riley (1961), and Chester (1965) for Recent pelagic carbonates vary from about 30 to 130. The averages for the Devonian pelagic sediments (98, 58 and

Explanation to Tables and Figures

Geochemical samples are referred to here as samples numbers 1 to 35, but the hand specimen(s) and peel or thin section for each sample for the Reading University archives have the numbers S 23001 to S 23035. The exact locations of samples are given in Appendix (p.380) and are also marked on the sections for the appropriate localities, e.g. ⑭.

The localities (Table 1, column 2) with the symbols used in the figures for the environment are as follows:-

■ = Basin. Samples 1 to 5, Innerstetal (IT)
X = Slope. Samples 6 to 11, Sparenberg (SB)
" 12 to 17, Lautenthal (LT)
⧗ = Upper slope. Samples 22 to 24, Margaretten Klippen (MK)
o = Schwellen. Samples 25 to 29, Aeketal (AT)
" 32 to 35, Bicken (B)
● = 'Special Schwellen'. Samples 18 to 21, Hühnertalskopf (HTK)
" 30 to 31, Hutthaler Widerwaage (HWW)

The age of the samples (Table 1, column 3; Table 3 and some figures) are the abbreviated German stratigraphical names:-

Gat	(*Gattendorfia*)	Lower Carboniferous
Das	(Dasberg)	*Clymenia* Stufe
Hem	(Hemberg)	*Platyclymenia* Stufe
Neh	(Nehden)	*Cheiloceras* Stufe
Adf	(Adorf)	Frasnian
Giv	(Givet)	Givetian

Table 1, column 4 gives the lithology:-

LgNod = large nodule, SmNod = small nodule, Nod = nodule and Lst = limestone.

No.	Loc.	Age	Lith.	Ca%	$CaCO_3$%	Mg%	$MgCO_3$%	Ca/Mg
1	IT	Das	LgNod	33.23	83.07	0.28	0.96	118.6
2	IT	Gat	LgNod	21.54	53.85	6.63	23.0	3.3
3	IT	Neh	Nod	26.87	67.17	0.33	1.14	81.5
4	IT	Hem	Nod	26.83	67.09	0.28	0.98	95.8
5	IT	Das	LgNod	29.31	73.28	0.41	1.41	71.5
6	SB	Giv	Nod	24.21	60.52	0.22	0.76	110.0
7	SB	Adf	Lst	28.73	71.82	0.33	1.13	87.1
8	SB	Adf	Lst	31.87	79.68	0.68	2.37	46.9
9	SB	Adf	NodLst	28.03	70.09	0.62	2.17	45.2
10	SB	Neh	Nod	14.34	35.85	0.50	1.74	28.7
11	SB	Hem	SmNod	29.02	72.56	0.23	0.81	126.1
12	LT	Giv	Nod	30.44	76.10	0.27	0.95	112.7
13	LT	Adf	NodLst	29.64	74.12	1.05	3.62	28.2
14	LT	Adf	SmNod	20.96	52.41	0.74	2.58	28.3
15	LT	Adf	NodLst	28.10	70.24	0.79	2.74	35.6
16	LT	Neh	SmNod	27.72	69.30	0.73	2.54	38.0
17	LT	Hem	NodLst	22.98	57.45	1.92	6.66	12.0
18	HTK	Giv	Lst	29.82	74.57	0.30	1.03	99.4
19	HTK	Adf	Lst	31.13	77.85	0.24	0.82	129.7
20	HTK	Neh	Nod	27.89	69.74	0.17	0.59	164.2
21	HTK	Hem	SmNod	22.95	57.38	0.47	1.62	48.8
22	MK	Adf	Lst	33.80	84.51	0.41	1.41	82.4
23	MK	Neh	SmNod	25.60	64.00	0.22	0.77	116.4
24	MK	Hem	NodLst	30.07	75.18	0.24	0.84	125.3
25	AT	Adf	Lst	31.70	79.25	0.40	1.38	79.3
26	AT	Adf	Lst	35.33	88.32	0.31	1.08	114.0
27	AT	Adf	Lst	33.90	84.74	0.38	1.31	89.2
28	AT	Neh	Lst	33.38	83.45	0.31	1.08	107.6
29	AT	Hem	Lst	30.76	76.90	0.75	2.60	41.0
30	HW	Adf	Lst	26.94	67.36	0.24	0.83	112.2
31	HW	Hem	Lst	31.60	79.00	0.25	0.86	126.4
32	B	Adf	Lst	32.67	81.68	0.26	0.90	125.6
33	B	Adf	Lst	32.10	82.77	0.34	1.17	94.4
34	B	Adf	Lst	32.37	80.93	0.34	1.18	95.2
35	B	Hem	Lst	32.88	82.19	0.25	0.88	131.6

Table 6.1. Atomic absorption results

No.	Sr ppm	Sr/Ca(x10^3)	Mn%	Fe%	Mn/Fe	InsRes
1	256	0.77	0.258	0.078	3.31	15.90
2	-	-	1.50	3.237	0.46	8.25
3	234	0.87	0.446	0.079	5.65	30.27
4	231	0.86	0.147	0.048	3.06	31.76
5	272	0.93	0.172	0.114	1.51	27.49
6	-	-	0.340	0.236	1.44	37.85
7	-	-	0.192	0.265	0.73	26.29
8	-	-	0.170	0.568	0.30	16.13
9	-	-	0.213	0.714	0.30	25.11
10	231	1.61	0.449	0.908	0.49	56.05
11	328	1.13	0.223	0.085	2.62	26.17
12	-	-	0.218	0.296	0.77	22.29
13	-	-	0.461	0.961	0.48	18.39
14	-	-	0.343	0.813	0.42	40.24
15	-	-	0.408	0.992	0.41	20.04
16	281	1.01	0.706	0.832	0.85	23.47
17	200	0.87	0.221	2.240	0.10	29.94
18	-	-	0.073	0.213	0.34	23.19
19	-	-	0.101	0.186	0.54	19.87
20	-	-	0.108	0.176	0.61	27.11
21	-	-	0.091	0.380	0.24	38.10
22	-	-	0.118	0.313	0.38	11.72
23	243	0.95	0.446	0.259	1.72	32.03
24	300	1.00	0.096	0.238	0.40	22.57
25	-	-	0.055	0.156	0.35	18.69
26	-	-	0.044	0.145	0.30	10.30
27	-	-	0.060	0.099	0.61	13.26
28	284	0.85	0.103	0.184	0.56	15.43
29	276	0.90	0.079	0.295	0.29	19.72
30	-	-	0.038	0.125	0.30	31.80
31	225	0.71	0.043	0.062	0.69	17.95
32	-	-	0.050	0.168	0.30	17.18
33	-	-	0.023	0.158	0.15	14.38
34	-	-	0.032	0.206	0.16	16.97
35	-	-	0.064	0.152	0.42	15.71

Table 6.2. Atomic absorption results

Geo-Chem No.	Locality	Age	Calcite	Quartz	Dolomite	Feldspar
1	Innerstetal	Das	x	x	-	-
2	"	Gat	-	x	x	-
3	"	Neh	x	x	-	-
4	"	Hem	x	x	-	-
5	"	Das	x	x	-	-
6	Sparenberg	Giv	x	x	-	?
9	"	Adf	x	x	trace	-
10	"	Neh	x	x	-	trace
13	Lautenthal	Adf	x	x	trace	-
-	Marg. Klippen	Adf	x	x	-	-
18	Hühnertalskopf	Giv	x	x	-	-
19	"	Adf	x	x	-	-
25	Aeketal	Adf	x	x	-	-
28	"	Neh	x	x	-	-
29	"	Hem	x	x	-	-
-	Langestal	Adf	x	x	-	-
31	Hutt. W.W.	Hem	x	x	-	?
-	Dunscombe Farm	Adf	x	x	-	-

Table 63. X - Ray Diffraction Results

	RISE FACIES Aeketal and Bicken		SLOPE FACIES Sparenberg and Lautenthal		BASIN FACIES Innerstetal	
Ca %	32.8	(1.32)	26.3	(4.93)	29.1	(3.0)
$CaCO_3$ %	82.3		65.4		72.9	
Mg %	0.37	(0.15)	0.67	(0.47)	0.33	(0.06)
Mg CO_3 %	1.29		2.34		1.12	
Sr ppm	280	(5.7)	260	(56.3)	248	(19.4)
Ca/Mg	97.6		58.2		91.8	
Sr/Ca ($x10^3$)	0.87		1.16		0.86	
Mn %	0.057	(0.024)	0.330	(0.158)	0.256	(0.135)
Fe %	0.174	(0.054)	0.742	(0.567)	0.075	(0.027)
Mn/Fe	0.35		0.75		3.38	
Mn/Ca ($x10^3$)	1.7		12.4		8.8	
Fe/Ca ($x10^3$)	5.3		28.2		2.6	
Insoluble Residue %	15.9		28.6		26.4	

Table 64: Table of averages for Devonian pelagic sediments from Germany. Standard deviations in brackets.

92) fall well within this range.

The Sr/Ca ratio decreases with increasing geological time through diagenetic removal of Sr (Wolf et al, 1967; Dodd, 1967), and it is for this reason that the Sr values for the Schwellen limestones are so low compared with Recent pelagic sediments which have Sr contents in the region of 1000 ppm (El Wakeel and Riley, 1961). Although this is based on whole rock analyses, Graf (1960 111) considers that most Sr is substituted for Ca in calcite. Similarly, figures for Mn and Fe are based on whole sediment analyses and this results in the modern pelagic deposits having a higher concentration than the German limestones. Values given for Mn (El Wakeel and Riley, 1961 and Graf, 1960) average 0.3% though lower values are recorded e.g. 0.02% for an Atlantic Globigerina ooze (Graf, 1960) showing that the Devonian values are of the right order of magnitude. Mn and Fe are also depleted during diagenesis (Gavish and Friedman, 1969).

The analyses then, show that the Devonian limestones compare well with modern pelagic sediments, although their geochemistry has been modified during diagenesis.

Section 6.4 Comparisons with other limestones

The literature on the geochemistry of limestones is rather limited and there are very few analyses of pelagic limestones. Only Audley-Charles (1965) gives a few analyses of manganiferous and ferriferous limestones of late Cretaceous age from Timor, and Hallam (1967) mentions the composition of Jurassic pelagic limestones from the Alps. Many thousands of limestones of PreCambrian to Caenozoic age have been analysed from the Russian Platform (Vinogradov and Ronov, 1956) and this forms the major work in limestone geochemistry. Some analyses of other limestones are given in Table 5 for comparison.

	Ca	Mg	Ca/Mg	Sr	Mn	Fe
Schwellen Limestones	32.8	0.37	97.6	0.028	0.057	0.174
Up.Dev.-Lr.Carb. Veevers (1969)	31.7	0.90	35.0	0.037*	0.035	0.115
Up.Dev.-Lr.Carb. Russian Platform	28.0	6.00	4.7	0.045	0.047*	2.700*
Up.Cret. Timor Fe rich lst.	24.4	0.45	54.2	0.040*	0.660*	-
Mn rich lst. Aud-Charles (1965)	25.5	0.63	40.5	0.038*	-	-
Tethyan Jurassic Hallam (1967)	28.3*	0.66*	42.9	-	0.021*	1.700*
Average Limestones Graf (1960)	-	-	-	0.048*	0.050*	-
Average Pel. Carb. El Wakeel & Riley	20.0	1.03	19.4	0.111	0.210*	2.700*
High Carb. Pel. Sed.	31.9	0.42	75.9	-	0.050*	1.500*
Average pel. Carb. Chester (1965)	22.7	0.85	26.7	0.100	0.240*	2.100*
Glob. Ooze Graf (1960)	31.8*	0.50*	63.6	-	0.420*	-
Lith. Pterop. Ooze Milliman (1969)	33.0	1.19	27.7	0.640	0.210	0.510

* denotes whole rock analysis. Ca and Mg values are for carbonate phase.

Table 6.5. Some analyses of limestones and Recent pelagic sediments for comparison with the Devonian Schwellen limestones.

6.4.1 Calcium and Magnesium

The analyses of limestones (mostly shallow water) of Upper Devonian/Lower Carboniferous age from the Russian Platform (Table 5) show a similar Ca content (28%) but a much higher Mg content (6%) in comparison with the rise sediments (0.37% Mg). The Ca/Mg ratio for the Russian carbonates is really low (8) compared with 98 for the Schwellen limestones. The Devonian pelagic limestones then, with reference to the Russian Platform carbonates are deficient in Mg by a factor of 20, and one must consider whether this is significant, i.e. does it reflect an original low concentration of Mg, or has Mg been lost diagenetically, a phenemenon illustrated by the analyses of Stehli and Hower (1961) and Gavish and Friedman (1969) for Recent and Pleistocene shelf carbonates.

Veevers (1969) analysing Upper Devonian/Lower Carboniferous platform sediments from Australia (Bonaparte Gulf Basin) obtained an average Mg content of 0.9% and a Ca/Mg ratio of 35. Since this ratio is much higher than the Russian value for rocks of the same age, Veevers considered that such a low Mg content could only be explained by 'some factor preventing hypersalinity, such as a wet climate causing low evaporation, or continuous mixing of the shelf water with the deeper oceanic water'. The latter hypothesis is applicable to the Schwellen limestones, that it is 'oceanic' water that is causing the low Mg concentration. It is usually thought that Mg is concentrated by restricted marine conditions (Veevers, 1969) and clearly it could be the reverse of this, open oceanic conditions, that has caused the low Mg content of the Schwellen limestones.

Various authors (e.g. El Wakeel and Riley, 1961; Dodd, 1967; Billings and Ragland, 1968) have suggested that the $MgCO_3$ content depends on the type of calcareous organism predominant in the sediment. Most calcareous pelagic sediments today are oozes formed of foraminifera and/or coccoliths, both of which have tests made principally of low Mg calcite (Blackmon and Todd, 1959; Stehli and Hower, 1961). It is possible then that the Schwellen limestones were also formed by low Mg carbonate skeletal material which has resulted in the high Ca/Mg ratio.

It has often been written in the literature that the Ca/Mg ratio for limestones decreases with geological time (e.g. Wolf et al, 1967). This increase in Mg probably results from the longer time available for contact with connate and meteoric waters. This is shown in Fig. 6.2, redrawn from Vinogradov and Ronov (1956), and based on numerous analyses, mostly of shallow water limestones. It can be seen clearly that the Ca/Mg values for the Devonian rocks do not fit in with this scheme. They are comparable rather with those of the Cretaceous, which have a high Ca/Mg ratio since chalk, so widespread at this time, is comparable with modern pelagic oozes and is composed mainly of coccoliths (low Mg calcite).

6.4.2 Strontium

Graf (1960) on whole rock analyses of Palaeozoic limestones gives Sr as 420-490 ppm. For late Cretaceous pelagic limestones however, Audley-Charles (1965) obtained lower values, 300-460 ppm, and possibly low Sr is a feature of pelagic limestones. One might expect this since low Mg calcite, the dominant mineral in pelagic organisms, does not hold as much Sr in the lattice as the other carbonate minerals, high Mg calcite and aragonite (Siegel, 1961; Billings and Ragland, 1968), which are characteristic of warm shallow water carbonate areas. However, other analyses for Sr show that there is a wide range of values. Sternberg et al (1959) recorded Sr values of 60 to 1600 ppm for the Jurassic Steinplatte Reef in Austria, and found an increase in Sr in the deeper water limestones. Kahle (1965a) obtained values between 230 and 400 ppm for Silurian and Lower Carboniferous oolitic limestones. The average for the Devonian Schwellen limestones (280 ppm) falls well within the range reported for other limestone types, showing that Sr is not significantly different in these pelagic limestones.

6.4.3 Manganese and Iron

The Mn and Fe values for the Schwellen limestones are higher than most published analyses of limestones, which are all shallow

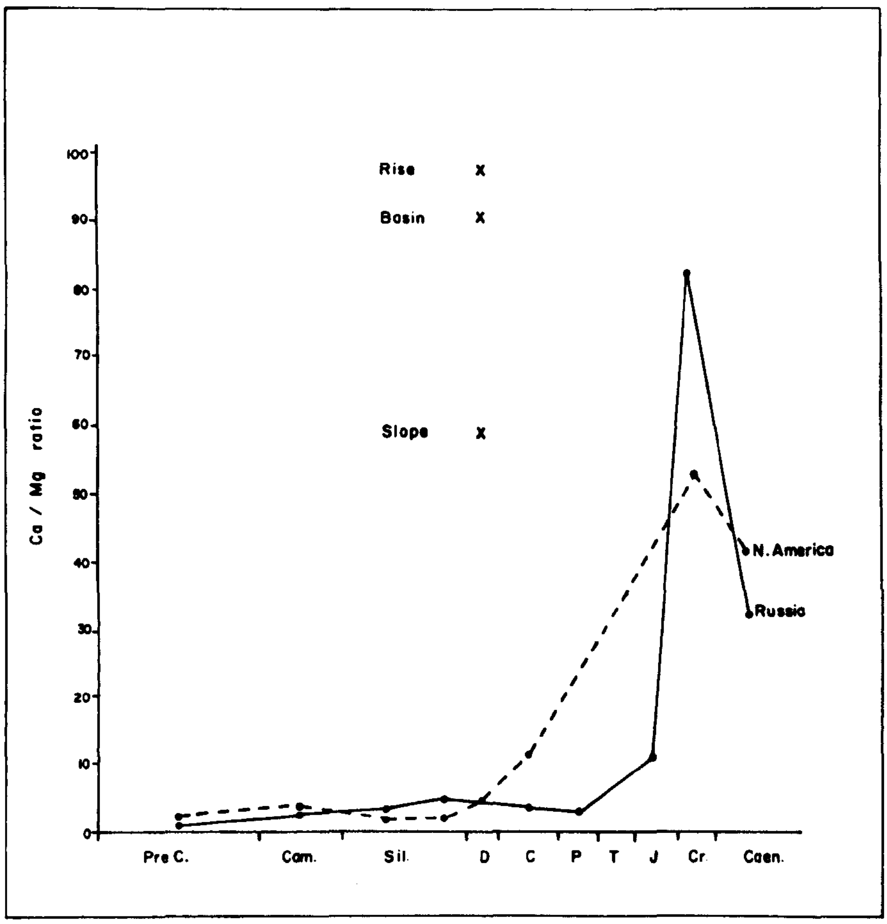


Fig. 6.2 Graph showing variation of Ca/Mg ratio with time, (redrawn from Vinogradov and Ronov, 1956). Ca/Mg ratios for the rise, slope and bainal facies are shown for comparison.

water or reef carbonates (e.g. Veevers records 0.035% Mn and 0.115% Fe, and values obtained here are 0.057% Mn and 0.174% Fe). The values for these elements from the Russian Platform carbonates are similar for Mn (0.047%), but much higher for Fe (2.7%). However whole rock analyses were made and much of the Fe and some of the Mn must be in the clay fraction. By comparison the Jurassic pelagic limestones from the Alps (Hallam, 1967) have quite low Mn (0.021%) and Fe (1.7%) contents on whole rock analyses. The analyses of Audley-Charles (1965) were of Cretaceous pelagic limestones especially rich in Mn and Fe, and cannot be compared.

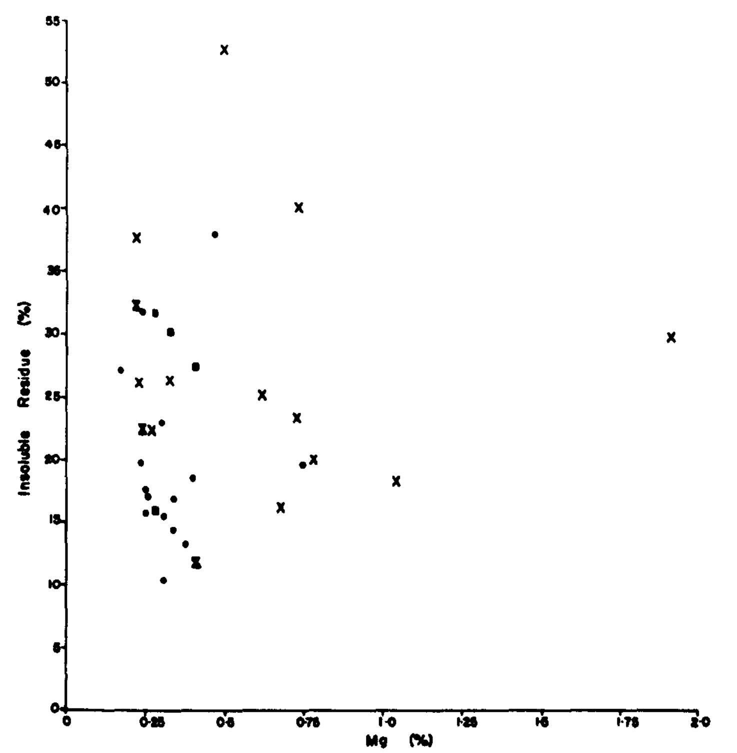
Generally then, the Mn and Fe contents of the Devonian pelagic limestones are higher than limestones of other environments. These elements may be derived from continental run-off or volcanic sources (Krauskopf, 1967). An enrichment in these elements is typical of pelagic sediments, through the slow rate of deposition.

Section 6.5 Discussion of Results for Devonian Pelagic Sediments

6.5.1 Calcium and Magnesium

From Table 4, it can be seen that there is less Ca and more Mg in the slope sediments, and hence that the Ca/Mg ratio is quite distinct. The insoluble residue is lower for the rise and higher for the slope and basin. One would expect this from greater current activity on the rise, sweeping clay material towards the basin. The lower Ca average for the slope sediments is somewhat surprising, but examination of the data shows that it is certain nodule samples (10, 14 and 17) which lower the Ca value and produce a high standard deviation. This could in fact be a sampling error if some of the shale from around the nodules was included in the analyses. Collecting only the central part of nodules in the slope facies (to obtain a sample free from weathering and without the surrounding shale) is at times rendered difficult by the small size of the nodules. The size of nodules decreases up-slope, but nodule bands are closer together (ch. 4). However, the amount of insoluble residue does not affect the Ca/Mg ratio, and the average for the slope (58) is appreciably lower than the rise and basinal values (98 and 92 respectively). Leaching of Mg from clay minerals during digestion of the sample is not considered important, since only 3% acetic acid was used. Hirst and Nicholls (1958) consider that even 25% acetic acid does not cause significant leaching.

Figs. 6.3 and 6.4 show the plots of insoluble residue and Ca against Mg. From Fig. 6.3 it is seen that about ⅝ of the samples have Mg concentrations between 0.2 and 0.4%, and that the rest are randomly scattered. Similarly in Fig. 6.4 about ⅝ of the samples occur together – the rise and basin samples, with a few from the slope. The rest (10) are quite widely scattered and these are mostly samples from the slope (hence high standard deviation for Mg in the slope facies, Table 4). Plotting Ca/Mg ratio and $CaCO_3$ content against distance across the West Harz Schwelle for Adorf, Nehden and Hemberg samples (Figs. 6.5 - 6.10) again show the variations in the slope region (outcrops Sparenberg and Lautenthal) but they also show that the variations are not consistent at a particular locality. Fig. 6.11 of Ca/Mg ratio against insoluble fraction is interesting since the samples fall into two groups, those with high Ca/Mg ratios and those with lower values, which are the samples giving the wide scatter in Figs. 6.3 and 6.4. Samples with high ratios are all the rise limestones except one, all the basinal nodules, and some of the slope sediments – those from the upper slope and some lithologically similar to basinal nodules. Of the samples with a low Ca/Mg ratio, i.e. rich in Mg, seven samples (8,9,13,14,15,16 and 17) come from beds which are associated with, or close to slumped or reworked sediments. Other samples with low Ca/Mg ratio (10,14 and 21) are nodules from dominantly shaley parts of the succession. The samples with higher Mg are of mixed lithology, it is not as if they are all nodules. There is no correlation between Ca and Mg (Table 6) for the rise or slope facies. To explain the enrichment of Mg for some samples then, there are at least four possible causes, 1) Diagenetic migration of Mg from clay minerals to the calcite (c.f. Kahle, 1965b). This could account for samples



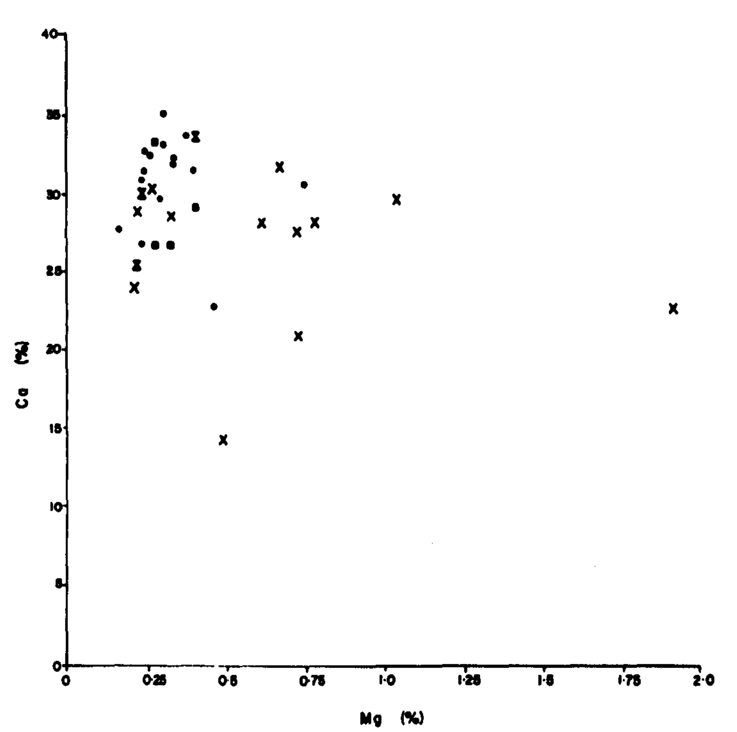
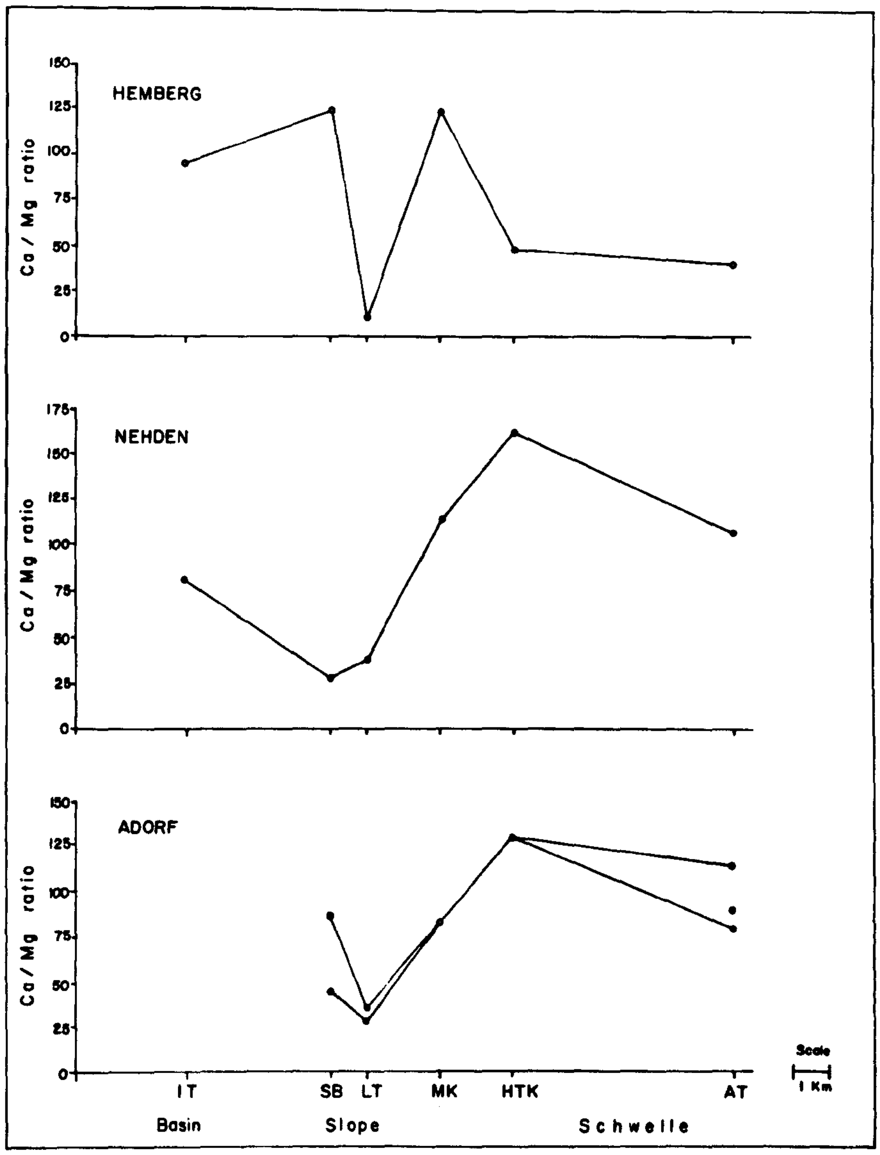
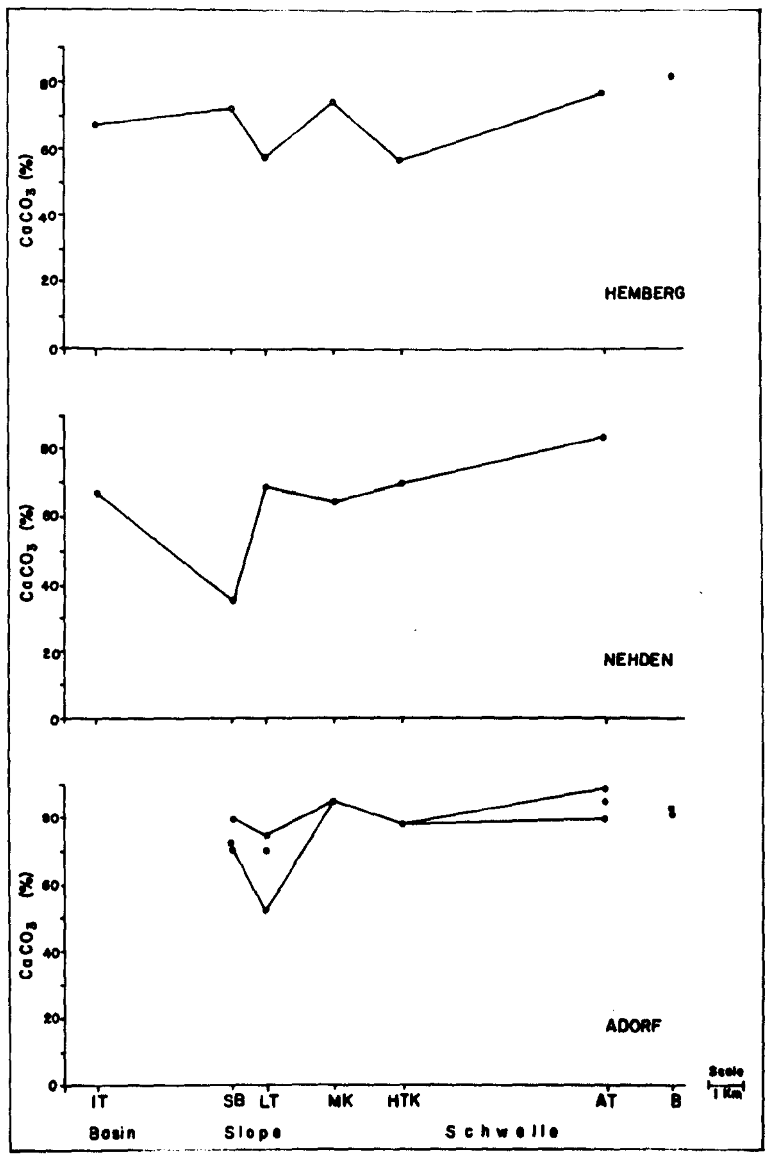


Fig. 6.3 Graph of Ca plotted against Mg.

Fig. 6.4 Graph of insoluble residue against Mg.
(Abbreviations as in Table 1 p.263).



Figs. 6.5, 6.6 and 6.7 Ca/Mg ratios plotted against distance across the North West Harz. (Abbreviations as in Table 1 p.263).



Figs. 6.8, 6.9 and 6.10. $CaCO_3$ content plotted against distance across the North West Harz.
(Abbreviations as in Table 1 p.263).

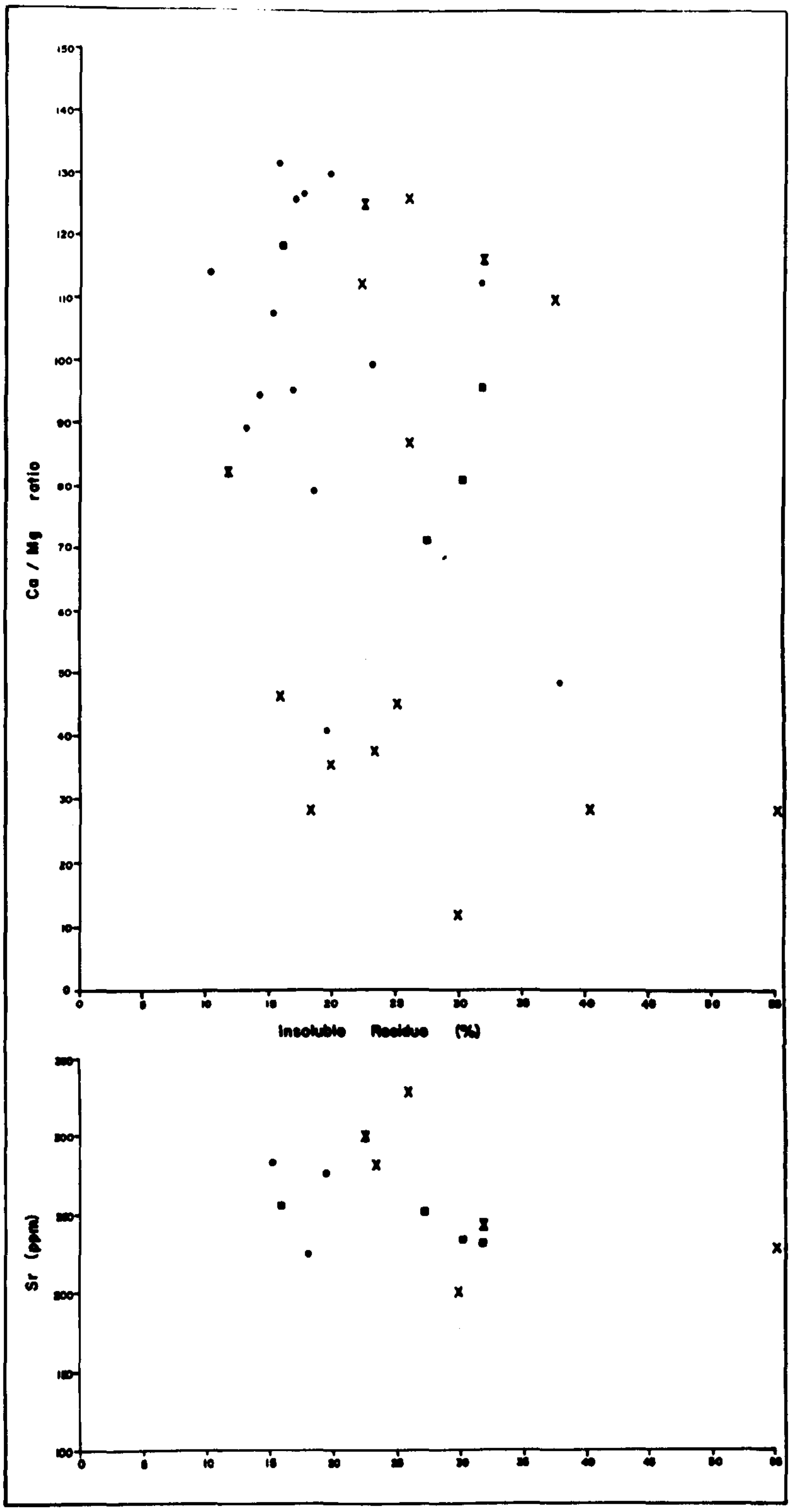


Fig. 6.11 Ca/Mg ratio plotted against insoluble residue.

Fig. 6.12 Sr plotted against insoluble residue.
(Abbreviations as in Table 1).263).

