SEDIMENTOLOGICAL STUDIES OF THE ELLER BECK BED AND THE LOWER DELTAIC SERIES IN NORTH-EAST YORKSHIRE

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ABSTRACT

In this study of the Lower Deltaic Series emphasis has been placed on the field relationships and characteristics of the various lithofacies types and their correlation with the observed facies of the present day deltaic environment. Deposits representing channel, levee, and interdistributary marsh environments have been recognised. The delta is though to have advanced very rapidly across a surface of Dogger sediments which were near or even above sea level. Three cycles of channelling are believed to be related to phases of uplift in the hinterland. Analysis of directional data indicates transport from the north and west in the west and from the north and east in the east. This pattern probably reflects the local distribution of main distributaries to the east and west, while the ultimate source is believed to have been a Scandinavian landmass.

The Eller Beck Bed transgression spread from the south-east, along an E.S.E. - W.N.W. axis of maximum subsidence. The fauna indicates that fully marine conditions prevailed during deposition of the basal ironstone and of the lower part of the shale. Upward passage from shale into sandstone appears to represent gradual readvance of the delta, culminating in the colonisation of the top of the sandstone by plants. Subsidence along the trough axis appears to have produced a clastic trap, so that to the south much of the sandstone is represented by shale. Upstanding sand-banks developed in the shale unit provided local environments suitable for the formation of chamosite ooliths.

The ooliths from the Winter Gill ironstone indicate formation under very gentle current conditions.

The petrography of the sediments has been studied with emphasis on the iron-rich types. Grain-size analysis of sandstones has been carried out and the results interpreted in terms of the hydrodynamic environment of deposition.

A C K N O W L E D G E M E N T S

I would like to express my appreciation to Professor J. E. Hemingway for having introduced me to this research topic and for his subsequent help both in the field and in discussion. I also thank Professor T. S. Westoll for the use of the research facilities of the Department of Geology, University of Newcastle upon Tyne, and Professor H. B. Whittington for the use of the facilities of the Geology Department, University of Cambridge, during the preparation of this thesis. The research was carried out under the tenure of a N.E.R.C. grant. I am indebted to my wife for having typed this thesis.

PREFACE

The purpose of this work is three-fold:

1. To study the sediments of the Lower Deltaic Series and their distribution in order to reconstruct the delta in terms of present day delta facies;
2. To study the sediments of the Eller Beck Bed and their vertical and lateral distribution in order to relate them to the marine transgression and regression that this bed represents;
3. To consider both formations together in order to determine the factors which influenced the general pattern of sedimentation.

Much of the environmental reconstruction has been carried out as a result of field observation but laboratory techniques, particularly the study of thin sections, have been used to determine both the original constitution of the sediments and the effects of diagenesis. Following the three-fold division given above this thesis is written in three parts: Part I deals with the Lower Deltaic Series, Part II with the Eller Beck Bed, and Part III with those aspects of the study which involve both formations. A list of sections of the Eller Beck Bed and accounts of laboratory techniques used are given in the Appendices.

PART ONE

THE LOWER DELTAIC SERIES

INTRODUCTION

I STRATIGRAPHICAL NOMENCLATURE

The subdivision and particularly the nomenclature of the Middle Jurassic strata of north-east Yorkshire have undergone several changes in the history of the geological investigation of the area. The subdivision was originally two-fold, with an upper and a lower group of sandstone, shale, and coal separated by the Grey Limestone Beds. Subsequently a three-fold division was set up in which the old lower division was subdivided at the horizon of the Millepore Bed. In order to standardise the nomenclature of the non-marine beds Fox-Strangways (1880, p.3) introduced the terms Upper, Middle, and Lower Estuarine Series. A detailed account of the earlier nomenclature is given in Fox-Strangways (1892 pp. 197,217,252), and only a list is given below of the terminology relevant to the beds under consideration:

Lower Carboniferous Series - part (between the Dogger and Grey Limestone)	William Smith (see Williamson,1837 p.229)
Sandstone Shale and Coal - part (between the Dogger and Grey Limestone)	1822 Young and Bird p.109
Coal Formation - part (between the Dogger and Kellaways Rock)	1826 Sedgwick p.16
Lower Shale and Sandstone	1829 Phillips p.33
Lower Sandstone and Shale	1837 Williamson p.235
Lower Sandstone and Shales - part (between the Dogger and Grey Limestone)	1860 Wright p.30
The Lower Shale and Sandstone	1874 Hudleston p.306
Lower Estuarine Series	1880 Fox-Strangways p.3

Following the Survey Memoirs the terminology remained unchanged for some time, although it became clear that the beds were non-marine and had been deposited under deltaic rather than estuarine conditions. In 1949, however, Hemingway proposed that the Estuarine Series should be renamed the Deltaic Series. At the same time he proposed that the boundary between the lower and middle divisions should be drawn at the Eller Beck Bed and not the Millepore Bed since the former was traceable over most of the outcrop, and the latter very restricted in outcrop. The Middle Deltaic division was separated into an upper Gristhorpe Sub-series and a lower Sycarham Sub-series for areas in which the Millepore Bed or its equivalents were developed. Following this Sylvester-Bradley (1949) proposed locality names for every non-marine division suggesting, in addition to Hemingway's sub-series names, that the Upper Deltaic Series be termed the Scalby Beds and the Lower Deltaic Series the Hayburn Beds. No name was suggested for the Middle Deltaic Series where the Millepore Bed is absent. Bate (1959, p.154) uses a combination of nomenclatures but later (1967, p.118) uses Hemingway's classification in a slightly modified form in that the terms Lower and Upper Middle Deltaic Series are used in place of the Sycarham and Gristhorpe Beds respectively.

The term Lower Deltaic Series is used here as it is in current usage. It is, however, appreciated that geographical location names would be required if the nomenclature should be revised to conform to the code of the American Commission on Stratigraphic Nomenclature (Ashley et al., 1933). With specific reference to the Lower Deltaic Series, however, the term Hayburn Beds is not considered to be suitable. Sylvester-Bradley's choice of name is based on the occurrence of a plant-bed at Hayburn Wyke, where as, in the code referred to above, completeness of exposure should be the most important feature of the standard locality. On this basis the term Hawsker Beds would be more suitable since the whole of the Lower Deltaic Series including the upper and lower boundaries are clearly displayed in the cliffs east of Hawsker.

II HISTORY OF RESEARCH

A. FIELD RELATIONSHIPS

All of the early authors describe the unusual variability of the non-marine strata but there is little attempt at any specific description. The erosional nature of the base of the thick sandstone lenses was first appreciated by Simpson (1868, p.39), but no detailed description of the sandstone bodies appeared until the papers by Hepworth (1923) and Black (1928). Hepworth describes contortion of strata beneath a sandstone channel-fill which he ascribes to earthquake shocks while Black, describing somewhat similar features, considers them to be due to bank collapse and compactional effects. Black was also able to trace along the foreshore the meandering form of some channels in the Upper Deltaic Series (1929, fig. 2). Erosion at the base of the Lower Deltaic beds affecting the Dogger and the top of the Lias had been mentioned earlier (Fox-Strangways, 1892; pp.154-163; Rastall, 1905; pp.450, 452; Tonks, 1923, p.30).

The preservation of Equisetum in vertical attitude was noted at an early date but Murchison (1832) was first to appreciate that such beds represented growth in place. The presence of abundant rootlets is frequently referred to, but of particular interest are the supposed tree roots preserved in the Dogger (Kendall and Wroot, 1924: p. 402). Preservation of dinosaur footprints in mud surfaces overlain by thin sandstones are recorded in Kendall and Wroot (pp 349-350) from both the Lower and Upper Deltaic Series, and similar beds are said to exhibit sun-cracking.

Although the sandstones are frequently referred to as being false-bedded or current-bedded there is no detailed description of the internal structures of these sediments. Cross-bedding and ripple-mark orientation was, however, studied by Sorby (1852).

B. PALAEONTOLOGY

Practically every publication on the Middle Jurassic of this region describes the abundant flora. Recent work, with accurate localities and descriptions, which includes the Lower Deltaic flora, is that of Lane and Saunders (1909), Thomas (1913a, 1913b, 1915), and Harris (1942-53, 1952, 1953). Harris (1953) discusses the palaeoecology of the flora.

The fauna of the non-marine beds is very restricted. Freshwater bivalves have been recorded from the Upper Deltaic Series, and these often occur in growth position (Leckenby 1864, 75; Hudleston 1874, 318-319). A similar occurrence is recorded by Kendall and Wroot (1924. p.351) from the 'footprint' bed at Saltwick. The position of this bed was discovered by Kendall, but fallen blocks with dinosaur footprint casts on the lower surface were discovered by Brodrick (1909). Footprint beds are also known from the Upper Deltaic Series

and Wroot
(Kendall/, 1924; pp.349-350). No ostracods are known from the non-marine beds.

C. PETROGRAPHY

Little work has been published on the petrography of the non-marine beds, and only the heavy mineral assemblages have been studied in any detail. Rastall (1932) commented on the small number of mineral species present and on the absence of metamorphic minerals. Smithson (1934, 1937, 1939, 1941) describes the heavy minerals and their distribution. He also describes cartographic methods of representation. A summary of his results (1942) includes a discussion of the possible source of the sediments and the effects of post-depositional solution on the heavy mineral assemblages.

D. SOURCE OF DETRITUS AND PALAEOGEOGRAPHY

A study of the drift-bedding and ripple-marks led Sorby (1852) to conclude that transport had been from the north-north-west. This was accepted by Fox-Strangways who also pointed out that the non-marine strata were coarsest and most arenaceous in the north-west. Kendal and Woot (1924) considered that a far northerly source was probable, but Black (1934) thought that the delta encroached from the east with minor amounts of material being derived from the west. Smithson (1942) considered that most of the sediment came from the northwest with minor amounts from the east and west.

Conclusions as to the palaeogeography are first found in

E. ENVIRONMENT OF DEPOSITION

The non-marine beds have been variously described as marine, lacustrine, estuarine, and deltaic in origin. Early workers, (Young and Bird 1822, p.296) considered that alternation of fresh-water and marine beds was impossible and it was Murchison (1832) who first clearly states that the beds were deposited in freshwater with subaerial phases during which plant colonisation took place. Murchison also describes the land surfaces as being buried under the accumulation of an estuary. Wright (1860; p.30) describes beds containing Unio as being lacustine. The first clear representation of conditions is that of Simpson (1868; p.39) in which he compares the deposits to those of a delta with sand-filled distributary channels. Phillips (1875; p.110) describes the beds as river sediments but still refers to them as estuarine. Fox-Strangways (1892; p.393) having introduced the term Estuarine Series, describes the beds as forming a land surface with much standing water and with meandering river courses. It appears that the term Estuarine Series was used not so much through ignorance of the conditions of accumulation as through misuse of the word 'estuarine'. Even Kendall and Wroot (1924) although clearly stating the beds to be the deposits of a delta still frequently use the word 'estuarine' in a descriptive sense (e.g. p.309). Hepworth (1923) and Black (1928) stressed the deltaic origin of these beds and finally in 1949 Hemingway proposed his revised nomenclature to clarify the situation.

III METHOD OF PRESENT RESEARCH

A. FIELDWORK

The area over which the Lower Deltaic Series has been studied is shown in figure 1.1. The degree of exposure is very variable and the extensive exposures on the coast permit an approach to research that is entirely different to that which is possible for the inland exposures. In particular the coastal exposures consist of both fine-grained and coarse-grained sediments whereas inland exposures are almost always confined to the more resistant sandstone bodies.

The full thickness of the Lower Deltaic Series is visible for much of the Whitby to Robin Hoods Bay cliff section while to the north and south the lower and upper parts respectively are best exposed. Natural inland exposures other than sandstone bluffs are found in some of the dale heads, especially those of Great Fryupdale and Rosedale. Several alum quarries along the northern escarpment provide good exposures of the lowest beds and their relationship to the underlying marine strata, but the only quarries of any importance at a higher level are the building-stone quarries of Aislaby.

The method of study has been to observe general features, such as current-bedding directions and channel distribution, over the whole area and to confine detailed petrographical studies to the beds exposed on the coast. The few comprehensive inland sections indicate that the facies represented differ in no way from those of the coast, and in the coastal exposures facies change is so rapid that there is little doubt that all lithologies and sedimentary association types

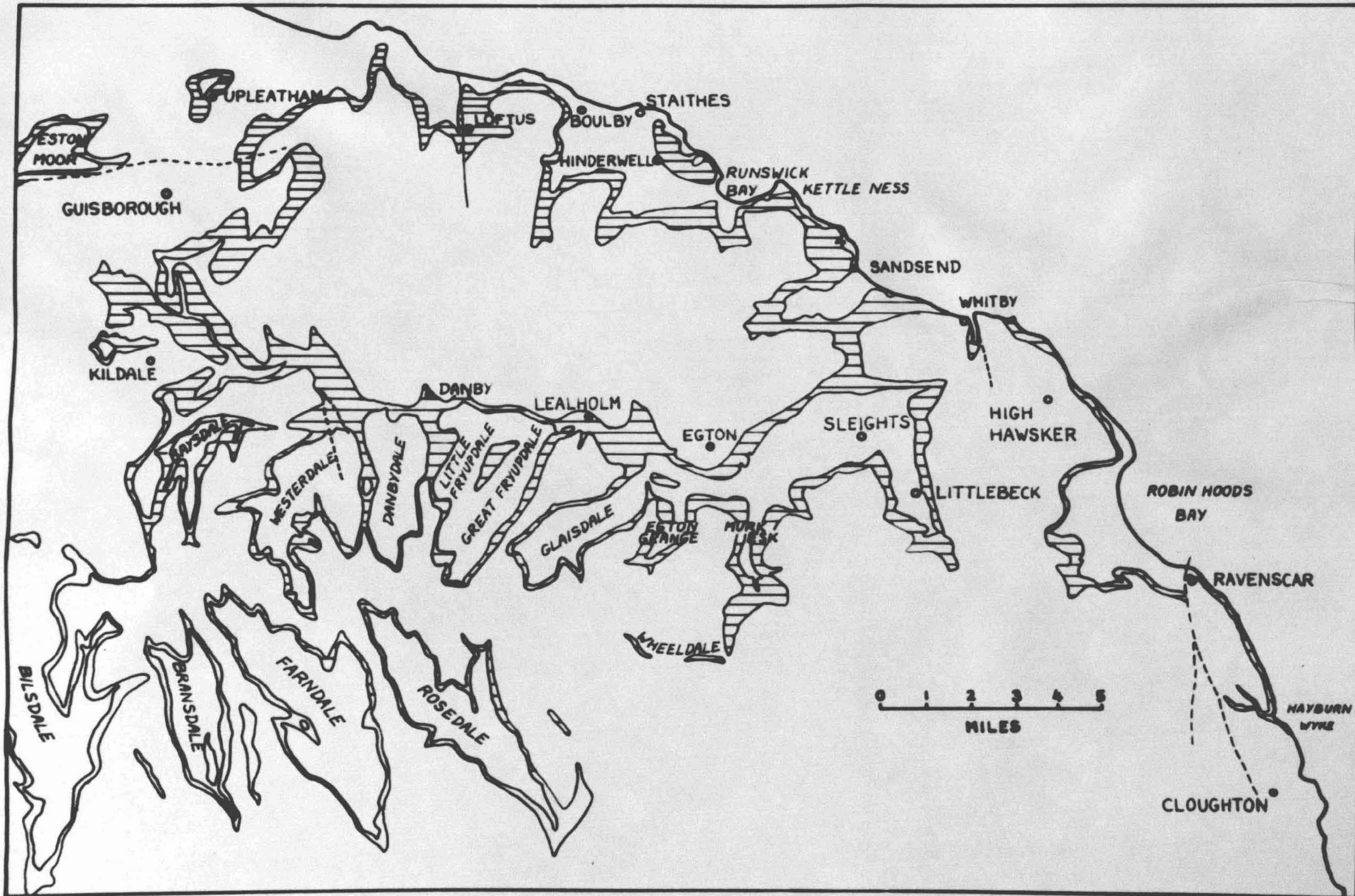


FIG. 1.1 OUTCROP OF THE LOWER DELTAIC SERIES; AREA OF STUDY SHADED

are represented. The beds are so variable that a stratigraphical sequence within the Lower Deltaic Series cannot usually be detected, and the nature of these beds is represented here by illustrating examples of all the principal lithofacies types and then commenting on any trends that might be present in their distribution.

B. LABORATORY WORK

The beds have been studied primarily with regard to their field relationships but the sandstones and ironstones have been studied under the microscope. Mechanical analysis of sandstones was also carried out, and a brief study made of the heavy mineral content.

CHAPTER I LITHOFACIES TYPES

I DISTRIBUTARY CHANNEL FACIES

A. SMALL-SCALE ELONGATE SAND BODIES

These include units ranging from a few inches to about twenty feet in width. There appears to be a fairly distinct natural division between these and the more common larger-scale sandstone bodies. On the basis of cross-sectional form groups may be distinguished.

1. SYMMETRICAL UNITS WITH NON-EROSIONAL BASE

These have been observed on a very small scale in a thin coaly horizon lying between two clay siltstone units (fig. 1.2). The sand occurs in tiny lenses which are all well separated and which in part lie within the coal seam. No roots are present below the coal and it is clearly of drifted type. The biconvex cross-section of the sand bodies has resulted from loading at the base and by compaction of the surrounding sediment. On a larger scale truly symmetrical sand bodies are not developed. This is due to channel migration and to the development of marginal facies which pass into the surrounding fine-grained sediments and which are always developed unequally on either side of a channel.

2. SYMMETRICAL UNITS WITH EROSIONAL BASE

The sandstone unit shown in figure 1.3 is of this type. Erosion

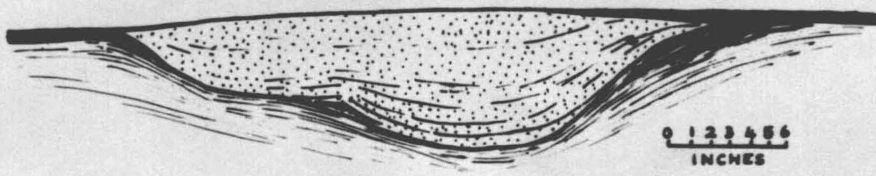


FIG. 1.2 SMALL-SCALE NON-EROSIONAL CHANNEL WITH LOADING AT BASE

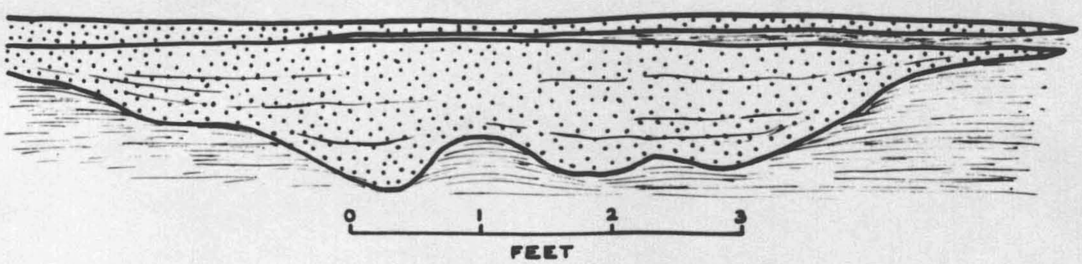


FIG. 1.3 SMALL-SCALE SYMMETRICAL EROSIONAL CHANNEL SANDSTONE

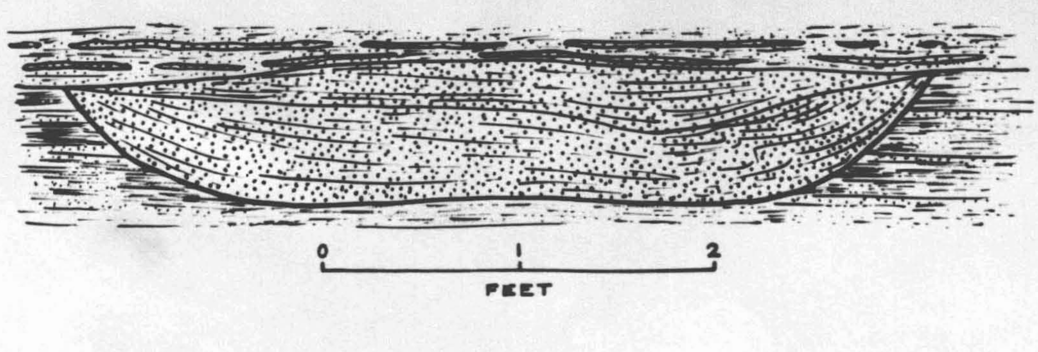


FIG. 1.4 SYMMETRICAL EROSIONAL CHANNEL

of the base is exaggerated by loading. Sandstone-filled channels occurring within sandy horizons are generally of this type (fig.1.4). The lack of marginal sedimentation allows the sandstone to retain the symmetry of the original erosion channel.

3. ASYMMETRICAL TRANSGRESSIVE UNITS

These generally occur at one end of a series of horizontally bedded sandstone which passes in the opposite direction into silts. They are characterised by an initial flat base which then transgresses upwards and becomes of an erosional nature. The top of the unit roughly parallels the base, producing a unit which in section is elongate at an angle to the horizontally bedded strata (figs.1.5,1.6, 1.7).

4. ASYMMETRICAL NON-TRANSGRESSIVE UNITS

These are similarly associated with sandy horizons; at one side the passage into more silty beds occurs while on the other a small amount of erosion may occur (fig.1.8,1.9). Erosion of the base is not pronounced, but loading may occur (fig.1.8).

5. INTERNAL STRUCTURE

The sandstone in all types may be massive (fig.1.9). More commonly horizontal bedding or cross-bedding is developed. Cross-bedding is usually of a distinctive low-angle type in which the foreset dip is perpendicular to the sand body axis. The bedding dips towards the erosional side of migrating channels and shows both

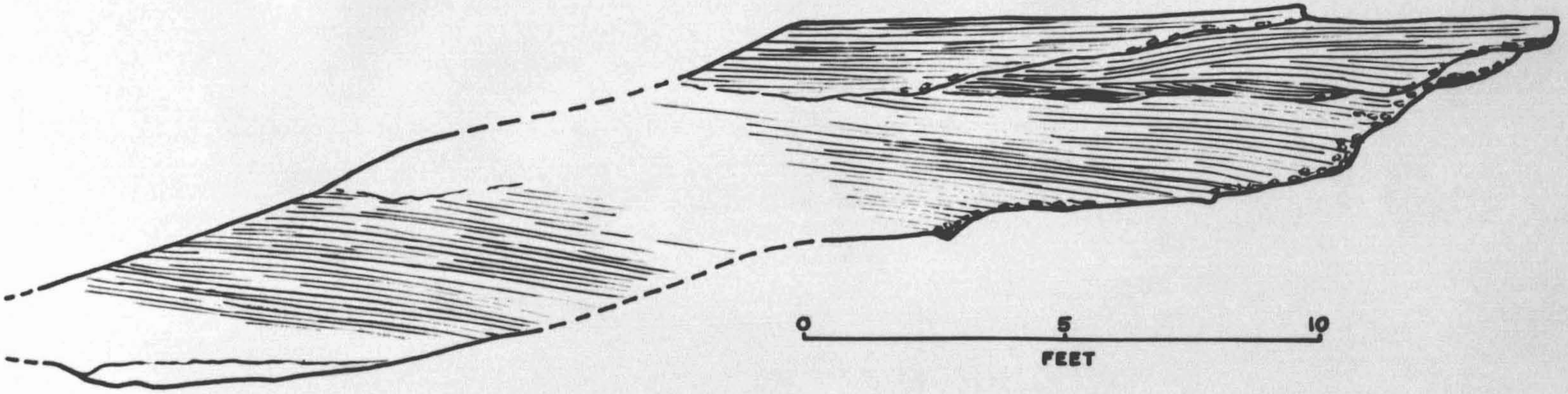


FIG. 1.5 SMALL-SCALE SIMPLE ASYMMETRICAL TRANSGRESSIVE CHANNEL SHOWING TRANSVERSE CROSS-BEDDING (AISLABY QUARRY)

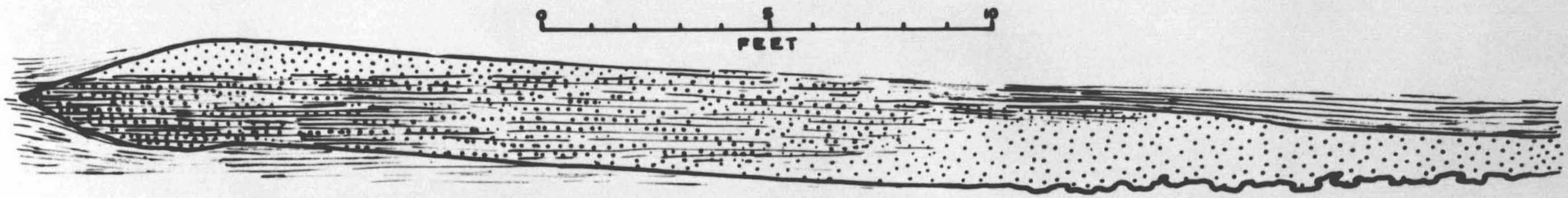


FIG. 1.6 ASYMMETRICAL TRANSGRESSIVE SANDSTONE

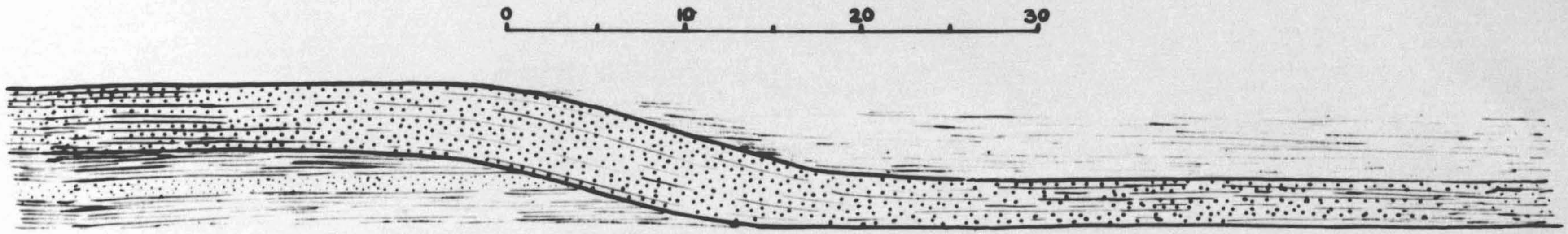


FIG. 1.7 TRANSGRESSIVE EROSIONAL SANDSTONE

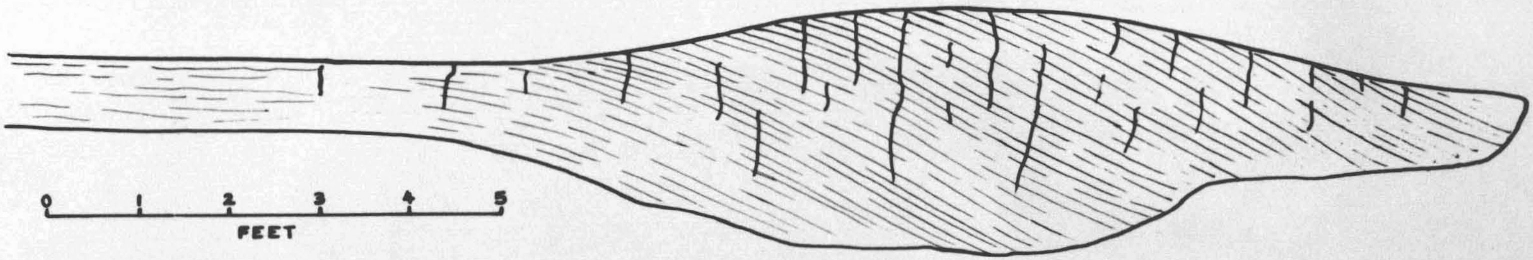


FIG. 1.8 ASYMMETRICAL NON-TRANSGRESSIVE SANDSTONE, TRANSVERSELY CROSS-BEDDED

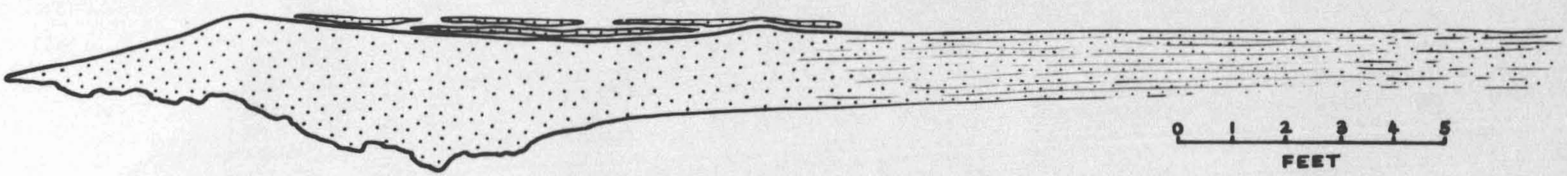


FIG. 1.9 ASYMMETRICAL NON-TRANSGRESSIVE SANDSTONE WITH LOADING AT THE BASE

horizontal and vertical accretion when the channel itself shows vertical migration. The laminae are laterally extensive and reflect only minor differences in grain-size so that the sandstone weathers out as a single mass as in figure 1.5. A single cross-bedded unit may constitute the whole mass, but small units showing reversal of dip are common in the upper part. In one case small-scale cross-bedding was developed within the individual beds of the low-angle unit such that a section parallel to the sand body axis showed a series of climbing ripples. This proved that current flow was perpendicular to foreset dip of the low-angle beds. Normal down-stream cross-bedding, which is described later in relation to large-scale sand bodies, is rare.

All structural types may suffer deformation by loading at the base of the sandstone. The scale of the loading is usually small and separation of loaded masses has not been observed. The loading occurs only on non-erosional surfaces.

6. MODE OF FORMATION OF SMALL SAND BODIES

It is clear that in all cases the sandstone has been deposited in small distributaries. In some cases erosion of the underlying strata has not occurred and the sand is built up from a level base. In such cases finer sediments are deposited on one or both flanks of the main sandstone depending probably on whether the distributary possessed a meandering or a straight course respectively. The predominance of asymmetrical sedimentation suggests that meandering of streams was common. In some cases after perhaps an initial flood which delineated the margin of sediment accumulation, the distributary adopted a fixed course which was maintained until choked with sediment. In other cases a somewhat longer life is indicated, during which a certain amount of lateral incision took place, accompanied by deposition on the inner meander margins which is represented by the low-angle cross-bedding.

The unusual sand-body shown in figure 1.7 may have had a different origin. Sand deposition appears to have taken place in relatively quiet current conditions after the initial phase of erosion since the bedding planes parallel the erosional surface. The sand may have been introduced by currents flowing laterally into the erosion hollow.

7. SMALL COMPOSITE ELONGATE SAND BODIES

These differ from the above only in that the distributaries were sufficiently long-lived as to show considerable vertical migration. In figure 1.10 a compact association of channels of similar type is illustrated while in figure 1.11 a more variable group is shown which indicates less stable current conditions. The internal structures are similar to those associated with the simple sandstone bodies. The scarcity of small-scale composite sandstone bodies reflects the ephemeral nature of the small streams from which the small-scale units were deposited.

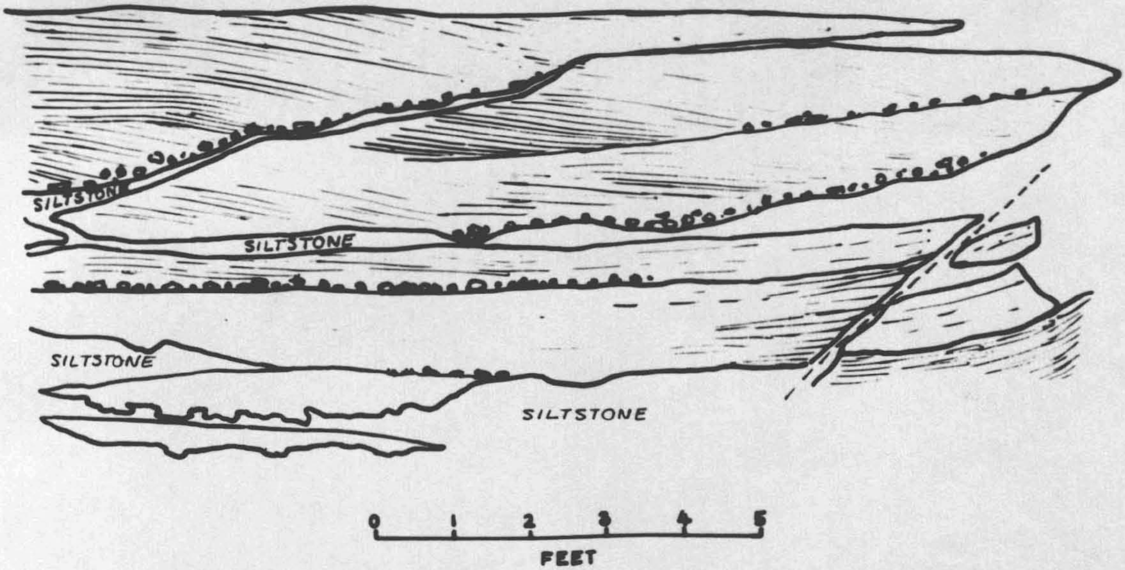


FIG. 1.10 SMALL-SCALE COMPOSITE SANDSTONE BODY WITH TRANSVERSE CROSS-BEDDING IN PART

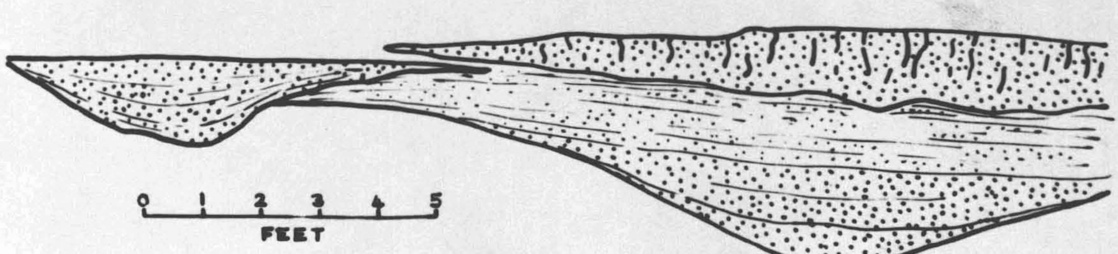


FIG. 1.11 SMALL-SCALE COMPOSITE SANDSTONE BODY

B. LARGE-SCALE ELONGATE SAND BODIES

1. SAND BODY FORM

The small channel-fill sandstones have been described in detail because they appear to represent in miniature form the nature of the larger sandstone bodies. However, where large distributaries have developed short lived phases were apparently rare and simple channels therefore are not seen. The composite nature of these channels is variable; in some cases a unidirectional lateral transgression is involved, giving a compact sandstone body, while in other cases the complexity reflects a more varied history.

The sandstone shown in figure 1.12 is of a relatively simple character. Strong erosional features occur at the south-eastern end and to a lesser extent at the base. Partings are present at the north-western end and these all dip to the south, indicating that accummulation took place in that direction, presumably concommitant with erosion along the southern margin. At the top the beds become more thinly bedded and show small-scale cross-bedding; silty partings are present. To the south thin-bedded sandstones extend beyond the channel itself and form a thin horisontally-bedded sandy horizon which extends into the interdistributary deposits.

A more complex series is shown in figure 1.13 which shows part of a channel horizon which extends for over 1000 yards along the Hawsker Cliffs. Sandstone distribution may be obtained in three dimensions in Aislaby quarries (fig.1.14) in which the principal channel direction is from east to west.

Some channel sandstones show a composite nature on a small

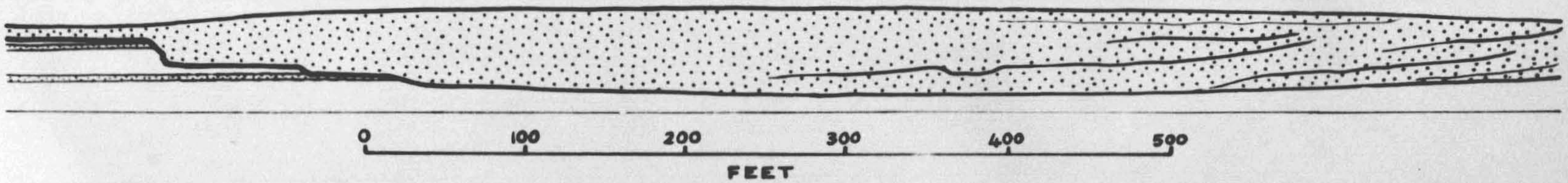


FIG. 1.12 LARGE-SCALE SANDSTONE BODY WITH EVIDENCE OF LATERAL ACCRETION
(STOUPE BROW)

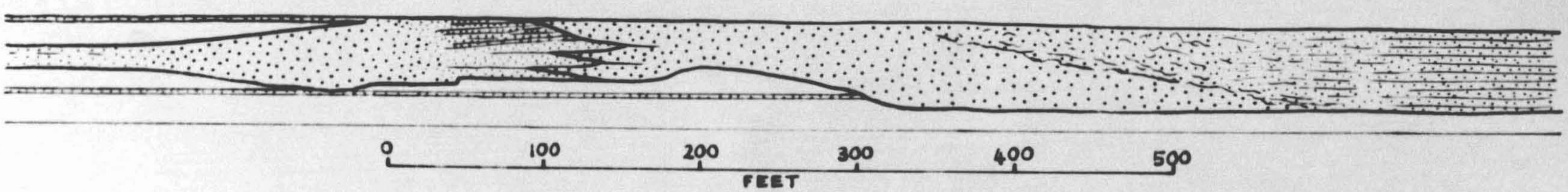


FIG. 1.13 LARGE-SCALE COMPOSITE CHANNEL SANDSTONE
(HAWSKER)

scale, with the bulk of the mass belonging to one phase. The sandstone at Mallyan Spout is remarkable for its great thickness and apparent homogeneity. The sandstone is at least 55 feet thick and unusually massive; horizontal bedding planes are seen at the base. To the south the sandstone becomes flaggy with partings of silty shale and locally a conglomerate of angular silty shale fragments is present at the base.

In addition to the sandstone at Mallyan Spout a large sandstone body is developed at Thomason Foss in Eller Beck which is at least 85 feet thick, the base being obscured. This sandstone is less homogeneous than that of Mallyan Spout and comprises bedding units of up to ten feet thick separated by thinner-bedded, and sometimes pebbly, partings. The topmost beds consist of fifteen feet of horizontally bedded flaggy sandstones with silty shale laminae. An upward and a lateral passage into thin-bedded sandstone with silty partings is characteristic of all channel sandstones. This sandstone appears to represent a continuous phase of deposition since no substantial silty partings are present.

Another thick sandstone development occurs in the West Cliff of Whitby. This is seen to comprise a complex series of sandstone bodies and is particularly interesting for the development of laminated beds to the side of the main channel sandstone (see p. 36). In this case a definite phase of sedimentation of fine-grained sediments occurs between the two main phases of sandstone deposition, though it may be of only a local nature.

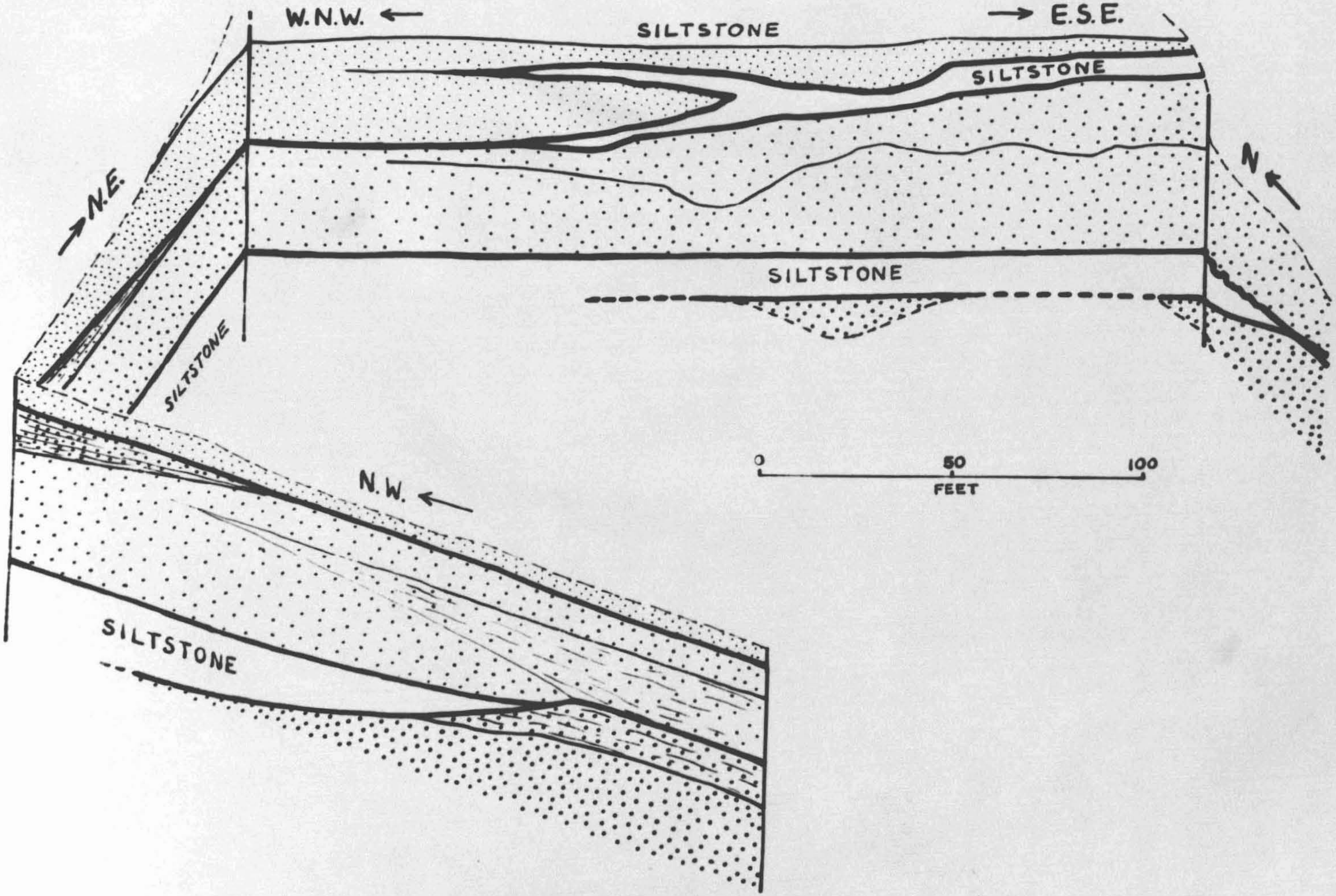


FIG. 1.14 THREE-DIMENSIONAL DISTRIBUTION OF CHANNEL SANDSTONES (AISLABY QUARRIES)

2. STRUCTURES ASSOCIATED WITH THE BASE

a) Normal erosion and its effects on eroded sediments. In many channels the erosion of the side walls has produced a relatively smooth surface. In other cases pronounced irregularities may occur which are sometimes related to the lithology of the wall sediments (fig.1.15) but which more often appear to be a reflection of variation in current strength. The irregularities often assume a step-like form which may be rounded or very sharp (fig.1.16). For small distances the wall may be nearly vertical, but overhanging walls have not been observed.

The nature of the contact appears to depend on two factors: the degree of lithification of the wall sediments prior to erosion and the time-lapse between erosion and sand deposition. In most cases the steepness of the contact indicates that compaction has been considerable and in the example illustrated in figure 1.16 erosion appears to post-date the formation of siderite nodules in the shale. This latter example also serves to illustrate the importance of the time factor; it is unusual in that the contact is very sharp and angular fragments of the eroded sediment occur in the adjacent part of the channel-fill sandstone. In contrast a few examples have been observed in which the sediments have been too unconsolidated to form an upstanding erosion surface. In figure 1.17 silty shales are seen to have slumped downwards,dislocating lumps of sand in the process; at the same time rafts of peaty material, which are common in the sandstone, were incorporated marginally.

b) Elongate Scours. A characteristic feature of many channel

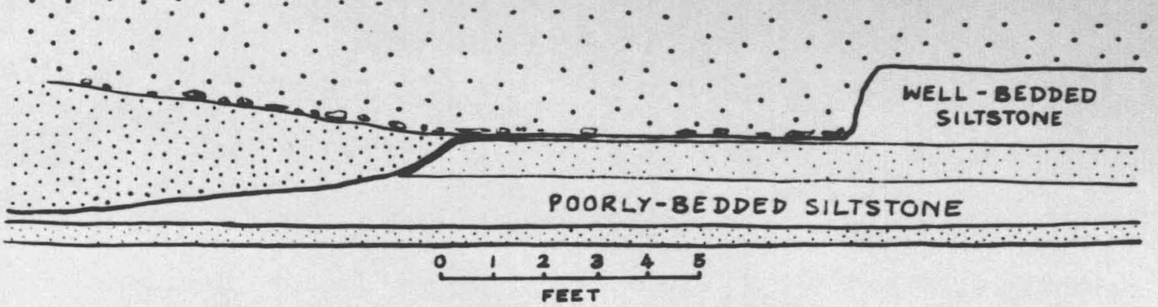


FIG. 1.15 INFLUENCE OF SEDIMENT TYPE ON EROSION

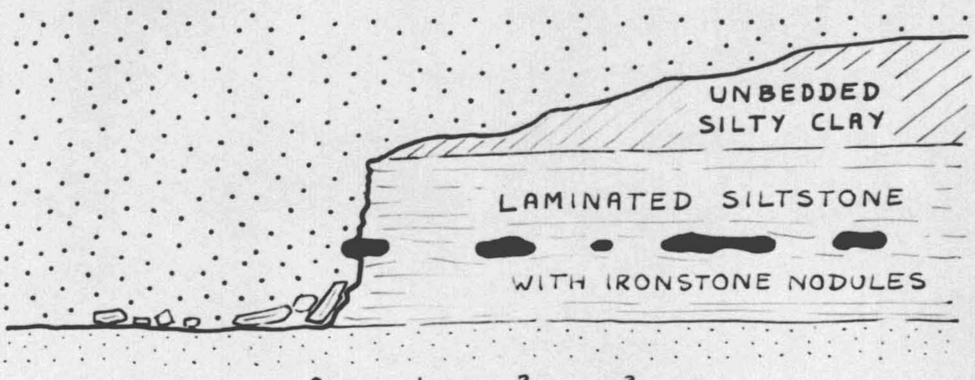


FIG. 1.16 EROSION OF LAMINATED SILTSTONE AND NODULES

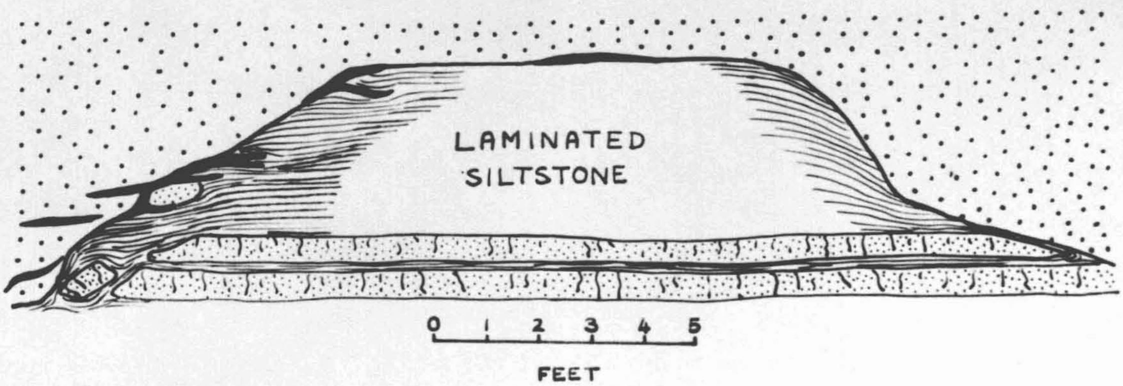


FIG. 1.17 SLUMPING OF SEDIMENT DOWN CHANNEL SIDES

sandstones with flat erosive bases is the presence of elongate erosional scours. These are developed when the sandstone overlies poorly bedded silty clays. They have a smooth outline and are always straight; in cross-section they vary from V-shaped to U-shaped and commonly show some degree of asymmetry. The scours range from a few inches to several feet in width but are generally in the order of six inches to one foot across. The scours have never been observed for more than about three feet along their length, but their constancy in these exposed portions implies considerable longitudinal extent. They are always parallel or sub-parallel with the sides of the overlying channel where observed and in any one group show little variation in direction. A typical set of scours is shown in cross-section in figure 1.18, showing the shape of the scours and their proximity to one another.

The fill of the scours shows some variation. In some cases they are massive and rather coarser-grained than the remainder of the sandstone; horizontal bedding may be seen and a horizontal parting is often present separating the scour fill from the overlying sandstone. Normal downstream cross-bedding is never present, but in some sandstones the grooves show lateral fill (fig.1.19).

c) Linguoid flute casts. These have been observed at only one locality where they are associated with erosional scours (fig.1.20). They are arranged in a narrow zone running parallel to the channel axis and from the current bedding direction in the overlying sandstone it is clear that the narrow end points upsteam. Individuals reach about two feet in width and have a relief of three to four inches.

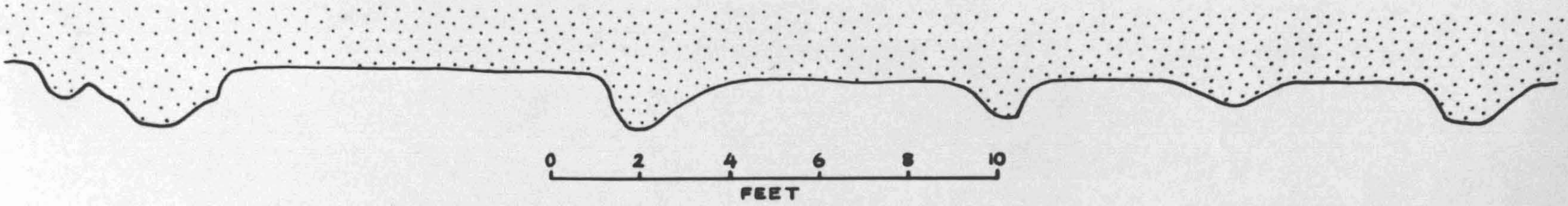


FIG. 1.18 SECTION OF SCOURS AT BASE OF CHANNEL (SALTWICK)

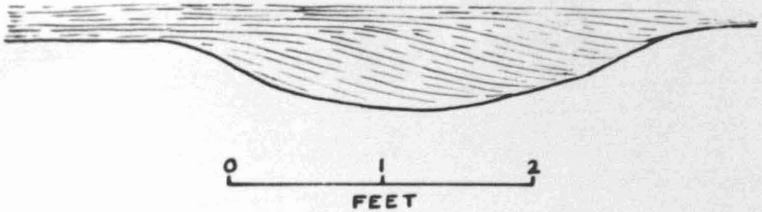


FIG. 1.19 TRANSVERSE CROSS-BEDDING IN SCOUR FILL

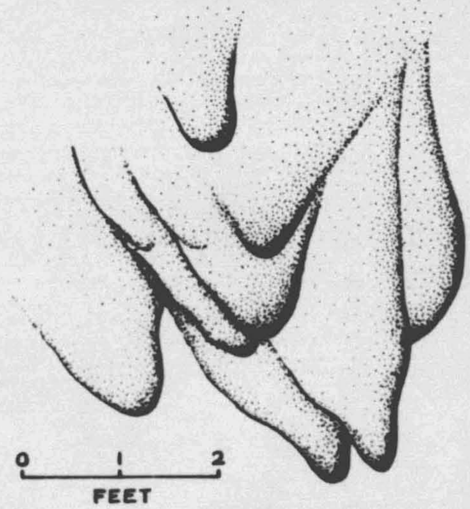


FIG. 1.20 FLUTE-CASTS ON BASAL SURFACE OF CHANNEL

d) Load casts. Loading is not a common feature of the base of large channel sandstones, presumably because erosion exposes compacted beds. Occasionally irregular loading at the base does occur in the marginal parts of the channels. More commonly loading occurs within composite units in which silt intercalations separate different sand phases. In one case the load casts reach 1ft. 3in. by 2 feet in cross-section, and appear to show a commonly oriented elongation; these are believed to have been initiated as erosional scours whose sandy fill has foundered into the little-compacted silty interdalation. The separation of between 2 and 12 feet compares with that of scours.

3. INTERNAL STRUCTURES

a) Cross-bedding. The description of cross-bedded units was first standardised by McKee and Weir (1953) and their classification was expanded to cover all distinguishable types of cross-stratification by J.R.L. Allen (1963). The nature of the cross-bedding in the sandstones studied here is restricted to perhaps five large-scale and two small-scale types on Allen's classification. Since Allen's types are clearly defined and illustrated they provide a useful reference for structures observed in the field and they have therefore been used in the following discriptions.

Cross-bedding is a common feature of many but by no means all channel-fill sandstones. Often where it does occur it is only a subsidiary structural type. In some of the larger sand bodies cross-bedded units of up to 3 feet in thickness are common and occasionally units of 8 and even 12 feet are present. More commonly the units are variable in thickness and less than 3 feet thick, with very irregular outlines due to erosion by succeeding units.

Trough cross-bedding (pi-type) is the commonest form, the troughs varying between one foot and several feet in width. Typical cosets are shown in figures 1.21a and 1.21b which represent sections perpendicular to and parallel to the general flow direction respectively. Considerable erosion of individual units is characteristic. Figure 1.21b illustrates a feature found in the upper and lateral parts of many sandstone bodies which is the development of planar units (beta-type). These units are usually thin and show progressive downstream and lateral thinning. They are clearly seen to climb at a low angle and this climbing effect is accentuated by the development of extensive 'bottom sets' at the expense of the 'foresets', until finally a flat laminated unit results. These flat laminae, however, themselves developed small-scale cross-bedding (lambda-type).

In some sandstones troughs are developed on a larger scale, when they are few in number and occupy a single horizon. These generally range in width from 3 to 6 feet and in depth from 6 to 12 inches, but occasionally horizons of larger scale troughs reaching twenty feet across (fig. 1.22) are present. Figure 1.23 illustrates more complex large-scale trough bedding.

Solitary trough cross-bedded units (theta-type) have been observed cutting down into both flat-bedded and cross-bedded units. These range in width from 5 feet to over 50 feet and the maximum erosional depth observed is 18 feet. The sandstone fill shows characteristic bedding in which all units thicken towards the axis of the

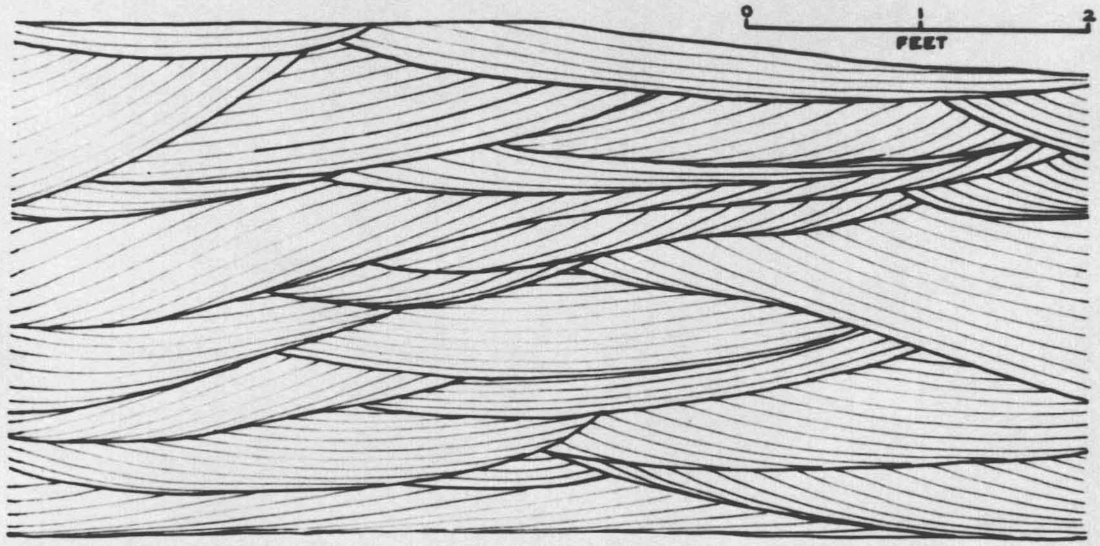


FIG. 1.21A SECTION OF CROSS-BEDDING PERPENDICULAR TO THE AVERAGE FLOW DIRECTION

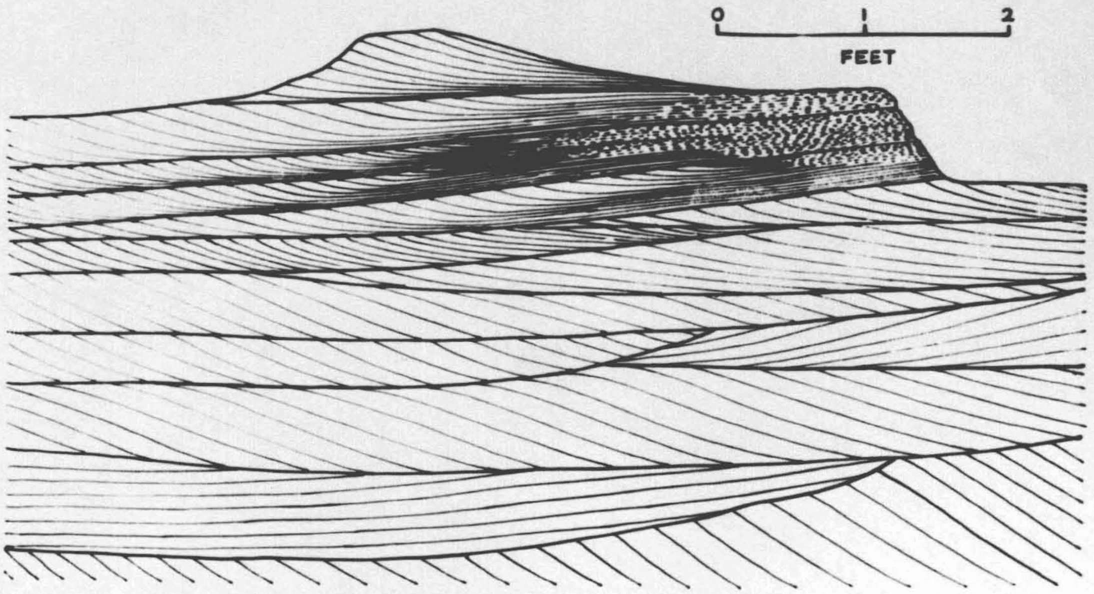


FIG. 1.21B SECTION OF CROSS-BEDDING PARALLEL TO THE AVERAGE FLOW DIRECTION

trough (fig. 1.24).

Planar cross-bedding is found rarely in grouped sets (omikron-type) but generally occurs as solitary beds at the base of large channels (alpha-type). Such beds may attain thicknesses of over ten feet and in one case a bed has been traced along the foreset strike for two hundred feet. As mentioned above thin tabular sets may occur in the uppermost part of trough-bedded cosets.

Low angle cross-bedding of the type described on page 11 is found to a limited extent in the larger channels. It is in particular developed throughout a large member of the composite channel at the base of East Cliff, Whitby, which will be described later. A large number of channel sandstones, though thickly bedded, display no internal structure. Presumably the sand must have been laid down in laminae although lack of grain-size variation and a constant current regime may produce a homogeneous bed. The most likely form of bedding to show this feature is the low angle lateral type, since this displays a more constant grain size than normal cross-bedding.

Small-scale cross-bedding is developed extensively at the top of channel sandstones and, together with silty intercalations, may form up to fifteen feet of thin-bedded sandstone. The cross-bedding is sometimes of the lambda-type but is more usually of the kappa-type. Occasionally small-scale cross-bedding is developed in the main part of a sandstone body and is then usually of the nu-type. Individual troughs usually measure between three and twelve inches across with depths up to one inch.

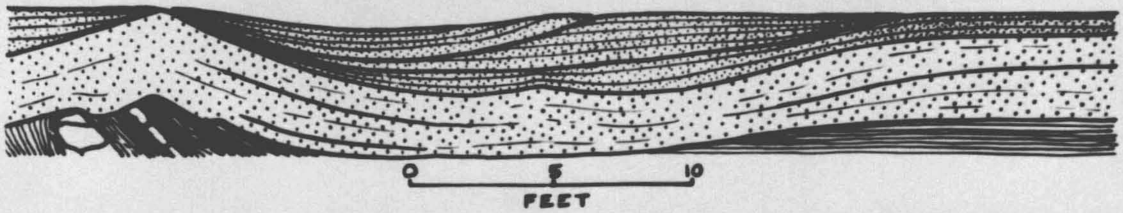


FIG. 1.22 TROUGH CROSS-BEDDING AT BASE OF SANDSTONE (WATER ARK FOSS)

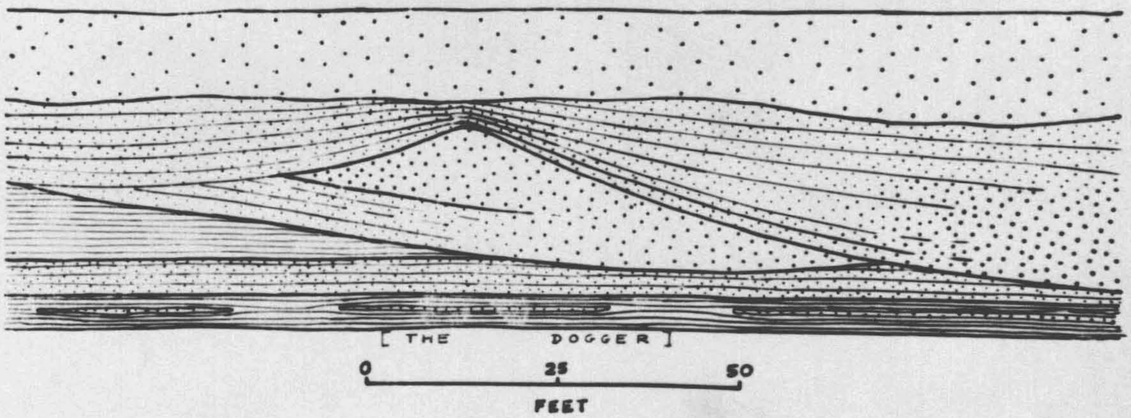


FIG. 1.23 LARGE-SCALE TROUGHS (BECK HOLE)

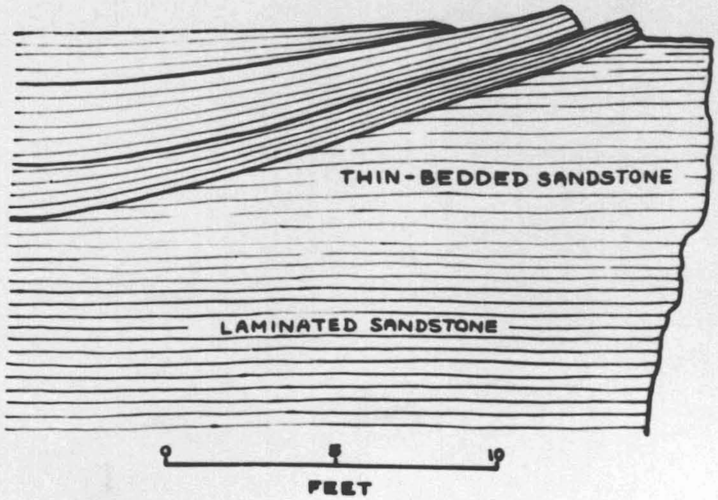


FIG. 1.24 PART OF ISOLATE TROUGH ERODING FLAT-LYING SANDSTONE

b) Parting lineation (Primary current lineation). Parting lineation is very rare, having been observed in only one case, where about one foot of flat-laminated and parting-lineated sandstone was overlain immediately by small-scale nu-type trough cross-bedding.

c) Horizontal bedding. This is of common occurrence in the upper and lateral parts of channel sandstones but is rarely observed elsewhere. It occurs at the base of one large sandstone body which shows a downward passage into flat-laminated sandy siltstone.

d) Inclined bedding. Inclined bedding may be distinguished from cross-bedding in that it involves the sedimentary units themselves. Where these are developed they generally run at a moderate angle from the top to the base of a channel sandstone and appear to strike parallel to the channel elongation. In the East Cliff sandstone (fig. 1.46) and in a channel on a similar horizon at Kettleness inclined planes can be seen in individual sand bodies and in fact the sand bodies themselves show a similar relationship to one another. The problem of the origin of such beds will be dealt with later (p. 32). More usually the inclined beds may make up part of normal channel sandstones (figs. 1.12, 1.25)

e) Erosion surfaces. Internal erosion surfaces are a feature of all composite sand bodies. Where a small time interval of non-deposition is represented the erosion surfaces are within a single sandstone body and usually die out rapidly downwards and laterally. In other cases the erosion is of a more extensive nature and a period of laminated silty sand deposition may have occurred during a temporary migration of the main channel flow. Erosion surfaces into silty intercalations may show all of the bottom structures already described; those into earlier sandstone beds have a smooth general outline with local irregularities. Ferruginous concretions and woody detritus are commonly associated with the base of the sand above the erosion surface. The erosion in many cases is haphazard but when developed on a relatively large scale a trend may usually be observed which reflects progressive migration of the main channel flow. The erosion planes sometimes appear to represent successive stages of bank incision, generally accompanied by slow vertical migration (fig.1.25). At other times a series of erosion surfaces occurs on the accumulative side of a channel. In figure 1,27, for example, erosion surfaces closely parallel the sedimentation surfaces and serve to accentuate them. Some of the prominent partings in the channel shown in figure 1.12 are also slightly erosive. Such erosion surfaces are in fact a guide to the nature of the original sedimentation surfaces in channels whose sandstone fill is otherwise apparently structureless, and add further support to the suggestion that some of the 'massive' sandstones may have been deposited by lateral accretion (see page 34).

f) Pebbles. In a few cases fragments of laminated silty sediment are associated with erosion surfaces. These are often angular and appear to be very locally derived. They have only been observed in massive or inclined-bedded sandstone. The only occurrence of a thick bed amounting to a clay-silt conglomerate is in the lateral part of the lowest beds exposed of the Mallyan Spout sandstone. Pebbles which appear to represent ironstone concretions derived from the

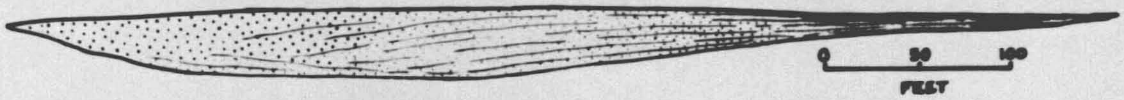


FIG. 1.25 INCLINED NON-EROSIONAL BEDDING PLANES

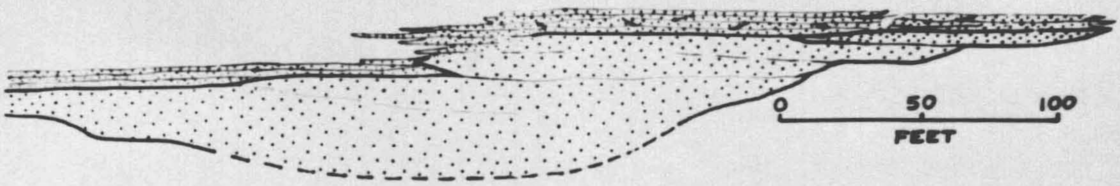


FIG. 1.26 VERTICAL AND LATERAL MIGRATION OF CHANNEL

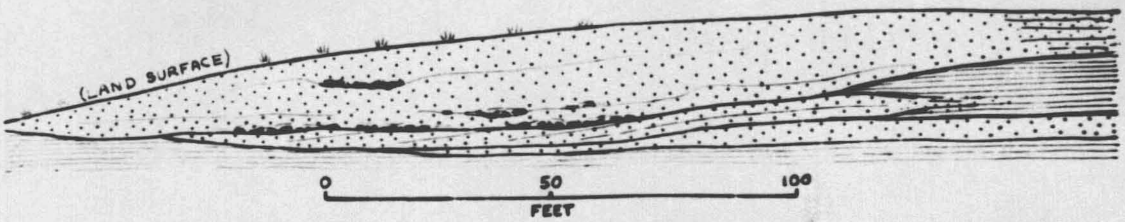


FIG. 1.27 EROSION ACCENTUATING INCLINED BEDDING PLANES

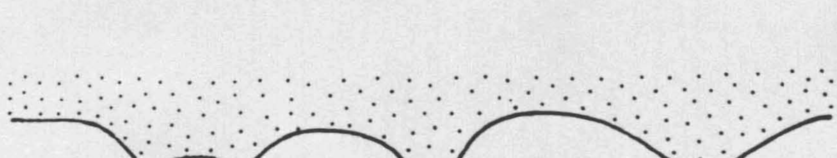


FIG. 1.28 LOG ASSOCIATED WITH ELONGATE SCOURS

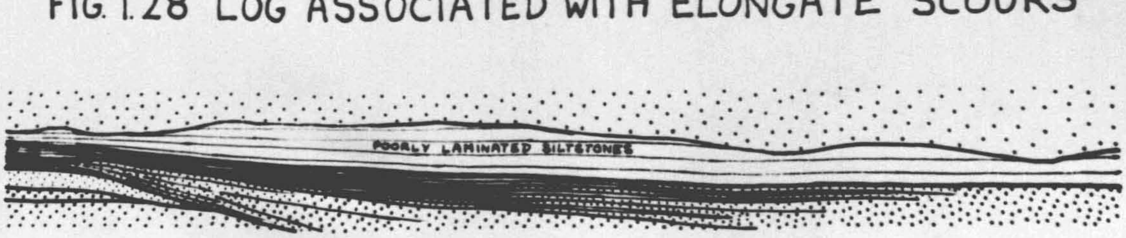


FIG. 1.29 DIAGRAM OF LATERAL PASSAGE FROM CHANNEL SANDSTONE INTO OVERBANK SILTSTONES; DIP PRODUCED BY COMPACTION OVER EARLIER CHANNEL LENS (WHITBY W. CLIFF)

surrounding sediment are concentrated along bands, usually erosive, in some channel deposits. These usually have diameters of about one inch and rarely exceed three inches. Minute pebbles have been observed along foreset boundaries in one cross-bedded unit, more commonly they occur along erosional trough bases. The pebbles are usually sparsely distributed along any one plane; these pebbly planes tend to occur in particular channels while in many sandstones they are absent. Occasionally larger ferruginous nodules occur; these are usually highly weathered and contain much silt and fine sand in laminae. In some cases they are seen to have rather diffuse boundaries and are believed to represent replacement of fine-grained sediment after incorporation into the sandstone.

g) Logs. Logs are widely distributed in channel sandstones. The largest, which may reach two feet in diameter, are usually found at the base of the erosion channels. They are often associated with elongate scours and their own relationship to the underlying beds implies that they lay on the channel bottom and acted as loci to the development of scouring (fig.1.28). Alternatively logs may occur in disoriented masses at the base of erosion planes which are usually internal and of little extent. The fragments are variable though rather small in size and appear to have been deposited rapidly with little subsequent sorting.

h) Distribution of iron. The highly ferruginous nature of some sandstones and the localised distribution of the iron minerals is very striking. The iron occurs as a hydrated oxide and varies between chocolate brown and dark red in colour. The distribution of the iron oxide is never haphazard, but shows relationship to bedding and is highly selective. Particular sandstone units may be affected while surrounding units are not, and within a single unit variation on a smaller scale is usually present. The foresets of cross-bedded units may show internal variation in the form of a gradation in iron content from one surface to the other; usually the lower part is most ferruginous, but reversals are not uncommon. In low angle cross-bedded units a downward decrease in iron content appears to be more common, and this also applies to large-scale lateral bedding units. The sandstone occurring above local erosion surfaces is often iron-rich, especially where it is rubbly due to the presence of stems and logs.

On a large scale it appears that only a relatively small proportion of channel fill sandstones show this feature, and only rarely are large masses of sandstone affected. On both the large and the small scale coarse grained sandstone, having a relatively high porosity, appears to be selected. There may also be some regional control on the occurrence of iron-rich sandstone since they are mostly found on the coast north of Whitby. Leaching may account for their absence inland but this is improbable, since ferruginous sandstones in the Eller Beck Bed, though often highly weathered, retain their iron in the oxide form.

D. MARGINAL FACIES ASSOCIATED WITH CHANNEL SANDSTONES

All channel sandstones, both large and small-scale, show a certain amount of lateral passage into finer grained sediments. As a general rule these beds contain a considerable proportion of sand proximally with subsidiary amounts of interbedded and interlaminated shaly siltstone. The proportion of silt and clay to sand increases away from the channel and this is normally accompanied by a slow decrease in thickness.

The nature of these beds may vary to some extent. The lateral passage in some cases is confined to the top of the channel; this occurs in channels in which the massive sandstone is replaced upwards by impersistent beds of flaggy sandstone, often showing small-scale cross-bedding, with silty intercalations. This top facies commonly spreads over a wide area in the final stages of deposition.

Another form of passage beds is rather rare but always developed on a large scale. This consists of well defined and continuous beds of sandstone and silty shale which are seen to thicken by swelling of sandstone beds towards the main channel. The laminated beds at the base of the West Cliff, Whitby are of this type, and similar beds have been observed in Rosedale. Unfortunately in both cases the relationship to the channel sandstone is not clearly seen.

The Whitby beds are seen to possess a moderate dip to the west, but this is believed to be due to compaction around the sandstone mass developed at a lower horizon to the east. The beds are seen in the east as laminated silty shales with sandy ribs in the lower part; the basal beds become dominantly sandy, with ripple drift and finally pass into massive beds with some current-bedding. The overlying beds have become sandy with ripple drift in the lower part and these in turn appear to pass into more massive sandstones. The uppermost beds remain silty or shaly. The passage from fine-grained laminated beds to massive and cross-bedded sandstones is represented diagrammatically in figure 1.29.

The laminated beds show a variety of delicately preserved structures. The lamination is sometimes interrupted by minor scours and a few larger channels (fig.1.30). Load structures are also developed, possibly associated with scouring (figs, 1.31, a & b). Vertical sandstone bodies occur along certain horizons and in plan form a polygonal pattern (fig.1.32a). There is always a parting between the base of the sandstone fill and an underlying sandstone where present (fig. 1.32b), and usually the top is confluent with a sandstone bed or laminae (fig. 1.32c). These features, together with horizontal internal laminae in some cases, suggest sandfilled dessication cracks rather than sandstone dykes.

The interbedded silty shale and sandstone series in Rosedale is on an even larger scale, but a similar thickening of first the lower sandy beds and then the upper is seen with the development of thick sandstone units which are flat-bedded with only minor cross-bedding. One horizon in the upper part is characterised by the development of very large irregularly shaped load casts penetrating the underlying shales to a depth of up to three feet. In one series of flaggy sandstones marginal to a more massive development structures resembling bivalve resting burrows are present along one horizon.