Rise (7 degrees of freedom)

Mg	-0.59		
Mn	0.10	+0.28	
Fe	-0.67	+0.75	+0.28
	Ca	Mg	Mn

Slope (10 degrees of freedom)

Mg	-0.14		
Mn	-0.26	0.1	
Fe	-0.35	+0.96	+0.15
	Ca	Mg	Mn

Basin (2 degrees of freedom) (not significant)

Mg	-0.14		
Mn	-0.15	0.1	
Fe	+0.30	+0.89	0;1
	Ca	Mg	Mn

Table 6.6. Correlation Matrices for Rise, Slope and Basin Samples. Correlations significant at 95 per cent level underlined.

10, 14 and 21. 2) 'Secondary' enrichment of Mg from seawater to the pore-waters in partly consolidated sediments caused by down-slope movement of sediment - this could account for seven of the samples. 3) Migration of pore-waters within the slope sediment. and 4) Migration of connate waters, upslope, through compaction and pressure solution of the basinal sediments. This will be discussed further in section 6.5.3.

6.5.2 Strontium

Unfortunately only thirteen samples were analysed for Sr, and this is not sufficient to draw any definite conclusions (Fig. 6.12). Basinal samples are quite constant in acid-soluble Sr (average, 248 ppm). The slope samples however, again show a wide range (330 to 200 ppm) and have a high standard deviation. For the rise, the two samples analysed give similar values (280 ppm). The slightly higher Sr content of the rise sediments could reflect less diagenetic activity there, than in the basinal or slope sediments.

6.5.3 Manganese and Iron

The three facies, rise, slope and basin can be distinguished on their Mn and Fe contents and Mn/Fe ratios (Table 4). The rise limestones are low in Mn and Fe, slope facies are high in Mn and Fe, and the basin facies are high in Mn and low in Fe. From Fig. 6.13 (graph of Mn and Fe) it can be seen that 1) the rise sediments have similar Mn and Fe contents, and they all cluster together around Mn 0.05% and Fe 0.17%, 2) the basin sediments are enriched in Mn relative to Fe and 3) the slope sediments show a high but variable Mn and Fe content, giving high standard deviations (Table 4). Figs. 6.14 and 6.15 of Mn and Fe plotted against Ca, show clearly the rise sediments clustering together, and the wide scatter for the slope sediments. Figs. 6.17, 6.18 and 6.19 of Mn/Fe ratio against distance across the West Harz Schwelle show clearly the rise of the ratio towards the basin.

Two problems arise from the Mn and Fe determinations, 1) why is there more Mn than Fe in the basin samples and 2) why is there

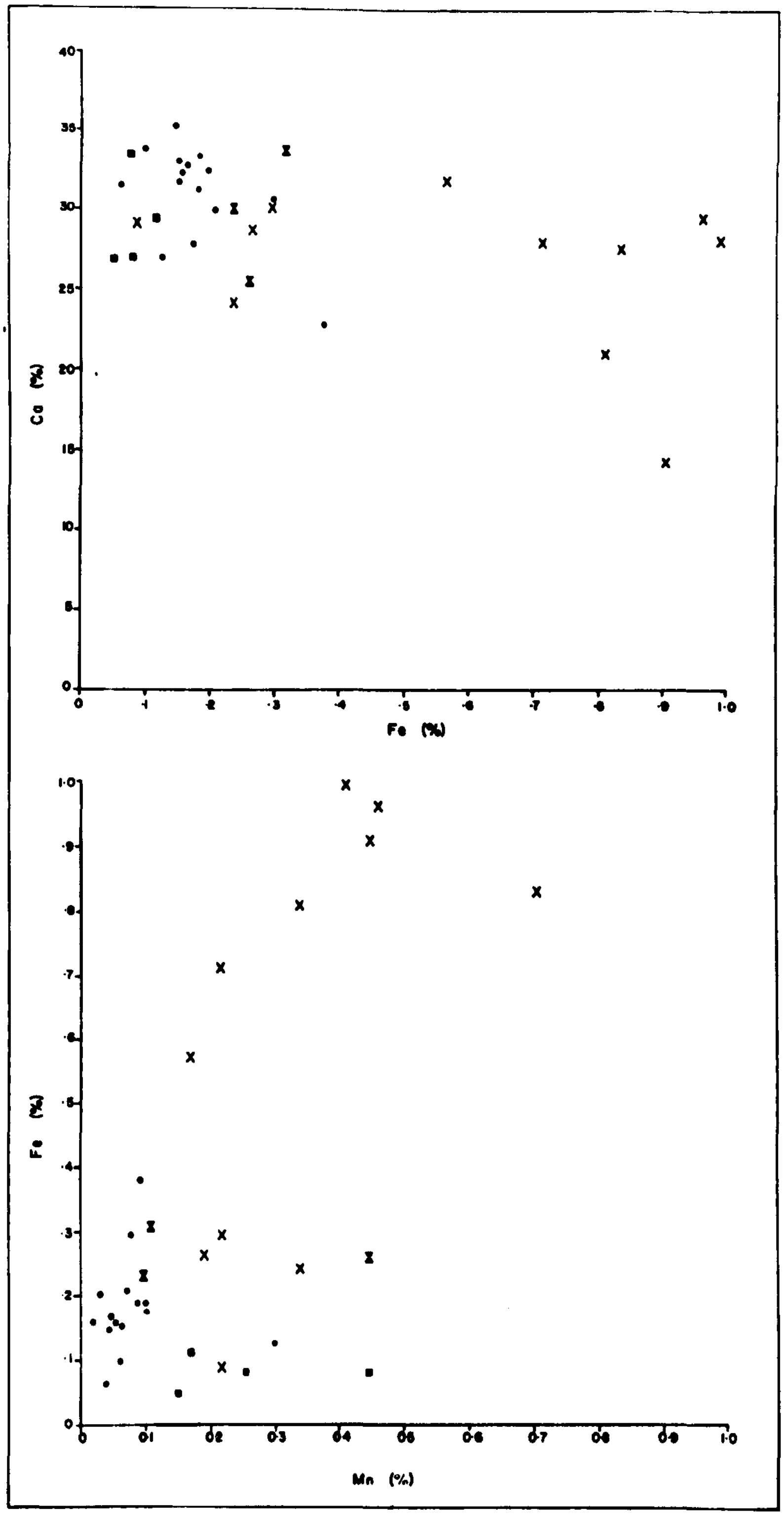
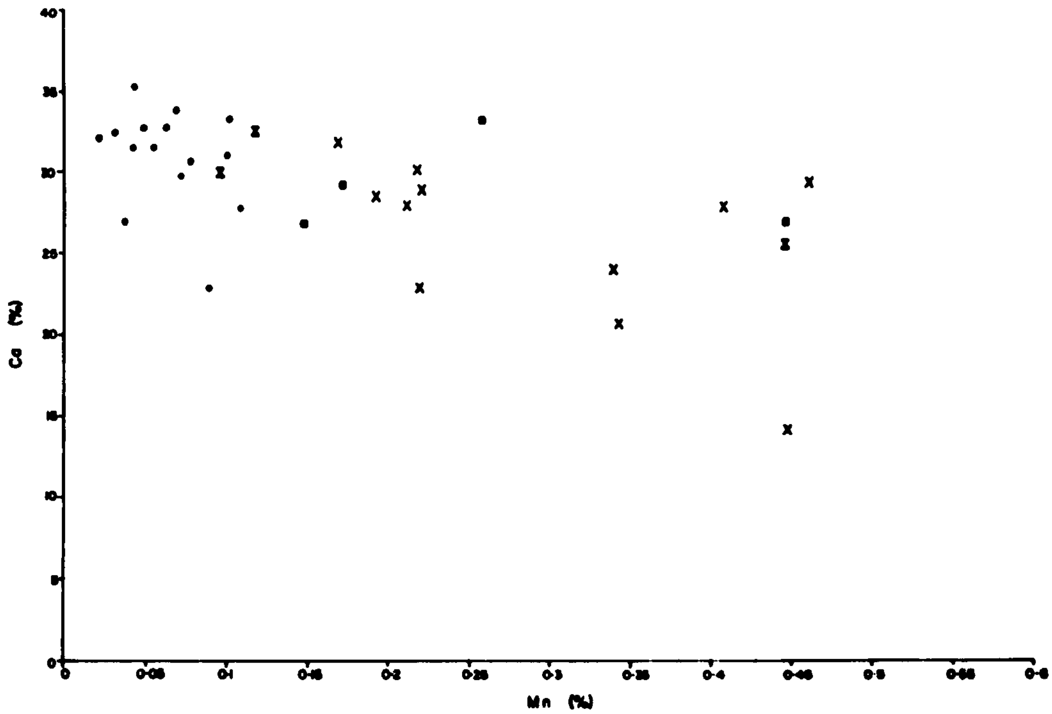


Fig. 6.13 Graph of Fe plotted against Mn.

Fig. 6.14 Graph of Ca plotted against Fe.
(Abbreviations as in Table 1, p.263).



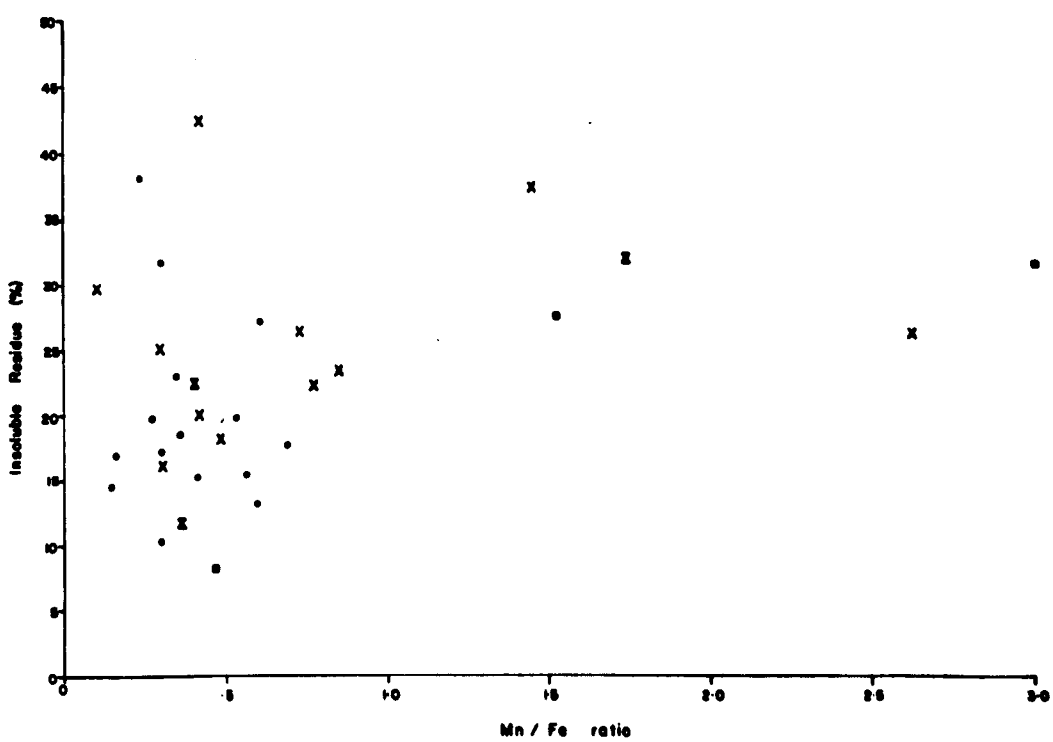


Fig. 6.15 Graph of Ca plotted against Mn.

Fig. 6.16 Graph of insoluble residue plotted against Mn/Fe ratio. (Abbreviations as in Table 1, p.263).

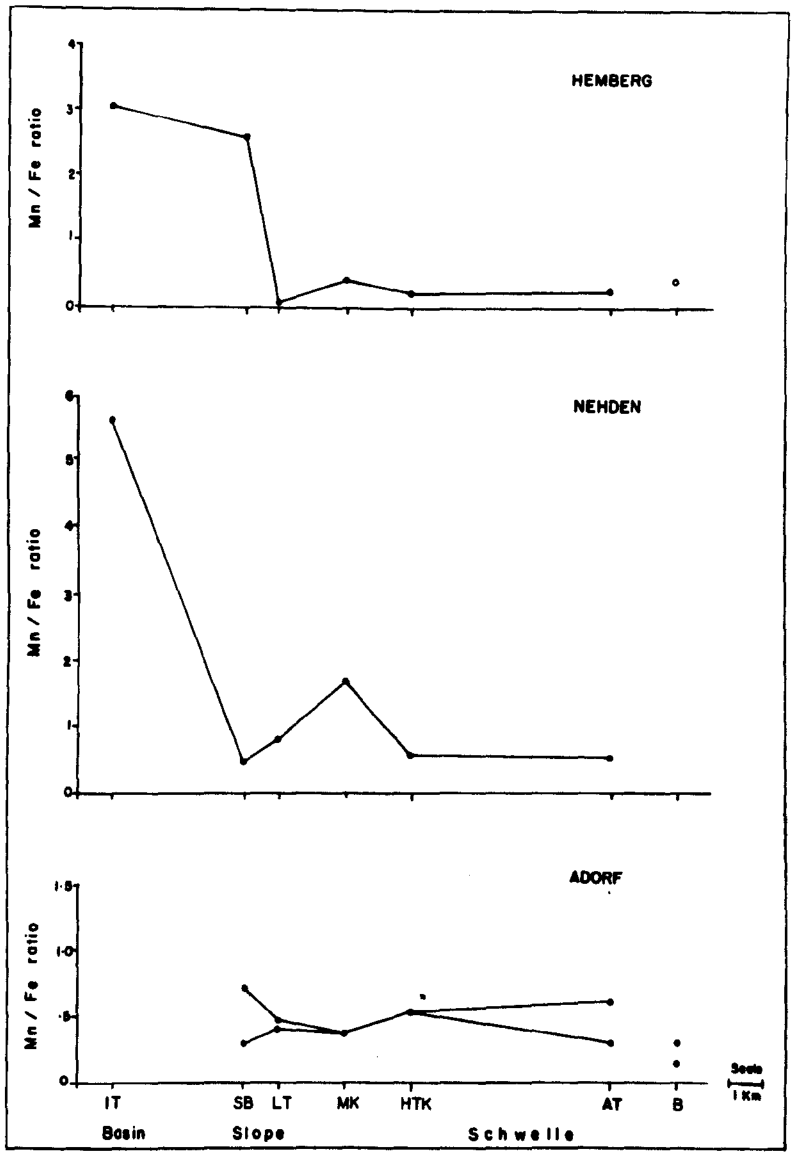


Fig. 6.17, 6.18 and 6.19. Mn/Fe ratio plotted against distance across the North West Harz.
(Abbreviations as in Table 1, p.263).

such a scatter for the slope sediments.

The enrichment of Mn relative to Fe in the basinal nodules may be a question of lithology. The basinal samples are quite large calcareous nodules (up to 30 cm across) and presumably took longer to accrete than the smaller nodules generally occurring in the slope region (chapters 4 and 5). More pore-water would have been involved in the formation of the larger nodules, and since Mn goes into solution more readily than Fe (Lynn and Bonatti, 1965) this would lead to an enrichment in Mn over Fe. Manganese nodules are thought to be enriched by the same process. The rise sediments are generally rather low in Mn and this presumably is also due to the upward migration of Mn, caused by dissolution in the reduction zone of the sediment.

In connection with the second question, the samples giving the scatter are the same ones as those with high and variable Mg contents described in section 6.5.3. The suggestions put forward there may apply in the case of Mn and Fe too. Since these sediments have insoluble residues mostly falling within the range of the rise and basin samples, it is likely that the higher Mg, Fe and Mn contents are original enrichments, and are not due to leaching from clays during digestion.

The enrichment of these elements is probably due to the effects of the slope itself. It has been suggested by a number of authors that compaction and pressure solution in basinal sediments leads to a migration of connate waters upslope, providing much of the carbonate for cementation of limestones on shelf areas (Chilingar et al, 1967; Trurnit, 1968a; 1968b). The Schwellen and Becken situation provides a test for this hypothesis since there is no evidence of emersion to complicate matters. The evidence from the geochemistry and sedimentology points against the migration theory. The Mg, Mn and Fe contents of the rise sediments are low compared with the slope facies suggesting that migration has not occurred in this direction. The carbonate in the Schwellen limestones, as revealed by staining and X-Ray diffraction, is always calcite. Dolomite and ferroan calcite, which would be present if a migration of cementing fluids had taken place, are completely absent. Also, there is evidence for early lithification of the limestones (chapter 3). Ferroan calcite and ferroan dolomite do locally occur in limestones lower down the slope (chapter 5), where they would certainly be expected from the higher Mg and Fe contents.

These results suggest then that connate waters have not migrated from the slope to the Schwellen region (unless some physical barrier such as a fault plane prevented this). The similarity of the basin sediments to the rise sediments (apart from higher Mn in the basin) would suggest that it is movement of connate waters within the slope sediments that is causing the enrichment. Perhaps then there is a greater diagenetic leaching of clay minerals in the slope region leading to the enrichments in Mg, Fe and Mn.

Goldberg and Arrhenius (1958) suggest that the rate of deposition of marine sediments is inversely proportional to the Mn content. This is certainly not illustrated by the analyses here, the Mn content does not decrease downslope as the sedimentation rate increases. There has presumably been too much diagenetic redistribution of Mn, but also only the Mn in the calcite has been determined and account has not been taken of Mn in the insoluble fraction or in the shales associated with the nodules.

Section 6.6 Conclusions

The results presented above show that indeed the Devonian pelagic sediments can be differentiated from limestones of other facies on their chemistry. and that they show similarities with Recent pelagic sediments.

The Schwellen limestones are characterized by low Mg and high Mn and Fe, in the acid-soluble fraction. Following Graf (1960) and Veevers (1969) these elements are considered to be in the calcite lattice. Low Mg (apart from diagenetic depletion) is either due to low Mg calcite skeletal material forming the limestones, or the result of oceanic conditions. These two possibilities are obviously interrelated. High Mn and Fe are to be expected for pelagic sediments. The Sr concentration does not differ much from other limestones.

The geochemical data also shows that there are differences between the rise, slope and basin facies, which can be ascribed to original topographic relief and diagenesis. The slope sediments are rather variable in Ca, Mg, Mn and Fe. The basin and rise samples are more constant, although Mn is enriched relative to Fe in the basinal nodules. The variable but high Mg, Mn and Fe for the slope samples is attributed to movement of connate waters within the slope sediments. There is no evidence for ascending connate waters from the basin to the rise, leading to the cementation of the Schwellen limestones.

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Chapter 7

Ferromanganese Nodules from Devonian pelagic sediments

Manganese nodules are characteristic of present-day pelagic sediments but also occur in marginal marine (Manheim, 1965) and freshwater environments (Ljunggren, 1955). A slow rate of sedimentation, enabling manganese and iron to be concentrated, low organic content and oxidising conditions appear to be the main factors in their formation (Price and Calvert, 1970). Volcanism and continental run-off are considered the main sources of the iron and manganese.

Ferromanganese nodules and encrustations around shell fragments and limestone clasts are described here occurring in the Upper Devonian pelagic limestone facies (griotte) of the Montagne Noire, S. France. They were only found at one locality in the Rheinisches Schiefergebirge, Germany. The nodules have been collected mainly from Combe D'Izarne, a valley section 4 km S.W. of Cabrières (Hérault) at the eastern end of the Montagne Noire. They have also been found at Mont-Peyroux in the central part, near the village of Causses-et-Veyran. Localities of the Montagne Noire are shown in Fig. 7.1. Conodonts give Famennian ages for the nodule horizons at Combe D'Izarne, and lower Frasnian for Mont-Peyroux (Appendix). At the other localities examined in the Montagne Noire (Coumiac, Caunes Minervois, and St. Nazaire de Ladarez) nodules are not so common. At various horizons in the Upper Devonian diagenetic enrichments of iron and manganese occur along pressure solution planes and fossils have been hematized (p. 77). Epigenetic mineral veins are developed at Caunes Minervois, at the western end of the Montagne Noire.

In the sections examined in the Pyrenees, west of Seo de Urgel, where a similar Upper Devonian pelagic facies is developed, ferromanganese encrustations around shell fragments were only occasionally seen.

Fossil ferromanganese nodules have previously only been described from the Cretaceous of Timor (Audley-Charles, 1965) and the Tethyan Jurassic of Sicily (e.g. Jenkyns, 1967, 1970) and the Alps (e.g. Wendt, 1969; 1970; Germann, 1971). The Cretaceous nodules from West Timor

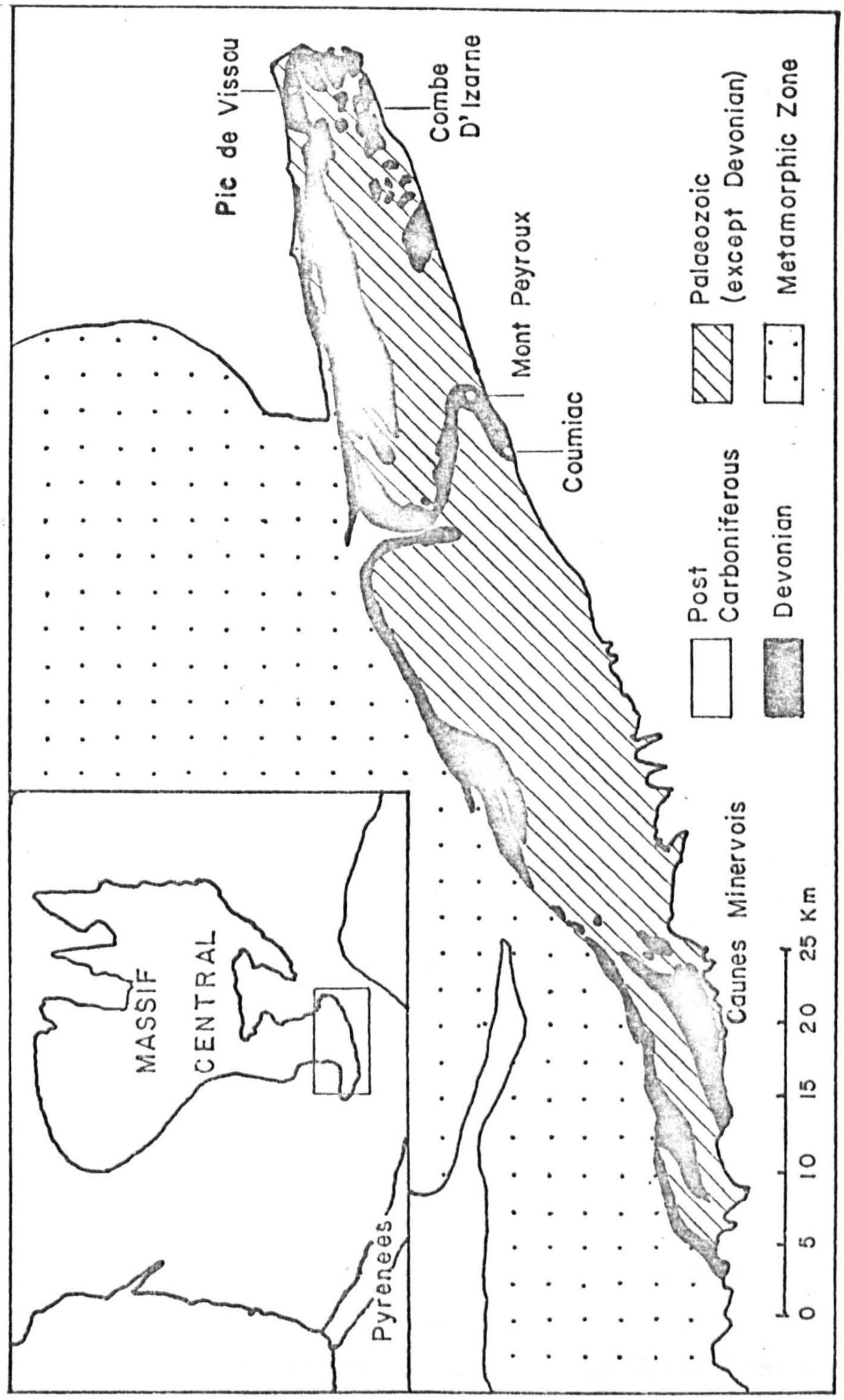


Fig. 7.1 Geological sketch map of the Montagne Noire (S. France) showing position of main localities (After Boyer et al, 1968)

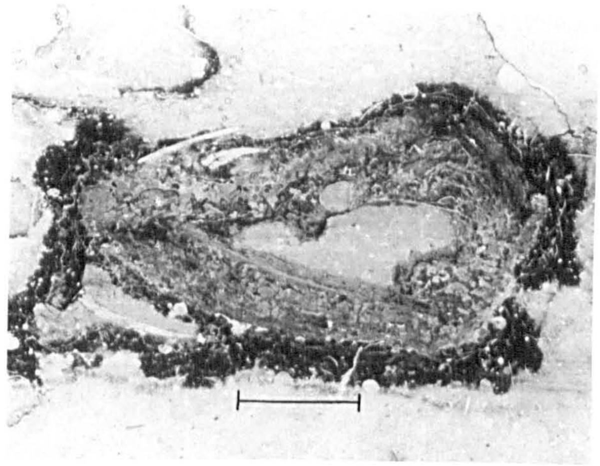
are considered to be deep water oceanic nodules, on their rich trace element content; those from East Timor are more akin to marginal nodules (Price, 1967). Jenkyns and Torrens (1971) consider the Jurassic nodules to have formed on current-swept seamounts within the photic zone (on the evidence of algal stromatolites interbedded with the nodules). Perseil (1968) has studied the mineralogy of manganese minerals in the south of France and has described these occurring as cements in Permo-Triassic sandstones and Carboniferous quartzites, and as nodules and lenses in the Upper Devonian griotte of the Pyrenees (Massif de l'Arize) and the Montagne Noire near Caunes Minervois. Perseil considers the Devonian manganese minerals to have formed by 'sedimentary concentration, where later enrichment by secondary processes is manifest'. Although Perseil says the nodules must have been formed penecontemporaneous with sedimentation, she does not appear to have recognised these nodules as the fossil equivalents of the manganese nodules which characterize modern pelagic sediments.

Section 7.1 Ferromanganese nodules and encrustations

A brief description of these nodules is given in Tucker (1971), (appended). The nodules take the form of encrustations around limestone clasts and skeletal fragments and range in size from a few millimetres to 4 or 5 cm in diameter (Figs. 7.2 and 7.3). They are generally a reddish-brown colour though some are almost black. Shells, particularly bivalves, are frequently coated in a 1 mm thick ferromanganese crust, which usually shows the development of colloform structures (the limonitic cauliflowers of Jenkyns, 1970a). These are up to 0.5 mm across and consist of an alternation of black and yellow laminations (Figs. 7.4 and 7.5) each about 10μ thick. These structures have been interpreted as organic in origin (Farinnacci, 1967) but Cronan and Tooms (1968) have described them from Recent nodules and furnished evidence to show that it is caused by diagenetic movement of the manganese and iron. This process works from the centre outwards, replacing the original concentric banding. In some cases in the Devonian nodules this has gone to an extreme and complete

Fig. 7.2 Limestone clast encrusted with a ferromanganese crust, containing foraminifera and skeletal fragments. Black 'halo' around the nodule, replacing the sediment, illustrates the mobility of iron and manganese. Lower Famennian. Combe D'Izarne, Montagne Noire. Thin section S 22965. Scale bar = 0.5 cm.

Fig. 7.3 Limestone clast with ferromanganese encrustation. Colloform structures are developed around the outside and complete segregations occur further into the crust. Lower Famennian. Combe D'Izarne, Montagne Noire. Thin section S 22964. Scale bar = 0.5 cm.



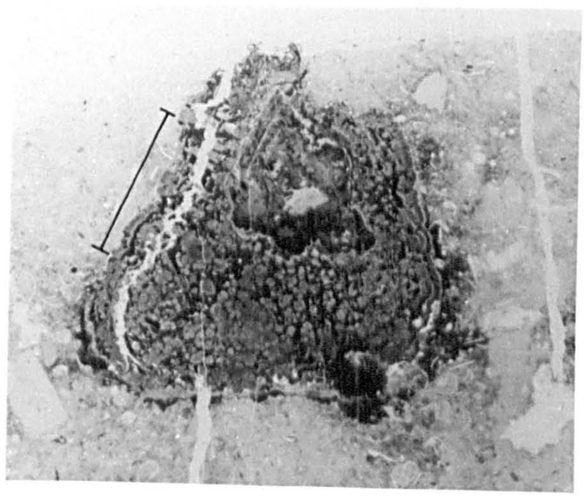
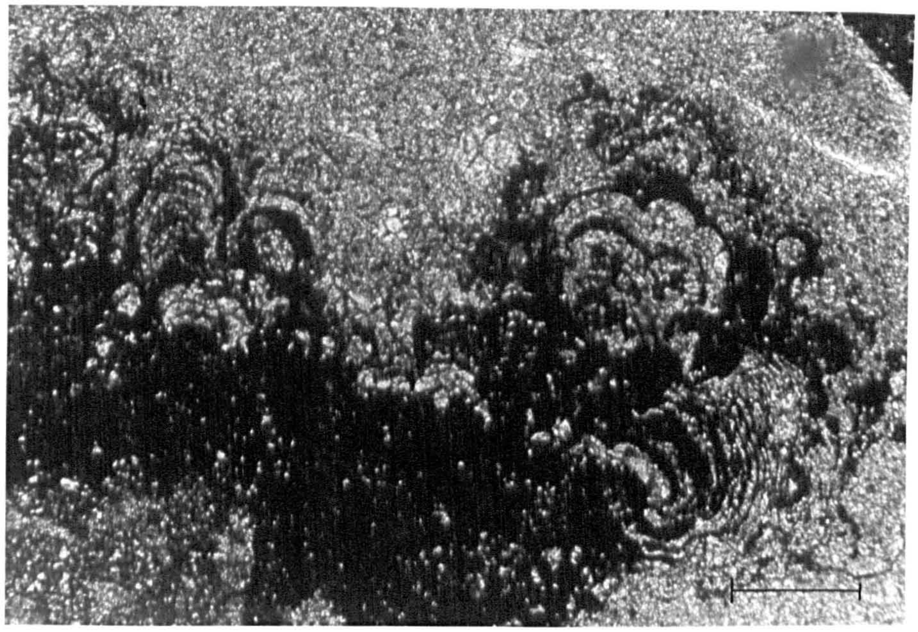
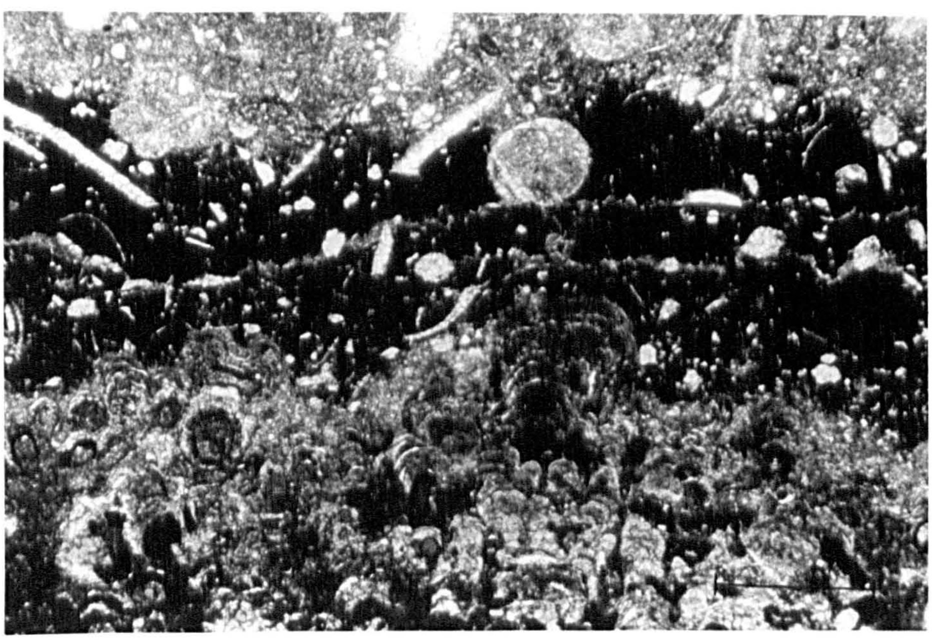


Fig. 7.4 Colloform structures developed in a ferromanganese coating of a limestone clast. Lower Frasnian. Mont Peyroux, Montagne Noire. Peel S 22958. Scale bar = 200μ.

Fig. 7.5 Ferromanganese crust with band of iron and manganese enrichment outside the crust, which has partly replaced the host sediment. Colloform structures and foraminifera developed in the crust. Lower Famennian. Combe D'Izarne, Montagne Noire. Thin section S 22965. Scale bar = 0.5 mm.





segregations of yellow or brown carbonate have developed with opaque areas between the segregations (Figs. 7.6 and 7.7). A black or dark brown band is developed around some nodules (Fig. 7.2 and 7.5) where ferromanganese has migrated outwards and replaced the sediment. Evidence can be found that the colloform structures too are replacing earlier fabrics in the matrix. A concentric banding of light brown and black bands commonly occurs (Figs. 7.8 and 7.9).

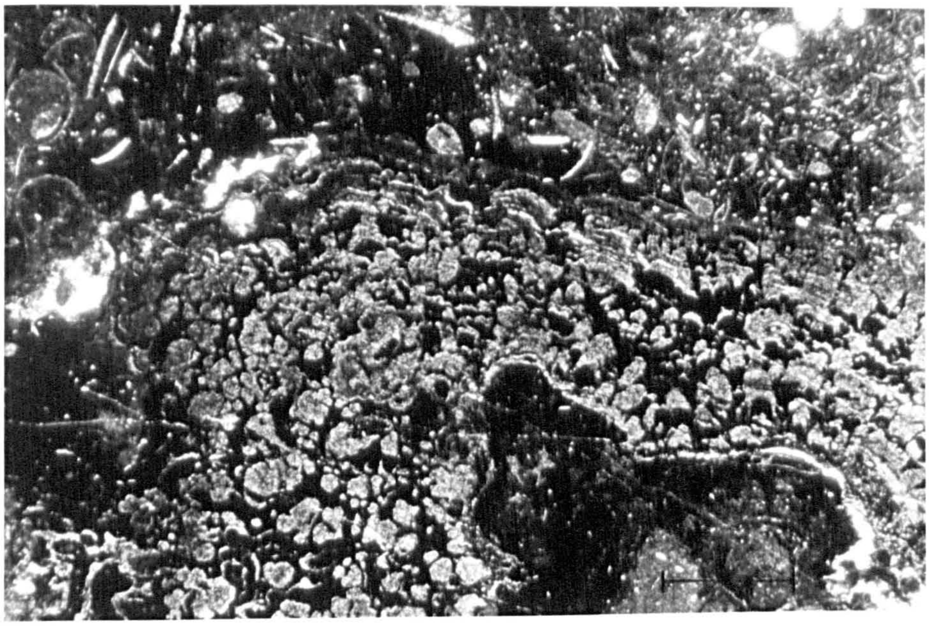
Shell fragments and limestone intraclasts form the centres to the nodules and these are usually intensively bored, particularly the shell pieces (Figs. 3.105 (p.183), 3.113 (p.195) and 7.8). The borings mostly have a diameter of about 0.1 mm and are usually filled by a brownish micrite (p. 193). Limonite is commonly developed around the walls of the borings (3.113 (p.195)) indicating that the shell fragment was lying on the sea floor for sometime before the ferromanganese crust began to develop.

The limestone fragments in the centres of nodules are usually identical to the host sediment, but the clasts may be of a slightly different lithology. Volcanic fragments have not been observed as centres (cf. Jenkyns, 1970a) and volcanism is absent in the Montagne Noire during the Upper Devonian (Gèze, 1949). The limestones in which the nodules occur is a typical pelagic facies, a condensed, hematitic sediment containing a dominantly pelagic fauna of goniatites, conodonts, cricoconarids, thin-shelled bivalves and restricted benthos.

One of the most interesting features of these ferromanganese nodules is the intimate association with the encrusting foraminifera Tolypammina sp. This foraminifera is common at some horizons in the Schwellen limestones of Germany (p.187, and Eickhoff, 1970). Tolypammina is found encrusting the shells and clasts within the ferromanganese coating (Figs. 7.9 and 7.11), showing that the ferromanganese coating is syngenetic. The test of this fixosessile foraminifera is very irregular in arrangement and consists of a meandering tube, occasionally branching, with a diameter of 100–200 μ . They are circular in cross-section. Normally the test is composed of quartz grains (p.189) but here, the tests usually consist of iron minerals (Fig. 7.12) which may have replaced original quartz grains. Shell fragments, conodonts and crinoid ossicles commonly occur within the manganese crust.

Fig. 7.6 Ferromanganese encrustation around limestone clast with segregations of yellowish brown carbonate surrounded by opaque material. Colloform structures developed around the outside. Lower Famennian. Combe D'Izarne, Montagne Noire. Thin section S 22964. Scale bar = 1 mm.

Fig. 7.7 Complete segregations of yellowish brown carbonate developed in a ferromanganese encrustation. Lower Famennian. Combe D'Izarne, Montagne Noire. Thin section S 22964. Scale bar = 0.75 mm.



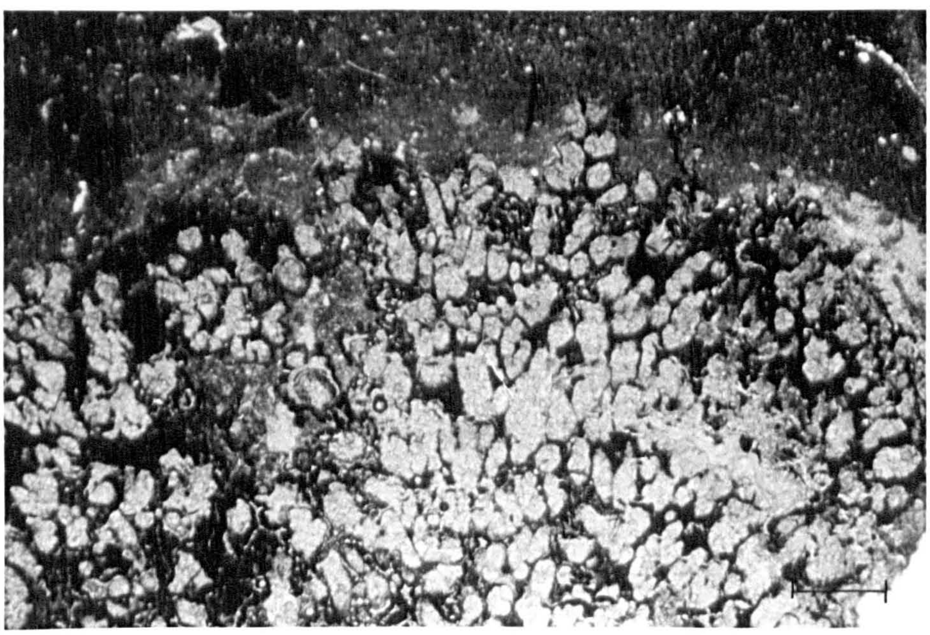
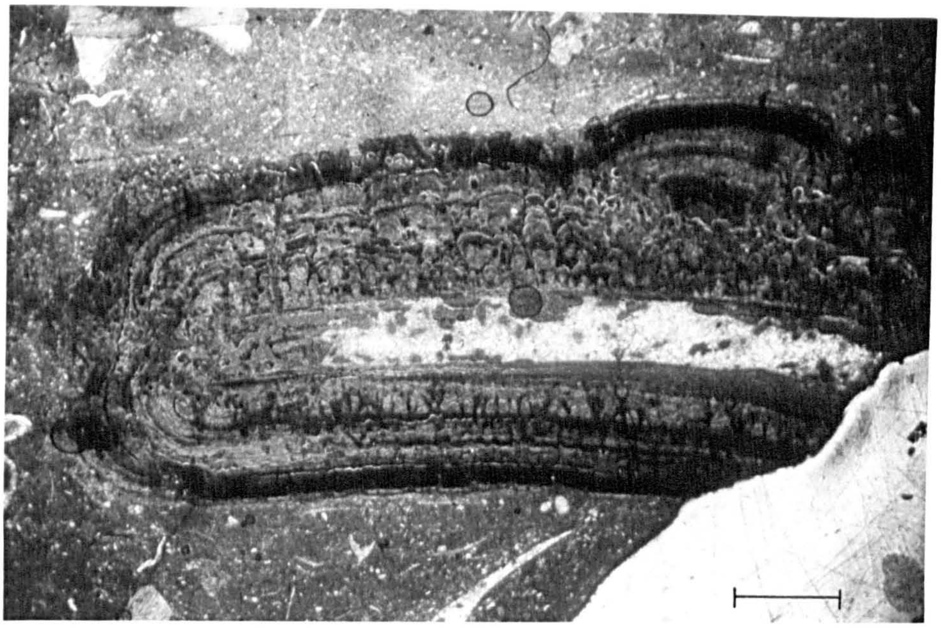


Fig. 7.8 Bored and partly micritized shell with a ferromanganese crust. Light and dark banding well developed in some parts, but nearer the shell this is replaced by colloform structures. Lower Famennian. Combe D'Izarne, Montagne Noire. Thin section S 22968. Scale bar = 1 mm.

Fig. 7.9 Part of a ferromanganese crust with light brown/dark brown banding. Foraminifera present within some of the bands. Lower Famennian. Combe D'Izarne, Montagne Noire. Thin section S 22967. Scale bar = 0.5 mm.



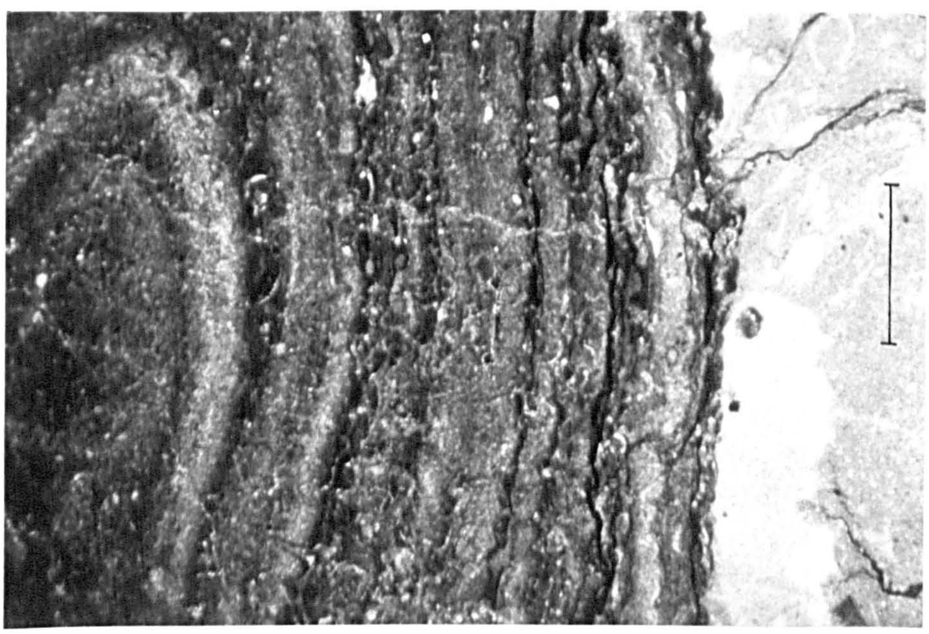


Fig. 7.10 Shells bored and partly micritized, with ferromanganese encrustations. Concentric banding and colloform structures are present. Lower Famennian. Combe D'Izarne, Montagne Noire. Peel S 22952. Scale bar = 1 mm.

Fig. 7.11 Ferromanganese and foraminifera encrustations around a bored bivalve. Some of foraminifera have tests composed of quartz grains, others have tests of ferromanganese minerals. Lower Famennian. Combe D'Izarne, Montagne Noire. Thin section S 22967. Scale bar = 1 mm.

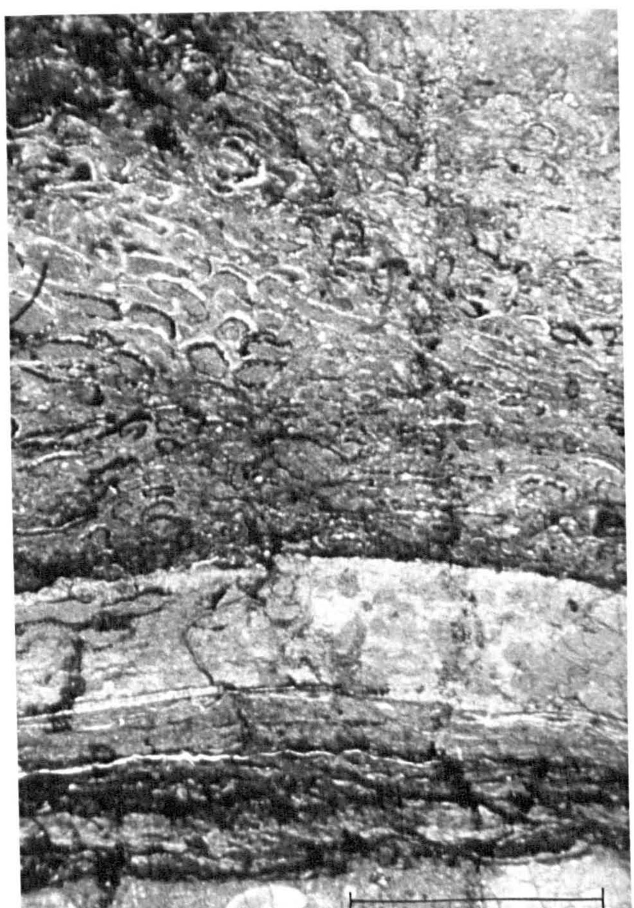
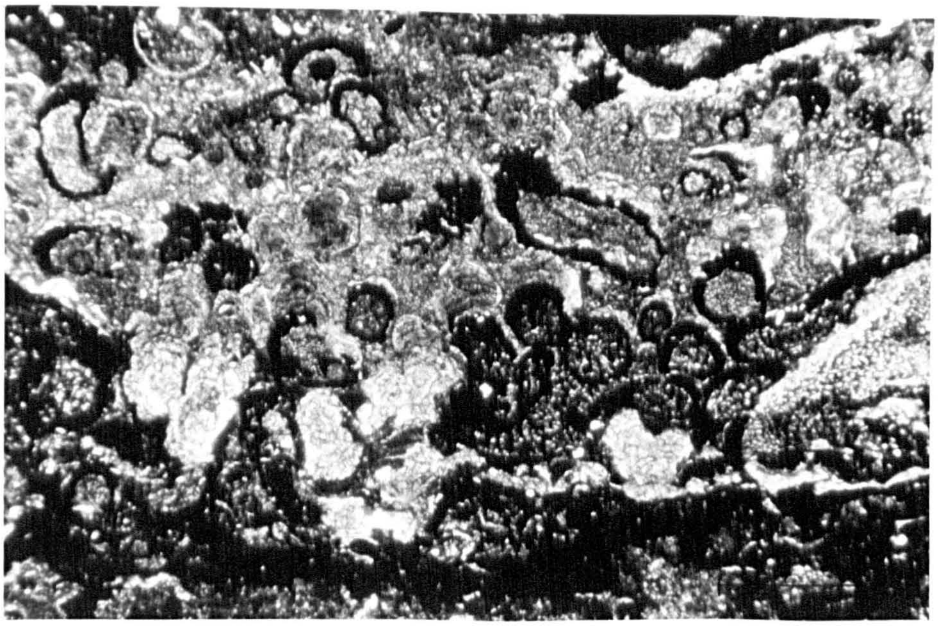
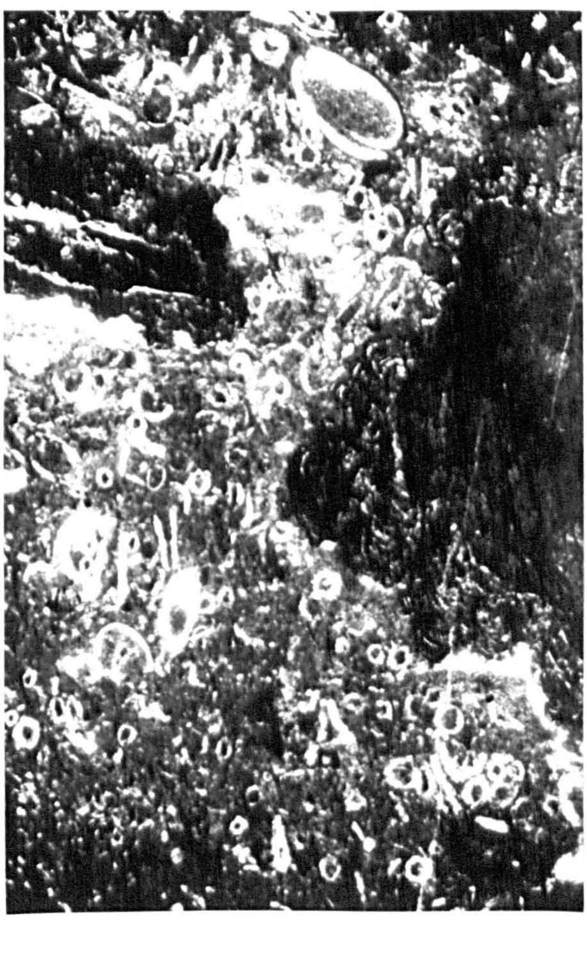


Fig. 7.12 Foraminifera within a ferromanganese coating with tests composed of ferromanganese minerals. Lower Famennian. Combe D'Izarne, Montagne Noire. Thin section S 22965. Scale bar = 0.5 mm.

Fig. 7.13 Ferromanganese encrustations with colloform structures developed around a limestone intraclast and skeletal fragment. Lower Frasnian. Bicken, Dill Syncline, Germany. Thin section S 22950a. Scale bar = 1 mm.





Some of the shells have a very irregular surface on which the crust has been deposited (Fig. 3.113 & 195) and the limestone clasts too may be angular (Figs. 7.2 and 7.3). Although there is no other evidence available, these irregular surfaces could be the result of subsolution (Hollmann, 1964), which normally occurs as the result of very slow sedimentation under oxidising conditions. Iron stained shells have been reported from the Atlantic (El Wakeel and Riley, 1961) in an area of negligible deposition where the sediment is in a very oxidized state. The same sort of conditions - slow sedimentation and an oxidizing environment must also have existed during the deposition of the griotte.

Wendt (1969) described 'microreefs' of Tolypammina, occurring on hardgrounds encrusted with limonite, in the Jurassic Hallstatt facies of Steiermark (Austria). They occur at very condensed horizons where subsolution of the seabottom had taken place. From comparisons with other fossil occurrences of encrusting foraminifera and from Recent examples, Wendt considered a depth of deposition less than 200 m (infra-circa littoral zone) for this Jurassic pelagic limestone. A similar depth is envisaged for the nodules from the Montagne Noire.

Ferromanganese encrustations from the Schwellen limestones in Germany were only found at one locality, Bicken in the Dill Syncline. The nodules, occurring in a cricoconarid-rich grey Flaser limestone, are associated with a hardground (p. 77) and are present in the top few centimetres below a corrasion surface. These encrustations show the same features as those from the Montagne Noire. The ferromanganese encrusts limestone intraclasts and shell fragments (Fig. 7.13) and foraminifera are also present. Pyrite cubes occur in the sediment with the nodules.

Section 7.2 Geochemistry of the nodules

Twenty-one nodules and encrustations were analysed by X-Ray fluoresence to determine the amount of iron, manganese and nickel present. The material presented difficulties since the nodules are embedded in a hard limestone and a fair amount of this was unavoidably incorporated in the sample. The nodules too are rather small and not enough material was available to make the tetraborate beads which one normally uses for X.R.F. work. Consequently, determinations were made on the whole powdered sample and then the limestone was dissolved out using dilute acid. The Fe, Mn and Ni values were then recalculated for the acid insoluble fraction. Results are shown in Table 1.

It is immediately apparent that the Fe and Mn values are very variable and some of this is clearly the result of different clay and silt contents of the samples. However, two analyses of a Jurassic ferromanganese encrusted hardground from the Sonnwendge-birge, Austria (Fig. 7.14) gave different results. The Fe content was 30% in one case, and 2% in the other. Much of the variation in the Devonian nodules could be a true reflection of original compositional differences. Previous workers have remarked on the great variation even between adjacent nodules (Price, 1967, Jenkyns, 1970).

The average Fe and Mn contents of the nodules are 1.8% and 9.2% respectively and the Mn/Fe ratio is 0.2. Comparison with analyses for Recent nodules (e.g. Mero, 1965; Price and Calvert, 1970) shows that the Devonian examples are depleted in Mn relative to Fe. Some fossil nodules from Sicily have low Mn contents (Jenkyns, 1967; 1970) as do the two analyses of Jurassic age made here (Table 1). Post-depositional mobility of Mn has been well established (Lynn and Bonatti, 1965) and it is probably this which gives rise to Mn depletion. This occurs particularly where reducing conditions exist in the sediment, with the result that buried nodules end up with a lower Mn/Fe ratio than their original value. Black and brown 'halos' occurring around some nodules indicate migration. The Ni content (0.035 %) is also low compared with other anaylses.

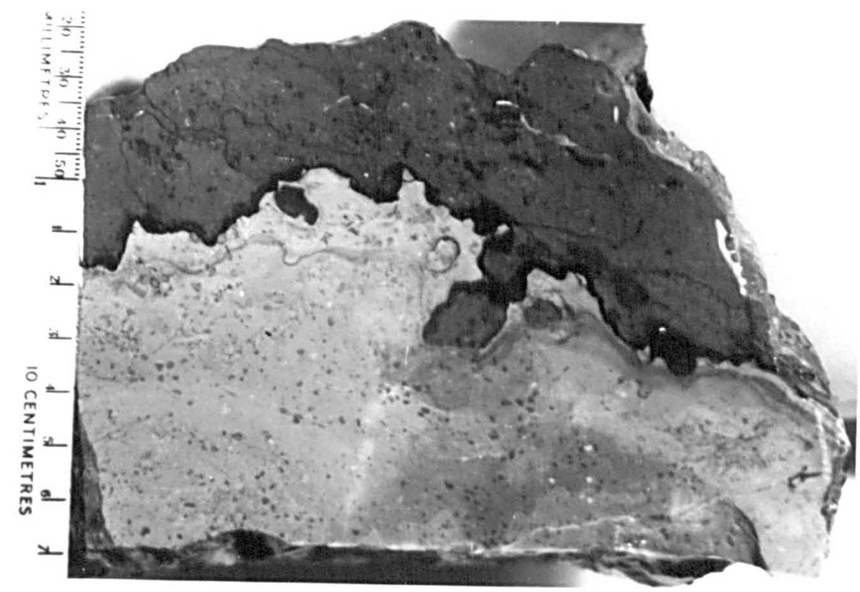


Fig. 7.14 Ferromanganese encrusted hardground surface. Subsolution of the underlying shallow water limestone has occurred and produced the irregular surface. Red pelagic limestone overlies the subsolution surface. Jurassic. Maurach, Sonnwendgebirge, Austria. Polished surface S 23068. Scale as shown.

Locality	No.	Fe %	Mn %	Ni %	Mn/Fe	Mn/Ni
Combe D'Izarne	1	15.33	0.81	0.120	0.05	6.7
"	2	7.30	1.10	0.024	0.15	46.8
"	2	7.25	0.91	0.017	0.13	54.0
"	2	6.47	1.18	0.020	0.18	59.9
"	2	9.82	0.98	0.022	0.10	43.8
"	2	8.33	0.99	0.022	0.12	45.7
"	2	4.51	2.00	0.038	0.44	53.2
"	2	11.25	1.18	0.036	0.11	32.9
"	2	12.75	3.49	0.044	0.27	79.3
"	2	9.45	1.72	0.031	0.18	54.9
"	2	8.15	1.37	0.036	0.17	38.4
"	2	10.51	2.09	0.041	0.20	50.9
"	3	5.55	2.11	0.027	0.38	78.6
"	3	5.00	2.09	0.038	0.42	54.4
"	3	6.35	1.56	0.023	0.25	67.0
"	3	4.98	0.55	0.103	0.11	5.4
"	4	7.61	1.42	0.017	0.19	82.1
"	4	9.30	1.09	0.023	0.12	48.4
"	4	9.32	1.46	0.028	0.16	51.1
Mont-Peyroux	32	8.85	3.94	0.055	0.45	71.2
"	32	31.23	5.08	0.040	0.16	127.2
Alps (Jurassic)		1.96	0.96	0.053	0.49	18.1
"		30.96	1.17	0.087	0.38	13.3

Table 7.1. Results of XRF analysis of ferromanganese nodules from the Montagne Noire and the Alps.

Correlation coefficients for Fe, Mn and Ni (with 19 degrees of freedom) are as follows:-

Mn	<u>0.605</u>	
Ni	0.175	-0.063
	Fe	Mn

Cronan (1969) reports a positive covariance between Mn and Ni but there is no indication of this here. There is however a positive correlation between Fe and Mn (significant at the 99% level). However, as shown below, on the microscale the covariance is negative. Since the ferromanganese encrustations exhibit banding or colloform structures, the inter-element relationships are best examined with the electron probe. Traverses were made across five nodules to determine the variations of Fe, Mn, Si and Ca. The results with the locations of the traverses marked on photomicrographs are shown in Figs. 7.15 to 7.19.

The traverses show clearly that in the ferromanganese crust, Fe is positively correlated with Si, but negatively correlated with Mn. Mn covaries positively with Ca. The crust is commonly made up of light and dark brown bands, and the microprobe analyses show that the light brown bands are generally high in Ca and Mn, but depleted in Si and Fe. The dark areas on the other hand are enriched in Si and Fe. Traverses across complicated textural areas, where colloform structures are present still show these inter-element relationships. There is a greater variation of Si generally and this must be due to the presence of arenaceous foraminifera within the crust. The probe traverses of Jenkyns (1970) across Jurassic ferromanganese nodules and crusts show a similar negative correlation of Fe and Mn. Cronan and Tooms (1969) for Recent nodules with colloform segregations showed that Mn (and Ca) is concentrated within the segregation and that Fe is more evenly distributed. Further diagenetic processes in the Devonian nodules may have accentuated the Fe and Mn distribution. X-Ray diffraction analyses of these nodules did not conclusively reveal the Fe and Mn mineralogy. Hematite was definitely present, but for Mn only a trace of rhodochrosite was obtained in one sample.

Fig. 7.15 Electron microprobe traverse across a ferromanganese encrustation around a bored shell fragment. The full scale deflection for calcium is approximately 43%, for silicon 14%, for iron 2% and for manganese 4%. Lower Famennian. Combe D'Izarne, Montagne Noire. Thin section S 22967. Scale as shown.

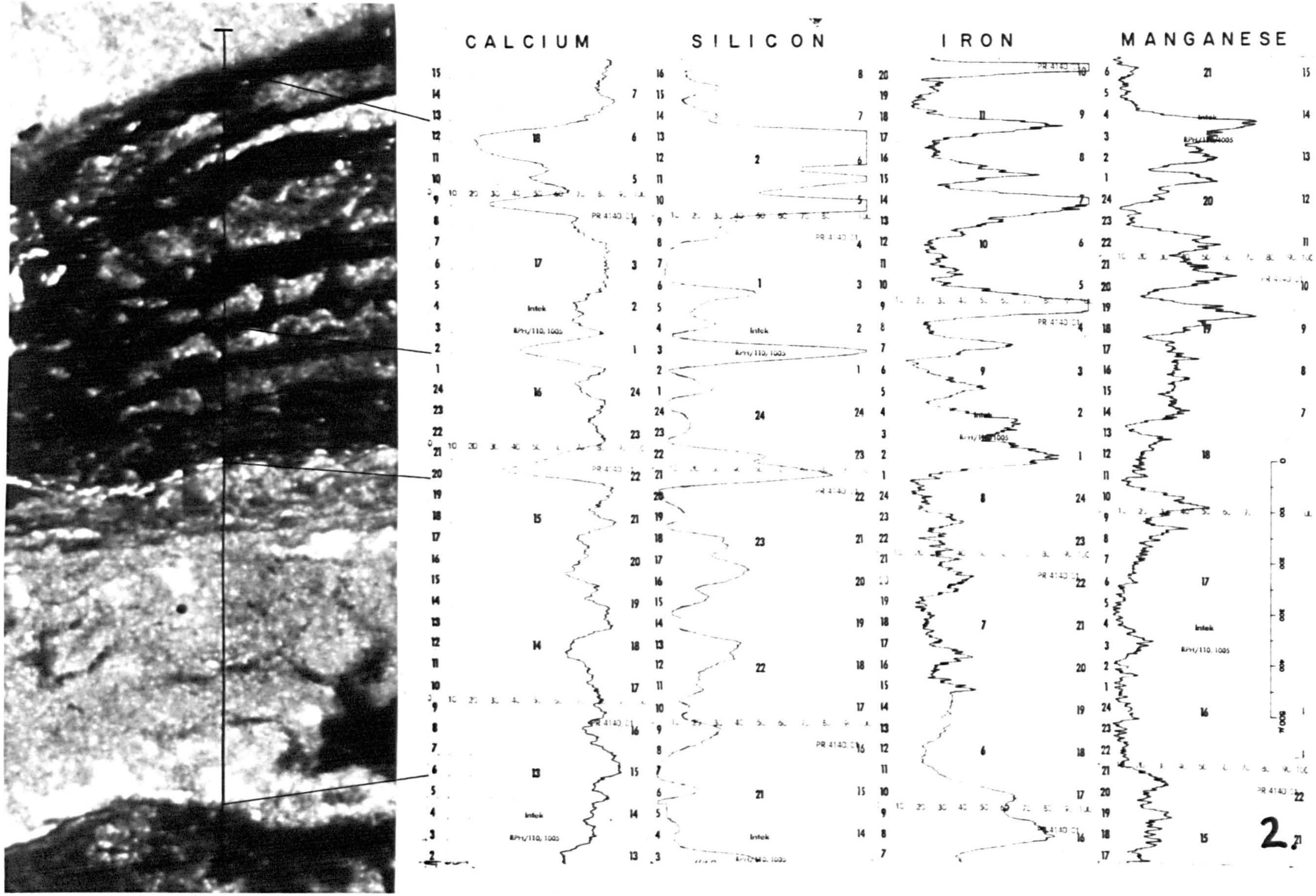

CALCIUM
SILICON
IRON
MANGANESE
2

Fig. 7.16 Electron microprobe traverse across a ferromanganese encrustation around a bored shell fragment. The full scale deflection for calcium is approximately 43%, for silicon 14%, for iron 2% and for manganese 4%. Lower Famennian. Combe D'Izarne, Montagne Noire. Thin section S 22967. Scale as shown.

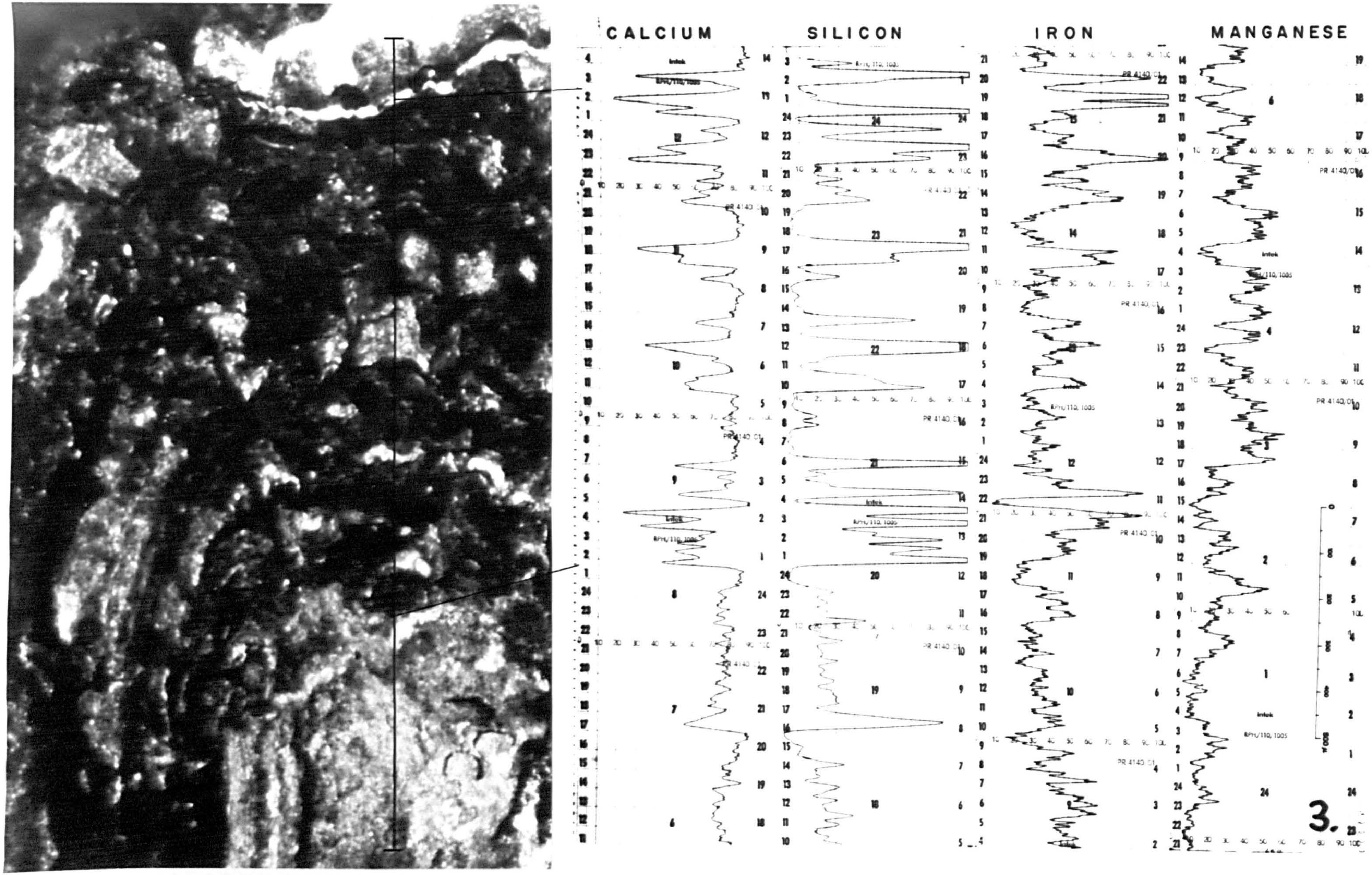


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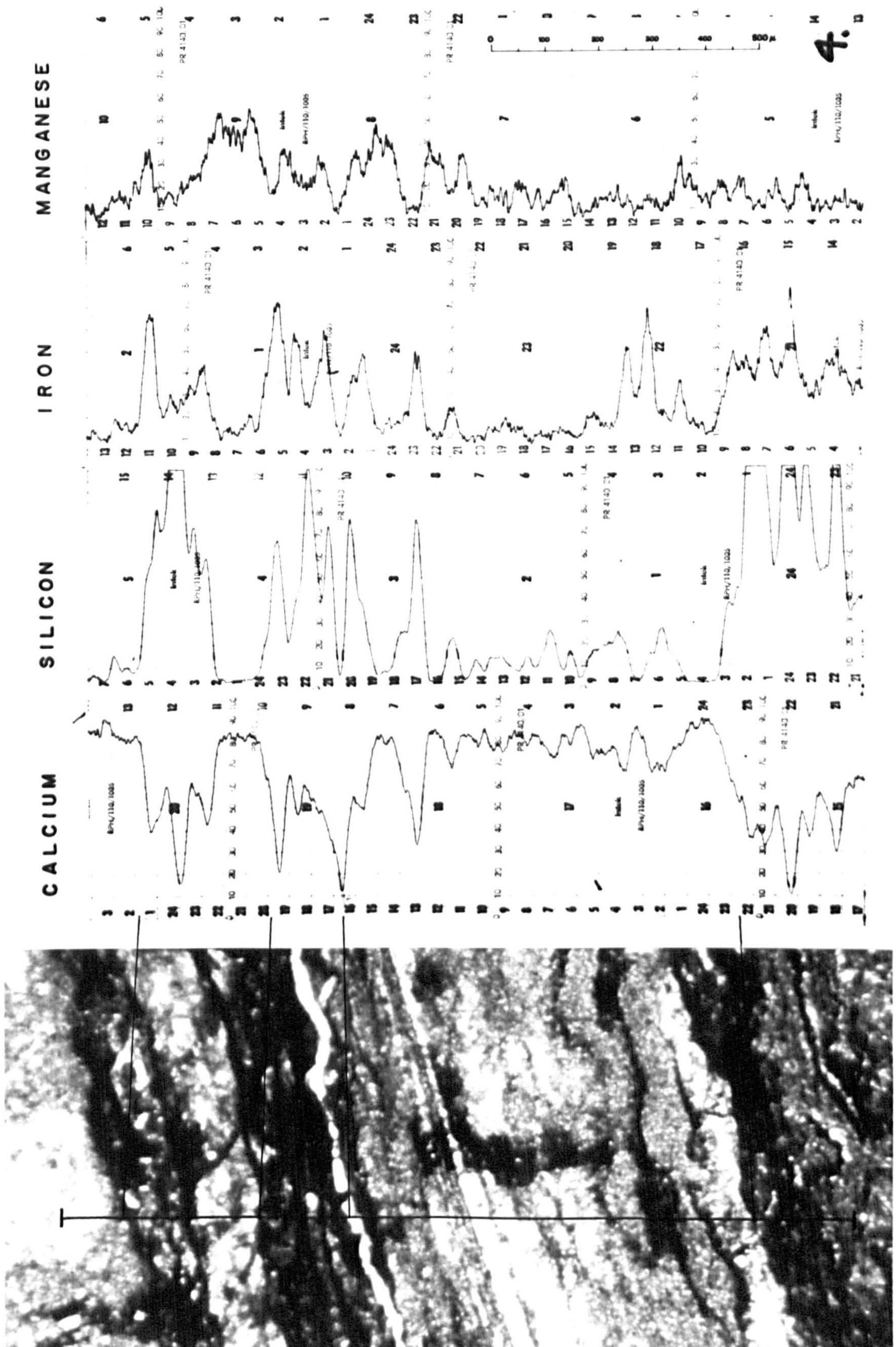
Fig. 7.17 Electron microprobe traverse across a ferro-manganese encrustation around a bored shell fragment. The full scale deflection for calcium is approximately 43%, for silicon 14%. for iron 4% and for manganese 4%. Lower Famennian. Combe D'Izarne, Montagne Noire. Thin section S 22967. Scale as shown.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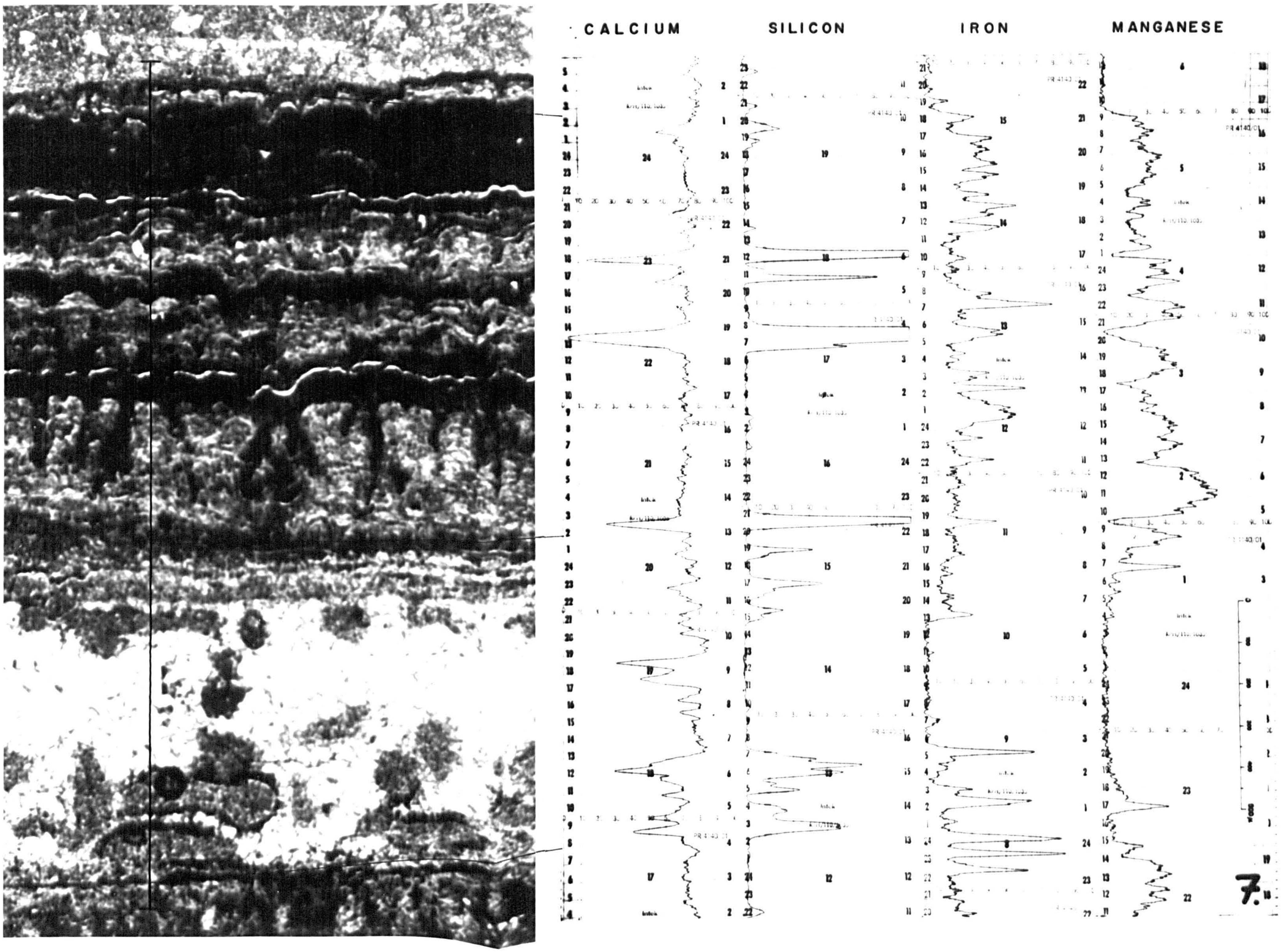

Fig. 7.18 Electron microprobe traverse across a ferromanganese encrustation around a bored shell fragment. The maximum deflection for calcium is approximately 43%, for silicon 67%, for iron 4% and for manganese 8%. Lower Famennian. Combe D'Izarne, Montagne Noire. Thin section S 22968. Scale as shown.