A third lateral facies type also consists of laminated sandstone and silty shale but the interbedding is generally on a small scale with diffuse boundaries. Units of this nature are associated with

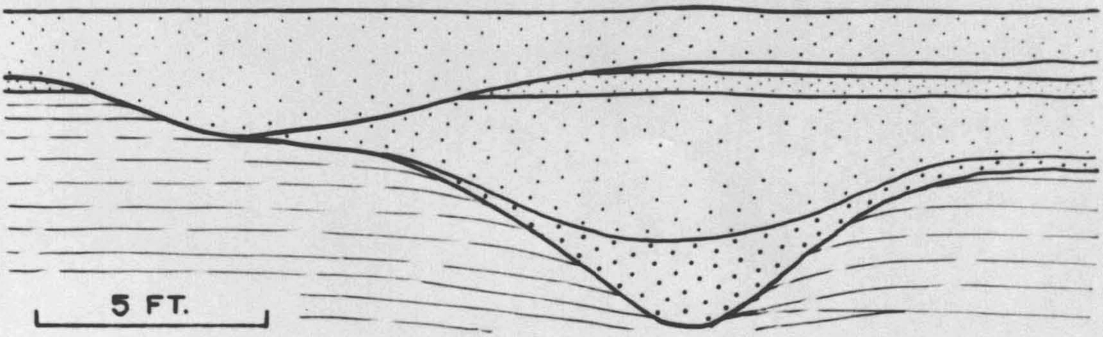


FIG. 1.30 SCOURING AT BASE OF SANDSTONE HORIZONS

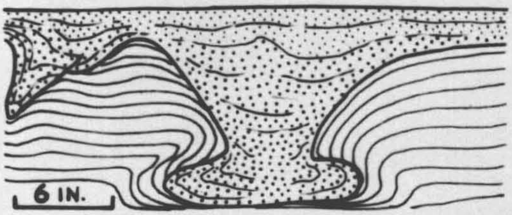


FIG. 1.31A SOLITARY LOAD CAST

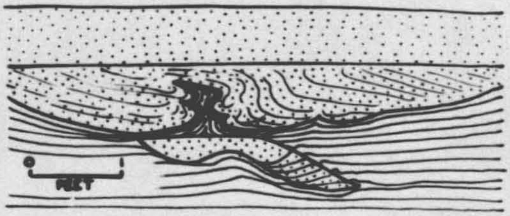


FIG. 1.31B LOADED CROSS-BEDDING

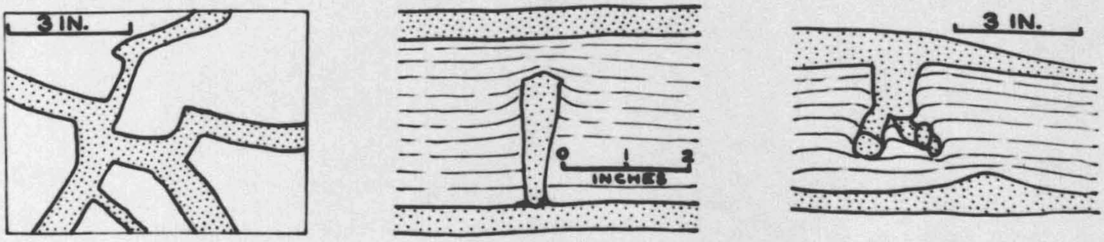


FIG. 1.32 DESSICATION CRACKS: A) PLAN ; B) SECTION ; C) COMPLEX

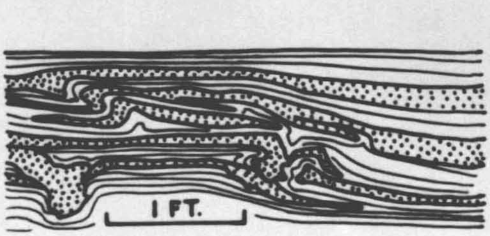


FIG. 1.33 CONTORTIONS IN LAMINATED SANDY SILTSTONE

both small and large channel sandstones. In the former the channel sandstone breaks up into laminated beds on one side which themselves pass laterally into increasingly shaly sediments. Small irregularities of bedding are common and sometimes more pronounced contortion of bedding is developed (fig.1.33). Carbonaceous material and carbonised plant fragments are often present in the finer grained laminae.

Beds of similar lithology are associated with larger channels; these may reach ten feet in thickness whereas those associated with smaller channels rarely exceed three feet in thickness. Another striking feature of these larger-scale units is the development of high depositional dips in the proximal part, which gradually decrease away from the channel until flat-bedding is attained, with thinning and with an increase in the proportion of fine grained sediment. These units are developed on only one side of a channel phase which corresponds to the side of maximum erosion, or the direction of migration of the channel. Where the channel shows little composite nature the passage into sandstone and silty shale may be preserved (fig. 1.34) but where the channel is of a composite nature erosion of the dipping beds is often shown (fig.1.35). Owing to the composite nature of most channels such erosion is common. The examples in figures 1.36 and 1.37 are unusual in that a continuous dipping series has developed during more than one channel phase; more commonly the dipping units appear to belong to one phase only.

The sediments in these dipping units show very poor overall

FIG. 1.34 LATERAL PASSAGE OF CHANNEL SANDSTONE INTO SILTSTONE/SANDSTONE UNIT SHOWING INCLINED BEDDING

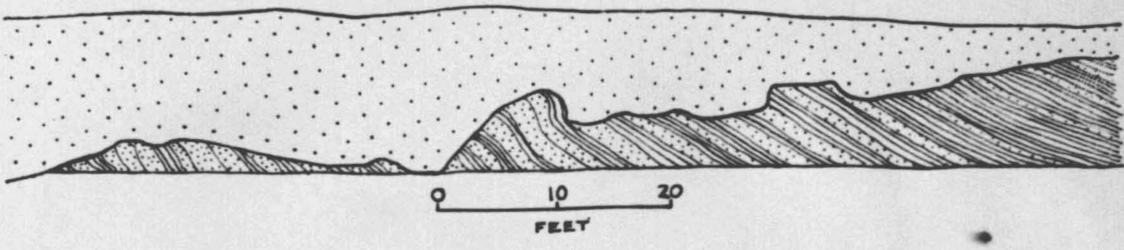


FIG. 1.35 EROSION OF INCLINED BEDS DUE TO LATERAL MIGRATION OF THE HOST CHANNEL

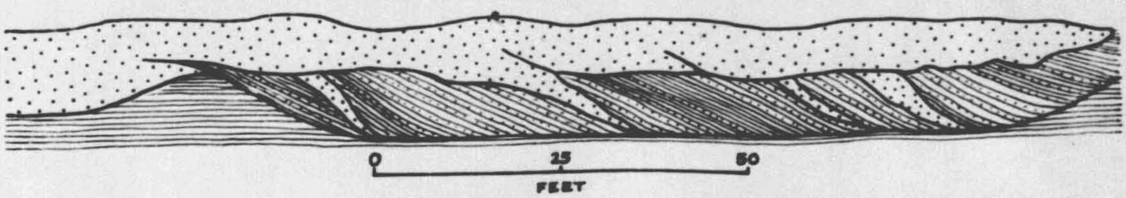


FIG. 1.36 INCLINED BEDS SHOWING A) MIGRATING CHANNEL BOTH FEEDING AND ERODING THE UNIT; B) DEPOSITION INTO EROSION HOLLOW

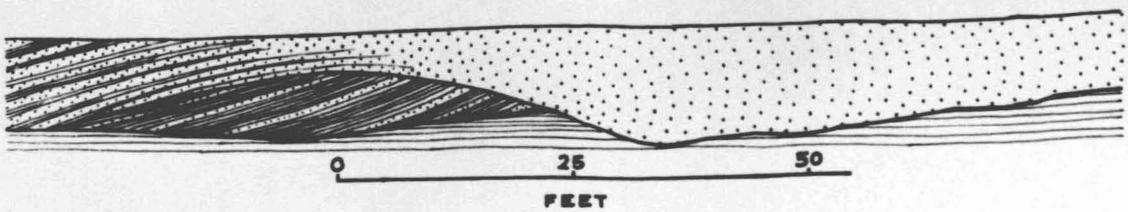


FIG 1.37 CHANNEL SANDSTONE AND ASSOCIATED INCLINED BEDS SHOWING AN EROSIONAL CONTACT WITH EARLIER INCLINED BEDS

sorting but individual beds and laminae are better sorted. Plant remains may be abundant in the finer-grained laminae as in the example along The Scar at Whitby. Small-scale faulting and sometimes folding is common; downslope movement is indicated, the faulting being both normal and reverse. The dip of the laminae normally reaches 25 or sometimes 30 degrees. In examples where the lateral passage is preserved the unit is associated with the uppermost part of the main sandstone body, or one of its phases. It also appears that the units develop on the erosional side of migrating channels. These two factors account for the common erosion of the unit by later channel phases.

E. COMPACTIONAL EFFECTS ASSOCIATED WITH SANDSTONE BODIES

The most obvious compactional effect associated with channel sandstones is the relative compaction of the fine-grained sediments which usually lie to either side of the sandstone body. The nature of these effects will depend largely on the degree of compaction of the beds into which erosion has taken place. The shape of the sandstone body will also influence compaction. The large-scale sandstone bodies are characterised by a moderately low channel ratio and by a tapering rather than abrupt lateral termination. The time involved in the formation of these sandstones appears to have been sufficiently great for compaction of the underlying beds to be fairly complete even where loading at the very base of the unit indicates that the topmost sediments were unconsolidated at the initiation of sandstone deposition. In most cases the nature of the basal erosion indicates a considerable degree of compaction. As a result severe compactional disturbances seldom occur. Where the marginal sediments are products of early lateral deposition by the channel itself compaction may be more evident, and disturbance of bedding in the dipping unit shown in figure 1.38 is believed to be the result of compaction between two sandstone units.

Severe compactional effects have also been observed associated with the small-scale composite channel illustrated in figure 1.10. It consists of reverse fault which indicates that the upper marginal beds have been forced down the channel margin (fig.1.39). The faulting results from the abrupt nature of the channel margin, and the development of a reverse rather than a normal fault appears to be due to the shape of this margin, which is inwardly inclined. Some horizontal movement parallel to the channel axis is indicated since the beds on either side of the thrust are not of the same thickness.

The effects of compaction on a larger scale may be seen in all cases. The shape of most channels in cross-section is inevitably lensiform and the thickest, central part displaces eroded strata which may be present beneath the thinner lateral parts. Since there is more compactable sediment beneath these lateral parts subsidence occurs, with the result that the upper surface of the sand body becomes curved, and the cross-section biconvex. This outward dip at the top of channel sandstones cannot be original in view of the nature of deposition. I t is noticeable that where troughs are developed in sandy beds or where channels are filled with fine-grained sediment (see p. 55) that the upper surface is flat, and the lower

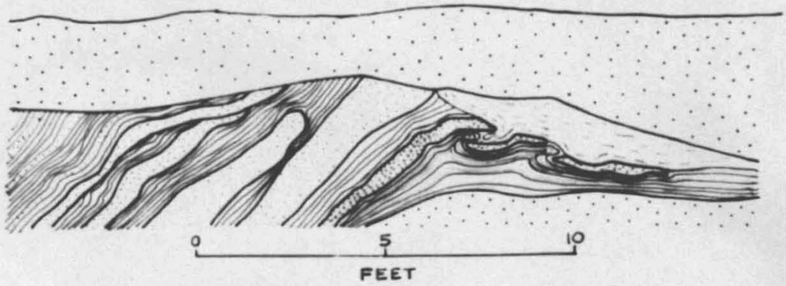


FIG. 1.38 CONTORTED BEDDING IN INCLINED BEDS DUE TO COMPACTION BETWEEN TWO MASSIVE SANDSTONES

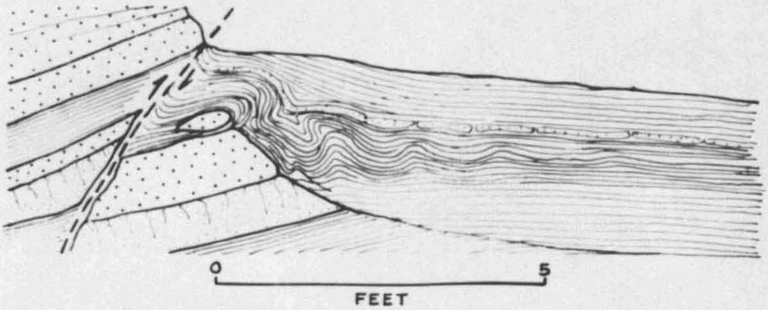


FIG. 1.39 OVERTHRUSTING AT CHANNEL MARGIN (SEE FIG. 1.10)

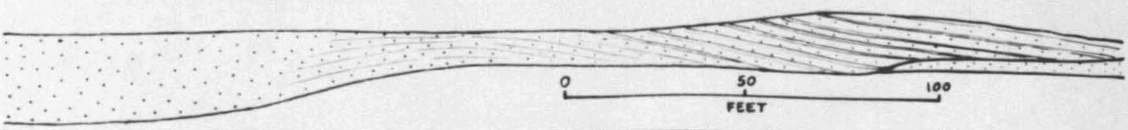


FIG. 1.40 RECONSTRUCTION OF FIG. 1.34 ALLOWING FOR COMPACTION

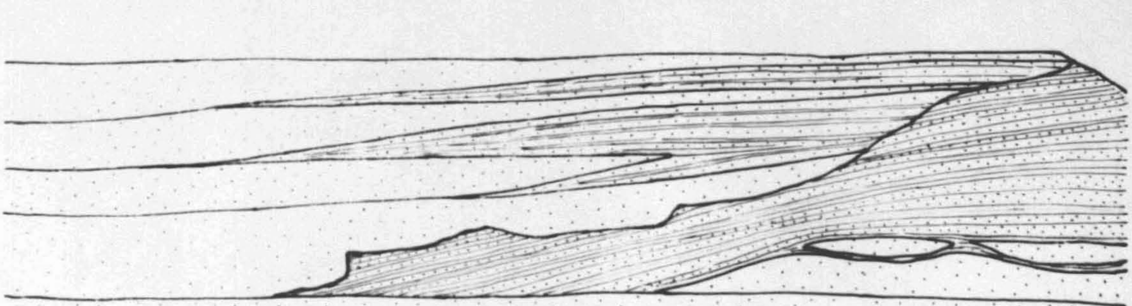


FIG. 1.41 SPION KOP, WHITBY : THICK SERIES OF INCLINED BEDS CUT INTO BY LATER SERIES, WHICH PASSES INTO SANDSTONE

surface strongly curved. These probably reflect the original cross-section of all sediment-filled channels.

An attempt has been made in figure 1.41 to reconstruct original channel form by redrawing the present dimension assuming a flat upper surface. A good example of compaction of beds over a channel is provided by the interbedded sandstone and shale unit of West Cliff, Whitby (fig.1.29) these beds dip down towards their host sandstone since they ride up over the top of an earlier channel. The rapid thinning of these beds over the earlier sandstone may indicate that compaction had produced some surface relief at the time of their deposition, and it may be this very effect which is responsible for the sudden shift in channel course that appears to have taken place frequently in these beds.

F. DISTRIBUTARY FACIES SEDIMENTATION

The deposits associated with the distributary sandstones can for the most part be grouped into several facies types representing the micro-environments which are known from present-day fluviatile deposits. This is made possible by the field distribution of the facies types and by direct comparison with the recent sediments.

1. CHANNEL ENVIRONMENT

With one exception deposition in the large-scale channels appears to have been preceded by erosion of the previously deposited strata. The depth of erosion varies considerably and is difficult to estimate owing to compactional effects and to later erosion during lateral migration of the channel . Probably initial erosion occurred to a máximum depth of 15-20 feet. A small amount of deposition involving pebbles and large logs may take place at this stage. After the initial cutting down the development of elongate scours may represent a period of less violent erosion. A lag deposit of medium to coarse sand may occur within these scours and in the lowest few inches of the main sandstone.

The principal structural type associated with channel bottom deposition is trough cross-bedding. This is believed to have developed through the migration of trains of large-scale and small-scale asymmetrical ripple-marks with pronouncedly curved crests (Allen, J.R.L. 1962, p.110). Both Allen (1966, p.183, Table 4) and Simons and others (1961 p.59) consider the ripple-marks to have formed under the tranquil flow regime. Simons and others (Fig.22) relate the distinction between small-scale and large-scale forms to water depth and velocity; Allen considers hydraulic roughness or smoothness to be the critical factor. Many channels show some trough cross-bedding and in a few it is extensively developed, but it is significant that a great deal of the sandstone in the channel fills are totally devoid of cross-bedded units. Possibly this lack of structure is due to very rapid deposition from suspension but in some cases, at least, it appears to be the product of lateral accretion.

Parting lineation, ascribed to the upper flow regime, is rare, but an interesting feature observed in one case was the upward change-over from horizontal bedding with parting lineation to small-scale trough cross-bedding with no angular shift in orientation. On

Allen's hypothesis this represents a direct switch from rapid flow and hydraulic roughness to tranquil flow and hydraulic smoothness.

2. SAND BAR ENVIRONMENT

Large-scale solitary planar cross-bedded units are ascribed by Allen (1963, p.101) to building of solitary banks with smooth or curved leading edges above slip-off faces. Cross-bedding produced by this process is described from beach sand bars (Thompson, 1937; Plate 7 Fig. 2) and from fluviatile sand bars (Leopold and Wolman, 1957; Ore, 1963). A well exposed cross-bedded unit which may be of this type is traceable for over 200 feet along the foreshore. It shows a maximum dip of 30 degrees to the north and is overlain by trough cross-bedded units indicating flow to the west and north-west. The steep foreset dip distinguishes this unit from others which show foreset dips perpendicular to the channel axis (see p. 11) and which are believed to be point bar deposits. Ore (p.10) states that the angle of dip of the bank forsets represents the angle of the rest of the sand, whereas cross-bedding produced by lateral accretion of point bars show dips corresponding to the angle of inclination of point bar surfaces which (Allen, J.R.L. 1966, p.177) is lower than the angle of the rest of the sediment.

3. POINT BAR ENVIRONMENT

Low-angle cross-bedding of two principal types is observed in the channel sandstones ~~(see fig.)~~. The more delicate type is confined mostly to small-scale channels but has been observed in larger ones. Such units appear to represent a single phase of migration of a small stream; both lateral and vertical components may be evident. The delicate nature of the laminae, their uniformity and the low contrast in grain size between laminae suggests deposition under very stable conditions. Such channel phases are probably short-lived. Where stream development is more persistent a composite body may be built up of such units (fig.1.10).

The second type of low-angle bedding is on a larger scale, each inclined unit measuring between a few inches and several feet, with no visible internal structure. Sometimes a certain degree of grading is present within each inclined unit, and this is often of the reversed type. Planes of erosion sometimes accentuate the partings and represent successive limits of the channel sides indicating progressive lateral accretion; only the last erosion plane on the non-depositional side is preserved.

4. CHANNEL TOP ENVIRONMENT

Point bar sediments often pass laterally in the upper part into thin-bedded sandstones with silty shale intercalations. The sandstone ribs are generally discontinuous and may show small-scale channelling at the base. Small-scale cross-bedding is common together with horizontal lamination. Similar beds are found at the top of the whole sand body though they may be preceeded by more substantial small-scale cross-bedded sandstones. Small ironstone nodules are often developed. These top channel deposits may extend laterally beyond the limits of the main sandstone indicating the last stages of channel fill, with expansion of the depositional area onto the surrounding marsh sediments. Rootlets are commonly present. In a few cases cessation of sand deposition has been more abrupt, and thick-bedded sandstone is overlain directly by fine-grained sediments.

The laterally deposited sandstones are identified by inclined bedding planes, sometimes accentuated by erosion. At other times sandstones that are principally massive possess inclined erosion planes only and in some sandstones even these are absent. This suggests that all such sandstones, i.e. most of those that do not show normal cross-bedding, may have accumulated by lateral accretion. The occurrence of bedding planes and erosion planes is sufficiently irregular as to indicate that they are by no means necessary features of lateral accretion.

5. LEVEE ENVIRONMENT

This facies is perhaps the most difficult to identify. This is due partly to lack of previous records from fossil sediments and to the lack of any detailed work on the internal structure of recent levees. The following description of the sediments composing leveees is given by J.R.L. Allen (1965, p.571) and is the most detailed to appear in the literature:

"The typical levee deposit is a layered one of alternating poorly to moderately sorted very fine sand and brown to grey clayey silt with plant shreds. The sand layers, generally micaceous and with plant shreds, persist laterally for long distances in the river banks and may reveal either even lamination or small-scale cross-stratification.

Plant rootlets and animal burrows give rise to pronounced mottling bordered by re-brown oxidised rims. The levae deposits fine away from the river channels until at the toes of the levees clayey silts with rootlet mottlings become predominant."

This description is taken from the levees of the Niger delta, but corresponds closely with the description of those from the Mississippi delta (Fisk, 1944, p.18; 1952, p.76) and from the Conneticut Valley (Jahns, 1947, pp.48 and 49). Records of recent flood deposits (Mansfield 1938,pp.702-722; Jahns,1947,p.88) consisting of interbedded silts and ine sands of a few inches in thickness, show close correspondence with the descriptions of levee deposits.

The lithology of the alternating silt and sand inclined-bedded units (p. 25) shows a striking similarity to the above description, and this implies a levee origin. Furthermore the inclined units are found on the non-depositional side of the channels and in recent rivers levees are most prominent on the convex bends of meandering rivers (Fisk,1944, p.18).

The Mississippi levees vary between 10 and 25 feet in height (Fisk 1944 p.19) and the widest levees are associated with the widest channels. From maps of the delta (Fisk,1952) many of the secondary streams are seen to possess levees of only a few hundred yards in width. The natural levees of the Rhone delta (Kruit, 1955, p.378) range in width from 2 to 4 kilometers and in height from 4½ to 1 metre. Those of the Niger Delta are described by J.R.L. Allen (1965, p.370) as narrow levees reaching two metres in height. Owing to erosion the inclined units are rather variable in thickness but it is estimated that in most cases the original maximum thickness was between 5 and 10 feet. The inclined beds may persist laterally for a hundred feet or more, with gradually decreasing dip. The extent of the flat-lying distal part is difficult to judge owing to the gradual passage into silts and clays, but the laminated beds are usually distinguishable for a hundred yards or more. A possible exception to these general dimensions is the thick series of silty sandstones developed at Spion Kop, Whitby, (fig. 1.40). Except for a high sand to silt ratio these beds are similar in type to those of the smaller units. The total vertical thickness is, ~~however, in~~ at least thirty feet, but although this appears to exceed that of Mississippi levees it has been pointed out by Fisk (1944, p.19) that subsidence of levees is a common feature and that the total thickness of sediment may be much greater than that suggested by the topographical feature produced. The sandy nature of practically the whole vertical sequence at this locality implies that a major distributary is represented and large-scale levee deposits might therefore be expected.

If these units can be ascribed to levee deposits then some explanation must be found for their internal structure. As a general rule the units are associated only with sandstones which do not show a great deal of erosion at the base, and which presumably represent shallow streams. Such streams might be expected to be more prone to overspill than those in deeper channels. Where the levee structure is not destroyed by later channel migration (fig.1.34) the maximum thickness is seen to occur a short distance away from the channel margin and up to this point truncation of the inlined bedding appears to indicate erosion. Immediately adjacent to the channel are flat-lying beds which pass into the main sandstone body. This relationship is believed to represent erosion of the inner levee margin during times of flood while deposition occurred on the outer levee margin. During recession of the flood fine-grained material would be deposited on the sloping inner levee margin and would pass into the sand being deposited on the stream bed itself. The time relationship must have been somewhat distorted by compaction and an attempt has been made to correct for this (fig.1.41) by assuming an original flat channel surface (see p. 29).

The steep dip of the beds in the proximal part must be due to rapid initial sedimentation, and where levee development has continued for some time, a gentler depositional slope is developed which more closely resembles those observed at the present day. Unfortunately no information is available in the literature which is concerned with the large-scale internal structure of levee deposits.

Since levee deposits are principally associated with the outside of meanders lateral migration of streams tends to destroy ~~the~~ the proximal part of previously formed levees. The example illustrated in figure 1.36 is unusual in this respect since the channel has migrated over the inclined beds with little erosion and is even confluent with one sandy unit. This appears to have resulted from the early development of a shallow erosional channel parallel to and possibly related to the main sandstone-filled channel. The levee deposits have in this case not built up as a bank but have filled this trough. Deposition within the trough accounts for the limited extent of the unit. The sandstone channel is very shallow and it appears that during flood stages some bedload sediment has been deposited in the trough, producing unusually coarse-grained and thick units. The lack of relief of the levee deposits has permitted lateral migration of the channel with little or no erosion below.

The rate of deposition of these units is not clear, but it is f significant that rootlets are not present in the proximal inclined portions, which may again indicate rapid initial accumulation.

G. SANDSTONE UNITS OF UNCERTAIN ORIGIN

Two sandstone bodies are described here which are unique as far as present exposures are concerned. One of these consists simply of a set of inclined strata which lies discordantly on horizontally bedded strata, (fig.1.42). The bedding of the discordant beds is parallel to the erosion plane which dips at an angle of 20 degrees. Individual beds are of very uniform thickness. The section is obscured to the north of this exposure but it seems probable that the dip of the upper beds flattens out and that no corresponding erosion surface is present that might suggest a simple channel. Erosion might have been produced by lateral migration of a stream but the bedding in the sandstone is unlike any observed in other channels, and appears to have accumulated by vertical accretion in a standing body of water.

The second structure is better exposed but more complex. It a appears to consist basically of a trough cut into horizontally bedded sandstones (fig.1.43). Two sides of this trough are seen and

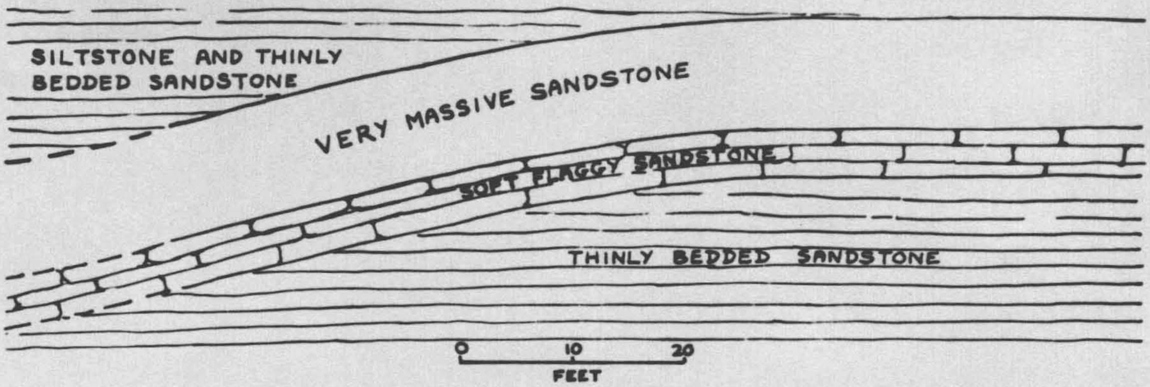


FIG. 1.42 PLUNGING SANDSTONE OF UNIFORM THICKNESS (GREAT FRYUPDALE)

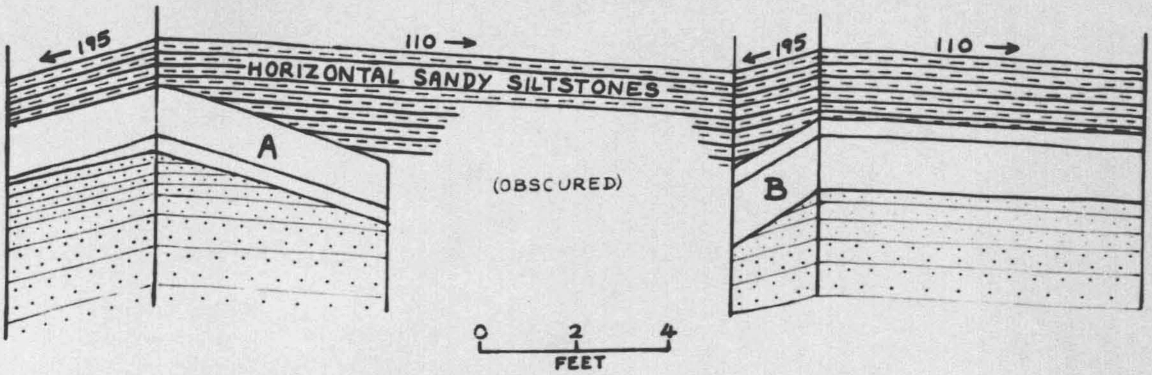


FIG. 1.43 PLUNGING 'CHANNEL' ; AISLABY QUARRIES (VIEWED FROM BASE OF CLIFF)

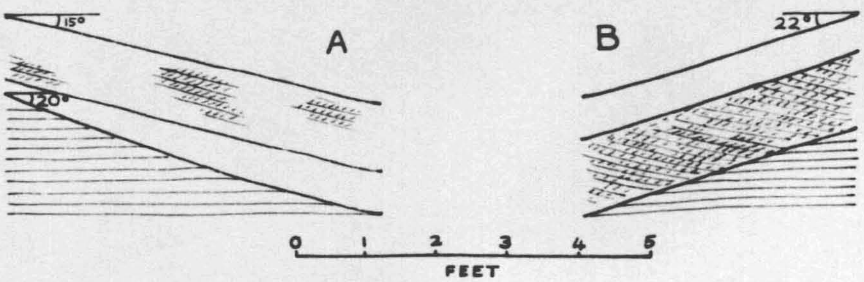


FIG. 1.44 DETAIL OF FACES (A) AND (B) IN FIG. 1.43

their dips are perpendicular to one another, indicating that rapid closure must take place near the present outcrop. On one limb the base of the channel-fill deposits consists of massive sandstone with grooves indicating a current running parallel to the strike of the limb. This is succeeded by a unit showing ripple drift bedding which is of roughly uniform thickness throughout. The ripple drift indicates deposition from a current which affected both limbs similarly, with the result that on one limb the ripples were migrating down the slope and on the other limb, along the slope. The succeeding beds comprise flat-bedded sandy silts. A trough appears to have been eroded and a thin lag sand deposited. Later currents flowing diagonally across the trough introduced sand which formed climbing ripples. The direction and nature of the sand transport must be responsible for the uniform thickness of the unit. Subsequent deposition appears to have been of the vertical accretion type, and consists of a fining upward sequence.

H. CHANNEL ASSOCIATIONS

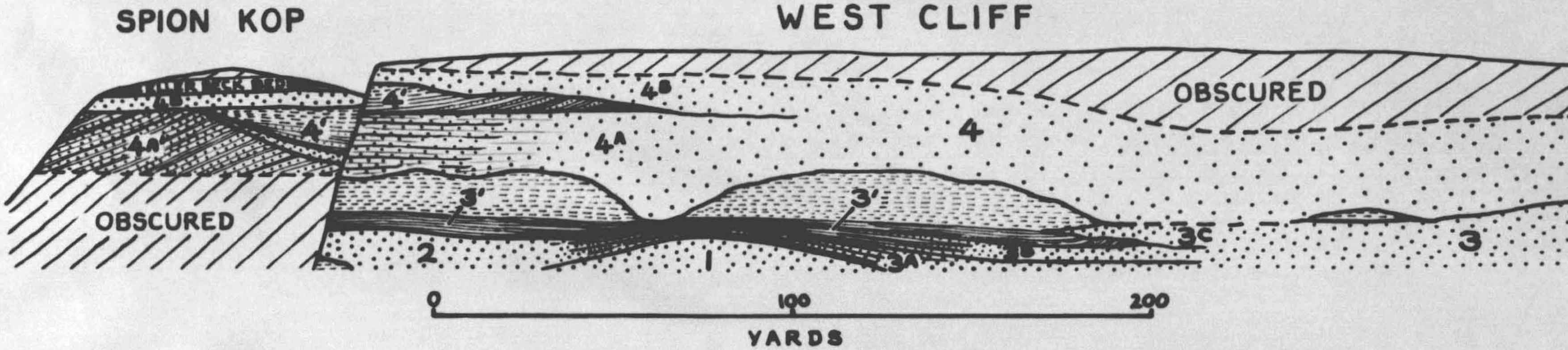
1. WEST CLIFF, WHITBY

The section in the West Cliff, Whitby illustrates the association of the various channel facies and at the same time illustrates the mutual relationships of several channel bodies which may represent continuous distributary sedimentation. No other coastal section shows such a complete development of sandstone throughout the vertical sequence and it is therefore believed to mark the site of an important distributary.

The history of the channel development may be determined from the nature of the channel contacts. The sandstone bodies have been numbered in their probable order of formation (fig.1.45), and each number is also applied to the respective lateral facies sediments.

The Dogger is exposed on the foreshore only at the lowest tides and when the beach is at a low level, and it is not known whether sandstone is developed directly on top of the marine beds or not. The approximate level of the Dogger has been included in the diagram. The earliest sandstone (1) consists of large-scale cross-bedded units; the base is not seen but the top is sharply defined and strongly convex. By comparison with other sandstone bodies this convexity may indicate channelling into fine-grained sediments rather than deposition direct onto Dogger sediments. A small exposure of shale is preserved to the east. Some degree of compaction of the sediments below the channel sandstone appears to have occurred before the next phase of sandstone deposition since the overlying beds show pronounced thinning towards the axis of sandstone.(1).

Sandstone (2) consists of massive and large-scale cross-bedded sandstones in the lower half but in the upper half consists of thick units of fine-grained sandstone with climbing ripples. The ripple-bedded sandstones extend westwards and show considerable thinning over sandstone (1). These lateral beds are in turn overlain by thin-bedded ripple-marked sandstone with considerable erosion at the contact. The thin-bedded sandstone is overlain by siltstone with sandy ribs in the lower part. These beds represent the lateral deposits of sandstone (3). The thin-bedded sandstones thicken



1-4 : TIME SEQUENCE OF CHANNEL DEVELOPMENT
A,B,C : PHASES WITHIN AN ENTIRE ACTIVE PERIOD
3',4' : LATERAL FACIES OF RESPECTIVE SANDSTONES

FIG. 1.45 SKETCH OF FACIES CHANGE DISPLAYED IN WEST CLIFF AND SPION KOP (WHITBY) rapidly away from the axis of sandstone (1) and pass into cross-bedded sandstone (3a). The lower part of the overlying silstone develops sandstone ribs which thicken to the west and also pass into thick-bedded sandstone (3b). This last sandstone is overlain by a dipping laminated siltstone and sandstone levee deposit which passes eastwards into siltstones and this in turn is overlain, with some erosion, by thick-bedded sandstone (3c). Sandstones (2) and (3) and their lateral deposits dip away from the axis of channel (1), and the dip increases upwards indicating that compaction beneath this channel continued throughout.

The sandstones described so far seem to form a closely associated group and it is interesting to observe the control of sandstone (1) on the location of sandstones (2) and (3). Sandstone (3) consists of three visible phases; the first two represent westward migration while the last represents a minor eastward migration, preceded by the development of a small levee deposit. The siltstone and shale deposited after sandstone (3) appears to be less localised than the previous beds, and culminates in the development of a coal seam. It is not certain whether this represents a complete cessation of distributary sedimentation in this area or lateral migration of the channel. The thick development of sandstone (3) in the extreme west may indicate migration in this direction with a coal seam being deposited locally to the east.

Sandstone (4) is a widespread poorly-bedded sandstone with a strongly erosive base. The time interval between this and the underlying beds may be judged to some extent by the lack of relation-

ship

ship of the irregularities at the base to the underlying sandstones. The equivalent of this sandstone on Spion Kop appears to be the dipping laminated sandstone unit which itself develops sandstone lenses in the lowest part. The thickness of this unit relative to any other exposed is indicative of the amount of overbank deposition that must have been associated with this particular channel. A final sandstone phase (5) appears to have transgressed over this dipping unit with some erosion at the base. Before deposition of sandstone (5) a channel, twenty feet deep at the axis, was eroded into the laminated beds: this appears to have run almost parallel to the sandstone channel (4). This channel has been filled by vertical sedimentation with sandstone at the base and siltstones and shales above. Sandstone (5) is succeeded by the Eller Beck Bed, sometimes with a thin shale bed between.

In general these beds are characterised by an unusual amount of overbank sedimentation. Their lateral extent cannot be determined but massive sandstone occurs a few hundred yards to the west, and some or even a major part of the distributary sandstone may be obscured in the west by slipping boulder clay.

2. THE SCAR? WHITBY

This sandstone association is apparently unique, and is described because of the several problems which are associated with it. The section is represented in figure 1.46. It comprises a series of thick sandstones with bedding planes dipping to the south-east. Sedimentation appears to have proceeded from the north-west

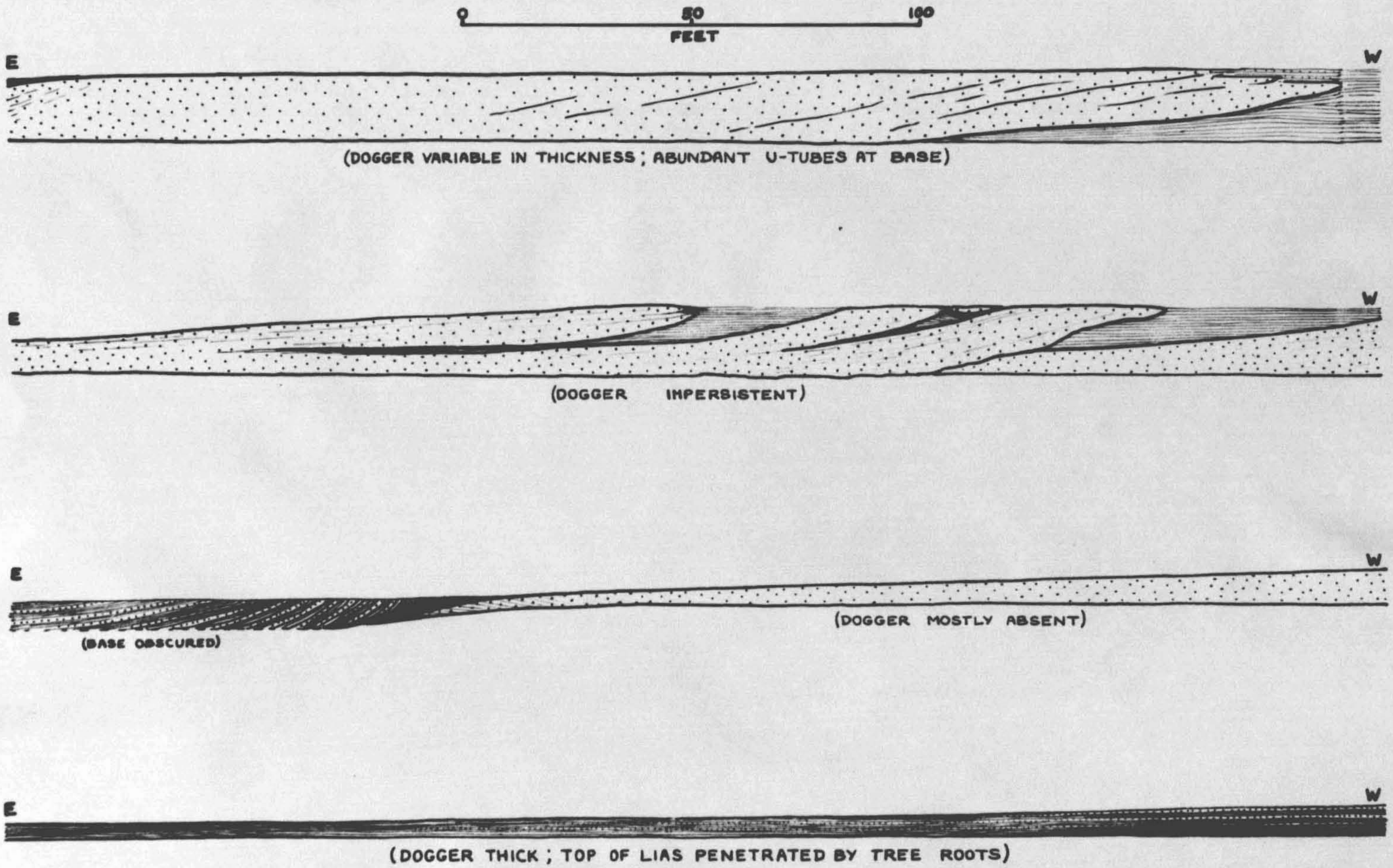


FIG. 1.46 SECTION OF SANDSTONES AND LATERAL PASSAGE BEDS ; THE SCAR, EAST OF WHITBY

to the south-east as may be seen from some of the erosional contacts. Two main phases may be distinguished, the first of which constitutes the greater part of the sandstone mass. The early sandstone does not appear to be strongly erosional at its north-western end and the bedding in the underlying silts is parallel to the base of the sandstone. The upper part of the sandstone at this end passes laterally into silty sandstones. The main sandstone body is made up of a series of units of varying thickness and definition, all of which dip gently to the south-east. The upper surface at the south-eastern end of the sand body shows a similar dip and is overlain by horizontally bedded laminated sandy siltstones.

The shales which pass under the first sandstone at the north-western end thin out slowly sothwards and finally the sandstone rests directly on the Dogger. The thickness of the Dogger varies considerably even beyond the limit of the sandstone, and at the base of the sandstone a thin pebble bed is locally developed, containing phosphatic pebbles derived from the Dogger.

The siltstones and the lowest part of the sandstone are strongly eroded by the sandstone of the second phase. From this point southwards the first phase is entirely absent.

The sandstone rests directly on the Dogger and to the south-east cuts through to the Alum Shale. At the north-western end siltstone fragments occur within the sandstone at the base of the steep erosion surface. On this surface an ironstone nodule, which formed part of a band of nodules within the siltstones, was seen to project into the sandstone, indicating that nodule formation preceded the erosion. The second sandstone is less homogeneous than the first, and contains one horizontally bedded siltstone intercalation in the early part and several thin intercalations in the later part whose bedding parallels that of the sandstone beds. Silty laminae become increasingly common in the later sandstones, especially at the top, and the bedding becomes thinner. The first sandstone of the second phase is very massive weathering but close examination reveals that a fine lamination is present throughout whose dip is almost parallel to that of the upper and lower surfaces of the sandy body. This resembles the low-angle cross-lamination characteristic of small-scale channels.

A pebble bed of varying thickness is developed at the base of the second phase sandstones. It appears to be associated with the base of the finely laminated sandstone and it is not eroded by the later beds. The pebble bed and the base of the sandstone are tough and possess a calcareous matrix. In the extreme south-east the sandstone shows low-angle cross-bedding dipping to the north-west. The coarse sandstone and underlying pebble bed can be traced further to the south-east than any other of the sandstone units. The overlying sandstones appear to thin to the south-east; this seems to be a primary feature though to some extent accentuated by erosion.

At the south-eastern extremity of the coarse sandstone a well defined thrust plane is developed (fig. 1.47). This brings Dogger sandstone and crushed Lias shale on top of the sandstone. The top of the Dogger and shale slice are eroded and a thin band of distorted pebbles of sideritic sandstone lies on the eroded shale surface.

Since the Dogger is absent beneath the pebble bed the thrust must have had a horizontal movement of at least ten feet, and as the Dogger in the slice is a foot thick the movement may have been considerably greater. The Dogger - Lias contact is itself thrusted and the bottom end of the Dogger rests on a Lias nodule. The Dogger is overlain by a thin development of the pebbly sandstone followed by laminated sandstones with thin partings of silty shale in which the bedding is nearly vertical. These laminated beds reach a maximum of seven feet thick, and show a gradual decrease in dip to the south-east. This dipping unit has all of the characters of the supposed levee deposits described earlier, and the near vertical attitude of the beds at the western end are presumably due to oversteepening by the thrust. In the south-east the laminated beds become finer grained and rest without erosion on 1ft.4in. of Dogger. The beds dip towards the thrust zone on both sides, but the Dogger on the south-eastern side is at a level several feet below that on the north-western side.

Several features distinguish this sandstone association from all others observed in the Lower Deltaic series:

(i) Sandstone deposition developed at an earlier stage than observed elsewhere.

(ii) Sandstone developed over an area in which the Dogger underwent considerable erosion which was not associated with channelling.

(iii) Normal cross-bedding types are absent; all sandstone units have been deposited by lateral accretion proceeding from north-west to south-east, with the exception of the lowest pebbly sandstone.

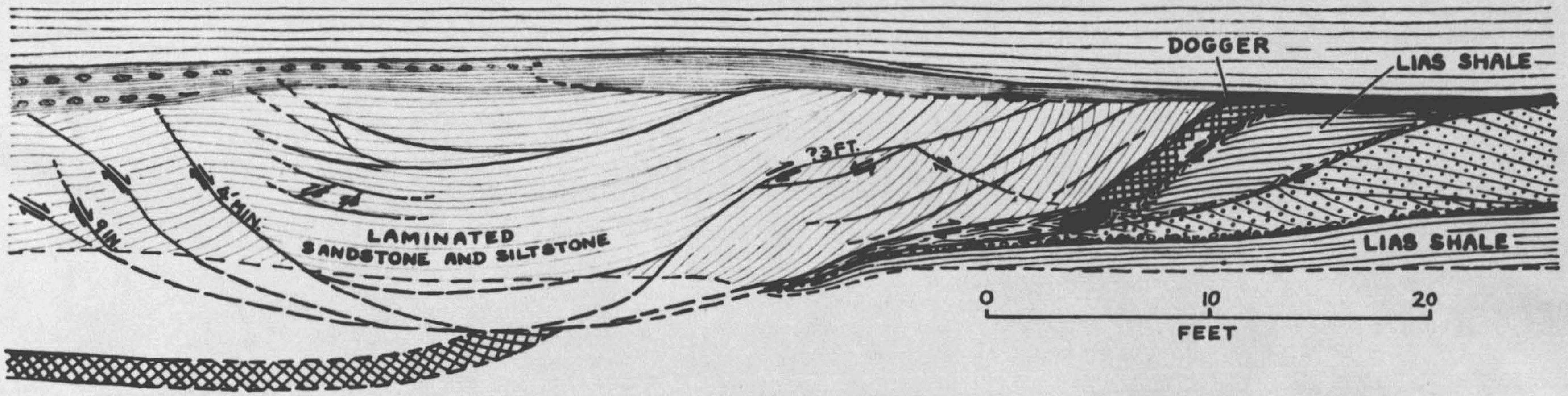


FIG. 1.47 DETAIL OF THRUST IN FIG. 1.46, SHOWING ROTATIONAL FAULTS AND THEIR PROBABLE EXTENSION BELOW BEACH LEVEL

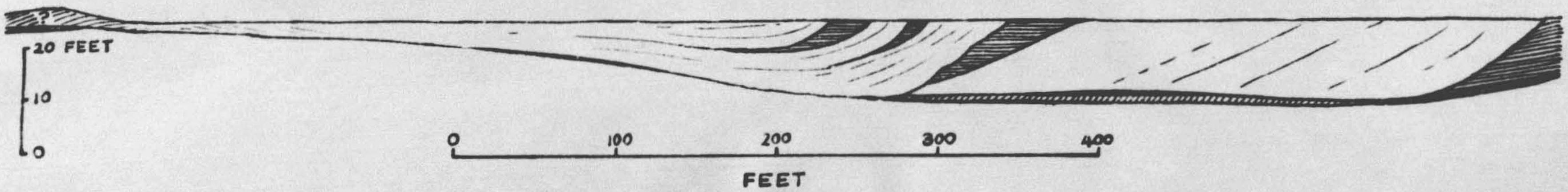


FIG. 1.48 THE SCAR SECTION RE-DRAWN WITH HORIZONTAL TOP (VERTICAL EXAGGERATION)

(iv) Long periods of vertical accretion of fine-grained sediments separated some sandstone units

(v) Each successive unit represents the subaqueous bank of the depositing stream, but there is no trace of an opposite bank until the development of levee deposits at an apparently late stage.

(vi) The development of the thrusts after deposition of the sandstone but before deposition of the overlying coaly shale.

(vii) A possible association with the present axis of the Whitby syncline.

All of these points appear to be significant with regard to the reason for the development and peristence of sandstone deposition at this particular point. Item (v) suggests the presence of some sort of trough which acted as a channel course while to either side little or no deposition was taking place. The levee unit and its lateral equivalents to the south-east rest directly on the Dogger, but in the thrust zone a thin representative of the pebble bed is seen on top of the Dogger. The first sandstone phase at least is not represented and yet reaches fifteen feet in thickness two hundred feet away; the pre-sandstone shales are similarly not represented. This trough could not have been produced by erosion since the Dogger is thinly developed in parts and the chocolate mudstone nodules horizon at the top of the lias is usually present. Local downwarping with an axis a short distance to the north-west of the present synclinal axis seems more probable. In support of this the abundance of U-tubes penetrating the top of the Lias beneath the sandstone units indicates some localised control on sedimentation and ecology even in pre-Dogger times. U-tubes are not seen elsewhere on the coast at the base of the Dogger.

This trough could have acted as a focus for influxes of fresh-water causing considerable erosion of the Dogger, and followed by deposition of non-marine muds. Subsequent phases of increased water flow seem to have favoured a similar cou e and this has usually been accompanied by erosion of the muds followed by sand transport and deposition with intermittent quieter phases. The development of levees would presumably occur only when sedimentation began to overtop the trough margins and apart from slight modification through thrusting and erosion the present sandstone body must reflect the shape of the original trough, which may be reconstructed as in figure 1.48 . The final phase of water flow occurred before quiet water deposition of coaly shale and must have been responsible for erosion following the thrusting.

Sand appears to have been deposited by lateral accretion throughout, but it is not clear why deposition should always have been on the same side. This may be related to the cross-sectional shape of the trough, of sinuosity of the trough (possibly resembling that of rivers), or of continued subsidence with migration to the south-east. No entirely satisfactory explanation can be offered here.

The development of the thrust is clearly related to the tectonic and sedimentary conditions that preceded it. A foreshortening of e perhaps twenty feet is involved which may have been produced by renewed tectonic movement or by superficial movement of the non-marine sediments. Once again conclusive evidence for any one mechanism cannot be found. A factor that may be important is that the present axis of the Whitby syncline coincides with the estimated lower limit of the thrust flexure. If the syncline were produced by compression thrusting could possibly have been developed in the superficial sediments, but it could also have developed as a result of gravitational sliding of the sediments down the limbs of the syncline.

Internal faulting is very evident in the laminated sandstone unit and from the nature of these faults it would appear that the movement of the unit has to a large extent been rotational, resulting in the detatchment of a portion of Dogger and Lias and its being carried up to a higher level and suffering subsequent erosion. Most of the lateral relative movement involved appears to have taken place on this side. There is, however, evidence of movement on the other limb; at the base of the siderite mudstone nodule layer at the top of the Lias one or two planes of brecciation are developed. There is evidence of movement along these planes in that the U-tubes which descend from the base of the Dogger are truncated by the planes and in some cases dislocation of the U-tubes indicates a south-eastward movement of the overlying beds. This movement has probably been accompanied by incompetent movement of the shale and may be responsible for the uplift of the downthrust side of the fault.

In conclusion the sandstone developed at this locality may owe its origin and many of its unusual features to minor earth movements of a very localised nature which appear to have been closely associated with what is now the axis of the Whitby syncline. All movement appears to have ceased before the fine-grained coaly beds were laid down on the sandstone unit.

II INTERDISTRIBUTARY FACIES

Although many different lithological types may be distinguished in the field, lateral and vertical gradation of one type into another is so prevalent that it is impossible to break the series down into well-defined categories. In the following descriptions the more distinct lithological types have been considered separately and the remainder considered as a single, highly variable group.

A. SHALES, SILTSTONES AND ASSOCIATED SANDY HORIZONS

The most uniform and recurrent lithological type consists of pale to dark grey tough clay siltstones. Bedding is either poor or absent and the rock usually weathers to small irregular fragments. The constituent grains range from fine sand to clay, forming a more or less unsorted association. Small mica flakes, showing only slight bedding orientation are sometimes common and finely divided carbonised plant material is usually present. Rootlets, preserved as vertical carbonaceous streaks occur in many beds and are locally abundant. Since rootlets may be absent, however, it is clear that they are not responsible for the lack of bedding. This would seem to be a primary feature and not produced by vegetation or burrowing activity since in the coarser siltstones fine sandy laminae are sometimes present and these show no disturbance of bedding. Colour variation is common within the siltstone units, the coarsest-grained beds being light grey and the finest dark grey or nearly black. The variation follows the bedding and although inversion or repetition is often observed the usual upward sequence is from coarse pale-grey siltstone to fine dark siltstone. Macroscopic plant material is more common in the darker siltstones but the colour appears to be due to much more finely divided material, presumably of organic origin. The top of the dark siltstones is usually carbonaceous, and thin impersistent bands of bright coal may be present. This ideal upward passage is not always complete; the coaly siltstones often have a sharply defined base which in some cases is seen to be erosional. Rootlets sometimes occur in the beds beneath the coal, but in most cases they are absent.

A second type of upward passage is from grey siltstone to very pale unbedded sandy siltstone which commonly develops small, well-separated lenses of less silty white sand. Rootlets are usually abundant and show sharp truncation at the top of the unit. A coaly shale often overlies these units and may represent local plant growth though evidence of redistribution is usually present. The sand lenses may show a little channelling.

The base of the grey silt units sometimes passes down into alternating siltstone and fine-grained sandstone of a lithology very similar to that of the levee deposits described earlier. These beds show very delicate lamination which in the sandy layers often shows small-scale convolution (fig.1.47). The base of these units is sharp and sometimes mildly erosive. Small channels may occur within the unit and the channel fill is of a similar laminated lithology. In one such channel the top is filled with a grey shale in which a thin bed crowded with plant material is present. The plants comprise entire leaves and fronds which are in a mummified state, having retained a considerable amount of elasticity. In the cliffs east of Hawsker a thick unit is present a short distance above the Dogger. This consists of thin bedded laminated sandstone which may possibly be occupying a channel. It is of particular interest in that it contains horizons with abundant lamellibranch resting burrows and a few bedding planes crowded with Unio. The shells are preserved in siderite mudstone when articulated or as a ferruginous film on a sand cast when disarticulated.

The usual order of thickness between sharp junctions in the fine-grained sedimentary series is from 2 to 4 feet. Some siltstone units may reach 15 feet, especially in sections in which large channel sandstones are absent. The sandy developments are of a local nature and very variable in character. The siltstones, except where they develop sandy horizons are more uniform, but many instances of discontinuity and erosion within the siltstone have been observed.

B. THIN SHEET SANDSTONES

These form prominent features in the siltstone series by virtue of their resistance to weathering and their continuity. They are characteristically fine-grained and well sorted and they usually occur in units of up to one foot in thickness which are separated by thin intercalations of shale and in some cases of siderite mudstone. Internally they may be structureless but commonly display small-scale planar cross-bedding. Erosion of underlying beds is rare, but isolated load casts may occur and are usually on a large scale.

These beds occur at all levels and are traceable for several hundreds of yards in most cases. They die out laterally through thinning and do not show passage horizontally or vertically into the fine-grained beds except by intercalation.

One of these horizons is of particular interest since it is the bed from which Unio kendalli and dinosaur footprints were obtained (p. 4). In the present study several other features have been noted. Firstly the sandstone rests on a grey laminated shale which is of a lithology very unlike that of the normal fine-grained sediments. At the base of the sandstone and sometimes in the middle a thin sideritic siltstone is often developed. Siltstone partings may occur and the unit may comprise one, two, or three beds of sandstone. The maximum thickness is three feet; the minimum is ten inches. Climbing ripples are characteristic, but the basal part may be structureless. The bed is usually 20 to 25 feet above the Dogger.

At Saltwick the bivalve Unio kendalli occurs on the lower surface of the ironstone layer which typically weathers reddish and shows strong polygonal jointing on a small scale. A bennetitalean fructification was also observed. In addition, structures resembling prod marks occur within the ironstone; these consist of narrow vertical sandy sheets which are continuous with the overlying sandstone, (fig.1.50). They show a blunt and a tapering end in plan view and they form a sharp ridge on the lower surface of the ironstone. Their origin is not clear, but they may represent lamellibranch resting burrows which, owing to the unconsolidated

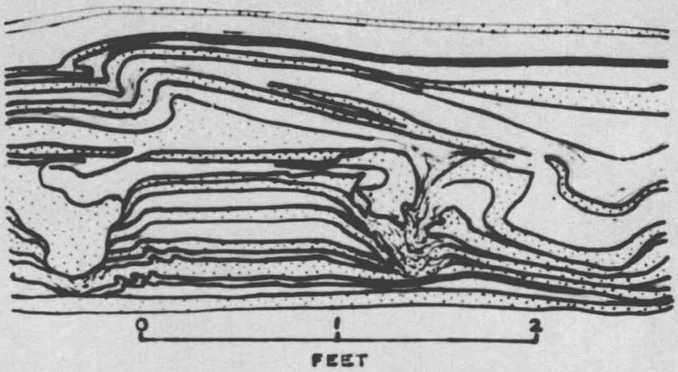


FIG. 1.49 CONVOLUTION IN LAMINATED SILTSTONE/SANDSTONE UNIT

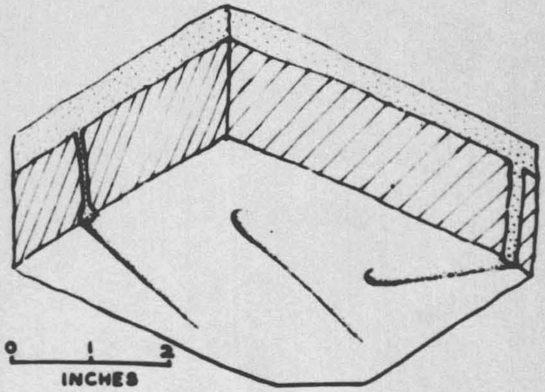


FIG. 1.50 ?LATERALLY COMPRESSED CASTS OF RESTING BURROWS IN IRONSTONE AT BASE OF UNIO BED

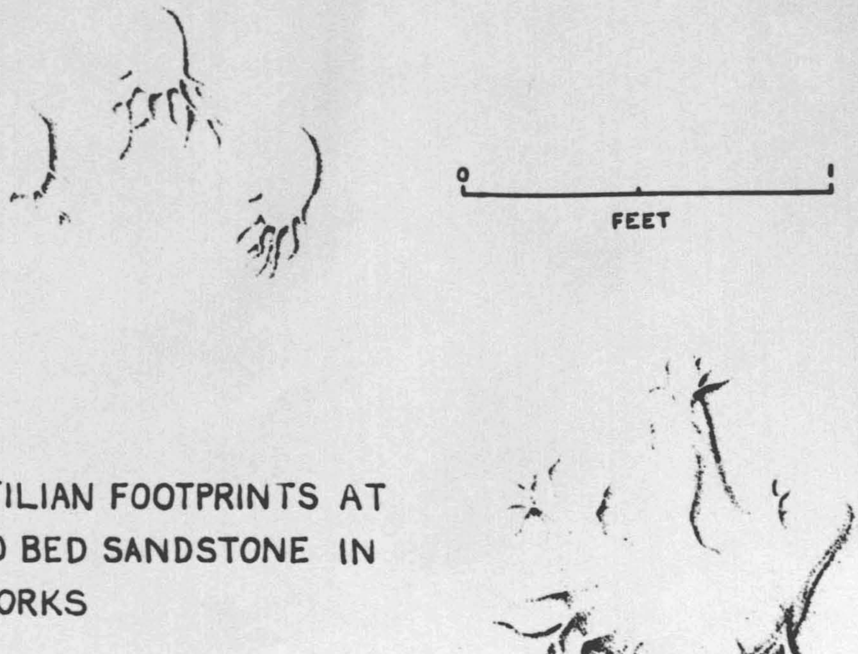


FIG. 1.51 REPTILIAN FOOTPRINTS AT BASE OF UNIO BED SANDSTONE IN PEAK ALUM WORKS

nature of the sediment, have become compressed on evacuation. They are rather deep relative to their length and breadth to be prod-marks.

This horizon has been traced, with variation in thickness and interbedding, to the east and west of Saltwick. It is traceable to the west as far as the East Cliff. To the south-east it is more or less continuous, apart from occasional erosion by channels, until Hawsker. Specimens of Unio are found in places on the basal surface and sometimes on the lower surface of the upper unit where a silty parting is developed. Ironstone is often not developed and the sharp groove structures penetrate the laminated shale, but are not as deep as those in the ironstone. This may result from subsequent compaction. In the Peak Alum Works a sandstone occurs on a similar horizon. Ironstone is not present and the sandstone is impersistent and locally overlain by a silty variable sandstone. Sharp grooves are present at the base but no lamellibranchs have been found. Dinosaur footprints are present (see Fig.1.51) and the small drainage runnels indicate that the mud must have been moist but not under water. Shallow suncracks have also been seen. The bed has not been identified south of the Peak Fault.

Ripple marks at Saltwick indicate flow towards south-west and the shap grooves show corresponding alignment with their blunt ends upcurrent. This may feflect the orientation of the shells in their resting burrows. The bed is referred to as the Unio Bed in the subsequent text.

C. NON-SANDY CHANNEL FILLS

Examples of erosion surfaces within siltstone series have been mentioned on page 52 . In the example described on page 27 the fill is of levee-type sediments and a similar feature is seen in the Hawsker Cliffs (fig.1.52). The steeply dipping unit becomes more sandy to the south-east and is associated with a large channel sandstone. The small plant-bearing channel described on page is shale-filled in the upper part.

The most striking example of a siltstone and fine-grained sandstone channel fill is seen in Spion Kop, Whitby (fig.1.53). Channelling has occurred to a depth of twenty feet into steeply dipping laminated sandstone. The initial fill consists of sandstone with bedding sub-parallel to channel walls and in which each bed thickens towards the axis. The sand has built up to a fairly level surface on which siltstones with thin sandy ribs have been laid down. The siltstones pass upwards into dark silty shales. The channel runs along the strike of the dipping unit and is parallel with the axis of the main sandstone body to the west. The channel shows a great contrast in shape to the sandstone-filled channels in that the top is flat and the erosional surface strongly convex. This is ascribed to the absence of the effects of relative compaction that are seen where sandstone bodies occur among shales.

D. COAL SEAMS

Coal seams in these beds are thin and impersistent and coaly material usually occurs as thin bands in dark carbonaceous shale.

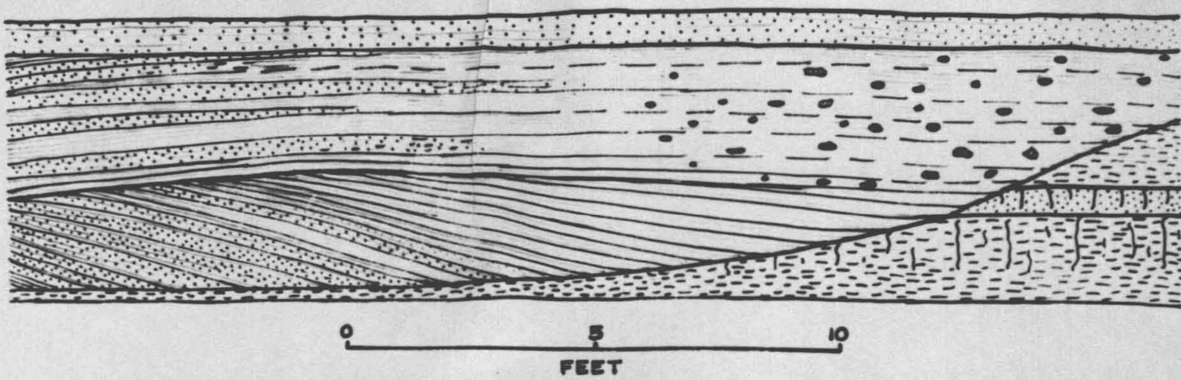


FIG. 1.52 EROSION FOLLOWED BY LEVEE SEDIMENTATION (HAWSKER)

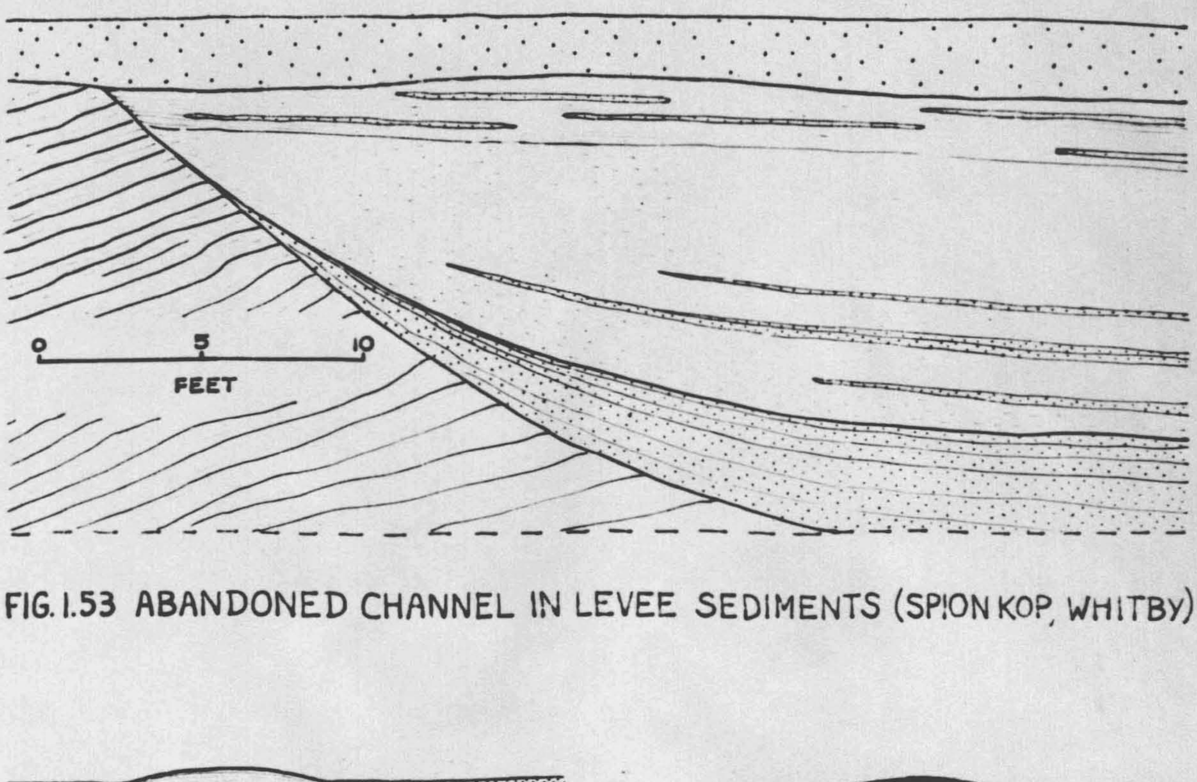


FIG. 1.53 ABANDONED CHANNEL IN LEVEE SEDIMENTS (SPION KOP, WHITBY)

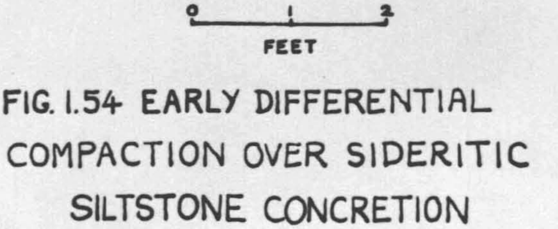


FIG. 1.54 EARLY DIFFERENTIAL COMPACTION OVER SIDERITIC SILTSTONE CONCRETION

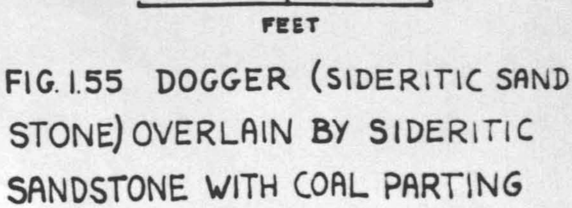


FIG. 1.55 DOGGER (SIDERITIC SAND-STONE) OVERLAIN BY SIDERITIC SANDSTONE WITH COAL PARTING

These shales often have a sharp base and do not appear to be associated with underlying rootlet beds. One of the most persistent coaly horizons is found in the cliffs between Saltwick Bay and Hawsker where about three feet of silty shale with one or two thin coal seams lie immediately on Dogger sandstone. Coal of more limited extent is often associated with the top of the thin beds which occur between a channel sandstone and the levee deposits, as in the Scar section of Whitby.

E. PLANT BEDS

Plant accumulations occur in several types of environment, but in most cases there is little evidence for growth in situ. The plant-beds have not been studied in detail as it is clear that a thorough knowledge of both the macroscopic and microscopic properties of the plants involved would be necessary in order to evaluate the growth conditions of the various floras. Sedimentologically the plant beds consist of the following types:

(i) Laterally persistent horizons, generally in shale or silty shale, in which small fronds are preserved on many bedding planes. The plants are probably locally derived and deposited under quiet conditions (e.g. Hayburn Wyke plant bed).

(ii) Small sandstone channels containing well preserved large fronds at their base. These may be preserved as carbonaceous material or less commonly in siderite mudstone, in which case the whole mass is probably derived.

(iii) Larger channels containing abundant logs in association with local erosional hollows. Plants may occur in flat-bedded sandstone in the lateral parts; in one instance fronds of _Baeiria_ occurred in moderate concentrations in the troughs of ripple-marks.

(iv) Small plant fragments associated with the fine-grained laminae of levee deposits, (e.g. Whitby plant bed).

(v) Very localised small-scale channels with fine-grained sediment fill; preservation may be extremely good with little or no carbonation, (see p. 52).

(vi) The only undoubted evidence of in situ plant growth is afforded by beds containing vertical _Equisetum_. Preservation is in the form of cylindrical sandstone moulds with a carbonaceous imprint of the plant on the outer surface. Rootlet beds occur beneath the stem-bearing bed. Several examples of _Equisetum_ beds are known and in some cases they appear to be associated with distal levee sediments. The lithology of these beds is however somewhat peculiar. The stems are typically preserved in fine grained sandstone with delicate lamination produced by thin films of micaceous silty shale. Thin shales of coal may be present. A typical sequence is given below:

	ft.	in.
Laminated sandstone with vertical Equisetum stems and associated rootlets; recumbent stems at base	3	3
Black coaly shale with compressed Equisetum		3
Grey sandy siltstone	1	2
Dark silty shale with sandy rootlet casts passing down into blacky grey siltstone	3	0

This probably represents more than one phase of growth; the plants of the earlier phase rotted and collapsed and may have been transported whereas the later plants were preserved through inundation by sand.

A different mode of preservation is found in which the stems, or at least their lowest parts, are preserved in coaly shale. The upper parts of the stems which escaped burial were apparently removed by currehts which intpoduced the sand in which the base of the stems are cast. Compaction of the shale has led to crumpling of the stem casts. A section of this type of Equisetum bed is given below:

	ft.	in.
Grey unlaminated clay siltstone		
Dark carbonaceous shale, coaly at top, with sand-filled stems		9
Tough white sandstone with rootlets	1	0
Dark carbonaceous shale, sand laminae in lower part; sand-filled stems	1	0
Tough white laminated sandstone with rootlets.	1	3
Grey siltstone	1	6
Dark silty shale, coaly at top and at base, with compressed stems.	2	6
Tough white sandstone with rootlets	1	6

F. IRONSTONE

Ironstone occurs in three prinicipal forms comprising sideritic sandstone, siderite mudstone, and sphaerosiderite.

1. Sideritic sandstone

Some of the very thin sandstone horizons have a pinkish or brownish colour which on close examination i s seen to be due to the presence of abundant small siderite crystals. A similar feature is seen in sandy lenses developed within silty horizons suggesting selective sideritisation of the coarser grained rock. The ironstone bands developed in the Unio bed are of similar type, but are extemely sideritic; this appears to be a reflection of the finer grain of the original sediment which may have been more prone to replacement.

Nodules of a silty or sandy nature are developed in siltstones and sandy siltstones. These usually have asmooth regular outline and show some internal lamination! Relative compaction of the surrounding sediments is not great unless the replaced material constitutes a local coarse-grained lens. This situation is seen in figure 1.54 and has resulted in small-scale faulting which appears to have preceded deposition of the overlying beds.

1. Siderite Mudstone

Siderite mudstone nodules are usually dark grey in colour with a reddish weathered zone. The outline is commonly irregular and shows little relationship to bedding, unlike the sandy nodules described above. Another distinguishing feature is that the sandy concretions commonly contain rootlets when they are present in the surrounding siltstones, while the siderite mudstone nodules do not. The nodules are usually small, and seldom exceed three inches in diameter. The distribution of the nodules rarely follows any general rule, but they are usually associated with silty shales. The distribution is related to bedding in a general way; in that a whole bed or part of a bed is characterised by scattered nodules showing no preference for any particular bedding plane. In a few cases large isolated nodules occur along a particular horizon. Abundant nodules are generally found in the silty intercalations that occur in the channel-top thin-bedded sandstones and in the thin-bedded lateral facies.

3. Sphaerosiderite

Sphaerosiderite is a common, though minor constituent of the shales. Three varieties may be distinguished:

(i) Large discrete concretions of spherulitic siderite are occasionally developed on a particular horizon. The concretions are roughly equidimensional and reach one foot in diameter; they may be separated by distances up to ten feet. The outer boundary is sharp and the surrounding shales show no spherulitic bodies. Rootlets may be present as in all forms of sphaerosiderite. Compactional features are strongly developed around the concretions, usually with the development of slickensiding.

(ii) Siderite spherulites occur disseminated throughout some poorly bedded shale horizons. Separation of individual spherulites may be considerable but in other cases they coalesce to form loose clayey aggregates. Such aggregates occur in the middle of shale beds, and their margins are always diffuse. The aggregates are usually persistent for a few feet laterally, and occur at intervals along a particular horizon.

In one or two cases there is strong evidence for spherulites being washed out of such horizons and subsequently being redeposited Thin sandy sphaerosiderites which lack a clay matrix and which overlie sphaerosideritic shales appear to have been formed in this way. The spherulites in such beds are partially oxidised and do not show interlocking relationships.

(iii) Occasionally sphaerosiderite occurs in more continuous beds which may reach two feet in thickness. The spherulites are set in a white clay matrix and the whole bed is distinct from those above and below. Such beds have been observed at Iron Scar and in Winter Gill, and it may be significant that in both cases they occur at or near the top of the non-marine beds.

G. INTERDISTRIBUTARY SEDIMENTATION

Studies on interdistributary sediments of recent rivers and deltas (Fisk, 1944, 1947, 1952; J.R.L.Allen, 1965) indicate that the predominant sediment type is fine grained and poorly laminated or unlaminated. The sediment is deposited from suspension during floods; the water is both transported into the area and drained off along a network of shallow channels, some of which may follow old distributary courses (Fisk,1947 p.43). The annual increment is very small judging by data obtained from recent floods (Jahns, 1947) in which an average of less than 0.5 inches are deposited in each flood away from the levees. These figures apply to alluvial flood plains as opposed to a true delta environment.

The bulk of the shales and siltstones described above compare closely to the descriptions of the modern sediments, while the coarser laminated beds probably represent the distal parts of levees.

The thin sheet sandstones appear to represent deposition in some sort of lake. The laminated shale which underlies the Unio Bed indicates that an area in which one or more large distributaries were normally present was receiving only the finest sediment, which unlike the fine sediment previously deposited showed delicate lamination. This seems to indicate slow deposition in an area of widespread shallow water. Colonisation of the muds by freshwater bivalves was widespread but with increasingly shallow water conditions these must have died out. Finally the surface muds became sufficiently dried out to support the weight of small reptiles, whose footprints show that the mud was moist and cohesive. Occasional sun-cracks are evidence of the dessication of the mud flat surface. The overlying sandstone indicates a renewed influx of water but this time with currents capable of transporting fine sand. Local reversion to muddy bottom conditions are indicated, together with renewed colonisation by lamellibranchs.