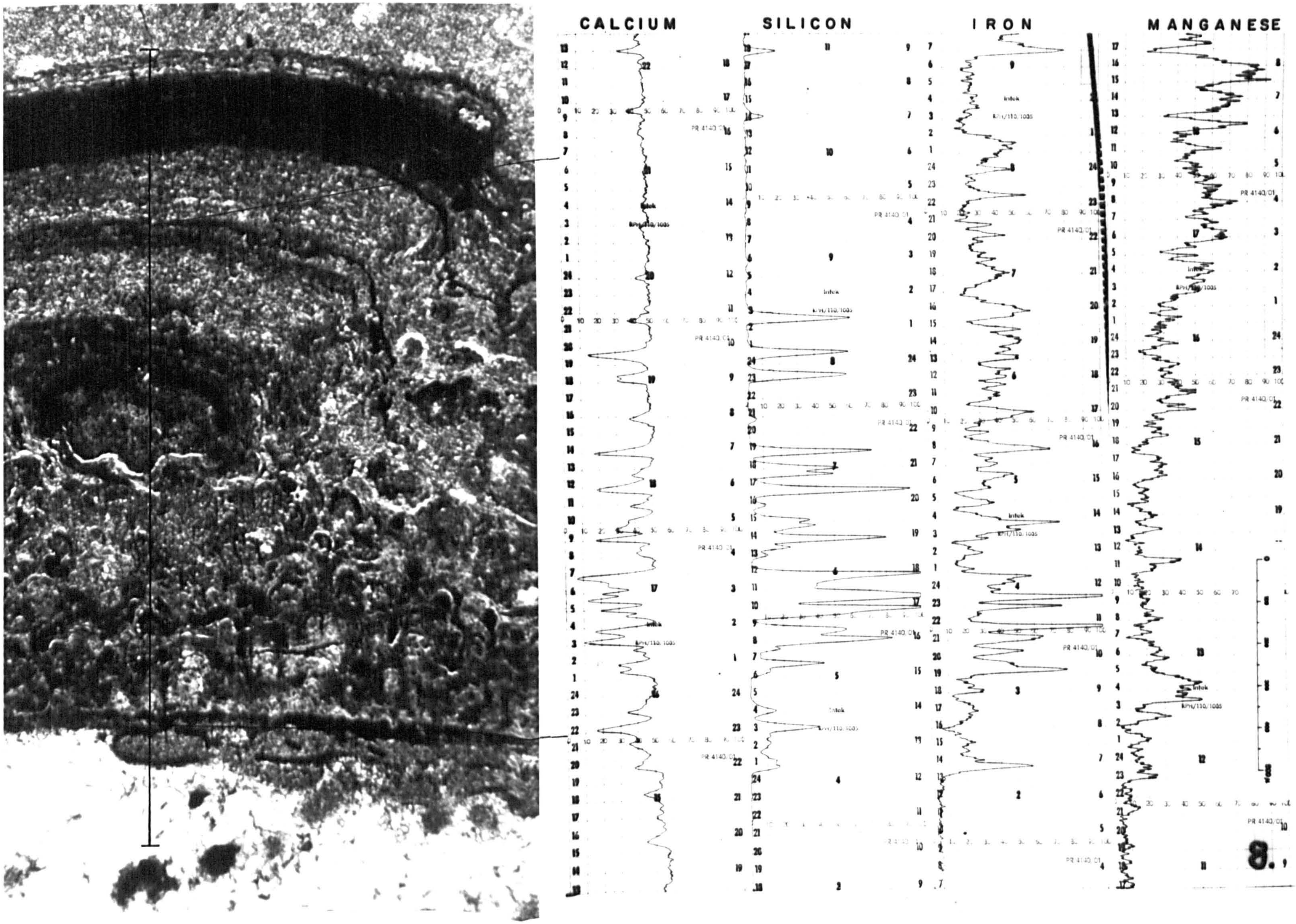

Fig. 7.19 Electron microprobe traverse across a ferromanganese encrustation around a shell fragment. The maximum deflection for calcium is approximately 86%, for silicon 27%, for iron 4% and for manganese 4%. Scale as shown.

The concentration of the ferromanganese minerals must be below the detection limit for X-Ray diffraction (about 10%). The positive covariance of Mn and Ca might further indicate the presence of Mn in the carbonate phase. Perseil (1968) has recorded Mn minerals from the Montagne Noire, which are found in Recent Mn nodules (todorokite and manganite).

The microprobe traverses also revealed interesting information about the shells which are encrusted. Normally, the shells are very low in Si, Fe and Mn, but where the shell has been micritized (? through boring organisms) then Fe is higher and Si may have some strong peaks (Fig. 7.18). Where there is a little Fe or Mn present in the shell, these still vary negatively (Fig. 7.16). Borings in the shell have lower Ca than the rest of the shell, and higher Si. Commonly, the borings have a brown border (p.294) which shows a clear enrichment in Fe, and to a lesser extent in Mn (Figs. 7.18 and 7.19).

Section 7.3 <u>Sedimentological significance of the ferromanganese encrustations</u>

Direct comparisons with Recent nodules are not really possible, since the mobilities of Mn, Fe and Ni are different and the ratios of these elements have altered. The sediment containing the nodules however would point to an open 'oceanic' environment some distance from the influence of land. Price and Calvert (1970) consider the rate of sedimentation a most important factor in the formation and geochemistry of Recent nodules since it is this which determines the amount of organic matter in the sediment. Most of the griotte is coloured red through the presence of hematite, which is probably a diagenetic alteration product of goethite (as considered by Hallam (1967) and Jenkyns (1970b) for the red limestones of the Alpine Jurassic). The presence of hematite in the griotte, indicating oxidising conditions, suggests that there was very little organic matter in the sediments originally. This could be caused by low organic productivity in the overlying waters, or a current swept sea-floor.

It is not possible to determine the rate of sedimentation for the Combe D'Izarne section since the exposure is not continuous and the succession has not been dated accurately with conodonts. Prof. M.R. House (pers comm. 1970) gives a figure of 28m for the griotte (Frasnian and Famennian) at La Serre, one km south of Combe D'Izarne, and this would give an approximate rate of 2.5mm/1000 years for the Upper Devonian. This figure is comparable with pelagic sedimentation rates today.

Nodules and encrustations were only observed at one locality (Bicken) in the Schwellen limestones of Germany. This is somewhat surprising since the sedimentation rates are similar or in some cases much lower. The limestones there are normally grey and contain pyrite. Where hematite is present (e.g. Adorf am Martenberg (p.26)) then this is diagenetic and has come from hydrothermal solutions connected with a basalt extrusion below (Bottke, 1965). Shells are frequently bored in the German limestones and encrusting foraminifera are present (p.187) but apart from the Bicken examples, they are not encrusted with ferromanganese. The Bicken encrustations are a special case since they are associated with a hardground (p.77). Enrichments of Mn and Fe might be expected in such a situation, and occur at a similar hardground in the Montagne Noire (p.77). One possible explanation which could account for the absence of Fe and Mn is that there was a much higher organic content of the sediment. On burial, this would create a reducing environment in the sediment which would enhance the mobility of manganese (and iron to a lesser extent) (Lynn and Bonatti, 1965). Dissolution on burial, has been suggested by Manheim (1965), and several authors have noted the apparent absence of manganese nodules in the reduction zone of Recent pelagic sediment cores (e.g. Bonatti and Nayudu, 1965). Organic productivity in seawater is highest in areas of upwelling, particularly around land areas, though the organic content of marine waters today, below 300m, is uniform (Menzel, 1967). Perhaps then, the proximity of the Old Red Continent to the north of the Rheinisches Schiefergebirge and the Harz Mountains affected the Schwellen sediments in this way. With the Montagne Noire griotte, low organic content and perhaps greater current activity, could account for the marked enrichment in ferric iron.

Two sources are generally considered for the manganese, 1) volcanism, 2) continental run-off (Arrhenius et al, 1964; Bonatti and Nayudu, 1965). Volcanics are developed at a number of horizons in the Upper Devonian of Germany and the Schwellen sediments sometimes occur directly on top of pillow lavas. On the other hand, volcanism is absent in the Montagne Noire (Gèze, 1949). Ronov and Ermishkina (1959), from many analyses of limestones on the Russian platform, showed that the Mn content decreases with distance from land, and that most Mn is concentrated in the coastal region. The pelagic facies in the Harz Moutains and Rheinisches Schiefergebirge was deposited a few hundred kilometres or less, from the shore-line (Schmidt, 1962). The presence of volcanics and proximity to a continent should favour the development of ferromanganese nodules. A high organic content is the most likely factor to have prevented the formation of ferromanganese nodules in the German Upper Devonian.

The Schwellen limestones of Germany are unlike many Recent pelagic sediments (and fossil ones) which are enriched in certain minerals, mainly as a result of the slow deposition. An enrichment of iron and manganese in the soluble carbonate fraction does occur (chapter 6) compared with other limestones. The absence of ferromanganese nodules has been considered, but other minerals, glauconite and phosphate, as occur in the condensed Ordovician *Orthoceras* limestones (Lindström, 1963) and at certain horizons in the Chalk, are also rare or absent in the Schwellen limestones. Phosphatic nodules were only found in one limestone sample (from Eibach, a volcanic rise in the Dill Syncline) and glauconite was not found at all.

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CHAPTER 8

Conclusions

A detailed sedimentological analysis has been presented of Devonian pelagic sediments and apart from the new material specifically on these sediments (summarized in this chapter) results of wider implication include the following:-

1. Lithification of the Schwellen limestones occurred during early diagenesis and in some cases took place on the sea floor, with the formation of hardgrounds. Contact with meteoric water, which has been previously invoked as necessary for cementation of carbonates, did not occur until uplift during the mid-Carboniferous.

2. Evidence has been presented on the origin of fibrous radiaxial calcite (a common void-filling calcite in many limestones, which has previously been considered a primary precipitate from fresh, marine or connate water as calcite or aragonite) showing that it is a replacement of an early acicular cement (a note on this is in press, Kendall and Tucker, 1971 and appended, p403 to 405).

3. The presence of Devonian ferromanganese nodules (the oldest so far recorded) shows that these nodules are not dissolved through burial as suggested by various authors.

4. The analyses of the Devonian pelagic sediments and the absence of ferroan calcite and dolomite in Schwellen limestones suggest that migration of connate waters from the basin to rise did not occur. Such a process has been proposed to account for the lithification of shelf carbonates.

From the papers appended on S.W. England, perhaps the most important contribution is that the German conodont zonation for the

Upper Devonian is applicable to the sediments of S.W. England (Tucker and Straaten, 1970). Straaten and Tucker (1971, in press) show that detailed micropalaeontological and sedimentological work may reveal that geological situations are more complicated than at first appear. Tucker (1969) describes crinoidal turbidites with a central dune bedded division (not visible in the field) which is apparently absent in terrigenous turbidites.

The main points concerning the Schwellen limestones, slope and basin sediments arising from the work presented here are as follows:-

1. Devonian cephalopod limestones accumulated on submarine rises within the Rhenish geosyncline, a depositional trough from Lower Devonian to Lower Carboniferous which is considered to be intracratonic. The pelagic limestones were developed in three situations, above basement rises (geanticlines), submerged 'reefs' and on volcanic rises. Thicker deposits of shales with nodules and nodular limestones locally with slumped and reworked beds were deposited on the flanks of the Schwellen. Silty shales, locally with turbidites, were deposited in the basins.

2. The depth of deposition of the pelagic limestones probably did not exceed a few hundred metres and in some cases was 50 m or less. The shallower depths are suggested from foraminiferal/algal nodules and the occurrence of fixosessile arenaceous foraminifera indicate depths down to 200 m. The slope region probably varied in width and in the Harz Mountains was about 20 km wide. The basinal sediments probably accumulated at depths in the region of 1000 m.

3. The origin of the carbonate sediment is not known but by comparison with Recent carbonates it was probably biogenic . Many of the limestones are homogeneous microsparites, and the original textures have been largely obliterated by recrystallization. In some cases the original sediment appears to have been composed of comminuted skeletal material, comparable with the silt-sized pelagic carbonate sediments occurring on the Yucatan Shelf today. The Jurassic pelagic carbonates of the Alps on the other hand are generally finer grained (micrites) and

are comparable rather with Recent coccolith oozes.

4. Current activity in the Schwellen limestones is indicated by the presence of laminated carbonates and fossil concentrates, Cricoconarids, thin shelled bivalves and ostracods may form micro-coquinas, and cricoconarids may be packed one inside the other also indicating current activity. Thin units of terrigenous silt and clay commonly occur in some limestones and represent deposition from low density suspension currents or nepheloid layers.

5. Hardgrounds locally occur in the cephalopod limestones and indicate synsedimentary cementation. Corrasional hardgrounds, similar to those in the British Jurassic and the Russian Ordovician and Devonian, cut sheet cavity cements, skeletal material and ferro-manganese encrustations. Hematization of microfossils and the host sediment has occurred in the case of one hardground. Cryptohardgrounds, having surfaces with a relief of up to 1 cm, which are encrusted with fixosessile foraminifera, show evidence of subsolution. They were only detected in thin section since the sediment above and below the cryptohardground surface is similar and did not allow differential erosion to take place.

6. Sheet cracks of various shapes and sizes, some resembling Stromatactis sheet spars occur in the Schwellen limestones and are filled by internal sediment, now microsparite, and radiaxial calcite. The cracks are considered to have formed by shear failure on a slope. In one instance, where a sheet crack is cut by a corrasion surface, radiaxial calcite has nucleated from the surface, and has not been cut by the abrasion, showing that the radiaxial calcite has replaced an earlier cement which filled the cavity. A note on this is in press (Kendall and Tucker, 1971).

7. Early cementation of cricoconarid microcoquinas occurred through the development of syntaxial overgrowths, analogous to the cementation of pteropod oozes at the bottom of the Red Sea. The fibrous overgrowth crystals showing some similarities with radiaxial fabric, are a replacement of an early acicular cement or the host sediment. Thin sheet cracks formed through this cementation are filled by kite-shaped fibrous crystals and internal sediment also rich in cricoconarids.

8. Neptunian dykes of various sizes penetrate the Schwellen limestones and are commonly filled by internal sediment similar to the host sediment. In some cases the host sediment was lithified before the filling of the dyke. Dykes may have formed in a similar way to the sheet cracks, or through small scale tectonic movements.

9. Lack of compaction and certain sedimentary features (hardgrounds, limestone intraclasts, sheet cracks and neptunian dykes) indicate early lithification for the cephalopod limestones in the submarine environment. Contact with meteoric waters did not affect these pelagic limestones until uplift at the end of the Lower Carboniferous.

10. The late diagenesis of the Schwellen limestones, is similar to that for other fine grained limestones. Recrystallization has led to the development of microsparite and patches of coarse microspar and pseudospar. Aggrading neomorphism may have been important in reducing skeletal fragments to microsparite. Syntaxial overgrowths occur around cricoconarids and some crinoid fragments.

11. Late diagenetic and tectonic pressure solution, the latter coincident with the development of cleavage, have given rise to numerous thin streaks of clay and stylolites in the Schwellen limestones.

12. Ferromanganese nodules and encrustations were discovered in the Upper Devonian griotte of the Montagne Noire and at one locality in Germany. A concentric light and dark banding is commonly developed and in some cases this is replaced by colloform structures and segregations. The encrustations are depleted in manganese, iron and nickel relative to Recent nodules. The encrusting foraminifera *Tolypammina* occurs within the ferromanganese crusts. Glauconite was not found and phosphatic nodules are rare in the Devonian limestones.

13. The fauna of the limestones is predominantly pelagic, consisting mainly of ammonoids, thin shelled bivalves, cricoconarids and conodonts. Foraminiferal/algal nodules occur at one locality and suggest shallow depths (about 50 m). Arenaceous foraminifera encrust shells, conodonts and cryptohardgrounds, and suggest sublittoral depths less than 200 m.

14. The slope facies is characterised by slumped and reworked sediments and in most cases the angle of slope was 5° or less. Steeper slopes and fault escarpments existed locally. Calcareous nodules, common in slope lithologies and locally in the basinal shales, were formed by early diagenetic movement of $CaCO_3$. The shales, locally rich in ostracods, typically have a silty lamination although at certain horizons this is absent and black or dark grey laminated carbonaceous shales are developed.

15. The chemical analyses of the Schwellen limestones compare well with Recent pelagic carbonates in being low in magnesium and high in iron and manganese. The Devonian limestones are also significantly different from limestones of other facies. The slope sediments have a more variable chemistry than the rise and basinal samples and tend to be enriched in magnesium, iron and manganese. The enrichments are attributed to the effects of the slope itself and the movement of connate water.

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APPENDIX 1

Details of Localities

The important localities in Schwellen, slope and basin sediments of the Variscan geosyncline are briefly described with map and literature references. Sections of the most complete sequences are given in Figs. A.1 to A.13, generalized from field sections. Localities are arranged alphabetically within the regions Harz Mountains, Rheinisches Sciefergebirge, Montagne Noire, Pyrenees and S.W. England.

Harz Mountains

Acketal, 6 km S of Goslar. Grid Reference R35993 H57455, Fig. 3.3, p. 49.

A small quarry in the Staatsforst Schulenberg exposes the whole Famennian as 5 m of grey Flaser limestone. Frasnian limestones are badly exposed in the bank of forest track nearby. The sediments contain thin graded units of terrigenous material and coarse mosaics occur between some graded bands. References : Born (1912), Fuhrmann (1954), Ziegler (1962).

Buntenbock, 4 km S of Clausthal Zellerfeld, Grid Ref. R35919 H57379.

Badly exposed Upper Devonian occurs in a disused iron-ore pit. The sediments (Flaser limestones, black shales and cherts) were deposited in a depression on the Oberharzer Diabas-Zug.

Hessenweg, Granetal, 5 km SW of Goslar, Grid Ref. R35948 H57499.

An exposure of Upper Devonian slope sediments with sedimentary slumping. Reference: Stoppel (1968).

Hühnertalskopf, 7 km SW of Goslar, Grid Ref. R35941 H57499, Fig. A.1, p.356.

A poor exposure in the bank of forest road shows a very condensed Upper Devonian succession (12 m thick) which is mainly shale ('Special Schwelle' of Früh, 1960). Reference: Müller-Steffen (1965).

Hutthaler Widerwaage, 5 km SE of Clausthal Zellerfeld, Grid Ref. R35960H57398.

A very poor section in the bed of a stream where the Upper Devonian is condensed to 98 cm in grey Flaser limestone with two disconformities. Reference: Meischner and Schneider (1970).

Innerstetal Reservoir (I), 6 km S of Langelsheim, Grid Ref. R35893 H57521.

Exposure of Frasnian basinal shales in railway cutting. Dark grey and black carbonaceous shales locally rich in cricoconarids. Reference: Müller-Steffen (1962).

Innerstetal Resevoir (II), 2 km SW of Langelsheim, Grid Ref. R35884 H57506, Fig. 4.2, p.203.

The railway cutting by the south end of the resevoir contains a good exposure of Famennian lower slope/basinal facies. There is no evidence of sediment movement. The section extends from the Cheiloceras Stufe to the Lower Carboniferous. References: Eickhoff (1961), Müller-Steffen (1962).

Junkernberg, Innerstetal, 2 km SW of Langelsheim, Grid Ref. R35907 H57551, Fig. 4.1, p.202.

A disused quarry showing top part of Famennian and lowest Carboniferous developed in basinal facies. Clymenia Stufe as grey shales with large calcareous nodules. References: Müller-Steffen (1962), Stoppel (1968).

Langestal, 8 km S of Oker, Grid Ref. R36012 H57454.

A good exposure on the north side of an arm of the Okertal reservoir shows Givetian and Frasnian Flaser limestone but the Famennian is absent. A neptunian dyke penetrating 30 cm occurs in the Frasnian limestone. Many limestones are rich in cricoconarids. References: Stoppel (1968), Mohr (1962).

Lautenthal, 10 km S of Langelsheim, Grid Ref. R35890 H57486, Fig. 5.1, p.225.

The section alongside the River Innerste, 500 m NE of the town, shows the Givetian and Upper Devonian developed in slope facies. The lithology is variable and sedimentary slumping and reworking occur. References: Eickhoff (1962), Stoppel (1968), Stoppel and Zscheked (1963).

Margaretten Klippen, Granetal, 4 km S of Goslar, Grid Ref. R35947 H57508, Fig. 5.2, p.226.

This section along the forest road from Granetal to Goslar and in the woods above is in Upper Devonian slope facies, mostly shales with nodules and nodular limestones. Good examples of slump folds are present. References: Meyer (1965), Stoppel (1968), Stoppel and Zscheked (1963).

Rabenklippe, opposite Romkerhalle Waterfall, 6 km S of Oker, Grid Ref. R36017 H57469.

Hillside exposure of Givetian to Lower Carboniferous. Sediments of slope facies showing slumping and reworking. References: Wackendorf (1961), Stoppel and Zscheked (1963), Stoppel (1968).

Riesenbachtal, near Oberschulenberg, 8 km S of Goslar. Grid Ref. R35978 H57447.

Upper Devonian Flaser limestones deposited on the West Harz Schwelle are poorly exposed in the bed and banks of a small stream and in an overgrown quarry nearby, 2 km west of Okertalsperre. A cryptohardground was found in one of the limestone beds. Reference: Fuhrmann (1954).

Romkerhalle Waterfall, 6 km S of Oker, Grid Ref. R36023 H57468.
Fig. A.2, p.357.

Givetian to lower Carboniferous section mostly in grey Flaser limestones. Slightly metamorphosed by the Oker granite, 1 km away. References: Wachendorf (1961), Stoppel (1968).

Sparenberg, 2 km N of Lautenthal. Grid Ref. R35889 H57489.
Fig. A.3, p.358.

Eifelian to upper Famennian strata are exposed on the east bank of the River Innerste, south of the road to Wolfshagen. It is mainly a shaley succession with calcareous nodules and nodular limestones. Slumped blocks of Givetian/lower Frasnian Flaser limestone occur in Frasnian shales, and limestone breccias occur in the Upper Frasnian.
References: Eickhoff (1962), Stoppel (1968), Stoppel and Zscheked (1968).

Sternplatz, 3 km W of Lautenthal, Grid Ref. R35869 H57483.

Poor exposures of Frasnian Flaser limestone, and Famennian shales with nodules. Slumping and reworking of sediments has occurred.
References Stoppel (1968).

Rheinisches Schiefergebirge

Adorf am Martenberg, 19 km ESE of Brilon (Sauerland)
Grid Ref. R34873 H56918, Fig. 3.2, p.48 and Fig. 3.37, p.95.

The quarry at Martenberg is a classic locality for the Frasnian 'Cephalopodenkalk', which was deposited in a small depression on a volcanic ridge, the Hauptgrünsteinzug. Crinoidal limestones pass up into red and grey Flaser limestones containing foraminiferal/ algal nodules, 'cryptohardgrounds', sheet cavities and neptunian dykes. At the top of the quarry ostracod shales of Cheiloceras age occur. Recent references: Ziegler (1959), Bottke (1965).

Belecke, 5 km N of Warstein (Sauerland), Grid Ref. R34538 H57045.

Vertical Frasnian nodular limestones are overlain by horizontal Carboniferous cherts in a small quarry by the junction of the B55 and B516 roads, one km north of the town. The 'unconformity' is due to slumping down the Brilon 'reef' Schwelle during the lowest Carboniferous, connected with that at Drewer. Reference: Schmidt and Plessmann (1961).

Beul, near Eisborn in Honnetal, 7 km SSE of Menden (Sauerland), Grid Ref. R34213 H56938.

Red and grey Flaser limestones deposited above the Arnsberg 'reef' are poorly exposed in thick forest. Typical pelagic carbonate facies. Reference: Paeckelmann (1924).

Bicken, Benner Steinbruch, 8 km SE of Dillenburg (Dill Syncline), Grid Ref. R34576 H56163. Fig. A.4, p.359 and Fig. A.15, p.371.

At this classic locality on the north side of the road between Bicken and Offenbach, the whole Middle and Upper Devonian is condensed to some 60 m of grey Flaser limestone which was deposited on the Ense Schwelle (a subsidiary ridge of the midgeosynclinal rise). The quarry is now mostly filled by water and only the upper Givetian and Upper Devonian part is accessible. Structures present in the Frasnian sediments include ferromanganese nodules, cricoconarid microcoquinas with fibrous overgrowths, a corrasion hardground surface, sheet cracks and neptunian dykes. References: Kockel (1958), Wittekindt (1965).

Blauer Bruch, 2 km SE of Bad Wildungen (Kellerwald), Grid Ref. R34315 H56635

A disused quarry on the road to Wenzigerode exposes Givetian and Frasnian Flaser limestone. The Frasnian limestones are thinly bedded and contain horizontal and cross laminations. Reference: Kockel (1958). Fig. A.5, p.360.

Bonzel am Rhenert, near Grevenbrück, 8 km E of Attendorn, Grid Ref. R34315 H56635.

Givetian and Frasnian Flaser limestones exposed in an overgrown quarry 500 m east of Bonzel contain thin detrital bands of displaced shallow water material, derived from the Attendorn 'reef'. Sheet cracks and fibrous overgrowths around cricoconarids are well developed. 1 m of this Cephalopodenkalk is equivalent to 200 m of 'reef' limestone deposited during the same time a few kilometres away. References: Schmidt and Plessmann (1961), Ziegler (1965).

Calvarienberg, near Kallenhardt, 6 km E of Warstein (Sauerland), Grid. Ref. R34616 H56914.

In thick woods 2 km east of Kallenhardt grey Platyclymenia Flaser limestone is cut by a neptunian dyke 1 m wide by at least 10 m, which is filled by dark grey shales with nodule bands of Clymenia Stufe. References: Schmidt (1921), Schmidt and Plessmann (1961).

Dernbach, near Gladbach, 18 km WSW of Marburg (Dill Syncline), Grid Ref. R34647 H56252.

Famennian basinal shales with large nodules 15 cm in diameter and 3 cm thick are exposed in a road cutting 1 km south west of Dernbach, on the road to Wommelshausen.

Drewer (Provincial Steinbruch), 6 km N of Warstein (Sauerland), Grid Ref. R34557 H57050, Fig. A.6, p.361.

A very large disused quarry on the eastern side of the Drewer to Belecke road exposes sediments of Frasnian to Lower Carboniferous. The Upper Devonian is developed in slope facies, nodular limestones interbedded with black shales. Mass movement of sediment off the Brilon 'reef' Schwelle occurred during the lowest Carboniferous, forming slump folds and sedimentary slides. Culm cherts overlie the moved deposits. The nodular limestones are rich in macrofossils, particularly clymeniids. References: Schmidt and Plessmann (1961), Stascheh (1968).

Dunsbachtal (Eberstein), 20 km SSW of Marburg (Lahn Syncline), Grid Ref. R69310 H11910.

An Upper Devonian succession of Flaser limestones is developed above Givetian and lower Frasnian 'reef' limestone. Typical Cephalopodenkalk. Reference: Bender and Brinckmann (1970).

Eibach, 4 km E of Dillenburg (Dill Syncline), Grid Ref. R34530 H56216.

An overgrown quarry on the south side of the village, exposes Upper Devonian Flaser limestone and nodular limestone which accumulated on a volcanic Schwelle. Phosphatic nodules and sheet cracks occur in the limestones. Reference: Krebs (1960).

Eulenspiegel, 6 km ENE of Warstein (Sauerland), Grid Ref. R34615 H57024.

A very large disused quarry in Upper Devonian pelagic facies on the north side of the road between Kallenhardt and Rüthen, exposes Frasnian Flaser limestones in the lower part and Famennian nodular limestones higher up. These sediments were deposited above the Brilon 'reef'. Neptunian dykes are present in the Frasnian limestones and slumping has occurred in the Famennian sediments. References: Schmidt (1921), Schmidt and Plessmann (1962).

Gaudernbach, 5 km WSW of Weilburg (Lahn Syncline) just north of the village, Grid Ref. R34436 H55913.

A large working quarry in Middle Devonian 'reef' limestone contains vertical neptunian dykes filled by Upper Devonian Cephalopod limestone. Fragments of 'reef' limestone and clasts of fibrous cement occur in the dykes. References: mentioned in Krebs 1968a and 1969.

Grevenbrück, 9 km E of Attendorn, Grid. Ref. R34313 H56653.

An overgrown quarry on the south western side of the town off the B236 road exposes Givetian and lower Frasnian Schwellen limestones which contain thin detrital bands of material derived from the Attendorn 'reef'.

Haingrube, near Haddenberg, 13 km S of Bad Wildungen (Kellerwald), Grid Ref. R35050 H56524, Fig. A.7, p.362.

A poorly exposed Upper Devonian succession in forest cutting shows pelagic sediments developed on top of a volcanic ridge. Some shales and limestones are rich in ostracods and a cryptohardground is developed in a limestone band of upper Platyclymenia age. References: Bering (1967), Schneider (1969).

Kallenhardt, 7 km E of Warstein (Sauerland), Grid Ref. R34609 H56998.

An overgrown quarry on eastern side of the village exposes the Schlagwasserbreccia, a synsedimentary fault breccia containing blocks and conodonts of Middle Devonian to upper Famennian. References: Schmidt and Plessmann (1961), Staschen (1968).

Kattensiepen, 3 km NE of Warstein (Sauerland), Grid Ref. R34580 H57020.

A large quarry along the road between Warstein and Rüthen is working Famennian nodular limestones which were deposited on the flanks of the Brilon 'reef' Schwelle, Sedimentary slumping is developed on a minor scale. Reference: Schmidt and Plessmann (1961).

Kleine Leuchte, near Fischbach, 10 km SW of Bad Wildungen (Kellerwald), Grid Ref. R35057 H56548.

Very poor exposures on the hillside north of Fischbach are of red shales with rare bands of calcareous nodules. These were deposited in the trough between the Ense and Hundsdorfer Schwellen on the midgeosynclinal rise. Reference: Schneider (1969).

Langenaubach, Konstenze Quarry, 7 km SW of Dillenburg (Dill Syncline), Grid Ref. R34433 H56183.

A small quarry (Mamorbruch) at the eastern end of the main

basalt quarry exposes the Upper Devonian in Flaser limestone which was deposited on a volcanic ridge in the Dill Syncline.
Reference: Kockel (1958), Krebs (1966).

Langenholthausen, 17 km S of Menden (Sauerland), Grid Ref. R34219 H56853.

A badly weathered section along a road cutting in Famennian nodular limestones which were deposited on the flanks of the Arnsberg 'reef' Schwelle. Reference: Schmidt and Plessmann (1961).

Löhnberg, 2 km N of Weilberg (Lahn Syncline), Grid Ref. R34476 H55948.

Upper Devonian basinal facies are exposed in an old quarry on the main road from Löhnberg to Weilberg on the right bank of the River Lahn. The Famennian red and green shales are locally rich in ostracods and posidonid bivalves.

Mellen, near Balve, 14 km SSE of Menden (Sauerland), Grid Ref. R34220 H56868.

An overgrown quarry just off the main road 1 km west of Mellen shows Famennian nodular limestones and calcareous nodules in red shales. There is no evidence of slumping and nodules are present in closely packed bands. The sediments accumulated on the flanks of the Arnsberg 'reef' Schwelle. Reference: Schmidt and Plessmann (1961).

Obermarpe, near Cobbewode, 18 km NE of Attendorn, Grid Ref. R34194 H56946.

The Upper Devonian is exposed along a road cutting, and consists of shales with nodular limestones and calcareous nodules. This succession was deposited on the flanks of the Arnsberg 'reef' Schwelle. References: Schmidt and Plessmann (1961), Ziegler (1962).

Öse, near Hemer, 5 km SW of Menden (Sauerland), Grid Ref. R34154 H56952, Fig. A.8, p. 363.

A section on the north side of the B7 road, in Famennian slope facies (similar to that at Oberrodinghausen) consists mainly of calcareous nodules and nodular limestones in red and green shales.

References: Schmidt and Plessmann (1961), Ziegler (1962).

Padberg, 14 km E of Brilon (Sauerland), Grid Ref. R34842 H56945.
A road cutting between Padberg and Bredelar exposing Famennian basinal facies, red and green shales with rare calcareous nodules.
Reference: Schmidt and Plessmann (1961).

Pfaffenhardt, near Erda, 24 km SW of Marburg (Lahn Syncline), Grid Ref. R34659 H56245.
Middle and Upper Devonian grey shales are exposed in a disused quarry in thick woods. Clasts of Flaser limestone occur in Givetian shales and nodule bands in the Frasnian shales have slumped.
Reference: Bender (1965).

Silberstollen, near Densberg, 15 km SW of Bad Widungen (Kellerwald), Grid Ref. R35056 H56476.
Very poor exposures in thick forest of Flaser limestone of Lower to Upper Devonian. Deposited on a basement Schwelle, the southern ridge of the midgeosynclinal rise.
References: Stoppel (1959), Kockel (1958).

Steinbruch Schmidt, near Braunau, 5 km S of Bad Wildungen (Kellerwald), Grid Ref. R35091 H56594, Fig. A.9, p.364 and Fig. A.14, p.371
Steinbruch Schmidt is a very important locality where pelagic limestones of lower Frasnian to upper Famennian age are exposed. These are mainly thin-bedded limestones separated by shale partings. Two Kellwasser horizons are well developed in the quarry. These sediments were deposited on the Ense Schwelle, one of the ridges of the midgeosynclinal ridge. Limestones contain thin graded units of clastic silt and clay. Neomorphism has led to formation of coarse mosaics in some limestones.
References: Ziegler (1959), Kockel (1958).

Steinbruch Syring, near Oderhausen, 3 km S of Bad Wildungen (Kellerwald) Grid Ref. R35091 H56608, Fig. A.10, p.365, also smaller quarry south of

Steinbruch Syring, Grid Ref. R35092 H56603, Fig. A.11, p.366

These two quarries expose Givetian and lower Frasnian limestones very similar to those at Steinbruch Schmidt. Thin platey limestones and Flaser limestones are separated by shale partings. Some beds are rich in cricoconarids. References: Kockel (1958), Ziegler (1959).

Weilburg (Lahn Syncline), Grid Ref. R34478 H55919.

Upper Devonian slope sediments are exposed in a road cutting just south of the town by the canal tunnel. Shales with calcareous nodules and nodular limestone are slumped. Reworked nodules also occur.

Weipdtshausen, 17 km SW of Marburg (Lahn Syncline), Grid Ref. R34713 H56184.

A disused quarry situated 500 m west of the village exposes Givetian and lower Frasnian Flaser limestone, locally rich in cricoconarids. These sediments were deposited on a geanticline in the Lahn Syncline. Reference: Bender (1965).

Wocklum, near Balve, 14 km SSE of Menden (Sauerland), Grid Ref. 34221 H56886.

The type locality of the Wocklumeria Stufe is a quarry exposing a dark grey nodular limestone on the right bank of the River Borke. These sediments were deposited on the Arnsberg 'reef' Schwelle. Reference: Schmidt and Plessmann (1961).

Montagne Noire (S. France)

Caunes Minervois, Fontaine de la Santé, 19 km NE of Carcassone, Fig. A.12, p. 367.

A complete Upper Devonian succession in Flaser limestone and nodular limestone (griotte) is exposed in the valley of a stream 700 m ENE of the town. Some horizons of the limestone are rich in goniatites. Of note is the presence of 10 m of breccia consisting of Flaser limestone clasts and shale fragments. Epigenetic manganese mineralization is developed in veins. References: Gèze (1949), Boyer et al (1968).

Combe D'Izarne, near Cabrières, 9 km SE of Clermont l'Hérault.

Famennian Flaser limestones crop out on the north side of the D-15 road, 2 km west from the junction with the D124, 2 km south of Cabrieres. Red and grey limestone locally rich in goniatites, contains ferromanganese encrustations around skeletal material and limestone intraclasts. Reference: Gèze: (1949).

Coumiac, north of Cassenon, 10 km NE of St Chinian.

An old quarry (marked 'Carriere de marbre' on the Michelin map 83) situated between Cessenon and Caunes-et-Veyran on the north side of the D136 road. The Upper Devonian is developed as red and grey Flaser limestone with some beds containing many goniatites. Ferromanganese nodules not very common.
References: Gèze (1949), Boyer et al (1968).

Mont Peyroux, north of Causses-et-Veyran, 13 km NE of St Chinian, Fig. A.13, p. 368 and 369.

Upper Devonian griotte is exposed on the hillside to the north of Causses-et-Veyran. A corrasion hardground surface is developed and ferromanganese nodules and encrustations are present. Hematization of some limestones has also occurred.
References: Gèze (1949), Boyer et al (1968).

Pic de Vissou, near Cabrières, 7 km WSW of Clermont l'Hérault.

Pic de Vissou, a mountain 3 km north of Cabrières has an exposure of Upper Devonian griotte towards the top. Brecciated horizons, consisting of angular limestone clasts one centimetre across are developed. Reference: Gèze (1949).

Smaller exposures of griotte in the Montagne Noire were examined at La Serre, Vailhan, Mas Rolland, Roquebrun, St. Nazaire de Ladarez and north of Caunes Minervois.

Upper Devonian sediments of the same facies (Flaser and nodular limestones) were also examined in the Mouthoumet inlier, situated between the Montagne Noire and the Pyrenees. (Localities given in Ziegler, 1959a).

Spanish Pyrenees

Flaser and nodular limestones (griotte) are developed at various localities in the Pyrenees and exposures in the Seo de Urgell region (eastern part of the Spanish Pyrenees, 20 km south of Andorra) were briefly examined. Griotte localities are mentioned in De Sitter and Zwart (1961) and Ziegler (1959). The Upper Devonian sediments are very similar to those in the Montagne Noire and Germany, although deformation has been greater in the Pyrenees.

Espahent, Rio de Castellas, 15 km SW of Seo de Urgell.

A small quarry in red Famennian Flaser limestone is situated on the plateau above the junction of the Rio de Castellas and Rio de la Guardia, 2 km south of the village of Espahent. The griotte here contains goniatites and other skeletal material and is cut by numerous tectonic stylolites.
Reference: De Sitter and Zwart (1961).

Gorge Rio Castellas, 16 km SW of Seo de Urgell.

A working quarry by the River Castellas, situated 1 km from the junction with the Guardia river. Upper Devonian grey and red Flaser limestone with poorly developed bedding. Goniatites are rare although styliolinids are present in the lower parts.
Reference: De Sitter and Zwart (1961).

Trejubell, above the Rio de la Guardia, 18 km west of Seo de Urgell.

Griotte of Middle and Upper Devonian age outcrops along the track past the village of Trejubell. Brecciated horizons are developed, consisting of limestone clasts up to 30 cm across.
Reference: De Sitter and Zwart (1961).

Gerri, 6 km S of Sort, Palleresa Valley.

Just north of the village of Gerri, Middle and Upper Devonian Flaser limestones and nodular limestones are exposed. In parts, the limestones are badly recrystallized and coarse mosaics have developed. Reference: Ziegler (1959a).

S.W. England

Chudleigh, N of Newton Abbot, S. Devon.

Schwellen limestones are developed above Givetian and lower Frasnian 'reef' limestone at Dunscombe Farm (SX 886791). The limestone is rich in conodonts and goniatites and some of the latter have been filled by red calcisiltite containing conodonts.

Exposures in Upper Devonian shales with nodules and nodular limestones occur in the Kiln Wood region (SX 861779). The Upper Devonian sediments in the Chudleigh area are described in Tucker and Straaten (1970) which is appended ~~(p. 388 to 402)~~.

Saltern Cove, S of Torquay, S. Devon.

Upper Devonian shales with bands of calcareous nodules and tuffs are developed from the centre of Saltern Cove to the north end of Waterside Cove. Slumping of sediment has occurred and clasts of Schwellen limestone are present in the shales of the Saltern Cove Goniatite Bed. These sediments and the slumping are described in Straaten and Tucker (1971, in press), the manuscript of which is appended (p.388 to 402).

Padstow, North Cornwall.

The sediments of the 'St. Minver synclinorium' are mainly Givetian and Upper Devonian shales with ash bands and turbidites developed at various horizons. A very brief summary of the sediments is given in Tucker (1969) (appended). The Marble Cliff Beds of upper Givetian/lower Frasnian age are crinoidal turbidites interbedded with black shales. These are described in Tucker (1969) which is included as part of the Appendix. ~~chapter 4 on the basinal facies~~.

ooOoo

Fig. A.1. Upper Devonian succession at Hühnertalskopf (Harz)

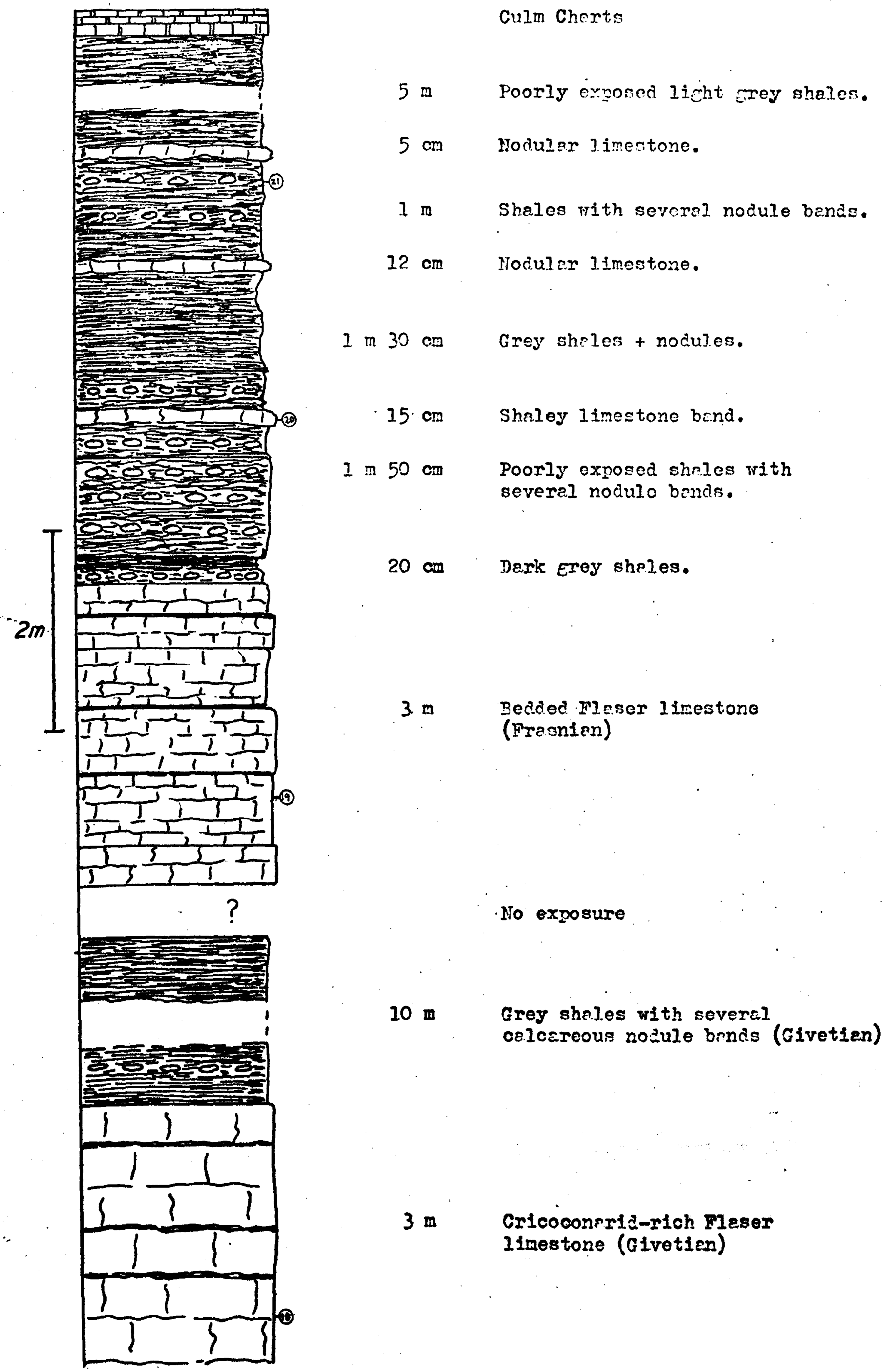
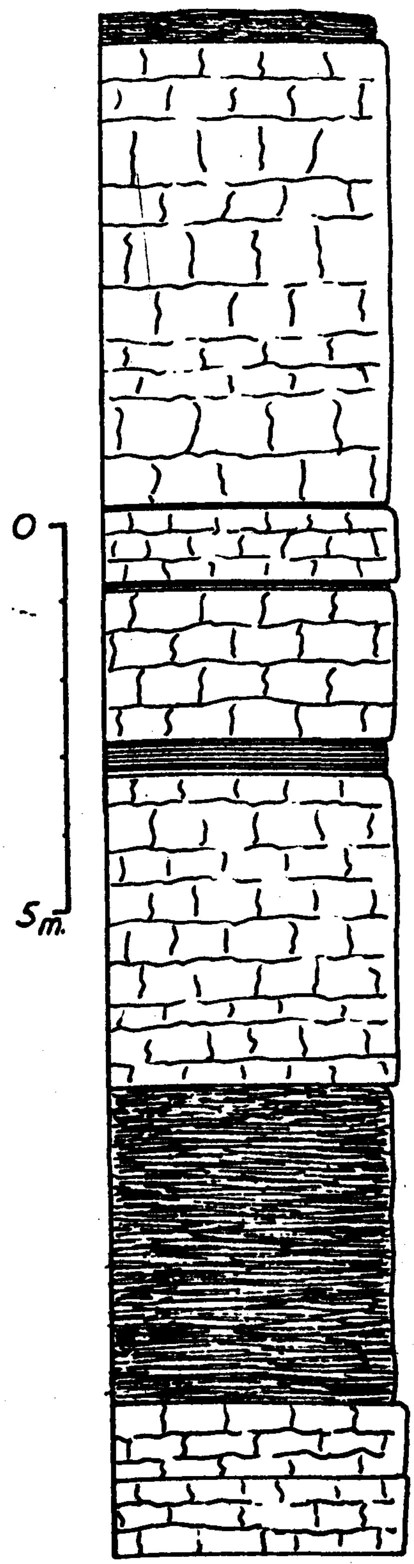
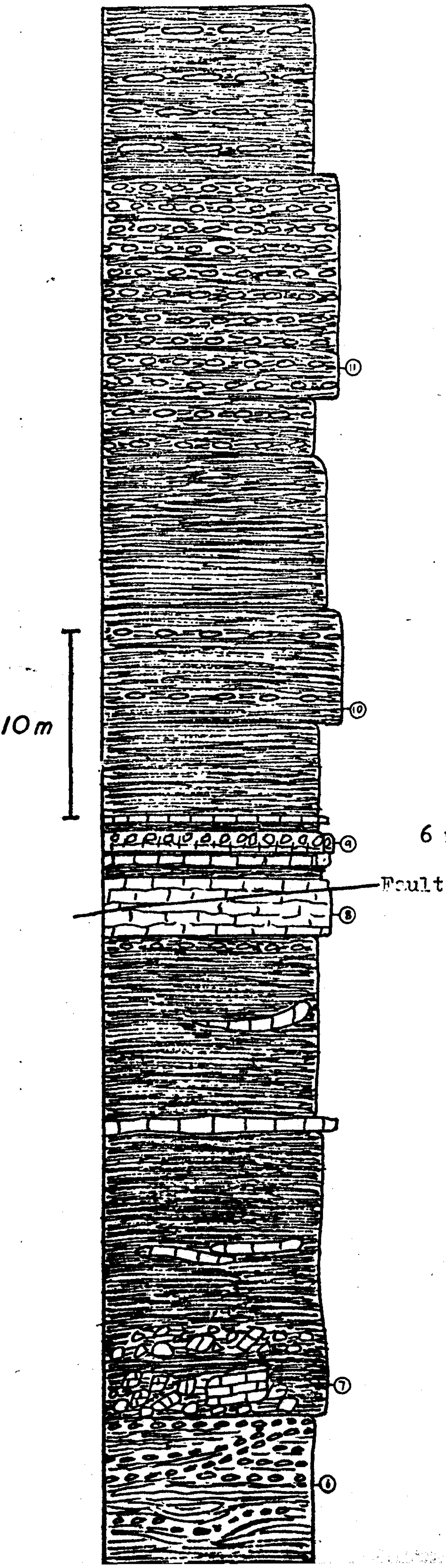


Fig. A.2. Givetian to Carboniferous succession at Romkerhalle Wasserfall (Harz)



	Dark grey shales (Culm)
6 m	Grey Flaser limestone (*Cheiloceras* to *Wocklumeria* Stufen)
3 cm	Black calcareous shale.
1 m	Grey Flaser limestone.
10 cm	Kellwasser horizon.
2 m	Grey Flaser limestone (Frasnian) + thin black bands. Goniatites present.
40 cm	Kellwasser horizon.
4 m	Grey Flaser limestone + styliolinds. Poor exposure.
4 m	Dark grey laminated slates.
2 m	Grey Flaser limestone (Givetian)

Fig. A.3. Givetian/Famennian slope succession at Sparenberg (Harz)



10 m	Dark grey shales with thin nodules. Ostracods and bivalves common. (*Platyclymenia* Stufe)
12 m	Red shale with many bands of small nodules. (*Platyclymenia* Stufe)
3 m	Red shale + few calcareous nodules.
8 m	Greenish grey shales with some red patches.
6 m	Grey shales with several nodule bands (*Cheiloceras* Stufe)
5 m	Poorly exposed grey shales.
6 m 30 cm	Flaser limestone, breccias, grey shales and Kellwasser horizon (Frasnian)
10 m	Grey shales with several nodule bands and limestone lense. (Frasnian)
30 cm	Limestone bed rich in styliolinids.
15 m	Dark grey shales with derived limestone clasts at base, and disturbed limestone bed higher up. (Givetian/Frasnian)
+15 m	Dark grey shales with nodule bands. Slumping and reworking has occurred. (Givetian)

Fig. A.4. Upper Givetian/Frasnian succession at Bicken (Dill Syncline)

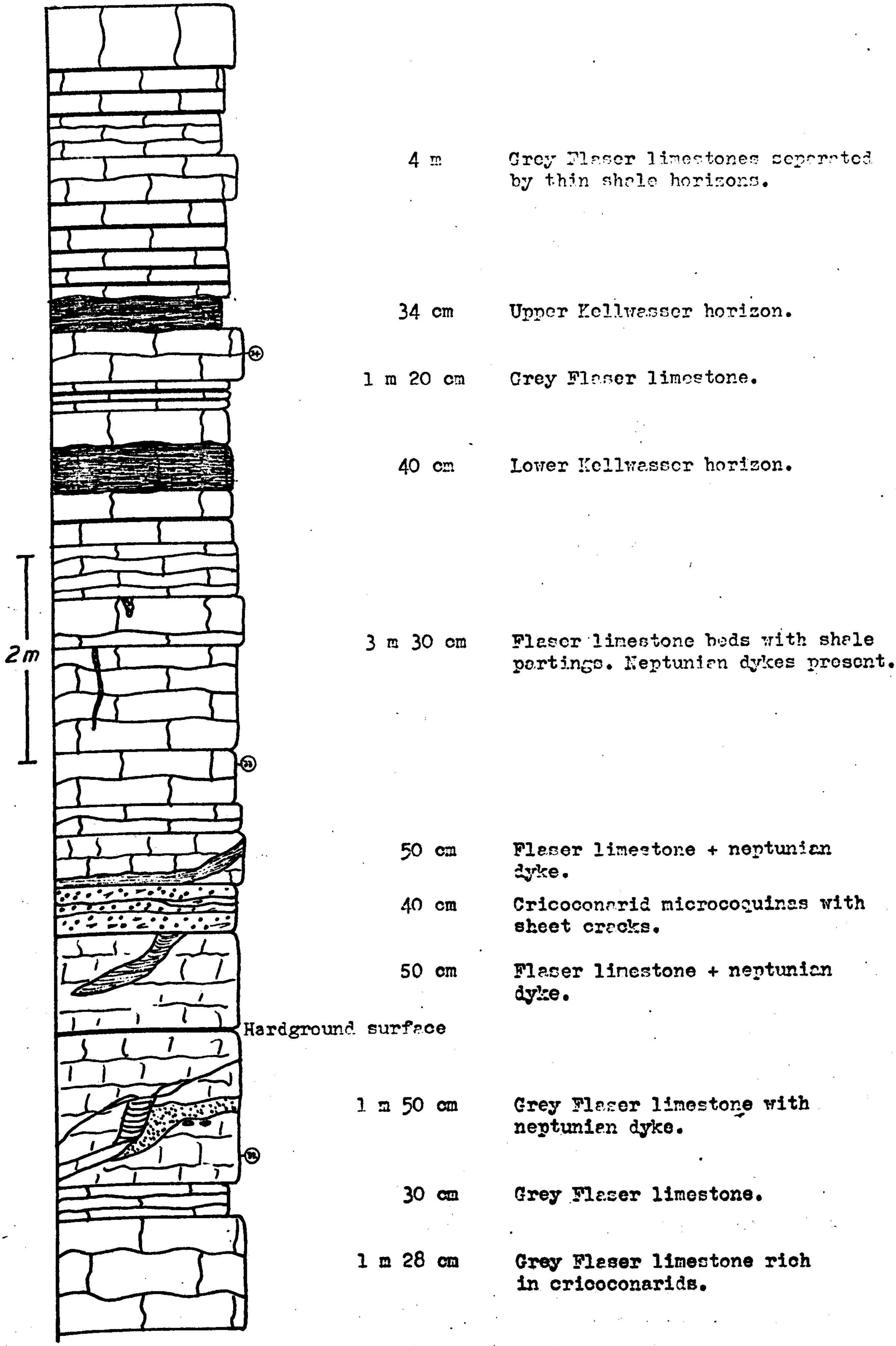
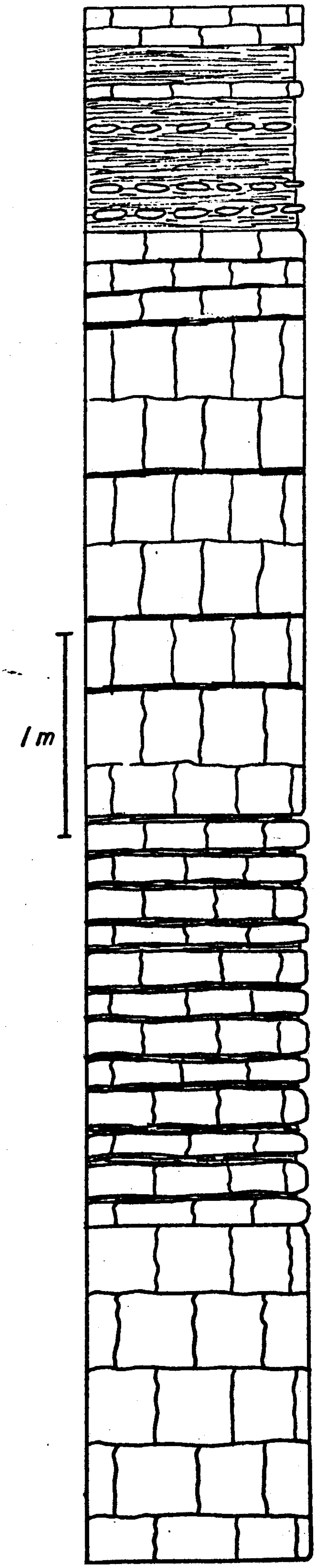


Fig. A.5. Givetian/Frasnian Succession at Blauer Bruch (Kellerwald)



20 cm — Grey Flaser limestone.

1 m — Grey shales and nodular horizons.

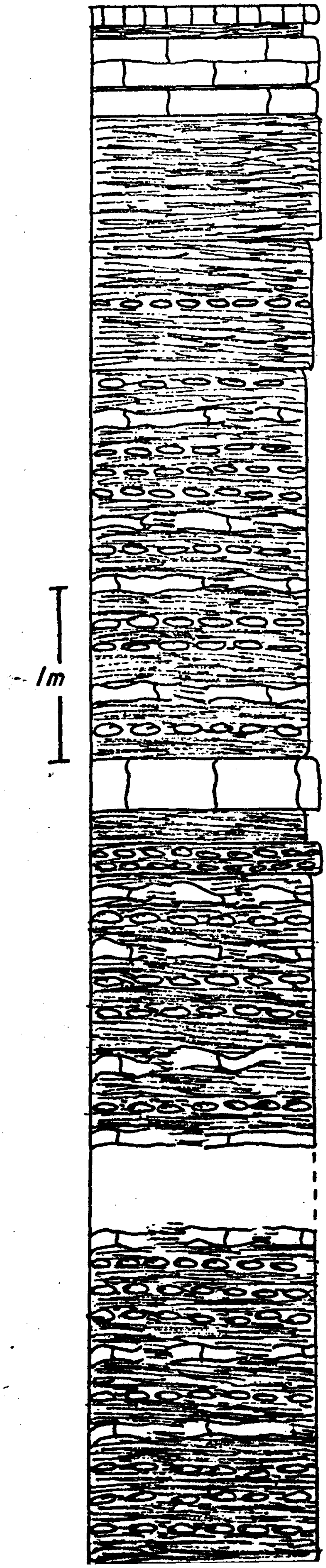
Thin Flaser limestones with shale partings.

3 m — Grey cricoconarid-rich limestone with thin shale bands and cut by numerous Flasers.

2 m — Thin limestone bands 5-15 cm thick, with horizontal and cross laminations, separated by thin shale partings.

1.5 m — Grey Flaser limestone rich in cricoconarids.

Fig. A.6. Famennian/Lower Carboniferous Succession at Drewer



	Culm Cherts
40 cm	Erdbach limestone.
65 cm	Alaunschiefer :- black shales + phosphate nodules.
65 cm	Black shales + calcareous nodules.
2 m 20 cm	*Gattendorfia* limestone. Nodular limestones and calcareous nodules in black shales.
35 cm	Hangenberg limestone (*Gattendorfia* Stufe).
23 cm	Hangenberg slate (black shale).
20 cm	Grey shales with calcareous nodules.
25 m	Nodular limestones and calcareous nodules in pyritic black shales. Rich in macrofossils (*Cheiloceras* to *Clymenia* Stufe).

Fig. A.7. Upper Devonian succession at Haingrube (Kellerwald)

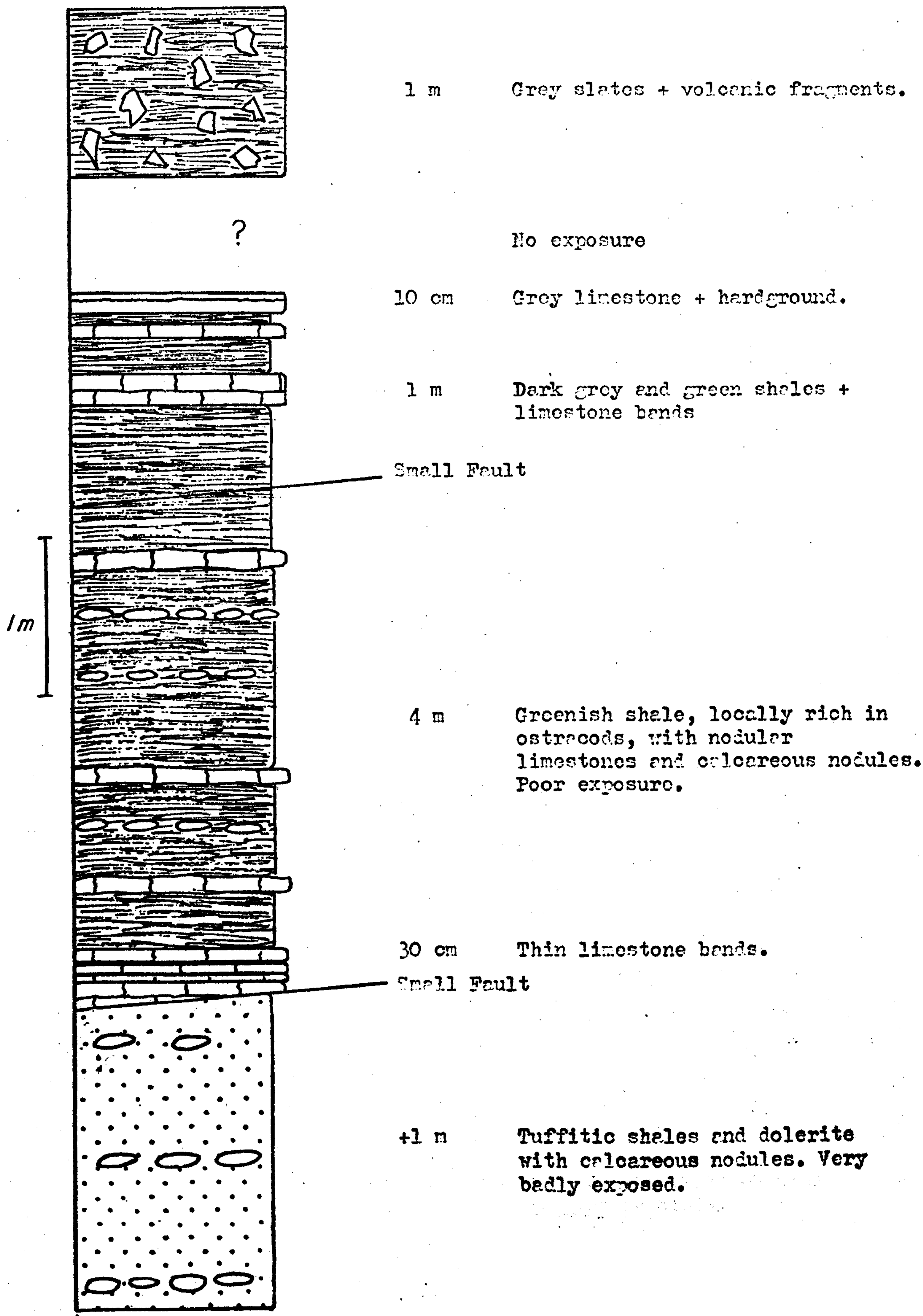
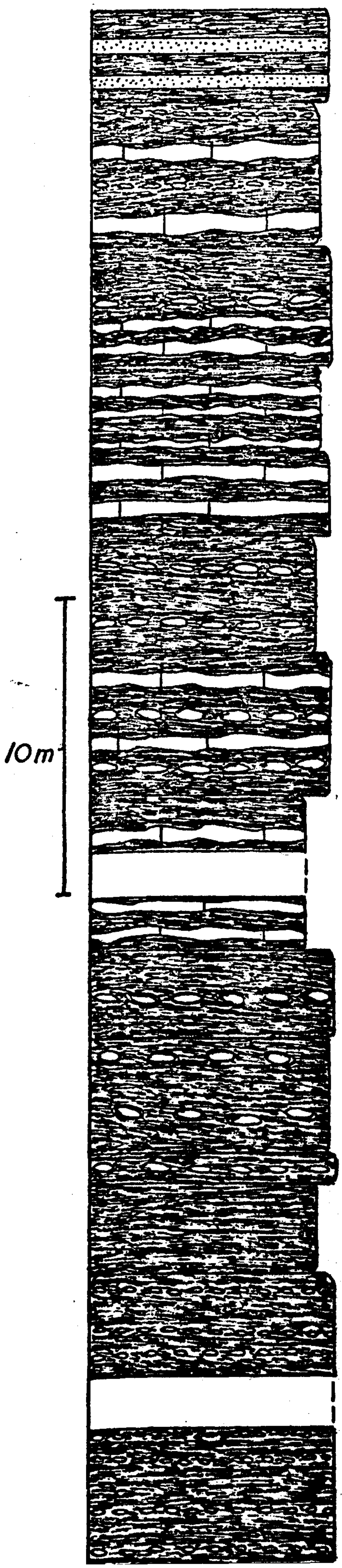
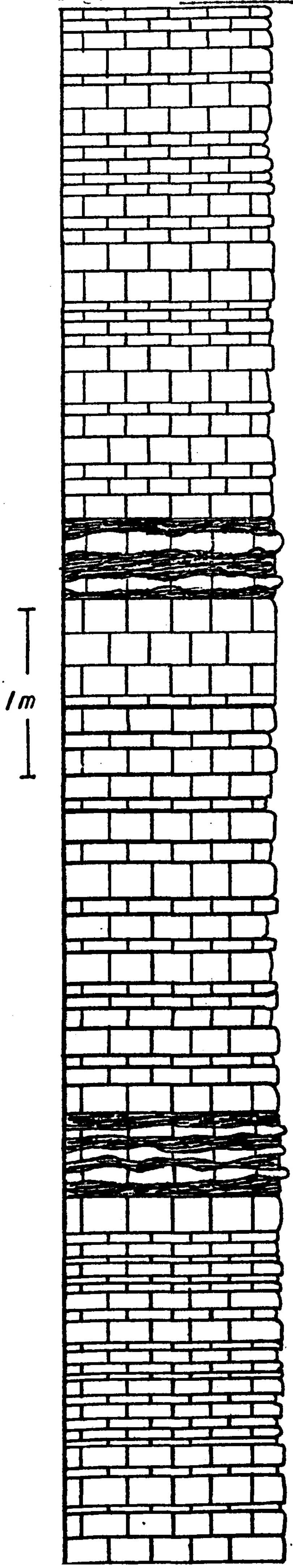


Fig. A.8. Famennian Succession at Öse (Sauerland)



3 m	Grey shales and graded sandstones (*Wocklumeria* Stufe)
5 m	Grey green shales and nodules and nodular limestones.
7 m	Grey shales and calcareous nodules (*Clymenia* Stufe)
3 m	Red shales and nodular limestones (*Platyclymenia Stufe*)
4 m	Green shales and calcareous nodules.
5 m	Greenish shales and nodular limestones and nodules.
10 m	Red shales and nodular limestones.
3 m	Red shales and nodules.
4 m	Green shales, red in parts with few nodules.
1 m	Green shales and nodules.
3 m	Greenish-brown shales.
10-15 m	Red shales with small nodules (*Platyclymenia* Stufe)

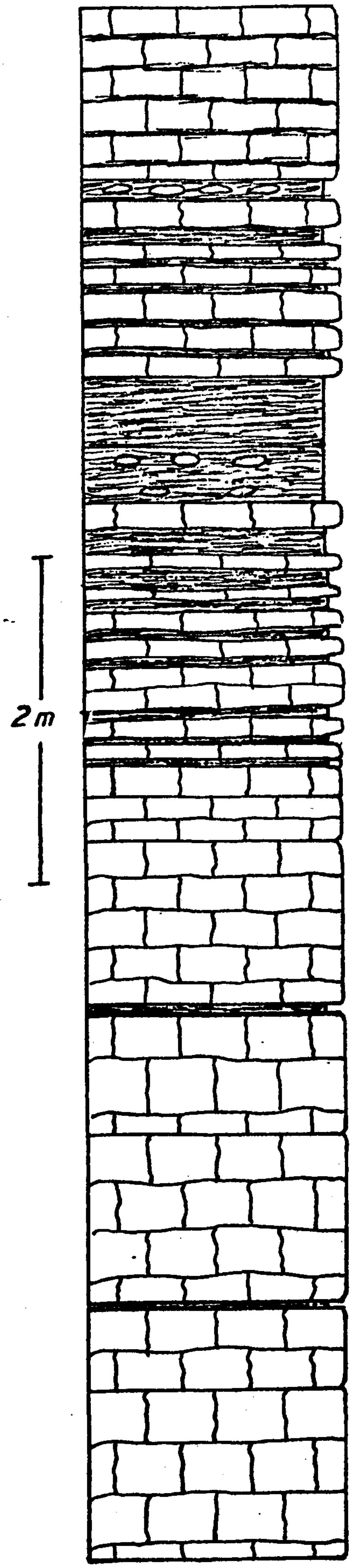
Fig. A.9. Steinbruch Schmidt, near Braunau (Kellerwald)



3 m 70 cm	Thin bedded Flaser limestones with thin shale bands between.
45 cm	Upper Kellwasser horizon.
3 m	Flaser limestone beds 2 to 20 cm thick, separated by thin shale partings.
52 cm	Lower Kellwasser horizon (black shales + nodular limestone)
+2 m	Flaser limestone bands, varying in thickness from 2 cm to 30 cm, separated by shale partings.

Fig. A.10.

Givetian/Frasnian succession at Steinbruch Syring near Oderhausen (Kellerwald)



1 m	Thin limestones, up to 6 cm thick + shale partings.
10 cm	Red shales + calcareous nodules.
1 m 10 cm	Thin limestone bands, up to 20 cm thick.
70 cm	Dark grey shale + few nodules.
1 m 60 cm	Thin Flaser limestones 5 to 25 cm thick + shale partings.
1 m 50 cm	Flaser limestones, thin shale partings.
8 cm	Red shale.
1 m 80 cm	Grey Flaser limestone rich in styliolinids.
3 cm	Grey shale
1 m 60 cm	Grey Flaser limestone.

Fig. A.11. Givetian/Frasnian succession in quarry south of Syring (Kellerwald)

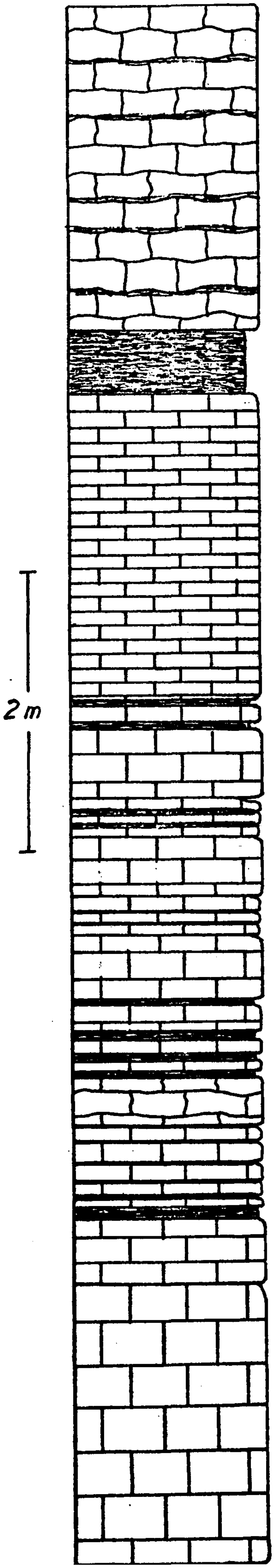
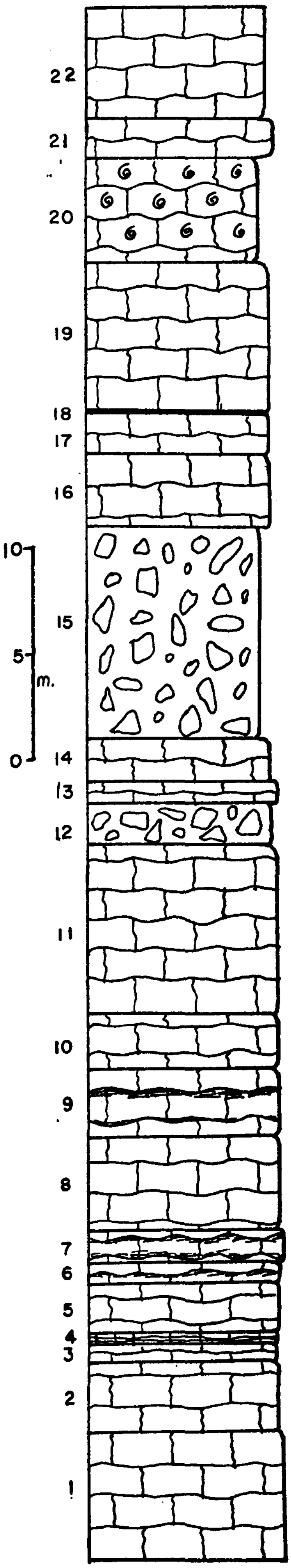


Fig.AJ2. Upper Devonian Griotte Succession at Caunes Minervois, Montagne Noire



22	5 m	Grey Flaser limestone.
21	2 m	Red Flaser limestone.
20	5 m	Red griotte, rather shaley + goniatites.
		(Lower Famennian)
19	7 m	Red & grey Flaser limestone + nodular limestone.
18	12 cm	Kellwasser limestone. (Frasnian)
17	2 m	Grey Flaser limestone.
16	3.5 m	Red Flaser limestone.
15	10 m	Breccia, limestone fragments and nodules in red shales. Fe/Mn enrichment.
14	2 m	Pink nodular limestone.
13	1 m	Grey Flaser limestone.
12	2 m	Coarse breccia.
11	8 m	Pink/grey Flaser limestone + shaley partings. (Frasnian)
10	2.5 m	Pink & grey Flaser limestone.
9	3 m	Grey & pink nodular limestone.
8	4.5 m	Grey/Pink Flaser limestone + shaley partings.
7	1.5 m	Red nodular/Flaser limestone.
6	1.07 m	Grey limestone + shaley partings.
5	2.20 m	Grey Flaser limestone.
4	60 cm	Thin limestones + shale partings.
3	1 m	Grey Flaser limestone
2, 1	+ 9 m	Grey Flaser limestone (Givetian)

Fig. A.13 Upper Devonian Griotte Succession at Mont Peyroux.