The Unio Bed and similar sandstones seem to represent periods in which sedimentation has been reduced to a minimum, with only the finest muds being deposited, followed by a period of reintroduction of coarser sediments. One other horizon bearing foot-prints has been observed in the uppermost few inches of the channel top sediments in a section in Danbydale; sun-cracks are also present. Inferences that may be drawn as to the physiography of the delta surface during these occasional incursions by reptiles are considered later.

Small-scale sandstone channels and abandoned channels which are associated with the more regularly bedded fine-grained deposits probably represent the channel network present in modern backswamp areas. The large channel shown in figure 1.53 probably represents a flood-water distributary cutting through the earlier levee deposits and may therefore represent a crevasse developed while the river was at high flood level and subsequently filled by normal levee and flood-plain deposits. The small sandstone channel fills may represent small sediment distributaries within the interdistributary areas, which functioned only during times of flood; alternatively they could represent winnowing of coarser detritus during drainage of floods. The discordant nature of many of the siltstone and shale contacts may represent erosion during the height of flood activity, followed by normal deposition.

Although rootlet-bearing horizons are common a large number of normal siltstone and shale horizons show no signs of plant growth in place. Similarly many of the sandstone horizons are devoid of rootlets. Rootlets are best developed in upward coarsening siltstone sequences which generally culminate in fine-grained silty sandstone horizons, often with minor channelling. Equisetum beds in which the stems are preserved in an upright position are associated with sediment types not found in other associations.

The lack of lamination in many silty beds may be a primary feature or it could result from bioturbation due to small animals or even lowly plants such as the mosses. The records of unlaminated beds in modern deltas come from heavily vegetated regions and so it is not clear whether it is a primary feature of not. The absence of rootlets on many horizons in the Yorkshire delta does, however, indicatethat conditions were adverse to the development of dense vegetation. Two factors could account for this: rapid sedimentation and sediment redistribution or the presence of large submerged areas in which only specialised plant types could flourish. Evidence obtained from modern delta environments indicates that sedimentation is very slow in the interdistributary areas and is therefore very unlikely to smother plant growth.

The presence of a water cover over parts of the interdistributary areas is more plausible, and the preservation of Equisetum in growth position favours the presence of such conditions. Sub-aerial conditions are indicated by rootlet beds from which the associated plants have been removed and, since these are associated with relatively coarse sediment, deposition in such cases may have been sufficiently rapid as to bring the sediment surface near to or above water level.

The presence of fronds of plants which are believed to have had a land-dwelling habitat indicates clearly that subaerial conditions prevailed in part and the subaqueous environment was probably restricted to those areas furthest from active distributaries.

Two principle types of cycle may be detected. The upward coarsening cycle which ends in the development of silty sandstones has been suggested above as indicated increasing sedimentation leading to the development of plant-colonised surfaces. The second type consists of an upward decrease in grain size coupled with an increase in the amount of carbonaceous matter and sometimes culminating in a coal seam.

The common absence of rootlets and the impure nature of the coaly horizons indicate that the consituent plant matter is of drifted origin. A decrease in detrital sediment supply is indicated together with an increased supply of plant material in a finely divided state. The rhythmic nature of the siltstone and shale sedimentation must clearly by related to the distribution and nature of the distributary channels and the correlation between the two will be considered in the next section.

III RELATIONSHIP BETWEEN DISTRIBUTARY AND INTERDISTRIBUTARY FACIES

In the preceding section some conclusions have been reached as to the immediate depositional environment of the fine-grained sediments but their relationships to the associated channel sandstones have not yet been discussed. A guide as to the relative proportion of channel sandstone to interdistributary sediments may be obtained from inspection of the cliffs between Whitby and Hawsker Bottoms ~~(see enclosure)~~. Generally speaking on these coastal exposures channel sandstones constitute 15 to 20 percent and interdistributary sandstone about 15 percent of the total, with fine-grained sediments making up the remaining 65 to 70 percent.

The relatively high proportion of fine-grained sediment can be accounted for in two ways:

(i) The sediments are deposited locally by the associated distributaries which are either long-lived or deposit flood sediment at a high rate. Most of the channels observed do not show signs of having been long-lived since lateral migration occurs only on a small scale and vertical migration is not common. Neither is a rapid rate of sedimentation evident since very few fine-grained deposits are clearly seen to be associated with channels. Compaction may have distorted the field relationships to some extent, but those deposits flanking channel sandstones and which do not show an increase in sand content towards the channel cannot have been deposited by overbank flooding from that channel. The sediments which are certainly the product of overbank deposition associated with adjacent channels constitute less than five percent of the total.

(ii) The sediments are deposited by a few large and persistent distributary channels with only subsidiary amounts being derived from the many short-lived and mainly erosional channels. Since most of the channels are observed to make only small contributions to the fine-grained sedimentary facies it is possible that the latter represent the deposits of relatively widely spaced major distribuaries, which might be expected to be represented by a complex sandstone body extending through the whole veritical sequence. Thick sandstone sequences are developed in Eller Beck, at Mallyan Spout, in Dambydale, Lofthouse Beck and in the West Cliff, Whitby. Unfortunately the lateral facies of these sandstones and their relationship to the fine-grained sediments in surrounding areas are seldom visible. The West Cliff sandstone complex, however, shows the development at more than one level of thick sandstone units which clearly pass laterally into finer sediments. It must be pointed out, however, that even these sandstones appear to have developed after the deposition of deltaic shales and siltstones. Erosion may bring the base down to the Dogger but, with the possible exception of the channel on The Scar, large-scale sandstone deposition does not appear to have taken place in the earliest phase of deltaic sedimentation. It is therefore possible that although the above-mentioned sandstones represent relatively persistent distributaries they may themselves have been subordinate to an even higher order of distributary channel. If this hypothesis is correct then the nature of the lesser distributaries and their relationship to the delta framework as a whole has to be determined. This, together with other conclusions as to the general nature of the delta, is discussed in the next chapter.

CHAPTER II THE DELTA

I. HISTORY OF THE DELTA

A. INITIATION OF DELTA CONDITIONS

South of the Peak Fault the Dogger is overlain by dark silty shales with sand grains. In a section close to the fault this shale is thin, and is overlain by one foot six inches of coarse sandstone current-bedded from the north and penetrated at the top occasional U-tubes. This is followed by more silty shale. To the north of the fault the contact is exposed in several disused quarries where it is extremely variable. In the Stoupe Brow Alum Works the Dogger is represented by a pebble bed of between 7 and 15 inches in thickness which is overlain by a thin grey sandstone with carbonaceous partings. To the south-east, in Peak Alum Works the pebble bed is thin and overlain by 4 ft. 6in. of grey sandstone, penetrated by U-tubes at the top, which contains silty laminae, especially at the base. Further to the south-east the pebble bed thickens and develops into a pebbly sideritic sandstone with U-tubes at the top which is overlain by grey silty sand with occasional U-tubes. The sideritic sandstone reaches a maximum thickness of 4ft. 6in. when it is so crowded with U-tubes as to produce a columnar structure; the overlying grey sandstone shows slight thinning. In this entire section the upper grey sandstone is overlain by black silty shales which are sandy at the base. At one point the sideritic sandstone is overlain by a local thick sandstone sequence from which a poorly preserved

?Gresslya was obtained. The section reads:

	ft.	in.
Yellow sandstone, penetrated by rootlets		10
Sandy siltstone with occasional siderite mudstone ribs	2	9
Sandstone, reddish weathering; lamellibranch 22" from top	6	3
Dark silty shale		5
Sideritic sandstone with abundant U-tubes		

In Howdale sideritic sandstone is overlain by carbonaceous siltstones which are sandy at the base. The Dogger varies greatly in thickness between Howdale and Hawsker Bottoms where the Dogger reappears in the cliffs (Rastall and Hemingway 1940 p.183). In the cliffs east of Hawsker the Dogger varies in thickness from 3 to 6 feet and is overlain by a thin grey sandstone which locally reaches 1ft. 3ins. in thickness where it shows interdigitation with coaly material (fig.1.55). This grey sandstone is overlain by a thin coal followed by shales with another coaly horizon above. Rootlets do not occur in the sandstone. Near Black Nab the following section was recorded:

	ft.	in.
Silty shale with coal at top, occasional small ferruginous pebbles at base	2	3
Irregularly bedded (?pebbly) silty sideritic sandstone; penetrated by rootlets	1	2
Massive sideritic sandstone with occasional silt-filled U-tubes; penetrated by rootlets	3	8

At Saltwick the Dogger is again locally pebbly at the top; elsewhere massive sideritic sandstone is overlain by grey clay with local small-scale channelling. Between Saltwick and Whitby the Dogger is penetrated by large roots filled with carbonaceous silty material and is locally removed by erosion. ~~(see p. ——)~~.

In practically all of the sections described the Dogger appears to show signs of reworking with the deposition of sandy beds of similar appearance to the original Dogger but less sideritic. The occurrence of U-tubes and even a lamellibranch in the Peak Alum Works indicates saline conditions but the silty and carbonaceous laminae and partings reflect the influx of terrestial sediment. In the north roots and rootlets indicate colonisation of the Dogger surface by plants but subsequent erosion, which is not associated with channel sandstones, has removed much of the Dogger in places and at Black Nab Dogger material appears to have been redeposited as soft pebbles on an eroded surface with truncated rootlets. Some fluctuation in conditions is indicated, with the non-marine influence most pronounced in the north.

North of Whitby the Dogger is again variable, comprising essentially two blocks of ferruginous sandstone, the lower of which is commonly fossiliferous. The upper block passes northwards into shales with rows of siderite mudstone nodules and the base of the non-marine beds is difficult to define. Further north a thin sideritic sandstone overlying a pebble bed is overlain directly by non-marine silty shale. The sandstone finally thins out altogether and only the pebble bed remains and this is often incoherent, having a shaly matrix. Although sandstone channels sometimes cut down to near the pebble bed it is clear that they are not responsible for any erosion of the Dogger that might have taken place. The thin pebble bed is said by Rastall and Hemingway (1940 p.195) to have developed on a low submarine ridge.

From Cat Beck northwards non-marine shales rest conformably on the Dogger, but local washouts are responsible for deep erosion of the Dogger as in Boulby Alum Works. In the northern escarpment exposures are poor but in the Guisborough region dark non-marine shales rest directly on marine shales. At Roseberry Topping the age relationship of the _Thinnfeldia_ Plant Bed to the overlying beds is not clear; though it is overlain by typical non-marine shales. A sandstone cuts rapidly down to the south and may be seen resting directly on Alum Shale. On the main escarpment the Dogger is represented by a rotten shelly limestone with a pebble bed at the base and an ironstone at the top; this is similar to the development near Kildale. These beds are overlain by tough silty shale which passes up into dark grey siltstone with sandy laminae and ribs (12 feet) with a rootlet-bearing sandstone at the top. A further eight feet of siltstone lies above this sandstone and becomes laminated at the top with some partings crowded with small delicate bivalves. These are probably fresh-water forms and so do not appear to represent an interdigitation of marine and non-marine conditions, but the lowest shale may belong to the _murchisonae_ beds. Elsewhere in the north-western escarpment the ironstone and limestone appears to be absent and non-marine shales or sandstones rest on the _murchisonae_ beds.

Exposures in the dales tend to be very limited in extent and the large-scale nature of the marine - non-marine contact is not clear. In all cases, however, the Dogger is overlain by shales or fine-grained sandstones and erosion of the Dogger is not evident.

The non-marine beds are therefore for the most part concordant on the Dogger where it is developed although a certain amount of redistribution of sediment may have occurred in the present coastal region before and during the initial non-marine sedimentation. Colonisation of the Dogger sediment by plants is evident between Whitby and Black Nab and probably preceded the erosional phase. In the extreme north and west the murchisonae shales are developed in most places with occasional developments of carbonates below. The variation between sections in this region appears to be due to the uneven nature of the eroded surface of Lias shale, a feature which is more pronounced to the south (Black, 1934 p.273).

A striking feature of the basal contact over the whole area is the development of typical interdistributary deposits directly onto the marine beds whether or not erosion has occurred. This indicates that the Dogger surface was either exposed or covered with very shallow water, so that no delta foreset beds could be built up.

B. STRATIGRAPHY OF THE DELTA SEDIMENTS

As far as can be determined from available sections the first non-marine sediments to be laid down consisted of horizontally-bedded shales or siltstones with occasional sandstones. Coarse sand grains derived from the Dogger are commonly present in the lowest few inches. With the exception of the Scar section east of Whitby channelling does not appear to have occurred until several feet of fine-grained sediment had been laid down, although the associated erosion may remove the Dogger entirely. It is clear from the coastal exposures and from many inland exposures that extensive channel development took place after about ten feet of shale and siltstone had been deposited. As previously mentioned little sedimentation of fine-grained sediment appears to be associated with these channels but beds of siltstone and shale often pass over the sandstone bodies with apparent continuity. In the Whitby to Hawsker Bottoms section these channels occur within a limited vertical range and the Unio Bed is developed with little interruption throughout, and even extends to the Peak Alum Works. Above the Unio Bed channels were not developed until nearly ten feet of sediment had been laid down, although as with the Dogger the Unio Bed is occasionally removed by erosion at the base of the channels. Again these channels appear to occur within a limited vertical range. The highest beds are not continuously exposed but are better displayed in the section between Hayburn Wyke and Peak. The distribution of channel sandstones is represented diagrammatically in figure 1.56 where two phases may be distinguished.

It appears that phases of relatively widespread interdistributary sedimentation were interrupted by several phases of channel erosion and sedimentation, apparently with little overbank deposition. The channels must represent offshoots from larger, more permanent distributaries, since the coarseness of the detrital fragments and the presence of large logs indicates that they cannot represent drainage courses within the interdistributary area. During several distinct episodes the large distribuatries appear to have developed numerous short-lived subsidiary channels. Several factors might affect a previously stable distributary regime;

(i) increased rainfall in the drainage area: this would result in increased run-off and increased erosion of the hinterland. As a result of the high discharge crevassing would probably develop along the distributaries, with the development of subsidiary channels.
(ii) eustatic lowering of sea level: this would be expected to result only in the incision of pre-existing channels.
(iii) Uplift of the hinterland: this would result in increased erosion and run-off rate; it could also cause the increased rainfall suggested in (ii). Since the channels are largely erosional and non-depositional some relative uplift of the delta area itself is indicated.
(iv) If the uplift phase were followed by continued normal subsidence then channel sedimentation would increase rapidly while the run off would decrease as erosion of the hinterland proceeded. Finally the pre-channelling conditions would again become established.

It seems likely, therefore, that channelling phases resulted from climatic changes in the hinterland and that these changes were tectonically controlled. In contrast, conditions of low hinterland relief appear to be represented by the development of lakes between the channel phases. Low sedimentation rates reflected by the fine-grained laminated shales suggest that even the major distributaries showed a considerable decrease in capacity. In the case of the Unio Bed the lake was clearly extensive and of sufficient depth to permit colonisation by bivalves but prevent growth of plants. A gradual drying of the climate is indicated by the presence of dessication cracks locally. It is also significant that physiographical conditions were suitable for the migration of small reptiles across the area, indicating that no major distributaries prevented their movement from the more stable land surface which they presumably occupied during the active delta phases.

The development of lakes at intervals within the the interdistributary areas is only an extreme form of the general rhythmic variation which occurs in the fine-grained sediment. As mentioned on page this comprises upward coarsening and upward fining sequences, the former culminating in a sandy rootlet bed. These rhythms presumably represent small-scale climatic fluctuations causing variation in frequency and intensity of overbank sedimentation. The upward fining sequences seem to represent phases of decreasing sedimentation rate producing fine-grained muds rich in organic material. The upward coarsening sequences represent phases of increasing sedimentation rate resulting in the rapid build-up of interdistributary sediments to such a high level that conditions suitable for plant colonisation were attained. The plant growth would presumably continue for some time since until continued subsidence came into effect subsequent flood stages would be unlikely to increase the newly established sediment level by significant amounts.

Discontinuities within the fine grain ed sequences appear in many cases to be due to erosion followed by deposition of fine-grained material. Erosion in these cases presumably occurred during flood phases or during the recession of floods. Considerable local variation may occur on horizons which are devoid of channel sandstones and it cannot therefore be attributed to mutual interference of overbank sediments originating from different channels.

Considering the region as a whole, variation in the proportion of sandstone to the fine-grained beds is not very evident. However in the westernmost dales and in the western escarpment an almost continuous sandstone appears to be present near the base. The sandstone characteristically lacks normal cross-bedding and is either massive or displays low-angle lateral cross-bedding.

6. END OF DELTA SEDIMENTATION

The upper boundary of the deltaic beds and their contact with the Eller Beck Bed is much more uniform than the lower boundary. Generally speaking the uppermost beds are free of channel sandstones and this is well illustrated in the Hayburn Wyke to Peak section (fig.1.56) in which channels are rare in the top forty feet. North of the Peak Fault channel sandstones occur not far below the Eller Beck Bed but the maximum level of channelling is about twenty feet below the contact. Sandstone does occasionally underlie the Eller Beck Bed, as in Kilton Beck and Danbydale but probably a phase of non-deposition occurred over these channels which compaction and sedimentation of shales occurred to wither side. The Danbydale section indicates dessication of an upstanding channel ridge; sun-cracks are developed in a thin siltstone intercalation within a fine-grained laminated sandstone which immediately underlies the marine beds. Fine purple coloured laminae in the sand may even indicate contemporaneous oxidisation of iron. Reptilian footprints are associated with the sun-crack horizon.

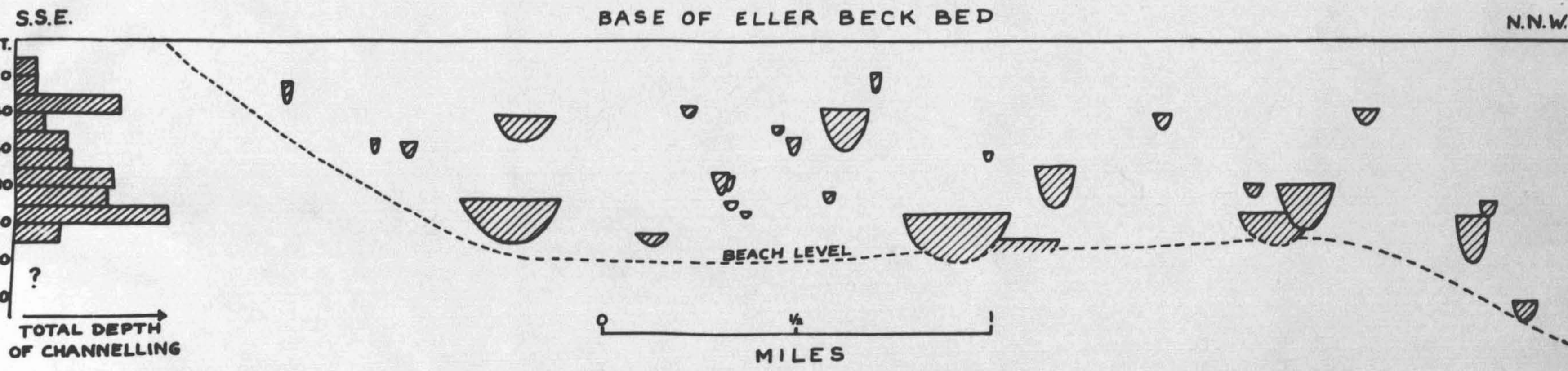


FIG. 1.56 DIAGRAM SHOWING CHANNEL DISTRIBUTION, IRON SCAR TO PEAK

Elsewhere the uppermost sediments comprise shales and silty shales. These are poorly bedded for the most part but in south-western districts are laminated and characterised by the presence of Baeiria. On the coast south of the Peak Fault the usual grey silty shales are absent in the uppermost beds which consist of a thick clayey sphaerosiderite overlain by buff laminated shale with comminuted plant material and occasional leaves. The leaf, which has not been identified, is found in the Eller Beck shales in the same area and also in Rosedale. A thin grey shale with a sharp lower contact is locally developed beneath the marine ironstone at Iron Scar. This has been included in the Eller Beck Bed by Bate (1967, p.131), but the thin impersistent ironstone referred to in that paper appears to result from burrowing at a higher horizon. Marine fossils which are common at the base of the true Eller Beck Bed shale, are absent, and since the shale is softer than the marine shale and contains carbonaceous material it is here included in the deltaic beds. It may represent a local brackish water pool which preceded the main incursion. Spherulitic shales are quite commonly developed in the uppermost beds in inland sections.

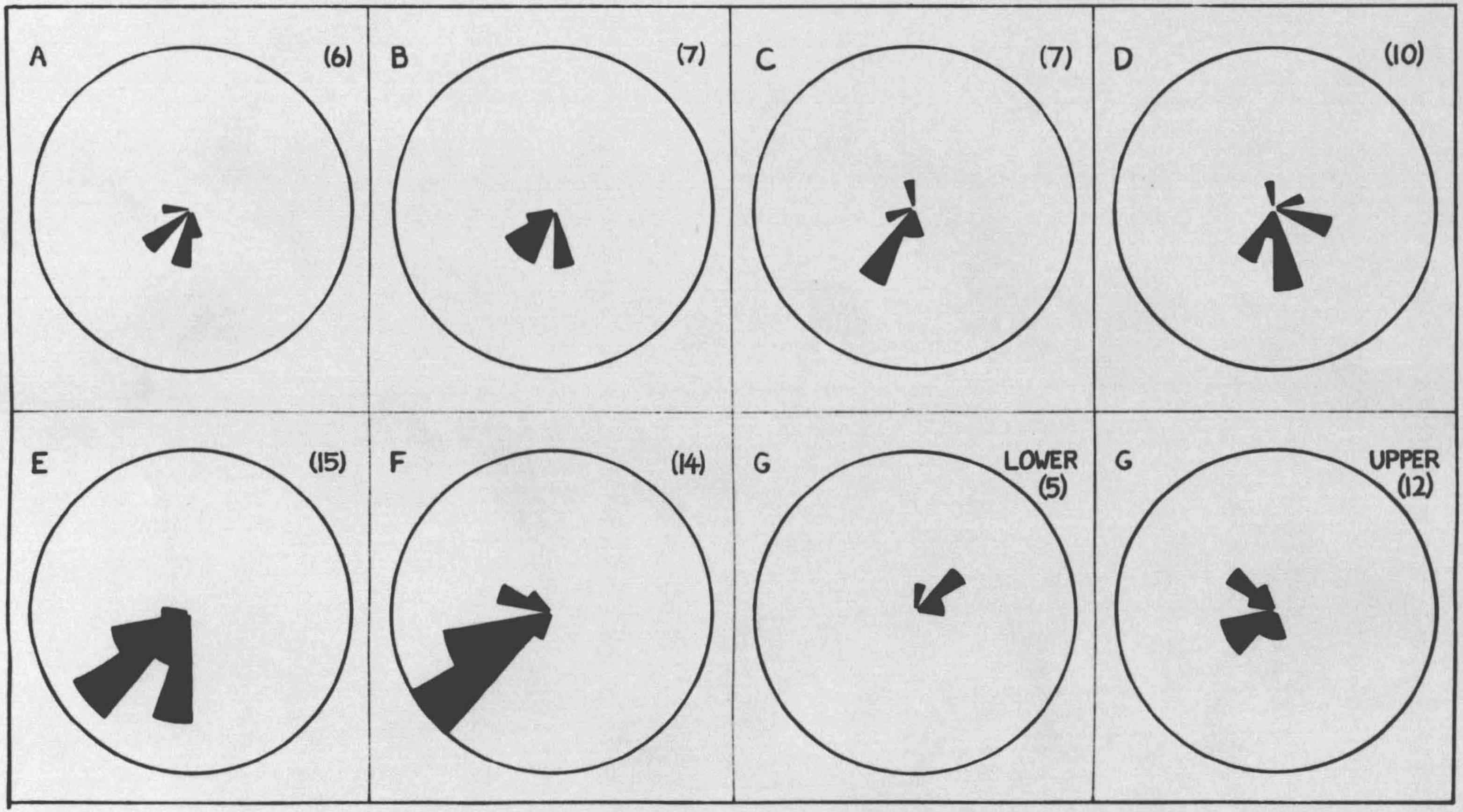
As far as can be determined channel activity seems to have ceased for some time before the marine transgression and the extensive fine-grained sediments have probably been deposited by widely spaced distributaries. The Whitby distributary may still have been active at this time but probably the deposits are related to the higher order distributaries referred to previously (p. 66).

II SEDIMENTARY STRUCTURES AND THE DIRECTION OF TRANSPORT

A. RELATIVE VALUE OF STRUCTURES AS INDICATORS OF TRANSPORT DIRECTION.

i) Cross-Bedding. With the exception of a few planar units which may represent lateral migration of sand bars the cross-bedding is believed to reflect the downstream current direction at the time of deposition. Owing to the common absence of cross-bedding and to the inaccessibility of the greater part of many sandstone bodies the number of recordings for any one sandstone are few. The maximum number obtained was sixteen, but more commonly as few as four to six readings are obtainable. Measurements were made only on units exceeding one foot in thickness; thinner units are often extremely variable in direction and in many cases represent units which have been eroded down to the bottom-sets, so that the angle of dip is low.

Variability of cross-bedding within any one sandstone may be of two types. One reflects the variability produced by a current with a constant azimuthal mean, the other reflects variation of the mean itself as well as variation around the mean. Hamblin (1958, fig.28) found standard deviations of modern streams to vary between 20 and 80 degrees, depending on the degree of meandering, and compared these values with those obtained from Cambrian sandstones. The variability in several Lower Deltaic sandstones is represented in figure 1.57 by rose diagrams and by the standard deviation value. The cross-bedding generally lies within a 100 to 200 degree sector and the standard deviation varies from 26 to 66 degrees, though most values are below 40 degrees.



READINGS GROUPED IN 20 DEGREE SECTORS ; EACH CONCENTRIC DIVISION = ONE READING ; TOTAL NO. OF READINGS GIVEN TOP RIGHT

FIG. 1.57 ROSE DIAGRAMS OF CROSS-BEDDING DIRECTIONS IN INDIVIDUAL CHANNEL SANDSTONES

The standard deviation values cannot unfortunately be directly compared with those of modern streams as recorded by Hamblin. The modern streams were measured along their length, thus including meanders, while in this study measurement has been made in a vertical cross-section. The cross-section would be expected to present a lower variability than the plan, especially if the section is perpendicular to the channel axis. However, meanders may be represented as a vertical or lateral change in direction in some sandstone bodies. Another difficulty is the small number of readings obtainable for any one channel; in figure 1.57 c, for example, variability is greatly increased by a single reading which has a profound effect on the standard deviation value.

It would appear that although cross-bedding in these sandstones may be significant in relation to the regional transport direction it is of little value in determining the nature of the streams. In a few cases, however, meandering is definitely suggested. The sandstone represented in figure 1.57 g possesses an upper and a lower portion whose mean cross-bedding directions are almost directly opposed to one another. A similar change in direction is suggested in sandstone (2) of the West Cliff section as shown in figure 1.58 where the cross-bedding directions in channels (1) and (3) are also represented. Channel (3) which as earlier stated is believed to be more long-lived than the others shows the greatest variability.

ii) Ripple-marks. Ripple-marks are seldom easily measured in these beds. Where they occur in close association with the channel sand they show rough downstream orientation, but often with a pronounced lateral component. Ripple-marks in overbank marginal facies bear little relationship to the directions obtained from structures within the associated sandstone.

iii) Elongate scours and logs. As described earlier scours and large logs commonly occur together at the base of some sandstones. Smaller logs occurring at levels within sandstone bodies are generally disoriented, but the large basal logs show good preferred orientation which corresponds to that of the associated erosional scours. The large logs always appear to be oriented along, rather than across, the channel axis. Figure 1.59 illustrates the relationship between scours and logs in a single channel. Figures 1.60 a, b illustrate the relationship of scours and logs to current bedding in the overlying sandstone; the two directions are roughly comparable in the second but not in the first. It is probable that in such cases the bottom structures are a more reliable guide to the channel direction than the cross-bedding since they represent a relatively long period of erosional flow which presumably took place in a rather straight channel. Subsequent cross-bedding, representing the depositional phase, is more likely to be affected by meandering, as appears to be the case in figure 1.60 a.

B. DIRECTION OF TRANSPORT

Cross-bedding and other directional structures have been plotted on to a Schmidt stereographic net for various regions and for the whole area (fig.1,59 a-f). The dip, and not the pole of the dip, of each reading is plotted so that a rose diagram of log and scour orientation may be incorporated in the middle. Current directions

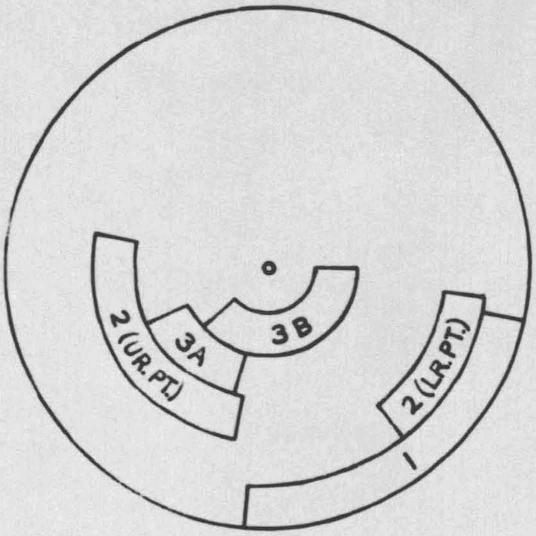


FIG. 1.58 VARIATION IN RANGE OF CROSS-BEDDING DIRECTION WITHIN & BETWEEN CHANNELS (SEE FIG. 1.45)

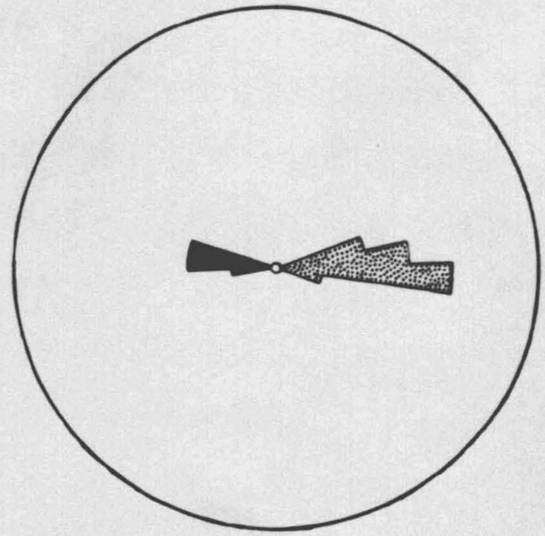


FIG. 1.59 COMMON ORIENTATION OF LOGS (STIPPLED, 10 RDGS.) & SCOURS (BLACK, 5) AT CHANNEL BASE

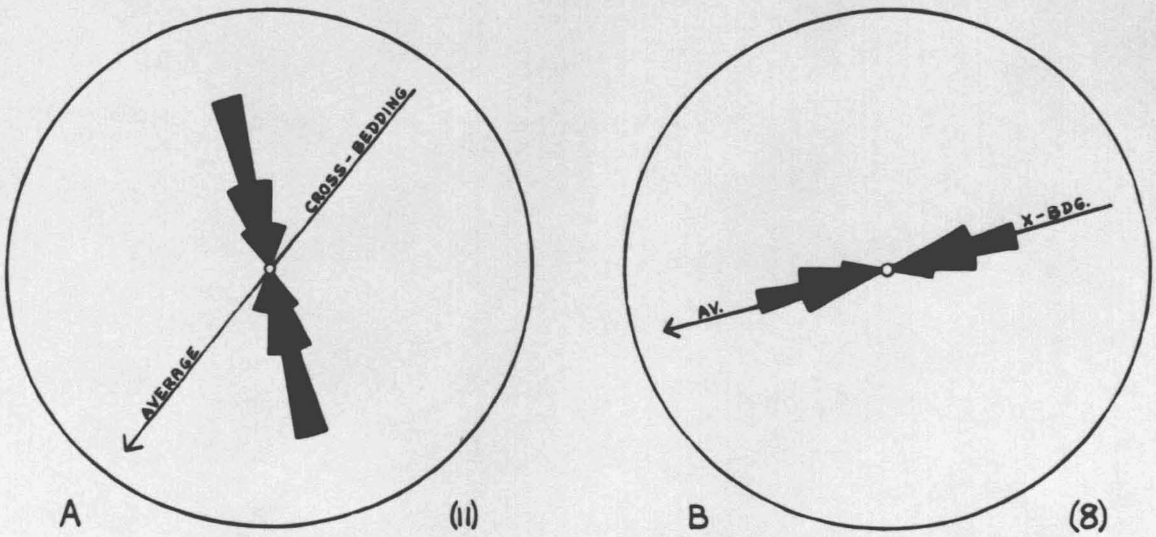


FIG. 1.60 COMPARISON OF ORIENTATION OF LOGS AND SCOURS WITH ORIENTATION OF CROSS-BEDDING IN OVERLYING SANDSTONE

Current directions obtained from trough axes and ripple-marks are shown along the margin of the stereogram.

The two coastal sections show similar patterns; the northern part shows a higher proportion of currents from the west, which is partly due to the lowest sandstones in the West Cliff section. The lower Eskdale region and Rosedale also indicate currents from the north and east but in Upper Eskdale the current direction becomes more variable, being on average from the north but including transport to the east and to the west. This variability increases to the west and north with transport predominantly to the east. The number of readings in the east far exceeds those from the other regions so that a summary of currents (fig.1.61f) directions indicates transport predominantly from the north and north-east. This is due not only to the better exposures on the coast as compared with the inland regions but apparently to the predominance of low angle cross-bedding or lack of directional structures in many of the sandstones in the north and west. The region dominated by transport to the soth and east may be of similar dimensions as that in which transport appears to have been to the south and west, but conditions of sedimentation in the former appears not to have favoured the development of normal downstream cross-bedding.

The cross-bedding readings are shown as rose-diagrams in their regional setting in figure 1.62. As discussed earlier structures at the base of channels, since they are associated with the initial erosion phase, are more likely to reflect the true direction of water flow

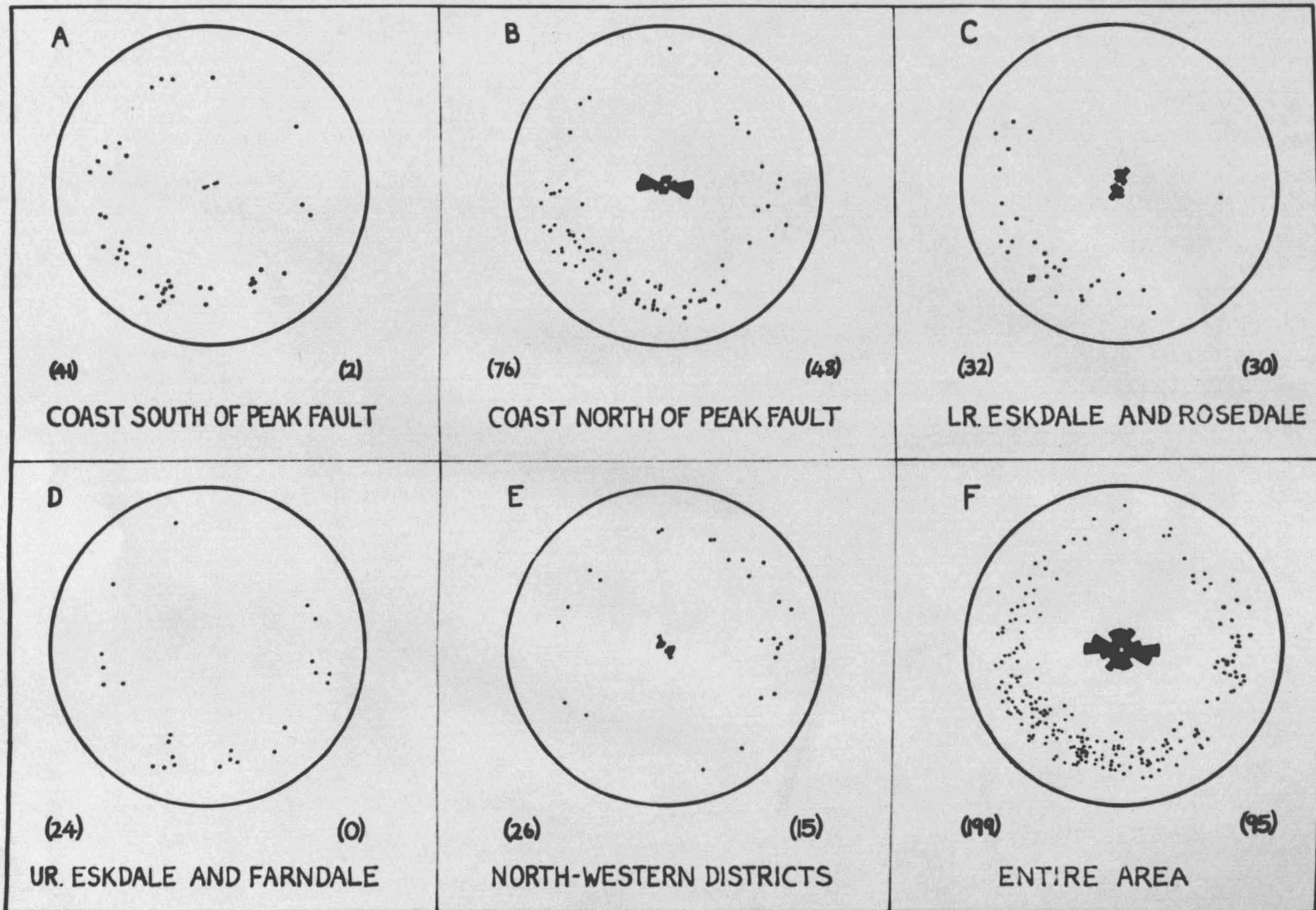


FIG. 1.61 STEREOGRAMS OF REGIONAL CROSS-BEDDING DIPS, WITH ORIENTATION OF LOGS AND SCOURS
(NO. OF READINGS CROSS-BEDDING GIVEN BOTTOM LEFT; LOGS AND SCOURS BOTTOM RT.)

which may be masked by meandering in the depositional phase. On a regional scale erosional structures will tend to run perpendicular to any surface slope and therefore give a true reflection of the palaeoslope which could otherwise only be obtained by a large number of cross-bedding readings from a considerable length of any one channel. Rose diagrams for individual localities within the region are shown in figure 1.62 and show remarkable lack of variability. In the east and south the two lines of evidence correspond closely but divergence between the two increases to the west owing to the increased variability of the cross-bedding orientation and to the intermittent exposure.

The cross-bedding or 'drift bedding' was investigated by Sorby in 1852, and his results have often been quoted in subsequent work. He concluded that: "...the drift bedding indicated that the currents had been from points varying from E.N.E. to W.N.W., the mean being from N.N.W. to N.W. The direction indicated by the ripple-marks varied from N.E. by E. to N.W. by W., the mean being from N. to N.N.W. Whence the mean derived from these two sources combined is N.N.W.". The mean of readings in the Lower Deltaic sandstones in this study lies between north and east, but had exposure been more uniform the mean would probably have been between the north-west and north-east. Since Sorby's readings were obtained from the entire range of the 'oolitic sandstone' a more balanced sample may have been possible.

The transport direction as inferred by Smithson on zircon grain size throughout the deltaic beds corresponds closely with the cross-bedding data for the Lower Deltaic Series. His conclusions have been

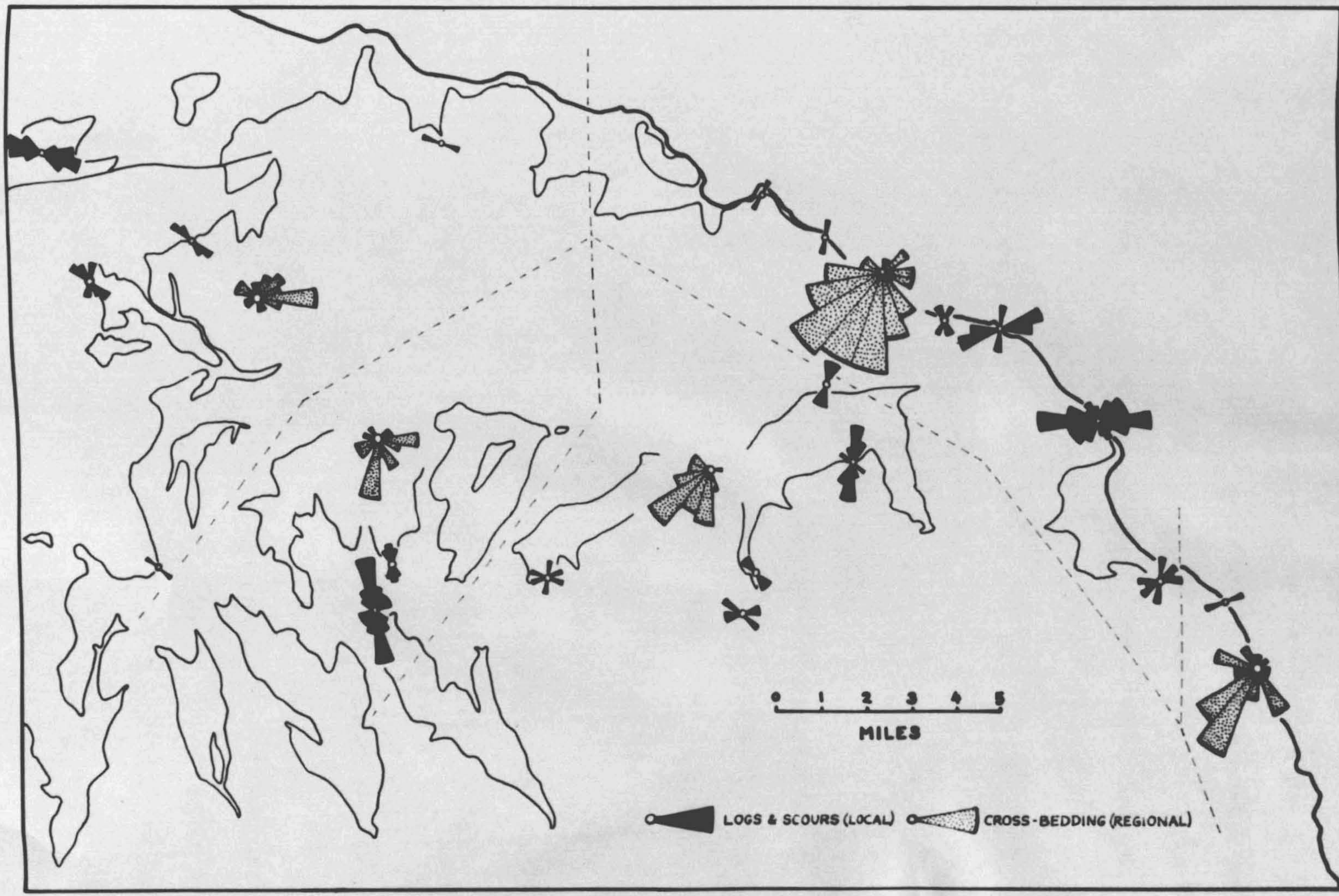


FIG. 1.62 REGIONAL ORIENTATION OF DIRECTIONAL STRUCTURES

criticised (Hastall, 1942) in that the grain size might be a reflection of conditions of deposition rather than indicating downstream attrition. Considering the distances involved this seems more probable, and the tendency for coarsening towards the margin of the present area might reflect proximity to major distributaries. Channels which are as short-lived as most of those exposed appear to have been might be expected to show strong size grading with dumping at the end near the major distributary.

III SOURCE OF SEDIMENTS

As outlined in the introduction several hypotheses have been put forward as to the source of the Middle Jurassic sediments and as to the palaeogeography of the surrounding region. The present study has indicated that currents flowed into the area of the present outcrop from the north, east,and west and drained off to the south. This flow pattern is well defined in the erosional stages of channel formation and therefore probably reflects an original slope. Such a slope could have been produced in two ways:

i) by tectonic subsidence of central part of the area

ii) by tectonic uplift of hinterland

iii) by low sedimentation in the central part as compared with the margins.

If tectonic subsidence were the controlling factor it would be expected that a greater thickness of sediment would have accumulated in the central area of maximum subsidence than at the margins. The vertical distance between the Dogger and the Eller Beck Bed has been estimated by means of an Abney Level at several inland sections. On the coast of Whitby the Lower Deltaic beds are just over one hundred feet thick, and a similar thickness was estimated for Little Beck. In Rosedale and Great Fryupdale, however, the thickness is reduced to about eighty feet, while it again reaches one hundred feet at Burton Howe on the western escarpment. The beds therefore become thinner towards the centre of the outcrop and relative uplift rather than subsidence is indicated.

The preferred orientation of streams could have been a reflection of the distribution of the source outcrops; if slopes were sufficiently steep to produce fast-flowing and straight streams oriented at right angles to the outcrop boundary the orientation might be retained in the delta area. In order that such a pronounced convergent stream pattern could occur in such a small area this would require that the same outcrops lay very close to the margins of the present area. This has been suggested by Smithson (1942, fig.15), but recent mapping of the bed of the North Sea (Donovan 1968, fig.1) has indicated that Jurassic strata extended eastwards for at least fifty miles; beyond this limit they have been cut out by the Cretaceous unconformity.

The distributary pattern of the present outcrop is best explained as resulting from the occurrence of major distributaries running on either side of the region in an approximately north-south direction. If these large rivers underwent little change in course then an inter-distributary area would become established between them. The beds preserved in the present outcrop are believed to have occupied a central position in this interdistributary area, so that in times of increased run off secondary distributaries would enter the area from both the east and west, the convergence being due partly to the distribution of the large rivers and partly due to the relatively low elevation of the interdistributary area. On reaching the centre of the interdistributary area the streams would then have turned southwards to flow down the regional slope which controlled the direction of the larger rivers.

Close proximity to land is therefore not required in a palaeogeographical construction. The similarity in the sediments and in the nature of the lowest and highest beds over the whole area favours derivation from a single source. Smithson's work on heavy minerals, however, indicates that garnet is present in the western outcrops but not in the east. This is attributed to interstratal solution, in the central and eastern part, of garnet which was originally deposited over the whole area. If, however, the presence of garnet in the west were in fact a primary feature it could have been introduced by small streams draining a low-lying Pennine landmass and acting as tributaries to the proposed western deltaic distributary.

If the immediate source of the Yorkshire sediments is the major deltaic distributaries in the area the ultimate source has to be further afield. It is probable that this source lay somewhere between north and east of the present outcrop. The general north-south flow indicated by current-bedding readings only represents local flow direction which may have resulted from the deflection of rivers from the east against the Scottish-Pennine landmass. Arkell (1933 p.596 and Spath, quoted in Smithson, 1942 p.50) agree that a connection with an Arctic sea must have been present to the east of a Scottish-Pennine landmass during the Lower and Upper Jurassic, and therefore the existence of a northern land-mass as proposed by Kendal and Wroot (1924 p.311) is improbable. The nature of the heavy mineral assemblage which consists essentially of the stable minerals zircon, rutile and tourmaline together with various opague minerals indicates derivation from pre-existing sediments. Smithson (1942,p.48) compared the composition and grain size heavy mineral assemblages of the Middle Jurassic sediments with those of the Carboniferous, Permian, and Trias of northern England and, on the assumption that garnet had undergone subsequent solution, concluded that Carboniferous sediments could have supplied most of the assemblage, with small additions from the sediments of the other two systems. Whether these results could be applied to possible source beds to the east is not certain, but the Carboniferous heavy mineral assemblages might be expected to be fairly constant over an area such as the northern part of the North Sea.

The most probable palaeogeographical reconstruction would therefore involve a Scottish landmass with a probable low lying southward extension into the Pennine area, and a substantial landmass to the east or north-east of Britain in which the surface rocks belonged to sedimentary formations, possibly of Carboniferous age. This eastern landmass would have been drained by a river or rivers which flowed into the low lying area formerly occupied by the sea, and built up a delta which extended right across to the landmass of northern Britain. The western limit of such a Scandinavian landmass cannot be determined but the area covered by the delta was probably large. If conditions preceding delta formation over the whole area were like those in Yorkshire, with shallow water or even emergence, a delta could build up extremely rapidly from a river of moderate size.

The suggestion that a Scandinavian landmass provided the deltaic sediments is in accordance with the conclusions of most recent workers, including Arkell (1933, p.596), Black (1934, p.279) and Wills (1952). Bate (1967, p.137), on the other hand, follows Smithson in proposing an arcuate land mass to the west, north, and east of the area and very close to the limits of the present outcrop.

CHAPTER III PETROGRAPHY

I SANDSTONE

A. GRAIN SIZE ANALYSIS

The laboratory method in preparing samples for mechanical analysis is described in Appendix 2a .

i) Statistical Measures used in Analysis. The parameters which are here used to represent the grain size distribution are:

Mean Grain Size $M\phi = \frac{\phi16 + \phi50 + \phi84}{3}$

Standard Deviation $\sigma\phi = \frac{\phi84 - \phi16}{2}$

Skewness $\alpha\phi = \frac{\phi84 + \phi16 - 2\phi50}{\phi84 - \phi16}$

The more comprehensive formulae used by Folk and Ward (1957) for standard deviation and skewness are not used, since, owing to the high percentage of material finer than 4.75o, the 95th percentile was not reached in any of the plots obtained. For the same reason kurtosis could not be determined. The use of arthmetical probability paper allows the distribution to be checked for log normality. In this and in later sections the log normal part of the cumulative curve has been treated separately and given its own parameters, distinguished by the suffix 1 :

$$M_1\phi; \sigma_1\phi$$

In addition to these standard parameters the percentage at which the inflection of the curve occurs is recorded. This is known as the Inflection Percentage and is represented by the symbol I%.

ii) Results. The mechanical analysis was carried out on specimens of varying grain size and structural features. The results obtained are shown in table 1.1. Some of the cumulative curves from which these results were obtained are shown in figure 1.68.

i) Mean Grain Size (see table 1.1) ranges up to 1.7ϕ with the exception of the three coarse sandstones (15 to 17). These last three were each collected from the base of a channel sandstone which rests directly on the Dogger sandstone, and contain grains considerably coarser than any found in other parts of the Lower Deltaic succession. Since the Dogger sandstone frequently contains very coarse grains, it is thought that the coarse sand has been incorporated into these channel sandstones as a lag deposit. 16 and 17 are unimodal, and 15 bimodal, perhaps reflecting a mixture of the lag and normal bed-load populations. On the number of analyses available, the only other significant feature is the slightly coarser average mean grain size of the structureless sandstones (group D) relative to the cross-bedded sandstones (group B).

ii) Standard Deviation is variable but on average differs little from group to group.

iii) Skewness shows more significant variation than does standard deviation. The average value for group E is particularly high (0.57), while the remainder, in decreasing order of average values, are: group A, 0.42; group C, 0.39; group B, 0.31; group D, 0.09.

iv) Inflection Percentage shows on average significant variation between groups:

A	78%
B	85%
C	80%
D	89%
E	68%

Since the cumulative curves are log-normal at the coarse end, the

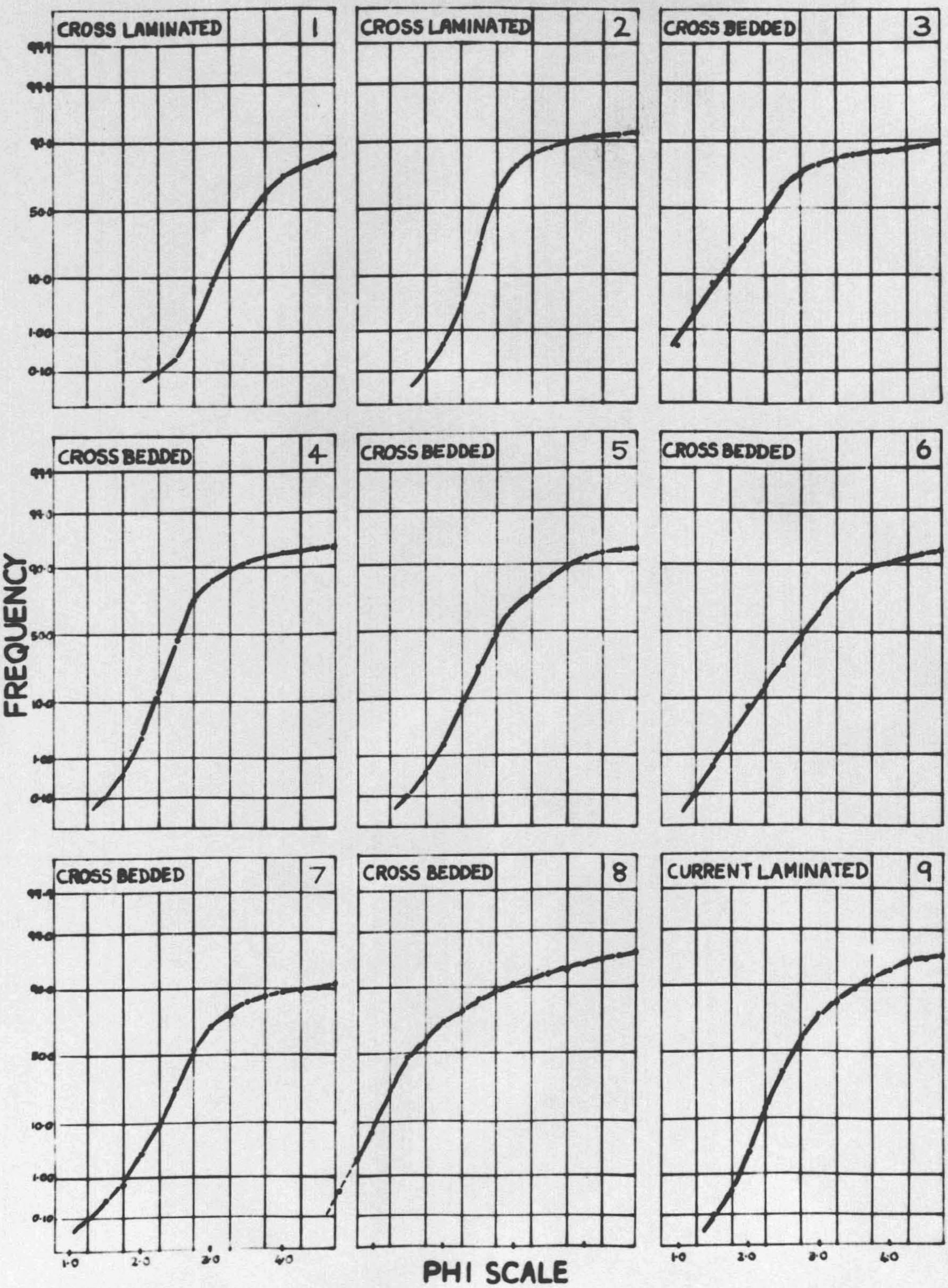


FIG. 1.63 (i) CUMULATIVE CURVES FOR DELTAIC SANDSTONES

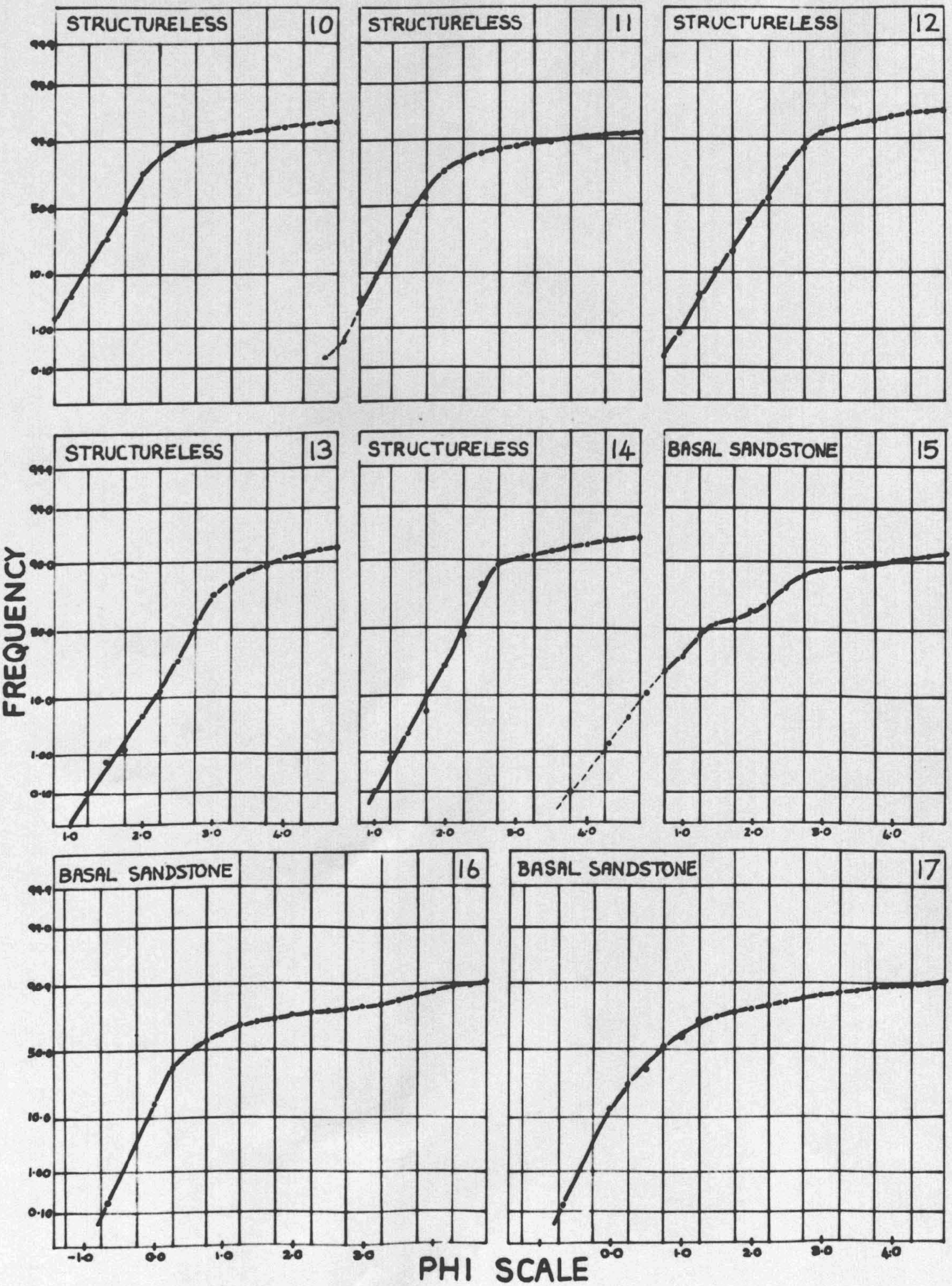

STRUCTURELESS 10
STRUCTURELESS 11
STRUCTURELESS 12
STRUCTURELESS 13
STRUCTURELESS 14
BASAL SANDSTONE 15
BASAL SANDSTONE 16
BASAL SANDSTONE 17
FREQUENCY
0·10
1·00
10·0
50·0
1·0
2·0
3·0
4·0
-1·0
0·0
PHI SCALE

FIG. 1.63 (ii)

STRUCTURE	NO.	$M\phi$	$\sigma\phi$	$\alpha\phi$	I%
A. SMALL-SCALE CROSS-BEDDED	1	2·75	0·45	0·33	81
	2	3·93	0·83	0·50	75
B. LARGE-SCALE CROSS-BEDDED	3	2·52	0·83	0·39	80
	4	2·58	0·35	0·14	88
	5	2·88	0·55	0·36	87
	6	2·83	0·55	0·10	85
	7	2·88	0·60	0·46	84
	8	1·72	0·65	0·38	87
C. CURRENT LIN$^{\underline{o}}$	9	2·75	0·83	0·39	80
D. STRUCTURELESS	10	1·82	0·53	0·05	89
	11	1·78	0·70	0·29	85
	12	2·07	0·43	0·00	92
	13	2·78	0·55	0·10	88
	14	2·25	0·40	0·00	90
E. SANDSTONES WITH DERIVED DOGGER GRAINS	15	1·67	1·05	0·24	–
	16	1·40	1·83	0·70	66
	17	1·23	1·40	0·57	70

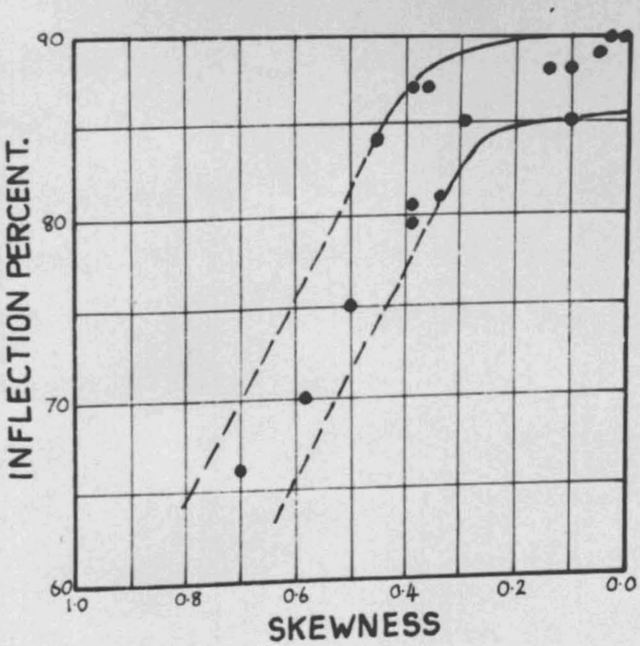
TABLE 1.1 STATISTICAL PARAMETERS OF LOWER DELTAIC SANDSTONES

value of skewness is entirely dependent on the nature of the fine end of the curve, or more precisely, on the point at which the curve ceases to follow a log normal course, together with the angle of inflection. When the log-normal curve continues to any extent above the 84th percentile, the skewness will equal 0. Skewness increases rapidly with the distance at which the termination of log-normality lies below the inflection.

Graphs have been drawn (fig.164) to show the varying degrees of dependence of skewness and standard deviation on the inflection percentage. The latter apparently reflects the decrease in curvature more accurately than the point at which the curve ceases to be log-normal, which does not reflect the rate of change of curvature. The plot of α against I% shows a definite relationship, the small scatter being the result of the varying slopes of the fine 'tail' of each curve. The somewhat greater scatter on the plot of σ against I% results from the influence of the slope of both sections of the curve.

From the above observations it may be concluded that while the parameters $M\phi$, $\sigma\phi$ and $\alpha\phi$ reflect quite accurately the nature of the curve, a more informative series of parameters would comprise $M_1\phi$, $\sigma_1\phi$ (mean grain size and standard deviation of the log normal part of the curve), the inflect percentage, and some measure of the sorting of the fine section of the curve. The means by which these parameters may be obtained are dealt with below.

<u>Curve Separation</u> is carried out by a method based on the assumption that a log normal distribution is the most superior that a sediment may attain. Deviations from the log-normal distribution would



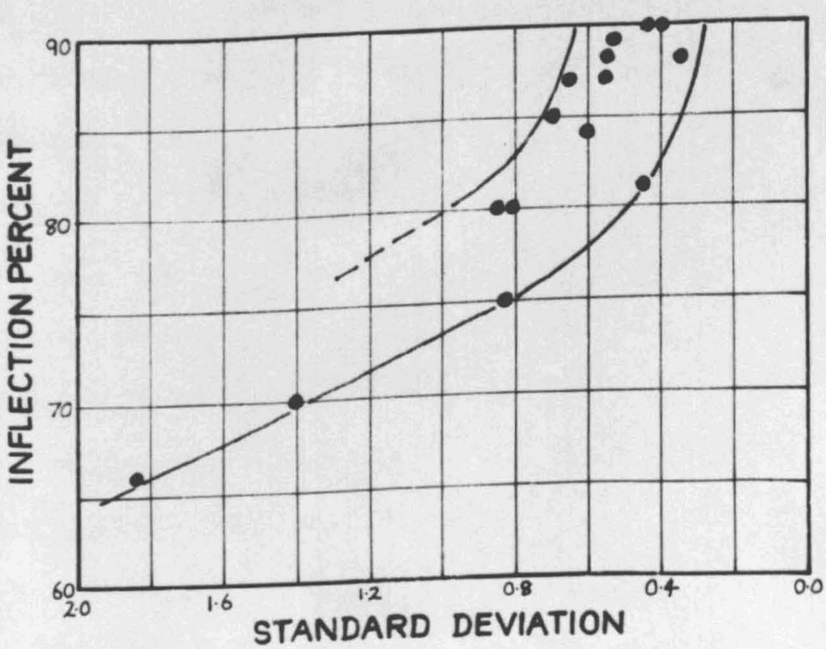


FIG. I.64 RELATIONSHIP OF SKEWNESS AND STANDARD DEVIATION TO INFLECT PERCENT therefore reflect a lack of uniformity in the conditions of accumulation of a sediment. This assumption, as stated by Tanner (1964, p.157), has no theoretical basis, but it has been adopted here in view of the ubiquitous tendency towards log-normality in the sediments studied, and by virtue of the significance of the separated log-normal curve found as a result of the study.

Curve separation has been carried out by a quick graphical method in which the non-cumulative percentage of each size fraction is plotted on arithmetical probability paper to form a size frequency curve. The trace of the log-normal part of the curve is then produced towards the fine end of the scale so that a symmetrical log normal curve is produced. The numerical values of this curve are then read off at the usual $\frac{1}{4}\phi$ intervals and recalculated to 100%. The fine fraction is similarly treated, and both fractions plotted as separate cumulative curves. The method is outlined in figures 1.65 a,b.

Owing to the high percentage of very fine material the plotted part of the fine fraction rarely reaches the 50th percentile, and parameters have not been obtained for these curves. It is obvious from the nearly horizontal trend of the fine end of many of these curves that a sharp upward inflection must take place in the unseen portion of the curve, probably in the region of clay sized particles. It is concluded in the following section on microscopic petrography that much of this clay may be secondary, in which case the nature of the fine separated curve would be considerably altered from the original. Certainly the sorting of these curves is extremely poor when compared with those derived from analyses of similar sands from

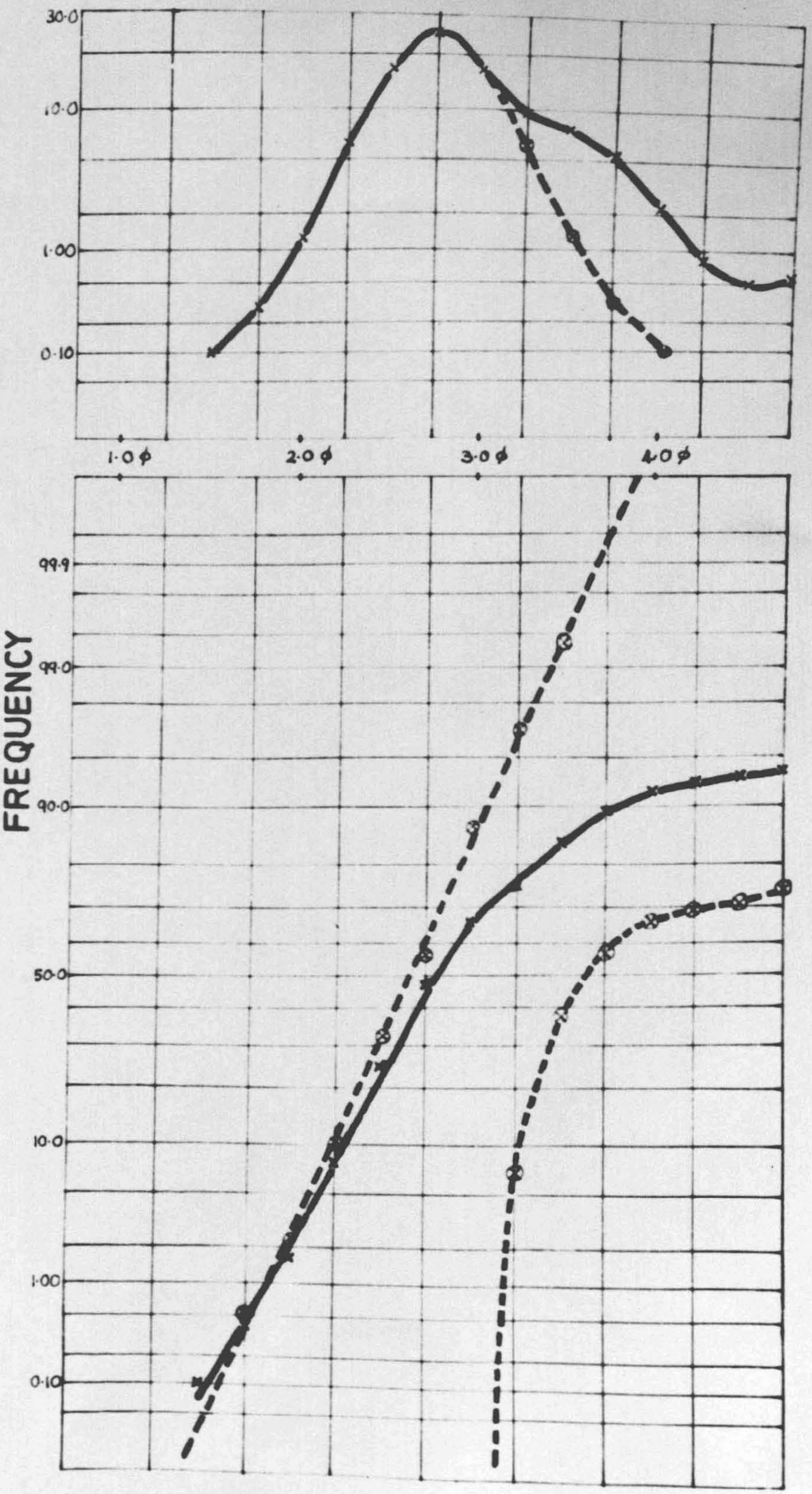


FIG. 1.65 CURVE SEPARATION

present day fluviatile environments (see for example Visher, 1965, p.126).

The results obtained from curve separation are shown in table 1.2 where they are compared with those obtained from the original curves: The new values reduce both mean grain size and standard deviation numerically, and considerably reduce the variability of the standard deviation in the group as a whole. The σ_ϕ values for the structureless sandstones are less variable and, on average, higher than those for the current bedded sandstones. The significance of the new values, in particular those of σ_ϕ o. will be considered in Part III, when the interpretation of cumulative curves in a broader sense is discussed.

B. MICROSCOPIC PETROGRAPHY

a) Grain composition

i) Quartz is the dominant grain type and usually makes up over ninety-five percent of the grains . In a few sandstones some other grain type shows local concentration and the proportion of quartz is therefore reduced. The original outline of the grains is very much obscured by intergranular solution and by secondary growth. The best guide as to the original surface features is the 'ghost' surface which is accentuated by inclusions beneath authigenic overgrowths which are clearly not derived. Such relict surfaces indicate that the grains were originally of moderate sphericity and sub-rounded.

ii) Feldspar is normally a minor constituent of the sandstone, making up between one and five percent of the grains. Both potash feldspar and plagioclase feldspar are present, but the former is more

STRUCTURE	NO.	$M\phi$	$M_1\phi$	$\sigma\phi$	$\sigma_1\phi$
A. SMALL-SCALE CROSS-BEDDED	1	2·75	2·60	0·45	0·25
	2	3·93	3·53	0·83	0·38
B. LARGE-SCALE CROSS-BEDDED	3	2·52	2·18	0·83	0·50
	4	2·58	2·48	0·35	0·28
	5	2·88	2·60	0·60	0·30
	6	2·83	2·73	0·55	0·48
	7	2·88	2·60	0·60	0·38
	8	1·72	1·28	0·65	0·30
C. CURRENT LIN$^{\underline{o}}$	9	2·75	2·60	0·83	0·30
D. STRUCTURELESS	10	1·82	1·65	0·53	0·38
	11	1·78	1·45	0·70	0·40
	12	2·07	2·05	0·43	0·45
	13	2·78	2·60	0·55	0·35
	14	2·25	2·18	0·40	0·35
E. SANDSTONES WITH DERIVED DOGGER GRAINS	15	1·67	-	1·05	-
	16	1·40	0·20	1·83	0·33
	17	1·23	0·50	1·40	0·50

TABLE 1.2 COMPARISON OF STATISTICAL PARAMETERS OF SEPARATED AND UNSEPARATED CURVES

abundant. The feldspar grains are fresh for the most part, but the plagioclase in particular may show varying degrees of sericitisation. The original surface form of the grains is destroyed in the same way as that of the quartz grains.

iii) Rock fragments in most sandstones consist only of compound quartz grains, and form only a small proportion of the grains. The constituent quartz particles are usually very tightly interlocking and the grains appear to be of metamorphic origin. Sandstones whose base hase scoured the Dogger contain grains, and sometimes pebbles, of phosphatic material which in some cases includes ooliths.

iv) Accessory minerals which are observed in most sandstones include muscovite, zircon, leucoxene, rutile, and tourmaline, in order of decreasing abundance. Muscovite is never very common and the other minerals are conspicuous only when they are concentrated along certain laminae.

v) Intraclasts are rare. The only grains which appear to have been derived from previously deposited non-marine beds are entire or fragmented siderite spherulites, and other sideritic or limonitic grains. These are locally abundant and usually occur in lag deposits at the base of channels.

vi) Plant fragments include various cellar carbonised material; larger fragments may show the development of shrinkage cracks.

b) Matrix Composition

i) Clay is the normal matrix constituent. It usually occurs as random aggregates of crystals which range between 5 and 10 microns in diameter. These crystals are often seen to consist of stacked sheets in a vermicular habit and in some cases larger vermicular crystals occur which may reach 50 microns in length. The crystal habit suggests that the clay consists principally of kaolinite and this has been confirmed by x-ray diffractometry (see Appendix 2B, no. 1).

ii) Siderite occurs in many sandstones in association with a clay matrix. When unweathered it is seen to consist of clear yellow brown crystals with well-developed rhomb faces; these may occur as isolated individuals, as clusters, or as more continuous interstitial aggregates. Some crystals occur as overgrowths on what are apparently detrital grains. Some of those shown in figure 1.66 show pronounced zoning which may indicate more than one phase of reworking. Usually the original structure of the siderite is obscured by oxidation which causes a considerable amount of redistribution of iron; rhombic outlines can, however, often be made out in some of the limonitic matrices.

iii) Calcite is confined to the basal part of the channel in the Scar section at Whitby, and to a few interdistributary sandstones. The calcite, as shown by staining (see Appendix 2C, no. i), is usually ferroan, but occasionally non-ferroan calcite is developed at the centre of the cemented interstices. The calcite is clear and often poikilitic.

c) <u>Diagenesis</u>

i) <u>Sandstones with a clay matrix</u>

The most pronounced diagenetic feature is the reoganisation of the original grain framework by intergranular solution. This leads to the development of tightly interlocking clusters of grains from which all interstitial space has been excluded;the grain contacts vary from straight to rather irregular. Where a grain has penetrated another deeply a rounded crystal shape may be developed; this is apparently due to selective solution and not to overgrowth. Peripheral overgrowth, with the development in some cases of sharp crystal faces, is present; in particular it develops so as to fill interstices within the grain clusters. The larger interstitial areas between the grain clusters are clay-filled. The absence of small terrigenous grains together with the relatively large size and vermicular form of the kaolinite indicates that the clay is diagenetic and is the product of direct precipitation rather than recrystallisation. Grains which line these clay patches show ragged surfaces, even where overgrowths have developed. The proportion of clay matrix varies but may reach twenty percent; values such as this are not uncommon in channel sandstones, but interdistributary sandstones often show more intense suturing with consequent reduction in the proportion of matrix.