The Upper Devonian at Mont Peyroux has not been dated with conodonts by Boyer et al (1968). Four conodont samples (e.g. B) were obtained from the horizons indicated on the section opposite and determined as follows (provisionally by M.E.T., checked by P. van Straaten). o = cf. determination.

	A	B	C	D
Ancyrodella gigas YOUNGQUIST 1947		x		
Ancyrodella retundiloba BRYANT 1921		x		
Ancyrodella spp.		x		
Icriodus spp.		x		
Palmatolepis gracilis gracilis BRANSON & MEHL 1934			x	x
Polygnathus asymmetrica asymmetrica		x		
Polygnathus linguiformis HINDE 1879	x			
Polygnathus xylus STAUFFER 1940	x			
Polygnathus spp.		x		
Pseudopolygnathus spp.				x
Spathognathodus costatus costatus BRANSON 1934				x
Spathognathodus costatus spinulicostatus BRANSON 1934			o	x
Spathognathodus stabilis BRANSON & MEHL 1934			o	
Spathognathus strigosus BRANSON & MEHL 1934			o	
Spathognathus sp indet.		x		

The stratigraphic positions of the samples (from Ziegler, 1962) are as follows:-

Sample A Givetian/Frasnian

Sample B Lower Frasnian (asymmetrica zone)

Sample C Upper Famennian (costatus zone)

Sample D Upper Famennian (costatus zone).

Fig. A.13. Upper Devonian Griotte Succession at Mont Peyroux, Montagne Noire

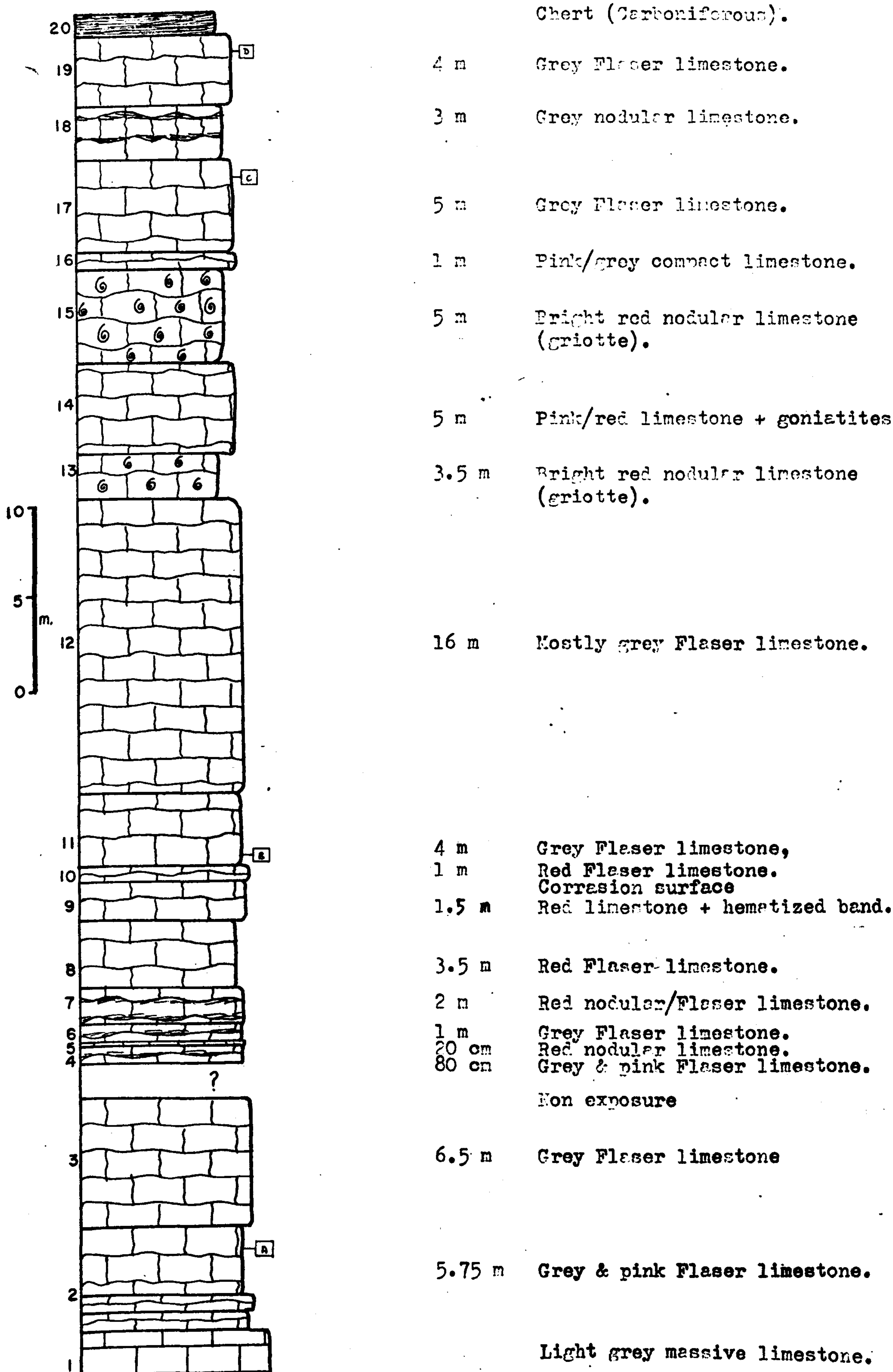
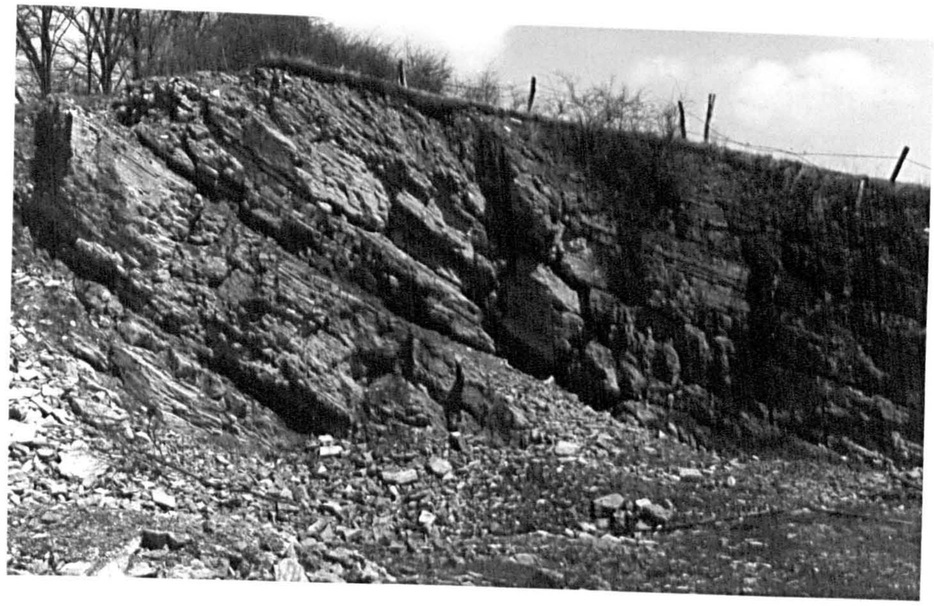
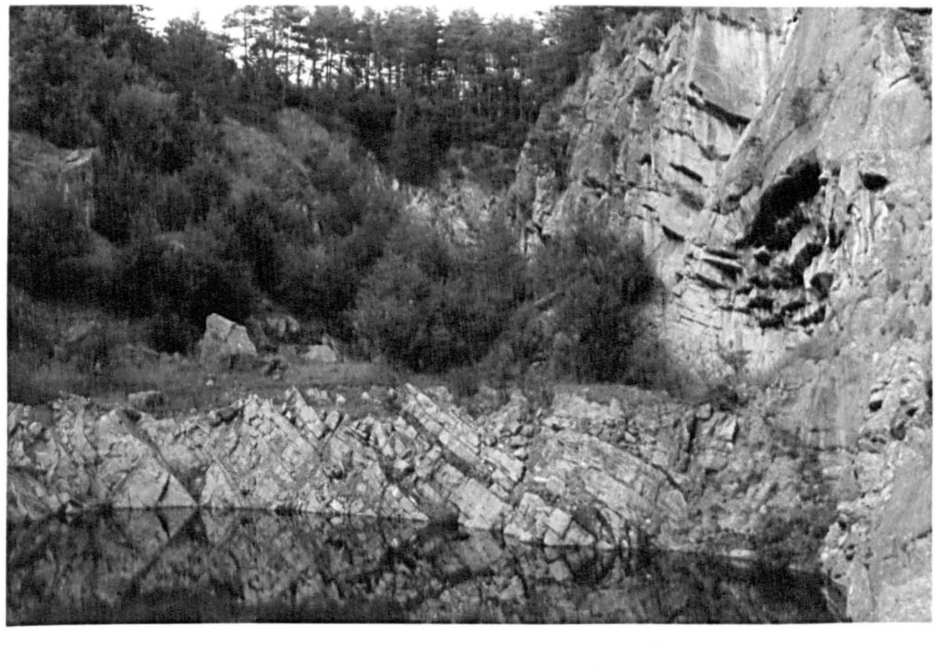


Fig. A .14. Steinbruch Schmidt (near Braunau) Kellerwald (looking east). The two prominent beds (on the left) which have been excavated further are the upper and lower Kellwasser Horizons. A fault in the centre of the quarry separates the Frasnian succession on the left, from the Famennian on the right.

Fig. A .15. Benner Steinbruch, Bicken (near Dillenburg) (looking east) Dill Syncline. Section immediately above lake is in Frasnian Flaser limestones, with two Kellwasser horizons exposed in the central part. Famennian cephalopod limestones are exposed in the cliff face on the right.





Locations of figured specimens

All specimens (thin sections, peels and hand specimens) figured in this thesis are archived in the Sedimentology Research Laboratory, University of Reading. Additional material has also been placed in this collection. Particulars about the localities are given earlier in this appendix. The abbreviations Rh.Sch., M.N. and S.W.Eng. are used for Rheinisches Schiefergebirge, Montagne Noire and S.W. England respectively.

Archive No.	Locality	Location of Specimen
S22950	Bicken	4m 70cm below lower Kellwasser limestone
S22951	Mont Peyroux (M.N.)	Corrasion surface located 55m NE of small quarry, 30m from beginning of the griotte, 20m stratigraphically above light grey massive limestone.
S22952	Combe D'Izarne (M.N.)	20m above and to N of stream, 10m of beginning of exposure, 30m west of track.
S22954	Combe D'Izarne (M.N.)	as S22952, sample collected 15m W of beginning of exposure, 35m W of track.
S22955	Combe D'Izarne (M.N.)	20m stratigraphically above light grey massive limestone, 60m NW of track. (25m N of sample S22952)
S22957	Mont Peyroux (M.N.)	As S22951, specimen from below corrasion surface.
S22958	Mont Peyroux (M.N.)	As S22951, sample 30cm below corrasion surface.
S22959	Mont Peyroux (M.N.)	as S22958
S22960	Mont Peyroux (M.N.)	Sample from entrance to small quarry om hillside.
S22962	Caunes Minervois (M.N.)	4m from base of brecciated horizon.
S22964	Combe D'Izarne (M.N.)	As S22954
S22965	Combe D'Izarne (M.N.)	As S22954
S22966	Coumiac (M.N.)	Higher quarry, 14m above Kellwasser horizon of Boyer et al,1968, 5m below upper Famennian grey Flaser limestone.
S22967	Combe D'Izarne (M.N.)	As S22952
S22968	Combe D'Izarne (M.N.)	As S22952

Archive No.	Locality	Location of Specimen
S22969	Haingrube (Rh.Sch.)	Limestone band below limestone containing cryptohardground (see section p. 362) 4m 30cm above dolerite.
S22970	Haingrube (Rh.Sch.)	Highest limestone before non-exposure, 4m 50cm above dolerite extrusion.
S22973	Blauer Bruch (Rh.Sch.)	Samples of thin bedded limestones, 60cm above poorly bedded grey Flaser limestone, exposed behind farmhouse.
S22974	Adorf am Martenberg (Rh.Sch.)	1m 40cm below ostracod shales at top of section. NE corner of quarry cliff.
S22975	Adorf am Martenberg (Rh.Sch.)	1m 30cm below ostracod shales at top of section. NE corner of quarry cliff.
S22977	Adorf am Martenberg (Rh.Sch.)	loose block.
S22979	Adorf am Martenberg (Rh. Sch.)	3m 70cm below ostracod shales at top of section.
S22983	Adorf am Martenberg (Rh.Sch.)	1m 50cm below ostracod shales at top of section. NE corner of cliff.
S22986	Dunscombe Farm Chudleigh (S.W.Eng.)	1m 50cm above top of massive limestone.
S22990	Steinbruch Schmidt (Rh.Sch.)	25cm below lower Kellwasser horizon.
S22992	Steinbruch Schmidt (Rh.Sch.)	1m 10cm above lower Kellwasser horizon.
S22997	Langenholthausen (Rh.Sch.)	Sample of nodular limestone, about 10m from base of section.
S22998	Grevenbruck (Rh.Sch.)	Sample 4m 30cm above base of section.
S23035	Bicken (Rh.Sch.)	9m above upper Kellwasser horizon, 3m 50cm from base of cliff section.
S23036	Aeketal (Harz)	3m 50cm below lower Kellwasser horizon.
S23038	Aeketal (Harz)	50cm below lower Kellwasser horizon.
S23041	Aeketal (Harz)	70cm above lower Kellwasser horizon.
S23042	Aeketal (Harz)	3m 40cm above lower Kellwasser horizon.
S23045	Langestal (Harz)	By telegraph post 26, 2m 10cm from base of section.
S23047	Gaudernbach (Rh.Sch.)	Loose block.
S23051	Eulenspiegel (Rh.Sch.)	Specimen from Frasnian limestone, W face of quarry, 4m from base of section.
S23052	Railway cutting Innerstetal Reservoir (Harz)	By railway post no. 66 collected 3 to 4 m from base of grey green shales with nodules.
S23054	Railway cutting Innerstetal Reservoir (Harz)	40m N of railway post no. 48, sample of black shales in bank.

Archive No.	Locality	Location of Specimen
S23059	Drewer (Rh.Sch.)	5m below Hangenberg limestone, near exit of quarry.
S23061	Drewer (Rh.Sch.)	5m below Hangenberg limestone, near exit of quarry.
S23062	Drewer (Rh.Sch.)	10m below Hangenberg limestone, 20m east of quarry exit.
S23063	Bicken (Rh.Sch.)	4m 70cm below lower Kellwasser horizon.
S23064	Bicken (Rh.Sch.)	3m 80cm below lower Kellwasser horizon.
S23065	Bicken (Rh.Sch.)	4m below lower Kellwasser horizon.
S23067	Bonzel (Rh.Sch.)	1m 50cm below top of quarry, approx. 3m above <u>Terebratula</u> <u>pumilio</u> bed.
S23068	Maurach Sonnwendgebirge Austria	Loose block from quarry, E of village.
S23069	Adorf am Martenberg (Rh.Sch.)	3m 90cm below ostracod shales at top of section. N part of quarry cliff.
S23070	Junkernberg (Harz)	Samples from exposure of red and green shales, in bank of forest track to Junkernberg quarry. 250m W of Langelsheim to Lautenthal road.
S23072	Hühnertalskopf (Harz)	3m 50cm above top of Givetian limestone.
S23073	Railway cutting Innerstetal Reservoir (Harz)	Exposure of grey greenish shales with nodules. Opposite railway post no. 64, 2m from base of section.
S23075	Adorf am Martenberg (Rh.Sch.)	60cm below ostracod shales at top of section. NE corner of quarry cliff.
S23077	Blauer Bruch (Rh.Sch.)	Thin bedded limestone 25 cm above grey poorly bedded Flaser limestone, exposed behind farmhouse.

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APPENDIX II

Appendix to Chapter 3: Electron Microscpe Preparations

One surface of a centimetre cube of the limestone sample was highly polished, finally using 1 micron powder. The polished surface was then etched in 1% hydrochloric acid for 14 seconds. Various acid concentrations and etch times were tried to find the best effect. Acetate peels were then taken of the surface and the second or third one was used in the replicating stage. The peel was shadowed at 30° with gold/palladium (40% Au) and then coated with carbon from three directions. A small piece of the acetate peel, about 3 mm across was placed on a grid and then the acetate paper dissolved away in acetone vapour. The replica was then ready for viewing with the electron microscope.

I am very grateful to Mr. John Forsdyke for the excellent preparation of replicas of 25 limestone samples.

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APPENDIX III

APPENDIX TO CHAPTER 6: Details of Analytical Techniques.

Preparation of Solutions for Atomic Absorption Spectrophotometer

For determinations of Ca and Mg 0.1 gms of oven-dried powdered sample were weighed and then digested overnight in a beaker with 200 mls of 3% acetic acid. The solution was transfered to a flask and made up to 1000 ml with de-ionised water. After shaking well, the solution was filtered into a 1 litre plastic bottle. This solution was the stock solution from which further dilutions were made.

5 ml of stock solution were pipetted into a 50 ml flask. To this were added 10 ml lanthanum chloride solution (5% solution La^{3+}) and the whole made up to exactly 50 ml. This solution in a small plastic bottle was then analysed for Ca and Mg ions. The purpose of lanthanum ions in the solution is to release the Ca ions and make them all available for atomic absorption. Cleanliness is essential throughout the preparation, glassware and plastic bottles must be washed several times with nitric acid, hot water and de-ionised water to prevent contamination.

Preparation of solutions for Sr, Mn and Fe

For determinations of these elements, a stronger solution is obviously needed, and so approximately 0.5 gm of powder were accurately weighed out and digested overnight as before, with 200 ml 3% acetic acid, and then made up to 250 ml. The solution was now ready for analysis of Mn and Fe, but a further stage is required for Sr. 20 ml of solution were withdrawn by pipette and placed in a small plastic bottle. 5 ml of sodium chloride solution (5% Na^{+}) were added from a burette, and then determination of Sr undertaken. Sodium ions have a similar effect on the strontium as the lanthanum on calcium, they release the Sr ions and prevent interference from calcium.

Preparation of Standard Solutions

To calibrate the atomic absorption spectrophotometer standards were made up for the various elements. The standard solutions in each case contained the same amount of acetic acid as the sample solutions.

Calcium For Ca a 1000 ppm stock solution was made by dissolving 0.2497 gm of dried Analar $CaCO_3$ in de-ionised water and making up to 1000 ml. By two dilutions standard solutions of 1,2,3 and 4 ppm Ca^{2+} were obtained. The correct amount of lanthanum chloride was also added.

Magnesium A 100 ppm stock solution was obtained by dissolving 1.0136 gm of $MgSO_4.7H_2O$ and making up to 1 litre. Solutions of 0.25, 0.5, 0.75 and 1 ppm were prepared by two dilutions.

Stontium 0.30431 gm $SrCl_2.6H_2O$ were dissolved and made up to 1 litre to give 100 ppm stock solution. From this, standard solutions of 1,2,3 and 4 ppm were made. The requisite amount of NaCl was also added.

Manganese and Iron These elements have not been analysed by atomic absorption before in Reading and so two sets of standards were made to test for matrix effects. To one set a certain amount of Ca^{2+} ions were added so that the standard solutions had a similar calcium concentration to the samples. Ca^{2+} ions were not added to the other set of standards. Standards ranging from 0 to 10 ppm were made for both Fe and Mn, all standards contain the same acetic acid concentration as the samples.

Precision

To check the precision of the method for Ca and Mg, 10 determinations were made of Standard Argillaceous Limestone 1 A (National Bureau of Standards) and 6 determinations were made of 2 samples from Germany (Innerstetal 2 and Bicken 34) one rich and one poor in Mg. For Sr, Mn and Fe precisions, ten determinations were made of a German sample, Aeketal 29. Standard deviations obtained for the precision tests are as follows:-

	Ca	Mg	Sr	Mn	Fe
Argill. Limestone 1A	0.32	.014	-	-	-
Bicken 34	0.78	.020	-	-	-
Innerstetal 2	0.29	.083	-	-	-
Aeketal 29	-	-	.011	.0017	.0069

The accuracy of the method cannot be determined since the Ca and Mg values given for the Standard Argillaceous Limestone are based on whole rock analyses, and results obtained here are from the acid-soluble fraction. Values given for the limestone by N.B.S., with those obtained by the author, and Mr. H.A.Guntilake (also on acid-soluble fraction) are as follows:-

	N.B.S.	M.E.T.	H.A.G.	Thompson et al 1970*
Ca%	29.53	29.17	29.59	-
Mg%	1.32	0.74	0.71	-
Sr ppm	1945	1880	-	2000± 117
Fe%	1.14	0.32	-	-
Mn ppm	-	0.037	-	-

*Thompson et al (1970) published trace element analyses on whole rock sample by emission spectrocopy.

Insoluble Residue

0.1 gm of the powdered sample were digested overnight in

200 ml of 3% acetic acid and the residue was obtained using a Gooch crucible and Buchner Flask, attached to a vacuum pump. The difference in weight between the oven-dried crucible and insoluble residue, and the oven-dried crucible before filtration gave the insoluble residue.

X-Ray Diffraction Analyses

A selection of the samples were analysed in the standard way on a Philips X-Ray Diffractometer to ascertain the major mineral phases present. Most powdered samples were scanned from 4° to 64°, using CuKα radiation.

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Locations of Geochemical Samples

The stage or Stufe to which the sample belongs is given in Table 1, chapter 6, and map and literature references to the localities are given in Appendix I. The sample horizons are marked on the vertical section for each locality e.g. -⑩ . Powder samples, numbered 1 to 35 have hand specimen(s) and thin section or peel numbered 23001 to 23035 in the Reading University Sedimentology Research Laboratory archives.

Locality	Sample No.	Exact location of sample	Archived Material
Junkernberg (Innerstetal)	1	Small exposure of brown shales with large nodules on SE side of quarry about 25 m below Lower Carboniferous dolerite.	HS, TS, P.
"	2	Main quarry, NW face, sample from band of calcareous nodules 2m 45cm below intrusive dolerite.	HS, TS, P.
Railway cutting S. end of Innerstetal Reservoir.	3	40 m S of railway post no. 64, sample collected 3m 50cm from base of grey shales with nodules lithology.	HS, TS, P.
"	4	Section by railway post no. 66, sample collected 2m 30cm from base of section in red shales with nodules.	HS, TS, P.
"	5	Sample collected 4m 50cm from start of grey shales with nodules.	HS, TS, P.
Sparenberg	6	Eifelian to upper Famennian section on E bank of the River Innerste. Sample from S end of section, collected 7 m below base of lower Frasnian shales containing large slumped blocks of limestone.	HS, TS, P.
"	7	Sample from block of slumped lower Frasnian limestone, 2 m above evenly bedded Givetian shales with nodules.	HS, P.
"	8	Sample collected 1 m below fault, 2 m above base of middle Frasnian limestone.	HS, P.
"	9	Sample collected 15cm below top of highest Frasnian limestone.	HS, P.

Locality	Sample No.	Exact location of sample	Archived Material
Sparenberg	10	Sample collected from nodular band in grey shales 8m 20cm above the highest Frasnian limestone.	HS, TS, P.
"	11	Sample collected 28 m above highest Frasnian limestone, from red shales with small nodules, 80 cm above 'U. Hem' (lower <u>Platyclymenia</u>) painted on rock.	HS, TS, P.
Lautenthal	12	Section situated along footpath on E bank of River Innerste. Sample collected from nodular limestone 10 m N of first fault, 5 m below first band of reworked nodules.	HS, P.
"	13	Sample collected immediately below first black Kellwasser horizon; 2 m S of second fault.	HS, P.
"	14	Sample collected 80 cm above first Kellwasser horizon.	HS, P.
"	15	Sample collected immediately above top Kellwasser horizon.	HS, P.
"	16	Sample collected from nodular band in grey shales, 15 m S along path from highest Frasnian limestone, first nodule band 6 m below fault.	HS, P.
"	17	Sample collected 4 m S of a brick structure, 5 m above a prominent fault.	HS, P.
Hühnertalskopf	18	Section along forest track, sample taken 1m 20cm up from base of first limestone, at the N end of the succession.	HS, TS, P.
"	19	Sample from Frasnian limestones 2m 10cm S of large tree, 80 cm from base of limestone.	HS, TS.
"	20	Sample from nodular band 1m 40cm above the Frasnian limestones.	HS, P.
"	21	Sample from nodular band 4m 20cm above top of Frasnian limestones.	HS, P.
Margaretten Klippen	22	Section along a forest track, sample taken from 50 cm above base of lowest Frasnian limestone.	HS, P.
"	23	Sample taken 9 m above the top of the Frasnian limestone, from shales with nodules lithology.	HS, TS, P.
"	24	Sample taken 25 m above Frasnian limestones, 9 m above the start of the <u>Platyclymenia</u> red and green shales.	HS, P.

Locality	Sample No.	Exact location of sample	Archived Material
Aeketal	25	Section along a forest track, sample collected 60cm below the Lower Kellwasser horizon.	HS, P.
"	26	Sample collected 50cm above the lower Kellwasser horizon.	HS, P.
"	27	Sample collected 1m 20cm above the upper Kellwasser horizon.	HS, P.
"	28	Sample collected from bank at back of quarry behind a fir tree, 50 cm below outcrop of first continuous limestone in quarry.	HS, P.
"	29	Sample collected 60cm above prominent bedding plane in central part of quarry, approx. 7 m above the upper Kellwasser horizon.	HS, P.
Hutthaler Widerwaage	30	Stream section, sample collected from 10 cm above a thick extrusive dolerite.	HS, TS.
"	31	Sample collected 20 cm above thin dolerite dyke of _Platyclymenia_ age.	HS, P.
Bicken	32	Flooded quarry. Sample collected 5m 20cm below lower Kellwasser horizon.	HS, P.
"	33	Sample collected 2m 60cm below lower Kellwasser horizon.	HS, P.
"	34	Sample collected 95cm above lower Kellwasser horizon, 25cm below upper Kellwasser horizon.	HS, P.
"	35	Sample collected approx. 9 m above the upper Kellwasser horizon, 3m 50cm from the base of the cliff section above the lake.	HS, TS, P.

HS = Hand specimen
TS = Thin section
P = Peel

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APPENDIX IV

Published work and work in press

The following papers and manuscripts are appended:-

Tucker, M.E., 1969. The sedimentological history of the Padstow area, North Cornwall (Abstract). Proc. Ussher Soc., 2 plll.

Tucker, M.E., 1969. Crinoidal turbidites from the Devonian of Cornwall and their palaeogeographic significance. Sedimentology, 13: 281-290.

Tucker, M.E., 1971. Devonian Manganese Nodules from France. Nature (Physical Science), 230: 116-117.

Tucker, M.E. and Straaten, P. van, 1970. Conodonts and Facies on the Chudleigh Schwelle. Proc. Ussher Soc., 2: 160-170.

Tucker, M.E. and Straaten, P. van, 1970. Conodonts from the Upper Devonian of the Saltern Cove - Elberry Cove area (Abstract). Proc. Ussher Soc., 2 p.159.

Straaten, P. van and Tucker, M.E. The Upper Devonian Saltern Cove Goniatite Bed is an intraformational slump. Palaeontology (in press).

Kendall, A.C. and Tucker, M.E. Radiaxial fibrous calcite as a replacement after syn-sedimentary cement. Nature (in press).

Two papers and an abstract of work published with Peter van Straaten are appended. Field work for the two papers was mainly carried out by the authors in October 1969. The areas had been examined previously by M.E.T. in March, 1969. Further work at Saltern Cove when more conodont samples were collected, was undertaken in May 1970 by M.E.T. and by both authors in January, 1971.

Conodonts from acid digestion were obtained by both authors in Reading and in Gottingen but all final determinations were made by P. van Straaten. All sedimentological work was undertaken by M.E.T. in Reading and both papers were written by M.E.T. Text figures and photographs were also prepared in Reading.

EXTRACT PROCEEDINGS OF THE USSHER SOCIETY
Volume Two, Part Two, 1969 p.111

The sedimentological history of the Padstow area, North Cornwall (Abstract): by M. E. Tucker.

The Middle and Upper Devonian bathyal sediments of the thrusted St. Minver synclinorium are exposed along the coast from Booby's Bay to Portquin, North Cornwall. The successions are significantly different on either side of the structure: sedimentation was influenced locally by positive rises.

The oldest sediments (Trevose Slates in the west) are finely laminated grey slates and distal turbidites, with a dominantly pelagic fauna. A series of allodapic limestones interstratified with black pyritiferous shales (Marble Cliff Beds) follows in the west. The limestones, interpreted as near-proximal turbidites derived from a rise to the south-west, are composed almost entirely of crinoidal and bryozoan debris, with no terrigenous material.

Volcanic activity followed giving rise to the Pentire Pillow Lavas and associated sediments in the north, which are partly equivalent to tuff and agglomerate bands in shales of the west (Longcarrow Cove Beds). A period of calm conditions then ensued giving variously coloured shales with goniatite bands (Merope Island Beds and Pentire Slates). A further rise effected local coarse sedimentation within the northern argillites (Gravel Caverns Conglomerates). Finally, stable conditions prevailed over the whole area, with the deposition of purple and green ostracod shales (Polzeath Slates).

Acknowledgements to G. Gauss (D. Phil. Thesis Oxford 1967).

Devonian Manganese Nodules from France

MANGANESE nodules are characteristic of present-day pelagic sediments. Price and Calvert[1] consider post-depositional processes to be the most important factors in the formation and geochemistry of the nodules, particularly diagenetic migration of manganese and related elements[2]. The solution of nodules on burial has been suggested by Manheim[3] for neritic nodules, and this could account for their scarcity in the fossil record. Fossil manganese nodules have been described from the Cretaceous of Timor[4] and the Tethyan Jurassic of Sicily[5] and the Alps[6]. In this article we report ferro-manganese nodules from Upper Devonian pelagic facies of the Montagne Noire, South France. Perseil[7] recently described some manganese minerals (including todorokite and manganite) from the Upper Devonian of Haute Pyrénées, Ariège and Aude (France), occurring as nodules, lenses and cements, concentrated by sedimentary and secondary processes.

The nodules from the Montagne Noire occur in similar lithological association to the younger manganese nodules, that is, with pelagic limestones, a facies in which they seem most likely to be preserved. The nodules are in the form of encrustations around limestone clasts and skeletal fragments, and have sizes from a few millimetres to 4 or 5 cm. Shell fragments, which are often intensively bored, are coated in a 1–5 mm thick ferromanganese pellicle. A concentric banding is occasionally seen but is usually replaced by mineral segregations and colloform structures (the limonitic cauliflowers of Jenkyns[8]). This process is attributed to a diagenetic rearrangement of Fe and Mn (ref. 9).

The composition of the nodules shows considerable variability. They are low in Fe and Mn, compared with recent nodules[1], chiefly because of the differing clay and silt content of the samples. XRF analyses of twenty nodules give the following averages: Mn, 1.8%; Fe, 9.2%; Ni, 350 p.p.m.; Mn/Fe, 0.2; Mn/Ni, 57. The low Mn/Fe ratio of these nodules is probably due to diagenetic depletion of manganese, and in fact a black or brown "halo" is often seen around the nodules (Fig. 1, lower nodule), attesting to the movement of iron and manganese.

The Devonian nodules are associated with red or, occasionally, grey nodular limestones (griotte) in the Famennian near Cabrières (Hérault) and Frasnian at Mont-Peyroux (Aude). As with the Mesozoic examples, the pelagic, condensed, hematite-rich micritic limestones yield a dominantly pelagic fauna of goniatites, conodonts, thin-shelled bivalves and pteropods, with little benthos. A further feature is that the nodules are intimately associated with the encrusting foraminifera *Tolypammina*. These occur around shell fragments and clasts, within the ferromanganese coating. In the Jurassic this foraminifera has been recorded forming "microreefs" on Mn and Fe encrusted hardgrounds[10] and has been taken to indicate a sublittoral depth (< 200 m). Similarly, Jenkyns[5] considered that the manganese nodules of Sicily formed at quite shallow depths on current-swept topographic

Fig. 7.3, p. 291.

Fig. 7.2, p. 291.

Fig. 1 Photomicrographs of iron–manganese encrustations around limestone clasts. In the upper nodule there are segregations and colloform structures (×5). The lower nodule has foraminifera (arrowed) and a black "halo" which has replaced the sediment around the nodule (×4). From Combe D'Izarne, Cabrières, Montagne Noire, France.

highs, and a similar situation is envisaged for the French nodules.

The Upper Devonian pelagic facies is also well developed in the Harz and Sauerland areas of Germany and at Chudleigh, South Devon[11], but manganese nodules seem to be absent. This could be due to a higher organic content of the original sediment, causing reducing conditions, which promotes manganese mobility. Volcanic fragments do not occur as centres to the nodules and there is no evidence of contemporaneous volcanism in the Montagne Noire during the Upper Devonian. Volcanism is present, however, in south-west England and Germany. This would suggest that volcanism is not the source of Mn, at least for these particular nodules.

I thank Dr A. Parker for analysing the nodules, and Dr R. Goldring and Dr R. Till for comments.

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EXTRACT PROCEEDINGS OF THE USSHER SOCIETY
Volume Two, Part Three, 1970, p.159

Conodonts from the Upper Devonian of the Saltern Cove—Elberry Cove area (Abstract): by M. E. Tucker and P. van Straaten.

Conodonts have been obtained from several red micritic limestone bands and blocks associated with the shales and tuff succession about the classic Saltern Cove goniatite bed. Blocks in, and immediately to the south of the goniatite bed, blocks within the limestone conglomerate, and in the first limestone band to the north of this conglomerate, have yielded conodonts of lower *quadrantinodosa* zone (to IIβ) together with older Givetian conodonts in the conglomerate. Conodonts include: *Palmatolepis distorta* Branson & Mehl, *Palmatolepis glabra elongata* Holmes, *Palmatolepis glabra pectinata* Ziegler, *Palmatolepis quadrantinodosa inflexoidea* Ziegler and *Palmatolepis quadrantinodosa marginifera* Ziegler.

The goniatites indicate an Upper Frasnian age (to Iδ) (House 1963). This surprising occurrence of Upper Frasnian goniatites with upper *Cheiloceras* zone conodonts could be due to 1). tectonic complications, 2). miscorrelation of the ammonoid and conodont chronologies when applied to south-west England or 3). to re-working of the goniatites and conodonts. Other work suggests that the first two possibilities are unlikely; it is thus tentatively suggested that the goniatites (and probably other elements of the main fauna) are derived.