It appears that after deposition of the sandstone solution and redepositionof silica has taken place throughout the rock. This process is somewhat localised and may have been most pronounced where grains were originally most tightly interlocking. Such areas would be expected to be subject to greater pressure than those in which the grains were more loosely packed. After the pressure solution had taken place conditions appear to have been present in the pore solutions under which silica chemically unstable, resulting in corrosion of the grains. Clay was deposited in the interstices at this or at some subsequent time.

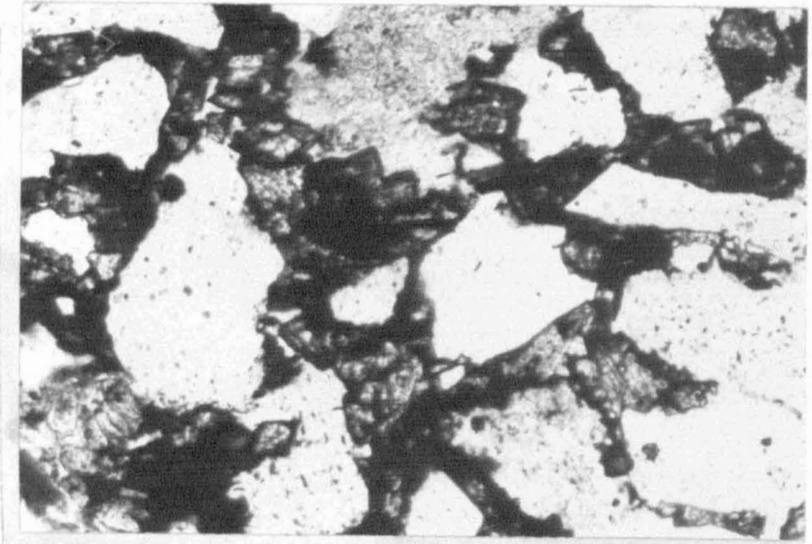
Mobilisation of iron is a common feature in the sandstones. The resulting siderite precipitation takes two forms depending on whether the interstices were voidal or filled with an original clay matrix. In the first case siderite develops as clear rhombic crystals or groups of crystals; penetration of sand grains by the siderite is common and euhedral faces are developed on both replacive and drusy surfaces. Siderite may be replacing clay in interstices but in most cases in these and the marine beds above, kaolinite seems to have been deposited subsequent to siderite formation. The lack of intergranular solution in sandstones with almost entirely sideritic cement supports this view, since it indicates that siderite cementation was probably one of the first diagenetic processes.

Fine-grained interdistributary sandstones commonly possess a matrix of clay and silt; when siderite occurs in these beds a spherulitic and not a rhombic habit is developed (fig. 1.67). The spherulites are very small, ranging up to sixty microns. At the centre of each spherulite is a grain of translucent amorphous reddish brown material; the outer surface is quite sharply defined and commonly shows rounded protruberances. Sometimes larger masses of similar appearance occur; these show continuous rims of radiating siderite crystals. The nature of these nuclei is not clear but they appear to be detrital; whether they represent flocculated particles or reworked ferruginous and oxidised siderite crystals is not certain. The later siderite rims have clearly grown in place, and show mutual boundaries in the same way as normal spherulites (p. 98). Siderite has apparently replaced much of the original matrix and the sand

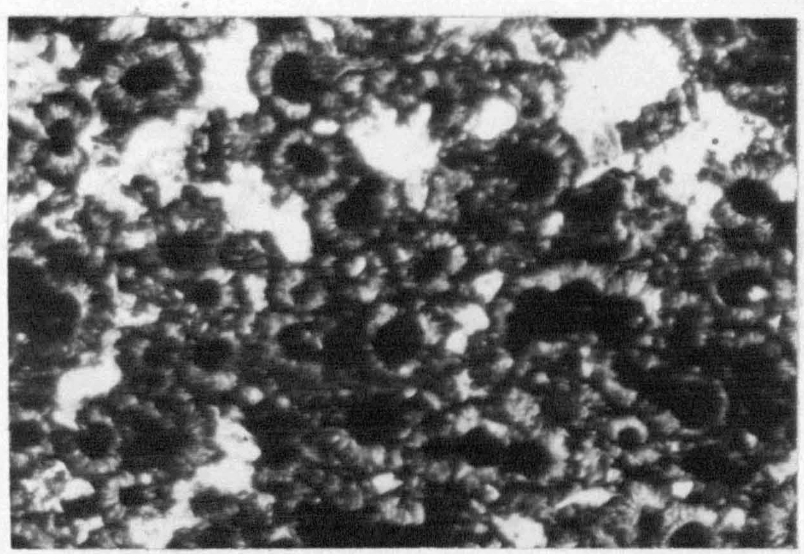
FIG. 1.66 Zoned siderite rhombs in sandstone matrix (X50)

FIG. 1.67 Spherulitic siderite growing off opaque cores in coarse siltstone. (X70)

FIG. 1.68 Large siderite spherulites in microspherulitic matrix. (X55)



1.66 500μ



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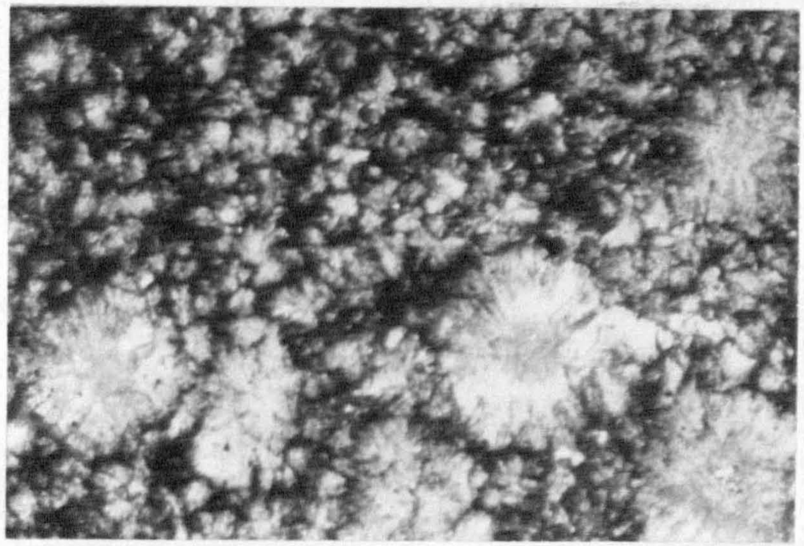


1.68 500μ

grains show varying degrees of surface erosion.

Accessory minerals appear on the whole to have been stable. Muscovite flakes sometimes show curvature, apparently due to intergranular compaction. Another feature common in sideritic rocks is the development of small siderite crystals along some cleavage planes, forcing adjacent flakes apart. In one sample a rounded tourmaline grain was observed to have a large overgrowth, but it is not clear whether or not the entire grain is derived.

ii) Sandstones with a calcareous matrix

In these sandstones the calcite cement makes up about thirty percent of the rock. Grain contacts are frequently seen, but no suturing has occurred. The grains usually have a smooth outline but at the same time may show deep indentations which appear to be the result of solution. Feldspars are very susceptible to replacement by calcite; this proceeds along cleavage planes and results in the formation of skeletal crystals.

The absence of grain suturing indicates that calcite, in the same way as siderite, is an early cement.

II SHALES AND SILTSTONES

Little petrographic work has been carried out on these beds. Some of the normal interdistributary siltstones were examined by X-ray diffractometry for their clay mineralogy content. The results (see appendix 2B, nos. ii) indicate that kaolinite is the dominant clay mineral; similar results were obtained from insoluble residues of siderite mudstones (nos. iii).

III. IRONSTONES

The siderite mudstone nodules present few features in thin section. The rock consists of a homogeneous mass of cryptocrystalline or microcrystalline turbid siderite with occasional small mica flakes and plant fragments. Speulitic ironstone is more variable; the variation occurs in the proportion of siderite to clay matrix and in the size and internal arrangment of the spherulites. In the most clayey, uncohesive types spherulites of similar size are developed throughout, with diameters ranging from 250 to 500 microns. In the tougher, nodular types, however, spherulites of similar diameters are accompanied by a groundmass of microspherulitic siderite (fig. 1.68) The microspherulites which range up to 100 microns in diameter are in close contact with one another, so that interstitial clay is absent.

The contact surface between spherulites is straight or gently curved, the curvature being convex towards the larger individual. Finely divided opaque material is concentrated along these boundaries. Internally the spherulites consist of radially oriented fibres of siderite which is mostly yellowish brown and slightly turbid. The centre of the spherulites is however occupied by a colourless and structureless zone of cryptocrystalline siderite. Blebs of pyrite may occur in this central zone and often form an irregular ring along its margin. This central zone never contains any traces of detrital grains but in the surrounding fibrous zone silt-sized grains are sometimes present; these grains are most common in the outer parts of the fibrous zone. In many cases the included grain has been replaced by siderite in optical continuity with adjacent fibres; larger grains retain their original composition. A crude concentric zoning is present in the fibrous zone with a tendency for an increase in turbidity and depth of colour outwards.

The matrix in the sparsely spherulitic rocks consists of little altered silty clay. With increased sideritisation and reduction in interstitial area a development of white mica becomes apparent. Where replacement by large spherulites is almost complete the interstices are crowded with mica flakes.

The development of sphaerosiderite appears to consist of crystallisation of siderite at a large number of points and subsequent growth from these centres. Replacement must have taken place but insoluble organic material and the larger detrital grains may resist replacement. There are indications that insoluble particules are pushed outwards by the growth of the spherulites. The concentration of mica in interstices seems to represent diagenetic crystallisation and not physical concentration of previously scattered flakes. The mica must represent in part the clay expelled by sideritisation. In the more compact varities an acceleration of siderite crystallisation is implied by the development of interstitial microspherulites.

The origin of the structureless zone at the centre of the spherulites is not certain but it is probably an original feature. The only clear alteration of spherulites is the recrystallisation of groups of fibres to produce larger crystals comprising four or more sectors of the spherulite.

PART TWO

THE ELLER BECK BED

INTRODUCTION

A. REVIEW OF LITERATURE

The term 'Eller Beck Bed' was introduced by Barrow (1877) to include a thin series of marine strata intercalated within the non-marine 'Lower Estuarine' beds. The marine bed comprises three units which in upward sequence are ironstone, shale, and sandstone. The name is derived from the section at Walk Mill Force in Eller Beck, near Goathland. In the paper referred to above, Barrow gave details of exposures along the coast south of Whitby and in southern and central Eskdale. These localities were added to in the subsequent Memoirs of the Geological Survey (Fox-Strangways and Barrow, 1882; Fox-Strangways Reid and Barrow, 1885; and Barrow, 1888), and a complete summary is given in Fox-Strangways (1892). The sections described are with rare exceptions available at the present day.

The Eller Beck Bed ironstone had been mentioned in the literature prior to the publication of Barrow's paper. Marley (1857, p.202) gives a section of the ironstone at the Ingleby Manor Mines of which he says: "From the above section it will be seen that a third seam comes into view for the first time as a working seam, and is higher geologically than the seam called the top seam of Cleveland and, therefore, I have called this seam the 'Ingleby' top or third seam," He later recognised the seam in Fangdale Beck (Marley 1858, p.158).

The ironstone was traced over a much wider area by Bewick (1861), who recognised the bed on the coast south of Whitby, through much of Eskdale, and in the north near Skelton.

A previously unrecorded section in Newholme Beck is mentioned by Hemingway (1951,p.120), and the clay mineralogy of the ironstone at Eller Beck has been investigated (Hemingway and Brindley, 1952).

The Eller Beck Bed has recently been briefly studied by Bate (1967), who was chiefly concerned with zoning the Middle Jurassic strata of the area by means of ostracods. The Hydraulic Limestone was dated as belonging to the Hyperlioceras discites zone, and the Eller Beck Bed, though yielding no ostracods, was assumed to belong to the same zone as the limestone. The palaeogeography of the time was also considered.

B. ECONOMIC HISTORY

Although a few small quarries have been opened into the Eller Beck Bed sandstone, probably for material for dry stone walling, the only deposits which have been of any economic importance are the ironstones. The ironstone unit appears to have been mined by the Romans at Killing Pits (Bewick, 1861; p.99) near Goathland, where it is known as Julian's line (Fox-Strangways, 1892; P.197). Of a similar date are the bell pits at Holey Intake which have been sunk into a local ferruginous development of the sandstone unit. In modern times the ironstone has been tested near Grosmont, in Winter Gill, near Ingleby Greenhow, and on Snilesworth Moor (Fox-Strangways, 1892; P.450). The Winter Gill ironstone is another local development within the sandstone unit. Although the ironstone is commonly very rich in iron, variation in thickness and in character is too great for it to have been of any great importance in modern times

(Bewick, 1861; p.67, and Fox-Strangways, 1892; p.450).

C. AIM OF RESEARCH

1) To trace the Eller Beck Bed in the field over the whole of the area studied.

2) To study the lithological variation within each of the three main sedimentary units, and, briefly the variation in faunal assemblages.

3) To evaluate the conditions of deposition of the sediments of each unit, and to trace their diagenetic history.

4) To determine as far as possible the nature of the marine transgression and regression, together with the palaeogeography of the time.

D. AREAL EXTENT OF THE ELLER BECK BED.

With the help of several new sections it has been possible to extend the area of known outcrop as given by the Geological Survey. The probable complete outcrop is given in figure 2.1, together with the proved outcrop and all localities. The numbered localities refer to the section given in Appendix I. Availability of outcrop varies considerably over the area. The coast section and the dale heads south of the Esk provide the most numerous sections. North of the Esk exposures are rare, due to the lack of deep dissection of the moorland as is found in the south. In the west of the area outcrops are again poor due to the dip carrying the Eller Beck Bed up to a high level on the moor top.

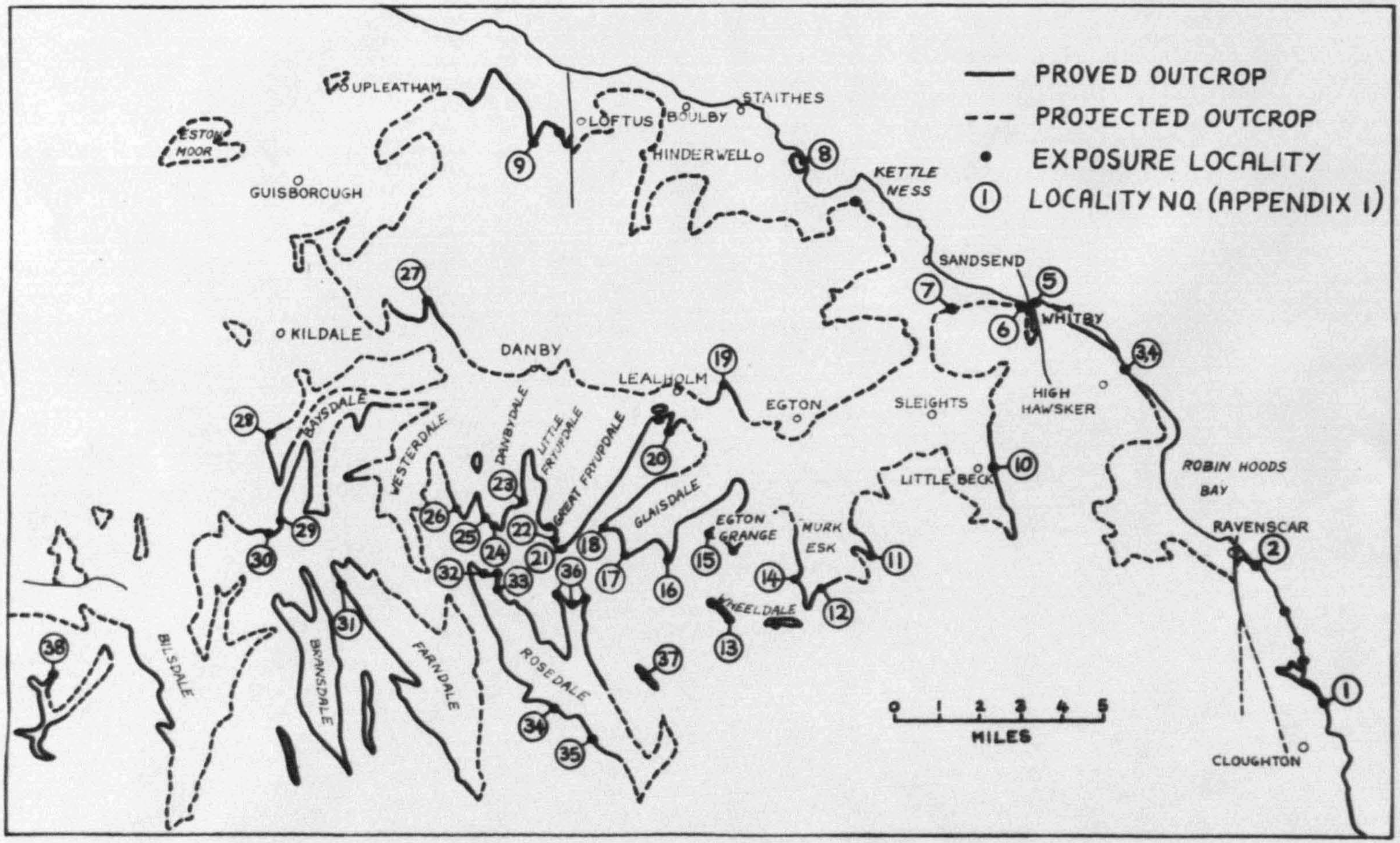


FIG 2.1 OUTCROP PATTERN OF THE ELLER BECK BED AND DISTRIBUTION OF EXPOSURES

CHAPTER IV THE IRONSTONE UNIT

I STRATIGRAPHY

A. FACIES DISTRIBUTION

Except in the north, where it becomes impersistent, the ironstone unit is a conspicuous though thin member of the Eller Beck Bed. The ironstone itself is commonly underlain by a shelly sandy siltstone which is so closely associated with the ironstone that it has been included in the same unit.

On the coast between Iron Scar (1) and Whitby (5) the ironstone unit comprises an oolitic siderite mudstone, with ooliths and terrigenous grains concentrated along laminae. At Iron Scar the ironstone is characterised by extremely abundant ironstone-filled burrows which extend from the base into the underlying non-marine shales. Scattered shells are present, and are entire. At several points along the section the ironstone has undergone erosion and the associated features will be described in detail at the end of this section. From Iron Scar northwards the ironstone becomes thicker, although locally removed by erosion, while the associated burrows become increasingly rare. Locally a second block of ironstone is developed above the main block and both have a similar lithology.

At Hawsker the ironstone is absent but the surface of the non-marine shale is penetrated by scattered ironstone-filled burrows. Locally (Section 3) even these burrows are absent where a broad trough has cut down to a maximum depth of about two feet (fig. 2.51), and in Whitby East Cliff (5) erosion has again removed the ironstone and the top of the non-marine shale, except where locally the ironstone is preserved and reaches a thickness of eight inches. In the West Cliff (6) the ironstone is greatly reduced in thickness with a maximum of two inches. The ironstone is a finely textured siderite mudstone with rare ooliths and shell fragments, contrasting with the shelly oolitic ironstone of the East Cliff. Burrows are not present at the base and two inches of dark sandy siltstone underlie the ironstone.

At Hinderwell (8) the ironstone and burrows are absent but in Newholme Beck (7) a fine textured pale siderite mudstone with abundant shells is present, overlying siltstone with local ironstone developments; burrows are present at the base. Specimens of shelly ironstone collected from Cat Beck, Kettleness though out of place were clearly nodular in type and it seems likely that the ironstone recorded by the Geological Survey (Fox-Strangways, 1892; p.195) is the same, and not the true basal ironstone. In Kilton Beck (9) the ironstone and thin siltstone are impersistent and locally show loading into the underlying sandstone. This sandstone is penetrated by irregularly distributed silt-filled burrows (probably _Taenidium_) in the top foot, and passes down into normal deltaic sandstones. The remaining section north of the Esk is at Commondale (27) where the unit is represented by a laminated silty shale passing up into shelly siltstone; shale-filled burrows occur at the base.

The most easterly section south of the Esk is at Littlebeck (10); the lowest marine unit is a thin dark siltstone with shell casts, but no burrows at the base. This is overlain by 2ft. 3in. of pale fossiliferous shale which becomes silty upwards and passes into a shelly silty sandstone. Near the base of the sandstone is a horizon of isolated dark grey fossiliferous ironstone masses, very similar in appearance to the basal ironstone of lower Eskdale. It is probable that the lower part of the section represents an early scour fill sequence and that the current involved was active throughout the deposition of the ironstone unit. The upward passage into shale suggests that these conditions continued into the shale unit. The ironstone masses may represent unconsolidated material derived from some nearby area of more normal deposition: this is supported by the unusual laminated nature of the ironstone and the absence of large shells. At Eller Beck (11) very shelly silty ironstone is developed overlying shelly siltstones with silt-filled burrows at the base. Comminuted shell material is abundant in the upper part and larger complete shells are more common in the lower part, including mud dwellers in growth position. The ironstone is similarly developed with minor thickness variations throughout Murk Esk (12,14) Wheeldale (13), Egton Grange (15) and Glaisdale (16, 17). To the north, in Crunkley Gill, Busco Beck (20) and Stonegate Gill (19) the ironstone becomes siltier with a small amount of fragmentary shell material; it attains its greatest thickness in this area.

In Great Fryupdale the ironstone is mostly cut out (21) but is locally preserved (22) where it is very shelly and calcareous, and rests directly on non-marine beds with a few ironstone-filled burrows. In Danbydale (24), Farndale (31) and along the Ingleby Escarpment (29, 30) the silts are similarly absent, as are burrows at the base; the ironstone is thin and becomes less shelly to the south and west. In Blow Gill the ironstone appears to be absent, but this cannot be

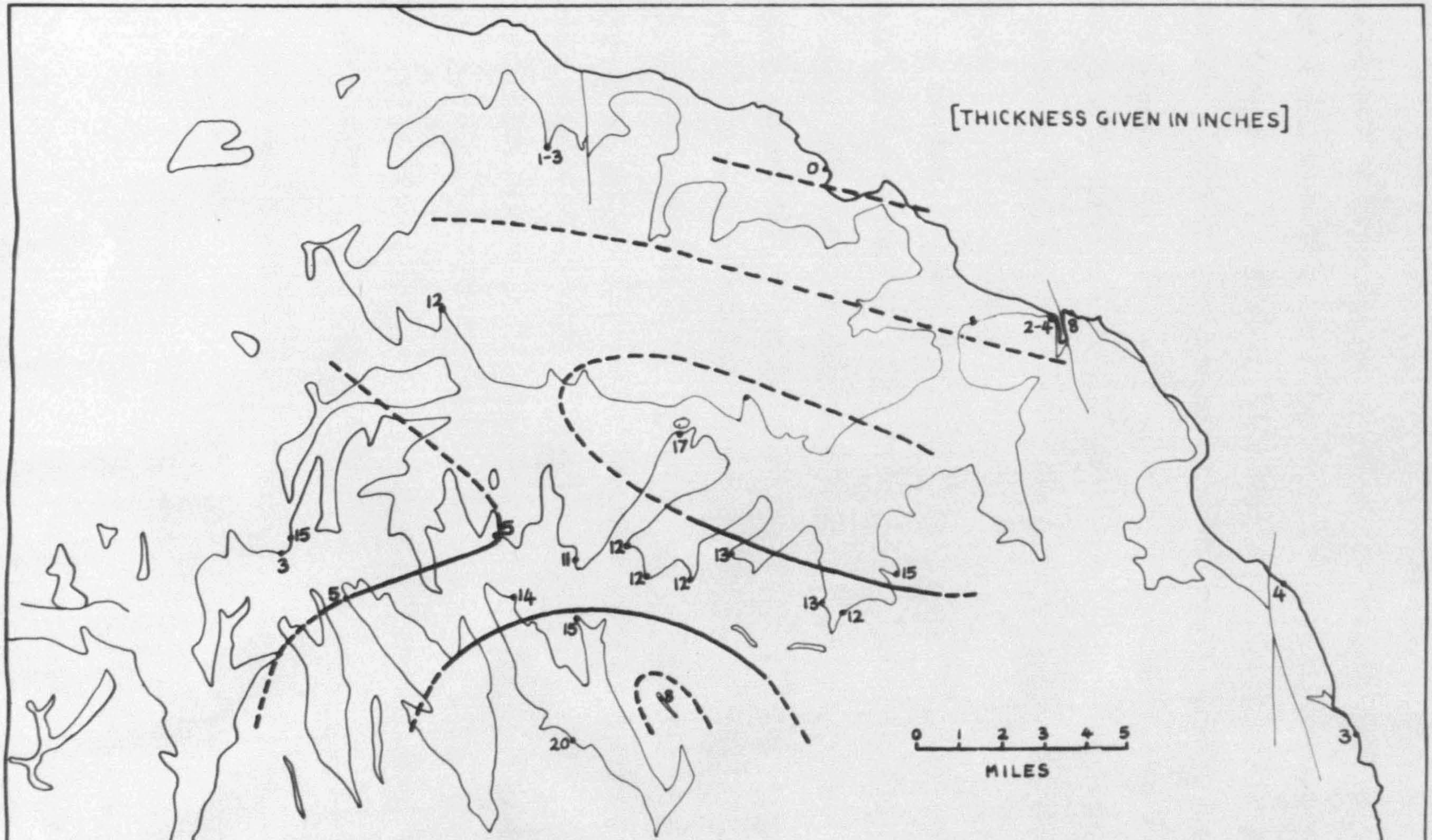


FIG. 2.2 ISOPACHYTE MAP OF THE IRONSTONE UNIT

certain until intermediate sections have been studied. In Battersby Crags (28) a siderite mudstone occurs beneath a thick sandstone and may represent redistributed material from the ironstone unit. A sandstone is locally developed beneath the ironstone which is burrowed by Taenidium. The sandstone differs from that of Kilton Beck in having a sharp lower contact with non-marine shales, and appears to be associated with the marine transgression although fossils are absent.

In Rosedale (32, 33, 34, 36) the ironstone unit is represented by a thick siltstone which is shelly in the north, and a thin impersistent ironstone. Burrows penetrate the underlying beds in northern Rosedale except where removed by local scouring. In Hartoft Beck the ironstone resembles that of lower Eskdale, with abundant fragmentary shell material and with burrows at the base. A thin oolitic layer occurs at the top, in which shells are rare; this very much resembles the ironstone of southern coastal districts.

The total thickness of ironstone and silt at each locality is given in figure 2.2, and isopachytes have been drawn to show the general trends. Figure 2.3 shows the distribution of siltstone within the ironstone unit.

The ironstone generally appears to be of a rather uniform nature, but where continuous sections are available on the coast considerable local variation is observed. At Iron Scar (1) the ironstone thickness varies due to erosion before shale deposition, resulting in the shale resting on a thin ironstone or on a surface of burrowed non-marine shale. The burrows are usually ironstone-filled but are locally silt-

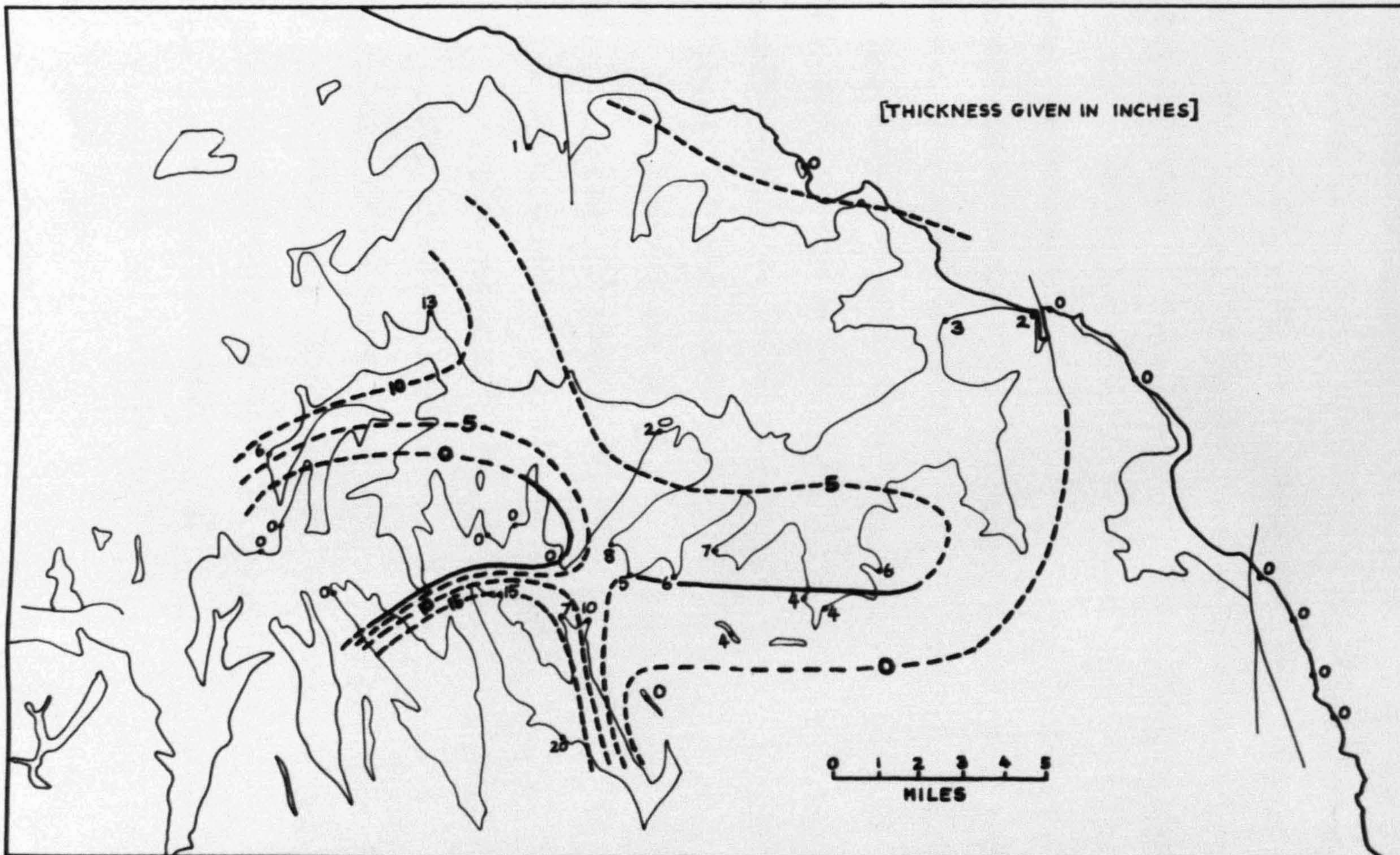


FIG 2.3 ISOPACHYTE MAP OF THE BASAL SILTSTONE

filled, indicating an earlier phase of erosion. In one case (fig. 2.4) a shallow trough has been eroded which is mostly silt-filled but includes a large mass of slightly oolitic ironstone with few fossils except for stem ossicles of *Pentacrinus* and *Turritella*, neither of which occur in the adjacent ironstone. A second example involves a larger area of scour (fig. 2.5) in which a mass of ironstone is present in a curious tilted fashion; the tilting appears to result from differential compaction in the underlying non-marine beds in which ironstone is locally developed. At several other points the ironstone unit is removed and replaced by very silty shale which passes up into normal shale. A tendency for the silt-filled burrows to be shorter and less abundant than the ironstone-filled burrows is attributed to erosion having removed the top few inches of the burrowed surface, thus eliminating the shorter burrows. Compaction around the ironstone has brought the eroded surface up to around the level of the original surface. The age relations of the burrows to the non-marine shale are to some extent shown by the development of slickensiding on ironstone filled burrows and by crumpling of silt-filled burrows. At Hawsker (3, 4) a somewhat similar feature is seen in the margins of the eroded trough, where the burrows are silt-filled; elsewhere they are preserved in ironstone although the ironstone bed has been removed throughout.

The features described above will be discussed later with respect to the diagenesis of the ironstone and its relation to the erosion prior to shale deposition.

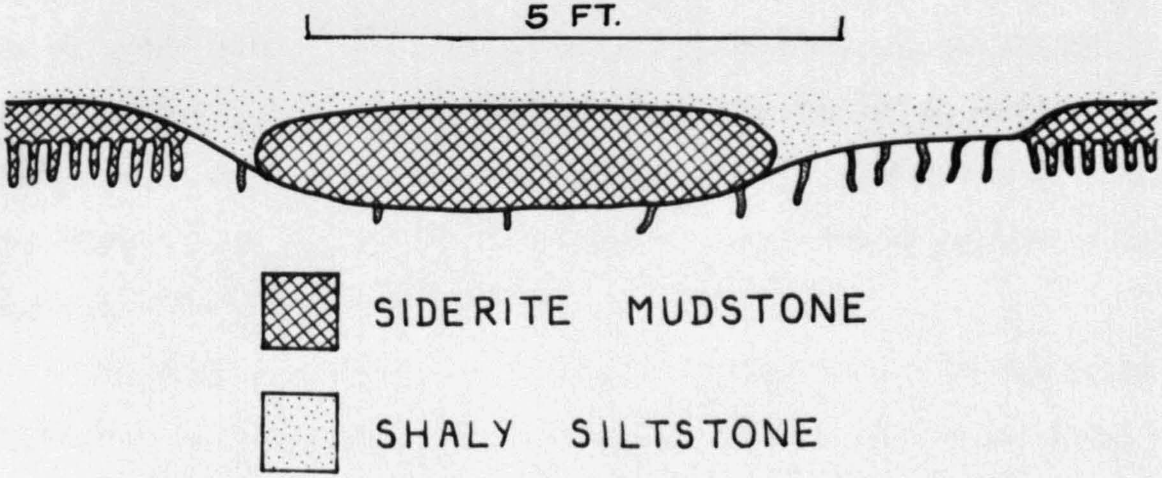


FIG. 2.4 DEVELOPMENT OF NODULAR IRONSTONE IN PLACE OF BEDDED IRONSTONE

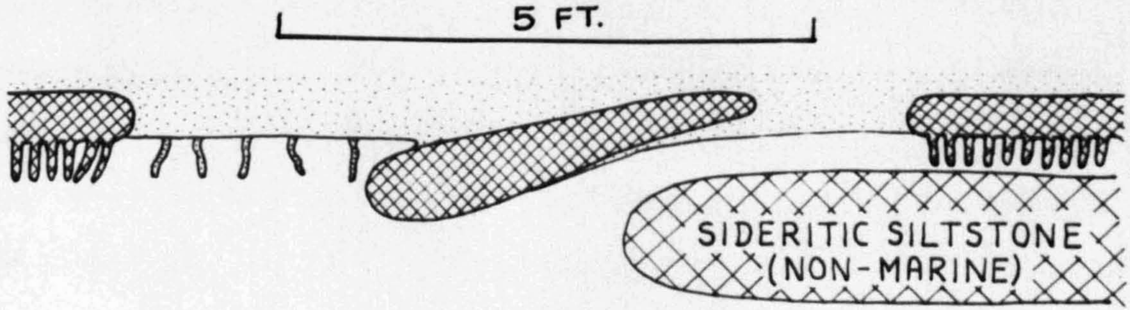


FIG. 2.5 NODULAR IRONSTONE DEVELOPMENT

B. PALAEONTOLOGY

The same basic faunal elements are present in varying proportions over the whole area. All are found in central and lower Eskdale while to the north, west, south and to some extent to the east, the faunal variation becomes reduced. The preservation of fossils as casts or moulds makes identification difficult, and for the purpose of this study the fauna has been considered in ecological groups rather than as individual genera or species.

Lamellibranchs constitute a large part of the fauna. Sessile forms include an abundant small oyster, Liostrea, together with larger forms such as Gervillia, Modiola, and Pinna. Active forms are an abundant small Astarte, Protocardia, Trigonia, Nuculana, and several other genera. The burrowing lamellibranchs Pholadomya and Pleuromya are characteristically found in vertical living position, and their distribution is shown in figure 2.6. Gastropods are represented mainly by a small form of Turritella. Other classes of mollusc, together with brachiopods, are absent. Ostracods have been observed in a few samples, but are probably widespread since they are commonly seen in thin section; alteration appears to make them difficult to distinguish from the matrix.

In central and lower Eskdale shells are abundant and all of the known forms are represented. Comminuted shell material is abundant, especially in the upper part, while burrowing forms may occur at any level. On the coast south of Whitby only larger complete shells are present, and small forms are rare; burrowing forms occur in the East Cliff of Whitby but are not seen to the south, where the fauna as a whole is reduced in numbers. In the West Cliff shells are rare and

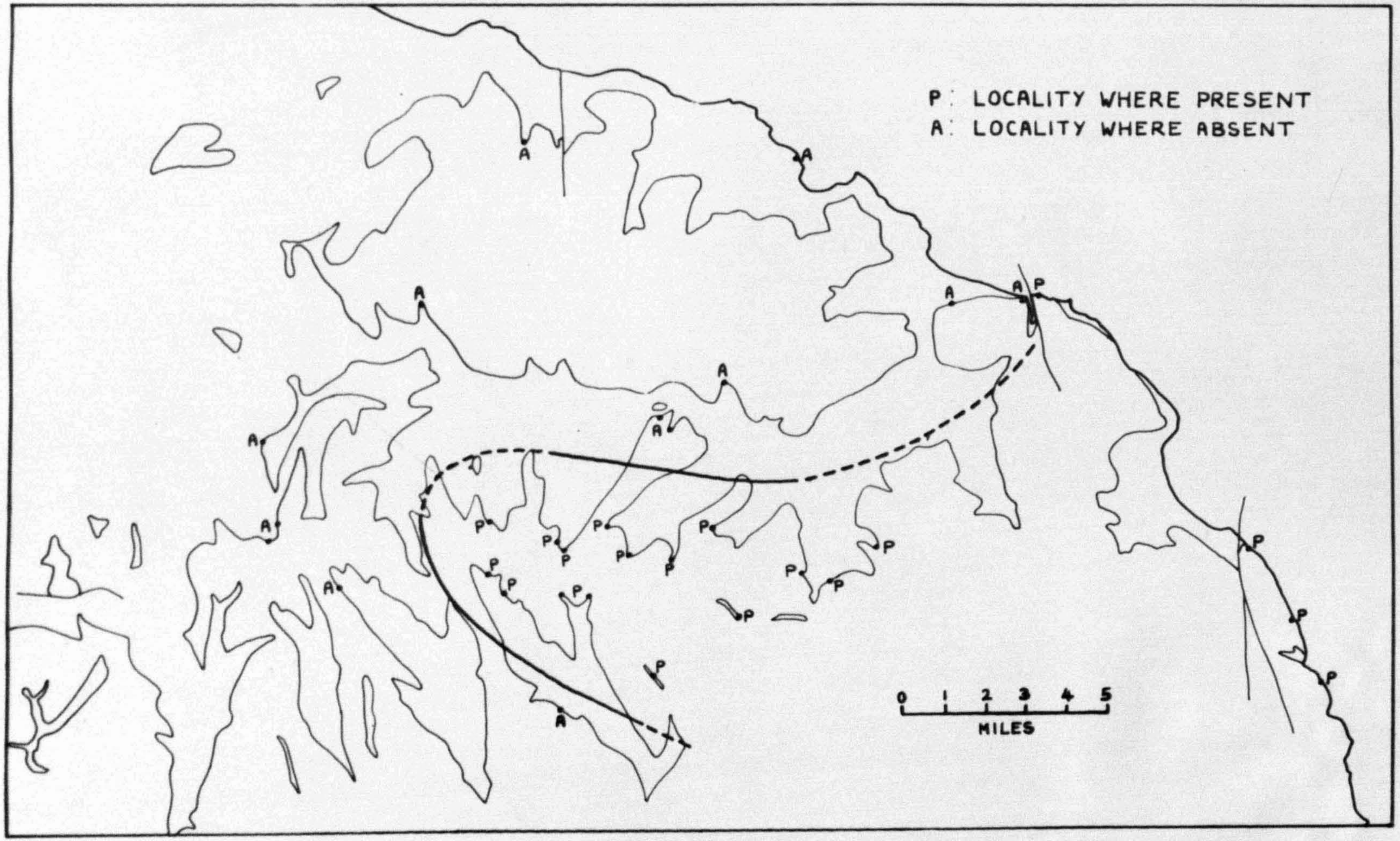


FIG 2.6 IRONSTONE UNIT: DISTRIBUTION OF BURROWING LAMELLIBRANCHS

fragmentary and to the north distribution of shells is patchy, with oysters being abundant and burrowing forms absent.

In Rosedale the ironstone contains small and fragmented shells but in the north the siltstone is very fossiliferous and contains the fauna characteristic of the ironstone in Eskdale, including burrowing forms. This contrasts with the siltstone of other areas which contains only scattered oysters, and suggests that these silts are equivalent to the ironstone of Eskdale. In Great Fryupdale shells are very abundant while to the north only small forms occur; to the west larger entire shells are present but become increasingly rare to the south-west where oysters become predominant.

In addition to the shelly fauna the ironstone unit is characterised by the presence of burrows, particularly at the base. These are preserved in siltstone or ironstone depending on the nature of the overlying sediment and are usually conspicuous in their non-marine shale setting. Burrows are most numerous and abundant at Iron Scar where six types have been distinguished:

(i) Vertical U-tubes with spreiten; large, 4in. by $\frac{3}{4}$in. in cross-section, and about 3in. long (Corophioides).

(ii) As above, but tubes inclined or horisontal and up to at least 6in. long (Rhizocorallium)

(iii) Vertical U-tubes with spreiten; small, $\frac{3}{4}$in. by $\frac{1}{4}$in. in cross-section and up to 8in. long (Diplocraterion).

(iv) Horisontal or gently inclined U-tubes; small, 2in. by $\frac{1}{4}$in. in cross-section; length uncertain (Rhizocorallium)

(v) Horisontal anastomosing tubes without spreiten, diameter between

½in and 1½ in (*Thalassinoides*).

(v) Horizontal tubes without spreiten, diameter about ½in. (not identified).

All but type six bear scratch marks on the surface, indicating crustacean activity (see under Rhizocorallium, Moore, 1962 p.W211). The interpenetrating relationships of the various types enables a rough order of colonisation to be drawn up:

(i) and (ii)
(v)
(vi)
(iii) and (iv)

To the north of Iron Scar the variety in type is greatly reduced, and throughout the greater part of the area only types (iii) and (iv) are present. The distribution of these burrows is shown in figure 2.7. In the extreme northern and north-eastern localities the burrows described are not developed, but in Kilton Beck *Taenidium* occurs in the top non- marine sediments; it also occurs at Battersby Crags. Over the whole area the ironstone and siltstone may show small-scale burrowing by *Chondrites*.

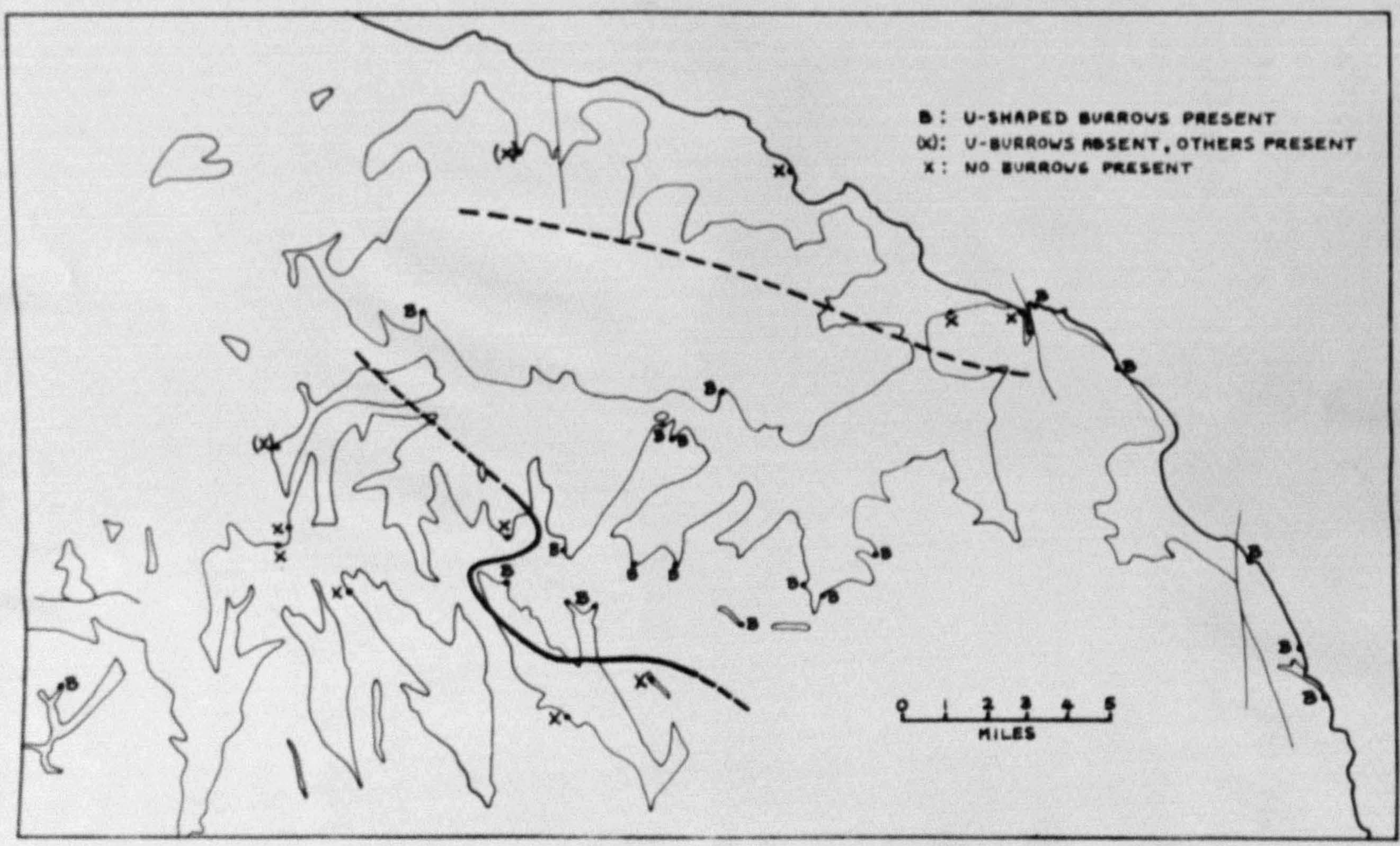


FIG 2.7 IRONSTONE UNIT : DISTRIBUTION OF BURROWS AT BASE

II PETROGRAPHY

The sediments of the ironstone unit comprise siderite mudstone and in many localities an underlying siltstone or sandstone. The latter are incoherent and usually weathered, and have not been studied in detail. They consist of terrigenous grains with a small amount of clay matrix together with cavities which appear to represent ooliths and shell material which have been removed by solution. The ironstone on the other hand is close-grained and can usually be obtained in a fresh state suitable for the preparation of thin sections, and it is with the ironstone that this chapter will be primarily concerned. The ironstone comprises three major constituent types: a matrix composed mainly of siderite, terrigenous silt, and sand grains, and allochemical grains derived from within the basin of deposition including ooliths, shells, and intraclasts. These constituent types are dealt with below.

A. MATRIX

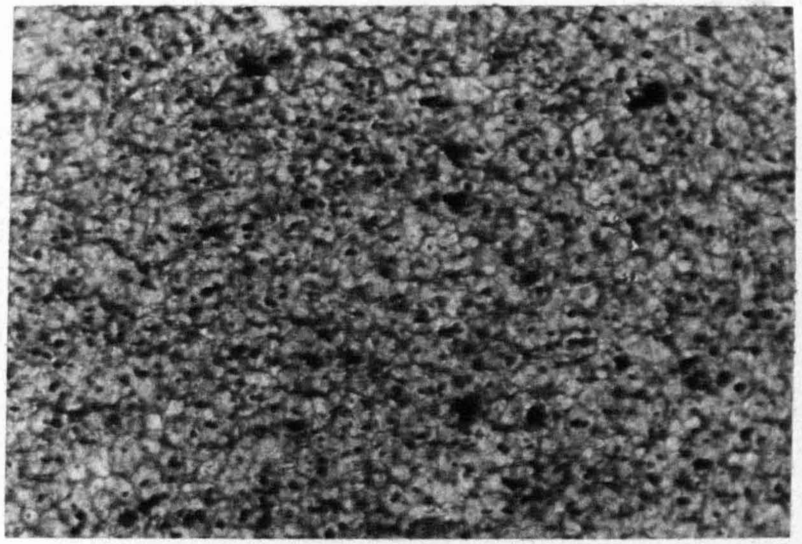
1. Siderite

The grain size of the siderite matrix varies between 3 and 50 microns, but crystals between 5 and 15 microns in diameter are most common. The crystals are anhedral or occasionally subhedral and usually appear to be tightly interlocking. Normally the crystals are equant or slightly elongated but in some burrows large subhedral crystals with elongations of up to 3 μ are present. The crystal size within any one specimen is often uniform but occasionally the mudstones contain laminae of contrasting crystal size which clearly reflect an original sedimentary fabric. Areas of mud enclosed by shell material or occurring as burrow infillings generally differ in crystal size from the general matrix.

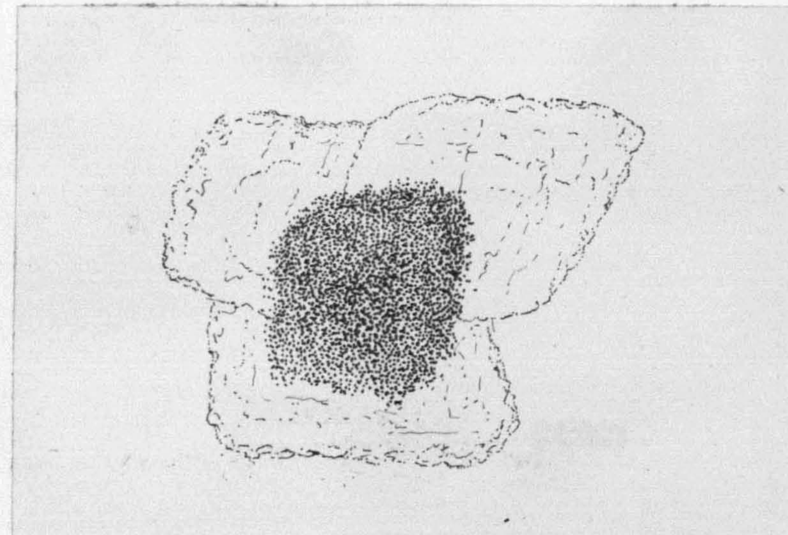
A characteristic feature of every sample studied is the presence of small, apparently discrete, particles within many of the siderite crystals (fig.2.8). Each particle occupies a central position within a single siderite crystal or, occasionally, within a group of crystals, and will be referred to as a 'core'. The cores are equant to lozenge shaped with a rounded outline; the elongation of the core usually corresponds to that of the host crystal, but is very rarely perpendicular to it. They vary in length between 2 and 5 microns, but are usually more uniform in size in any one section, the size of the core being independent of the size of the enclosing crystal (fig.2.14). The cores may be translucent but are usually opaque. When viewed at high magnification the core boundary is generally sharp, but occasionally the outer part or even the whole core is diffuse. On a small scale the distribution of cored crystals relative to uncored crystals may vary considerably, the percentage of cored crystals ranging from zero to perhaps eighty. In some rocks in which the crystals appear to be uncored the weathered zone produces dark brown turbid inner zones, which may represent cores that have been originally clear or which have been resolved into the surrounding crystal (fig.2.12).

In addition to the normal development of a single core within each crystal, relatively large core-like bodies may show radial development of several crystals. This is occasionally seen in coarser matrices (fig.2.9) but generally occurs in finely crystalline siderite within phosphatic cement. Another type of siderite development is found

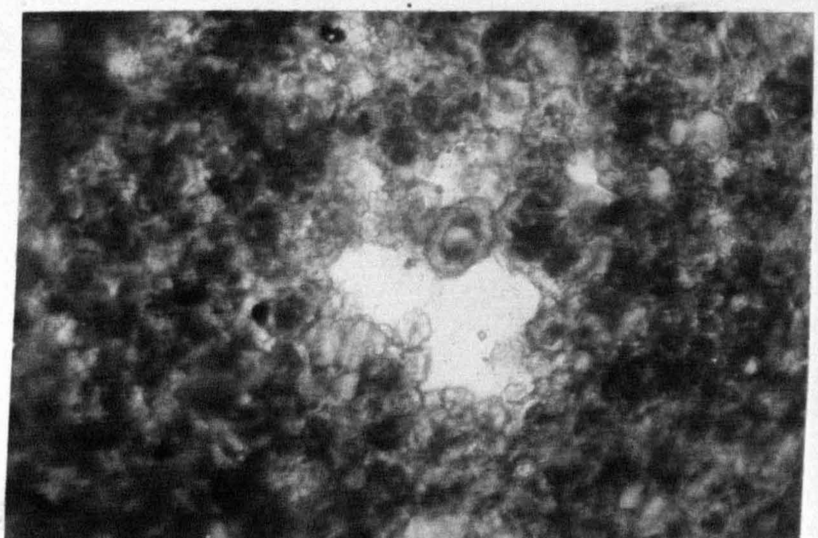
FIG. 2.8	Matrix of siderite crystals containing opaque cores	(x225)
FIG. 2.9	Several crystals growing around a single core	(x2500)
FIG. 2.10	Overgrowths of clear siderite on earlier, rather oxidised crystals	(x400)
FIG. 2.11	Overgrowth of euhedral siderite on elongate zoned crystal.	(x550)



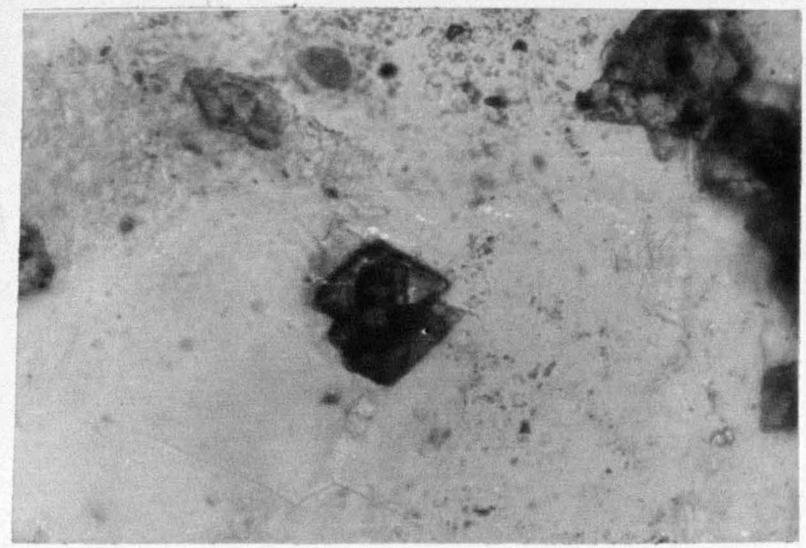
2.8 50μ



2.9 10μ



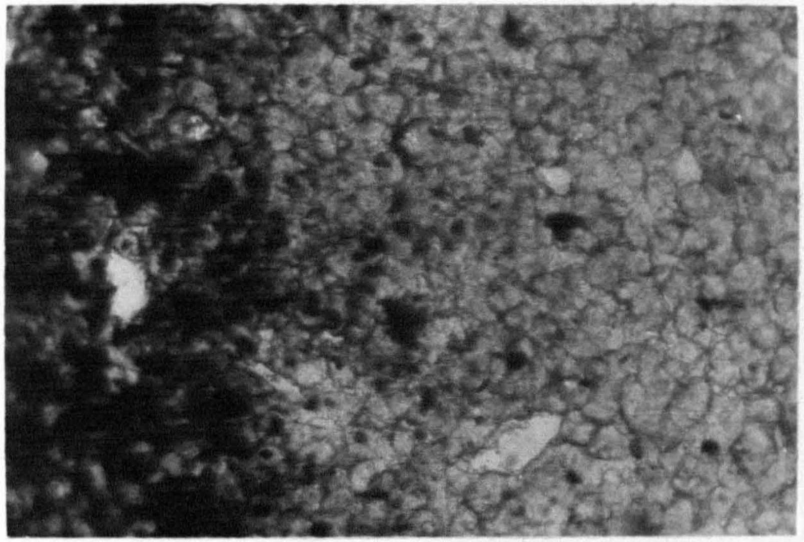
2.10 50μ



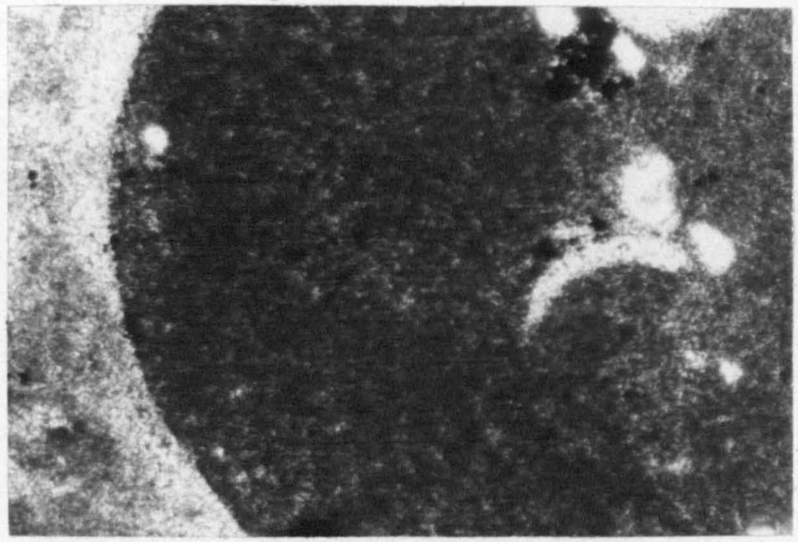
2.11 50μ

within the sandy laminae of the siderite mudstone of Battersby Crags. In the associated mudstone the crystal size varies between 5 and 20 microns and the larger crystals which are oval in shape, show large, opaque cores. Cored crystals of identical appearance are found within the sandy laminae where the smaller crystals are absent, and these show euhedral rhombic overgrowths of clear siderite (fig.2.10). Overgrowths also occur on a few siderite grains of a type not found within the associated mud; these are elongate grains with rounded cores enclosed within relatively turbid siderite, and the overgrowths consist of clear siderite of euhedral rhombic habit. The overgrowths are unusual in that the c-axes of the rhombs are perpendicular to the elongation of the host crystal, and two or more rhombs make up the overgrowth (fig.2.11). Both types of overgrowth occur only where the matrix comprises drusy kaolinite, and are absent in the presence of a siderite mudstone matrix.

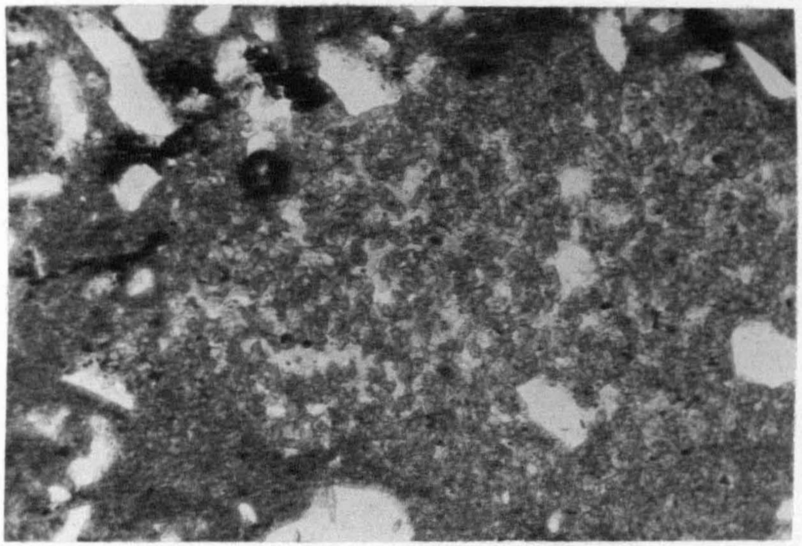
The nature of the siderite matrix shows regional variation to some extent. Finely crystalline matrices are characteristic of north-eastern localities, and two matrix types may be distinguished. In one type the matrix is turbid and extremely fine-grained, the turbidity being due to abundant opaque bodies identical to the cores already described (fig.2.13); in the other the siderite is clearer and slightly more coarsely crystalline, while cores are less abundant and less distinct than in the turbid matrix. The boundary between the two types may be gradational or sharp, and may follow distinct sedimentary structures, such as laminae, shells, or burrow margins, or may be related to the outline of the ironstone surface. A characteristic feature of the



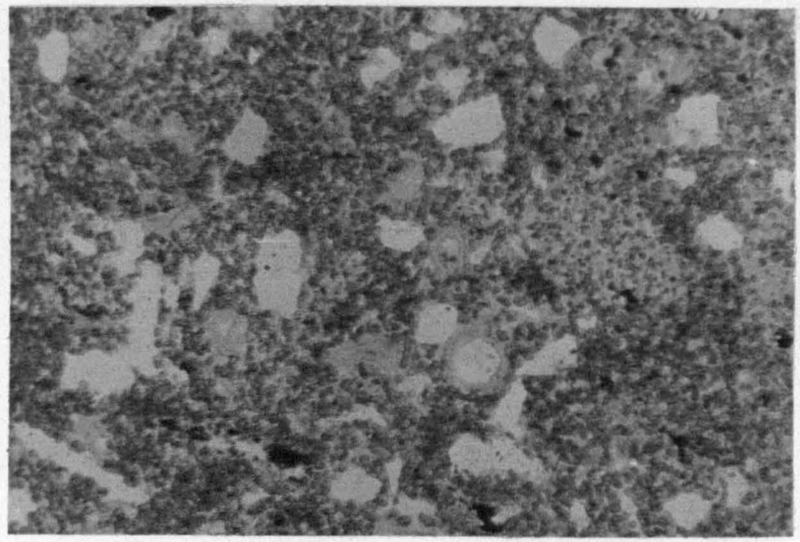
2.12



2.13



2.14



2.15

turbid siderite matrix is the development of a very fine network of septarian cracks (see p. 117). More coarsely crystalline may occur locally in these finely crystalline mudstones, the principle sites being shell interiors and burrow infillings. The former may contain a coarse granular mosaic of cored or uncored crystals; the latter generally include elongate crystals reaching 50 microns in length which are set in a matrix of clay (fig.2.14).

In the coastal sections south of Whitby Harbour the matrix comprises equigranular clear siderite crystals with diameters between 10 and 15 microns, and with only occasional cores. The internal mould of a Pholadomya in growth position collected from the East Cliff of Whitby, however, consisted of pale coloured fine grained mudstone in which several siderite crystals are clustered around each core. This mudstone later proved to be phosphatic (p.115). In Eskdale the matrix consists of rather turbid equigranular siderite crystals in which cores are fairly abundant and well defined. Cored crystals of subhedral habit occur in the ironstone of Hartoft Beck, where the crystals are not interlocking in parts due to an interstitial clay matrix (fig.2.15). The ironstone of the south-western localities is too highly weathered for the preparation of thin sections.

2. Clay

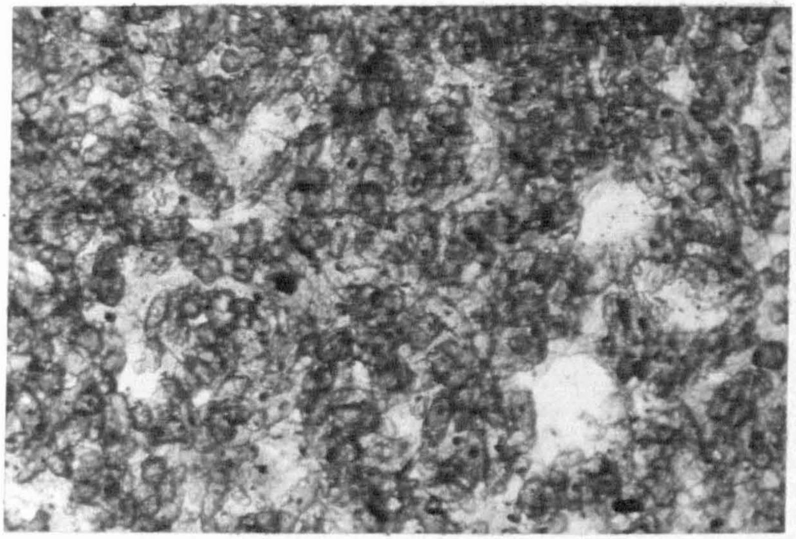
The matrix of the ironstone as observed in thin section usually consists entirely of siderite, and Hartoft Beck is the only locality in which a clay matrix is visible apart from in burrow infillings. Clay constitutes 10 percent of the rock and locally as much as 60 percent in burrows. It consists of slightly turbid greenish grey chamosite with scattered opaque dusty inclusions, and under crossed nicols is seen to comprise a disoriented mass of fine, flaky crystals with low or very low birefringence. The chamosite is locally replaced by small colourless kaolinite aggregates. Similar clay occurs interstitially to siderite in burrow infillings in north-eastern localities.

A small amount of kaolinite appears to be present along siderite crystal boundaries in the most sideritic matrices, since a kaolinite peak is obtained on x-ray diffraction traces of the hydrochloric acid insoluble residues. This clay is also encountered in the preparation of acetate peels, when some ironstones, particularly the finely crystalline pale coloured types yield a large amount of finely divided clay on etched surfaces. Variations in clay content on a small scale are detected in this way, as where the ironstone is laminated or burrowed.

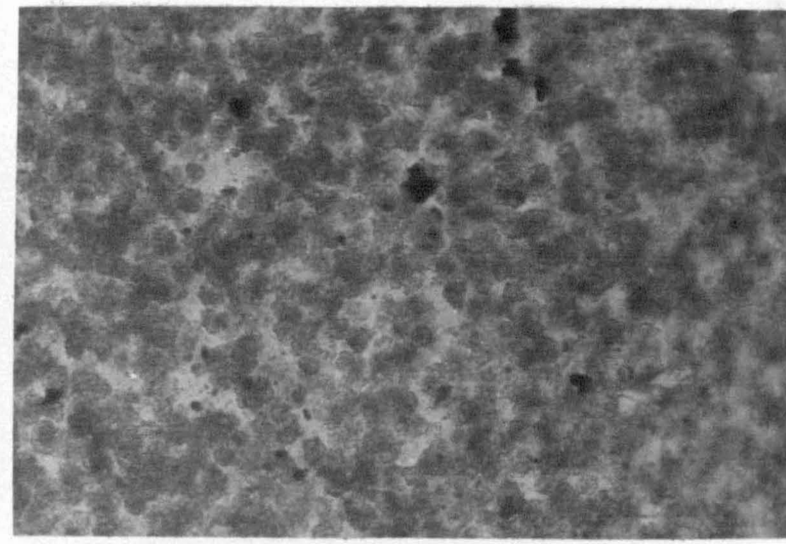
3. Phosphate

Phosphate is a major constituent in only a few samples from north-eastern localities, and may be associated with the general matrix or with specialised local environments such as burrows and the phosphatic shell mould already mentioned (p.114). The phosphate comprises isotropic brownish homogeneous material which is interstitial to siderite crystals when developed. The siderite crystals are usually in the form of poorly defined aggregates (fig.2.16). Phosphate may constitute as much as 50 percent of the matrix in some burrows, and in a few cases siderite may be absent (fig.2.17). Chemical analysis (Appendix 2, D) was carried out on the phosphatic internal shell mould and on the surrounding

FIG. 2.16	Elongate crystals in figure 2.14 showing opaque cores	(x170)
FIG. 2.17	Siderite crystals with irregular outer margins set in phosphatic matrix.	(x380)
FIG. 2.18	Disseminated pyrite replacement.	(x70)
FIG. 2.19	Massive pyrite replacement.	(x80)

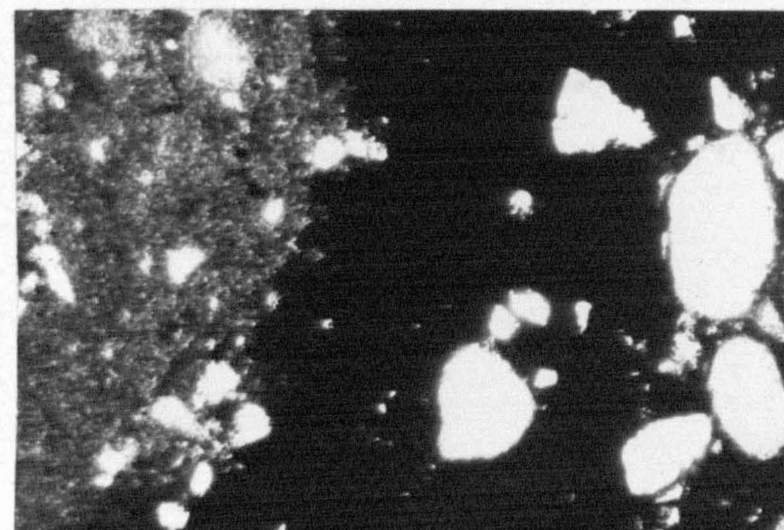


2.16 50μ



2.17 50μ

2.18 500μ



2.19 500μ

ironstone and the results are given below:

1. Shell mould 17.86% P_2O_5

1. Normal ironstone 1.20% P_2O_5

The figure of nearly 18 percent is equivalent to about 30 percent of calcium phosphate a figure which agrees with the estimated bulk percentage of phosphatic matrix. The figure of 1.2 percent corresponds to those given for comparable siderite mudstones (Fox-Strangways, 1892; pp.443-451). The shell interior has clearly acted as the locus for a considerable concentration of phosphate. The phosphate appears to be in the form of collophane throughout.

4. Pyrite

Pyrite occurs in disseminated form as small irregular patches with the siderite mudstone in coastal sections. It is commonly, but not always, associated with local concentrations of grains (fig.2.18). The pyrite consists of anhedral crystals about five microns in diameter which may form aggregates or even larger areas of 'massive' pyrite in which almost total pyritisation of the rock has taken place (fig.2.19).

5. Calcite

Ferroan calcite occurs as a cementing material in the shelly calcareous ironstone of Great Fryupdale. The cement is rare and is developed only within algal borings into shell fragments and in cavities produced by shell fracture during diagenesis.

6. Depositional Fabric

The ironstone is for the most part homogeneous, but in inland sections vertical variation in shell and sand content is common. In the West Cliff of Whitby the mud is laminated with contrasting types of siderite matrix, and to the south of Whitby the predominantly muddy ironstone contains ooliths and terrigenous grains concentrated along laminae.

7. Post-depositional Fabric

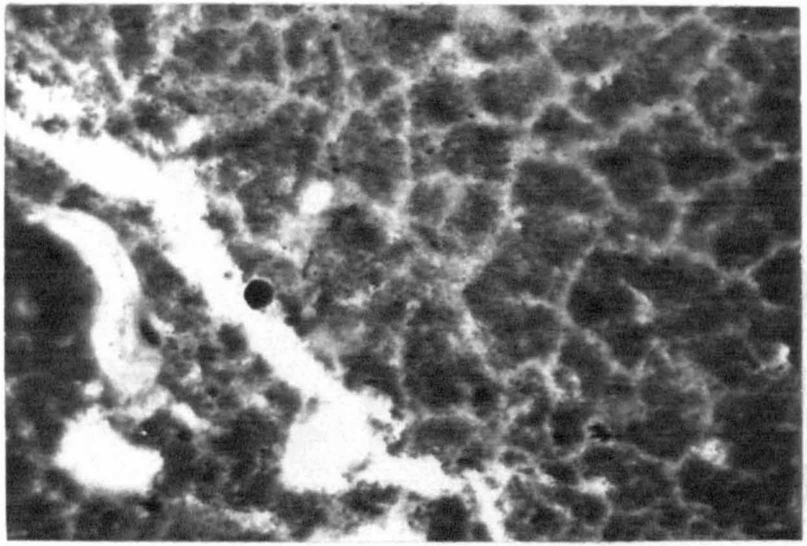
a) Burrows. In thin section only one type, possibly Chondrites, is prominent; the burrows appear circular or oval, depending on the section, with diameters varying between 250 to 500 microns. The burrows are distinguishable by their contrast to the surrounding matrix, and generally show a high clay or phosphate content.

b) Micro-septarian Structure. This term has been applied here to the network of very fine cracks which are characteristic of the dark fine-grained mudstone facies (fig.2.20). The cracks seldom exceed 10 microns across and are filled with siderite which is clearer and a little more coarsely crystalline than the surrounding matrix.

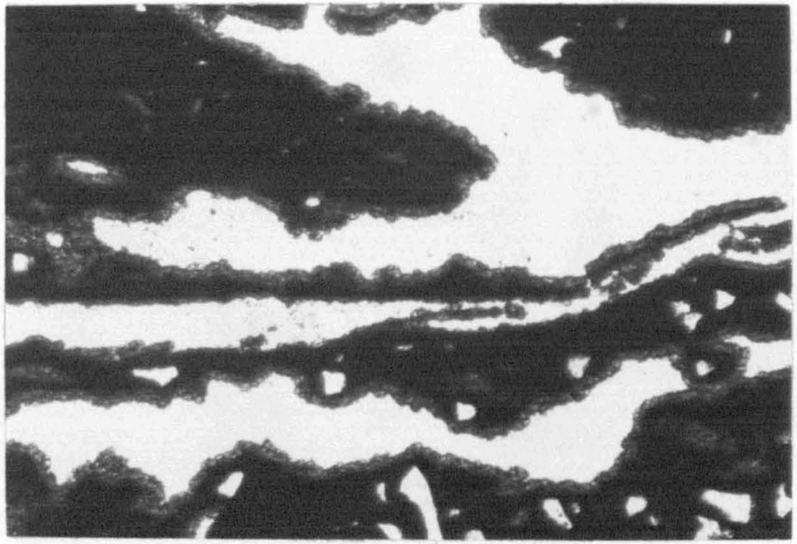
c) Septarian Structure. This is developed in several samples in which the siderite matrix is finely crystalline. Two varieties may be distinguished:

(i) One type is characterised by the irregular nature of the fissure walls and by the rapid 'pinching out' of the cracks. These cracks are usually associated with shell fragments (fig.2.21) and to a lesser extent with other grains. The shell surface may act as the plane of parting but more often a thin layer of mud lies between the shell and the crack; this layer is left unsupported when shell solution takes place and often suffers breakage. The maximum width of the cracks is about 350 microns and they extend for nearly the entire length of the shell; short and rapidly tapering transverse

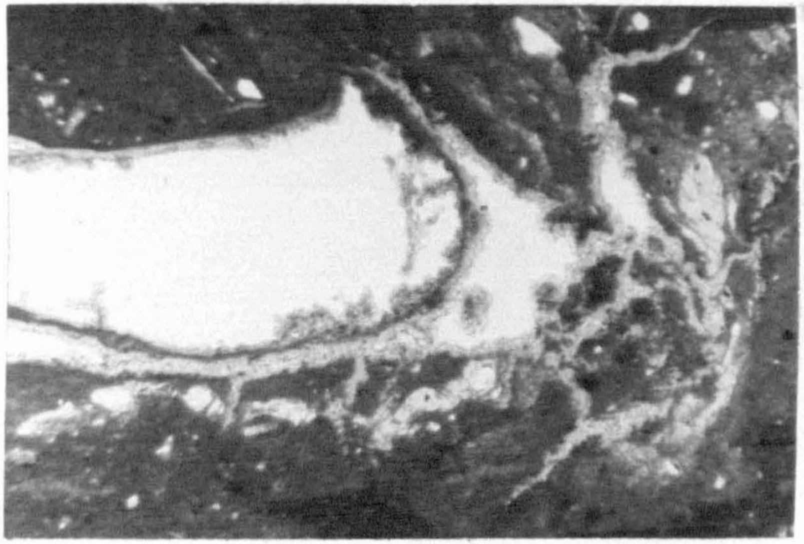
FIG. 2.20	Microseptarian structure	(x150)
FIG. 2.21	Type (i) septarian crack associated with shell margin.	(x55)
FIG. 2.22	Transverse terminal septarian crack	(x70)
FIG. 2.23	Type (ii) segarian crack cutting across shell and associated type (i) crack	(x55)



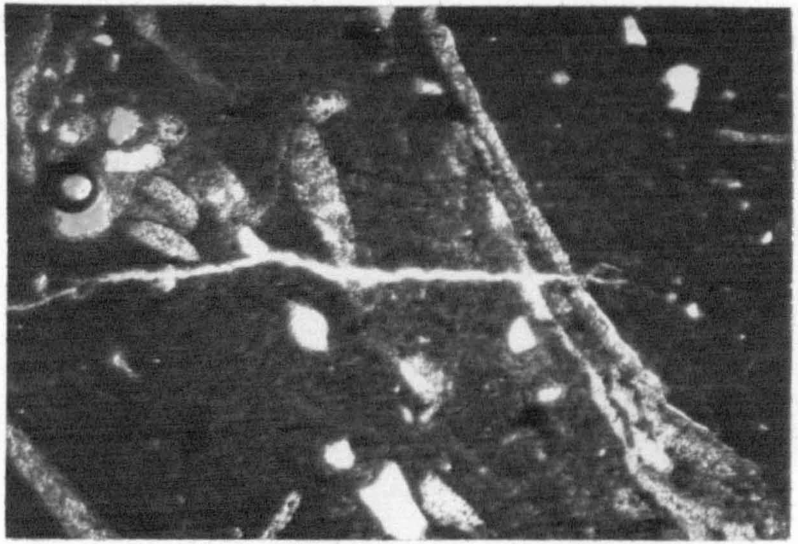
2.20 1500 μ



2.21 500 μ



2.22 500 μ



2.23 500 μ

cracks may develop and at the ends of a shell a highly irregular ramifying network of cracks is developed usually with a prominent transverse member (fig. 2.22). Where the septarian sturcture is not related to a grain it is usually developed as a series of short cracks arranged in echelon. The cracks possess a rim of drusy siderite and the central portions are filled by kaolinite. In Newholme Beck where the structures are best developed these septarian structures comprise four percent of the volume of the rock.

(ii) The second type is characterised by smooth and relatively extensive fractures which cut across all previous structures, including shell fragments and the first type of septarian crack together with the drusy siderite rims, (fig.2.23). These cracks lack siderite rims but are filled by kaolinite which is apparently continuous with that of the other septarian structures.

8. Diagenetic History

Since the siderite crystals seen to replace chamosite mud in Hartoft Beck are identical to those of some other localities it is probable that the original sediment was a chamositic mud throughout and has been replaced by siderite during diagenesis. The unusually finely crystalline matrix of the ironstone in north-eastern localities is attributed to the presence of interstitial phosphate which appears to have been an early diagenetic mineral which has inhibited growth of siderite crystals. In the case of the shell mould described on page 114 the mud within the mould is sandy and oolitic and almost certainly had the same original composition as the surrounding mudstone.

The siderite matrix shows little subsequent alteration; the relatively coarsely crystalline siderite in some matrices may be a product of recrystallisation but the presence of single sharp cores in each crystal suggests that the crystal size is an original feature. In some coastal sections the matrix shows local patches of partial or total pyritisation which appears to be of fairly late formation. The development of septarian structures presumably occurred immediately before final lithification; they are developed only where the ironstone is rather nodular in character and must be the result of shrinkage of the matrix within a resilient outer zone. The fractures are of two generations, one preceding the formation of drusy siderite rims and the other later than this; both occur before deposition of drusy kaolinite. During the first phase of cracking the mud appears to have had slight plasticity while in the second phase the mud has fractured in such a way as to suggest nearly complete rigidity. Siderite was deposited between the two phases and kaolinite at a later stage, when shrinkage was complete.

B. TERRIGENOUS GRAINS

1. Composition

The terrigenous grain fraction of the ironstone consists almost entirely of simple quartz grains, with less than five percent of other types, which include compound quartz grains, feldspars, and micas. The feldspars include sodic plagioclase, orthoclase, and microcline, all of which are very fresh. The micaceous grains are mostly muscovite, though a little chlorite may be present.

2. Distribution.

Percentages of terrigenous grains were determined by point-counting, five hundred counts per section being made to allow for inhomogeneity of the sediment. Regional variation in these percentages is shown in figure 2.24. The ironstone in Danbydale, Farndale and along the Ingleby Escarpment, though highly weathered, may be seen to have a low proportion of terrigenous grains. At Commondale and in Rosedale the ironstone is absent or impersistent, but at the former the unit has a high clay content. The two figures for Hartoft Beck refer to the upper and lower parts of the ironstone. From the distribution map a clearly defined coastal strip may be observed in which a low percentage of grains is present. In much of Eskdale a consistently high proportion of grains occurs, which probably increases to the south but decreases to the west and north. The upper part of the ironstone in Hartoft Beck has affinities with the southern coastal development, while the lower part resembles the Eskdale ironstone.

3. Grain parameters.

The grains were measured along linear traverses using a manually operated moving stage, care being taken to cover as large an area of the section as possible. Practically all grains belong to the coarse silt and very fine sand classes, but the northern and eastern sections show the presence of a coarse fraction which is superimposed on the modal class. Although this coarse fraction decreases the sorting, the average grain size is still lower than in Eskdale, where the modal class is coarser than on the coast. The size frequency distributions of the sections are included in figure 2.50.

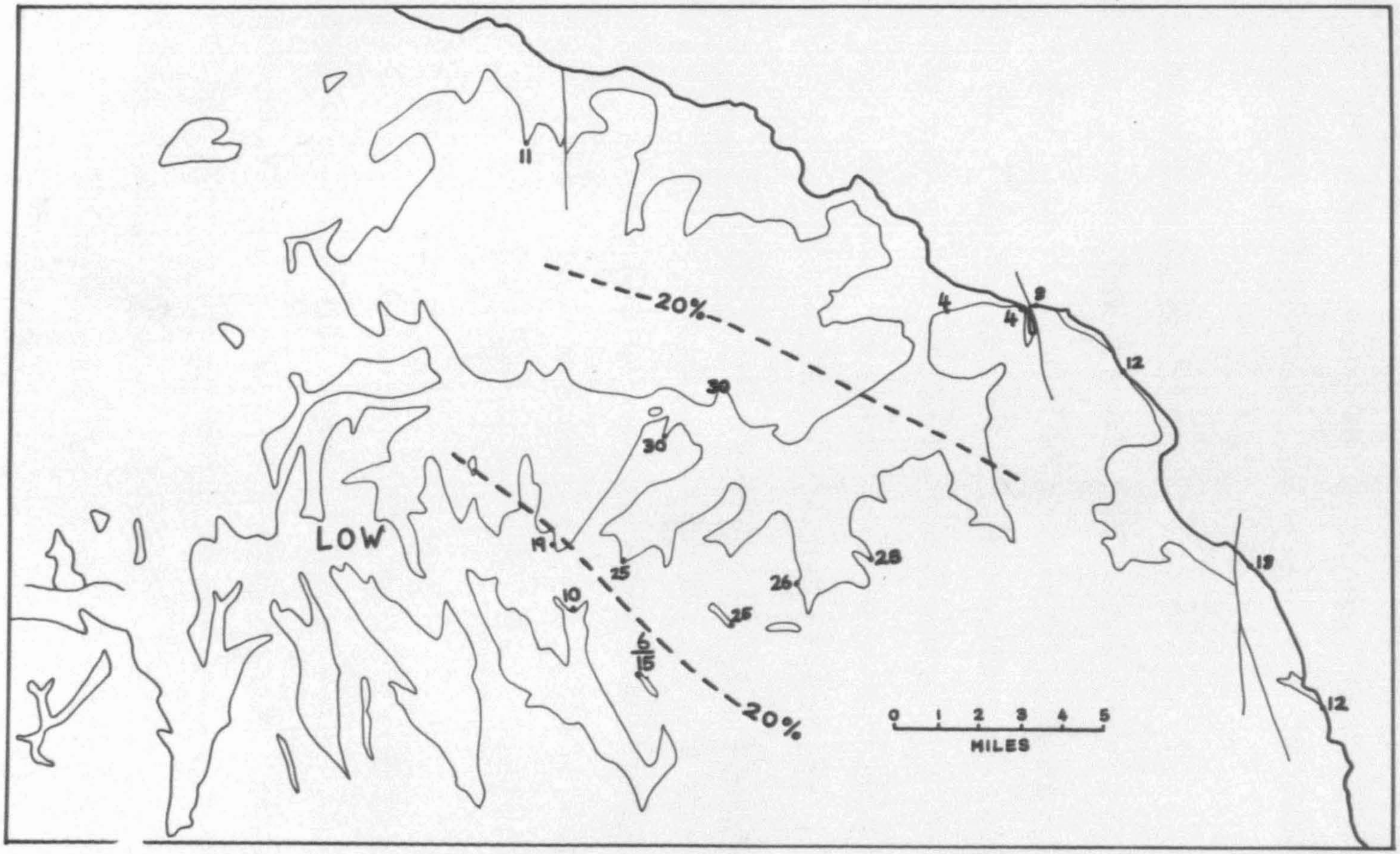


FIG. 2.24 REGIONAL DISTRIBUTION OF SAND GRAIN PERCENTAGE IN BASAL IRONSTONE

The grains frequently have an irregular outline due to etching (see later) and commonly show a low sphericity, with elongation exceeding 1:4 in some cases. Where etching has not occurred the grains are angular.

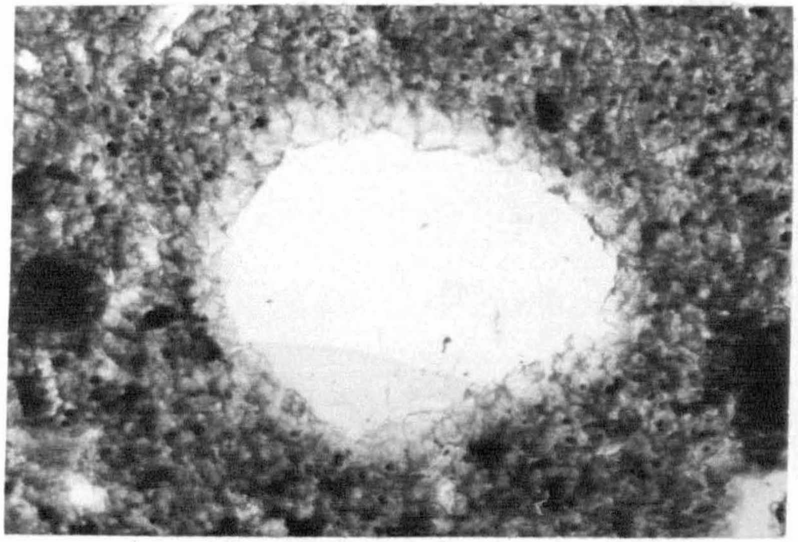
4. Replacement of Quartz and Feldspar

a) Siderite. Replacement of quartz grains by siderite is difficult to detect since a small amount of siderite is commonly superimposed on the margin of the quartz grain owing to the thickness of the section. Many grains, however, have highly irregular margins which are clearly not sectioning effects. The irregularities consist of projections and deep indentations, often with a rather smooth outline, and the siderite grains which are in contact with the grain margin are brownish, turbid, and often cored, and are indistinguishable from those of the matrix (fig. 2.26). A great deal of corrosion of quartz and feldspar grains which superficially appears to result from replacement by siderite may, therefore, have taken place before the final crystallisation of the siderite matrix.

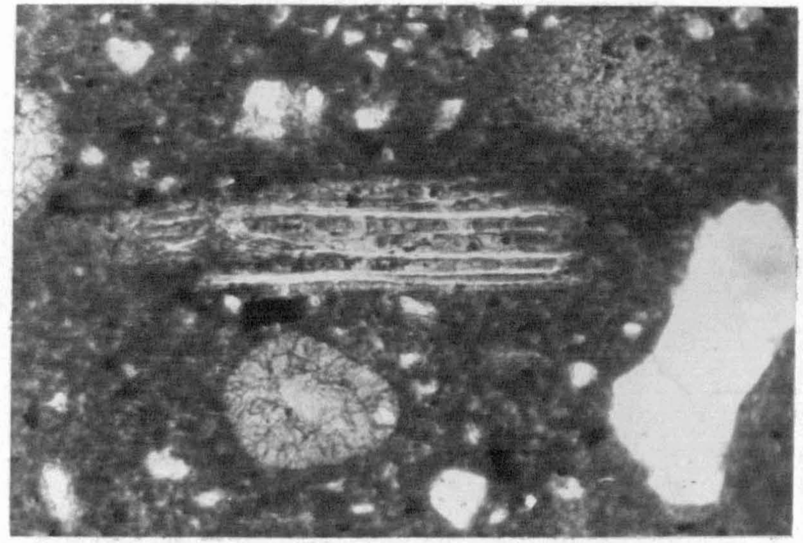
More definite evidence of replacement by siderite is seen when quartz grains are penetrated by clear and often euhedral siderite crystals, easily distinguishable from those of the matrix. The original outline of the grain may be seen in the boundary between the matrix and the replacing siderite. Such replacement is not common.

b) Calcite. In the calcareous facies of Great Fryupdale quartz grains are sometimes partly replaced by ferroan calcite. Compound grains in particular are affected, the replacement being initiated along the intergranular sutures.

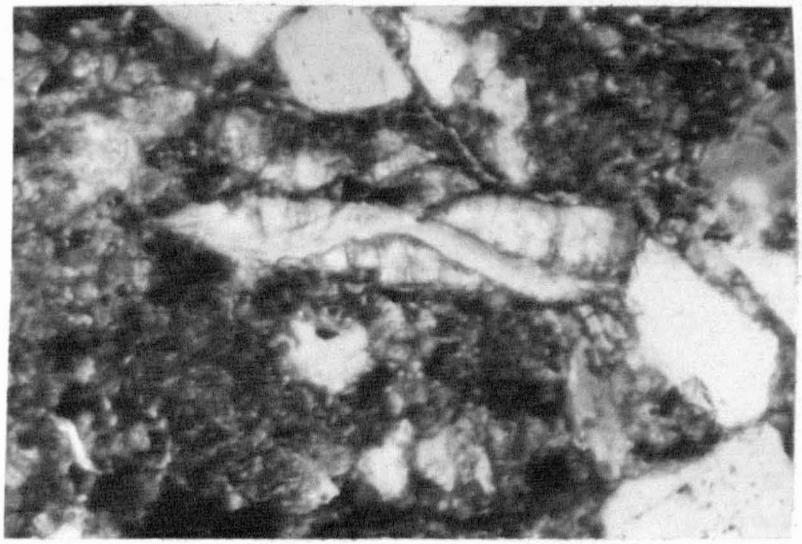
FIG. 2.26	Quartz grain in cored siderite matrix showing marginal replacement by clear siderite	(x200)
FIG. 2.27	Disruption of mica along cleavage planes by growth of siderite	(x130)
FIG. 2.28	As 2.27 with growth of dolomite	(x300)
FIG. 2.29	Replacement of non-ferroan calcite shell by dolomite crystal growing within flakes of clay mineral.	(x150)



2.26



2.27



2.28

2.29

5) Replacement of Micaceous Minerals

Muscovite and occasionally chlorite and chamosite flakes are often the site of growth of siderite, pyrite, or dolomite crystals. In all cases a small amount of replacement has taken place but this is generally insignificant in comparison with the amount of crystal growth.

a) Siderite. This is associated with mica flakes in two ways:

(i) between cleavages flakes of the original grain, which may result in a considerable separation of the constituent parts of the grain;

(ii) outside the grain, producing a fringe of clear pale brown siderite crystals which lack cores and are readily distinguishable from the matrix.

The internal grains often show a strong preferred orientation, with c-axes perpendicular to the mica cleavage. Both types may be seen in fig.2.27.

b) Pyrite. This occurs either alone or with siderite, and is developed only along internal cleavage flakes. The pyrite is in the form of small anhedral or subhedral bodies which may coalesce to form more continuous bands.

c) Dolomite. In the calcareous ironstone dolomite is associated with mica and chamosite flakes, to the exclusion of siderite and pyrite. Dolomite shows either internal or external development, and shows a much greater lateral expansion than either siderite or pyrite. The dolomite often grows so as to form a lens shape, whose width depends on that of the associated flake. Externally the dolomite particularly tends to occupy concave surfaces (fig 2.28). The crystals generally grow into the matrix, but where the mica flake has been in contact with a shell growth may take place into the shell material (fig.2.29); iron-free calcite shells only are affected. The growth does not seem to take the form of simple replacement as the dolomite is often not in contact with the calcite but is separated from it by fragments of the mica flake. Penetration may extend for nearly the whole width of the shell, with the mica flake being thrust deep into the shell.

All three types of replacement are often impersistent along the length of the grain, and the mica may be deformed into a network by the irregular displacement of its cleavage flakes.

6) Diagenetic History

Terrigenous grains, being on the whole stable, are subject to little diagenetic alteration. Considerable etching appears to occur before the crystallisation of the siderite matrix, while later replacement by siderite is only on a minor scale.

R Replacement of muscovite flakes, and the accompanying overgrowths cannot be satisfactorily dated, but may have taken place before full lithification of the matrix. In the case of dolomite, at least, it is clear that considerable pressure solution of resilient material occurs, apparently due to pressure of growth of the dolomite. This seems to provide clear evidence of force of crystallisation acting in such a way as to cause bodily movement of one grain and solution of another.

C. OOLITHS

1. Distribution

Ooliths are present in the ironstone unit over the greater part of the area but are absent in the western and south-western outcrops. They are most conspicuous in hand specimen in samples from the coast south of Whitby Harbour, where they are concentrated along laminae and replaced by white clay. The ooliths, however, comprise only two percent of the total rock. In lower Eskdale ooliths are present in similar small amounts, and are rendered less conspicuous by the presence of abundant fragmental shell material preserved in similar white clay. Ooliths are present in Glaisdale, but are absent in Great Fryupdale and in localities to the west. In the south ooliths are rare except in Hartoft Beck where a thin, slightly oolitic ironstone overlies the usual shelly ironstone. ~~The distribution of ooliths is summarised in figure 2.30.~~

2. Grain parameters

The average maximum diameter of the ooliths is greatest in the coastal sections south of Whitby Harbour. In the north it is somewhat less, and a regular decrease is seen as the ironstone is traced westwards (fig. 2.30). Sorting is similarly poorest on or near the coast and improves regularly when traced inland. The distribution appears to be unimodal in all cases. Size frequency diagrams (fig.2.50) show all of these features. The shape of the ooliths cannot be directly determined in thin section but maximum elongation of ooliths determined from four widely spaced localities is about 2:1. The wide scatter from this maximum will be due to sectioning away from the centre and to disorientation.

of the ooliths from the horizontal, either as a result of deposition or burrowing activity.

3. Composition

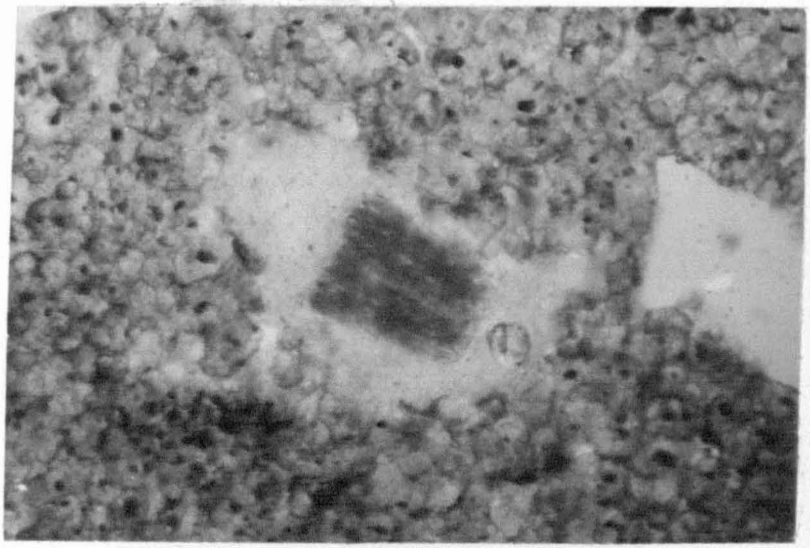
The ooliths may be divided into two main mineralogical groups: kaolinitic ooliths and chamositic ooliths. The chamositic ooliths show the usual tangential orientation of crystals within the envelope, but the kaolinitic ooliths show almost total loss of structure and are therefore considered to be the result of alteration of chamosite ooliths. The chamosite of the unaltered ooliths varies in colour from neutral to greyish green, and shows grey interference colours. In ordinary light the oolitic lamination often appears rather diffuse, and the chamosite turbid, but under crossed nicols the tangential orientation of flakes is seen to have been retained.

Nuclei occur in less than five percent of the ooliths. They generally consist of quartz grains, but also include grains of iron oxide and tubular chamosite crystals (fig.2.31). The latter are brownish green and show interference colours up to first order yellow, in contrast to the very low birefringent chamositic envelopes. Cleavage is generally parallel to the elongation of the crystal, but is occasionally transverse. The corners of the crystals are rather rounded.

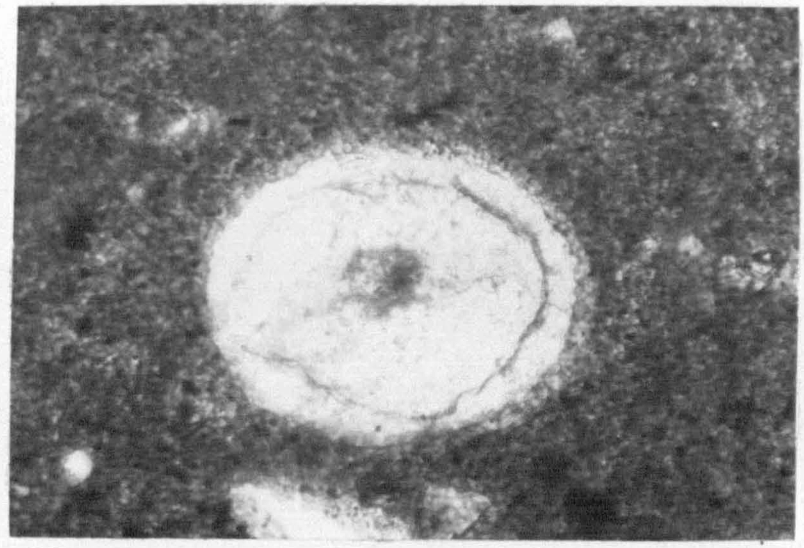
4. Replacement

a) Kaolinite. The 'structureless' ooliths already described consist of rather cloudy colourless material usually very fine grained and practically isotropic. Recrystallisation is not uncommon, however, and leads to the development of small 'books' of flaky crystals,

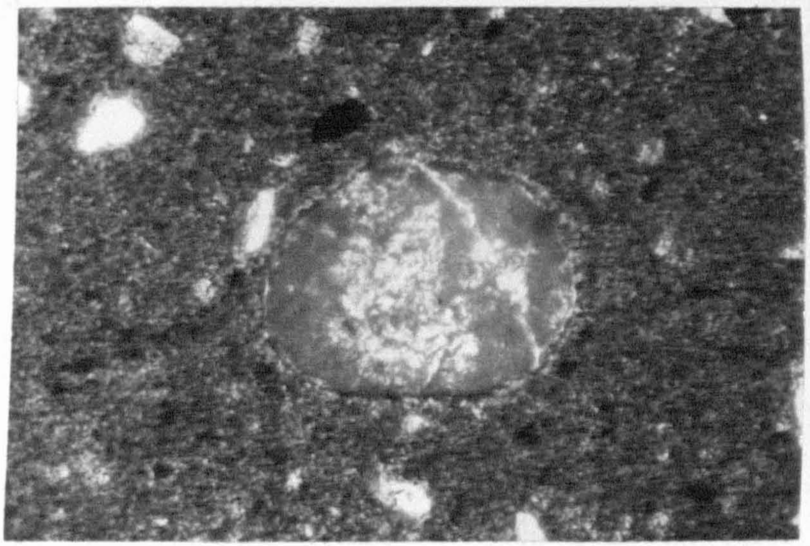
FIG. 2.31	Tubular chamosite nucleus in partially sideritised oolith	(x100)
FIG. 2.32	Kaolinitised oolith with remnant concentrated structure	(x160)
FIG. 2.33	'Opal' in kaolinitised oolith	(x150)
FIG. 2.34	'Opal' in chamosite ooliths	(x100)



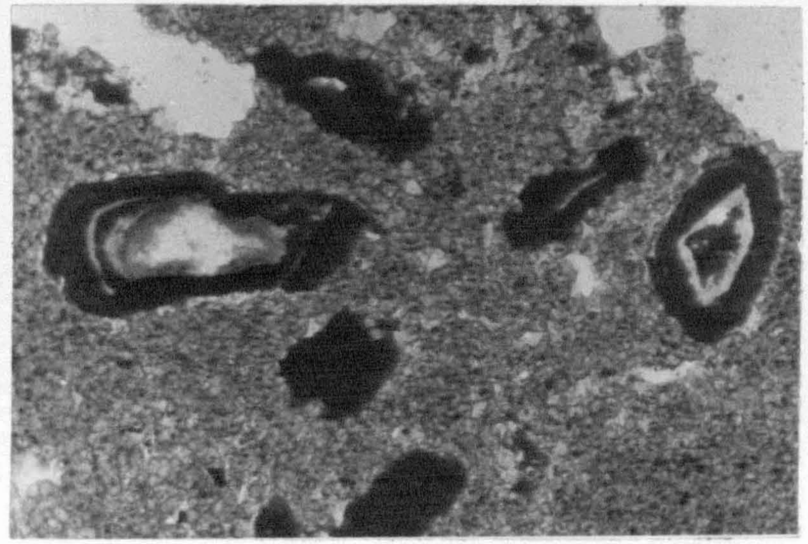
2.31 100 μ



2.32 100 μ



2.33 100μ



2.34 100μ

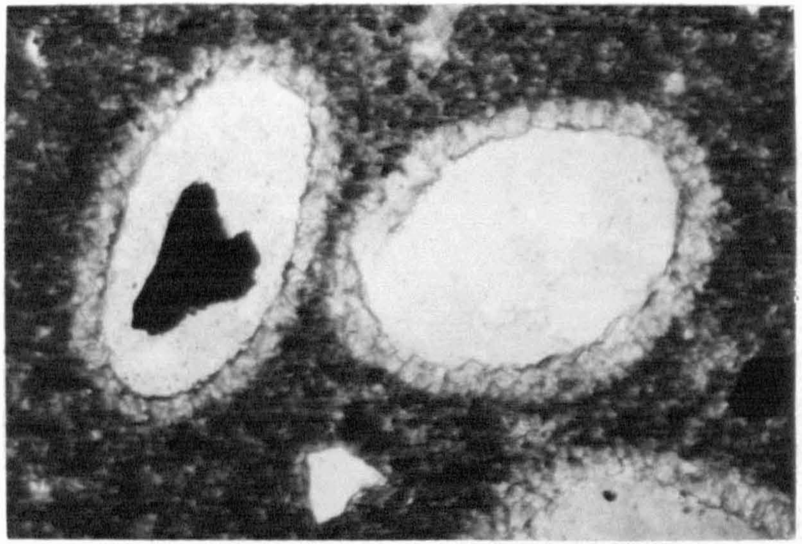
believed to be kaolinite. The distribution of larger crystals is very patchy and appears to bear no relationship to the outline of the oolith. Faint lines of inclusions reflect the original concentric structure of the oolith in a few cases (fig. 2.32) and nuclei, where present, occupy a central position.

b) ? Opal. A common feature of many ooliths is the presence of dusty inclusions. In kaolinitised ooliths these have an irregular distribution (fig. 2.33), but in laminated ooliths the cloudy material follows broad concentric bands in which it is clearly superimposed on well laminated tangentially oriented chamosite flakes (fig. 2.34). In reflected light the cloudy material gives a strong white reflection. The mineral has been described here as opal, as this has been the conclusion of others working on similar material (Whitehead et al. 1952; p.24).

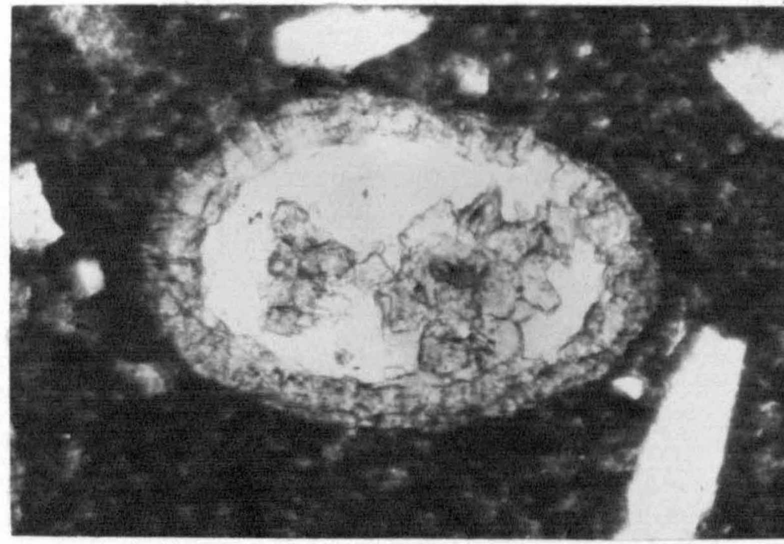
c) Siderite. Replacement by siderite is also a feature of most ooliths. The replacement may occur in two ways:

(i) marginal replacement, which is practically always developed, leads to the formation of a 'rind' of siderite along the periphery of the oolith (fig. 2.35). The thickness of each rind is generally constant, but the average thickness of rinds commonly varies from sample to sample, ranging from 60 microns to being thin and impersistent. No relationship between the rind thickness and the nature of the surrounding matrix could be determined. Poorly developed rinds are, however, characteristic of ooliths in which the structure is retained. The rinds are composed of a large number of rather elongated radially oriented crystals, each crystal generally extending through the total thickness of the rind. The inner margin of the rind is generally

FIG. 2.35	Marginal replacement of ooliths by siderite rinds (one oolith showing central nucleus within kaolinite).	(x180)
FIG. 2.36	Marginal siderite rind accompanied by internal replacement by siderite rhombs	(x180)
FIG. 2.37	Total replacement - combination of marginal and internal	(x220)
FIG. 2.38	Total replacement with retention of concentric structure (centre of oolith lost in sectioning)	(x130)

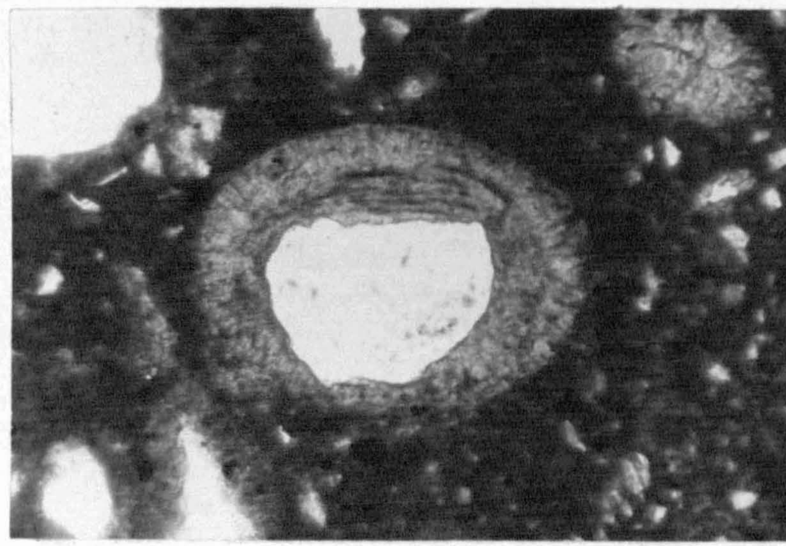


2.35 100μ



2.36 100μ

2.37 100μ



2.38 100μ

smooth in appearance at low magnification, but is seen to consist of a series of projecting crystal terminations at high magnification. The siderite is brownish and usually of a colour similar to that of the matrix, from which it is clearly distinguished by the radial orientation, and by the lack of crystal cores.

(ii) internal replacement, which is rarer than marginal replacement, leads to the development of irregular clusters of siderite crystals, some of which may show rhomb faces (fig. 2.36). The crystals are brownish in coulour, uncored, and are generally in the order of 50 microns in length. Such replacement generally occupies a central position but in some cases the crystals appear to grow from the inner margin of the siderite rind, from which it is distinguishable by a colour difference and a fine line of opaque inclusions. In such cases total replacement of an oolith commonly takes place (fig. 2.37) and the inner siderite consists of elongated crystals with radial orientation, but which are most strongly developed perpendicular to the maximum diameter of the oolith. A fine, irregular line of inclusions commonly follows the maximum diameter.

Very rarely internal replacement consists of a cryptocrystalline siderite aggregate in which the lamination of the oolith is to some extent preserved (fig. 2.38).

d) Pyrite. Replacement of ooliths by pyrite is limited to some of the coastal sections where small areas of pyritisation of both matrix and grains are present. The pyrite consists of irregular aggregates of pyrite crystals, 15-20 microns in diameter, which may coalesce to form massive pyrite. Except where pyritisation of the matrix is complete, the siderite rinds are generally unaffected while the enclosed portion may be partially or completely replaced (fig. 2.39). Where partial replacement has occurred the pyrite is densest near the inner margin of the siderite rind. The unreplaced material comprises granular clay similar to that found in associated ooliths which are unaffected by pyritisation. In some cases the ooliths are pyritised only where the matrix is completely replaced, but in others the ooliths are replaced independent of the degree of replacement of the matrix. In one sample the matrix is pyritised in a narrow zone around each oolith, while the ooliths themselves are unaffected.

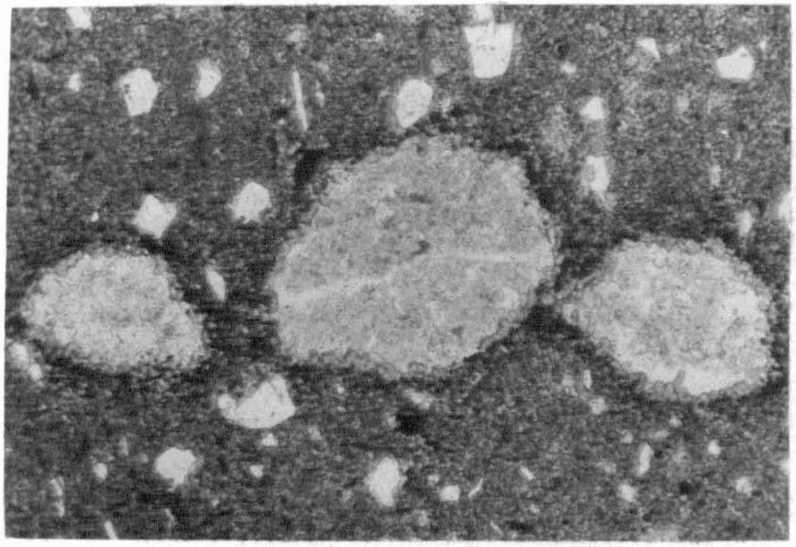
e) Sphalerite. Replacement by sphalerite is also restricted to sections on or near the coast. The sphalerite forms anhedral or subhedral crystals which may occupy a part or all of the area within the siderite rind (fig. 2.40); the rind itself is unaffected. The unreplaced material consists of granular clay.

f) Phosphate. Phosphatised ooliths have been observed only in the phosphate shell mould previously referred to. The ooliths are pale orange brown in colour with occasional clearer patches which show very weak birefringence in contrast to the bulk of the phosphate which is isotropic. Replacement is complete, except where a quartz nucleus is present, and marginal sideritisation is absent (fig. 2.41). The phosphatised ooliths are scattered and adjacent ooliths may be unaffected and show complete sideritisation.

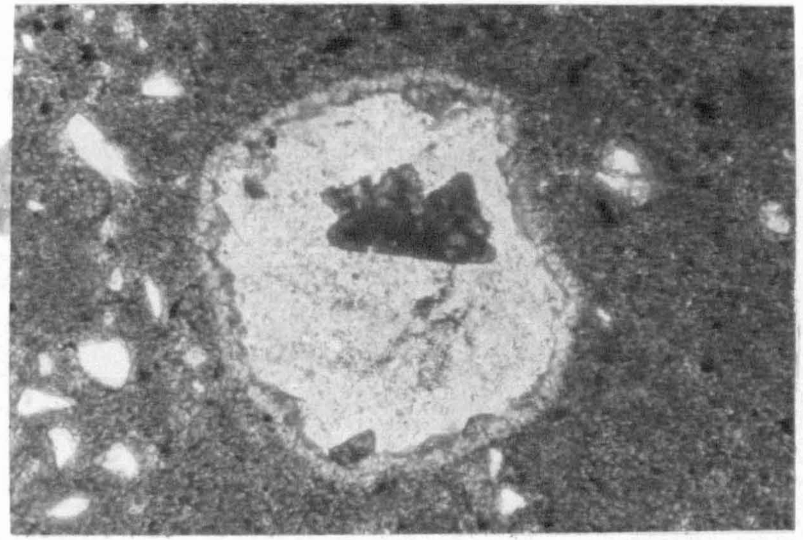
5. Solution

In a few localities in the north-east of the area the part of the oolith enclosed by the siderite rind may be void. Internal replacement

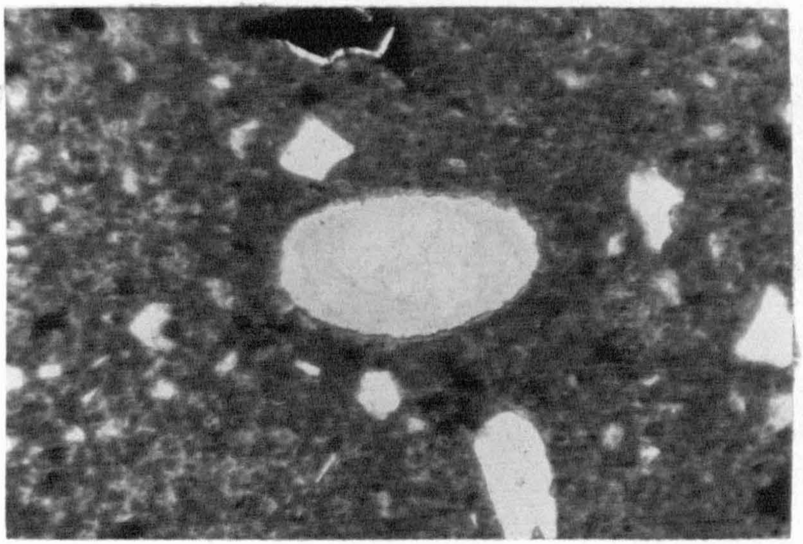
FIG. 2.39	Marginal pyritisation of ooliths	(x90)
FIG. 2.40	Sphalerite in kaolinitised intraclast	(x100)
FIG. 2.41	Phosphatised ooliths with faint concentric structure; note absence of typical rind.	(x130)
FIG. 2.42	Spastolithisation of chamosite ooliths.	(x150)



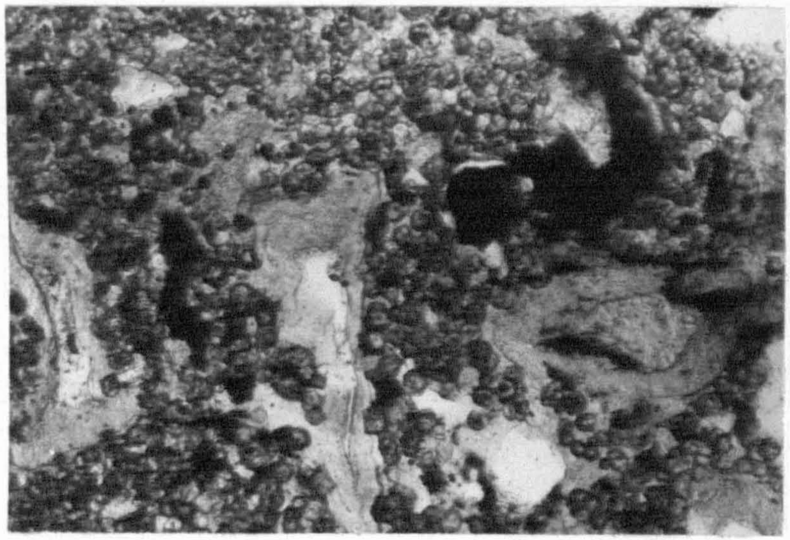
2.39 100μ



2.40 100μ



2.41 100μ



2.42 100μ

minerals such as siderite and sphalerite may occupy some of the space.

6. Deformation

The only deformed ooliths are those in the upper part of the ironstone in Hartoft Beck, which have undergone spastolithisation (Rastall and Hemingway, 1940; p.265). Deformation has occurred as a result of plastic flowage of the oolitic envelopes, which are lacking siderite rinds. The spastolithisation is apparently haphazard, showing no relation to bedding and no particular lines of deformation (fig. 2.42).

7. Diagenetic History

In the Eskdale localities the ooliths have undergone relatively little diagenesis while at localities along the coastal area they have undergone total reorganisation. It is probable that all of the ooliths originally had a laminated chamositic envelope, but in many cases this has been destroyed, the principal processes being the development of siderite rinds and replacement by kaolinite. There are two ways in which the kaolinite might have formed:

(i) by total solution of chamosite, with subsequent deposition of kaolinite from solution within the cavity;

(ii) by alteration of chamosite to kaolinite in place. In most ooliths there is little evidence to support either mechanism, but the occasional presence of concentric zones of inclusions, together with the fact that nuclei, when present, always occupy the centre of the ooliths, favours the second.

The development of siderite rinds clearly represents the replacement of the oolith from its outer margin inwards, and the similarity of the siderite within the rind to that of the surrounding matrix suggests that the matrix has a strong influence on the replacement. The rinds are poorly developed around chamositic ooliths, but are well developed where the ooliths have been converted to kaolinite. Similarly replacement by pyrite, sphalerite and internally by siderite is associated with kaolinite. This either suggests that the ooliths were at an early stage replaced by kaolinite, which was in turn susceptible to replacement by other minerals, or that the kaolinite may have been introduced later, the association of the replacing minerals simply reflecting the instability of the chamosite. The siderite rinds at least seem to precede the kaolinite. In a few rare cases internal (and therefore almost certainly later) replacement by siderite retains a crude concentric structure. The presence of voids within the rinds at some localities implies early rind formation, the alternative being replacement of chamosite by kaolinite, marginal replacement of kaolinite by siderite and finally total solution of kaolinite.

Kaolinite therefore appears to be a late replacement mineral, which was introduced subsequent to siderite and perhaps sphalerite replacement. The development of solution cavities in a few localities may have occurred near the time of kaolinite replacement elsewhere. Pyrite replacement is also difficult to date, but it is clearly later than rind formation and may have occurred after kaolinite replacement as a result of liberation of iron. The only remaining replacement, phosphatisation, appears to be very early, since in phosphatised ooliths siderite rinds are thin or absent.

Spastolithisation in the ironstone at Hartoft Beck may have occurred either before or after sideritisation of the matrix, but since no other samples show deformation of ooliths it may be that incomplete sideritisation has led to the development of local stresses during compaction. The lack of siderite rinds may have been a further factor.

D. SHELL MATERIAL

1. Discussion on original mineralogy

The original character of the shell material is almost totally lost except in the calcareous shelly ironstone of Great Fryupdale. In this ironstone the shells fall into two groups which are distinguished by their structure and by staining methods. The stain used distinguishes iron-free calcite from ferroan calcite (see Appendix 2E). In one group the shell is composed of iron-free calcite, and the structure which comprises fibrous laminated and occasionally prismatic elements is perfectly preserved. The other group is characterised by total or partial loss of structure, and the shell material is ferroan calcite whose texture clearly indicates replacement; any structure that is retained is of a regular laminated type. These features suggest that the first group consist of original calcite shells (including oysters and ostracods) while the second consist of shells that were originally aragonitic.

In the non-calcareous ironstone this distinction is seldom possible and shell types of known composition, such as ostracods and gastroppds are useful as guides. In the following sections an attempt will be made to refer to the original nature of the shell as far as possible.

2. Organic perforation of shells

The shell material is often highly fragmented and the iron-free calcite shells show extensive boring which affects both original and broken surfaces. The perforation may be so intense as to reduce the shell to a mere skeletal framework. The structure is virtually invisible in an unstained section, but is brought out well by staining since the perforations are filled with a cement of ferroan calcite. Two types of boring may be distinguished (fig. 2.43):

(i) diameter 0·5 to 1·0 microns, and rather straight

(ii) diameter 4·0 to 4·5 microns, and slightly sinuous

The first of these and possibly the second, may be ascribable to filamentous algae (Cayeux, 1914) The outline of the shells is highly irregular which gives the impression of replacement by the matrix, but it is apparently the result of these perforations. Similar structures are assumed to be present in the ferroan calcite shells, but cannot be brought out by staining.

3. Replacement

a) Ferroan calcite. The calcite which replaces aragonite shells is pale brown, and when stained is blue or purplish blue. Crystal size is variable but each shell is composed of a large number of crystals, whose size is roughly proportional to the size of the shell. The crystals may form an irregular mosaic but in some shells the original lamination has controlled the crystal boundaries, thereby preserving the basic elements of the shell structure.

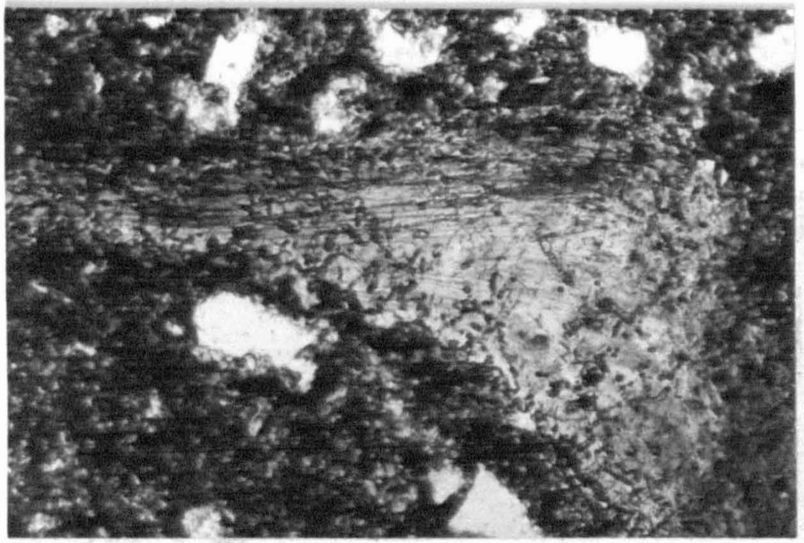
b) Pyrite. Shells replaced by ferroan calcite often show partial replacement by pyrite. Occasionally the pyrite is in the form of thin

FIG. 2.43 Algal perforations in calcite shell (accentuated by staining) (x15)

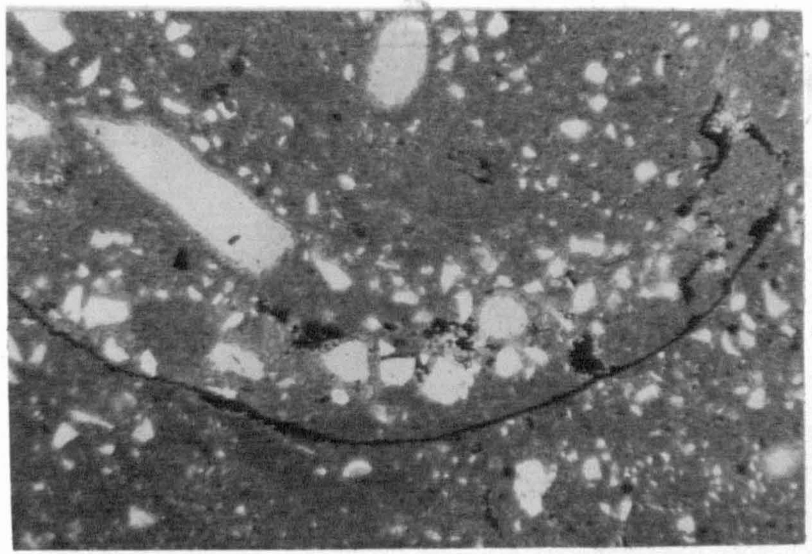
FIG. 2.44 Residual pyrite film. (x40)

FIG. 2.45 Sideritised shell showing retention of original structure and algal perforations (x160)

FIG. 2.46 Original calcite shell replaced by overgrowths of quartz on adjacent detrital grains (x460)

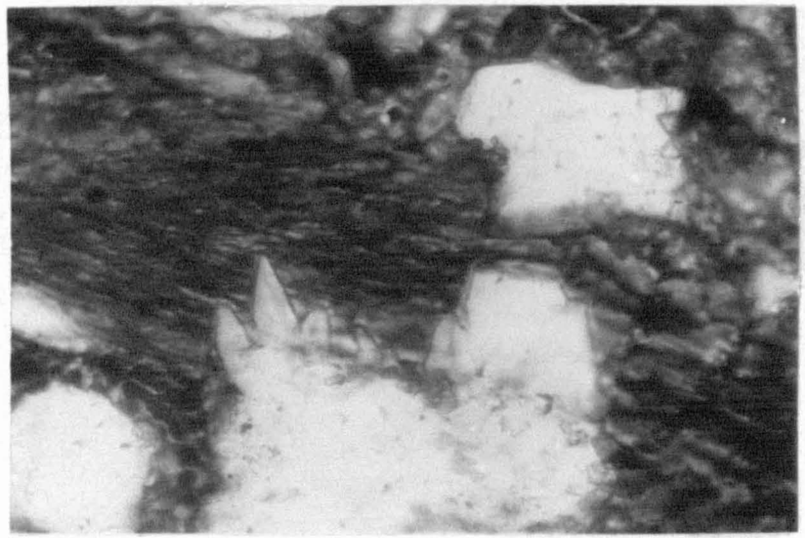


2.43 100μ



2.44 500μ

2.45 100μ



2.46 50μ

sheets of constant orientation, but it generally occurs as a zone of dusty inclusions near the shell margin. In both cases the pyrite appears to reflect the original structure of the shell, and shows no relationship to the calcite mosaic.

In the non-calcareous ironstone pyritisation has often been followed by total solution of the remaining shell material, leaving only a thin film of residual pyrite (fig.2.44). A similar feature, but with intermediate stages is seen in the ferroan calcite shells which have undergone pyritisation, and so shells represented by pyritic films were probably originally aragonitic. Pyritised shells seen in hand specimen are all aragonitic forms.

c. Siderite. Partial or total replacement of shell material by siderite is common in the non-calcareous ironstone. The original mineralogy of the shells is difficult to assess but probably both calcite and aragonite forms are affected. The replacing siderite is reddish brown and apparently cryptocrystalline. The internal structure of the replaced shell is preserved, and even the fine detail of algal perforations may be visible (fig.2.45). Where partial replacement has occurred the remainder of the shell has been removed by solution, and the sideritisation is clearly seen to have progressed along interlaminar boundaries.

d. Quartz. A feature which is common in original calcite shells but absent in ferroan calcite shells is the development within the shell of overgrowths from quartz grains with which the shell is in contact. Each overgrowth is in optical continuity with its respective detrital quartz grain and consists of one or more euhedral projections, (fig.

2.46). Inclusions are present within the overgrowths and reflect the structure of the replaced shell material. The overgrowths may extend from one side to the other but never pass beyond the limits of the shell.

e) Sphalerite. Original calcite is very rarely replaced by large crystals of sphalerite. Where quartz overgrowths also occur they are unaffected by the sphalerite.

4. Solution

Partially sideritised and pyritised shells have already been described as having undergone solution. In the first case solution generally leads to the development of a cavity which is later the site of deposition of drusy fillings, while in the second case the matrix has been able to accomodate the volume loss, leaving no cavity. Intermediate situations occur where the shell outline shows 'pinch and swell' in which the matrix has moved into the cavity to varying degrees along the length of the shell. Some cavities may have formed without any prior replacement, but most possess a thin wall of sideritised shell. Shell cavities may be voidal in part but generally they are filled with one or more minerals which have been deposited from solution. These minerals are considered below.

5. Cavity Filling

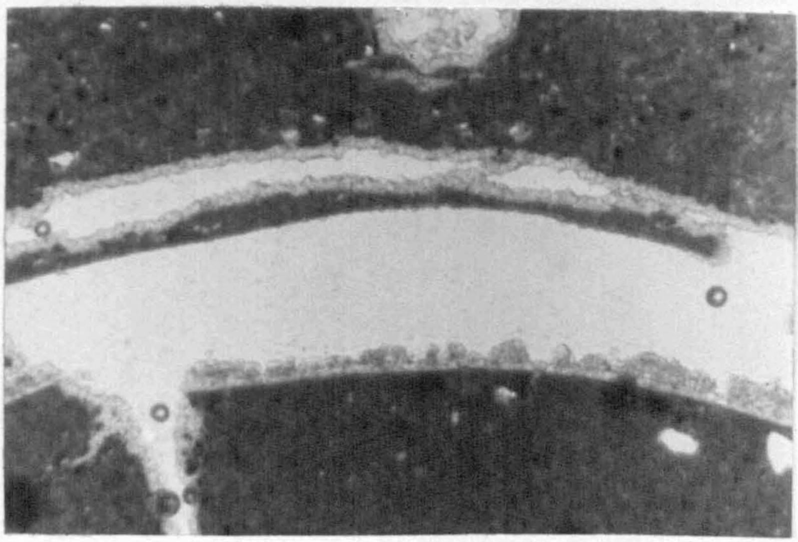
a) Siderite. Every shell cavity possesses a rim of siderite, though the thickness may be very small in some cases. The siderite is developed as clear slightly brownish crystals growing at right angles to the cavity wall. Ordinarily the rim is of even thickness in any one cavity and has a smooth outline, but in some cases the thickness of the rim varies on opposite sides of the cavity.

Siderite rims are seldom affected by deformation or breakage. Where differential solution has produced pinch and swell the siderite rim follows the irregular outline of the mould wall. In muddy rocks which show septarian shrinkage cracks around the larger grains more complex structures are produced. These cracks develop either along the surface of a shell or a short distance away from it, so that a thin film of mud may remain attached. On solution of the shell an extremely fine lamina of reddish brown material marks the margin of the shell and this apparently remains unbroken, even though support is totally lacking (fig.2.47). Rim depositon then occurs on either side of the film, into the shell and shrinkage cavities. Where mud has remained on the shell surface the film is less easily distinguished. The nature of this film is uncertain, but its extreme thinness and strength suggests that it may be the conchiolin periostracum of the shell. In many of the apparently sideritised shells similar fine laminae are visible and are responsible for the preservation of the shell structure, and it is possible that the siderite in such shells is not replacive but fills cavities between the conchiolin layers. In prismatic shell material the structure is again supported by a framework of thin reddish brown laminae, with siderite rims developed on both sides. This does not eliminate sideritisation as a process since in many shells the siderite is clearly not a rim development.

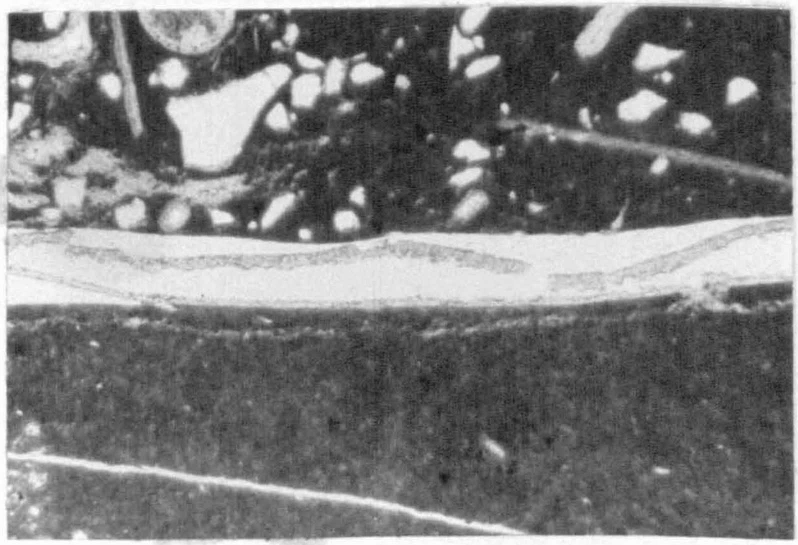
Although this film, together with the siderite rims, often shows remarkable resilience, breakage does occur. In some cases the film has been ruptured prior to the development of the siderite rim,

FIG. 2.47 Film of matrix suspended between septarian crack and dissolved shell fragment. (x70)

FIG. 2.48 Buckled siderite rims enclosed in kaolinite (x40)



2.47 500 μ



2.48 500μ

while in others the siderite rim itself is broken. Rupture of the film prior to the rim deposition only occurs when a layer of mud is present and appears to be due to shrinkage of the mud pulling apart the whole layer. No gravitational breakage seems to be involved. Fractures affecting the rim, on the other hand, appear to be the result of foreshortening along the shell. This may occur where a double rim is unsupported, or a normal rim not associated with shrinkage cracks may be affected. The latter case is shown in figure 2.48 where, since both rims are affected, gravity is clearly not a controlling factor.

Siderite rims seldom extend far into a cavity, but are usually followed by further cavity filling by siderite and kaolinite. The later siderite can be distinguished from the rim siderite since it develops as well formed rhombs which are scattered and patchily distributed.

b) Kaolinite and Dickite. The inner parts of shell cavities are almost invariably filled with colourless clay mineral. The grain size is usually too fine for optical tests to be carried out, although the mineral is clearly of the kaolinite family. X-ray diffraction of clay minerals from shell cavities has proved that two forms may be present, kaolinite and dickite (see Appendix 2B). Dickite has been recorded previously from the ironstone at Eller Beck (Hemingway and Brindley, 1952), and has now been proved to occur in much of lower Eskdale - at Eller Beck, Grain Beck, and in Glaisdale. Along the coastal area kaolinite is developed. The clay mineral clearly post-dates the siderite rims and where fracture of the rims has occurred the clay is undisturbed and shows no variation in grain size that could be related to such movements. Opal is often developed within the clay mineral when the associated ooliths are opaline. The opal has a random blotchy distribution.

c) Calcite. Exceptionally, in the phosphatic specimen from Whitby the cavity within the siderite rim is calcite-filled. This comprises a single crystal of clear ferroan calcite which extends for the complete length of the shell.

6. Deformation.

Several features of breakage and pressure solution are developed in the ironstone. Those relating to solution cavities have already been dealt with. Breakage occurs where thin shells have been compacted against resistant grains which may include quartz grains or even other shell fragments. In the calcareous ironstone the breakage can be roughly dated, since in some instances the fracture is filled by siderite matrix, and in others by ferroan calcite cement. Post-depositional breakage is not a common feature. In prismatic shells slip may be observed between the constituent prisms, and is accentuated by displacement of internal banding.

Pressure solution affects original aragonite shells. Where some structure is retained, as for instance where pyritic laminae are present, it shows plastic deformation by an intruding grain, while the associated ferroan calcite shows volume loss by solution. The intruding grain is usually quartz, but occasionally a shell may penetrate another when it is in contact with it end on. Calcite shells penetrate aragonite shells in this way and occasionally an aragonite shell penetrates another aragonite shell.

7. Diagenetic History

The first alteration to take place is pyritisation, which apparently affects only aragonite shells. This clearly precedes alteration of aragonite to ferroan calcite. Pyritised shells seem to have been particularly prone to early solution since in the calcareous ironstone the partially pyritised shells show partial or complete solution of aragonite prior to replacement by ferroan calcite, while the non-pyritised shells seem to have retained their original outline. In the non-calcareous ironstone shells are commonly represented by a thin lamina of pyritic residue while associated shells show cavity producing solution. It is not clear why this should by the case.

In order to date other processes with respect to the alteration of aragonite, the calcareous shelly ironstone of Great Fryupdale will be considered first. The distinct contrast between the effects of pressure by quartz grains on original calcite and original aragonite shells is very marked. The fact that the former have reacted by replacement and the latter by solution implies that the solution, at least, occurred before aragonite was converted to ferroan calcite, when the aragonite shells would be more soluble than the associated calcite shells. The development of overgrowths, however, implies a different phase of diagenesis and may have occurred after conversion of aragonite, the selective replacement of original calcite shells being due to the greater solubility of non-ferroan calcite compared with ferroan calcite. This difference in solubility is clearly seen on examination of an etched surface of a thin section, where non-ferroan calcite shows deep etching while the ferroan calcite is apparently only little affected. Sphalerite replacement of iron-free calcite shells post-dates the formation of quartz overgrowths and is probably the last chemical alteration to take place. Breakage of shells under compaction appears to have occured at two stages; early breaks are filled with siderite matrix, and later ones with ferroan calcite which is probably deposited during or after aragonite replacement.

In the non-calcareous ironstones there is little to distinguish aragonite from calcite shells. Partial sideritisation appears to affect both types and this is followed by solution of remaining calcium carbonate. The solution is followed by siderite deposition within the cavities, which are finally filled by clay mineral or, exceptionally, calcite.

The probable diagenetic sequence in non-calcareous and calcareous ironstones are compared in fig. 2.49.

E. INTRACLASTS

The term 'intraclast', though originally applied to limestones (Folk, 1959; p.4), has been adopted here to cover fragmental grains which appear to have been derived from within the basin of deposition, exclusive of shells and ooliths. The intracalsts though not abundant show significant regional variation.

1. Composition

In coastal districts south of Whitby harbour the intraclasts comprise

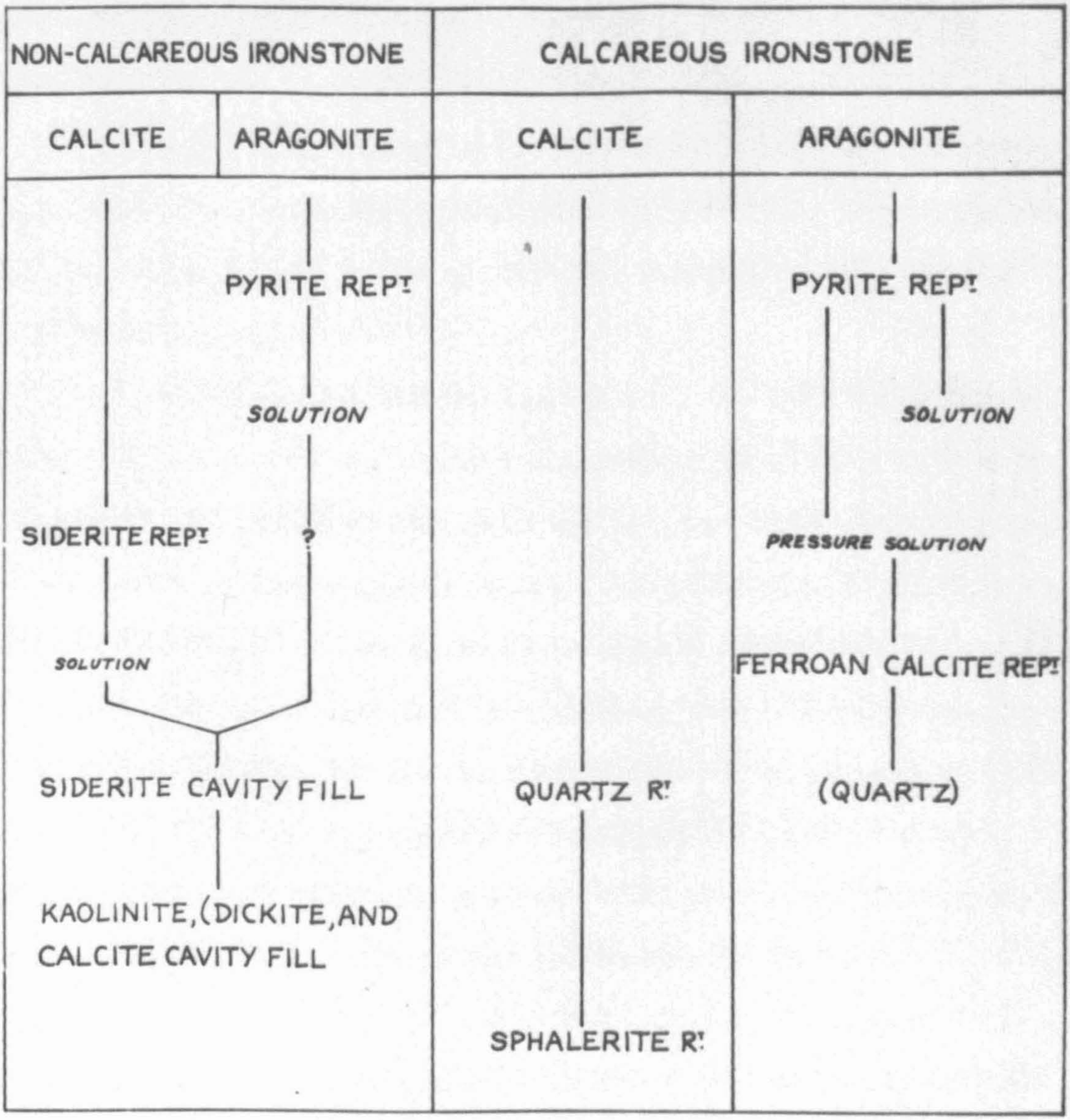


FIG. 2.49 DIAGENESIS IN SHELL MATERIAL IN CALCAREOUS AND NON-CALCAREOUS IRONSTONE

three types:

(i) fragments of similar size to the ooliths, or a little larger, and with a slightly irregular outline. The presence of sand grains sometimes serves to distinguish the grains from ooliths, but often identification is difficult.

(ii) larger fragments of shale occur in the lower part of the ironstone which are apparently derived from the underlying non-marine shale. The clay minerals show a strong preferred orientation which appears to be primary fabric.

(iii) siderite spherulites occur both in the siderite matrix and in the shale intraclasts. They are generally associated with sandy laminae and although they are unaltered and show little abrasion they do not appear to have grown in place. The underlying non-marine shale is a probable source since it is locally spherulitic.

In Eskdale the intraclasts are smaller, with a maximum diameter of about 300 microns, and are similar in size to the associated ooliths. In all cases they comprise greenish isotropic chamosite mudstone, commonly with included quartz grains. Rare phosphatic intraclasts are present, sometimes showing partial replacement by siderite.

<u>2. Replacement</u>

<u>a) Siderite</u>. On the coast many intraclasts have developed siderite rinds identical to those of the associated ooliths, and internal replacement is common, leading, in the phosphatic mudstone, to total replacement in some cases. Where the intraclasts are chamositic replacement by siderite is confined to the margin and is on a small scale.

<u>b) Kaolinite and Opal</u>. Total internal replacement by kaolinite and

opal occurs in intraclasts on the coast in a manner identical to that of the associated ooliths.

c) Pyrite. Shale fragments in coastal districts occasionally show replacement by bladed pyrite crystals.

d) Calcite. In the calcareous ironstone intraclasts which are physically similar to those of adjacent Eskdale localities are replaced by ferroan calcite. Occasionally replacement is incomplete, and the original material is seen to be chamosite.

3. Solution and Cavity Filling.

In phosphatic siderite mudstone siderite rind formation is accompanied by solution, leaving internal cavities which may be filled by a single calcite crystal in the same way as the accompanying shell material.

4. Diagenetic History

The fragments of non-marine shale which are found in coastal localities have undergone little diagenetic alteration. Replacement by siderite and pyrite is locally developed, and both have probably occurred during early diagenesis. Siderite spherulites appear to be unaffected. The remainder of the intraclasts were probably originally chamosite mudstone fragments. In Eskdale they are unaltered except where replaced by calcite in the calcareous ironstone of Great Fryupdale. On the coast, however, a sequence of replacement identical to that of the associated ooliths has occurred, with siderite replacement followed by recrystallisation of mud to kaolinite, or in phosphatic mudstones by solution of mud and later drusy filling by calcite.

Phosphatic intraclasts appear to represent early replacement of partially sideritised mud, similar to that observed in the matrix of the undisturbed phosphatic mudstones.

F. FAECAL PELLETS

Grains are occasionally seen in the ironstone which have an oval or oblong outline and which possess a diffuse internal structure resulting from the transverse orientation of numerous slightly elongate opaque bodies. Apart from these bodies the grains consist of dark brown turbid cryptocrystalline siderite. These grains are similar to those of faecal origin observed in other formation except for the faint internal structure. They measure between 300 and 500 microns in length and between 150 and 250 microns in width. The pellets show no sign of replacement in either the calcareous or non-calcareous ironstones. Similar grains occur in the sandstone unit in Rosedale and these will be described and figured later.

III CONCLUSIONS

A. DEPOSITION

1. Sequence

The first manifestation of marine or brackish conditions is the colonisation of the uppermost freshwater clays by various burrowing organisms, particularly the small forms of Diplocraterion. These burrows are present over much of the area, but become infrequent and die out to the north, west, and southwest (fig. 2.7). The absence of burrows in these areas may be due to low salinity or to insufficient water coverage. In central and southern districts the earliest marine sediment consists of a soft silty sandstone or sandy siltstone which is also the infilling material of the underlying burrows, (fig. 2.3). This is succeeded for the most part by siderite mudstone which extends over practically the whole area. In the south, however, the ironstone is rather imperistent and silt and sand deposition continued throughout.

2. Source of constituent materials

a) Clay. The clay fraction of the ironstone appears to have been chamositic. The few examples of chamosite which have not been replaced, however, are dirty in appearance and may contain a proportion of terrigenous clay. The impurity of the chamosite as compared with pure chamosite mud from other ironstones suggests that the chamosite may be derived, and not have been precipitated in place. If this is the case, there is little evidence as to the source of the chamosite, but it may well have been introduced with the ooliths (see below).

b) Terrigenous grains. These are most abundant in central and southern parts and show variation not only in abundance but in sorting and grain size as may be seen from the size frequency diagrams (fig. 2.50). In central Eskdale a single mode is present and the sorting good. To the west, north, and east this mode becomes finer and the probable source of these grains therefore lay to the south. In northern and eastern districts, however, coarser grains are also present which result in a long coarse tail in the size frequency distributions. These grains, being coarser in part than any found inland, must be derived from an easterly source.

c) Ooliths. The size frequency diagrams for ooliths are comparable in some respects to those for sand grains. In eastern localities the ooliths are large and poorly sorted, while the sorting shows a progressive increase westwards together with a decrease in grain size. Oolith abundance is variable and low in eastern and central districts, and in the west ooliths are absent. As with the coarse terrigenous grains an easterly source is again probable.

d) Shells. Shell material falls into two distinct categories; over the whole area entire shells are present to varying degrees while in central and southern areas fragmentary shell material is abundant. Most shells have probably undergone some transport with the exception of those forms which are found in vertical growth position, whose distribution is given in figure 2.6. The fragmentary shells are probably derived locally as there is no evidence of their having been introduced from surrounding parts.

e) Intraclasts. Intraclasts may comprise (very locally derived) fragments of sediment as in the case of the large clay fragments and spherulites in some coastal districts, but for the most part they

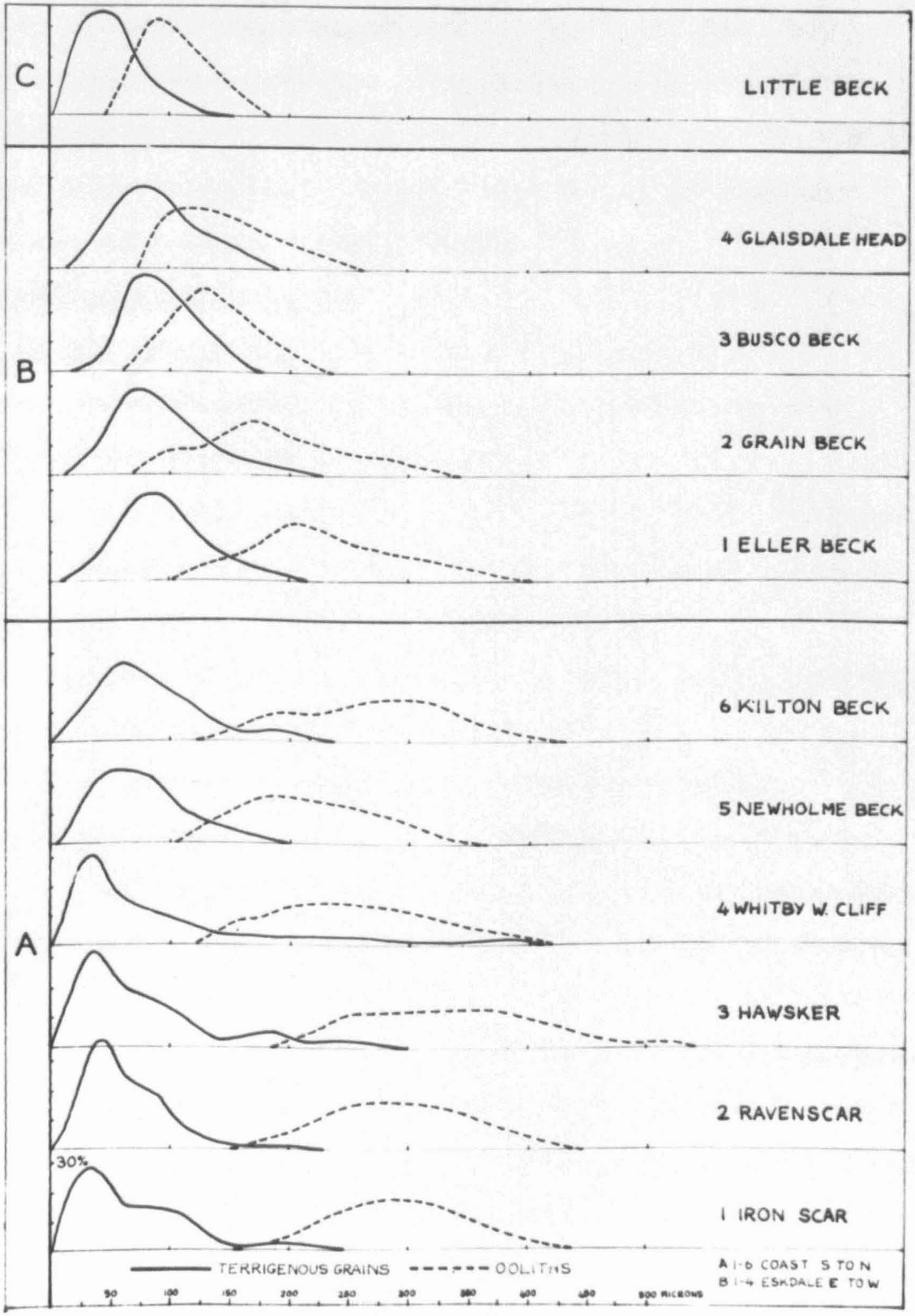


FIG. 2.50 SIZE FREQUENCY CURVES FOR OOLITHS & TERRIGENOUS GRAINS

appear to comprise particles of chamosite mud. The chamosite in these intraclasts is very pure and contrasts strongly with the chamositic mud relics in the matrix of some localities. The size of these intraclasts usually corresponds with that of the associated ooliths, and a similar easterly origin is proposed.