To the east of Elberry Cove, conodonts show that only the Frasnian is represented, up to lower *triangularis* zone (to Iδ), slightly younger than House (1963) obtained from goniatites below.

(Samples are preserved in the Dept. of Geology, University of Reading)

THE UPPER DEVONIAN SALTERN COVE GONIATITE BED IS AN INTRAFORMATIONAL SLUMP

by PETER VAN STRAATEN and MAURICE E. TUCKER

ABSTRACT. The Saltern Cove Goniatite Bed, which contains upper Frasnian conodonts, ostracods and goniatites, is shown also to contain blocks of Famennian limestone and to lie within a sequence of Famennian sediments. From the sedimentology it is considered that slumping of Famennian limestones caused the *en bloc* movement of upper Frasnian shales. A Schwellen area (rise) nearby is suggested by the nature of the derived limestone blocks.

Saltern Cove (Grid. Ref. SX 895585) 5 km south of Torquay, south Devon (text-fig. 1) is a classic locality for Upper Devonian goniatites. It was first described by Lee (1877) and subsequently by Ussher (1903), Anniss (1927), Donovan (1942) and House (1963). The goniatites are now ascribed to the Holzapfeli zone of the Frasnian (I δ) Upper Devonian. Recent work by the authors (briefly reported in Tucker and van Straaten 1970a) has shown that the stratigraphical context of the goniatites is not quite so simple as would first appear. Previously, all the Upper Devonian in Saltern Cove and Waterside Cove has been considered to be of Frasnian age, apart from the last 20 to 30 m at the north end of Waterside Cove which were attributed to the lower Famennian. Results presented here, show that all the succession from the north end of Saltern Cove is Famennian in age, though with several beds of older intraformational derived sediment. These include the Goniatite Bed.

STRATIGRAPHY AND STRUCTURE

The Devonian in South Devon is tectonically complicated and in the Torquay area, Richter (1969) has recognised four phases of structural deformation. In the Saltern Cove area the bedding is near vertical and youngs northwards. There are several small faults and a strong cleavage is present, which is almost horizontal or dips south-east at a few degrees.

In the central part of Saltern Cove a 3 m thick bed of massive limestone (dated on coral evidence as middle Frasnian by Scrutton 1967) is succeeded by about 20 m of alternating limestone and red shale bands. The limestones are composed of corals, crinoids and other carbonate debris and represent the talus from a carbonate producing area nearby. The northern part of the bay is occupied by some 25 m of purple and red mudstones, with rare crinoidal limestones, which are of lower Famennian age (*Cheiloceras* Stufe) at the top. These are terminated by a fault running ESE-WNW which forms the northern side of the cove (text-fig. 1). The succession from this fault to the north end of Waterside Cove is (youngest beds, most northern, at the top):

Lower Devonian

- - - - Fault - - - -

		Dated Range
40 m	Purple mudstones with tuff and shell bands (central part of Waterside Cove).	IIβ-IIIα.
1 m	Limestone conglomerate (described by Holwill 1966).	clasts: Givetian, I and IIβ-IIIα.
7 m	Red shales with nodules and tuff bands (southern part of Waterside Cove).	IIβ-IIIα.
1 m	Red shale with numerous derived limestone and tuff fragments.	IIβ-IIIα.
9 m	The Saltern Cove Goniatite Bed. Red silty shales with many limestone clasts. (most southern part of Waterside Cove and promontory between Waterside and Saltern Cove).	shales Iδ. clasts IIβ-IIIα.
20 m	Red and purple shales with many tuff bands (coastal platform between Waterside and Saltern Cove).	IIβ-IIIα.

- - - - Fault forming northern side of Saltern Cove - - - -

AGE DETERMINATIONS IN THE VICINITY OF THE GONIATITE BED.

Beds above and below the Goniatite Bed.

Samples from tuffs and calcareous nodules collected above and below the Goniatite Bed all yield conodonts of the quadrantinodosa zone (IIβ- IIIα, upper Cheiloceras - lower Platyclymenia, Famennian). Location of the samples is shown in text-figs. 1 and 2, and conodont determinations in ~~table~~ text-fig. 3. Limestone clasts from the conglomerate (Holwill 1966) give a range of ages; blocks with corals are Givetian-lower Frasnian (Holwill 1966), whilst fine grained clasts give Frasnian and lower Famennian ages on conodonts (samples 18, 19 and 20).

Immediately below the Goniatite Bed, and outcropping on the south face of the promontory, is a thin but continuous calcareous band (sample 6) which is particularly rich in conodonts (often broken), thin shelled bivalves, crinoid fragments and ostracods. The ostracods, with their stratigraphic range after Blumenstengel (1965) are Acratia spp., Amphissitis bispinosus (upper Cheiloceras-Platyclymenia, II β -III) and Ceratacratia cerata (II α -VI). Purple mudstones at the north end of Saltern Cove (sample 1, text-fig. 1) yielded Entomozoe (Nehdentomis) nehdensis Matern 1929 indicative of the lower Famennian (Cheiloceras Stufe).

The Goniatite Bed.

a) Shales. The goniatite fauna, recently revised by House (1963), belongs to the Holzapfeli zone (I δ) of the upper Frasnian. The fauna is dominated by the genus Archoceras. The goniatites are usually a centimetre or less in diameter and are haematised. Most have been obtained from the shales in the cliff section just north of the promontory (text-fig. 2). Goniatites also occur on the south face of the promontory (M.R. House, pers. comm., 1969). In addition, the authors have found goniatites at the same horizon in the vicinity of the large tuff block figured by Holwill (1966), 10 m west of the promontory. Other fossils present in the shales include Buchiola spp., orthocones, trilobites, brachiopods, crinoids (Anniss 1927) and haematised ostracods identified as Entomoprimitia spp. (upper Frasnian) and Entomozoe (Nehdentomis) tenera Rabien 1954 (Frasnian - lower Famennian, I-II). Conodonts occur in small 'reduction' centres: Palmatolepis gigas Miller and Youngquist 1947, Palmatolepis subrecta Miller and Youngquist 1947, Palmatolepis triangularis Sannemann 1955 and Polygnathus spp. These indicate an upper Frasnian age. The goniatites, ostracods and conodonts all indicate a similar age.

b) Calcareous clasts within the Goniatite Bed.

Within the Goniatite Bed are a number of angular and rounded limestone blocks (sample numbers 7 to 16, text-fig. 3). Most are about 20 cm x 10 cm x 10 cm, but some reach up to 2 m x 50 cm x 30 cm. The block of calcareous tuff figured by Holwill (1966) measures 3 m x 1 m x 50 cm (sample 8). All yielded conodonts of the quadrantinodosa zone (upper II β - lower III α) or rhomboidea - quadrantinodosa zones (II β - lower III α). Other fossils found in the limestone were crinoid fragments, orthocones, ostracods, gastropods and one goniatite (species indeterminate). The Goniatite Bed (red shales with limestone clasts) and the limestone conglomerate can also be found at the north west corner of Saltern Cove. Conodonts from limestones here (samples 2 and 3) again show the presence of the lower Famennian.

STRATIGRAPHICAL INTERPRETATION

The data presented above show the apparently anomalous situation in the Goniatite Bed where a shaley sediment contains blocks of reworked limestone clasts, two whole goniatite zones younger than the enclosing sediment and the adjacent shales. There are three possible explanations:

1) Mis-correlation of goniatites and conodonts.

The ostracods and conodonts in the shales are the same age as the goniatites. This shows that there is no local mis-correlation of the goniatite and conodont chronologies and that the German system (Ziegler 1962) is applicable to the area. Indeed, at Chudleigh, 30 km NNW of Saltern Cove, the conodont succession from the Cheiloceras Stufe shows complete agreement with the German chronology and with the goniatite succession (Tucker and Straaten 1970b).

2) Tectonic explanation.

From the bedding/cleavage relationship around the promontory (bedding vertical, cleavage horizontal) a fold cannot possibly be constructed so that the Goniatite Bed occurs in the centre of a tight anticline, unless such a fold was developed during an earlier deformational phase. However, the beds above and below the Goniatite Bed show normal grading (northwards), indicating that early folding did not take place. From text-fig. 4, it is clear that faulting and /or shearing has not occurred around the Goniatite Bed. The tectonic deformation is only such that the limestone blocks have been rotated by the cleavage, and once continuous ash bands are sheared. A complication of the tectonics then is most unlikely to have caused these Frasnian shales to occur in a Famennian succession.

3) Large-scale slumping of sediment.

A sedimentary explanation means that the youngest elements in the Goniatite Bed, the limestone clasts, slumped into older muddy sediments and that the whole then slumped into a muddy succession undetectably different in age to the clasts. Evidence that this was the case can be seen from an examination of the sedimentology of the succession.

<u>Lamination of sediments</u>. A petrographic examination of the shales of the Goniatite Bed shows that they are quite distinct from the shales above and below in one important respect, that good lamination is absent. In spite of the strong cleavage, thin continuous laminae of medium to coarse silt, 0.5 mm thick, occur in the shales above and below (text-fig. 5). In the goniatite shales, however, the same amount of silt is present but is either irregularly distributed or present in discontinuous lenses. Numerous graded tuff bands from a few millimetres to 40 cm thick occur at intervals of 20–30 cm above and below the Goniatite Bed and can be traced for up to 70 m across the wave-cut platform. A small channel structure is also exposed here. However in the shales of the Goniatite Bed macroscopic laminations of this type are completely absent. It is suggested

that penecontemporaneous movement destroyed the primary lamination. There is no evidence of any erosional structure between the Goniatite Bed and the underlying laminated shales.

Limestone clasts. The blocks of Famennian limestone and tuff are clearly derived. There is some concentration of clasts towards the base of the Goniatite Bed, but many occur randomly scattered throughout the shale. The limestones are micritic with an appreciable amount of silt. Some are almost completely dolomitized. Prominent pressure solution planes (stylolites) occur in some blocks and there are also cavity-fill structures. A few clasts contain a network of cavities filled first by ~~radiaxial~~ fibrous calcite, and then silty shale material. Many of the limestone blocks are reminiscent of the condensed pelagic facies (Schwellen facies) which at Chudleigh and commonly in Germany follows Givetian/lower Frasnian shallow-water massive limestones.

Another band with many tuff and limestone fragments occurs immediately above the Goniatite Bed, and conodonts suggest that the clasts are the same zonal age as the shales. The limestone conglomerate (7 m above the Goniatite Bed) described by Holwill (1966) contains limestone fragments of various types and ages. The bed is graded and Holwill suggested a proximal turbidity current origin.

COMPARISONS

Three types of sediment movement have occurred at Saltern Cove (a) of more frequent occurrence, movement of lithified shallow-water sediments (limestones) downslope into a deeper water environment, where the slumped material is generally older than the host sediments (limestone conglomerate) (b) slumping of lithified shallow-water sediments downslope where the slumped clasts are of the same zonal age as the deeper water sediments and (c) en bloc slumping of shales and pelagic limestone clasts where the shales are older than the host sediments and clasts. Slumping of the second type, with no stratigraphical break, is typical of most intraformational conglomerates and is known in a similar situation in the Upper Devonian of the Harz Mountains, Germany (Stoppel and Zscheked 1963). Slumping of the third type is less common. It is only detectable where either the stratigraphical break is large enough or sufficiently fine dating is possible. In the absence of the latter only the lack of lamination might give a clue to the detailed history.

SIGNIFICANCE OF SLUMPING AT SALTERN COVE

The two pelagic facies of the Upper Devonian, the condensed limestones (Schwellen or rise facies) and ostracod shales (Becken or basinal facies) are developed in south west England, though the rise facies is only well known at Chudleigh (House 1963, Tucker and Straaten 1970). However, it may be present at Petit Tor Combe (Grid. Ref. SX 926665, 4 km northeast of Torquay) where patches of nodular limestone containing middle Frasnian goniatites were recorded by Ussher (1890). The basinal facies is well developed in the Torquay district. The blocks of Famennian limestone at Saltern Cove indicate that pelagic carbonate was deposited in the vicinity on a topographic high. The well known Torquay and Brixham shallow-water limestones are likely source areas, or the massive limestone immediately to the west at Goodrington. Holwill (1966) considered volcanism as the immediate cause for the limestone conglomerate. However, for the Goniatite Bed, instability on a slope and movement of the underlying sediment is as likely an explanation. Movement may have been initiated by volcanic activity.

CONCLUSIONS

1) The Saltern Cove Goniatite Bed is a slumped deposit with Frasnian shales and Famennian limestone blocks within a sequence of early Famennian sediments.

2) The sedimentology of the finer grained sediments provides a clue to the stratigraphical interpretation.

3) The type of slumping is unusual since en bloc movement of shales has occurred, which are older than the contained limestone clasts and associated sediments.

4) The derived Famennian pelagic limestones indicate the former presence of a nearby Schwellen area.

5) Where tested the German goniatite, conodont and ostracod zonation is applicable and confirmed in South Devon.

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Text-Fig. 1. Geological sketch-map of Saltern Cove and Waterside Cove, South Devon (after mapping by the authors), showing location of conodont samples, 1,2,3 and 24.

Text-Fig. 2. Sketch of the promontory between Saltern Cove and Waterside Cove, viewed obliquely, showing location of conodont samples 4 to 23.

Text-Fig. 3. Conodont determinations from Saltern Cove and Waterside Cove, South Devon. The specimens are deposited in the archives of the Department of Geology, University of Reading, where they are numbered R.U.S 22901-22924. o = cf. determination.

The conodont zones represented by the samples (after Ziegler, 1962; 1965, and Klapper, 1966) are as follows:-

Sample 20 asymmetrica zone (I α).

Samples 1, 5, 7, 11, 12, 15, 16 rhomboidea -quadrantinodosa zones (II β - lower III α).

Samples 3, 4, 6, 8, 10, 17, 18, 19, 21, 24 quadrantinodosa zone (upper II β - lower III α).

Samples 13, 14, 22 lower quadrantinodosa zone (upper II β).

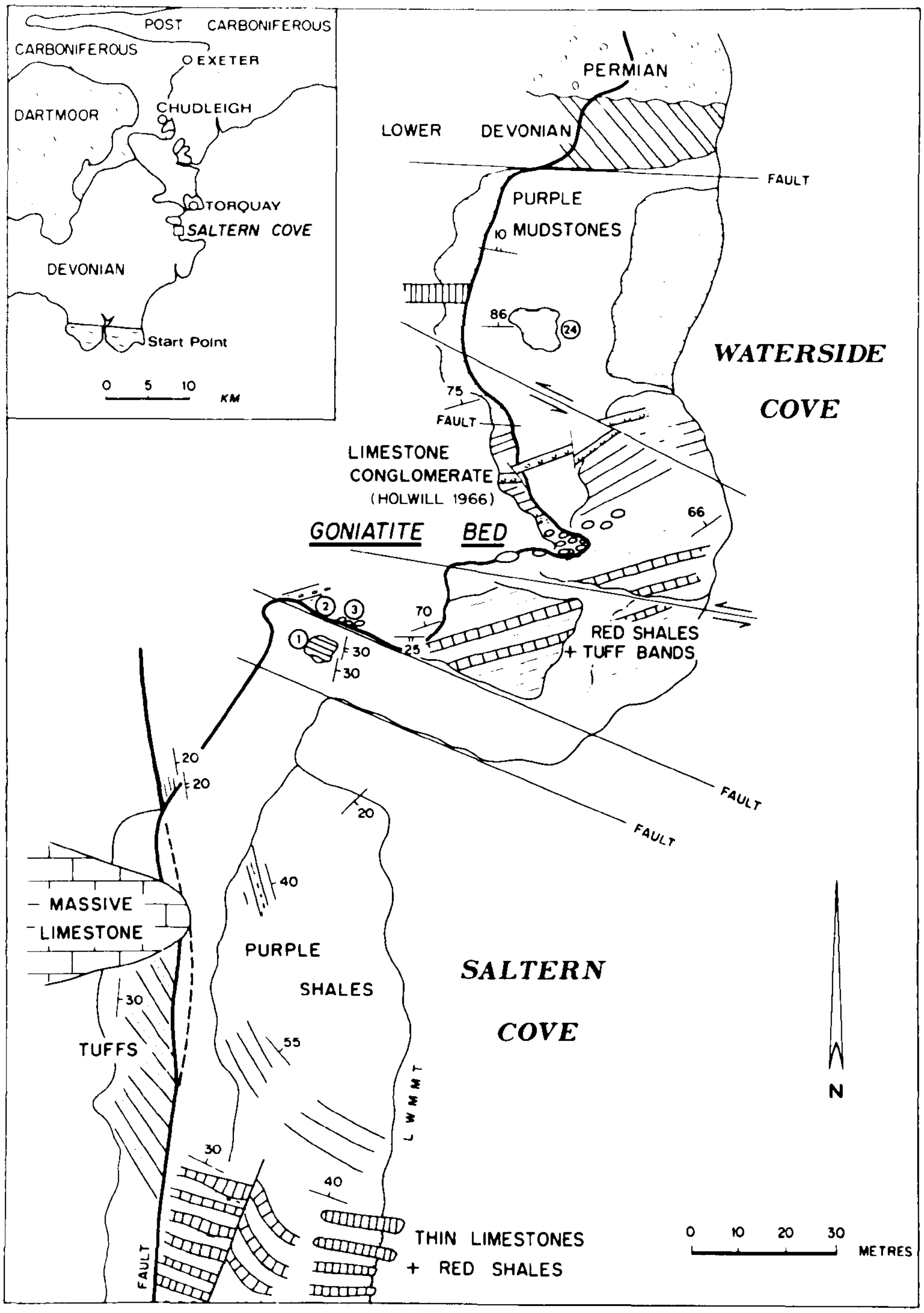
Sample 2 quadrantinodosa - middle velifera zones (upper II β - III β).

Sample 9 upper triangularis - upper velifera zones (I δ - III β).

Text-Fig. 4. The Saltern Cove Goniatite Bed. Location of samples indicated.

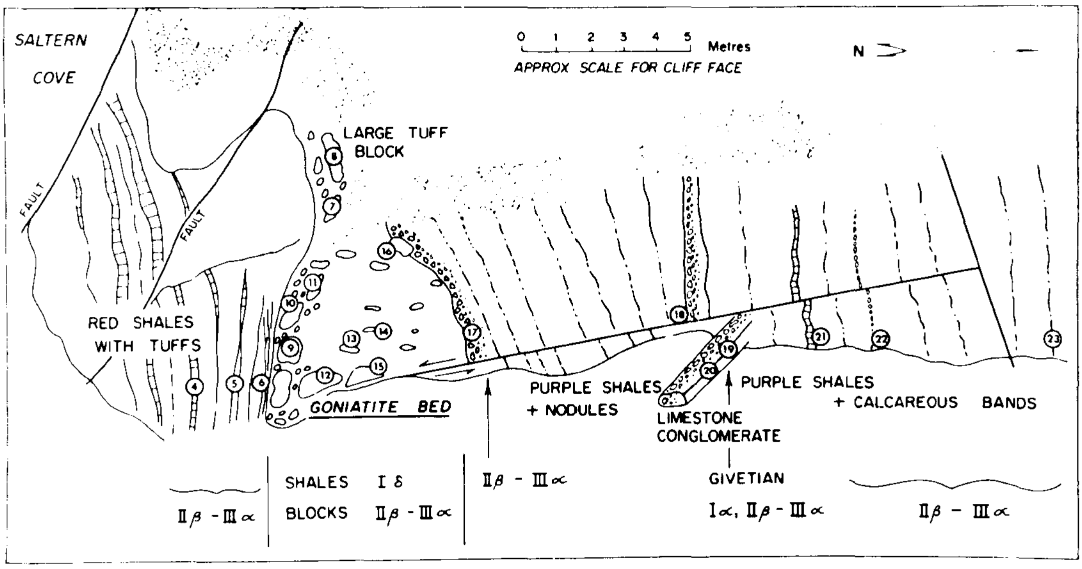
Text-Fig. 5. Photomicrograph of shales above the Goniatite Bed. Good lamination of silt is developed.

Straaten and Tucker (1971, in press)



Text - Fig. 1.

Straaten and Tucker (1971, in press)



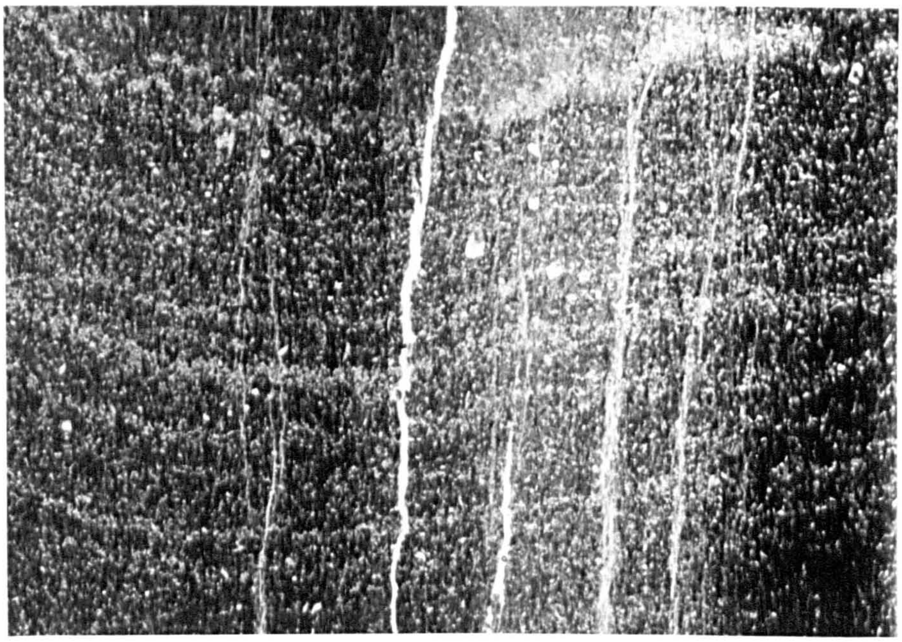
Text - Fig. 2.

	SALTERN COVE			BELOW GON. BED			LIMESTONE BLOCKS WITHIN GONIATITE BED										ABOVE GONIATITE BED							
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24
Ancyrodella rotundiloba BRYANT 1921																				x				
Icriodus spp										x												x		
Palmatolepis distorta BRANSON & MEHL 1934		x		x																				x
Pal glabra lepta ZIEGLER & HUDDLE 1969								x								o		x			x	x		
Pal glabra pectinata ZIEGLER 1960				x	x		x			x	x	x		x	x		o	o						x
Pal glabra prima ZIEGLER & HUDDLE 1969	x		x		x						x													
Pal gracilis gracilis BRANSON & MEHL 1934				x																				x
Pal minuta minuta BRANSON & MEHL 1934					x	x	x		x	x	x	x	x	x		x		x						x
Pal perlobata schindewolfi MULLER 1956				x	x								x						o					x
Pal quadrantinodosa inflexa MULLER 1956														x								x		
Pal quadrantinodosa inflexoidea ZIEGLER 1962			x										x											
Pal quadrantinodosa marginifera ZIEGLER 1960			x	x	x	x		x		x			o				x	x	x		x	x		x
Pal spp		x				x			x											x	x			
Polygnathus diversa HELMS 1959																								x
Pol glabra glabra ULRICH & BASSLER 1926										x									x					
Pol nodocostata s.l. BRANSON & MEHL 1934				x		x			x	x		x	x		x	x	x	x				x		x
Pol nodocostata cf *pennatuloidea* HOLMES 1928					x																			
Pol nodoundata HELMS 1961				x																	x			
Pol triphyliata ZIEGLER 1960													x	x	x									
Pol spp				x		x	x	x	x	x		x	x	x	x				x			x		x
Polylophodonta confluens ULRICH & BASSLER 1926																			x					
Spathognathodus amplus BRANSON & MEHL 1934															o									
Spathognathodus spp				x		x		x		x			x		x						x		x	

Text - Fig. 3.

Straaten and Tucker (1971, in press)

Text- Fig. 4.



Text - Fig. 5.

Nature — Letter — T 6199 R — 229 — 9-10 x20 halfems
RADIAXIAL FIBROUS CALCITE — *in situ* replacement origin after syn-sedimentary cement — — — — — — —

Radiaxial Fibrous Calcite as a Replacement after Syn-Sedimentary Cement

THERE is considerable controversy[1] about the origin of the coarsely crystalline calcite mosaics which have been found filling cavities in reefs and mudmound complexes of varying geological age around the world. The calcite mosaics are composed of fibrous crystals, and there is doubt about whether they represent primary precipitates (cements) or are replacive, and about the time and environment of their formation. Orme and Brown[2] consider some fibrous calcite to be replacive, but usually the external morphology of the mosaic and its association with internal sediments suggest a cement origin[3,4]. Direct precipitation of fibrous calcite from freshwater, marine water, or brines trapped in the sediments (or derived from neighbouring sediments) have been suggested[1]. Newell[5], Cotter[6] and Zankl[7], on the other hand, suggest that fibrous calcite filling cavities was originally aragonite.

In this article we consider only those fibrous calcite mosaics in which the crystals possess an internal structure termed "radiaxial" by Bathurst[3,8] and characterized by curved twin lamellae and glide planes (the "cleavage" if Bathurst), optic axes that converge away from the cavity walls and subcrystals which diverge in this same direction and thus cross cut the optic axis convergence. Such mosaics are present within sheet cracks and "Stromatactis"-like cavities within Upper Devonian pelagic limestones of the Rheinisches Schiefergebirge and the Montagne Noire. We believe that these mosaics represent replacements of earlier cements which were precipitated soon after deposition.

In the examples illustrated (Figs. 1 and 2), the lower parts of horizontal cavities with irregular roofs are filled by laminated sediments (internal sediment) and by radiaxial fibrous calcite, but the cavities are sharply truncated by smooth, planar and horizontal erosion surfaces. These surfaces also smoothly truncate the host-sediment, fossils and ferromanganese nodules (Fig. 1). The presence of the internal sediments with their horizontal upper surfaces (thus forming a geopetal structure) indicate that the cavity was once a void and also strongly suggests that the overlying fibrous calcite is a void-filling cement.

Fig. 1 Upper and side views of smooth, planar erosion surface truncating *Stromatactis*-like cavities (partially filled with internal sediment) and ferromanganese encrusted shells. Upper Devonian Griotte, Mont Peyroux, Herault, Montagne Noire. S 22951. Fig. 3.25, p. 79.

Fig. 2 Photomicrograph of truncated sheet-crack. Fibrous radiaxial mosaic nucleated from erosion surface (arrowed). The crystals are turbid with inclusions, increase in size away from erosion surface and contain *in situ* fills of borings. Upper Devonian Cephalopodenkalk, Bicken, Rheinisches Schiefergebirge. Thin section S 22950. Plane polarized light. The bar represents 1 mm. Fig 3.25, p. 92.

The sharpness and smoothness of the cavity truncation indicates that (1) the erosion surface was cut by abrasion rather than by solution; (2) pre-erosional lithification of the host-sediment took place; and (3) the cavities were formed and filled by internal sediment and a cement before erosion took place. On the other hand, the relation between the fibrous calcite filling the cavity and the erosion surface suggests that the first of these was formed after erosion and therefore cannot be of the pre-erosional (and syn-sedimentary) cement generation.

The size of the crystals in the fibrous mosaic increases away from the cavity walls, upwards from the upper surface of the internal sediment and downwards from the erosion surface. The last occurrence is anomalous because if the fibrous calcite were the pre-erosional cavity-filling cement, the zone of crystal size increase should have been associated with the original cavity roof and not with the later erosion surface. The original roof and its associated zone of finer crystalline cement should have been removed by erosion. The association of this zone with the erosion surface implies that the surface must have been present before the development of the fibrous mosaic (the crystals having nucleated from this surface). The cavities and their present fills of fibrous calcite are thus of different age—the cavities are pre-erosional and the fibrous calcite post-erosional.

Fig. 3 Diagrams illustrating the formation of an idealized truncated cavity and its later fill by fibrous radiaxial calcite. *A*, Formation of cavity within partially lithified pelagic sediment and partial filling by internal sediment; *B*, cavity completely filled by syn-sedimentary cement and lithification of host sediment; *C*, erosion of cavity and its cement fill; organisms penetrate erosion surface and their borings are later filled with fine-grained carbonate; *D*, sedimentation resumes followed by *in situ* replacement of cement fill to radiaxial fibrous calcite mosaic (seeded on cavity walls, surface of internal sediment and erosion surface); internal sediment recrystallization to microspar and boring-fills to micrite; erosion surface thus truncates cavity but not the fabric of the calcite fill.

Nevertheless, the cavities were filled by a cement before erosion, for they are smoothly truncated. The material causing the abrasion has not entered the cavities and the fibrous calcite must therefore be either a replacement of this pre-erosional cement or a cavity-filling cement which was precipitated after the earlier cement had been dissolved away. There is no evidence for a solution phase in the history of the cavities and so this explanation can be excluded as the fills of borings (which penetrated downwards from the erosion surface into the pre-erosional cement) are still *in situ*. (A void-stage would have allowed them to fall to the cavity-floor.)

The Upper Devonian pelagic limestone exhibit no evidence of subaerial exposure during their deposition and probably accumulated at depths of 50 m or more (M. E. Tucker, unpublished). The pre-erosional cement of the cavities was therefore precipitated from marine (or marine-derived) waters.

Another feature of radiaxial calcite crystals which suggests their *in situ* replacement origin is the presence within them of brownish inclusions—this feature seems to be characteristic of radiaxial fibrous mosaics in general. Recent submarine cements are commonly brown or yellowish and contain disseminated organic matter, clay or colloidal material. This non-carbonate material seems to have been retained during *in situ* replacement and forms the inclusions within the calcite crystals. In some crystals inclusions may be so numerous as to impart pseudo-pleochroism to the calcite, a phenomenon also known from *in situ* calcite replacements after skeletal aragonite[9-11].

The enlargement and systematic orientation of crystals within radiaxial fibrous mosaics are commonly thought to suggest a void-filling cement origin. Identical fabrics, however, are to be expected from any process that begins at a surface and progresses outwards[12]. For radiaxial fibrous calcite, the walls of cavities, the upper surfaces of internal sediments and erosion surfaces have acted as discontinuities on which the replacement calcite crystals have nucleated. The non-planar intercrystalline boundaries, which are also a characteristic feature of these mosaics, are more suggestive of replacement mosaics than of those with crystals that precipitated onto free surfaces[12].

We suggest that the evidence provided by the relationships between cavities, their radiaxial fibrous calcite fills and erosion surfaces can only be interpreted as demonstrating the *in situ* replacement origin of radiaxial fibrous calcite after a pre-existing and syn-depositional cement generation. The origin of the characteristic features of radiaxial fibrous calcite mosaics and the date of the replacement will be discussed elsewhere.

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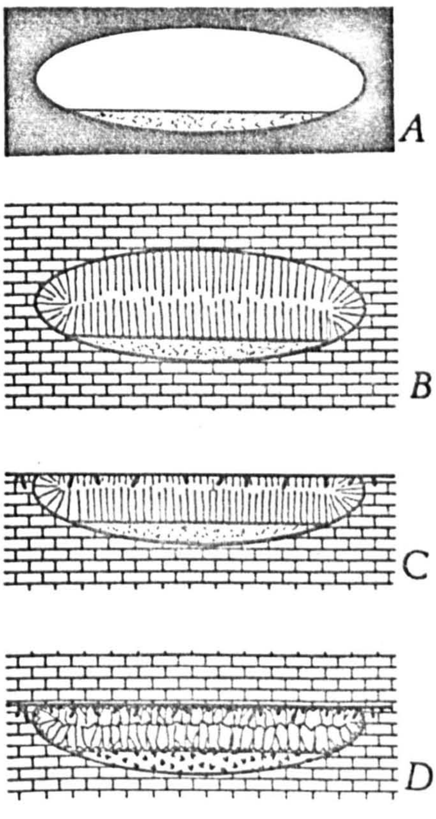
Fig. 3.



EXTRACT

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Abstract. Age determinations from the Chudleigh area S. Devon are presented, which show that the Kiln Wood Beds are equivalent in age to the Lower Dunscombe Goniatite Bed. The Kiln Wood Beds are interpreted as being deposits of a local deeper part of the Chudleigh **Schwelle.** Conodonts and facies are also considered for the lower part of the Mount Pleasant Series (Lower Famennian).

1. Introduction

For a consideration of facies changes, accurate dating is essential, and we are fortunate in the Upper Devonian that exact correlations can be made through conodonts, which are very common at this time. Good conodont faunas have been collected from Kiln Wood Quarry (SX 861779), the section in the road running past the quarry, and from the track leading to Winstow Cottages (Fig. 1). Other conodonts have been obtained from outcrops above Palace Quarry in the Riding Parks (SX 869787), and from the old quarry at Lower Dunscombe Farm (SX 886791).

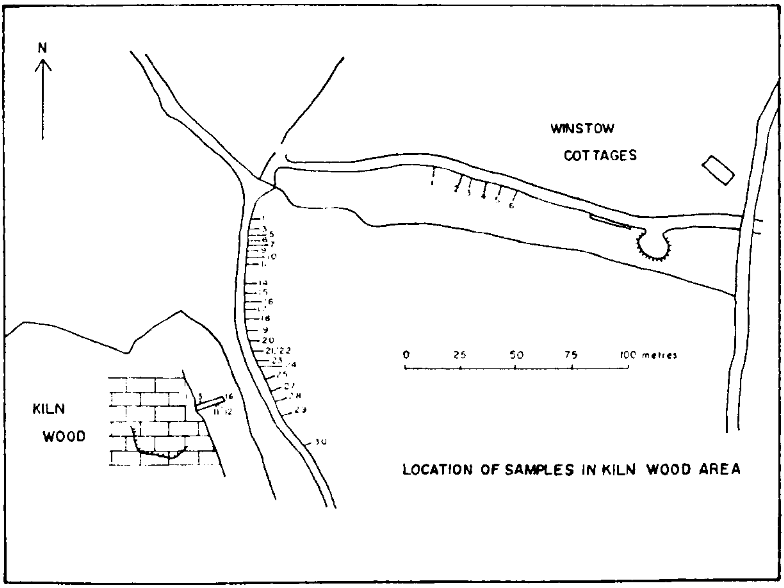

FIGURE 1. Map showing location of conodont samples in the Kiln Wood area, Chudleigh, South Devon.

A continuous but much reduced Upper Devonian sequence (House and Butcher 1962) of nodular limestones and shales occurs in the Chudleigh area on top of massive limestone, which is in part Frasnian in age. The thickness given for the Famennian is 53 m and it has been termed the Mount Pleasant Series by House (1963). House has described the ammonoid succession in the Chudleigh district and shown that all the German ammonoid *Stufen* are present, and that many of the species zones can be recognised too. House also shows that the Mount Pleasant Series is of *Schwellen* type (Schmidt 1926), that is a succession of condensed limestones, nodular limestones and shales with calcareous nodules thought to have been deposited on a submarine rise. In the Chudleigh case, the *Schwellen* sediments occur above Middle and lower Upper Devonian coral/stromatoporoid limestones. In contrast to the

UPPER DEVONIAN	FAMENNIAN	Wocklumeria	VI	Spathognathodus costatus	U
					M
		Clymenia – Stufe (to V)	V/VI		L
			V	Polygnathus styriaca	U
					M
					L
		Platyclymenia – Stufe (to III - to IV)	IV	Scaphignathus velifera	U
			III β		M
			III α		L
		Cheiloceras – Stufe (to II)	β	P. quadrantinodosa	U
					L
				P. rhomboidea	
			α	P. crepida	U
					M
					L
	FRASNIAN	Manticoceras – Stufe (to I)	δ ?	P. triangularis	U
					M
			δ		L
				P. gigas	U
			γ		L
				A. triangularis	
			β	Polygnathus asymmetrica	U
			α		M
					L

FIGURE 2. Chart showing correlation of ammonoid *Stufen* and conodont zones for the Upper Devonian (after Klapper 1966). "P." is abbreviation for *Palmatolepis*.

Schwellen facies, *Becken* sediments consist of ostracod shales sometimes with volcanic horizons and turbidites, and were deposited in deeper water (basinal) areas. These are also present in South Devon and occur in the Newton Abbot and Torquay districts (House and Selwood 1964). Scrutton (1969) has shown that the coral faunas in the early Frasnian massive limestone indicate a gradually deepening environment. This trend continues throughout the rest of the Upper Devonian.