3. Environment of Deposition

a) Physical Environment. The factors considered to be of most importance in elucidating the physical conditions under which the ironstone has formed are as follows:

(i) the lateral persistence of the ironstone despite its thinness, and the significance of variations in thickness of only a few inches with relation to the regional distribution (fig.2.2);

(ii) the poor overall sorting of the sediment, with a fairly high proportion of grains showing mud support;

(iii) the abundant though rather restricted fauna;

(iv) the fragmentation of shells in central parts of the area;

(v) the transport of bulk of terrigenous grains from south, and a few terrigenous grains together with ooliths and intraclasts from the east.

The first of these requires an extremely flat initial depositional surface followed by widespread deposition under relatively low energy conditions, with regular accumulation and little or no contemporaneous erosion. The second factor similarly implies accumulation under low energy conditions.The presence of an abundant fauna precludes stagnation of water, however, and the fragmentation of shells in the upper part of the ironstone in central Eskdale suggests a certain amount of reworking of the sediment in this area. Although accumulation clearly took place under quiet conditions the transport of grains into the area must require a certain degree of current movement. The terrigenous grains of central and southern regions may have been brought into the area during the transgression, by wave action, but the other components from the east may have been drifted in during temporary phases of higher energy conditions.

Figure 2.50 illustrates the nature of sand grain and oolith distributions in the ironstone from Eskdale and from northern and coastal districts. The distributions have been plotted on arithmetical probability paper rather than on a logarithmic scale in order to accentuate the variation which principally occurs at the coarse end (this is probably the effect of sectioning). The following features may be seen:

(i) Ooliths are poorly sorted in the east and north and in Eller Beck while to the west the sorting improves rapidly as a result of the diminution of the coarse 'tail'. At the same time the modal grain size decreases.

(ii) Sand grains show a similar trend to the ooliths as regards sorting, but the modal grain size in Eskdale localities is very constant and the two most northerly sections show similar modes. In the east and south-east the modal grain size is finer than to the west and north but is accompanied by coarse 'tail' which occasionally is of considerable length.

(iii) When the ooliths and sand grains are considered together they are better sorted relative to one another in Eskdale than on the coast as may be seen by the increasing proximity of the two curves from the east to west.

This implies a stable environment of accumulation in Eskdale, while on the coast the grain assemblages represent temporary phases of higher energy conditions.

It is difficult to account for the thickness variation of the ironstone unit; it may in part reflect variation in the pre-depositional surface, or it may represent slight downwarping of the central area. This will be considered again in relation to the entire Eller Beck Bed thickness variation. In the absence of any adjacent sections the unusual development near Littlebeck cannot be explained in terms of the known sections to the east and west, unless it lies on the site of a shallow channel which was the site of deposition which persisted through into the lower part of the shale unit.

b) Chemical Environment. The chemical conditions of deposition must have been such as to allow stability of chamosite and yet support an abundant fauna. It is probable, therefore, that slightly oxidising conditions prevailed. The stability of chamosite under such conditions is discussed later in Part III. The salinity may be estimated to some extent from the fauna; this is rather restricted in type, being confined to lamellibranchs, gastropods, and ostracods. The number of genera found throughout the ironstone amounts to about fifty, and so the faunal assemblage is not of very restricted type such as is associated with brackish water conditions. It is probable that the environment possessed open and shallow connection with the sea to the south but that an influx of fresh water from the north maintained a salinity slightly below normal.

B. DIAGENESIS

1. Diagenetic History

The diagenetic histories of the individual constituent types of the ironstone which have already been considered in detail will now be combined to give a broad outline of the diagenesis of the ironstone as a whole.

Sideritisation of the original clay is the most fundamental change that the original sediment has undergone, and in practically all localities replacement is complete. The replacement is clearly an early diagenetic process but when clay is the principal original component compaction is of particular importance and may be virtually complete during early diagenesis. The siderite may grow in the mud in two ways: crystals may grow into the interstitial cavities before compaction, or they may grow after compaction by replacement of the clay. In Hartoft Beck the siderite is clearly replacing a fairly well compacted chamosite mud and if this relationship may be extended to cover the rest of the ironstone then siderite replacement may be assumed to occur after the initial rapid compaction has taken place.

There are several features which suggest that compaction was not complete. Where chamosite has not been completely replaced the associated ooliths show a considerable amount of distortion and this is believed to result from strain during differential compaction of sideritic and chamositic patches within the rock. Another feature is the development of vertical slickensides on sideritic burrows at the base of the ironstone which proves that even the underlying shale underwent compaction after sideritisation; in non-sideritic burrows this is reflected in a slight crumpling of the burrows.

Another early diagenetic mineral is phosphate which in some north-eastern localities is clearly seen as an interstitial mineral in some siderite mudstones. The phosphate generally appears to be more or less contemporaneous with the siderite but occasional patches occur in which siderite crystals are rare, and which may therefore have preceded sideritisation.

Replacement of grains mostly occurs after sideritisation of the matrix. The etching of quartz grains, however, appears to have occurred either within the original mud or even during transport, though the highly irregular nature of some grains makes the second mechanism seem unlikely. Sideritisation of shells, ooliths, intraclasts and terrigenous grains is probably closely associated with sideritisation of the matrix, though it may not begin until the latter is complete. Pyritisation and sometimes solution of aragonite shells on the other hand probably took place in the original mud. Solution of shells and intraclasts followed partial siderite replacement while ooliths apparently were unaffected. In the calcareous ironstone no such solution takes place and the principal feature is the replacement of aragonite shells, intraclasts and quartz grains by ferroan calcite. The datesof these replacements are uncertain, but later replacements seem to be accompanied by a little compaction, and so calcite replacement may be roughly equivalent in time to solution in the other ironstones.

The solution cavities in non-calcareous ironstones are filled by siderite and later, kaolinite. A certain amount of shrinkage, represented by septarian cracks, and compaction occurred between the deposition of siderite and that of kaolinite. Rarely, calcite may take the place of siderite and kaolinite. In ooliths chamosite is replaced by kaolinite, probably at the same time as kaolinite is being deposited in cavities. Pyrite and possibly sphalerite appear to be the latest replacement minerals, the pyrite occurring in patches in which matrix and grains undergo partial or total replacement.

2. Chemical Environment

In order to save repetition the chemical environment in which diagenetic changes have taken place is discussed in a later chapter, in which the other Eller Beck Bed sediments are also considered.

C. EROSION

On the coast at Iron Scar the ironstone unit has suffered local erosion which appears to have occurred in two distinct phases:

(i) before sideritisation of the matrix; the non-marine shale is penetrated by unsideritised burrows, and overlain by silty shale;

(ii) after sideritisation of the matrix; the non-marine shale is penetrated by sideritic burrows, and overlain by shale.

The first type of erosion occurs as very local scours only a few feet across or as slightly larger scale features some tens of feet across. The limit of the siderite mudstone is always sharp and no gradual diminution in thickness occurs towards the scour. The silty shale which fills the eroded troughs may contain nodular ironstone (figs. 2.4, 2.5) which is distinct from the adjacent bedded ironstone.

The second type of erosion produces gradual thinning of the ironstone unit with occasional total removal, and is rather widespread in its effect. Although the sideritisation of the matrix must have taken place before erosion lithification was presumably not complete; this is implied by the smooth and rolling erosion surface and by the lack of pebbles.

At Hawsker erosion again appears to have occurred in two stages. In the first a trough about two feet deep was eroded into the underlying non-marine shales. On the margins of this trough burrows are preserved in non-sideritised silty clay. At a later stage erosion has affected the whole area removing the ironstone completely, but with the preservation of ironstone-filled burrows in the non-marine shale. This later phase of erosion is also seen in the East Cliff of Whitby. It may even by of slightly later date since the erosion is rather irregular, leaving patches of ironstone up to eight inches thick in places, while elsewhere erosion has cut down into the non-marine shale.

The development of the two phases of erosion is valuable in determining the time of siderite replacement relative to the deposition of the original mud and to the deposition of the overlying shale. The lack of sideritisation of the sediment in burrows underlying early erosional surfaces shows that sideritisation had not commenced, even at the base, until deposition was complete. The presence of the second erosional surface shows that a sufficient length of time elapsed for sideritisation to occur before erosion and deposition of the overlying shale.

CHAPTER V THE SHALE UNIT

I STRATIGRAPHY

A. FACIES DISTRIBUTION

The shale unit is usually easily recognisable throughout the area. It comprises dark grey laminated shale and silty shale which commonly contains ironstone concretions. Its lower boundary with the ironstone unit is sharply defined but its upper boundary is more arbitrary since there is a gradation upwards from shale into fine-grained sandstone with laminae of silty shale. The shales become increasingly silty upwards and near the top contain wisps or occasionally thin ribs of fine-grained sand. The overlying thin-bedded sandstone commonly shows a similarly slow increase in grain-size, with a gradual upward reduction in the number of partings of silty shale. The boundary between the dominantly shaly beds and the dominantly sandy beds is, however, usually quite sharp and may be readily picked out on weathered surfaces.

In the section from Iron Scar (1) to Ravenscar (2) the shale is developed in its typical form as outlined above, although ironstone concretions are virtually absent. Fossils are present in the lower part. At Hawsker (3,4), as described in the section on the ironstone unit, the base of the shale cuts down through the ironstone leaving only the burrowed non-marine surface. Locally the base cuts down further forming a shallow trough of about 2 feet maximum depth (fig.2.51) This trough is occupied at the base by a few inches of tough white sand with shaly partings which contains abundant small ironstone nodules in the lateral parts, and finally passes into shale with large unfossiliferous fine-grained ironstone nodules. Elsewhere the shale lies directly on non-marine beds. In the East Cliff of Whitby (5) there is further evidence of erosion at this level, since the ironstone is very irregularly developed. Except where patches of ironstone have remained the base of the shale is below the level of the base of the ironstone unit, sometimes to a depth of over nine inches (fig. 2.52).

The shale at Hawsker is fine-grained at the base but becomes very silty and carbonaceous in the upper part, and is tough and rather coarsely laminated. The silty shales pass upwards into fine grained silty sandstones, but the passage is marked by less interlamination than in other more typical sections. In the vicinity of the eroded trough at the base of the shale unit the shale shows features which are not seen elsewhere, and which are apparently limited laterally to about 100 feet of the section. In the lower, fine-grained part the shale contains scattered cross-laminated lenses of sand, apparently isolated ripples. The sand is fine-grained and very tough, and shows small scale loading at the base. In the upper part rafts of coaly material are very abundant. These measure up to 2 feet in width and up to about 2 inches in thickness.

In Whitby East Cliff (5) the shale is of more typical aspect, but fossils are again absent. Along the whole coastal section so far described the shale is relatively uniform in thickness, being between 4 feet and 4 feet 6 inches. In the West Cliff (6), however, the

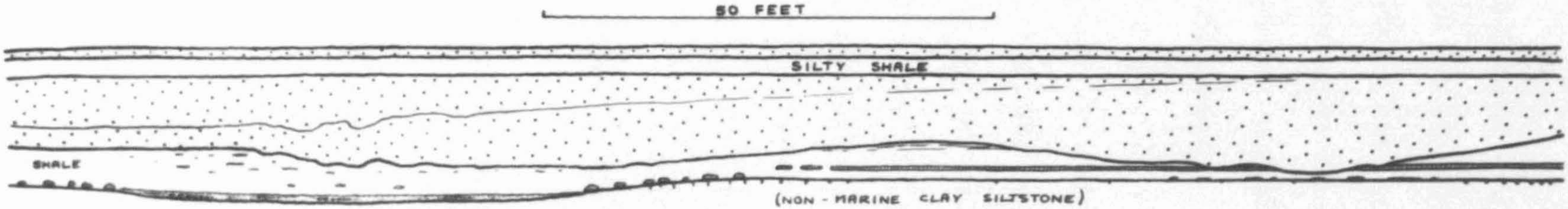


FIG 2.51 SECTION OF ELLER BECK BED EAST OF HAWSKER

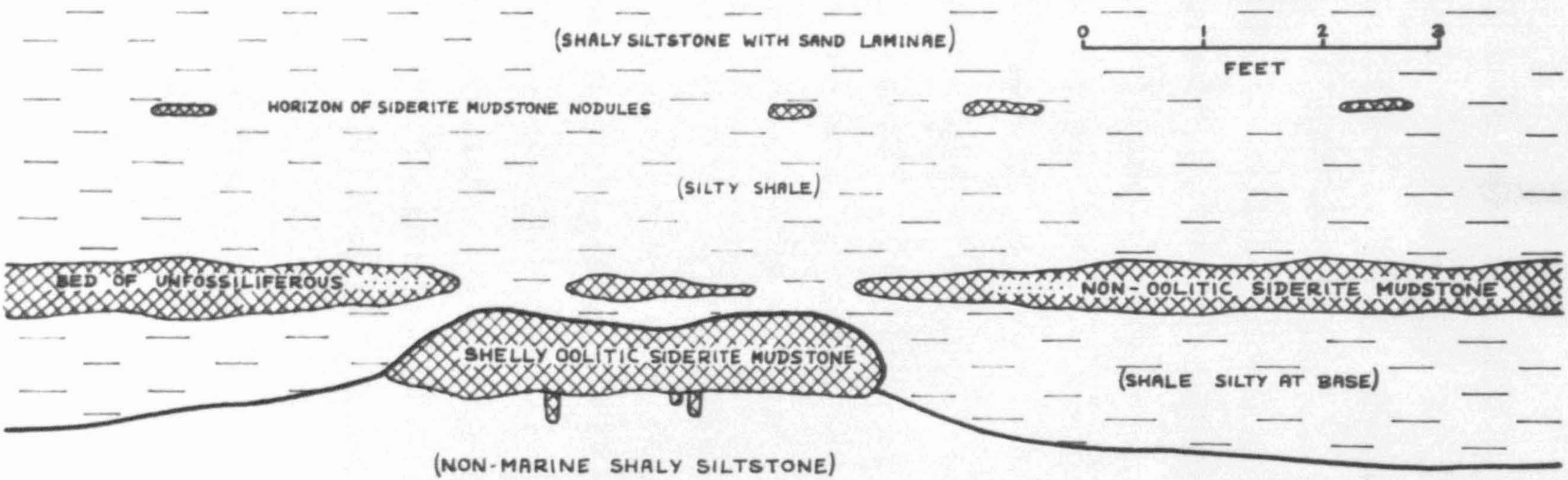


FIG. 2.52 EROSION OF IRONSTONE UNIT, EAST CLIFF WHITBY

shale thickness drops to 2 feet, though there is little lithological change and fossils are still absent. In Newholme Beck (7) the shale is still quite thin. The section in the waterfall is unweathered and the upper boundary not distinct, since the overlying thin-bedded sandstones contain abundant silty partings and appear dark due to the mdst conditions. In the similar section in the cliff near Hinderwell (8) the beds are weathered and the distinction between the sandy and shaly beds is much clearer. At Hinderwell the shale is only 1 ft.7 in. thick, but does contain rare shell casts near the base.

In Kilton Beck (9) the shale is very fine-grained throughout and does not show the usual upward passage into sandstone. The base of the sandstone is sharp, and since the lowest few inches contain flakes of shale it is probable that the upper part of the shale unit has been removed. In comparing the grain size of the shale with that of other sections it is possible that as much as two feet of shale, in which the upward passage into sandstone would have been present, may have been removed. Alternatively the shale may have been fine-grained throughout and erosion at the top only on a small scale. The overlying sandstone does not contain silty laminae and it is possible that the lower part of the sandstone is equivalent to the silty shale and thin-bedded sandstone as developed to the south.

At Commondale (27) the shale unit is again thin, and shows an unusual development. The shale is very fine-grained, splitting to thin laminae, but is interbedded with ribs of dark burrowed medium-grained sandstone. The upper contact with the sandstone unit is sharp but the lower part of the sandstone contains abundant partings of silty shale and is similar in lithology to the sandy ribs within the shale unit. Shell casts are very rare in the shale.

In the section near Little Beck (10) the shale horizon is partly occupied by a sandy facies. Above the horizon of ironstone nodules is a bed of pale coloured silty sandstone which is poorly bedded and contains abundant carbonaceous and micaceous laminae. Shells, chiefly oysters, are abundant locally. Above this sandy unit the normal shale is developed, being only 2ft. 6in. thick. The shale is silty and contains abundant laminae of fine sand in the lower part. The contact with the overlying sandstone is quite sharp and the sandstone is free from silty shale laminae although it is very thin-bedded.

Elsewhere in lower Eskdale the shale is typically developed as far west as Glaisdale. The shale varies in thickness but shows the upward passage into sandstone: shell casts are present in the basal part. The shale is about 4ft. 6in. thick in Eller Beck (11), but thickens to the west to about six feet (Section 12). This thickness is reduced in Egton Grange (15) and in the west of Wheeldale Gill (13) by channelling from above.

In Winter Gill (16) a thin shale development overlies the ironstone unit, comprising fine grained shale with no silty or sandy laminae. This shale is overlain by a pebbly sandstone with abundant wood fragments which passes upwards into a limestone composed almost entirely of shells. This, in turn, passes into a calcareous sandstone which is overlain by alternations of sandstone and pale grey fine-grained, unfossiliferous siderite mudstone. The lower beds show cross-bedding from both a northerly and southerly direction, there being no apparent preferred orientation. The sandy beds are overlain by laminated shaly micaceous siltstone shale with occasional shelly bedding planes, which passes quite rapidly into thin-bedded sandstone. The sandstone does not contain silty partings and becomes unbedded in the upper part, being a fine-grained, white, unfossiliferous, sandstone.

The limestone and calcareous sandstone almost certainly represent part of the shale unit and the upper sandstone probably belongs to the thin-bedded sandstone sub-unit (this is discussed later, p. 145). Since there is no section to compare with this, however, the micaceous siltstone might belong to either of these two groups, and as it contains shelly bands it cannot be correlated with the normal upper shale development. As the lower boundary is sharp and the upper gradational these silts are tentatively ascribed to the sandstone unit, the presence of shelly bands being supporting evidence for this. Calcareous sandstones overlain by shelly, thin-bedded sandstones occur in Great Fryupdale, and these may be analogous to those of Winter Gill.

In Glaisdale Head (17) the normal facies is present with only four feet of shale present. This thickens to about five feet in Birk Wath Beck. In Busco Beck (20) the shales are 6ft. 9in. thick and show the usual vertical lithological variation. They are, however, capped by a thin fine-grained sandstone with a sharp basal contact. Although the sequence above this is still shaly in part the base of this sandstone is taken as the upper limit of the shale unit. In Stonegate Gill the shale is 7ft. thick.

At the head of Great Fryupdale (21) the shale horizon is occupied by ferruginous green shelly sandstone, but passing westwards to The

Scar (22) the sandstone becomes more variable. In the most southerly exposure a foot of pale coloured silty shale occurs above the ironstone unit; this is overlain by a thin greenish pebbly sandstone and then a tough calcareous sandstone. These sandstones are in turn overlain by 2ft. 6in. of normal silty shale which passes upwards into thin-bedded shelly sandstone. When traced northwards the lower sandstones thicken by cutting down through the underlying shale and ironstone; at the same time the top of the sandstone rises and the overlying shale and even the lower part of the thin-bedded sandstone show progressive cutting out against the sandstone surface. The lower sandstone reaches a maximum thickness of seven feet, and as it becomes thicker cross-bedded units are developed with coarse-grained shelly horizons. The variation along this section is shown diagrammatically in figure 2.53.

In Danbydale the sandy development is maintained in Old Hannah's Nick (23) where seven feet of sandstone is overlain by 1ft. 6in. of silty shale with sandy laminae. Near the head of Danbydale the sandstone has broken up into thin units separated by siltstone and silty shale. The lowest bed is a soft pale shelly siltstone which is possibly equivalent to the pale silty shale of Great Fryupdale (22). Near the top of the exposed beds is a thin oolitic ironstone with occasional shells.

In Battersby Crags (28) the shale unit is absent, apparently due to erosion, but to the south in the Ingleby escarpment it is again developed. In Rud Scar below Burton Howe (29) the ironstone unit is overlain by shelly siltstones at the top of which is 1ft. 6in. of

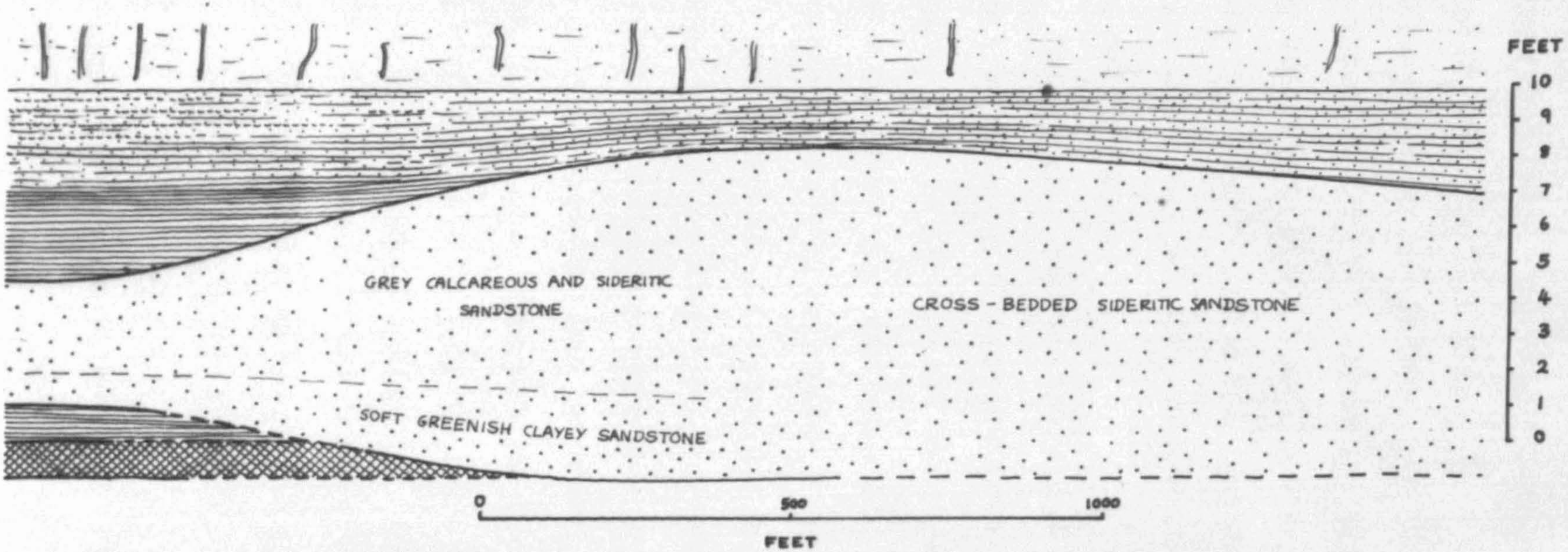


FIG. 2.53 DEVELOPMENT OF SANDSTONE LENS WITHIN SHALE UNIT, THE SCAR GREAT FRYUPDALE

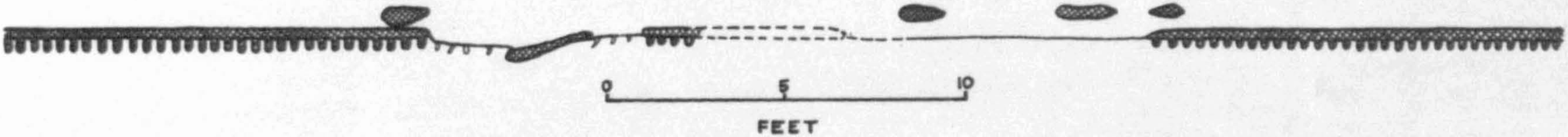


FIG. 2.54 IRONSTONE CONCRETIONS ASSOCIATED WITH EROSION OF THE IRONSTONE UNIT (I. SCAR)

oolitic ironstone. The total thickness of these beds is 5ft. 4in. compared with 5ft. 10in. between the ironstone unit and the top of the oolitic ironstone in Danbydale. In Bluebell Trough (30) the siltstones are reduced to 2ft. 8in. and the ironstone to 1ft. 2in.; in Farndale they have thinned further, to 1ft. 6in. and eight inches respectively. In the last two localities the siltstones are shelly, and pale in colour with a purplish tint, being similar to the lowest units in Rud Scar and Danbydale.

The overlying beds are not exposed in Rud Scar, but in Bluebell Trough seven feet of fine-grained unfossiliferous laminated shale are present, which in Farndale has thinned to 5ft. 9in. The section in Blow Gill, to the south-west shows 2ft. 9in. to 3ft. 3in. of oolitic ironstone overlain by twelve feet of shale.

In Rosedale the shale reaches a maximum thickness of 7ft. 8in. (Section 32) and shows the typical development of sandy laminae in the upper part. Beneath the shale is a thick series of siltstones with a thin oolitic ironstone near the middle. To the east (33) this ironstone is four inches thick with ten inches of silty shale below, while the overlying shale is silty with ribs of sandstone locally. The unit is exposed at several localities in Northdale in which the shale is about 4ft. 6in. thick. The oolitic ironstone is developed thinly in the south but dies out northwards while the underlying silty shale thins from 1ft. 6in. to less than one foot thick. In Rosedale Crags (34) the sideritic sandy beds which may represent the ironstone unit are overlain by dark, poorly-bedded fossiliferous silty shale which is capped by a thin slightly oolitic ironstone. The overlying shale is seven feet thick and contains sandy laminae in the upper part. The thin oolitic ironstone may be traced into Hartoft Beck where it overlies a thin shale in which a thin shelly ironstone is developed locally.

B. DISTRIBUTION OF IRONSTONE CONCRETIONS

Concretionary ironstone in the form of siderite mudstone nodules is fairly common though seldom abundant. The nodules are practically always small, up to about six inches in length, but under special circumstances larger concretions and even continuous concretionary bands are developed. Ordinary nodules are absent between Iron Scar and Hawsker but are present in the East Cliff of Whitby and in sections to the north.

In Eskdale isolated nodules occur within the shale and in Eller Beck and New Wath Scar lines of concretions are developed, with a locally continuous shelly ironstone in the former locality. In northern Rosedale small concretions are quite abundant locally, but are rare in the south. Ironstone concretions are totally absent in Farndale and in Bluebell Trough.

Unusual concretionary developments are seen in the coastal sections. One of these which occurs near the top of the shale at Iron Scar is associated with a compressed log (fig. 2.57), but the remainder are associated with erosion of the ironstone unit. Some of these have already been described (figs. 2.4 and 2.5), while the remainder consist of large isolated nodules which occur a short distance above erosion surfaces in which silt-filled burrows are present (fig.2.54)

At Hawsker large isolated nodules are developed above the erosion surface on the margins of the trough (fig. 2.51), where silt-filled burrows are again present. These nodules die out when ironstone-filled burrows appear and at the same time a continuous ironstone band is developed at a higher level about one foot from the base of the shale. A similar band occurs in the East Cliff of Whitby (fig. 2.52) at a maximum height of 1ft. 6in. from the base of the shale; this band thins and becomes discontinuous over patches of uneroded basal ironstone. The ironstone band is absent in the West Cliff, where the basal ironstone is not eroded, but reappears at Hinderwell though it is not continuous.

C. PALAEONTOLOGY.

The fauna of the shale unit comprises lamellibranchs and gastropods; ostracods are also seen in a few places. In shale the fossils are preserved as compressed moulds, occasionally with a thin pyritic film; in the calcareous sandy beds they are preserved in calcite or as uncrushed moulds. Identifiable forms include *Gervillia*, *Trigonia*, *Pleuromya*, *Pholadomya* and *Turritella*, all of which are found in south-eastern coastal districts and in northern Rosedale. As in other areas the fossils occur chiefly in the lowest part of the shale. In Eskdale fossils are rare, and on the coast at Hawsker and the East Cliff of Whitby they are absent. In Farndale and in the Ingleby Escarpment the shale is completely unfossiliferous except where a few intercalations of pale silty shale occur at the base.

Local concretionary ironstones may be very fossiliferous, as at Eller Beck where gastropods are particularly abundant. The sandy siltstone of Littlebeck is fossiliferous, the chief forms being *Liostrea* and *Pinna*. The limestones and sandstones of Wintergill and Great Fryupdale are characterised by oysters, though other undeterminable lamellibranchs are also abundant and a few ostracods have been observed. In the sandstone ribs and beds of Danbydale and Commondale show evidence of burrowing. The oolitic ironstone at Danbydale contains *Trigonia*, *Modiola*, *Liostrea*, and several other small lamellibranchs, and is extensively burrowed. A similar fauna occurs to the south and south-west.

In southern districts particularly a small leaf is common within the shale unit. The leaf which reaches about one inch in length has not been identified.

D. CORRELATION

The shale unit is usually easily correlated on its lithology but the development of siltstone and sandstone at some localities introduces several problems. At Littlebeck the silty sandstones appear to have developed during and after the deposition of the ironstone unit. The thinness of the overlying shale suggests that the sandstone formed an upstanding bank although initially a scour may have occurred on the site (p. 104).

The sandstones and limestones of Winter Gill and Great Fryupdale clearly post-date the initiation of shale deposition and the base of these beds are clearly erosional to some extent. No dating for the Winter Gill beds can be made, except that in the absence of any development of typical shale, accumulation of sand may have occurred until the end of normal shale unit deposition. The sandy siltstone overlying the sandstone is over seven feet above the ironstone unit and since in the neighbouring Glaisdale Head section the shale unit is only four feet thick it is probable that the sandy siltstone belongs to the sandstone unit.

In the Egton Grange section the shale is similarly eroded and overlain by four feet of rather coarse pebbly and shelly sandstone with impersistent beds of slightly oolitic siderite mudstone which are probably equivalent to the Winter Gill beds. The overlying finer-grained sandstones are typical of the sandstone unit. In Great Fryupdale and Danbydale the sand accumulation has clearly begun after the initiation of shale deposition, but in the area of The Scar, at least, has ceased early enough to allow deposition of over two feet of overlying shale. The shale preserved below the sandstone is unusually pale and silty and is probably of the same age as the basal siltstone of Danbydale. The remainder of the siltstones and sandstones appear to be equivalent to the sandstone of Great Fryupdale. Near the top of the section is a thin oolitic ironstone which in its general setting is similar to that of the south-western localities, and it is here suggested that the oolitic ironstone is continuous throughout. It is also traceable throughout Rosedale and into Hartoft Beck. The underlying beds thin to the east and in eastern Rosedale and Hartoft Beck are represented by about one foot of shale; in this case the ironstone and underlying beds are probably equivalent to the lowest foot of shale in the sections to the north-east. The distribution of these beds is represented in figure 2.55.

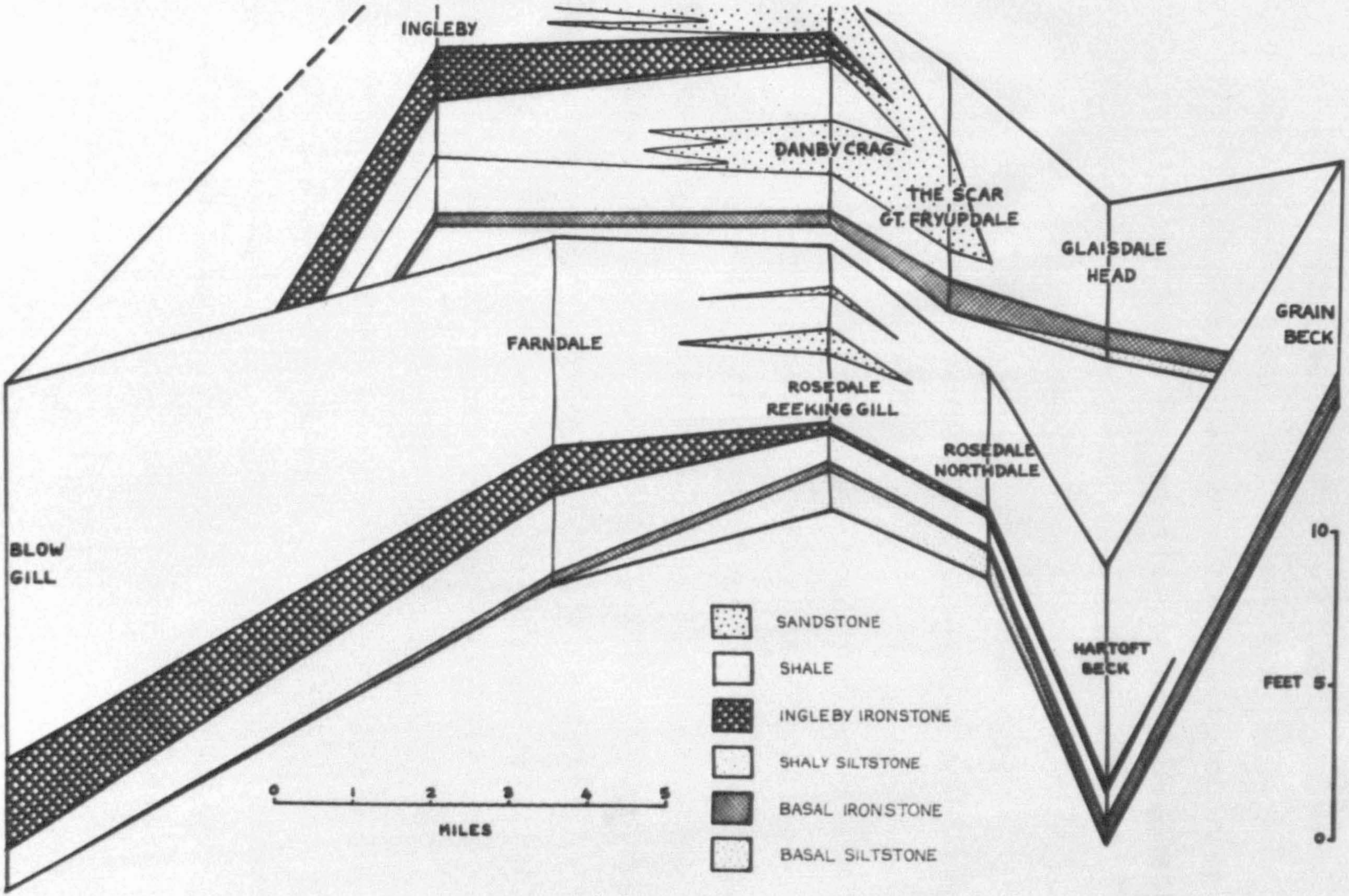


FIG. 2.55 DISTRIBUTION OF THE INGLEBY IRONSTONE SEAM

The ironstone clearly increases in thickness to the south-west, and probably correlates with the thick ironstone developed in Blow Gill (38). If this is the case the ironstone develops into a major unit to the south-west and had the area studied been extended in this direction, the ironstone and associated beds would have warranted separation into a separate unit. The ironstone will be referred to in this text as the 'Ingleby Ironstone' which follows Marley's nomenclature for the ironstone of Rud Scar, although that included the thin representative of the ironstone unit.

The shale at Hawsker is sufficiently different to that of other areas to suggest a different mode and possibly time of deposition. The basal scour is clearly post-ironstone unit in age and the loading at the base of the sandstone indicates lack of shale consolidation which suggests that deposition in this region was later than elsewhere. Deposition would have been more rapid and this is also suggested by the relatively poor sorting of these shales. A regime of water flow which removed the ironstone unit appears to have continued during shale deposition which was retarded in this area until rapid deposition brought the level of sedimentation up to that of surrounding areas.

Figure 2.56 represents the distribution of the shale unit excluding the Ingleby Ironstone and the associated silts. Numbers in brackets represent thicknesses which have been reduced by erosion, and the thickness for Great Fryupdale is taken from Section 22, where the sandstone bed is thinnest.

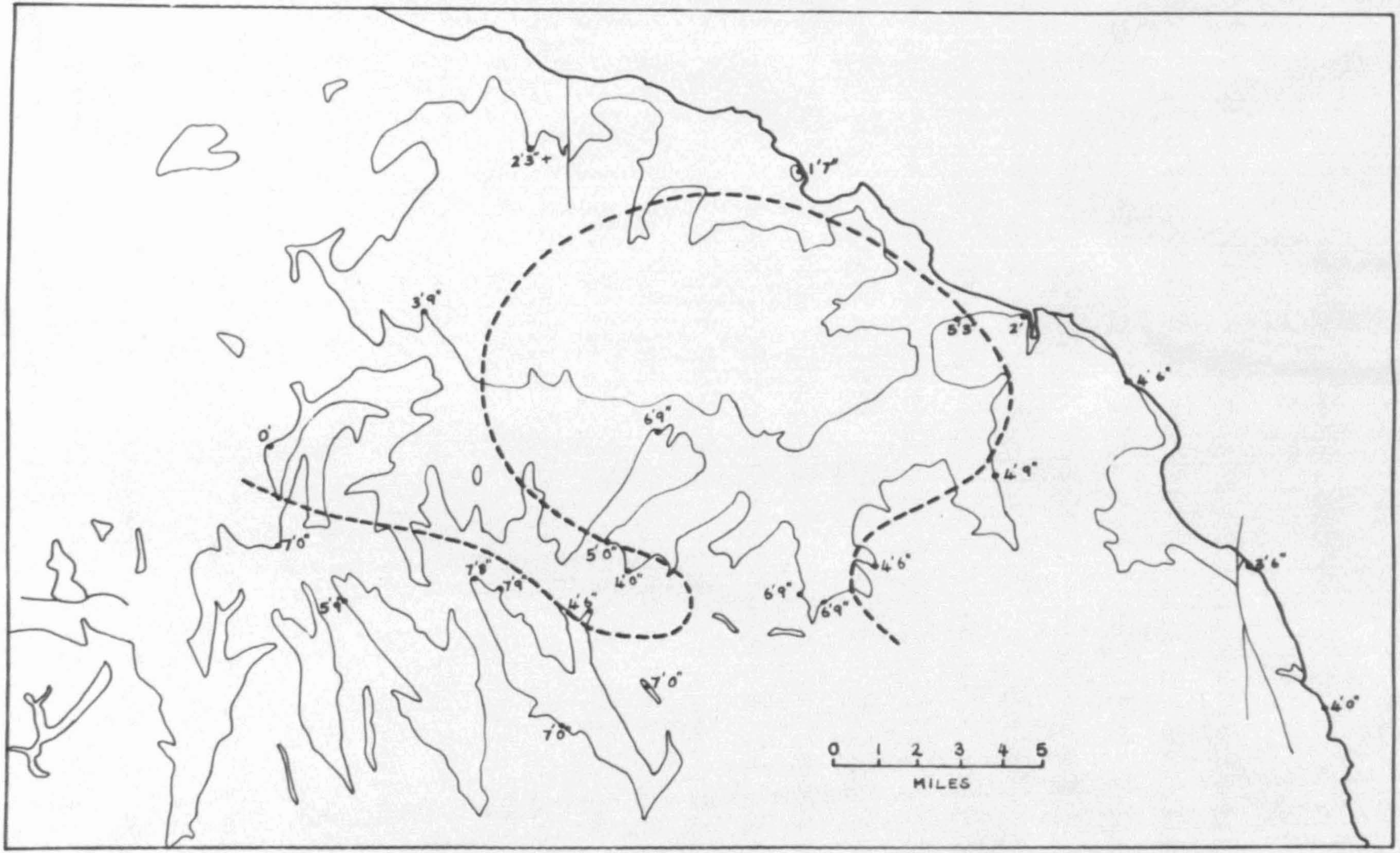


FIG. 2.56 ISOPACHYTE MAP OF THE SHALE UNIT

II PETROGRAPHY AND PARAGENESIS

A. SHALE

The shale has not been studied in thin section, but examination under a binocular microscope has provided the following observations on the texture and grain-size of this deposit. At the base of the unit the shale is fine-grained but contains thin laminae of silt; (exceptionally the basal few inches are very silty due to erosion of the ironstone unit.) The silty laminae become thicker and more abundant upwards, and in the upper part of the unit thin whitish-weathering laminae of silt and fine sand develop, while the associated shale becomes rather silty. The top of the shale is characterised by abundant laminae of silt and fine sand and occasionally thicker ribs of sand may occur: the unit then grades rapidly upwards into the base of the sandstone unit which comprises fine sandstone with thin laminae of silty shale. Although the term 'shale' has been used for field description the bulk of the sediment probably comprises clayey siltstone.

The clay material was identified by means of X-ray diffraction (Appendix 2 B iv) and proved to consist entirely of kaolinite. The bulk of the grains consist of quartz sand and silt together with a little fine-grained mica. Ooliths are apparently absent and the only allochemical constituents of any importance are shells.

Sedimentary structures are usually confined to horizontal lamination but occasional ripple-marked lenses of fine sand occur. Where these are in the finer grained shale, as at Hawsker, small-scale load casts may be developed at the base. Larger load casts, generally of a smooth nature, may occur at the base of thick sandstone ribs at the top of the unit and occasionally, as in Newholme Beck, these may characterise a particular horizon. Organic structures are confined to a few localities: at Eller Beck indistinct horizontal trace fossils are present and in southern coastal sections vertical sand-filled cylindrical tubes of Arenicolites occur.

B. IRONSTONE CONCRETIONS

Concretionary ironstone may develop at any level within the shale unit. Generally the distribution is haphazard, but in several cases the ironstone is clearly related to some other feature, such as shell beds or wash-outs. These more specialised developments will be described first.

(i) At Iron Scar ironstone nodules occur within silt-filled channels on the horizon of the bedded ironstone. The siderite matrix is very similar to that of the associated ironstone, comprising a granular mosaic of crystals averaging 15 microns in diameter, some having opaque cores. Terrigenous quartz grains are more abundant totalling about twenty percent of the rock, and are poorly sorted due to the abundance of very fine grains. Scattered ooliths and a few shell fragments are present, together with fragments of wood. The small wash-outs in which these nodules occur are believed to have occurred before sideritisation of the bedded ironstone (p.150). The channel-fill is therefore likely to be resorted mud from that horizon, and this is borne out by the presence of ooliths and the similarity of the matrix to that of the eroded ironstone. The nodules may therefore have formed at the same time as the ironstone unit became sideritised.

(ii) The section at Eller Beck is characterised by a rather persistent bed of concretionary ironstone. The matrix consists of finely crystalline siderite (3 to 5 microns) with occasional rather diffuse cores; silt grains are abundant, comprising fifty percent of the rock. Burrows are present which in contrast to the bulk of the ironstone are almost silt-free, and the siderite is very fine-grained (2 microns). Shell material is abundant, and shows partial sideritisation and clay mineral cavity filling similar to that already described for the ironstone unit. This ironstone appears to have formed as a normal nodular ironstone, its continuity being due to a shelly horizon having acted as the nucleus for siderite precipitation. The time of formation predates shell solution and compression as is seen in the surrounding shale.

(iii) In New Wath Scar another rather persistent ironstone horizon is present which has many features in common with the basal ironstone unit. The matrix consists of clear siderite crystals (10 microns in diameter) with abundant well-defined cores. Terrigenous grains form less than five percent of the rock and shell material is rare. Chamosite ooliths, most of which possess quartz grain nuclei, are present in small numbers together with intraclasts of chamosite mud. The low percentage of terrigenous grains and the presence of chamositic ooliths and intraclasts suggests that, like the ironstone unit, this horizon represents the sideritisation of a chamositic mud. The chamositic environment must have been very localised since the bed cannot be traced laterally into adjacent localities. It is unlikely that the ooliths are derived from a more distant source since in the known oolitic horizons the ooliths

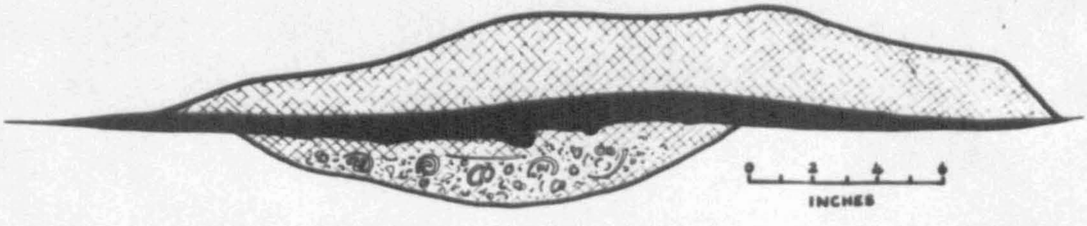


FIG. 2.57 IRONSTONE NODULE ASSOCIATED WITH COMPRESSED LOG, IRON SCAR

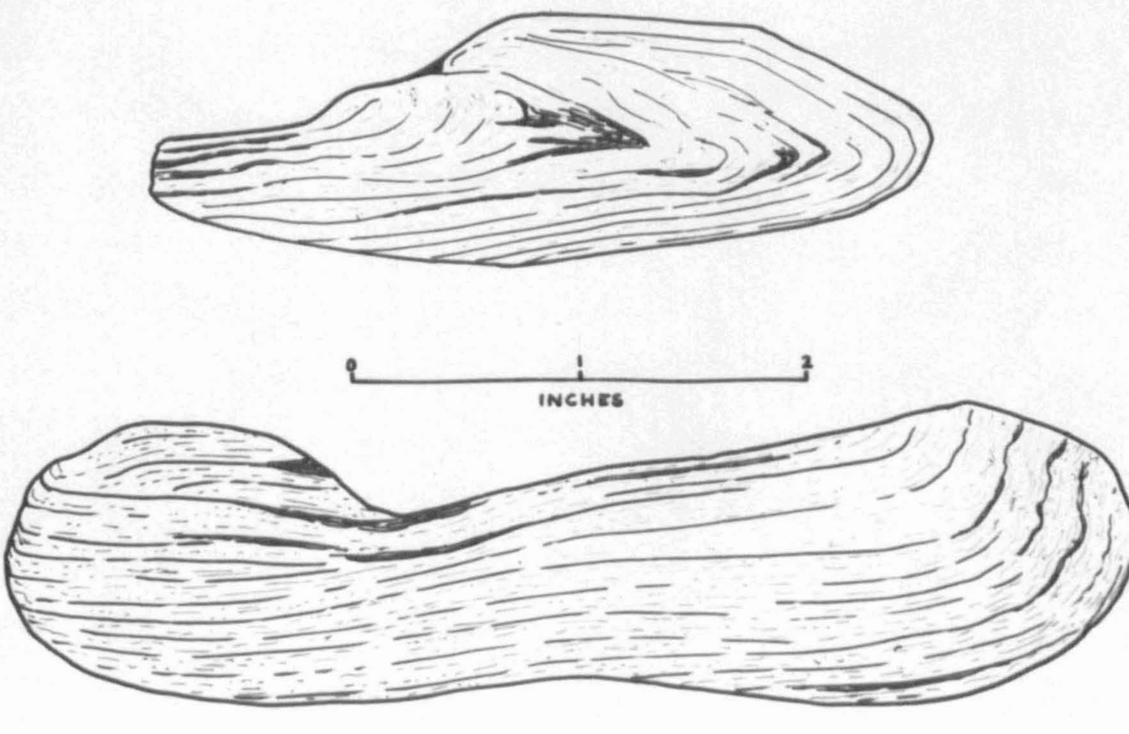


FIG. 2.58 CONTORTED SANDY LENSES, BLUEBELL TROUGH

ironstones possess rather finely crystalline siderite matrixes (2 to 5 microns) in which cored crystals may or may not be present. Lamination is present, and is chiefly due to variation in the proportion of terrigenous grains which range in grain size from coarse silt to very fine silt. These concretions clearly result from sideritisation of the normal shale. The relative compaction in the adjacent shales is usually less than thirty percent, indicating that siderite has developed in the later stages of compaction. The concretions clearly represent redistribution of iron from within the shale, but the exact nature of this process is difficult to assess. It is, however, interesting to note that the continuous ironstone bands and nodules which form above the areas of erosion of the underlying ironstone occur at some distance above the erosion plane. Where the erosion appears to have preceded sideritisation of the basal mud, scattered nodules occur a few inches above the erosion plane (figs.2.52 and 2.54), while where the erosion has post-dated sideritisation more continuous ironstone bands occur up to 1ft. 6in. above the erosion surface. Such continuous bands are seen at Hawsker, the East Cliff of Whitby, and Hinderwell; where remnants of the ironstone unit remain the band breaks up into nodules (fig.2.52). At Iron Scar such bands are not developed, even though the ironstone is eroded at several points along the cliff. The clear relationship of these bands with the underlying beds suggests that the migration of iron may be in an upward direction and considering the particular setting, the large amount of iron required to form such bands may be derived from the original basal mud which might have remained to some extent disseminated throughout the lower part of the shale unit. This implies that migration of iron in the shale must occur some considerable time after deposition.

The distribution of concretionary ironstone other than that related to the basal ironstone unit must be related to the iron content of the deposited mud. It shows significant regional variation as well as local variation, as may be seen from figure 2.58. The deposition and post-depositional migration of iron will be considered more generally in chapter IX.

C. SMALL SAND BODIES

The upper part of the shale is almost invariably characterised by laminae and lenses of fine silty sand; the following examples are considered in detail as they show more distinctive features.

(i) In the upper part of the shale in the section at Bluebell Trough (30) small lenses of coarse silt occur. These may show cross-lamination and always show clear evidence of loading, with the margin of the lens completely overturned (fig. 2.59). In thin section the grains, which average 50 microns in diameter and are very well sorted, show strong suturing in patches throughout the rock. Between these patches a clay matrix is present. These bodies appear to represent isolated ripple-marks which have foundered into the underlying shale; they may be derived from near the top of the unit since they are very similar to the overlying silty sandstones.

(ii) At Hawsker (3) rare sandy lenses occur near the middle of the shale unit in the region of the erosional trough. In hand-specimen these lenses are dark and tough; in thin section the dark colour is

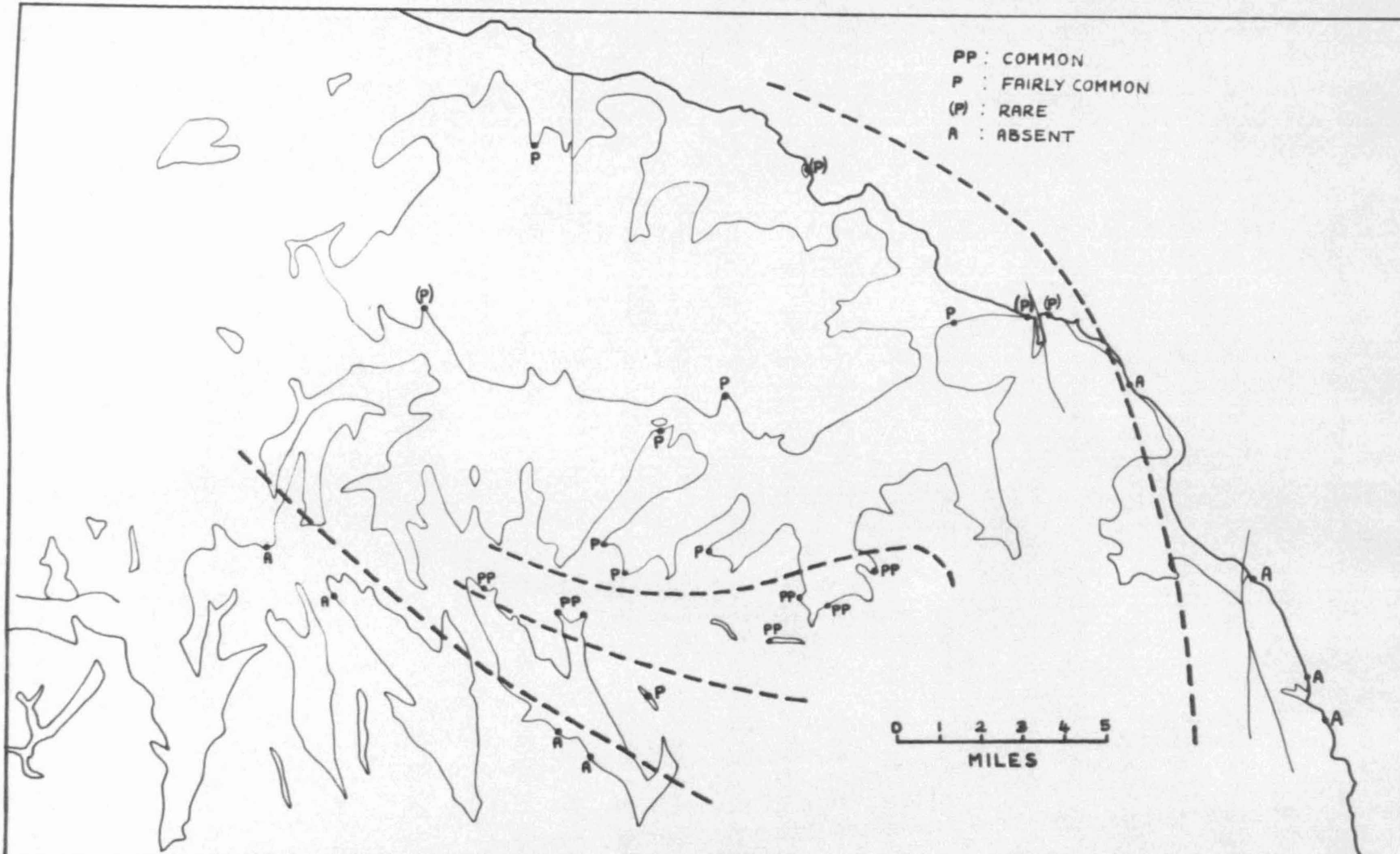


FIG. 2.59 DISTRIBUTION OF IRONSTONE NODULES NOT ASSOCIATED WITH EROSION OF IRONSTONE UNIT

seen to be due to abundant interstitial dusty pyrite and the toughness to intense grain suturing (fig.2.60). The sand grains average 150 microns in length, but many reach about 250 microns. Heavy minerals are fairly common, zircon and leucoxene being the chief constituents with very subordinate tourmaline and rutile. A broken chamosite oolith was also observed. These sand bodies appear to represent small lag deposits within what is otherwise a finely silty shale.

(iii) At the base of the channel referred to above, a thin sandstone is developed. In thin section the sandstone is very similar in grain size and texture to that described above; it does, however, lack the interstitial pyrite and is characterised by considerable concentrations of heavy minerals along certain laminae (fig.2.61). The heavy minerals again comprise zircon and leucoxene with minor amounts of rutile and tourmaline, but in addition slightly oxidised and fragmented siderite spherulites are present. Excluding siderite, the heavy minerals constitute over 5 percent of the total rock and in some laminae as much as 25 percent. This also clearly represents a lag deposit, the material being derived from the ironstone unit and the underlying non-marine beds.

D. WINTER GILL SANDSTONE

This constitutes one of the two areas of large-scale sandstone accumulation within the shale unit. Its areal extent is unknown since it is seen in only the one locality; it is absent at Glaisdale Head but may be represented by the lower part of the sandstone in Egton Grange, which is rather coarse-grained and contains oysters and intra-

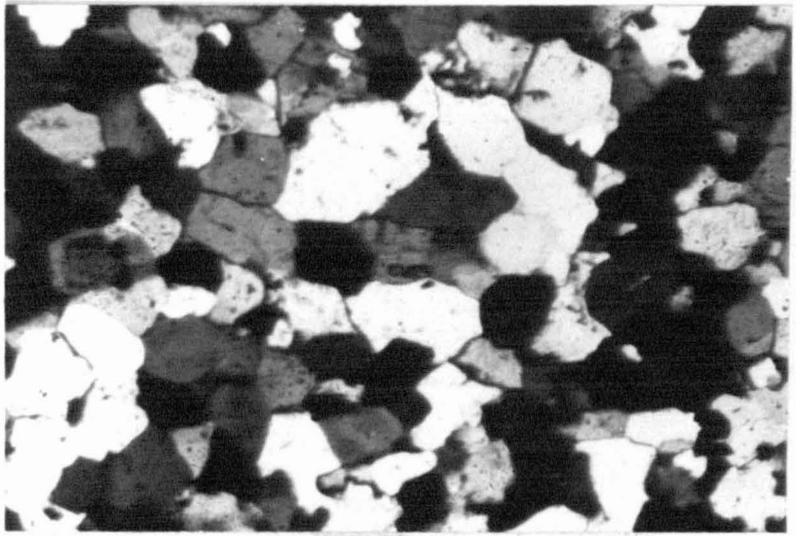
FIG. 2.60 Grain suturing in small-scale sand lens. (Nicols crossed; x70)

FIG. 2.61 Heavy mineral concentration in small sand lens (x140)

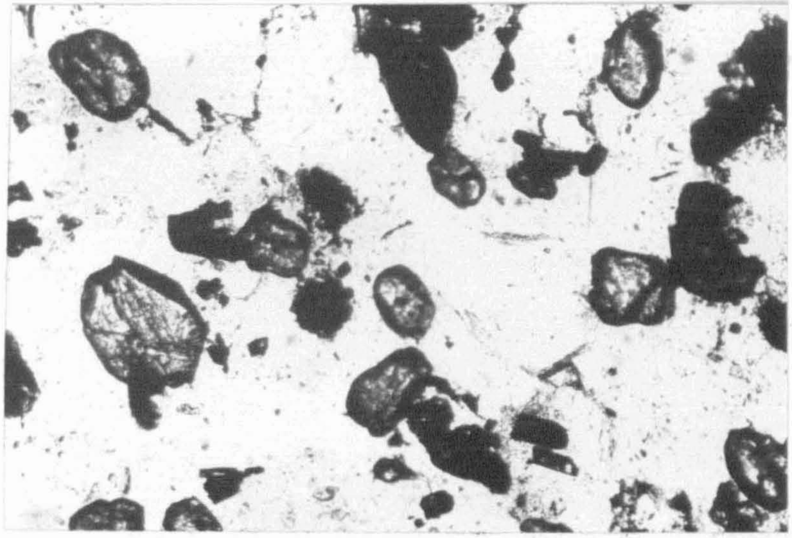
FIG. 2.63 Shattering of quartz grains, Winter Gill sand lens (x40)

FIG. 2.64 Cracking of quartz grain due to pressure against another (x80)

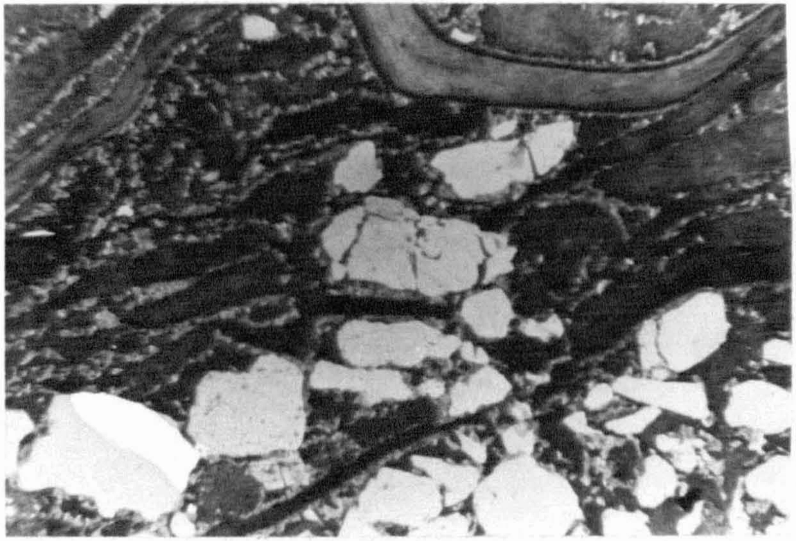
FIG. 2.60	Grain suturing in small-scale sand lens.	(Nicols crossed; x70)
FIG. 2.61	Heavy mineral concentration in small sand lens	(x148)
FIG. 2.63	Shattering of quartz grains, Winter Gill sand lens	(x40)
FIG. 2.64	Cracking of quartz grain due to pressure against another	(x80)



2.60 500μ



2.61 100μ



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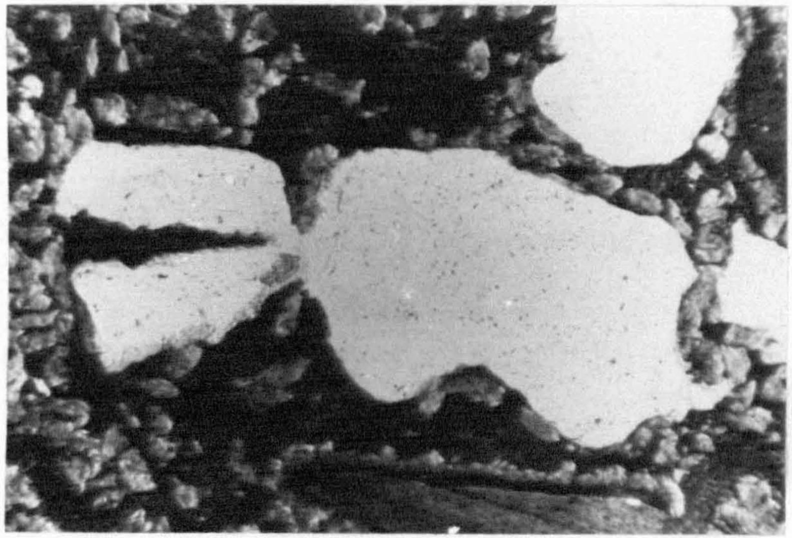


2.64 500μ

clasts. In both Egton Grange and Winter Gill the sandstone has an erosional base, 1ft. 10in. of shale being preserved in the former and 1ft. 4in. in the latter.

In Winter Gill the lowest beds show rapid lateral variation; in the downstream section they comprise highly weathered cross-bedded ferruginous sandstones overlying a thin rubbly bed, with shells, wood fragments and intraclasts, while upstream a shelly limestone is present. The shelly limestone passes upwards into a calcareous shelly sandstone which is locally oolitic and finally into sandstone with thin beds of siderite mudstone.

The basal limestone rocks owe their highly calcareous nature to the abundance of shells and to calcite cementation; no original calcareous mud is present. Apart from shells grain types include terrigenous sand, intraclasts and scattered ooliths. In many respects the petrography of the limestone resembles that of the calcareous ironstone of Great Fryupdale (p.150), to which reference will be made in order to avoid repetion. The Winter Gill limestone differs in having sparry cementation, comprising an early siderite rim followed by a ferroan calcite cement which has filled the remaining voids. The overlying sandstone is calcareous at the base, but for the most part has a sparry siderite cement.

1. Terrigenous Grains

a) Distribution and Composition At the base of the limestone terrigenous grains make up as little as five percent of the rock, but this increases upwards to about sixty percent. The average grain size decreases upwards and this is due to the presence in the basal units of a bimodal distribution of which the coarse mode decreases rapidly upwards (fig.2.62). The coarse mode is believed to represent a lag deposit marking the initial stages of the agitated conditions which must have prevailed in order to produce such a sand body. The fine, persistent mode appears to represent the incoming material from which the basal lag deposit is derived; it is well defined and persists to the top of the sandstone without any significant change in grain size. In view of the fact t at the nature of the deposit changes considerably, from a shelly limestone with a small amount of sand to a sandstone interbedded with siderite mudstone, it would seem that the constancy of the mode is a reflection of the nature of the source material or of the transport mechanism rather than of the depositional environment.

b) Diagenesis. In the Winter Gill limestone the sand grains show little replacement but some grains, particularly compound grains, show veining by ferroan calcite which is apparently associated with cracking since a few grains have apparently undergone disintegration in place (fig.2.63). Fracture and displacement is clearly the principal factor in the production of these features since fragments of the same original grain show varying positions of extinction. The crack may wedge out rapidly into the grain; the grain then shows different extinction on either side of the croack but where the crack dies out the grain shows strained extinction which is transitional between the two. These wedge-shaped cracks in particular are associated with the points of contact of quartz grains in which one grain has acted as a fulcrum in the breakage of the other. The strained portion of the broken grain apparently represents recrystallisation under pressure and very

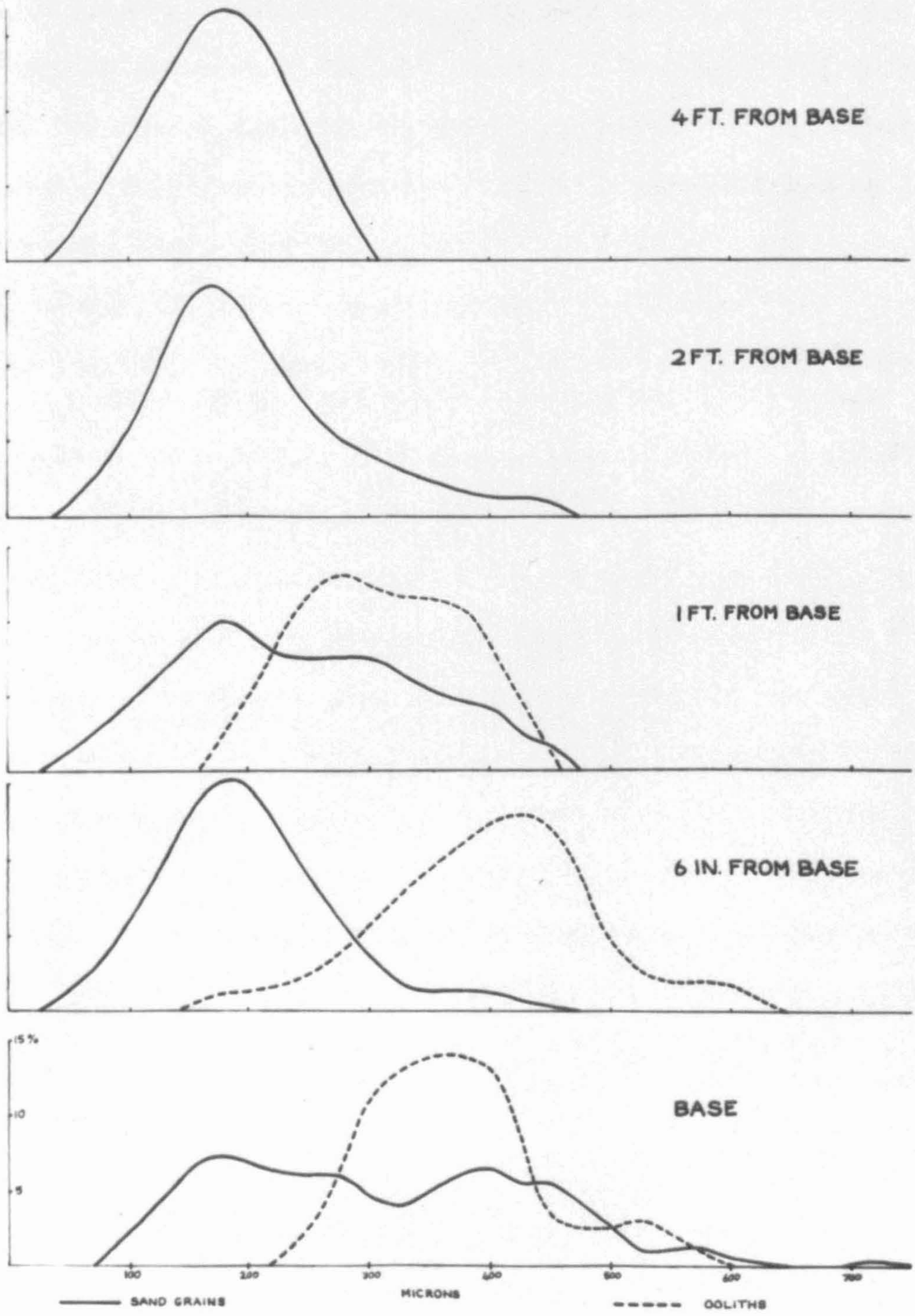


FIG. 2.62 GRAIN-SIZE FREQUENCY CURVES, WINTER GILL SAND LENS

commonly the two grains themselves show total fusion at the point of contact (fig.2.64). Radiating cracks may develop from the point of contact and it seems likely that all cracks are associated with grain contacts, the type of fracture depending on the immediate structural environment of the grains concerned. There are two mechanisms by which these fractures might be developed;

(i) by open fracturing under compaction pressure;

(ii) by partial replacement of the grains by calcite followed by crystal growth which causes fragmentation of the grain.

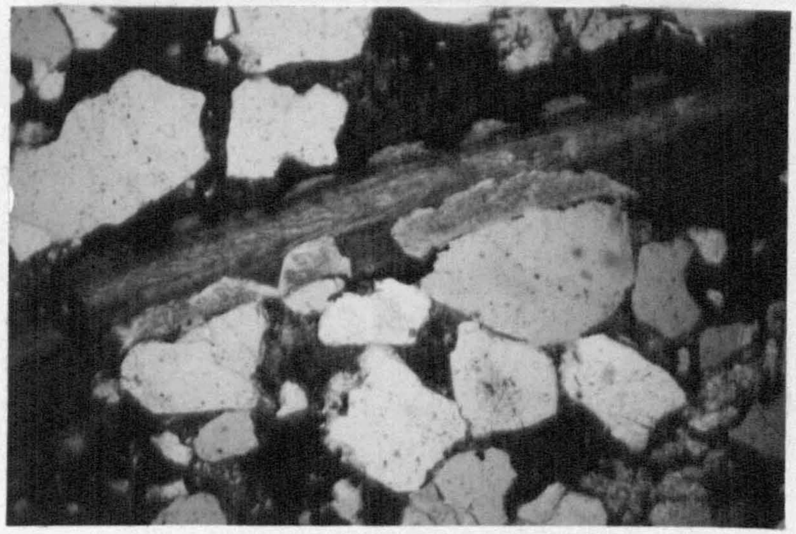
Since a calcite vein consists of a single crystal which is in optical continuity with the adjacent matrix it is probable that the fracture was an open crack, produced by compaction, which has been filled with calcite at the same time as the original voids. The association with grain contacts also favours grain fracturing. Siderite rims (see later) are broken by the cracks and the occasional crystals of siderite occurring within the fractures appear to have fallen in. Fracturing of quartz grains is confined to the lowest nine inches of the limestone and does not occur in the overlying more sandy beds. Breakage seems to result from pressure of overburden and is probably favoured by the inhomogeneity of the rock, with scattered resistant grains and a large original void space.

A certain degree of replacement of quartz grains by calcite does occur, and this often takes the form of long vein-like protrusions from the matrix which do not cause fracturing or displacement of the grains involved. Replacement may have taken place along incipient fracture lines. In the non-calcareous horizons quartz grains show irregular marginal replacement by siderite crystals. Feldspar grains to that of the calcareous facies of the ironstone unit including pyritisation of aragonite shells and replacement of original calcite shells by quartz (fig.2.65) and, rarely, dolomite (fig.2.66). The original structure of the replaced shell material is preserved in detail in the quartz overgrowths except in the outermost zone, where the quartz is clear. Quartz replacement post-dates siderite rim formation, since ghosts of the rim may often be seen, and pre-dates shell fragmentation, since the overgrowths may be dislocated from their host grains. The dolomite crystals are associated with a flaky mineral as in the basal ironstone; the example figured is particularly striking for the sharpness of the contact of calcite and dolomite which appears to reflect the exact path taken by the margin of the mineral flake. The dolomite crystals also show deformation of siderite rims and are in optical continuity with the adjacent calcite matrix. It appears that the dolomite has grown after siderite rim formation and after calcite cementation: pressure on the flaky mineral, produced by the crystal growth, has caused solution of the calcite shell and pressure on the siderite rim has caused solution of the calcite matrix.

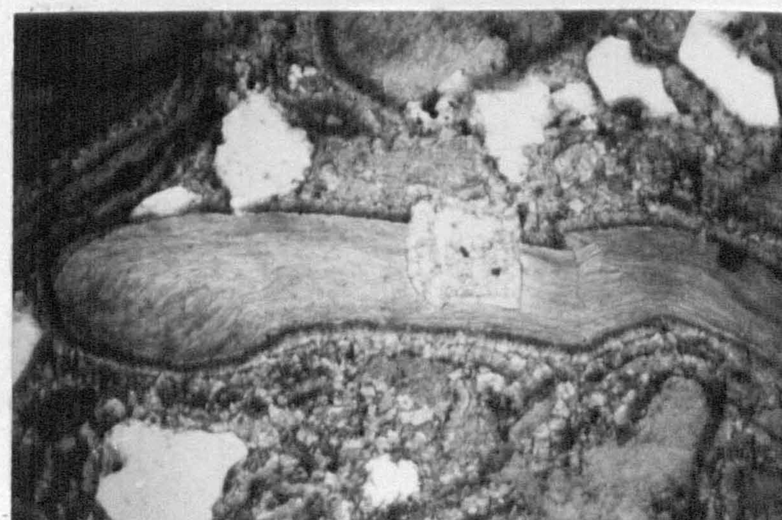
A small amount of sideritisation of non-ferroan calcite shells accompanies the formation of siderite rims; this is represented by a thin layer of dark brown cryptocrystalline siderite which occasionally projects into the shell. In the upper beds, where a calcite matrix is absent some shells are replaced by phosphate comprising a disoriented aggregate of finely acicular crystals reaching 10 microns in length, and which show slight birefringence.

FIG. 2.65 Quartz overgrowths, on detrital grains replacing non-ferroan calcite shell (pink) (x100)

FIG. 2.66 Dolomite (pale blue), "preceded" by clay mineral flake, replacing non-ferroan calcite shell (x100)



2.65 250μ



2.66 500μ

Judging by the nature of the siderite rims and the presence of zones of replacement, phosphatisation appears to have affected both aragonite and calcite shells after the formation of siderite rims. Phosphatised shells also show internal replacement by siderite rhombs. The tendency for phosphatised shells to occur in small groups indicates that replacement has occurred in place.

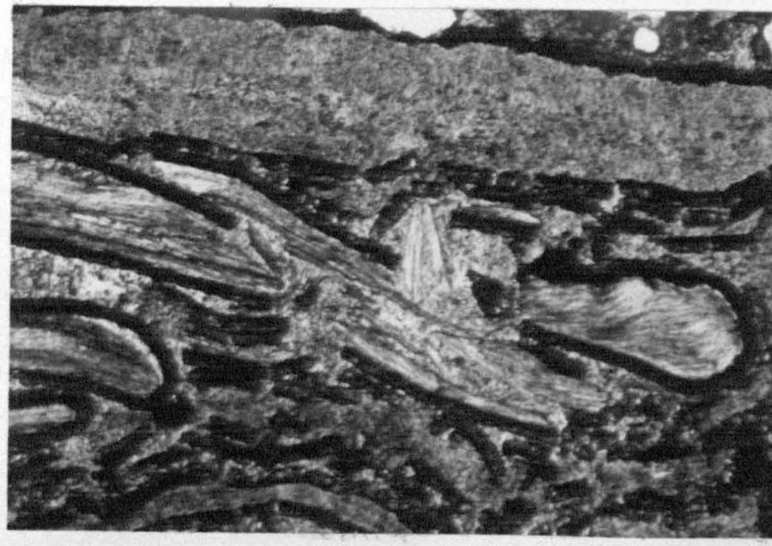
c) Solution. In the upper beds, where the calcite cement is absent, all but the phosphatised shells have been dissolved, and are represented only by siderite rims and, where present, inner replacement zones. No subsequent sparry filling has occurred except possibly the development of quartz overgrowths which form elongated prisms free from inclusions except where they pass through siderite rims. Since similar overgrowths are developed into primary voids subsequent to siderite rim formation, it is probable that the quartz is not replacive in these beds. In a thin zone between the limestone and the upper leached beds solution of aragonite shells only has taken place, and the voids so formed have been later filled with sparry ferroan calcite cement, together with the primary voids. The criteria by which replacive and drusy ferroan calcite may be distinguished are discussed in the section on cementation.

d) Deformation A feature confined to ferroan calcite shells is penetration by quartz grains as a result of pressure solution. The most distinctive deformation however, is shell fragmentation. This is developed most strongly in the very shelly limestone where as was pointed out with relation to fracture of quartz grains, the inhomogeneity of the rock has led to considerable strain on the component grains during compaction. The breakage may involve simple fractures but often leads to complete shattering of the shell (fig.2.67). In the limestone the fracturing of originally aragonitic shells can only be seen by displacement of the shell margin and by slight colour differences in stained and unstained sections: otherwise the cementing calcite tends to blend into the replacive calcite. Breakage also affects the siderite rims which may remain attached to the shell surface or may separate from it. Quartz overgrowths may be affected by shell breakage, as may be seen from their extinction which is no longer in continuity with that of the host crystal; in rare cases a fracture in the shell penetrates a quartz overgrowth for a short distance causing slight dislocation and then dies out. In the limestone fractures are filled with ferroan calcite cement, but in the upper beds siderite rims are continuous around diagenetic fracture surfaces, which are, however, rare.

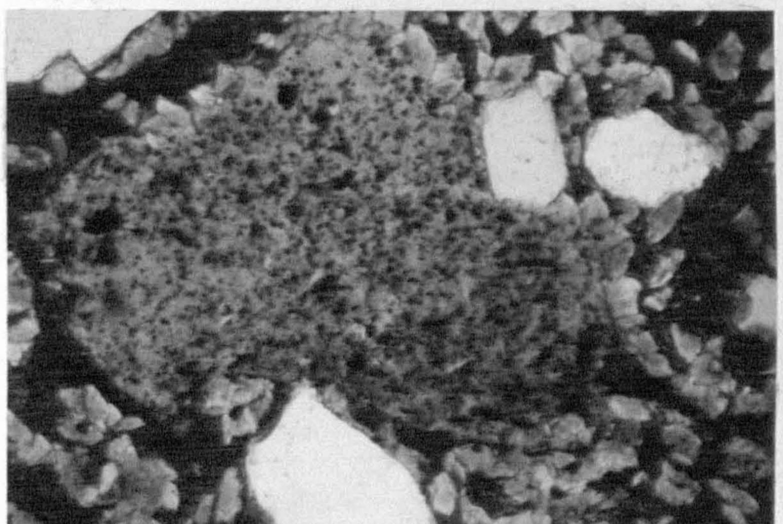
3. Ooliths

Ooliths are a minor constituent of both the limestone and the sandstone, comprising four percent of the rock at a maximum. They vary in shape from being slightly to moderately elongate (fig.2.73). In the limestone they show partial or complete replacement by ferroan calcite which may occur as a single crystal or as a mosaic. Unreplaced chamosite in recrystallised form may occur as thin bands within the calcite. In the non-calcareous beds the ooliths are unaffected by calcite, and consist of brownish chamosite with a rather diffuse concentric structure which may be accentuated by dusty opal; tangential orientation of the chamosite is seen under crossed nicols.

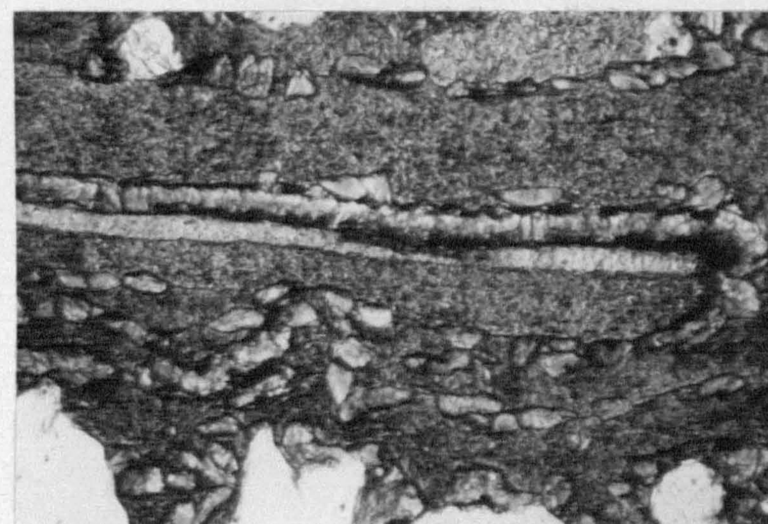
FIG. 2.67	Fracture of shells and siderite rims	(x50)
FIG. 2.68	Chamosite mudstone fragment with opaque inclusions	(x100)
FIG. 2.69	Shell consisting of (i) non-ferroan calcite (pink) with continuous rim of radial prismatic siderite crystals and (ii) ferroan clacite (blue) with impersistent rim of siderite rhombs	(x100)



2.67 500μ



2.68 250μ



2.69 250μ

In the intermediate zone of aragonite solution and calcite precipitation the chamosite ooliths are similarly unaffected by calcite replacement.

4. Intraclasts.

In the limestone intraclasts consist of phosphatic pebbles, irregular masses of chamosite mud and rare siderite spherulites. The phosphatic pebbles consist of silty mudstones replaced by collophane. The chamosite intraclasts consist of greenish isotropic mud in which abundant opaque bodies are present (fig.2.68); each of these particles is enclosed by a small anhedral or subhedral siderite crystals is a reflection of the shape of the opaque inclusions, which are in all respects similar to the 'cores' described from the matrix of the ironstone unit. The distribution of these particles is haphazard and they are quite variable in shape; the average size varies between intraclasts and sometimes within a single intraclast. The chamosite intraclasts show less rounding than the remainder. Intraclasts of unknown original composition which have been replaced by ferroan calcite are also present. In the non-calcareous beds intraclasts comprise large siderite mudstone fragments and phosphatised fragments in which the phosphate has the same acicular habit as in the associated shells.

5. Siderite Cement

Throughout the deposit all grains have been the site of siderite drusy rim formation. The rims are about 40 microns thick, and in the most shelly limestones where the original void space must have been very high the siderite constitutes nearly forty percent of the rock.

This percentage generally decreases upwards owing to decreasing shell content and to increasingly good packing of the grains; percentages still remain high in the non-calcareous beds since the thickness of siderite rims is greater (up to 60 microns) where the voids are sufficiently large and the siderite tends to fill practically all of the original voids. The siderite rims are of two types. In one type the siderite comprises a dark inner zone, probably replacive, followed by a thicker zone of radially oriented siderite crystals. These crystals are thin prisms about 30 microns long and up to 5 microns wide: they are brown at their proximal end and become clearer outwards. Such rims are confined to surfaces of non-ferroan calcite shells and when the calcite is composed of well defined crystals the siderite rim is in optical continuity with the adjacent calcite crystal; in the case of an echinoderm fragment the whole rim extinguishes at once and the siderite crystals show preferred rather than radial growth. In the second type, wheih affects all grains other than non-ferroan calcite grains, the rim consists of a series of rhombic crystals up to 20 microns across, which are sometimes cored, and which show little preferred orientation. In some parts these rims are continuous, with tightly interlocking crystals, and superficially resemble those of the first type; at other times the rhombs are very sparsely distributed. In a shell which comprises both types of calcite, the two types of rim are developed and are sharply delineated (fig. 2.69).

The contrast between the development of rims on non-ferroan and originally aragonitic shells implies one of two mechanisms of formation: (i) rim formation precedes replacement of aragonite by ferroan calcite and seeding of siderite crystals occurs more readily on the calcite, having a rhombohedral crystal structure, than on the orthorhombic aragonite;

(ii) rim formation follows replacement of aragonite and the seeding occurs more readily on the non-ferroan calcite owing to its greater solubility compared with the ferroan calcite.

The first mechanism is favoured by the development of rims which are in optical continuity with the adjacent non-ferroan calcite crystals, a feature which is never on ferroan calcite shells despite the fact that they comprise large, well-formed crystals. This suggests that the ferroan calcite had not been introduced at the time of rim formation; it is much more likely that the replacement of aragonite occurred at the same time as the ferroan calcite cementation which followed siderite rim formation.

In the basal lag deposit the siderite rims show severe breakage under compaction, while in the overlying beds, including the upper part of the limestone, siderite rims are continous around diagenetic fracture surfaces.

6. Calcite Cement

The limestone is characterised by the presence of ferroan calcite cement which fills in all depositional voids remaining after siderite rim formation and also the voids produced by fracture. Crystal size is usually in the order of one to three millimeters but in the thin zone of aragonite solution the cementing crystals reach twenty millimeters in length. The crystals are highly poikilitic and included ferroan calcite shells are in optical continuity with the cement. Replacive calcite may be distinguished from cementing calcite by its brownish colour and by the purplish blue stain compared with the clear blue of the cement; preservation of structure may, of course, be an additional feature. It is on this basis that the aragonite shells of the intermediate zone are thought to have been dissolved and later cemented; the very large crystal size of the cement may be a reflection of the lack of replaceable grains (removed by solution) which has reduced the number of seeding points. In all cases the seeding appears to have been extremely sparse; there is no development of a finely crystalline outer rim of calcite and each discrete cavity is filled by one or perhaps a few relatively large crystals. The calcite matrix forms between 12 and 20 percent of the rock, averaging about 15 percent. Replacement by ferroan calcite is presumably associated with cementation and the two processes probably mark the final diagenetic changes within the rock.

7. Diagenetic History. In order to outline the diagenetic history of this deposit it is necessary to consider it as comprising four zones. The first of these constitutes the basal lag deposit, about one foot in thickness, in which the development of quartz overgrowths, the penetration of aragonite shells by quartz, and the formation of the siderite rim all precede the main phase of compaction. The first two of these must have taken place under mild pressure, but the intense fracturing of shells and quartz grains which followed siderite rim formation represents the main phase of cemetation, by ferroan calcite; this is apparently accompanied by replacement of ooliths, some intraclasts and occasionally non-ferroan calcite shells. Dolomite has

probably formed at a later stage, but there is no other evidence for post-cementation changes. The overlying limestone shows a similar series of events but breakage, which is only on a small scale, precedes siderite rim cementation. In the beds overlying the ironstone the siderite cement is similarly later than the shell breakage and is succeeded by shell solution. Some shells and intraclasts replaced by phosphate and therefore not dissolved; the presence of the two types of siderite rim indicates that phosphatisation occurred after siderite cementation. Apart from solution little change took place after siderite cementation with the possible exception of quartz overgrowths into original and shell solution cavities. Ooliths are unaltered. In the basal few inches of these beds solution has affected only the aragonite shells and subsequently all voids have been filled by large ferroan calcite crystals. No replacement by ferroan calcite seems to have taken place.

The marked difference in the relative time of fracturing and siderite cementation between the basal bed and the remainder is attributed to the difference in rate of accumulation. The slow deposition of the basal lag deposit allowed the early chemical changes to take place without any substantial overburden; the subsequent more rapid deposition would have caused compaction in the basal bed and then in the successively higher beds which had not yet undergone the early siderite cementation. The development of quartz overgrowths replacing shell material are confined to the basal zone and seem to prefer conditions of prolonged gentle compaction within unstable structures. With more rapid deposition the main phase of compaction occurs at an early stage and the local strains in the structure are relieved by breakage and reorientation of grains, so that by the time that the chemical conditions suitable for quartz overgrowths have set in the structure is relatively free of local stress points, especially after siderite cementation had strengthened the compacted framework. The development of replacive overgrowths in the calcareous facies of the ironstone unit would similarly appear to have occurred under conditions of slow compaction under little overburden.

The combination of shell solution in the upper part and calcite cementation in the lower suggests that the calcium carbonate in the cement might be derived from solution of shells. To check this the average percentage of ferroan calcite in the cement and in replaced ooliths and intraclasts of the limestone, multiplied by a thickness factor of 2, was compared with the average percentage of shell voids in the upper beds multiplied by a thickness factor of 4. An extremely close correspondence was obtained which though probably to some extent spurious indicates that the solution of shells in the upper part could have provided the calcite for cementation of the basal beds.

The probable diagenetic sequence in the four zones is shown in figure 2.70.

E. THE GREAT FRYUPDALE SANDSTONE

This sandstone body occupies a horizon similar to that of the Winter Gill sandstone, but the sediments are more variable than at that locality. In Yew Grain and in part of the southern section of The Scar the sandstone is shelly, pebbly and non-calcareous with a

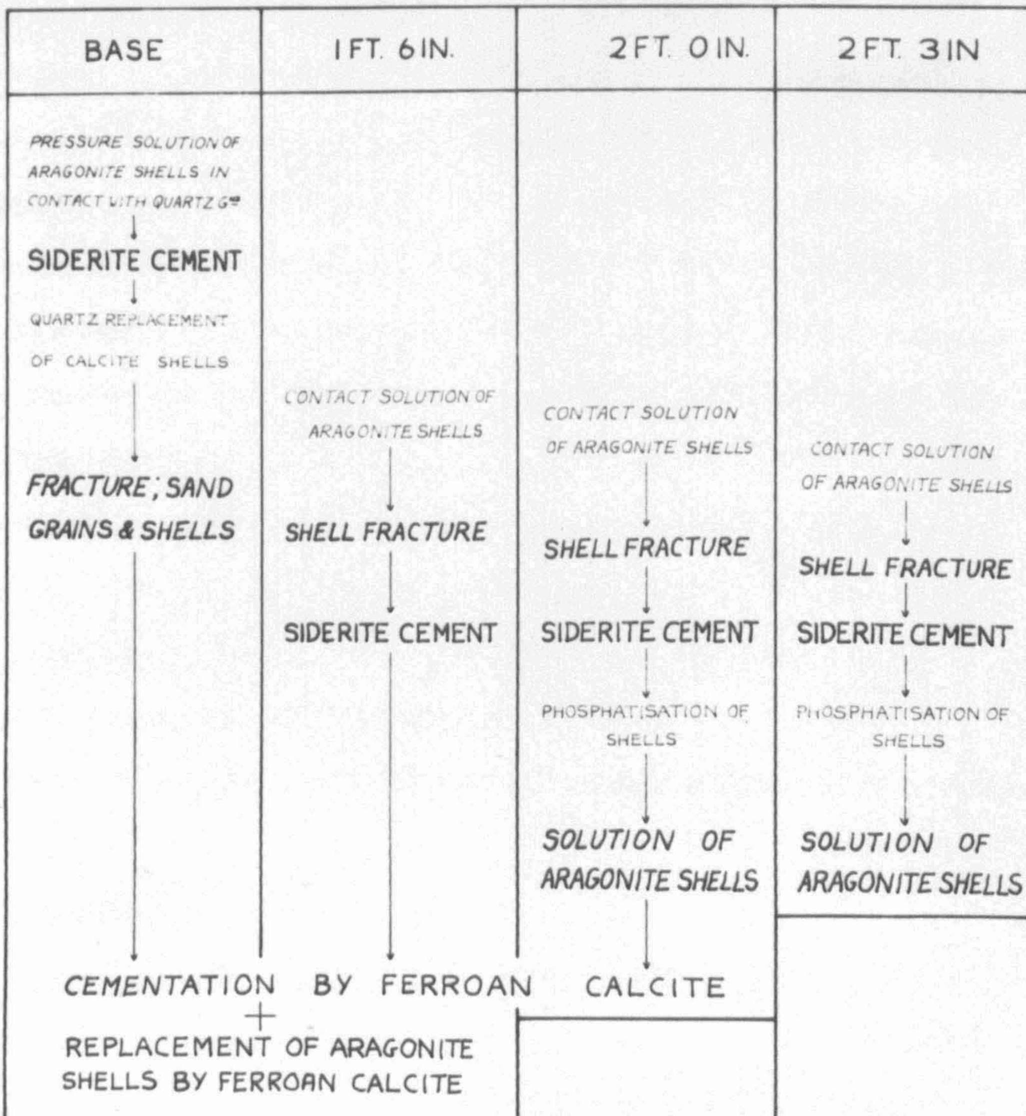
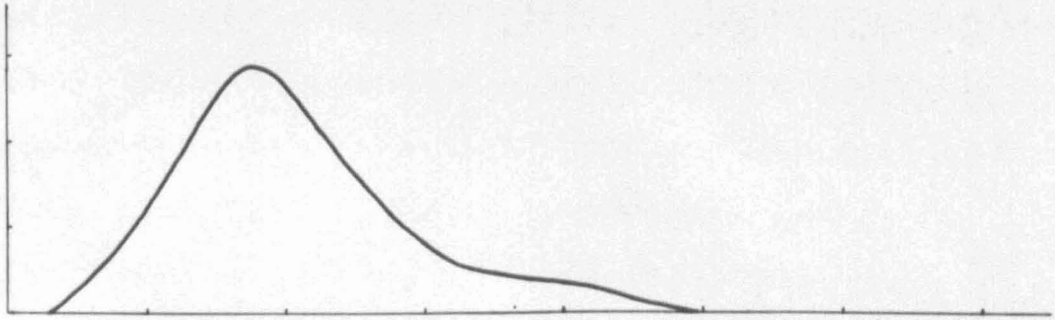


FIG. 2.70 DIAGENETIC SEQUENCE AT VARIOUS LEVELS IN THE WINTER GILL SANDSTONE

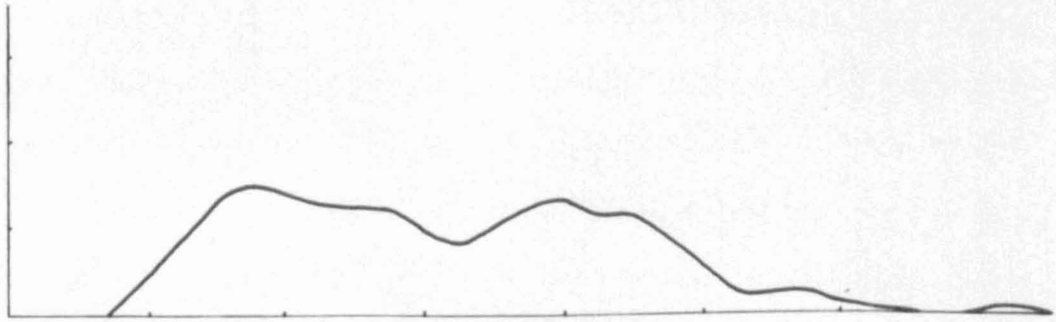
clay matrix. Along The Scar these beds give way to a calcareous sandstone which comprises terrigenous grains and rare shells in a ferroan calcite cement. Further to the north ferruginous, pebbly, cross-bedded calcareous sandstones occupy much of the lowest part of the section. In thin section this sandstone is seen to comprise shells preserved in ferroan and non-ferroan calcite, muddy intraclasts and abundant quartz grains, with early siderite rim cementation and later cementation by ferroan calcite. The siderite rims show contrasting development on ferroan and non-ferroan calcite shells; and the non-ferroan shells show no quartz overgrowth replacement but occasionally show partial replacement by ferroan calcite, especially in crinoid fragments. Although the siderite rims are rather thin (20 microns) the rock is petrographically very similar to the upper sandy limestone of Winter Gill. The grain size distribution of terrigenous grains corresponds closely to that of the Winter Gill sandstone and limestone, exclusive of the basal lag deposit (fig.2.71).

E. THE EGTON GRANGE SANDSTONE

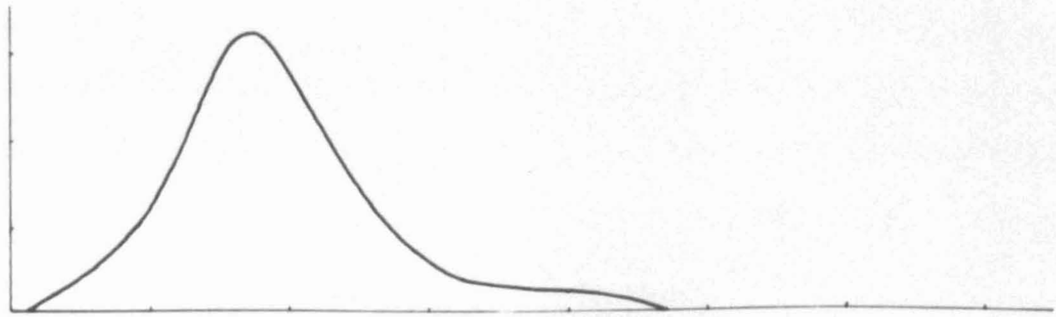
The lower part of the sandstone of the Egton Grange section is believed to be stratigraphically equivalent to the Winter Gill sandstone body. Petrographically the rock differs considerably from the Winter Gill samples in having abundant ooliths which are distorted and act as a matrix in part, no calcite or siderite. The matrix comprises a thin rim cementation of very finely acicular crystals which are oriented perpendicularly to the grain surfaces, and which are colourless to greenish and show very low birefringence. This mineral has



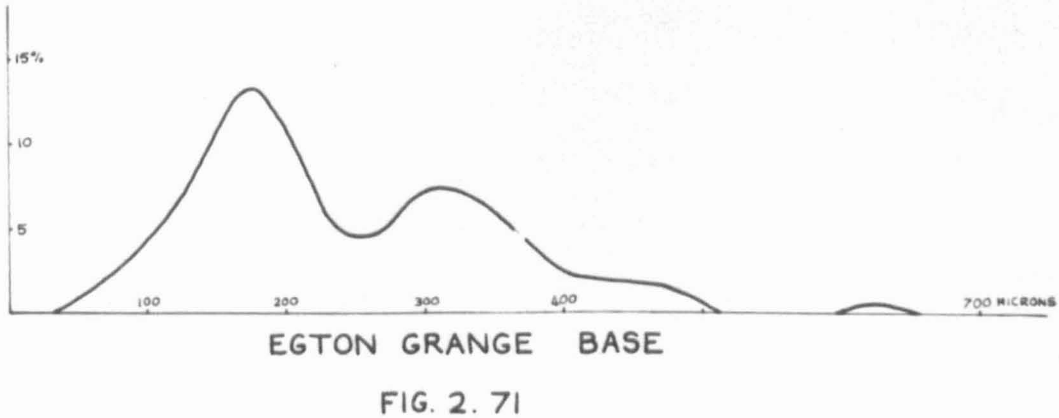
WINTER GILL AVERAGE EXCLUDING BASE



WINTER GILL BASE



GREAT FRYUPDALE AVERAGE



EGTON GRANGE BASE

FIG. 2. 71

not been identified. A later drusy filling by kaolinite has occurred in parts. Shells are represented by cavities into which protrude abundant quartz overgrowths in which the original shell structure is preserved. Occasional prismatic overgrowths occur which do not contain inclusions and which appear to be later drusy growths. In hand specimen the early replacive crystals are brownish and squat, while the drusy crystals are clear and elongate. Intraclasts of shale and chamosite mud are present together with abundant siderite spherulites. The grain size distribution is similar to that of the Winter Gill sandstone, even to the presence of a bimodal lag deposit at the base (fig.2.71), and it is probable therefore that this sandstone is the lateral equivalent of the Winter Gill beds. The diagenetic sequence appears to have been as follows: (i) replacive quartz overgrowths; (ii) compaction and spastolithisation; (iii) rim cementation; (iv) shell solution; (v) drusy quartz overgrowths; and (vi) kaolinite cavity filling.

G. THE INGLEBY IRONSTONE

Although the whole unit includes nearly six feet of siltstone in places the sections are so weathered that the ironstone seam is the only member which is suitable for thin sectioning. Although the matrix is mostly limonitised the constituent grains are readily distinguishable. Terrigenous grains are not abundant, reaching a maximum of 10 percent in Hartoft Beck, the most easterly exposure, while elsewhere they make up less than 5 percent. The modal grain size, however, is only about 75 microns in the east and increases to 150 microns at Danbydale and 175 microns at Rud Scar. Ooliths are most abundant in the west, constituting between 10 and 30 percent of the ironstone at Rud Scar, and less than 5 percent in Hartoft Beck. They are fairly large, averaging 400 microns in length and show little variation in size and shape between different localities (fig.2.72) which is a useful feature in confirming the correlation of the isolated outcrops. The concentric structure of the ooliths is always retained, and the envelope consists of chamosite with occasional opal; thin siderite rinds are developed. Shells are not abundant; in the west they comprise occasional small shells while in Hartoft Beck abundant very small shells, probably ostracods, are present. The shells may be replaced by siderite but more often occur as solution voids or kaolinite casts. The siderite matrix is variable. In Hartoft Beck it resembles the dark cryptocrystalline matrix of the ironstone unit (p. 113) and shows small scale irregular septarian cracking. Elsewhere the matrix is more coarsely crystalline with well defined cores.

This ironstone is distinguished from the basal ironstone in having a lower terrigenous grain content and in being more oolitic. The ooliths are quite large and elongate and the distribution of ooliths and terrigenous grains differs from that of the basal ironstone.

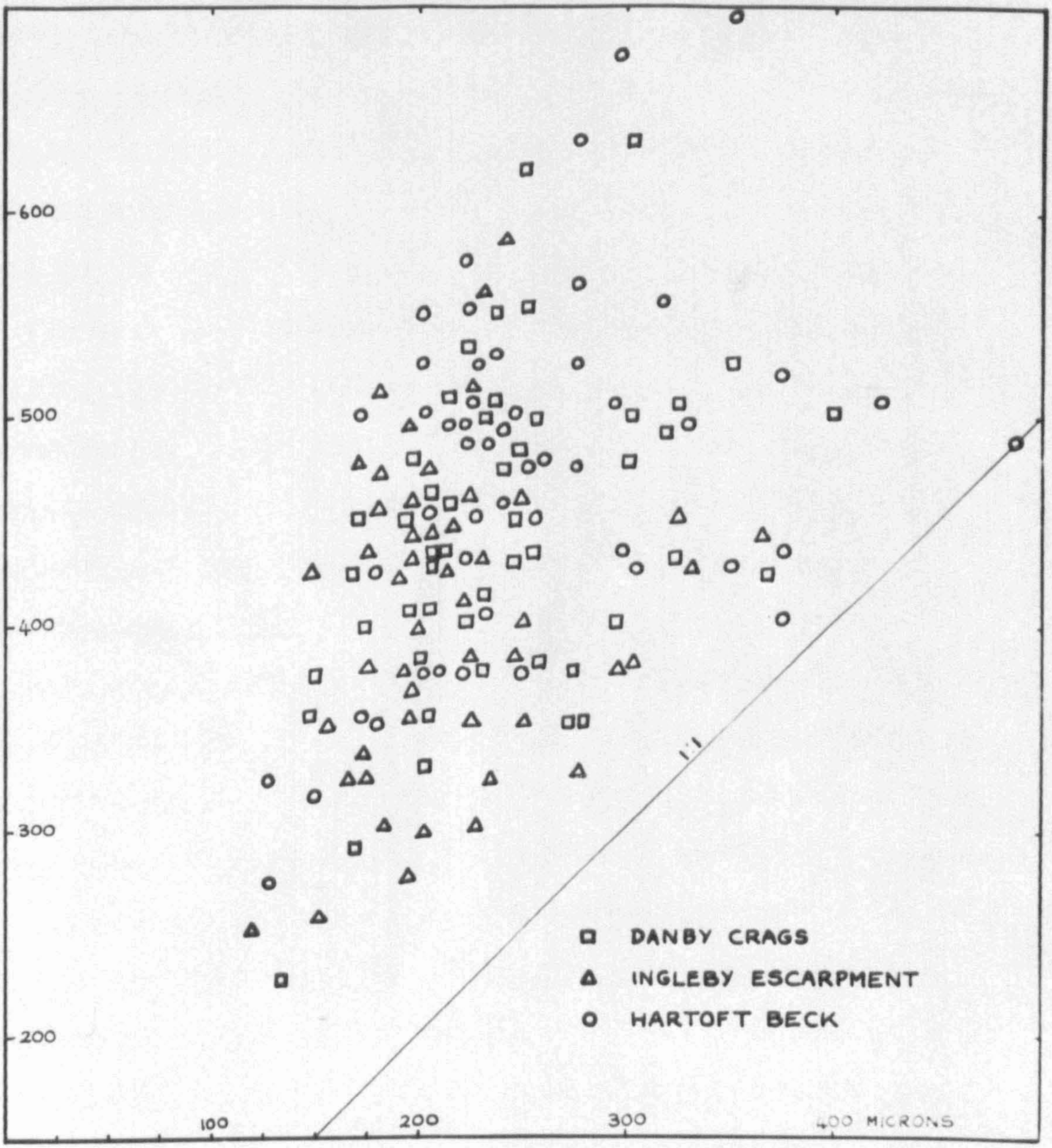


FIG. 2.72 APPARENT MAXIMUM AND MINIMUM OOLITH DIAMETERS FOR INGLEBY SEAM

III DEPOSITIONAL ENVIRONMENT

Over much of the area the ironstone unit was succeeded by deposition of clay. In the extreme east a certain amount of shallow erosion took place indicating that some interval occurred between the deposition of the two units. In the south-west coarser sediments accumulated which passed gradationally into the shales to the north and east. All deposition was probably in shallow water but while in the north and east the first sediments were probably deposited in relatively quiet water in the south-west currents must have been active bringing in silt and sand from the west. The silt deposition was then followed by the accumulation of fine muds, probably chamositic, which appear to indicate a return to quiet conditions, since they are overlain by fine-grained shale.

Locally in central Eskdale accumulation of thick deposits of well-washed sand occurred apparently after the deposition of silts to the south-west but for the most part before the deposition of oolitic muds. This suggests rapid accumulation which must have occurred either in a channel or on a bank. Although the erosional base may favour the former mechanism the presence of opposing north-south current-bedding directions combined with east-west elongation makes a bank seem more plausible. A channel would have to have developed at a late stage in the shale deposition, a possibility which is precluded by the correlations with the oolitic ironstone to the south-west, which is overlain by a thick shale. The sandstones appear to represent accumulation on a bank or banks immediately after deposition, with an early phase of scour and lag accumulation. The sand and probably much of the shell material is introduced, but some of the lag constituents such as spherulites and ooliths may be locally derived. The ooliths from the basal lag deposit of Winter Gill and also of the overlying beds are compared with those of the basal ironstone unit and of the Ingleby ironstone in figure 2.73; their elongation indicates derivation from the basal ironstone unit with the large ooliths from the lag deposit representing selective retention of the largest ooliths. Siderite spherulites could easily be obtained by the erosion of the top non-marine beds which in Winter Gill at least are highly spherulitic. The only probable source of sand is from the south-west or west since normal successions are developed in all other directions. The sand may have been sorted from the silts during agitated conditions and banked up [illegible] on the original northern limit of silt deposition or may have been introduced independently by long-shore currents. While the Fryupdale sandstone deposition ceased during shale deposition the Winter Gill bank may have continued till the end, since no normal shale is developed above; there is also a strong possibility that the sandstone bank has affected deposition of the overlying sandstone unit which is normally developed in Great Fryupdale.

Apart from these local developments the deposition within the shale unit was fairly uniform. The whole sequence is of a laminated nature with thin silty laminae in shale in the lower part and sandy laminae in silty shale in the upper part. This implies a certain degree of sorting throughout shale deposition and the shale was probably deposited as a compact mud, especially in the upper sandy

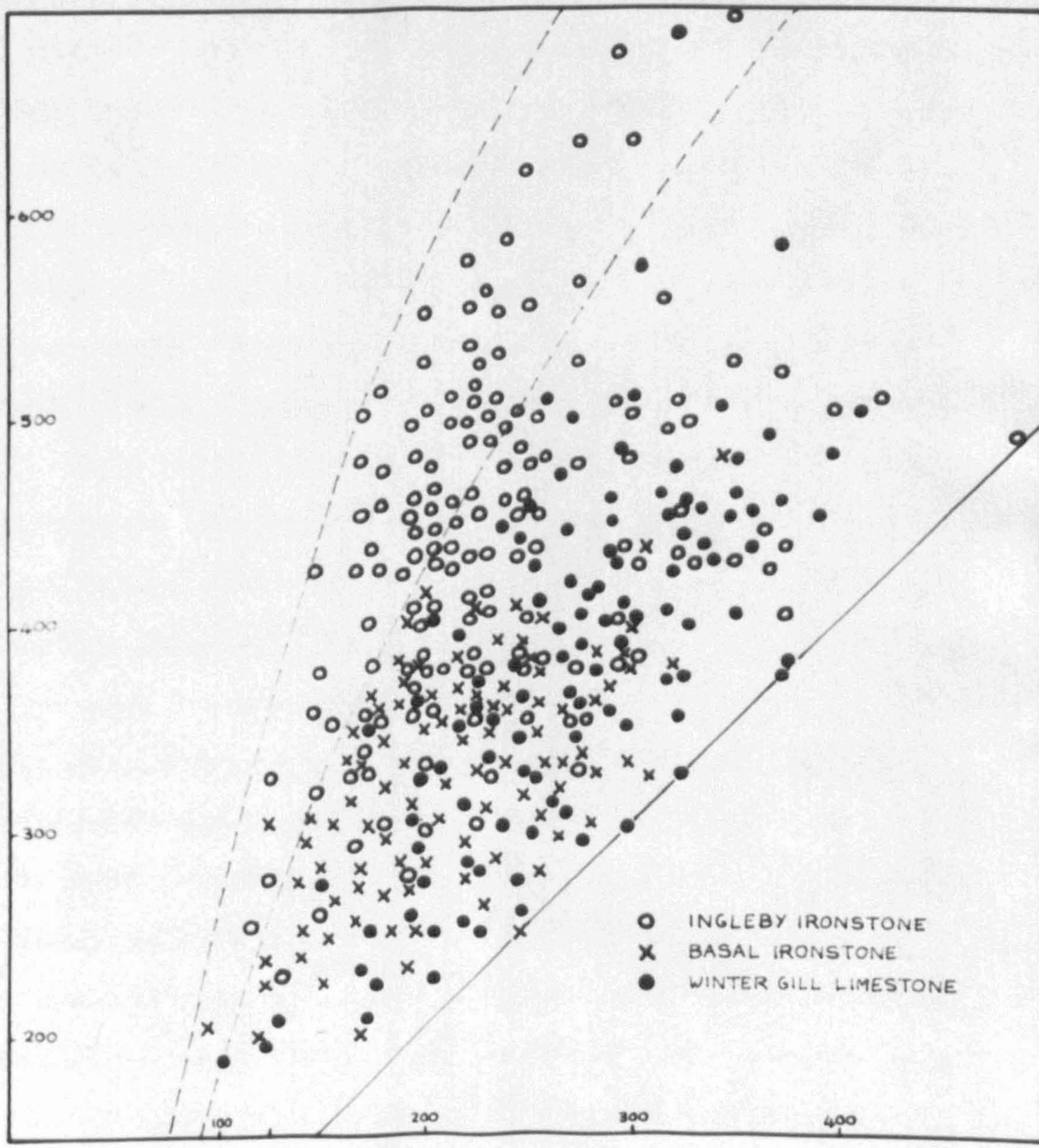


FIG. 2.73 COMPARISON OF OOLITH SHAPE: INGLEBY IRONSTONE, BASAL IRONSTONE, W^{r} GILL LIMESTONE

part. The influence of terrigenous grains on the depositional compaction is reflected in siderite mudstone nodules which in the upper part show practically no shale compaction around them, while in the lower part relative compaction reaches about fifty percent. The source of the terrigenous material in the shale appears to be to the east since in western, south-western, and northern localities the shale is fine-grained throughout and the contact with the sandstone unit relatively sharp. The only exceptional shale development is that of Hawker where in the region of the erosion at the base of the shale unit the shale itself shows unusual features, being mostly silty and poorly laminated with abundant comminuted plant material. A persistent off-shore current appears to have prevailed throughout the shale unit, probably from the east. Relatively slow deposition with winnowing of sand into small lenses apparently continued for much of the time followed by rapid deposition of poorly sorted silts and plant debris as is indicated by the heavy loading at the base of the overlying sandstone.

The chemical environment of deposition is difficult to evaluate in these rather variable beds. The Ingleby ironstone and underlying shelly siltstones appear to have accumulated under conditions similar to those of the basal ironstone unit. The sandbanks of central Eskdale must have accumulated under oxidising conditions but under conditions too vigorous to permit the development of a fauna; the shell debris, which includes crinoid remains, suggests fully marine conditions, but may be derived from outside the limits of the present outcrop, In the shale the presence of a benthonic fauna throughout, though very sparse in the upper part, indicates oxygenated conditions and forms such as Trigonia, which occurs near the base at Iron Scar and near the top at Eller Beck, indicate fully marine conditions. Fossils are common only in the lower part of the shale in the south-west and elsewhere are scarce or absent, especially in the upper part. This scarcity of benthonic forms may be due to too great a water depth or to too rapid depositon, with continually changing muddy bottom conditions. Even traces of buttowing forms are rare, and probably rapid deposition caused the fauna to be so restricted. Local shelly horizons such as at Eller Beck must represent local pauses in deposition which allowed a flourishing fauna to develop for a short time. The total absence of any fauna in the fine-grained shale of south-western localities is probably a reflection of deeper water and deposition of very fine mud unsuitable for benthonic colonisation.

6. DIAGENESIS

Diagenetic fabrics of the various members of the shale unit have already been described, and the implications as to the chemical environments involved are discussed in Chapter IX.

CHAPTER VI THE SANDSTONE UNIT

I STRATIGRAPHY

A. FACIES DISTRIBUTION

The sandstone unit is generally divisible, in the field, into two parts:

(i) Upper division: medium-grained sandstone; bedding units between six inches and several feet thick; horizontal lamination or sometimes cross-bedded; fossils rare or absent; penetrated at top by rootlets.

(ii) Lower Division: fine-grained sandstone; bedding units usually less than six inches thick; intercalations of silty shale, especially in lower part; horizontal lamination or ripple-marked; fossils locally abundant.

Between Iron Scar (1) and Ravenscar (2) the lower division is very fine grained, comprising lenses of laminated sandstone separated by more persistent silty shale partings; the upper part is still rather fine-grained and shows occasional ripple-marked bedding planes, but no silty intercalations. Impersistent beds of pale gray silty siderite mudstone are developed, especially in the lower part. Shelly bedding-planes occur in the lower division, and rare U-tubes in the upper division.

At Iron Scar the normal thin-bedded sequence is interrupted by a small channel, the top of which is just below the base of the upper division and which is about four feet deep. The sides of the channel are steep and the width probably not great; the present outcrop (fig.2.74) appears to represent the bifurcation of a single channel in an easterly direction. The channel fill consists of massive fine-grained sandstone with shale flakes and pebbles of ironstone similar to those in the eroded beds. This channel appears to be an isolated feature; no others have been observed although exposures are good in adjacent areas.

In the cliff east of Hawsker (3) the sandstone has an erosional base (fig. 2.51) and fills a series of small channels which run north-east to south-west. These channels are accentuated by loading into the shale. At this point the sandstone is divisible into two units, an upper rather massive sandstone and a lower division of variable thickness which is thin-bedded except in channels. When traced northwards the upper division thins and dies out while the lower unit thickens. The upper sandstone is heavily loaded into a thin silty shale which separates the two divisions. In the most northerly accessible section the channelling and loading is less pronounced and finally dies out. The sandstone is overlain throughout by a siltstone and a thin sandstone penetrated by rootlets; both of these units are persistent and may represent the top of the sandstone unit since the main sandstone below is unusually thin and lacks the usual penetration by rootlets.

In East Cliff (5) the sandstone is more normally developed, with a clear two-fold division; the West Cliff (6), however, shows a rapid change to a thick series of thin-bedded sandstone, while the top of the unit is not seen. In Newholme Beck (7) both divisions are again visible; trough cross-bedding is developed at the base of

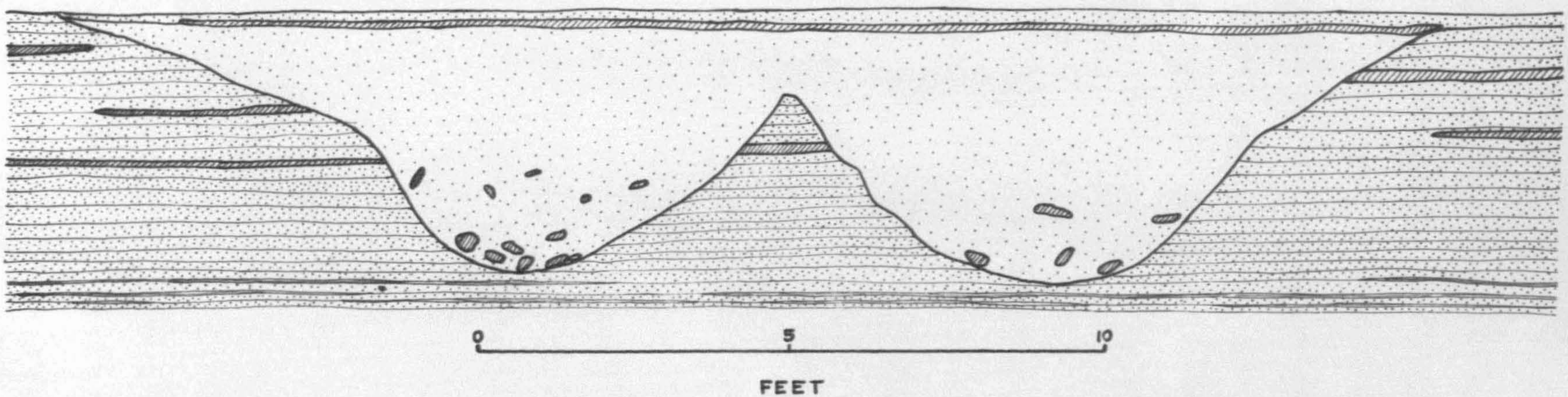


FIG. 2.74 CHANNELLING INTO THIN-BEDDED SANDSTONE, IRON SCAR

the upper unit which is rather massive and has probably cut down into the thin-bedded sandstone. At Hinderwell (8) the bedding is more regular. In Kilton Beck (9) the two-fold division is missing; the sandstone is flat-bedded but lamination is poor and silty shale intercalations absent. The base rests sharply on the shale unit and flakes of shale are included in the basal part of the sandstone, indicating a phase of erosion. At Loftus rootlets occur within the sandstone at several levels.

South of the Esk at Littlebeck (10) the sandstone is thin, flat bedded and extremely thinly bedded at the base, with no silty laminae. In Eller Beck (11), New Wath Scar (12) and much of Wheeldale the sandstone is thicker and exhibits a clear two-fold division. Flat bedding is present throughout with occasional ripple-marked bedding planes. Fossils are rare, mostly comprising U-tubes, but shelly beds occur locally, especially in Wheeldale. Sideritic sandstone is developed at several levels in the upper sandstone of New Wath Scar and Wheeldale Gill; ooliths may be present, particularly in the west. In the western inlier of the Eller Beck Bed in Wheeldale Gill (13) the sandstone unit is represented by a channel sandstone which removes the whole of the normal sandstone succession and much of the underlying shale. The channel fill comprises unfossiliferous and non-sideritic sandstone, with occasional cross-bedding. To the west the base rises slowly, revealing flat-bedded shelly sandstone of the lower division. In the Egton Grange (15) sandstone section the lower part is believed to belong to the shale unit and the upper part to belong to the sandstone unit and comprises non-ferruginous sandstone.

the top of which is apparently not seen. In Winter Gill (16) the development of the sandstone unit bears no resemblance to that of surrounding sections. As suggested earlier (p. 156) the grey sandy siltstone overlying the limestone-sandstone body may belong to the sandstone unit and be equivalent to part of the thin-bedded division. The overlying white fine-grained sandstone probably belongs to this division as well since the overlying beds are strongly sideritic and highly oolitic and would therefore appear to be equivalent to the sideritic and oolitic upper division of the sections to the east. The oolitic ironstone is overlain by a thin series of variable beds and finally a sandstone penetrated by rootlets.

The correlation of the oolitic ironstone with the sideritic sandstone to the east is supported by the presence of strongly sideritic sandstone in a section in the left bank of Winter Gill about a quarter of a mile from the outcrop, and by the abundance of fragments of very sandy laminated ironstone in the spoil heap of the Winter Gill mine. This sandy material is probably derived from the region of the air shaft where the ironstone is reported to have been thirteen feet thick and variable in character (Fox-Strangways, 1892; p.198). This correlation invlves a rapid lateral variation in thickness and in lithology and the probable reason for this is discussed in the next chapter. Figure 2.7.5 shows the proposed correlation of the Eller Beck Bed sections in this region.

At the head of Glaisdale (17) the thin-bedded sandstone is normally developed with a poorly exposed massive non-ferruginous sandstone above, and a similar section is seen in Birk Wath Beck (18).

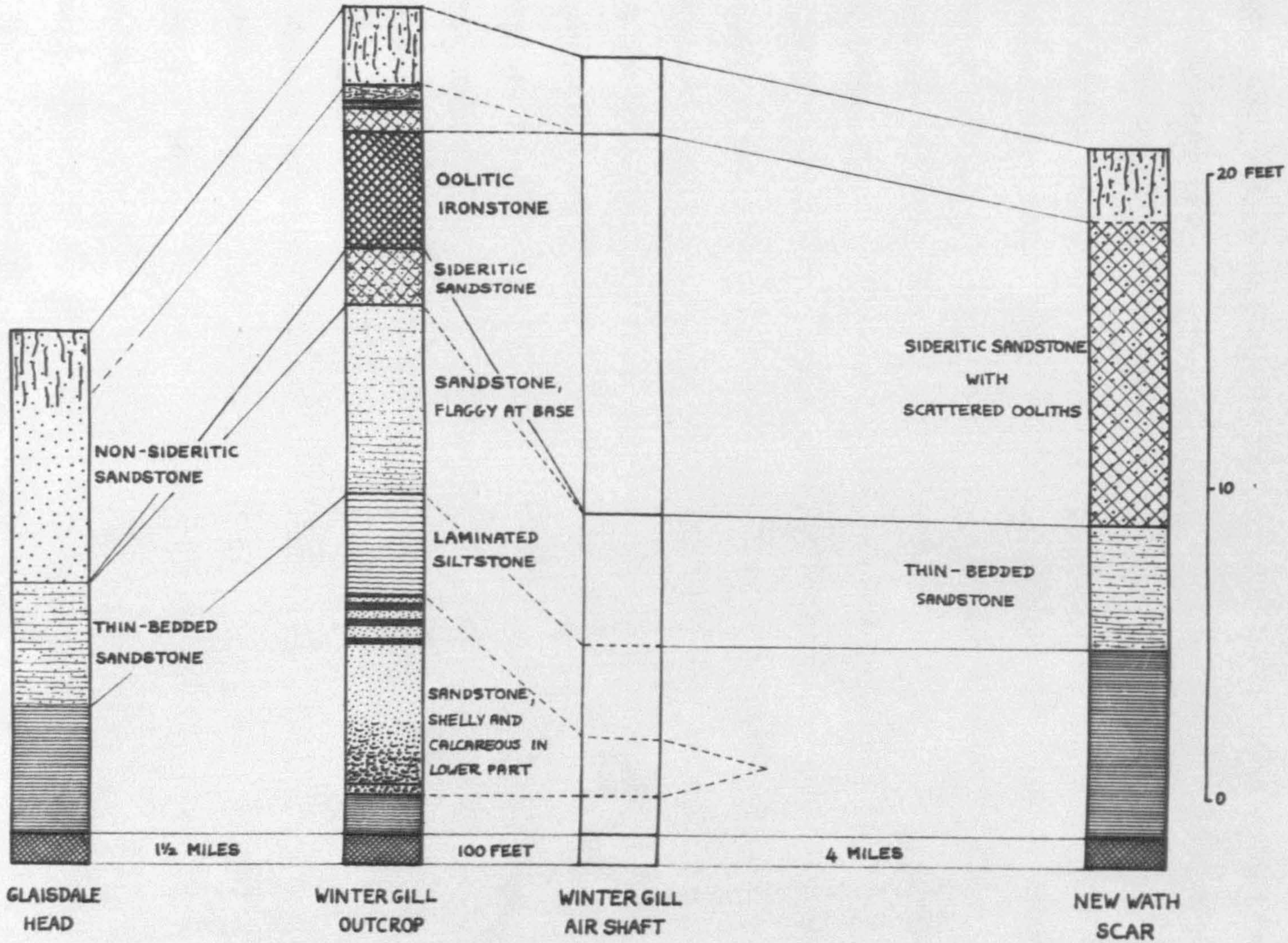


FIG. 2.75 CORRELATION BETWEEN WINTER GILL AND SECTIONS TO WEST AND EAST

To the north in sections close to the Esk the sandstone is variable in thickness. In Stonegate Gill (19) it is 3ft. 9in. thick and penetrated by rootlets at the top while in Busco Beck (20) the normal succession is entirely lacking and thin sandstones interbedded with siltstone and shale with a total thickness of five feet represent both divisions of the unit. Fossils are absent apart from occasional burrows in the sandstones which are locally sideritic and slightly oolitic. In Grunkley Gill the sandstone unit is more normally developed, though still rather thin, and is sideritic and slightly oolitic. The Busco Beck section may represent quiet water deposition in an eroded channel with some of the coarser material being derived from some adjacent more normal sequence. At the base of the unit is a thin tough white sandstone with low angle cross-laminae along which heavy minerals are concentrated; it is similar to the sandstone at the base of the channel in the shale unit at Hawsker and probably represents a lag deposit derived from erosion of the shale. Thin rootlet beds in the upper part of the sequence indicate fresh-water conditions for at least a part of the time.

Erosion of the sandstone unit and subsequent filling by fine-grained sediment is seen at one other locality on the southern end of New Wath Scar. The upper sandstone division has a rolling top in this area, with a maximum thickness of twelve feet, but is locally reduced to 2ft. 6in. by erosion in the form of a wide trough which is filled with fine-grained laminated dark grey shale which contains occasional concretions but is unfossiliferous. The shale becomes sandier at the top and passes into silty sandstone.

To the south of Glaisdale the sandstone is seen in Hartoft Beck (37) where it is only about six feet thick with thin-bedded sandstone and shale partings in the lower part and more massive sandstone of variable thickness in the upper part. Ironstone bands occur in the thin-bedded sandstone and resemble those of Iron Scar. In Northdale (36) the unit has thickned again and the lower division is particularly well developed, being finely laminated at the base with shale intercalations. In this basal part very shelly bedding planes are present. The remainder of the lower division consists of shelly sandstone which is unlaminated due to intense burrowing. Occasional large rather sp herical calcareous concretions occur in this upper part. The overlying sandstone contains scattered ooliths and bands of sideritic sandstone with ooliths and shell fragments. Locally the ooliths are abundant and beds of sandy oolitic ironstone occur which may be associated with siderite mudstone. At one point a shallow trough has removed the top of the sandstone and is filled by siderite mudstone, shale and at the top a thin sandstone with shells at the base.

In Great Fryupdale (21,22) the lower division is similar in lithology and fauna to that of Northdale. In the unburrowed laminated sandstone calcareous shelly bands up to two inches thick are present which show small-scale erosion at the base. The upper sandstone is mostly flat-bedded but near the top shows low angle cross-bedding locally. Also within two feet from the top a thin band of red-weathering ferruginous sandstone is developed. The total thickness of the sandstone unit is 27 feet, but for some distance along The Scar the basal part of the thin-bedded sandstone cuts out against the underlying sandstone of the shale unit.

In Danbydale the sandstone is poorly exposed but in the west (25,26) very ferruginous red-weathering sandstone develops in the upper unit. Ooliths occur scattered throughout the sandstone and also as local concentrations in very thin beds. Shells and wood fragments are found at the base of small washouts within the sandstone. The lower division of the unit is thin-bedded with silty shale partings; shells are not common. At Commondale (27) the sandstone is thinly and irregularly developed throughout. The base is sharp and the lowest sandstone is dark and silty with abundant shells and nodular sandy ironstone bands; a more persistent sandy ironstone with groups of large shells, possibly in growth position, occurs at the top of these lower beds. The overlying beds are non-ferruginous and unfossiliferous.

In Stockdale Beck a series of greenish sandstones of rather uniform lithology are developed; they contain occasional shells and include a thin shelly siderite mudstone band. In Battersby Crags (28) the sandstone unit is thick and apparently rests on an erosion plane. Sideritic sandstone with occasional shells is developed on two horizons. In western Rosedale the sandstone is again not typically developed; in Reeking Gill (33) a channel is present with well-developed trough cross-bedding indicating a current from the north-east. In the Castle Crag section (32) only the base of the sandstone unit is seen and it is again apparently erosional though not of the same type as in the previous section; the sandstone is flat-bedded or ripple-marked with shale fragments, shells and burrows. In south-west Rosedale

(34, 35) the sandstone is very massive in the upper part and thinner bedded at the base though shale laminae are absent.

In the Ingleby escarpment (30) and Farndale (31) the entire unit is represented by a thin tough fine-grained sandstone without fossils. The sandstone in Blow Gill (38) is of similar lithology and thickness.

B. PALAEONTOLOGY

The fauna of the sandstone unit is very restricted. Although shells are locally abundant in the thin-bedded sandstones they consist almost entirely of Liostrea and Aviculopecten. In the upper sandstones the few identifiable shells belong to the same two groups. The shells are generally concentrated in thin bands and are disarticulated indicating that they have undergone a certain degree of transport and sorting, though the transport is probably not over any great distance. In Great Fryupdale the shelly sandstone bands have slightly erosional bases which clearly cut across the horizontal lamination of the preceding non-shelly beds. The shells are usually preserved as casts, but in Great Fryupdale they calcium carbonate has not undergone solution. In the bed of sideritic sandstone at Commondale a different fauna is developed consisting of Gervillia and Modiola, and the shells occur in clusters which suggests that they are in position of growth. Shells, generally fragmentary, are found practically to the top of the sandstone but are usually visible only in sideritic sandstone where the white clay of the shells contrast with the dark matrix. Shells are absent in the north and in the south-west.

In the lower sandstone division trace fossils are locally abundant.

In Eller Beck and at New Wath Scar horizontal U-tubes are present in silty partings while in Northdale and Great Fryupdale the division is characterised by two forms probably Arenicolites and Taenidium. In the upper division U-tubes are found at several levels in many of the Eskdale sections and in Rosedale; these consist of the vertical form Corophiodes and an inclined form of similar character. A characteristic feature of the sandstone is the presence of rootlets in the topmost beds; in a few sections they are absent due to erosion while in the extreme south-west their absence appear to be original. In Lofthouse Beck the sandstone contains rootlets at more than one horizon within the unit.

C. CORRELATION

In most of the area the sandstone unit consists of the two divisions already mentioned, and it is probable that each division was laid down at about the same time in these areas. Since the base is gradational, however, passage of thin-bedded sandstone into shale is quite possible, and this may account for the very high shale unit to sandstone unit thickness ratio in the south west. At several localities the two-fold division is absent and unless lateral passage has occurred the sandstone must represent both divisions. In Kilton Beck and at Battersby Crags, where the base of the sandstone is erosional, it is possible that the lower division has been removed before deposition of the upper division. The sideritic and oolitic sandstones developed in several areas appear to be local facies varieties of the normal sandstone and are not of a different age.

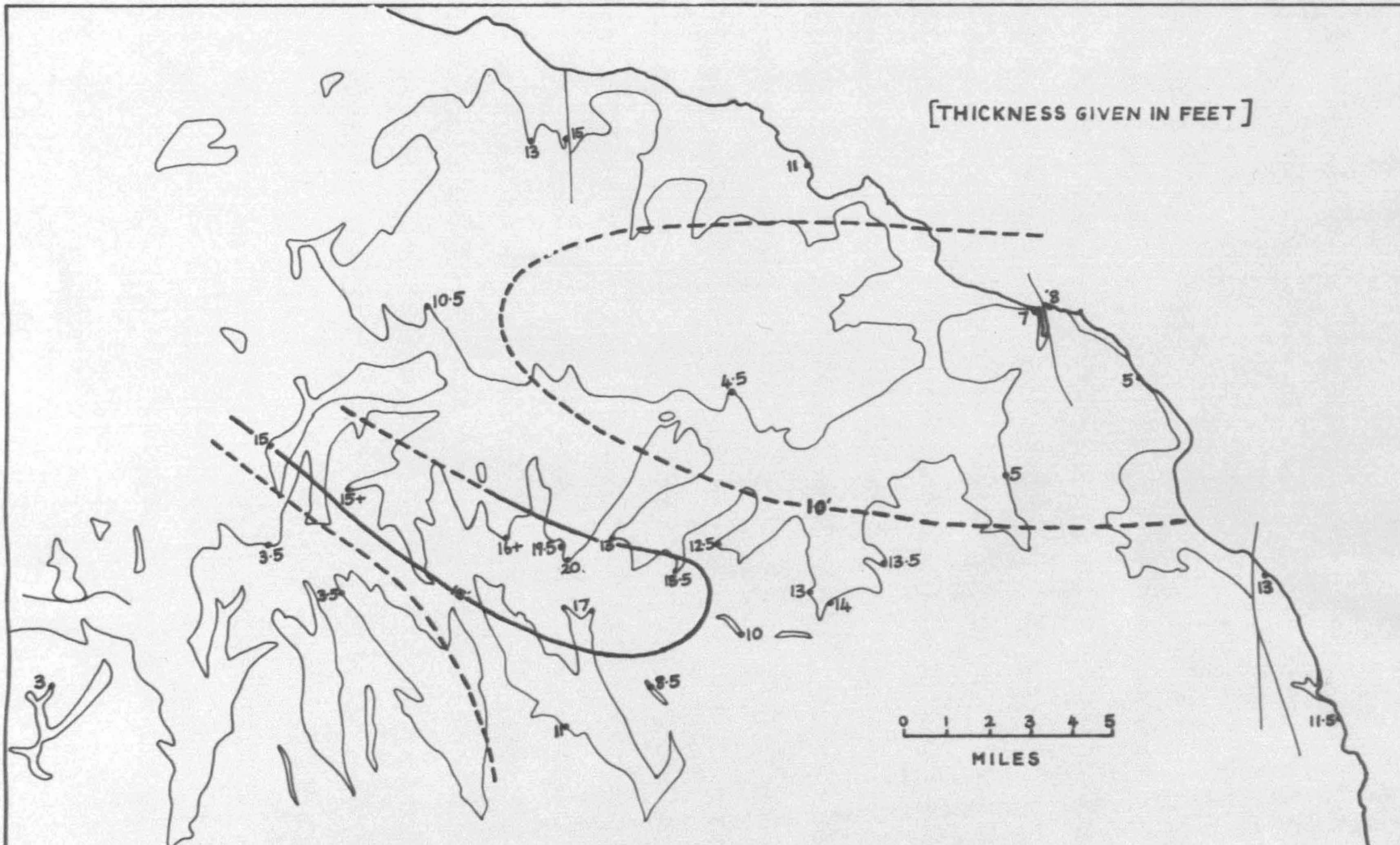


FIG. 2. 76 ISOPACHYTE MAP OF THE SANDSTONE UNIT

The total thickness of the sandstone unit is represented in an isopachyte map in figure 2.76.

D. SEDIMENTARY STRUCTURES

The sandstone unit is characteristically flat-bedded and horizontally laminated. Ripple-marks are developed on some bedding planes, chiefly in southern coastal districts but also in localities in eastern Eskdale and in Rosedale. These ripple-marks vary in cross-section from wavelength 3½ in. and amplitude ½ in. to wavelength 4½ in. and amplitude ¾ in. Although slight asymmetry may be detected in these ripple marks no significant preferred direction is present and they appear to be of oscillation type. The orientation of ripple marks at any one locality is very constant with a variance of less than 40 degrees. The regional variation in ripple-mark orientation is shown in figure 2.77. In the southern coastal localities the ripple-marks have sometimes persisted during deposition and when slight lateral migration has occurred cross-lamination is produced. The ripple-marks generally indicate a break in deposition and in the upper division the troughs may be filled with silty siderite mudstone.

Cross-bedding is rare in the sandstone unit but is seen in Great Fryupdale where the upper part of the upper division locally consists of small disoriented cross-bedded units. Deformation structures are rare; at Hawsker the sandstone unit shows heavy loading at the base and internally but elsewhere load-cast structures are isolated and on a very small scale.

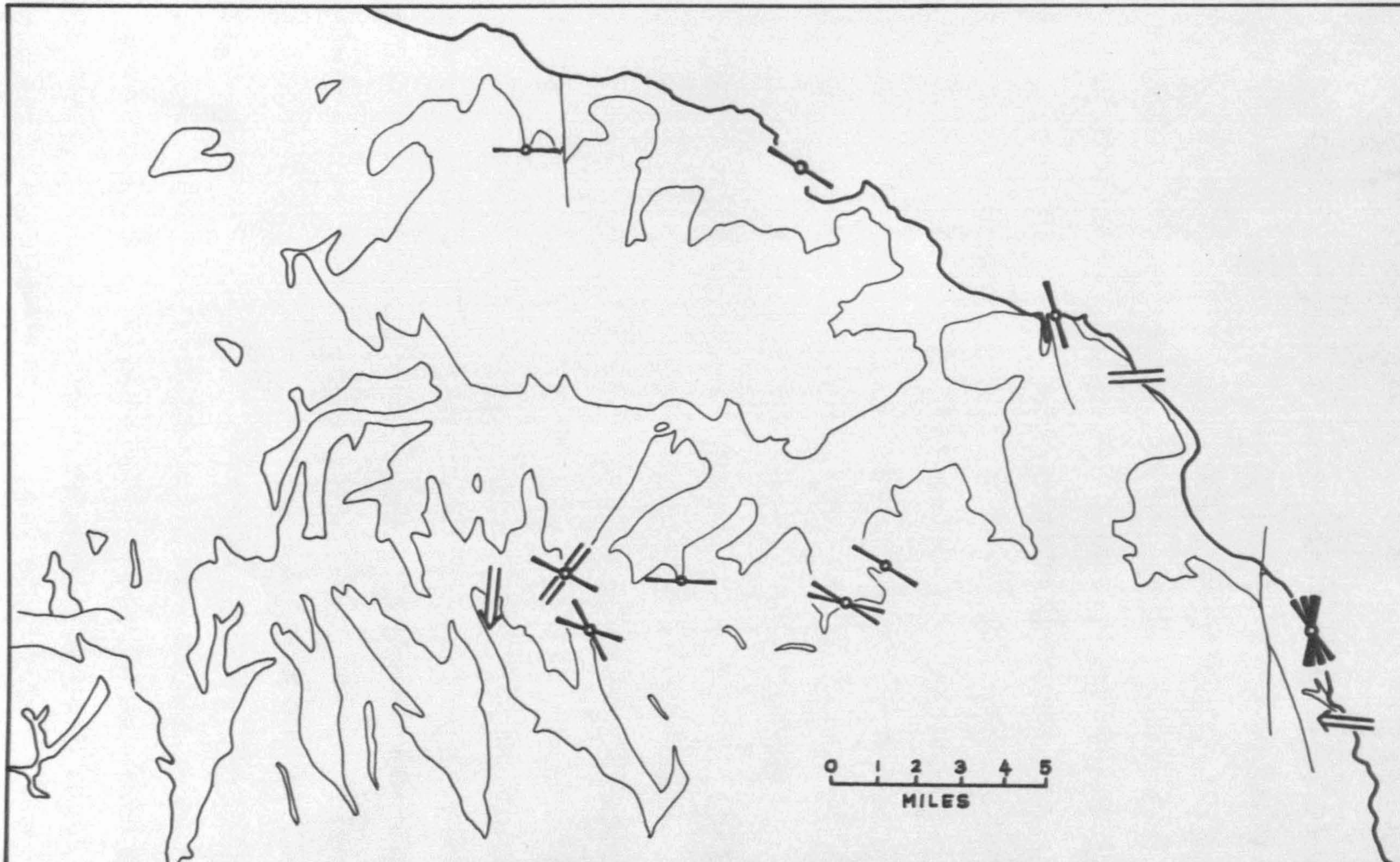


FIG. 2.77 RIPPLE-MARK AND CHANNEL ORIENTATION IN SANDSTONE UNIT

E. EROSION

Erosion of the sandstones is practically confined to the top of the unit. At Iron Scar, however, the lower division is eroded by a steep sided channel which apparently bifurcates to the east. In Wheeldale Gill a much larger structure cuts out the whole of the flat-bedded sandstone development; the channel is very shallow compared with its width which probably exceeds 500 yards. The central parts of the channel consist of structureless sandstone with occasional trough-like partings while to the west cross-bedding and cross-lamination are developed and indicate a northerly source. In Reeking Gill, Rosedale, the sandstone unit is again replaced by a channel sandstone which is probably not very wide. The sandstone displays well developed trough cross-lamination throughout; the individual troughs measure about 9 inches across and 2 inches deep and show a remarkably constant orientation, indicating a current from the north-east.

In other localities such as New Wath Scar and Northdale the top of the sandstone is locally eroded and the erosional troughs filled by fine-grained sediments. These include shale and mudstone with lithologies similar to the underlying marine beds. In the absence of any indigenous fauna these beds cannot be defindtly described as marine but are clearly not typical non-marine deposits. The shales and siltstones which completely replace the sandstone unit in Busco Beck may be of a similar origin, being deposited after erosion of the sandstone unit. Complete erosion of the unit in this area is not improbable since the sandstone in Stonegate Gill is only thinly developed.

II PETROGRAPHY

The sandstones may be divided into two main petrographic groups, those with and those without a siderite matrix and these will be described separately below. In addition to the sandstones, rocks are locally developed in which terrigenous grains are a minor constituent, and the various types will be considered under separate headings.

A. NON-SIDERITIC SANDSTONE

1. Matrix

a) Chamosite. In most samples of sandstone this is seen as a green-grey to golden brown aggregate of minute flakes which forms a film around all grains. The film is relatively thick in larger interstices but does not necessarily fill the whole cavity. The crystals may be disoriented but commonly show alignment parallel to the grain surface which results in the film showing spectral extinction. The birefringence of the chamosite probably reaches first order white but is generally masked by body colour. Chamosite of similar appearance is seen in some sideritic sandstones and rarely it forms small spherulites, about 60 microns in diameter, which contain concentric bands of inclusions. In the upper sandstone division some of the interstitial chamosite is derived from intense spastolithisation of ooliths. Incipient replacement by reddish-brown siderite rhombs in seen in some cases.

b) Kaolinite and Dickite. In cases where chamosite is absent or only partially fills interstices the remaining space is filled by colourless clay mineral, generally kaolinite but occasionally dickite (see

Appendix 2B).

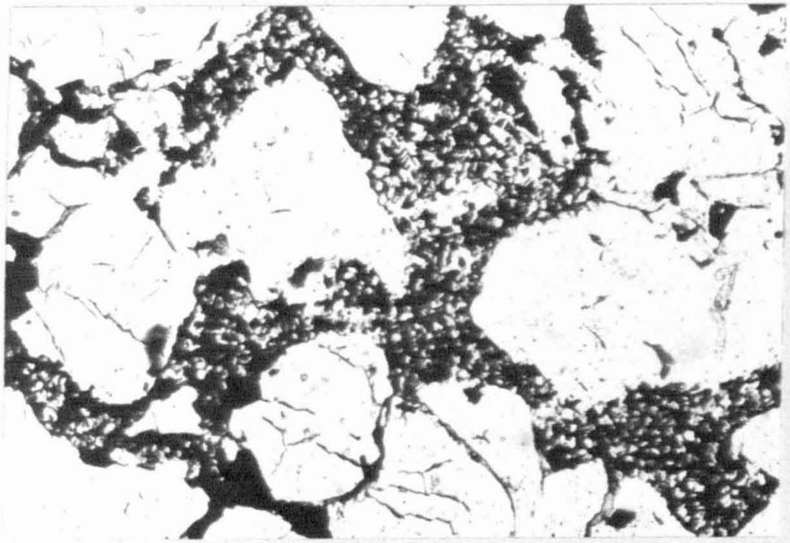
The distribution of these minerals varies considerably between samples but the general structure is similar in all cases. The clay is thinly distributed throughout much of the rock but is locally concentrated in patches; this feature is illustrated in figure 2.78 where both chamosite and kaolinite have been accentuated by iron oxide staining in the weathered zone of the sample. The weathering also accentuates the thin film of chamosite which occurs along inter-granular sutures, and which is difficult to detect in the unweathered rock. The percentage of matrix in these sandstones varies between 10 and 20 percent.

2. Grains

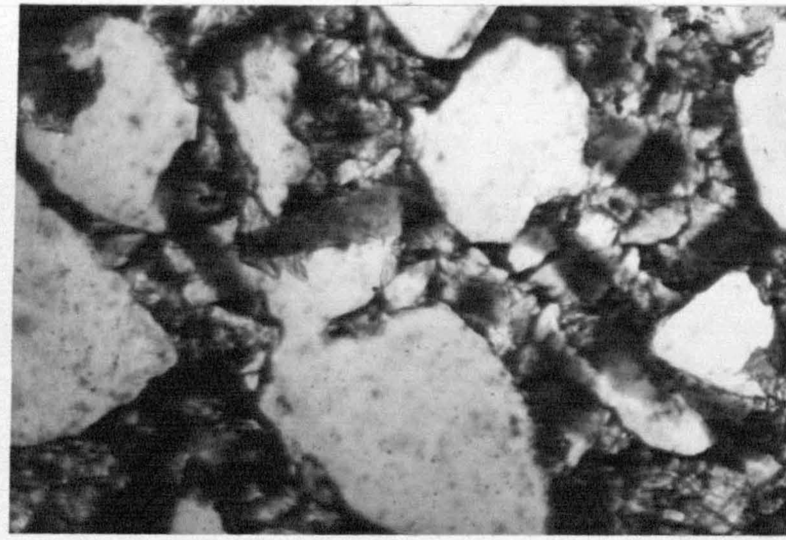
a) Terrigenous grains. The composition of the sand grains is similar to that of the grains in the other units, with quartz comprising over 90 percent of the total. The grains show the effect of pressure solution with the development of a tightly interlocking aggregate of sutured grains; at the same time, however, interstices develop which are clearly larger than the original depositional interstices must have been (fig.2.78). The sutured grain surfaces are generally smooth but quite often one grain penetrates into another. Grain surfaces which line interstices may show euhedral overgrowths in optical continuity with the host crystal, with a faint line of inclusions marking the original outline of the grain. Occasionally the overgrowths may take the form of small prismatic crystals projecting into the interstices. In all cases the overgrowths postdate the chamosite rim.

b) Shells. These are locally a major constituent in the lower sandstone

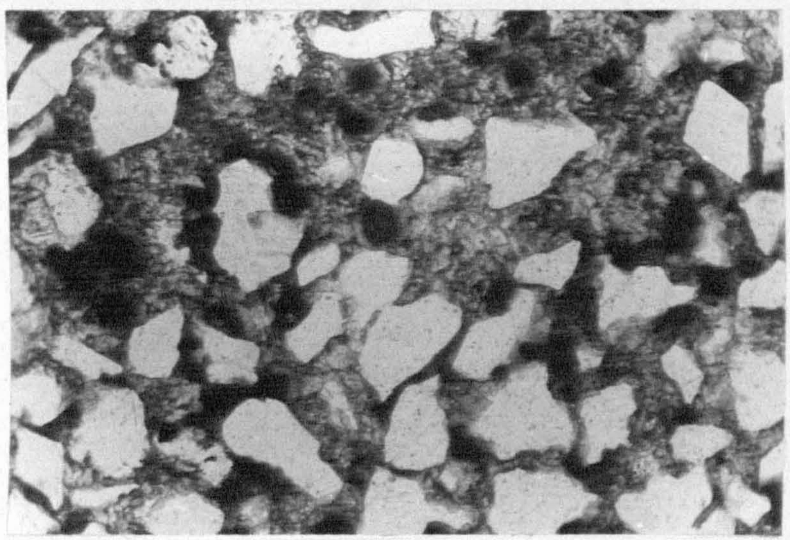
FIG. 2.78	Kaolinite in interstices in sandstone, accentuated by iron staining (sandstone unit)	(x140)
FIG. 2.79	Turbid prismatic siderite with overgrowths of clear rhombic siderite	(x140)
FIG. 2.80	Spherulites of turbid siderite	(x90)



2.78 100 μ



2.79 100 μ



2.80 100 μ

division and in Great Fryupdale the shells are preserved in calcium carbonate in the section along The Scar. The shells show some of the features already described for the other units (pp. 131, 174) including replacement of aragonite shells by ferroan calcite and replacement of original calcite shells by overgrowths from adjacent quartz grains. The replacement of calcite shells by ferroan calcite which is seen to a small extent in the Winter Gill limestone is more prominent, however. The replacement affects crinoid fragments and lamellibranch shells other than those of the oysters. In crinoid fragments replacement has occurred with no loss of the original single-crystal structure. The shells, on the other hand, are replaced by structureless cryptocrystalline ferroan calcite which in hand specimen has a chalky appearance, and which is porous enough to take up an iron oxide staining internally. Replacement clearly post-dates quartz overgrowth formation since replacive overgrowths are abundant and the inclusions retain the original shell structure. Overgrowths are less abundant in the oysters. The replacement by ferroan calcite is associated with cementation by the same mineral. In similar beds which lack the calcite cement the shells are preserved as voidal or clay-filled casts which often show abundant replacive quartz overgrowths in which the shell structure is preserved; these may locally have entirely replaced the shells.

In a thin calcareous sandstone locally developed at the base of the upper division in Northdale aragonite shells are present which have undergone total solution with partial collapse of the cavities and subsequent fill associated with the general ferroan calcite cementation.

c) Ooliths. The ooliths are generally not replaced though extreme distortion commonly takes place. In calcareous sandstone from the upper division they show replacement by single crystals of ferroan calcite, generally with abundant inclusions of clay which preserve the original concentric structure.

B. SIDERITIC SANDSTONE

1) Matrix

The chamosite matrix described above may show partial replacement by siderite in the form of reddish brown rhombs. This replacement may cause complete sideritisation of the matrix and such sideritic sandstones are developed within the upper sandstone division over much of the central part of the area. Much of the matrix in these sandstones consists of reddish-brown siderite which forms a rim around each grain but in some cases drusy siderite in the form of clear colourless subhedral rhombs grows off the replacive rim. Occasionally even the drusy siderite has not completely filled the interstices and the remaining voids are filled by kaolinite.

The replacive siderite rim generally consists of a layer of interlocking crystals of regular thickness but occasionally the siderite shows a more distinctive habit. The normal siderite rim may be accompanied by large prismatic crystals of reddish-brown siderite which reach 200 microns in length and which project into the interstices (fig.2.79). The later colourless siderite rhombs grow off the sides of the prismatic crystals in such a way that the c-axes of the two crystal forms are parallel. A second variety of rim type includes small spherulites of dark reddish-brown siderite in which colour zones and concentric bands of inclusions are present (fig. 2.80). The centre of the spherulites are often void or may contain an isotropic or very low birefringent mineral aggregate apparently of clay mineral, which may be anologous to the spherulitic chamosite which is occasionally developed in the matrix. In other cases the siderite has developed as a thin layer on what must have originally been a spherulitic mineral aggregate but which has subsequently been dissolved. The dark red sandstone of Danbydale contains abundant interstitial oxidised siderite and also unreplaced chamosite.

The matrix of these sandstones contrasts with that of the non-sideritic sandstones not only in mineralogy but in its proportion of the rock; in all cases it comprises between 40 and 45 percent as compared with less than 20 percent.

2. Terrigenous Grains. The sand grains are similar in composition to those of the non-sideritic beds but differ considerably in their fabric. Grain suturing is absent and all grains are separated by siderite except at occasional points of contact. Replacement by clear siderite, sometimes of prismatic form, occurs in some grains.

3. Ooliths. The ooliths may show a small amount of marginal replacement by siderite, but are otherwise unreplaced or replaced by kaolinite. The unreplaced chamosite is usually partially recrystallised with loss of definition of the concentric laminae in ordinary light; this is usually accompanied by a little distortion, but it is not extreme as in the non-sideritic sandstones.

4. Shell Material. Since sideritic sandstone only occurs in the

upper, poorly fossiliferous group, shell material is rarely seen in thin section. It is represented by voids or kaolinite moulds.