In this paper the conodont zonation of Ziegler and others for the Upper Devonian is used, and a full list of the references employed in the determinations is appended. Conodont zones and their relation to Upper Devonian orthochronology is shown in Figure 2.

2. Conodonts and Facies

Conodonts show that the Kiln Wood Beds (Goldring et al. 1967, fig. 4) of Middle/Upper Frasnian, are equivalent in age to the Lower Dunscombe Goniatite Bed, which contains (House 1963) the zone fossil *Manticoceras cordatum*. These two units both occur above massive limestone but are very different lithologically. They represent deposition in quite different environments, though both related to a *Schwelle*. In the field, the two facies occur only 1½ km apart, though there is probably some horizontal movement between the quite incompetent Kiln Wood Beds and the underlying massive limestone.

The nodular limestones of the Lower Dunscombe Goniatite Bed (1.7 m thick) are typical *Schwellen* carbonates, greyish-pink micritic limestones with many pressure solution planes and very thin irregular shaly horizons. These sediments yield a rich and varied fauna of goniatites, brachiopods, trilobites, rare corals (*Syringaxon* sp., identified by Dr. C. T. Scrutton) and bivalves including *Buchiola* sp. The dominantly pelagic fauna and restricted benthos suggests deposition at greater depth than in the case of the underlying limestone. There is evidence of small-scale sedimentary dyke formation. Dyke fillings consist of a coarse grained red carbonate containing crinoids.

Conodonts obtained at Dunscombe Farm show that the massive limestone (sample 1) below the goniatite bed is also Middle Frasnian in age (Table 1). Sample 4 from the top of the goniatite limestone shows that the upper *gigas* conodont zone is present, which is equivalent to the lower part of the *holzapfeli* goniatite zone of Upper Frasnian age (to Iδ).

TABLE 1. Conodonts from Lower Dunscombe Farm and Kiln Wood Quarry. Location of samples for Kiln Wood shown in Figs. 1 and 2. For Dunscombe Farm, samples 0 and 1 from bottom and top of massive limestone respectively. Samples 2, 3 and 4 from the goniatite limestone. Sample 4, 15 cm from top of quarry. (o = cf. determination).

Age of samples

Lower Dunscombe Farm

Sample 0: middle *asymmetrica* to lowermost *gigas* zone. 1: upper *asymmetrica* to lowermost *gigas* zone. 2: lowermost part of upper *gigas* zone. 3: lower part of upper *gigas* zone. 4: upper part of upper *gigas* zone.

Kiln Wood Quarry

Sample 1: as sample 1, Dunscombe Farm. 3: as sample 1, Dunscombe Farm. 11: *Ancyrognathus triangularis* to lowermost upper *gigas* zone. 12: lower to lower upper *gigas* zone. 16: *crepida* zone.

	Dunscombe					Kiln Wood				
	0	1	2	3	4	1	3	11	12	16
Ancyrodella curvata		x	x	x	x	x	x	x	x	
A. gigas		x					x			
A. ioides									x	
A. lobata	x					o				
A. nodosa							x		x	
Ancyrodella sp. indet. . .					x					
Ancyrognathus asymmetrica .			x	x	x		o			
A. triangularis			x	o				x	x	
Icriodus symmetricus . .			x						x	
Icriodus sp. indet.	x	x		x	x	x	x			
Palmatolepis crepida linquiformis					x					
P. gigas				x					x	
P. hassi								x		
P. minuta minuta										x
P. punctata	x	x				x	x			
P. punctata var. transitans .							x			
P. quadrantinodosalobata . .										x
P. subrecta				x	x		x	x	x	
P. tenuipunctata										x
P. unicornis					x				o	
Palmatolepis sp. indet. . .			x							
Polygnathus sp. indet. . .	x				x	x	x		x	

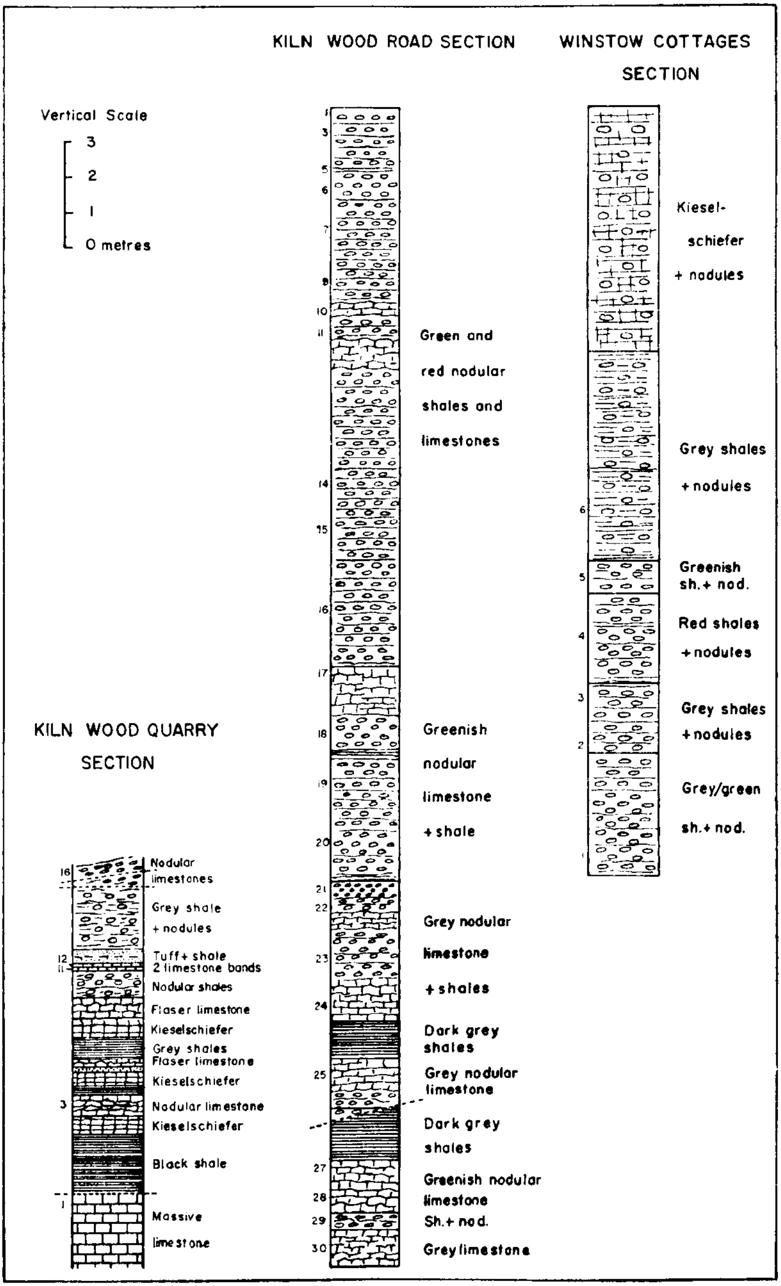

FIGURE 3. Upper Devonian succession in the Kiln Wood area, Chudleigh.

The Kiln Wood Beds (9 m thick) exposed in Kiln Wood Quarry consist of black and dark grey micaceous shales which contain siliceous horizons (*Kieselschiefer*) and three flaser limestones (Fig. 3). A flaser limestone is a relatively pure (85-95% calcium carbonate) sometimes nodular limestone containing irregular streaks or ' veins ' of clay material at all angles to the bedding (*Flaserkalk* of Gründel and Rösler 1963). The dark shales containing pyritised fossils indicate deeper water sedimentation of pelagic material under anaerobic conditions. The flaser limestones, which have been partly silicified, contain many small bivalves, mostly disarticulated, and crinoid ossicles, probably brought in by bottom currents. Fixosessile arenaceous foraminifera occasionally occur attached to the lamellibranch valves and attest to the presence of aerated bottom waters. Two tuff horizons are also developed. Conodonts show that most of the Kiln Wood Quarry succession is Middle Frasnian, including the top of the massive limestone (Table 1). At the top of this section the grey nodular limestone (sample 16) is not *in situ;* conodonts give a *crepida* age (to IIα).

It is suggested that the Kiln Wood Beds represent deposition in a localised deeper part of the *Schwelle.* Similar ' special basins ' occur along the Diabas-Schwelle in the Oberharz, Germany (Straaten 1969). Here, sediments traditionally regarded as basinal, occur with shallower water *Schwellen* limestones in local depressions on the volcanic rise. Local areas of lower relief were sites of accumulation of fine grained terrigenous material, though periods of higher and more widespread carbonate precipitation over the *Schwelle* gave rise to flaser limestones in the special basin succession. In the Chudleigh area, tectonics and lack of exposure do not permit an elucidation of the shape of the basin.

In the road running past Kiln Wood Quarry, there is a continuous outcrop for 150 m of Famennian nodular limestones and shales, belonging to the Mount Pleasant Series. This *Kalkknollenschiefer* succession dips and youngs north. The conodonts obtained belong to the *Cheiloceras Stufe* (to II), conodont zones *crepida, rhomboidea* and the lower part of the *quadrantinodosa* zone (Table 2). Rocks of this age and lithology also crop out in the Riding Parks above Palace Quarry, 1 km north-east of Kiln Wood, and yielded conodonts of the *crepida* zone. The conodonts from this nodular limestone group agree with ages obtained from ammonoids by Prof. M. R. House.

TABLE 2. Conodonts from the Kiln Wood road section. Location of samples shown in Figs. 1 and 2.

Age of samples

Samples 30 to 24: *crepida* zone. 24 to 21: lower to middle *crepida* zone. 18 and 17: upper *crepida* zone. 11 and 10: *rhomboidea* zone. 6 to 1: lower *quadrantinodosa* zone.

	30	29	28	27	25	24	23	22	21	20	19	18	17	16	15	14	11	10	9	7	6	5	3	1
Ancyrognathus sinelamina .		x										x	x											
Icriodus sp. indet.		x				x	x	x			x						x		x					
Palmatolepis crepida crepida .	x			x	x			x	x															
P. glabra elongata . . .																					x			
P. glabra glabra . . .											x	x	x	x	x							x		
P. glabra pectinata . . .												x		x			x				x			x
P. gracilis gracilis . . .																	x				x			x
P. minuta minuta . . .	x		x	x	x	x		x	x	x	x	x	x		x	x	x		x	x	x	x		x
P. quadrantinodosalobata .	x	x	x	x	x	x	x		x				x											
P. cf. regularis								x	x															
P. rhomboidea . . .																		x						
P. subperlobata subsp. a Helms .												x		x										
P. tenuipunctata . . .	x	x	x	x	x	x	x	x	x											x				
P. termini						x	x		x															
Polygnathus glabra glabra .												x	x					x			x			
P. nodocostata s. l. . .		x																	x				x	
Polygnathus sp. indet. . .								x			x				x			x	x	x	x	x		
Spathognathodus sp. indet. .																	x							x

Samples taken from outcrops along the track to Winstow Cottages (Fig. 1) yielded conodonts of the *styriaca* zone (Table 3). A more argillaceous succession occurs here containing less carbonate than the underlying *Kalkknollenschiefer,* and consisting of larger calcareous nodules (up to 10 cm across) in grey or green shales. Towards the top of the Upper Devonian (*Wocklumeria Stufe*) *Kieselschiefer* are present (House 1963), which continue into the Carboniferous (Matthews 1969).

TABLE 3. Conodonts from outcrops along track to Winstow Cottages. Location of samples shown in Figs. 1 and 2. (o = cf. determination). All samples are of *styriaca* zone.

	1	2	3	4	5	6
Palmatolepis distorta manca .				x		
P. gracilis gracilis . . .	x		x	x	x	x
P. gracilis sigmoidalis . .				x	x	o
P. helmsi				o	x	
P. maxima					x	
P. perlobata schindewolfi .	x		x	x		
P. perlobata sigmoidea . .				x		
P. schleizia				o		
Palmatolepis sp. indet. . .		x	x			
Polygnathus styriaca . . .	x			x	x	
Polygnathus sp. indet. . .		x				
Spathognathodus stabilis . .			x		o	
Spathognathodus sp. indet. .	x	x		x		x

The conodont work in the Chudleigh area illustrates that the conodont chronology employed in Germany can be applied with confidence in south-west England. In the 55 samples taken from the area, only two anomalies were encountered. 1) In the Winstow Cottages section, samples 4, 5 and 6 yielded a good fauna of the lower/middle *styriaca* zone (to IV-V), together with the species *Palmatolepis gracilis sigmoidalis* Ziegler. In Germany this conodont begins in the lower *costatus* zone, the conodont zone above.

2) Sample 7, of the *rhomboidea/quadrantinodosa* zone from Kiln Wood road section contained specimens of *Palmatolepis tenuipunctata* Sannemann. This species is normally found in the zone below (*crepida* zone), but could have been reworked into these beds of younger age.

3. Formation of Calcareous Nodules and Flaser Limestones

Calcareous nodule formation is best explained by diagenetic migration of carbonate very soon after deposition. In the reduction zone, a low pH (less than 7.8) causes the solution of the microcrystalline calcite which is disseminated throughout the muddy sediment. The dissolved carbonate is carried upwards by pore fluids into the oxidation zone, where the higher pH causes the calcium carbonate to be precipitated (Gründel and Rösler 1963). Field evidence shows that the size of the nodules and the distance apart of nodule bands both increase with decreasing carbonate content. From this and the process of formation outlined above, it follows that the size of the nodules is proportional to the rate of sedimentation, and that the distance apart of nodule bands is proportional to the amount of carbonate present in the original sediment. More continuous limestone bands and flaser limestones, representing periods of high carbonate production, are primary limestones which have been modified by early or late diagenesis, and/or tectonics. The shale streaks in flaser limestones can be formed in three ways (a) compaction, where the shaly material is injected into the surrounding, partly lithified limestone, (b) pressure solution (diagenetic or tectonic) where the flasers are residual seams, and (c) flasers produced by cleavage.

Trace fossils in carbonate nodules have been recently discovered by one of the authors (M.E.T.) in the Harz Mountains, in sediments of similar lithology and age to the ones described above. This find substantiates a very early diagenetic origin for the nodules. The trace fossils, a simple worm-burrow type of 1-3 mm diameter, occur at all angles to the bedding, but are mainly perpendicular to it. The bioturbation is rarely observed in the shales.

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CRINOIDAL TURBIDITES FROM THE DEVONIAN OF CORNWALL AND THEIR PALAEOGEOGRAPHIC SIGNIFICANCE

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SUMMARY

Devonian limestone turbidites from North Cornwall, composed mainly of crinoidal debris, are considered to have been derived from an oceanic rise or schwelle. Large-scale cross-bedding (representing division C) occurs in some limestone bands. The turbidites lack a pelitic division and there is usually a sharp contact with the shales above. Typical flutes are rare but broad grooves and channels are present on the soles.

INTRODUCTION

A series of turbidite units of crinoidal limestone has been recognized in the Devonian of the Padstow region, North Cornwall (Fig.1). The turbidite origin has been suggested by M. R. House (in GOLDRING et al, 1968). The structure of this area has been elucidated by G. A. Gauss (unpublished D. Phil. thesis, University of Oxford, 1968). The area was mapped by the Geological Survey (REID et al., 1910), and Fox (1903, 1905, 1906) and CRICK (1905, 1906) made contributions to the palaeontology. HOUSE (1956, 1960, 1963) has made more detailed stratigraphical and palaeontological observations on the area, involving goniatite studies. The Devonian rocks at Padstow occur in two structural blocks, and the successions in each block are significantly different, giving eastern and western successions. The western block is a recumbent syncline, with the axial plane dipping gently to the south and this is overthrust from the north, by an inverted northerly dipping block (GAUSS, 1968).

Localities are referred to by National Grid References from sheet 185 (1:63360).

GEOLOGICAL SETTING

The limestone turbidites with interstratified black shales have been named the Marble Cliff Beds (HOUSE, 1960). They succeed the Trevose Slates, a thick

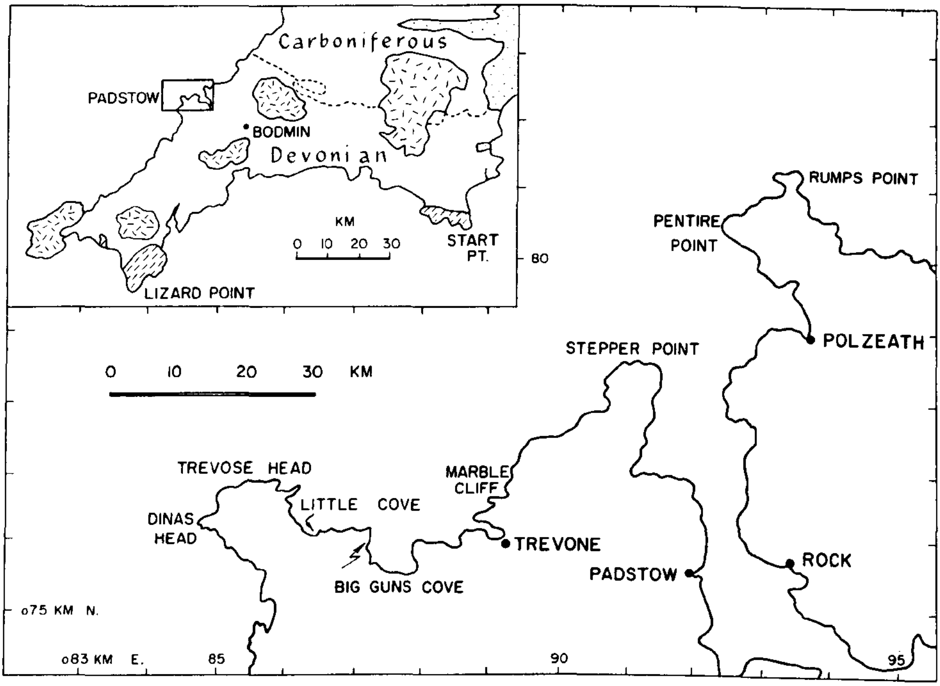

Fig. 1. Locality map of the Padstow District.

sequence of laminated dark grey shales with many light graded bands. The latter are probably of tuffaceous origin, and are regarded as distal turbidites. The Trevose Slates yield mainly pelagic fossils (goniatites, orthocones, and *Styliolina*, HOUSE and SELWOOD, 1964). The lithology indicates steady, vertical sedimentation in a quiet deep-water basin. Following the Marble Cliff Beds, another series of banded shales is developed within the western succession (the Longcarrow Cove Beds of GAUSS, 1968), which contain graded tuff bands and agglomerates partly equivalent to the Pentire Pillow Lavas of the eastern succession.

The Marble Cliff Beds thus make a very marked incursion into what is otherwise a fine-grained succession.

The type locality for the limestone turbidites is Marble Cliff, north of Trevone (National Grid Reference SW 891764), in the western succession, where the beds are inverted and form cliffs 50 m high. Further to the west, the limestones can be found at Big Guns Cove (SW 872761), near Cataclews Point; Little Cove (SW 865760), at the east end of Mother Ivey's Bay; and Dinas Head (SW 848763), near Trevose Lighthouse. On the eastern side of the Camel Estuary, the limestones are poorly exposed on the foreshore at Rock (SW 934756). In the eastern succession, turbidite limestones are not so well developed but the stratigraphic equivalents of the Marble Cliff limestones crop out at Rumps Point (SW 931812), north of Polzeath.

The Marble Cliff Beds at Trevone have been dated by Dr. W. T. Kirchgasser as Upper Givetian to Lower Frasnian on the basis of the conodont fauna. Conodonts of the *Polygnathus varca* Zone (Upper Givetian) occur in the Rumps Point limestones. The coral *Phillipsastrea hennahi hennahi* (LONSDALE), a characteristic Upper Givetian form, also occurs in the turbidites. A goniatite band occurs just below the turbidite sequence (Pentonwarra Point, SW 890760) and belongs (HOUSE, 1963) to the Upper Givetian zone of *Maenioceras terebratum*. The first dateable horizon above the turbidites is just above the top of the Longcarrow Cove Beds, where *Manticoceras cordatum* Zone goniatites occur (Middle Frasnian).

The shales interstratified with the limestone bands are black pyritiferous argillites containing the trace fossil *Chondrites*, commonly found on the soles of the turbidites. The shales, formed by relatively slow sedimentation, are unfossiliferous apart from very small brachiopods and goniatites (2–3 mm in diameter) found only at Rock.

THE LIMESTONE TURBIDITES

Some 80 limestone bands form Marble Cliff (Fig.2) where they vary in thickness from a few centimetres to a metre. Individual bands can be traced the whole length of the outcrop, a distance of about 200 m. The regularity of the beds is striking. The shale divisions between many bands are of similar thickness,

Fig.2. Limestone turbidites interstratified with black shales, Marble Cliff, near Trevone, North Cornwall. (SW 891764) The succession is inverted.

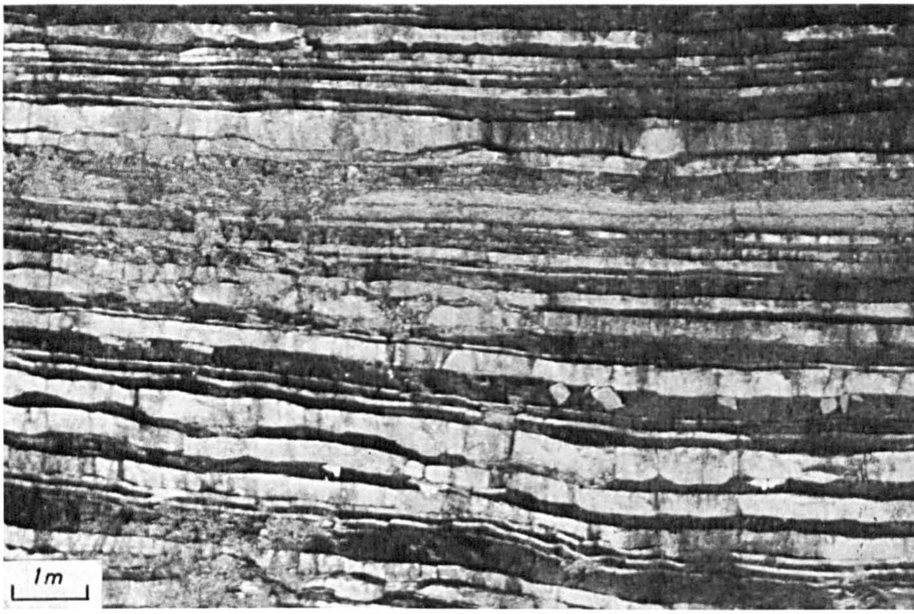

Fig.3. Large scour structures on soles of limestone bands at Marble Cliff. Shown in normal succession.

indicating periodic influxes of carbonate sediment. Continuous exposure is not available at the other localities, due to faulting and folding. At Rumps Point, however, the decreased size and number of the limestone bands is considered to be due to the more distal nature of the turbidites.

BOUMA (1962) has described the succession of structures found in sandstone turbidites. The structures developed in a turbidite must depend in part on the state and size of the grains present in the depositing current. The wholly calcareous Padstow turbidites display a sequence of structures not generally associated with terrigenous turbidites.

About 60% of the limestone units have a perfectly planar sole, but some of the thicker bands have irregular bases with scour marks (Fig.3). At Marble Cliff common scour structures are channels, box-shaped or rounded in cross-section, up to three metres wide and which cut down up to 30 cm, and may cut the underlying limestone band. Undersurfaces are poorly exposed but the channels can be followed for 2–3 m. The channels are filled with coarse-grained sediment, which is often horizontally laminated. Clearly no post-depositional deformation, such as load casting has taken place. It is not always possible to tell whether the channels (gouge-channels of KUENEN, 1957) are large groove or flute casts, but some are certainly the latter. Longitudinal groove casts are seen at a few localities. They are generally up to 10 cm wide and semi-circular in cross-section. All channels and groove casts have a southwest-northeast trend (see Fig.6). Normal flute casts are rare but tool marks (probably saltation marks) are occasionally seen.

The most frequent internal structure is a horizontal or nearly horizontal lamination. Limestones thicker than 30 cm invariably show a lower and an upper laminated division (representing divisions B and D of the Bouma sequence), with a central cross-laminated part (division C of Bouma) of similar thickness. This is characteristic of limestone bands which grade from a maximum of coarse sand grade (0.75 mm) to silt grade. The lamination is due to alternating coarse light coloured and fine darker laminae, each about 2 mm thick. In a few bands, where the upper horizontal lamination is absent, the top-most few centimetres may be reworked into current ripple, with local contorted lamination. The grain-size distribution was determined from thin-section for a 43 cm thick, well-laminated band from Marble Cliff using the sorting index of FOLK and WARD (1957). The sorting of the coarse laminae ($\sigma\phi = 0.58$) is greater than that of the fine laminae ($\sigma\phi = 0.89$). Fine laminae contain higher percentages of silt grade carbonate. The central cross-laminated division, up to 15 cm thick, is not obvious in hand specimen and can only be seen by etching and staining using the techniques of HAMBLIN (1962). The inclination of the lamination may reach 20°, and is again due to alternating coarse and fine bands. However, sorting into the two fractions is not as good as in the horizontally laminated divisions and individual cross-laminae are up to 7 mm thick. The grain size of this dune phase is of fine-sand grade (2.25–2.5ϕ)

Although most limestone bands do not show a separate graded division in the lower part, there is a continuous decrease in grain size up through the bands. Sorting decreases upwards from an average of $\sigma\phi = 0.70$ at the base, to $\sigma\phi = 1.36$ at the top. A 25-cm band at Big Guns Cove shows three graded units before passing into horizontal lamination. This "multiple grading" (KSIAZKIEWICZ, 1954) is thought to represent deposition from an immature density current in which sorting of grains had not yet been achieved (WALKER, 1965). In the few beds which do show a separate basal coarse grained division, fragments of corals and stromatoporoids, shells and reworked pieces of limestone up to several centimetres in diameter occur, and at one locality (see below) blocks of coral up to 40 cm across are found.

Limestone bands showing the complete Bouma turbidite sequence are uncommon at Trevone. However, at Rumps Point, one band does show an ABCD sequence (BOUMA, 1962), though there is a large grain size difference between the graded division A and the horizontally laminated division B (Fig.4). At the top of the graded division many grains of 1–2 mm diameter are present, in a coarse sand grade matrix. The B division contains mainly carbonate of fine and medium sand grade. Crude horizontal and cross laminations occur in the coarse grained band. It seems possible that this very coarse lower part has been deposited from a traction current (WALKER, 1965). Convolute bedding occurs at the top, with sharp "anticlinal" structures and axial planes at right angles to the bedding separated by a broader trough region.

The pelitic division (E) is never present, probably due to the absence of mud

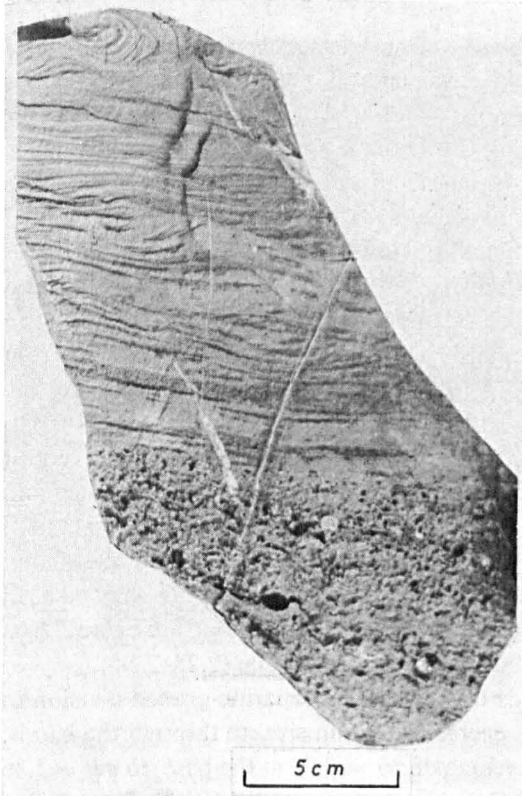

Fig.4. Limestone band from Rumps Point (SW 932812) near Polzeath.

and clay in the source area of the carbonate. The top of each limestone is either a distinct break or a gradational passage into shales within 10 mm. There is no evidence to suggest that the sharp break is due to diagenetic migration of carbonate accentuating the limestone/shale interface. The various features described in the limestones suggest that they are medial-proximal turbidites, deposited by currents still able to transport large blocks of material and to cut deep erosion channels.

COMPOSITION OF LIMESTONES

The limestones have been derived entirely from an area of carbonate sedimentation (Fig.5). Terrigenous quartz makes up less than 1% of the total composition, and some is certainly authigenic (quartz is often seen filling the cavity between ostracod valves). Crinoid fragments form the largest component of the limestones (up to 78%), but some have been converted to micrite by diagenesis. This masks the original crinoid content, which was probably a little higher, and also the original micrite percentage. The skeletal carbonate is well packed and so sparite cement is generally present only in subordinate amounts. However, it does

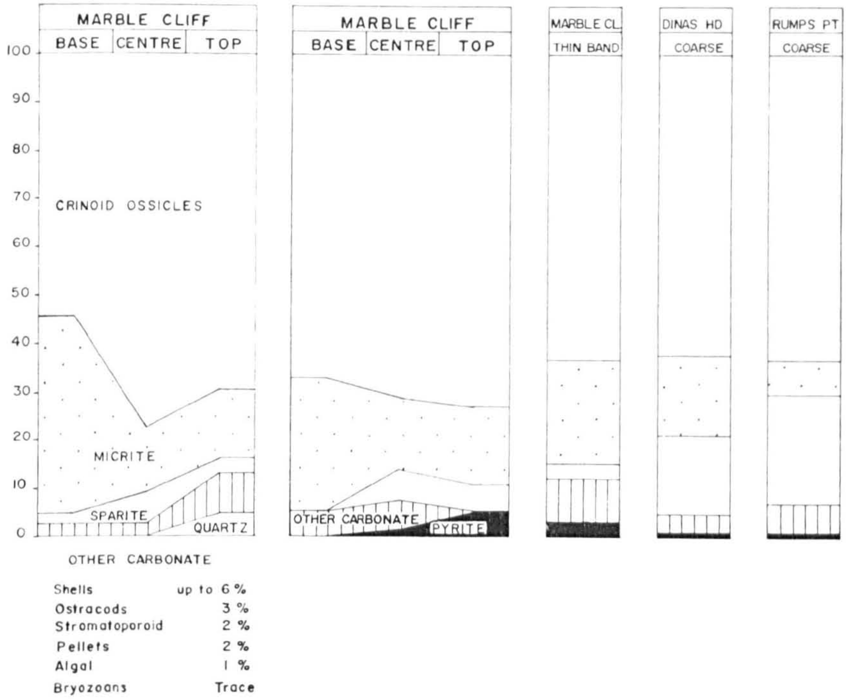

Fig.5. Petrographic modal analysis of five limestone bands.

form a significant amount in very coarse bands (around 20%). Sparite is often present as a syntaxial growth.

A coarse 40-cm limestone band exposed at Dinas Head contains some large allochthonous blocks of the coral *Phillipsastrea hennahi hennahi*. These reach 45 × 40 × 10 cm, the average size being about 12 cm square by 5 cm. The blocks are concentrated at the base of the band indicating that the corals were transported by the turbidity current and did not float into place. WELLS (1967) suggested that colonial rugose corals can be necroplotic and travel hundreds of miles before being deposited in foreign sediments. Associated with *Phillipsastrea* are fragments of *Stromatopora concentrica* up to 10 cm across, *Favosites* sp., and solitary corals. Minor constituents are small ostracods and occasional pieces of calcareous algae. Indeterminate fragments of thick shells are common, up to 3 cm long.

Insoluble residues (using 10% acetic acid) contain up to 60% pyrite which is disseminated throughout the limestone bands. Pyrite (probably epigenetic) is concentrated on the upper and lower surfaces of the turbidites. Pyritized centres of goniatites less than 1 mm in diameter are found in the residue, along with pyritized fragments of orthocones, gastropods, ostracods and styliolinids. Well preserved conodonts are common. Much of the original calcite and dolomite has been converted to ferroan carbonates, and most of the ferroan dolomite is restricted to the micritic matrix. Many of the crinoid fragments, particularly the smaller ones, have been converted to ferroan calcite. Ferroan calcite also forms the sparite cement and millimetre veins in the limestones.

PALAEOGEOGRAPHIC SIGNIFICANCE

The limestones are clearly the deposits of turbidity currents which brought pulses of carbonate detritus into a deep-water basin. This basin was a bathyal environment where argillaceous deposition prevailed. The carbonate came from a sub-littoral environment supporting crinoids and a reef community. *Phillipsastrea* indicates shallow seas. The crinoids probably lived on the slopes of the shallow water area and the corals and stromatoporoids higher up where light was available. Stick bryozoa and algae probably inhabited the fore-reef environment with the crinoids.

Although only a few current directions were measurable at Trevone, owing to the poor development of suitable sole marks and cross-bedding, the evidence does clearly indicate a south-westerly provenance (Fig.6). Immediately to the south older rocks are exposed, and although the stratigraphy of the Devonian of

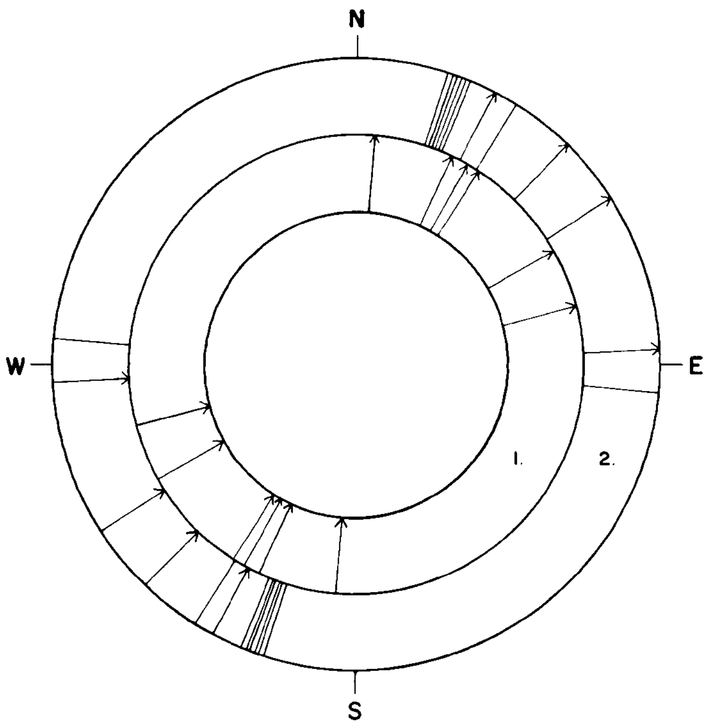

Fig.6. Current data from limestone bands at Marble Cliff and Big Guns Cove. *1.* Measurements from cross-bedding. *2.* Measurements from gouge channels, groove casts and tool marks.

South Cornwall is still little understood (GOLDRING et al., 1968) there is no suggestion of a nearby shore-line. The virtual absence of terrigenous material in the limestones, the predominance of crinoids, and the near-proximal origin of the turbidity currents, show that the turbidites must have come from a positive rise or schwelle in the ocean, uninfluenced by terrigenous erosion. There is no evidence that the swell rose above sea-level.

Schwellen are well-known in the Devonian of central Europe and are considered to be present in the Famennian of southwest England because of local developments of schwelle-type sediments. The crinoidal turbidites indirectly indicate their presence in the Middle Devonian of southwest England.

COMPARISONS

The limestone turbidites at Padstow show a number of atypical characters, related to their peculiar composition. Admittedly, large erosional channels have been recorded on the soles of sandstone sequences, but groove and flute casts, and various tool marks, are more usual. Lamination due to coarse and fine bands features the Padstow calcarenites, whereas in sandstone turbidites it is more often due to small-scale bedding. The dune phase is apparently absent in sandstone turbidites, but its presence here could be due to the different physical properties of the grains. The pelitic interval (*E* division of Bouma) is absent from the Marble Cliff Beds, but in other turbidites is usually present and can form up to 60% of the bed. A distinct upper boundary is commonly found at Padstow; in normal turbidite sequences there is a gradational passage into the background shale.

The Cornish limestone turbidites here described show many similarities with the allodapische Kalk of MEISCHNER (1964). Both derive from a reef environment, the Padstow turbidites being mainly crinoidal and the allodapic limestones more varied faunistically, including pelagic forms. However, the German calcareous turbidites, unlike those in the Marble Cliff Beds, often show a "prephase" unit of fine-grained marl below the coarsest part of the limestone. This is in fact calcified background sediment, formed by diagenetic migration of carbonate within the sediment.

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