C. DIAGENESIS OF THE SANDSTONES

The two types of sandstone clearly have a different diagenetic history but they probably differed little at the time of deposition. The development of siderite may have been related to chemical conditions or more probably to higher chamosite mud content. The contrast in amount of matrix between the two types, must, however, be due largely to diagenesis. The siderite which forms reddish brown rims is formed by replacement of chamosite and therefore chamosite was presumably an original matrix constituent in all of the sandstones.

The chamosite rims may have been derived by percolation of mud or by diagenetic chemical precipitation. The occurrence of rims consisting of either disoriented or tangential oriented crystals suggests that mud percolation is the more likely, forming rims of disoriented crystals which have later adopted a tangential orientation possibly under compactional stresses. Occasionally recrystallisation to a spherulite habit has taken place.

Where sideritisation has not occurred compaction has led to a profound reorganisation of the rock, with solution of chamosite rims and the development of sutured intergranular contacts together with occasional quartz overgrowths. Chamosite rims are preserved in scattered interstices but elsewhere chamosite occurs only as a thin film along intergranular sutures. Sideritisation of chamosite has occurred before compaction since the grains show only a loose packing and as a result the sideritic sandstone shows none of the solution effects described above. Siderite rim replacement may be followed by growth of drusy siderite which generally fills all remaining voids, but occasionally kaolinitic clay occurs at the centre of some interstices. The clay is more common in the non-sideritic sandstones.

D. MECHANICAL ANALYSIS OF THE SANDSTONES

An account of the laboratory method used in these analyses is given in Appendix 2A: the statistical parameters used have already been described (p. 88). Samples have been studied from five localities, which do not provide a complete coverage of the grain distribution over the whole area, but which provide examples of the principal types of vertical succession. Section A, New Wath Scar, represents the typical succession of thin-bedded sandstone passing up into thicker bedded sandstone. Section B. Egton Grange, represents a succession in which the lower part is unusually coarse and has an erosional base, which suggests affinities with the sandstone bodies of the shale unit. Section C, Battersby Crags, represents another more uniform sandstone with an erosional base and which occupies the whole Eller Beck Bed succession. Section D represents the flat bedded sandstone of the northernmost outcrops.

The cumulative curves are shown in figure 2.81 and the values for the statistical parameters obtained from these curves are given in table 2.1.

1. Mean Grain Size

In some sections the mean grain size shows some relationship to the vertical succession. In section A samples 1 to 3, representing

NO.	I%	$M\phi$	$M_1\phi$	$\sigma\phi$	$\sigma_1\phi$
A1	—	4.21	—	0.64	—
2	14	3.95	—	0.88	—
3	5	3.59	—	0.83	—
4	78	2.93	2.40	0.90	0.30
5	74	2.79	2.65	0.38	0.30
6	80	3.33	3.00	0.74	0.30
7	83	2.58	2.30	0.60	0.30
8	72	2.59	2.15	0.84	0.35
B1	83	2.33	2.23	0.41	0.28
2	88	2.58	2.40	0.44	0.30
3	74	2.43	2.20	0.51	0.30
4	84	2.45	2.45	0.31	0.25
5	85	3.14	3.10	0.31	0.25
6	93	2.53	2.15	0.32	0.30
7	78	2.53	2.35	0.43	0.25
8	76	2.83	2.65	0.46	0.30
9	93	2.71	2.65	0.31	0.29
10	76	2.75	2.50	0.48	0.25
11	85	2.94	2.80	0.45	0.35
CI	67	2.55	2.50	0.44	0.30
CII	92	2.68	2.67	0.50	0.38
D1	80	2.58	2.20	0.35	0.20
D2	88	3.27	—	0.45	—
D3	50	2.15	1.88	0.55	0.35
D4	84	2.58	2.35	0.58	0.35

TABLE 2.1 GRAIN SIZE PARAMETERS, SEPARATED AND UNSEPARATED CURVES

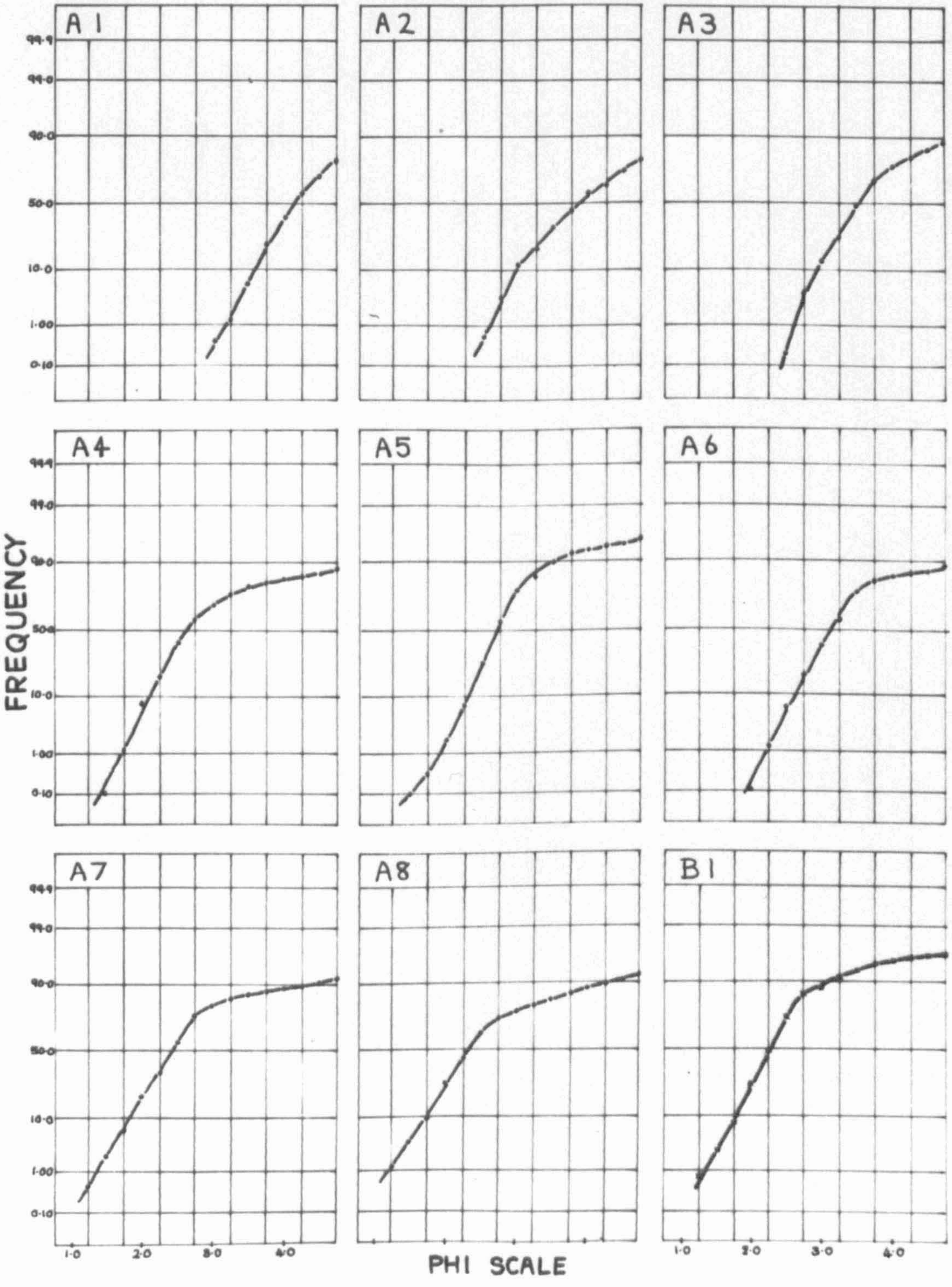


FIG. 2.81(i) CUMULATIVE CURVES FOR E.B.B. SANDSTONES

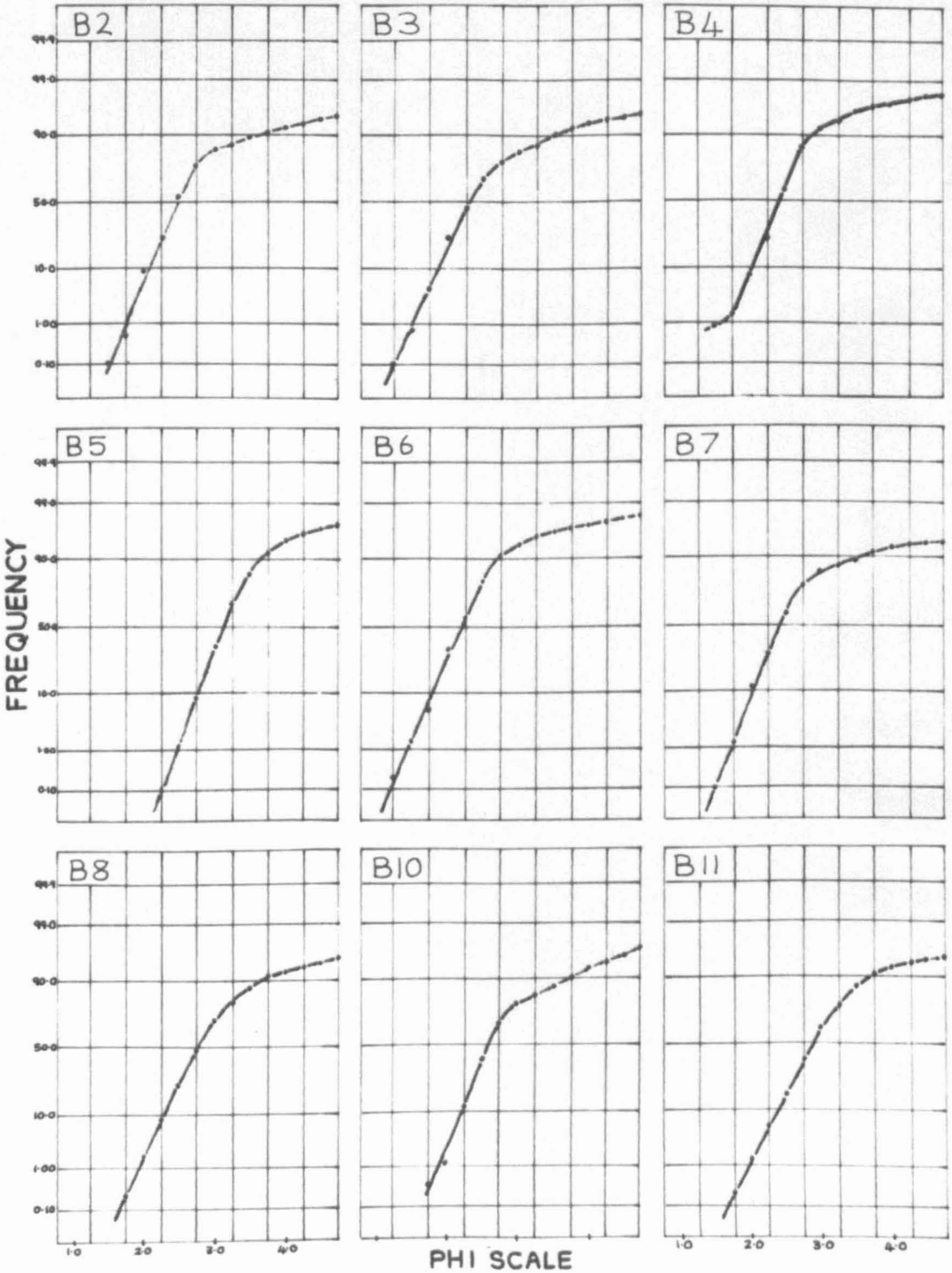

B2
B3
B4
B5
B6
B7
B8
B10
B11
FREQUENCY
99.9
99.0
90.0
50.0
10.0
1.00
0.10
1.0
2.0
3.0
4.0
PHI SCALE

FIG. 2.81 (ii)

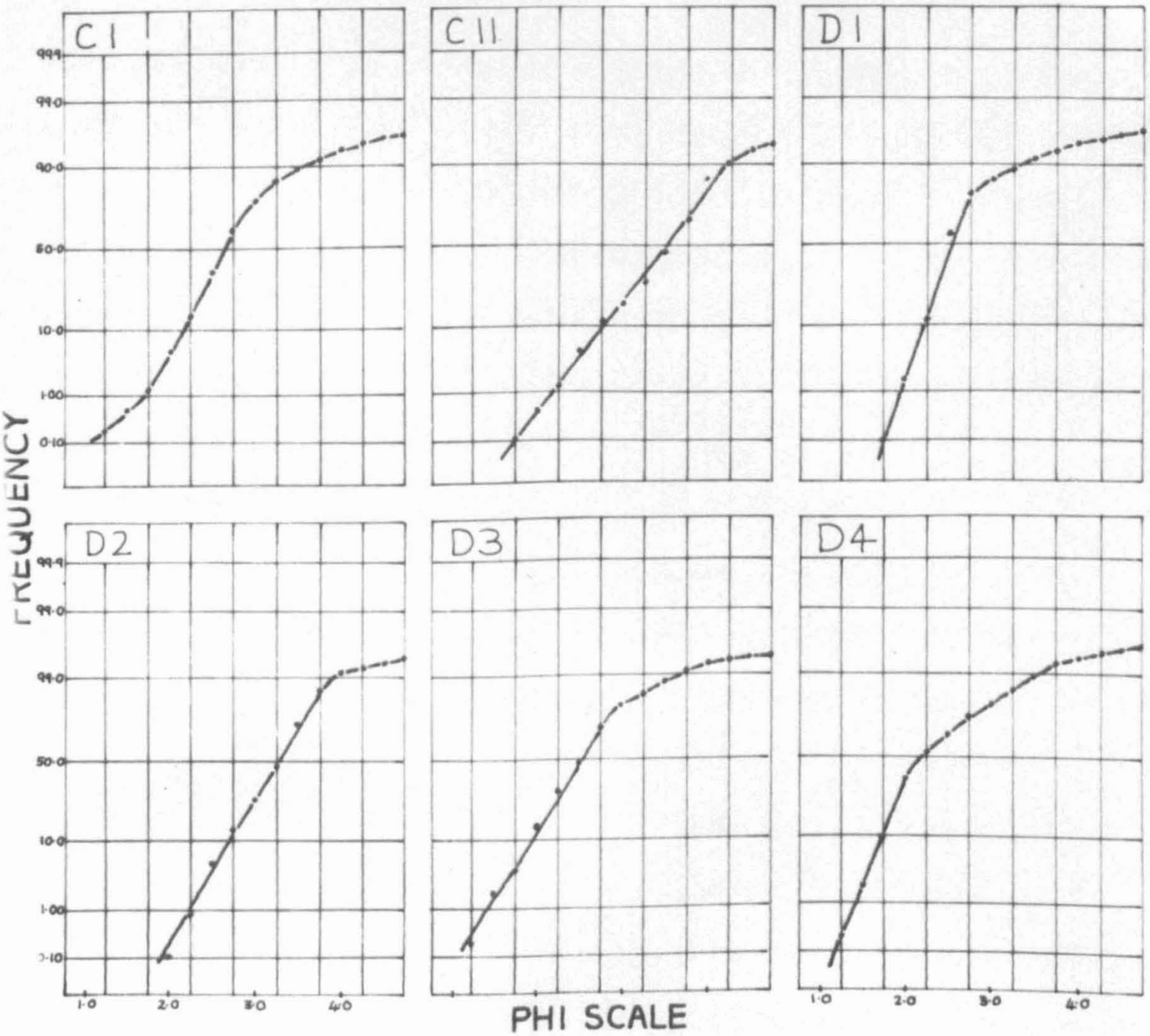
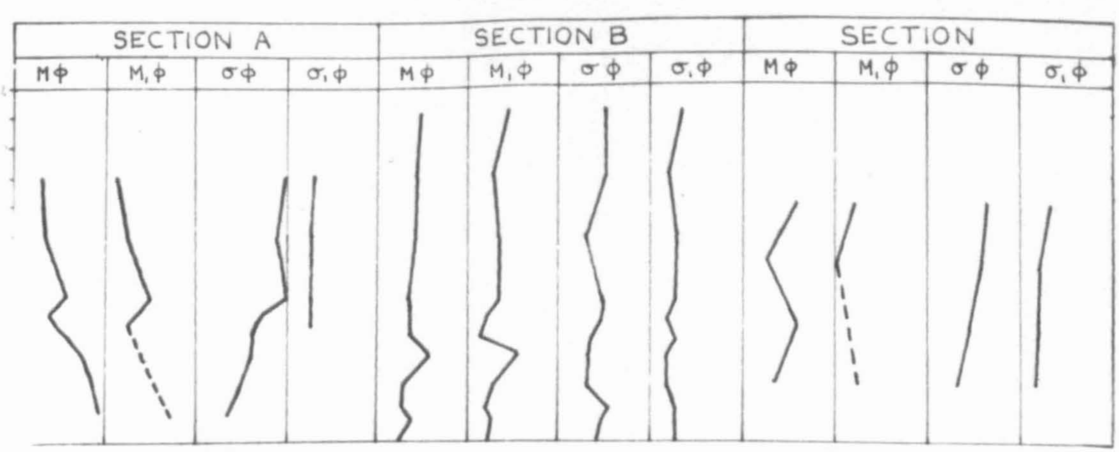


FIG. 2.81 (iii)



IG. 2.82 VERTICAL VARIATION IN PARAMETER VALUES

the lower sandstone division show a steady upward increase in mean grain size from 4.22 phi to 3.70 phi. In the upper division, in which the average mean grain size is 2.85 phi, a slight overall increase occurs, with a sharp increase from the underlying division. In sections B (average 2.63 phi) and C average 2.73 phi) the grain size is fairly uniform, with slight upward fining in section B. Section D (average 3.04 phi), is variable.

2. Standard Deviation

Standard deviation shows little relationship to the vertical succession, but does show variation between sections:

Section A	1 - 3: range	0.63 - 0.88;	average 0.73
A	4 - 8	0.45 - 0.90	0.72
Section B		0.33 - 0.53	0.42
Section C		0.40 - 0.45	0.43
Section D		0.35 - 0.70	0.70

Section A shows a high average and an increase in variability from the lower to the upper division. Section B shows a significantly low average and low variability.

3. Skewness

This is highly variable and shows no relation to vertical succession or geographical location.

4. Inflection Point

The above results suggest that as in the case of the Lower Deltaic sandstone analyses (p. 90) the variation in $\sigma\phi$ and $\alpha\phi$ may be for the most part a reflection of the position of the inflection point which, in itself, also shows no relation to the location of the sample.

Curve separation has therefore been carried out (see p. 90). The values of the parameters of the separated curve for the coarse fraction are given in table 2.1, where they are compared with the values for the unseparated curves and the inflect percent. The curves representing the log-normal fraction of the sandstones are shown in figure 2.81.

5. Mean Grain Aize of Log-normal Fraction.

The figures for this parameter differ from those of the mean grain size of the unseparated curves in having a lower numerical value, reflecting the relative coarseness of the log-normally distributed grains, and in having a lower variability. Vertical variation is slightly modified.

6. Standard Deviation of Log-normal Curves.

Standard deviation is clearly reduced numerically and at the same time the variation is greatly reduced. The new values also show some relation to the vertical succession. The thin-bedded sandstones of Section A show extremely good sorting and the overlying sandstones show uniform values. In Section B the sandstones show more variable $\sigma\phi_1$ values reflecting the rather variable lithology in this section.

The values for $M_1\phi$ and $\sigma_1\phi$ are shown in figure 2.82 where they are related to the vertical sequence in sections A, B and D, and compared with $M\phi$ and $\sigma\phi$. The parameters for the log-normal curves clearly show more relation to the vertical sequence than those for the unseparated curves. Since the parameters of mean grain size and standard deviation are intended to represent the original grain size distribution and hence the conditions of deposition, the values of $M_1\phi$ and $\sigma_1\phi$ would appear to be of greater significance than those of $M\phi$ and $\sigma\phi$.

The general lack of vertical trends in $M_1\phi$ and $\sigma_1\phi$ suggest relatively uniform conditions of deposition, except in the lower part of Section A where the thin-bedded sandstones show a general increase in grain size and very low standard deviation values. The topmost samples of sections A, B, and C show slightly higher $\sigma_1\phi$ than the remainder which may reflect a depositional environment which is closer to that of the fluviatile sandstones whose $\sigma_1\phi$ values (Table 1.2) are mostly higher than those of the marine sandstones. These top samples correspond to a bed of somewhat coarser sandstone which caps most of the sandstone sequences in Eskdale.

Mechanical analyses have not been carried out for the whole area but field observation clearly indicates that the finest-grained beds occur in the extreme south and the coarsest in the north.

E. IRONSTONES

In many sections local deposits of relatively non-sandy character are developed. These are extremely variable in character but are all characterised by being strongly ferruginous. The Winter Gill oolitic ironstone falls into this category but on account of its particular interest has been dealt with in a separate chapter.

1. Siderite Mudstone Bands

The lower division of Iron Scar and Hartoft Beck contains thin bands of silty siderite mudstone. In thin section these are seen to consist of terrigenous grains totalling about 30 percent, set in a matrix of very finely crystalline siderite mudstone. Muscovite flakes are abundant generally showing poor horizontal orientation. The ironstone pebbles from the channel at Iron Scar are similar petrographically.

2. Other Siderite Mudstones

At Commondale the shelly sideritic bed proved to be a sandy siderite mudstone in which the shells show partial sideritisation. Scattered ooliths and intraclasts of chamositic composition are present. A thin sideritic bed in Busco Beck consisted of siderite mudstone with very poorly sorted terrigenous silt and sand grains. As in the other mudstones terrigenous grains constitute about 30 percent of the rock. Ooliths are common and show replacement by kaolinite with a few residual chamosite laminae. In a third mudstone from a ferruginous oolitic sandstone in Northdale sand grains occur dispersed in a cryptocrystalline siderite matrix. There is little lamination and the sand grains which constitute between 50 and 75 percent of the rock, are regularly dispersed. Where the sand grains are least abundant heavy minerals make up 40 percent of the grains (fig.2.83).

3. Chamosite Oolite

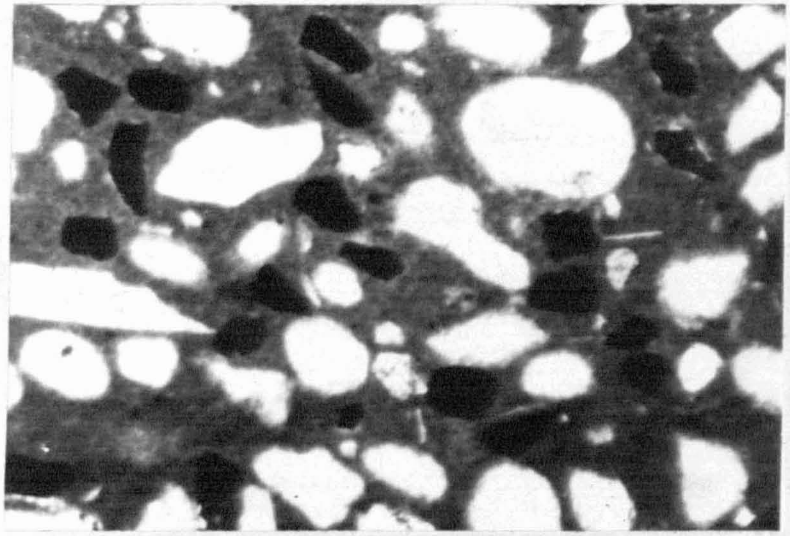
A thin band from the red sandstone of the Danbydale quarry section consists of about 80 percent of chamosite ooliths and nearly 20 percent of siderite matrix with scattered sand grains. The high percentage of ooliths is due to very good packing which has resulted from the flattening of the ooliths in the horizontal plane, often with more irregular distortion (fig.2.84). The chamosite is

FIG. 2.83 Heavy mineral concentration in localised siderite mudstone development (x160)

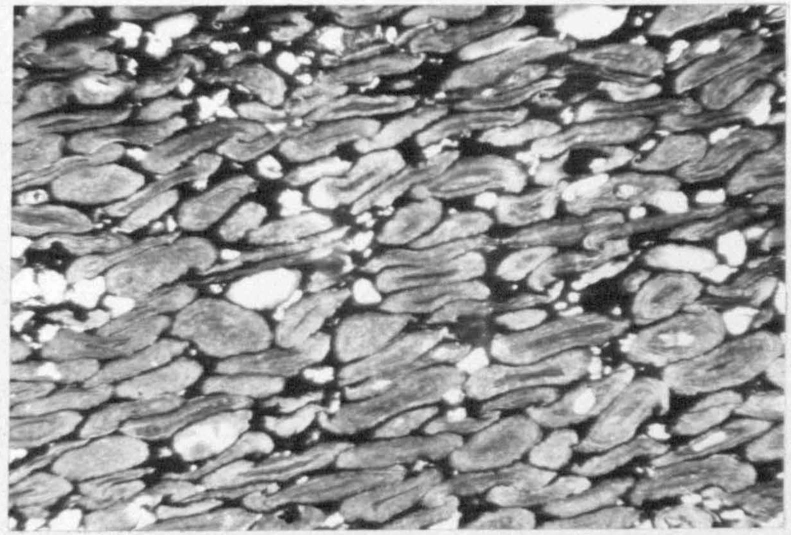
FIG. 2.84 Flattened chamosite ooliths from thin intercalation within sandstone, Danbydale (x50)

FIG. 2.85 Faecal pellet with diffuse transverse structure due to orientation of opaque bodies (x110)

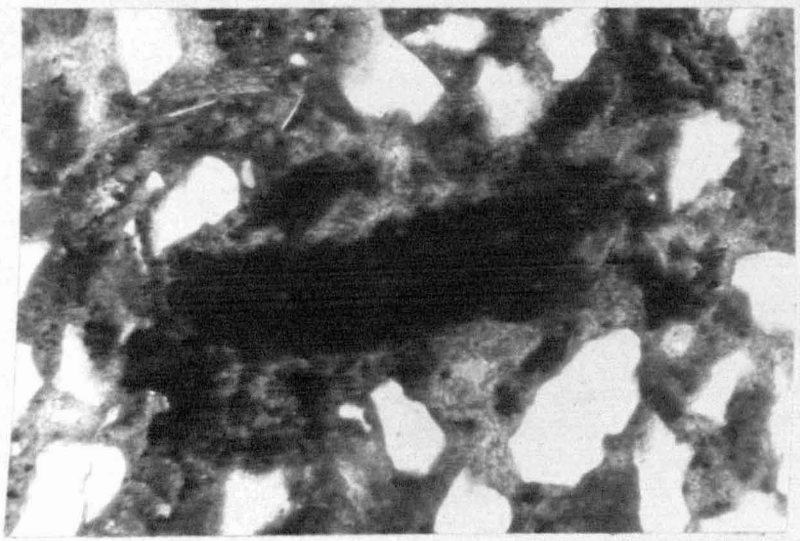
FIG. 2.86 Structureless pellets associated with muscovite (x40)



2.83 100μ



2.84 500μ



2.85 100μ

2.86 500μ

greenish and well crystallised with relatively high interference colours. Most of the ooliths have mud nuclei, though sand grains occasionally occur. The siderite matrix is highly weathered but consisted of clusters of small rhombs, and therefore seems to have been a cement rather than an original mud.

4. Pellet mudstones

In Northdale sideritic muddy rocks are developed at several levels within the sandstone unit, which are characterised by the presence of pellet-like bodies consisting of cryptocrystalline siderite. The pellets are of two types. The first type has been observed in a sandy mudstone from the base of the division; the pellets here are similar to those of the ironstone unit in reaching about 750 microns in length and consisting of slightly elongated opaque bodies (up to 10 by 5 microns in size) set in brown translucent cryptocrystalline siderite. The opaque bodies varying considerably in density of distribution and often show a tendency towards grouping into crude transverse laminae (fig.2.85). The rock is laminated and shows variation in percentage of pellets and terrigenous grains between the laminae.

The second type of pellet has been observed in two mudstones, one from a shallow trough in the top of the sandstone and the other from the middle of the sandstone at a locality where it is very ferruginous and oolitic. The two mudstones have identical lithologies, consisting of pellets, about 500 microns in diameter, of pale to dark brown strutureless cryptocrystalline siderite set in a pale brown cryptocrystalline siderite matrix (fig.2.86).

Terrigenous grains constitute less than 5 percent of the rock, except for muscovite flakes which are abundant and which occur in a disoriented manner. The similarity of the pellets to the matrix make point-counting very difficult but the pellets probably make up over 70 percent of the rock. In one case comminuted plant tissue is abundant.

The distribution of the pellets suggests deposition under winnowing currents for the first type while in both occurrences of the second type sorting is practically absent. The pellets in both cases are probably of faecal origin but there is no indication of the identity of the organisms by which they might have been produced.

III ENVIRONMENT OF DEPOSITION

As with the other units the chemical environment of deposition is considered separately in Part 3. In this case the physical conditions under which the sands accumulated are also dealt with later.

The Winter Gill ironstone which is to be described in the next chapter is believed to be of considerable importance in deducing these conditions, and the discussion is therefore left until the end of the next chapter.

CHAPTER VII THE WINTER GILL OOLITIC IRONSTONE

1 INTRODUCTION

The correlation of the Winter Gill ironstone with the more normal sandstone unit development has been discussed on page 195. It is interesting to note that this unique development of the sandstone unit which may be confined to a very small area is associated with the development of sandstone in the underlying shale unit, and the possible implications of this association are discussed later.

This chapter is primarily concerned with the description of the unusual petrographic features of the ironstone but the petrography of the associated beds in Winter Gill is also dealt with briefly.

II PETROGRAPHY

A. SIDERITE MATRIX

The matrix throughout the ironstone consists entirely of siderite. A great variety in crystal habit and grain size is displayed, each type being of localised occurrence in the vertical succession. A number of specimens collected from the spoil heap have provided additional matrix types not found in the present outcrop. Although the siderite is very variable several distinctive matrix types may be distinguished.

(i) In samples of laminated sandy ironstone from the spoil heap and in the ironstone from the top of the present outcrop the siderite consists of a mosaic of equigranular anhedral crystals. This mosaic includes a rim of a dark reddish brown colour which is developed around all grains and which is about 20-50 microns thick, and an interstitial mosaic of pale clear c ystals. In the samples from the outcrop and in the least sandy laminae of the other samples the matrix constitutes about 30 percent of the rock.

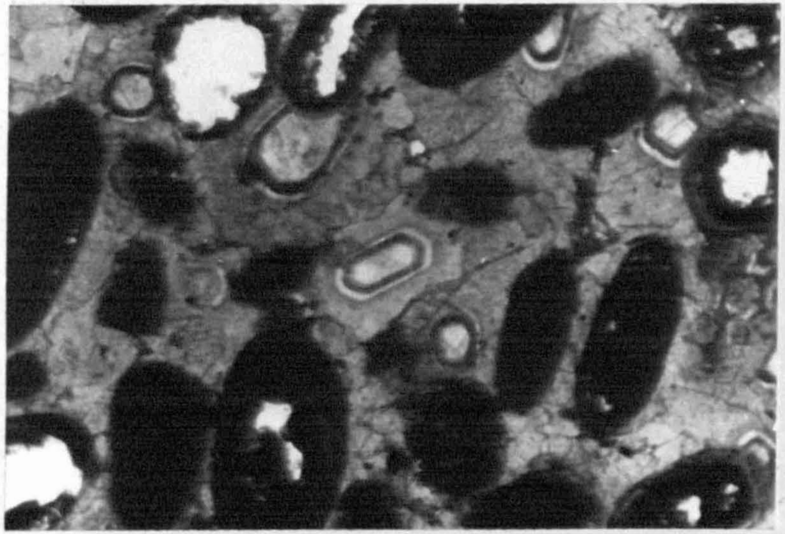
(ii) The crystal size of the siderite increases from the top of the ironstone downwards until crystals of between 50 and 150 microns are present. At the same time the crystals become less equigranular and isolated larger crystals are present which show faint zoning in the form of lines of inclusions and alternations of colourless and brown siderite. The rims of darker siderite are no longer developed. The matrix again consititutes about 30 percent of the rock.

(iii) At lower levels in the ironstone further changes take place. The development of relatively large isolated zoned crystals is more pronounced, and while the crystal size of the general matrix varies between 25 and 50 microns these large crystals may reach 600 microns in length. The smaller crystals which are reddish brown and turbid may themselves show a little zoning.

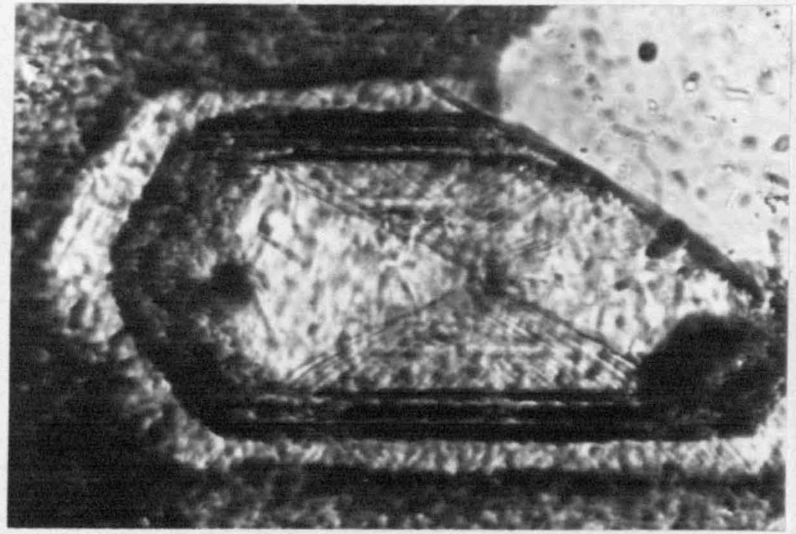
The large zoned crystals, of which examples are shown in figures 2.87 and 2.88, are seen under high magnification to be of complex structure. The outer margins of the zoned crystals are irregular due to inclusion of ooliths and other grains, and to the irregularity of the contacts with other crystals. The elongation varies, reaching a maximum of about 5 to 2. Except in sections perpendicular to the elongation the zones show a hexagonal form, in which two sides are relatively long. Zoning is produced by colour variation of the siderite, and both the dark and the pale siderite may contain lines of inclusions. In general colourless clear siderite predominates in the central part, while a poorly zoned area of brown siderite occurs on the outside. Adjacent grains often protrude into this outermost zone but never into the inner part. Considerable variation in the zonal sequence has been observed but it is very constant in any one sample.

In addition to the concentric zoning the crystals are divided into four sectoral zones which are well defined in the central parts. This produces a structure rather like an hour-glass except that the sector boundaries are rectilinear, and meet at the centre of the

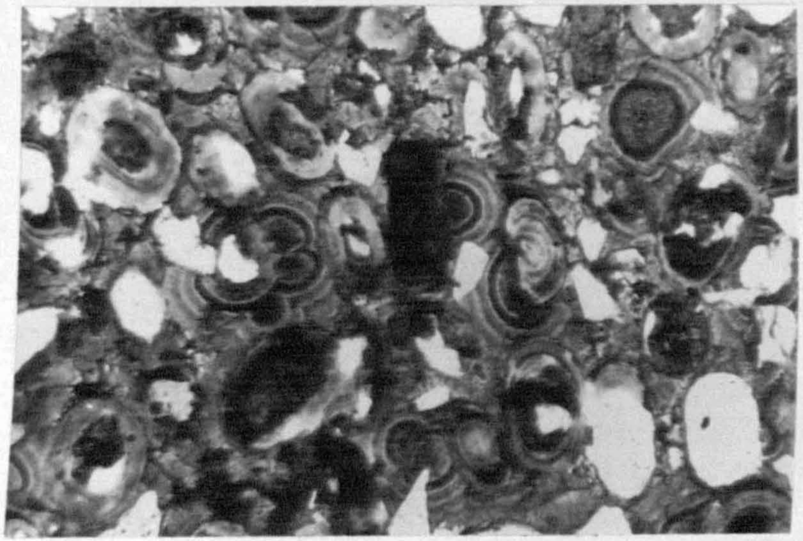
FIG. 2.87	Zoned prismatic siderite (Winter Gill Ironstone)	(x80)
FIG. 2.88	Zoned crystal showing internal structure	(x550)
FIG. 2.90	Zoned spherulitic siderite in matrix	(x30)
FIG. 2.91	Contrast between grain supported oolite (left) and mud supported oolite (right)	(x130)



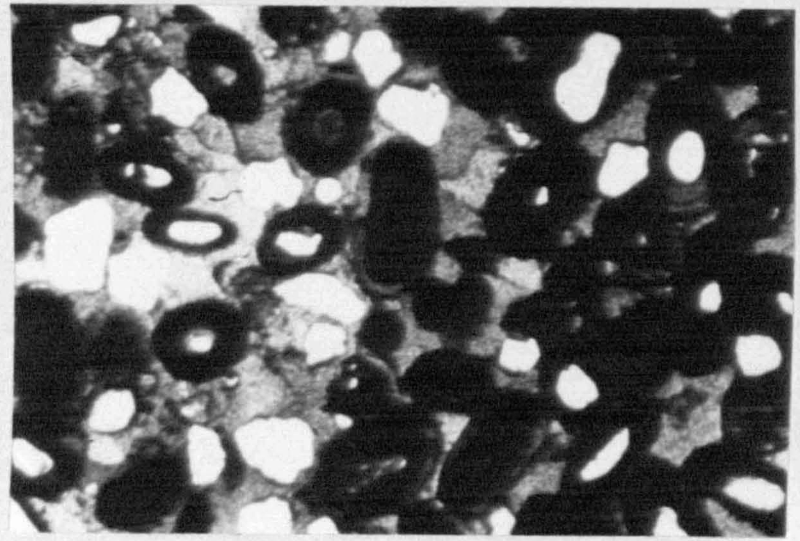
2.87 100μ



2.88 50μ



2.90 100μ



2.91 100μ

crystal. These are best developed in the lateral sectors, where two sets are present; in the end sectors only one set is present. ~~The arrangement of these lines is shown in figure 2.89.~~ Extinction varies through the crystal and is to some extent spectral, indicating partial radial structure. The radial crystallisation is further reflected in the curved cleavage planes which are concave towards the crystal centre. The elongation of the crystal lies in the obtuse angle of cleavage intersection. In some cases the outer zone of brown siderite is formed by four distinct crystals with irregular boundaries which conform roughly to the major and minor axes of the crustal, and in such cases the spectral extinction is confined to the inner zones of the crystal.

(iv) In one specimen small zoned siderite crystals of the type described above have acted as nuclei for truely radial crystallisation. Inclusions are confined to the inner part, but colour zoning, which becomes poorly defined outwards, is developed in the radial part. Since the inner crystal is elongated the whole aggregate is not truly spherulitic, but the crystal fibres in the outer part show true tangential orientation.

(v) In rocks showing the radially crystallised matrix described above small patches are sometimes present which show slight variation from the surrounding matrix. They may themselves vary, but all have the following features in common: a rim of slightly turbid reddish brown siderite is present on all grains; the remainder of the matrix is pale and free from inclusions; the grains are closely packed with the matrix constituting 30 percent of the rock. The turbid rims may be continuous and of even thickness like those described in type (i) but in some cases they consist of discrete subhedral crystals. In all cases the rims are in optical continuity with the spherulitic matrix. The percentage of matrix contrasts with that of surrounding rock which contains between 40 and 45 percent.

(vi) More perfectly spherulitic siderite has been observed from a thin band near the middle of the outcrop (fig. 2.90). These radial aggregates show very strong concentric zoning both by colour variation and by bands of inclusions. As in the other zoned forms the zones become more widely spaced and more diffuse towards the margins. Where two spherulites are in contact with one another they show the typical curved surface of contact which is convex towards the larger individual. There is no visible core to these spherulites and the centre is in fact void with the first zone of siderite having a smooth inner surface. The voids are hemispherical and when the rock is broken and viewed under a binocular microscope the voids are always seen to be situated on the surface of a grain with the siderite zones extending outwards from this point in all possible directions. The diameter of the voids reaches 250 microns, while the whole aggregate may reach 600 microns in diameter.

In the highly variable siderite matrix of the ironstone two distinct types may be distinguished which do not differ in crystal habit alone. In one case the matrix constitutes between 30 and 35 percent of the rock and consists of a rim around all grains together with a further matrix of paler clearer siderite which is free from inclusions, and in which zoning is of a very diffuse nature. The second type constitutes between 40 and 45 percent of the rock and comprises rather turbid reddish brown siderite which often includes large zoned crystals which in some cases at least grow directly off the grain surfaces. Where both types occur in the one sample the crystal habit is the same in both cases but the high percentage matrix has a dirtier appearance due to the presence of inclusions (fig 2.91).

B. TERRIGENOUS GRAINS

1. Distribution and Grain Size

In the ironstone at outcrop the percentage of terrigenous grains varies between 3.5 and 7.5, while in samples from the spoil heap it varies between less than one and about sixty. There is apparently no relationship between the percentage of terrigenous grains and the vertical sequence. The grain size is relatively constant in all samples with a persistent mode at about 200 to 225 microns (figs. 2.92 a - f) but in the top few inches at outcrop the sand is coarser, with a polymodal distribution and an average grain size of 350 microns, (figs.2.92. g,h).

2. Composition

Simple quartz grains are dominant with minor compound quartz and feldspar grains. Heavy minerals are occasionally seen and mostly comprise zircon, tourmaline, and opaque minerals.

3. Solution

A characteristic feature of many sand grains is a highly

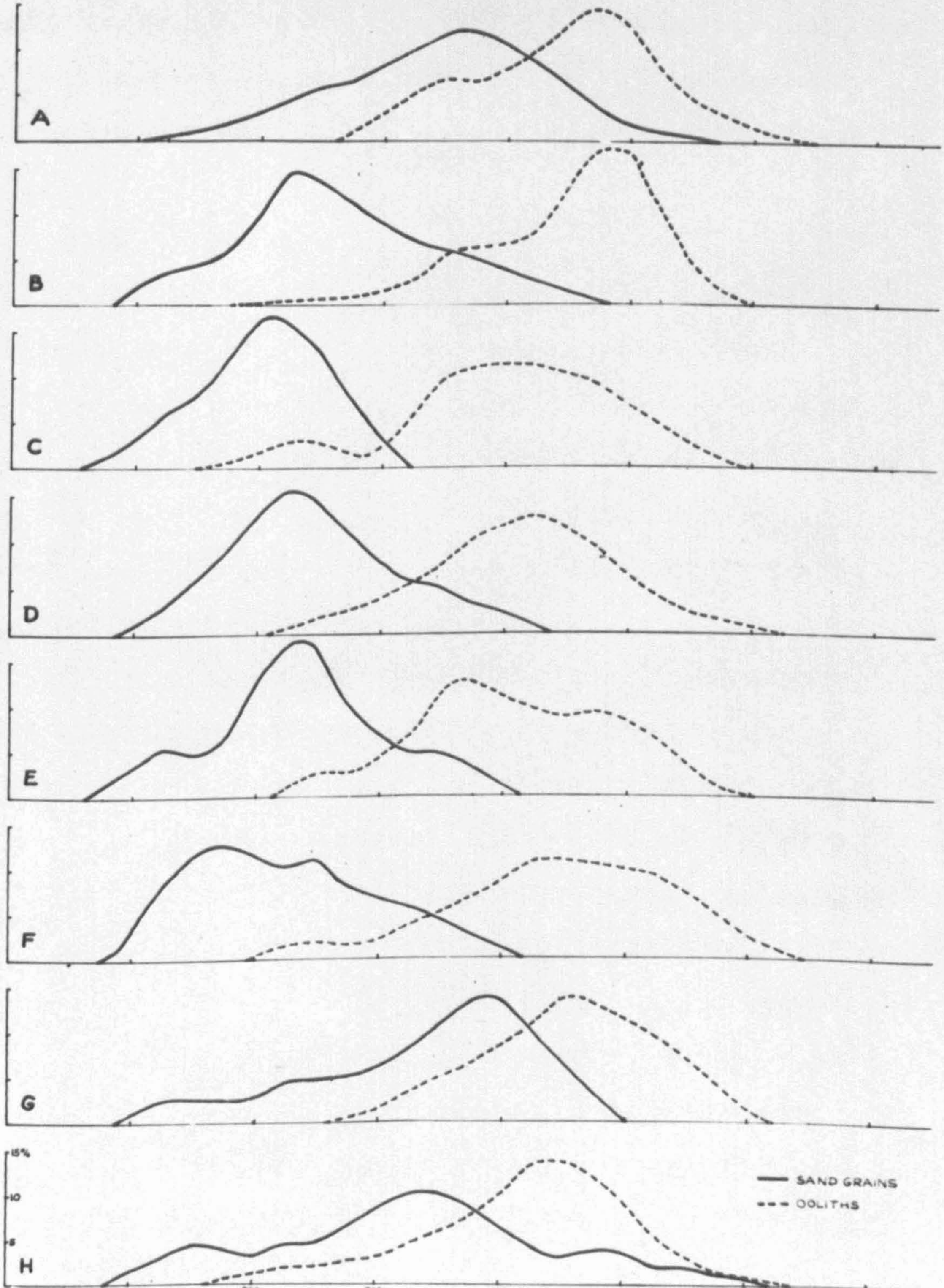


FIG. 2.92 SIZE FREQUENCY CURVES FOR OOLITHS AND SAND GRAINS
WINTER GILL IRONSTONE

irregular but rounded outline. This is also seen in grains which have acted as nuclei for oolith formation. The irregularity of the grains suggests that the rounding is the result of solution, and not abrasion.

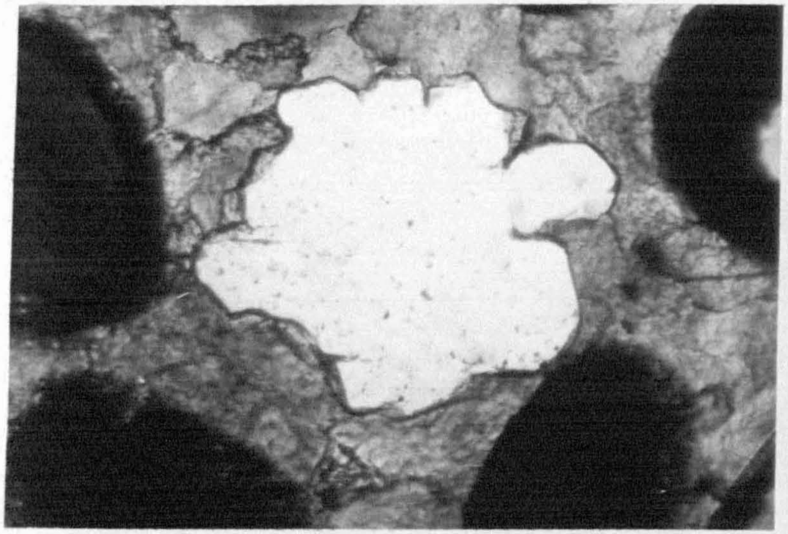
4. Replacement

a) Siderite. The sand grains usually show marginal corrosion by the siderite matrix. The replacing siderite crystals are in optical continuity with those in the adjacent matrix, but are paler and free of inclusions (fig.2.93). Replacement may also take the form of long tongues of siderite which penetrate the grain, presumably along lines of weakness, and which may completely fracture a grain, often with separation of the fragments.

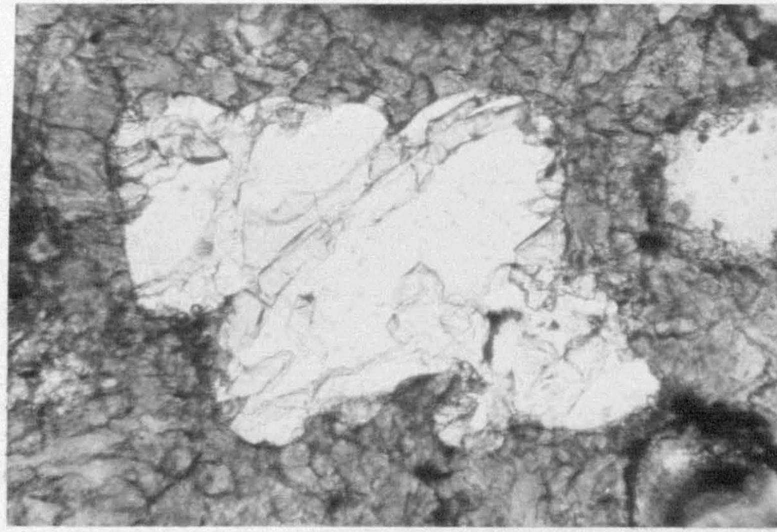
In the top few inches of the ironstone and in several samples from the spil heap the terrigenous grains are replaced by clear colourless crystals which commonly show well-defined rhomb faces and which may reach 0.1 mms. in length. Replacement begins at the grain margin (fig.2.94) and may proceed to completion. The siderite is commonly associated with authigenic feldspar (see below).

b) Pyrite. Replacement of sand grains by pyrite occurs throughout the ironstone but is most common in the lower part. The pyrite may occur in disseminated form as a dusting along the outer part of a grain or along grain sutures in compound grains. More commonly larger euhedral grains which sometimes show striations are associated with the margins of sand grains and replace the matrix and, to a lesser extent, the host grain.

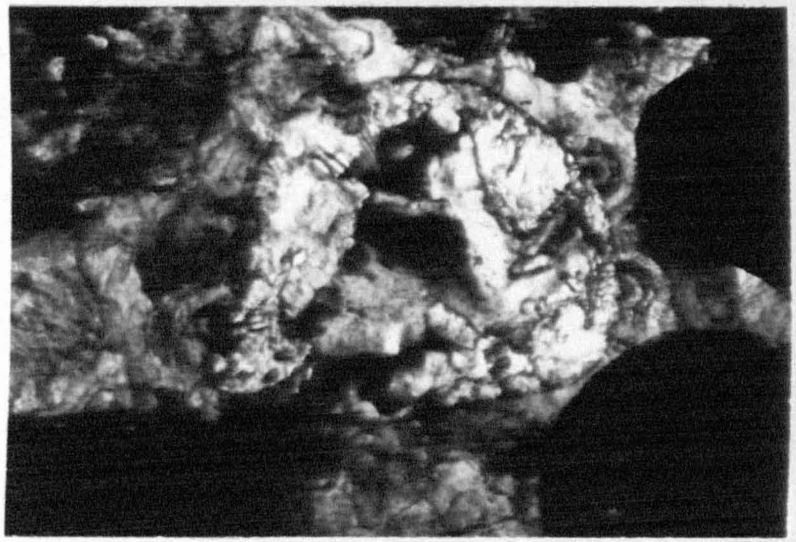
FIG. 2.93	Irregular sand grain (with marginal siderite replacement?)	(x140)
FIG.2.94	Sand grain showing internal and marginal replacement by clear siderite	(x140)
FIG. 2.95	Quartz grain showing replacement by feldspar and siderite	(x50)
FIG. 2.96	Chamosite spherulite acting as nucleus to oolith	(x60)



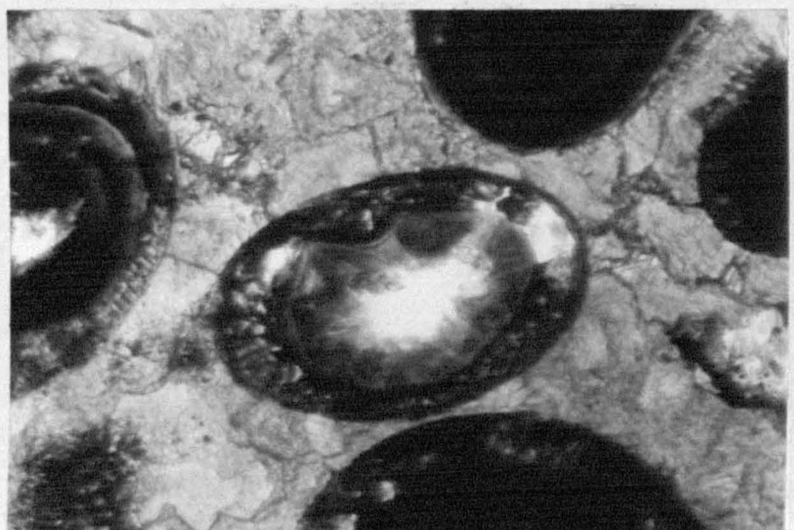
2.93 100μ



2.94 100μ



2.95 100μ



2.96 100μ

c) Phosphate Phosphatisation of sand grains is quite common in the lower part of the ironstone. Compound quartz grains in particular are affected and the phosphate replaces along intergranular sutures forming a network of irregular bands about 10 microns wide. Marginal replacement also occurs and affects simple quartz grains. In the intergranular sutures the phosphate occurs as structureless brown isotropic collophane, but occasionally colloform habit is developed. Individual grains or components of compound grains show internal replacement by fine needles. Replacement also affects sand grains which have acted as nuclei to the associated ooliths.

d) Feldspar. Quartz grains sometimes show replacement by feldspar (fig.2.95) which is confined to the top few inches of the ironstone and is associated with marginal replacement by subhedral siderite crystals. Usually the feldspar occurs as aggregates of small tabular crystals but occasionally isolated crystals occur within the quartz grains; the feldspar is clear and free from inclusions.

G. INTRACLASTS

These are rare and form less than one percent of the rock and include phosphatised mud grains, very rare grains of structure-less chamosite mud, and chamosite spherulites which are the most abundant.

1. Phosphatic grains

These are usually larger than the associated ooliths and reach 1mm. in diameter. The phosphate is orange-brown and consists of minute spherulites often with dark, opaque cores.

2. Chamosite spherulites

a) Composition. These are characteristic of all samples studied and are found both as free grains and as oolith nuclei (fig.2.96). The former are usually very rounded while in the latter sharp projections may be present on the surface. The chamosite is in all cases of a muddy turbid appearance with abundant minute opaque inclusions, and is totally isotropic. The basic structure consists of radiating fibres with or without a structureless core. In some cases the radial pattern is produced by a series of bunches which taper inwards giving a petaloid appearance in thin section. In most cases normal radiating fibres are associated with the bunches; occasionally two or three bunches alone are present and virtually consist of tabular fibrous aggregates which pinch out towards the centre of the spherulite. The structure is much less regular than that of siderite spherulites observed in this study. Where the spherulites act as oolith nuclei chamosite mud may be presence in concave surfaces.

b) Replacement Practically all of the free spherulites show some degree of phosphatisation, and similar replacement occurs in spherulitic nuclei to a lesser extent. The replacing phosphate is distinguished from the chamosite by being transparent and pale orange-brown in colour. The phosphate replaces along the chamosite fibres and itself takes on a fibrous form with fibres reaching 75 microns in width. Each fibre consists of a line of spherulites whose lateral growth is interrupted by the fibre margin. The residual chamosite is often very dark green and structureless. Siderite in the form of crystal aggregates may replace the spherulites, and shows no relation to the radial structure. In the top few inches of the ironstone large crystals of siderite occur, often associated with feldspar which may show a crude radial orientation.

e) Origin of spherulites. The muddy appearance and close association of these bodies with the ironstone suggests that they have formed authigenically within an original chamosite mud in the same way as siderite spherulites are believed to develop. They are, however, not directly analogous to siderite spherulites because each fibre comprises a microcrystalline aggregate.

D. SHELL MATERIAL

Owing to solution of calcium carbonate shells are represented by cavities which are partially filled by early replacing minerals. Phosphate of similar character to that replacing intraclasts is common and may show some relation to the original shell structure. In the uppermost few inches of the oolite phosphate is absent but large siderite crystals associated with feldspar may be present.

E. CONCENTRIC OOLITHS

This term is applied to ooliths whose nuclei are situated centrally within the oolitic envelope and has been used in order to distinguish these normal ooliths from a small number of ooliths in which the envelope is eccentric to the nucleus, and which are here termed 'eccentric ooliths'. These eccentric ooliths will be described at the end of the petrography section.

1. Size and shape

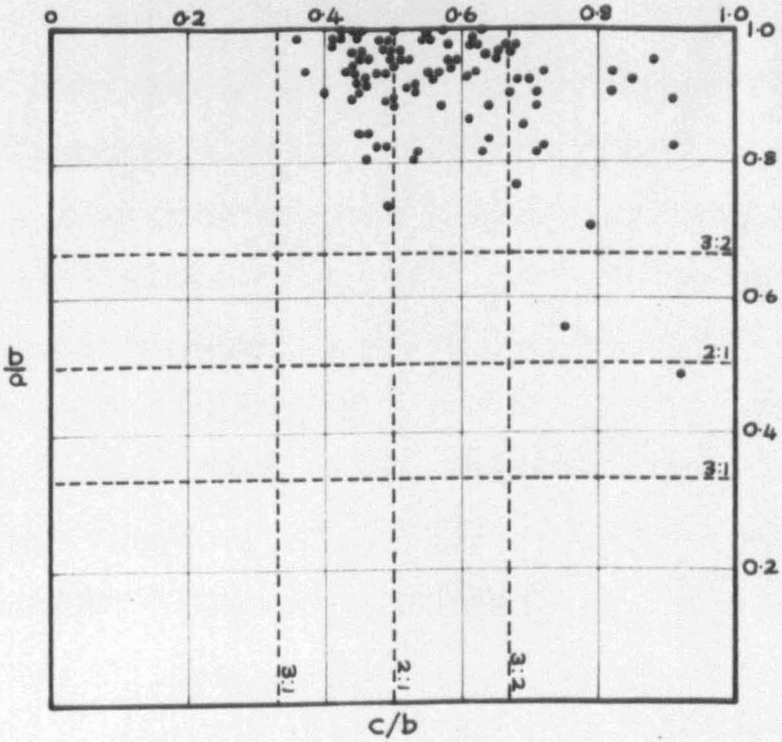
Ooliths were separated from two samples and one hundred from each measured along all three axes. In practically all cases the ooliths proved to be discoidal with two nearly equal long axes and one much shorter axis. The shape of the ooliths is summarised in figure 2.97 in which the upper and right-hand margins represent discs and rods respectively. The maximum observed a : b ratio was 2 : 1 and the maximum a : c ratio was 3 : 1. The length reached a maximum of 600 microns and a minimum of 320 microns with an average of 500 microns; the thickness (c axis) reached a maximum of 400 microns and minimum of 140 microns with an average of 250 microns. Ooliths measured in thin section naturally showed very variable shape parameters, but the maximum elongation observed was 3 : 1 as in the separated ooliths. The size frequency curves for long axes of ooliths measured in thin section are included in fig. 2.92.

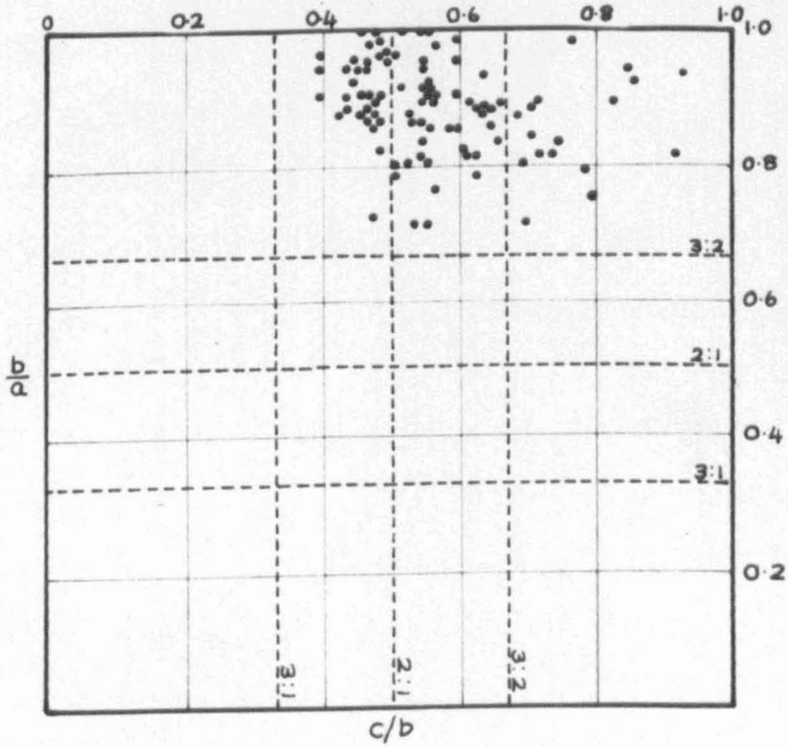
2. Distribution

In the ironstone at outcrop the proportion of ooliths varies from between 55 and 63 percent in the top few inches to between 46 and 53 percent in the remainder. In specimens from the spoil heap the proportion varies between 45 and 65 percent except in sandy ironstones in which the proportion may drop to around 30 percent. Occasionally the sandy ironstones consist of laminae which may be richer or poorer in ooliths than the rock as a whole.

3. Mineralogy

The ooliths vary considerably in appearance in thin section as will be described below. X-ray photographs of two ooliths of very





a = LONG AXIS b = INTERMEDIATE AXIS c = SHORT AXIS

FIG. 2.97 OOLITH SHAPE

different colour gave identical lines which corresponded very closely with the lines for ferrous chamosite (see Appendix 2c). Magnetite is not present, although the rock has been described as a magnetite-chamosite oolite by Hallimond (1925; p.76). When the rock is crushed, however, some fragments are affected by a magnet, and in all cases they are fragments of the siderite matrix; the ooliths are non-magnetic. Siderite from other sedimentary rocks and from hydrothermal veins was treated similarly and in all cases the siderite fragments were affected by a magnet.

4. Structure

Each oolitic envelope comprises several concentric zones which are distinguished from one another by variation in colour. The colour of the chamosite varies from pale green, through olive green to brown. Except in oblique sections these colour zones have sharp boundaries and they form convenient units for description of the oolitic envelopes. Each zone consists of laminae, varying between 0.5 and 8 microns in thickness, which show colour variation among themselves.

The inter-laminar colour variations are associated with variation in the structure of the laminae. The lighter clearer laminae are composed of minute tangentially oriented flakes, and under crossed nicols show first order white. The darker laminae are structureless and isotropic; the crystal size may be smaller than that of the tangentially oriented chamosite but the lack of birefringence may be due to the disorientation of the crystals. The finest laminae of the oriented type are about 5 microns thick, and many of the thicker laminae can be seen at high magnification to be composed of iner subdivisions in the oder of 5 microns. The difference in colour between oolitic zones is due to variation in the proportion of the two types of lamina.

The distribution of the components of the oolitic envelope will be described by using the nomenclature adopted by Chowns (1968) for similar discoid chamosite ooliths. The relatively flat surfaces of the disc are described as the polar surfaces and the curved circular surface as the equatorial surface (fig. 2.98). Since the flattening in the ooliths under consideration is quite considerable the polar and equatorial surfaces are distinguishable in practically every section.

The early oolitic layers are disposed around the nucleus in such a way as to rapidly increase the roundness and sphericity of the original nucleus, with each band thickening into concave surfaces and thinning or even becoming discontinuous over convex surfaces and projections. Thick layers of the dark structureless chamosite form a high proportion of the early oolitic deposits. Once the nucleus is completely enveloped the oolitic layers become more regular, but despite this most zones and many laminae are seen to vary in thickness, sometimes to the extent of being discontinuous.

The variation in thickness may occur in four ways.

(i) Relative thickening over the equatorial surface is the most common form. (fig.2.99). Both light and dark zones show this feature but it is particularly pronounced in the dark zones which are often discontinuous over the polar surfaces. A closer examination

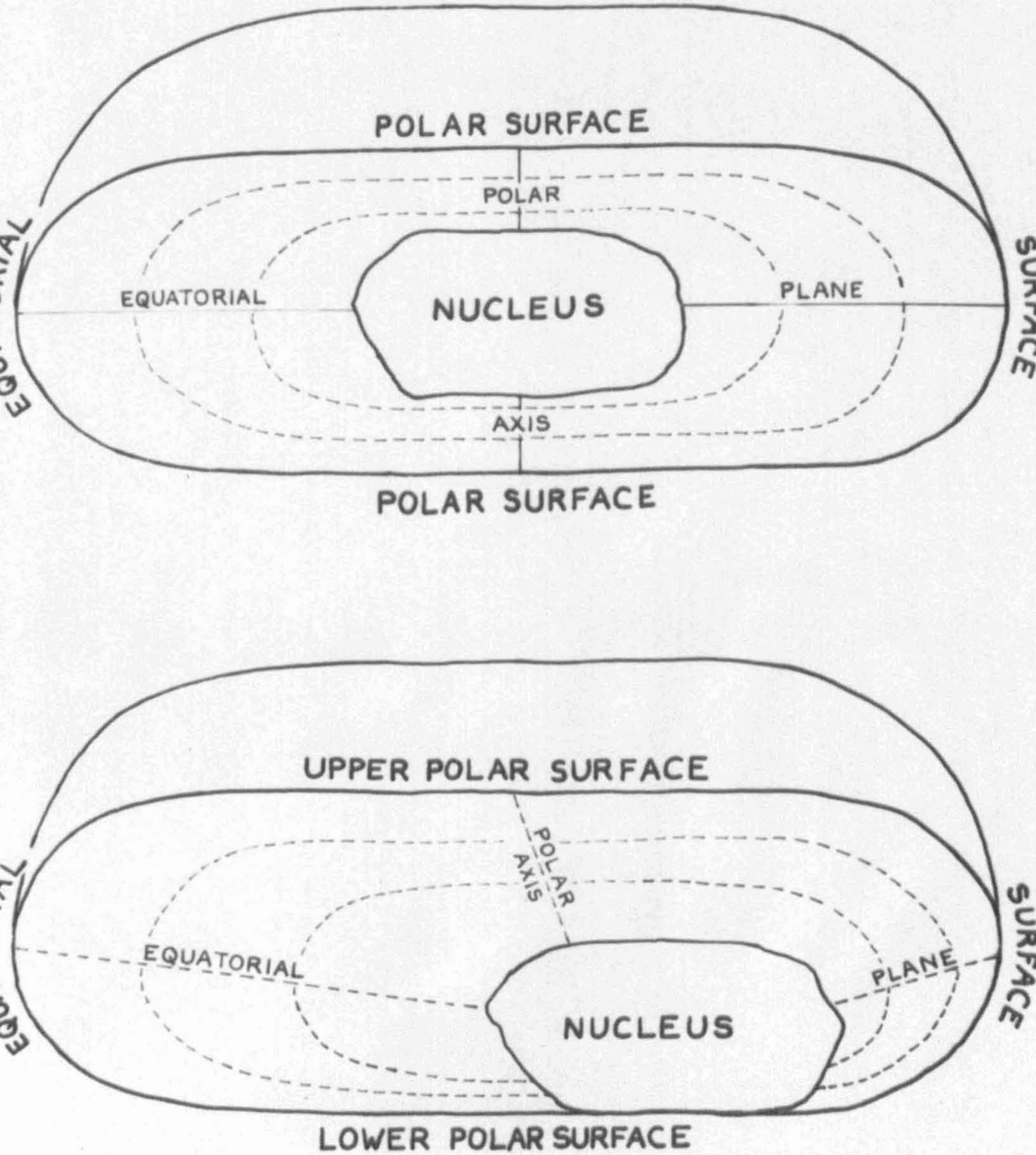
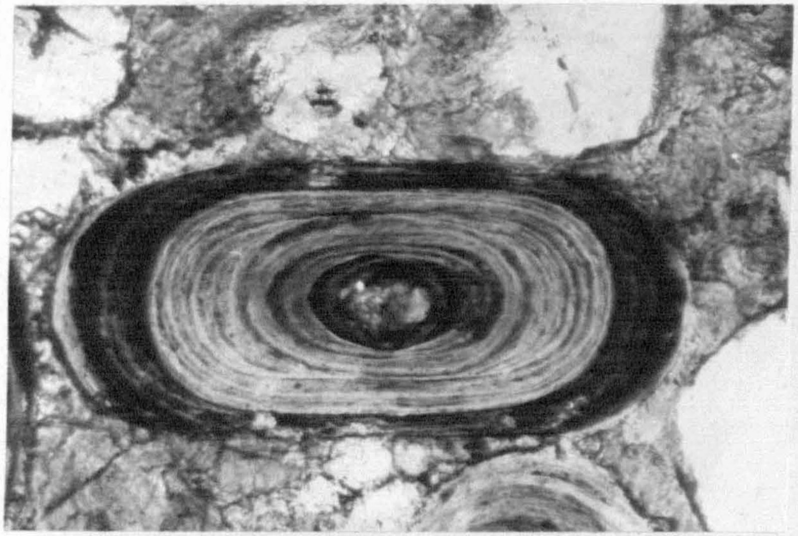
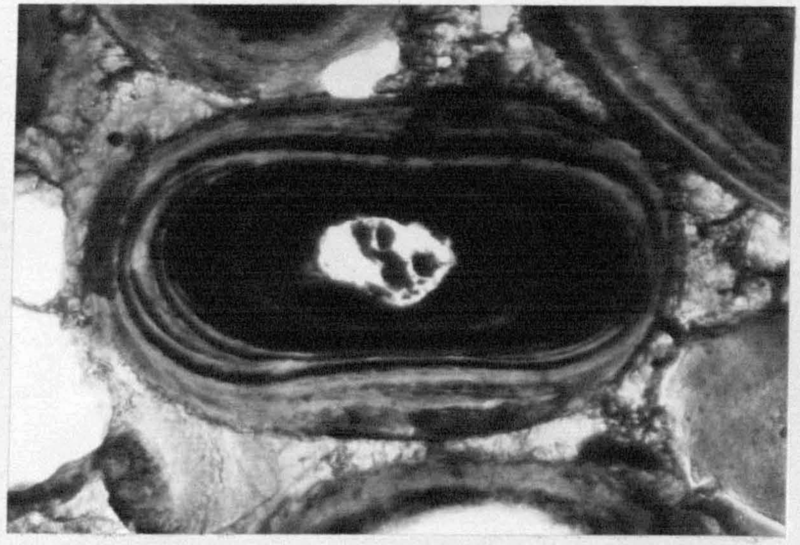


FIG. 2.98 DESCRIPTIVE TERMINOLOGY FOR CONCENTRIC AND ECCENTRIC OOLITHS

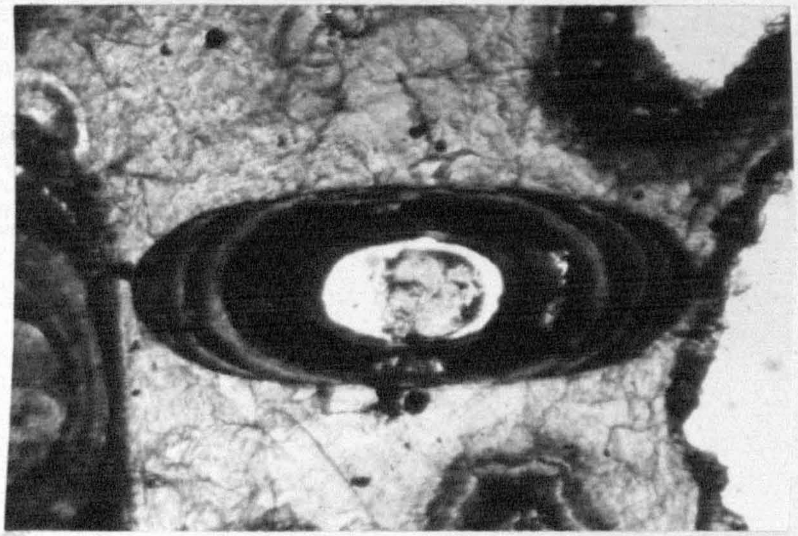
FIG.2.99	Chamosite oolith showing relative equatorial thickening of laminae	(x160)
FIG. 2.100	Oolith showing late-stage polar thickening	(x170)
FIG. 2.101	Oolith showing lack of polar deposition in late stages	(x180)
FIG. 2.102)	Shift in equatorial plane orientation (accompanied by discontinuity of laminae on one polar surface	(x140)



2.99 $\sqcup$ 100μ



2.100 $\sqcup$ 100μ



2.101 $\sqcup$ 100μ

2.102 $\sqcup$ 100μ

of both types of zone shows that the thickness variation is largely due to variation in the structureless laminae. In addition the dark zones may develop relatively thick 'lenses' of structureless chamosite which are sufficiently localised as to produce slight irregularities which are best seen in equatorial sections.

(ii) In many ooliths the equatorial surface itself shows slight variations in thickness of zones and laminae. Laminae are often confined to one side, as seen in thin section, but this preferential deposition normally affects both sides to an equal extent, so that the zone as a whole is fairly symmetrically developed. Exceptionally, preferential deposition on one side involves a series of laminae or even a complete zone, in which case the envelope becomes eccentric relative to the nucleus. Such ooliths are described more fully on page 245.

(iii) Relative thickening over the polar surfaces is (fig.2.100), resulting in an increase in sphericity in contrast to the normal process which, except possibly in the earliest zone, results in a decrease in sphericity.

(iv) An increase in thickness over one polar surface relative to the other may occur as a variation of type (iii), but in many ooliths the individual laminae are discontinuous over one polar surface though since both surfaces are affected to an equal extent the zonal distribution is normal. In contrast to the zones and laminae of type (iii) this type of discontinuous lamina shows the usual equatorial thickening.

Of these four types of unequal deposition the one which has the most profound effect on all ooliths is that involving relatively thick equatorial deposition. From the earliest stages the oolith increases in width relative to thickness and in many ooliths a stage is reached where polar deposition is entirely lacking (fig.2.101). This may be followed by a late stage of renewed polar deposition. Occasionally, where the nucleus is particularly large, the oolith has grown purely by equatorial accretion. In both cases the polar deposition in individual ooliths is somehow dependent on the growth of the oolith. Measurements in thin section of the length of the short axis in all ooliths showing polar non-deposition have been made in order to discover whether the axis length or thickness of the oolith could be the controlling factor. Only sections which passed through the nucleus were considered but oblique sections could still give an exaggerated length, so only the minimum values are of any significance. In both normal ooliths and those with unusually large cores polar deposition occurred until the minor axis reached at least 120 microns. If the diameter of the nucleus exceeded this value then equatorial deposition alone would have occurred, but it is possible that this minimum value does not apply in all cases.

In a few ooliths the equatorial axis shows a sudden change in orientation at an interzonal contact, and all the subsequent oolitic layers are deposited according to the newly established equatorial plane (fig. 2.102). In such cases the later set of zones may be somewhat eccentric to the earlier set and to the nucleus.

5. Oolith nuclei

Most of the nuclei consist of material which is found as loose grains within the ironstone. They include the following types;

a) Terrigenous grains. These consist of quarts with minor quartzite and rare feldspar. The grains are generally smaller than the associated free sand grains (fig.2.103) but have a similar shape.

b) Chamosite mud grains. These constitute about 75 percent of the nuclei. The chamosite is in the form of green to brownish-green structureless crypto-crystalline isotropic mud. The outer margin of these grains is often rather diffuse and merges into the surrounding oolitic envelope. The grains are highly rounded and have a moderate to high sphericity.

c) Radial chamosite aggregates. These are similar to the radial bodies found as discrete grains but are generally smaller. The outer margin may be smooth but is commonly irregular due to projection of radial fibres. The oolitic laminae show the same relationship to these grains as they do to nuclei which are definitely of detritall origin, and in no instances have projections been seen to transect the enclosing laminae.

d) Microcrystalline chamosite grains. These are distinct from those described above in being relatively clear and bright green in colour; they generally show microspherulitic habit. They are considered to result from replacement of other grain types and are considered more fully later.

e) Shell fragments. In some cases where a nucleus is represented by a void the outline is such as to suggest fragmented shell material.

f) Broken ooliths. These consist of ooliths, similar in character to the remainder, which have usually been broken parallel to the

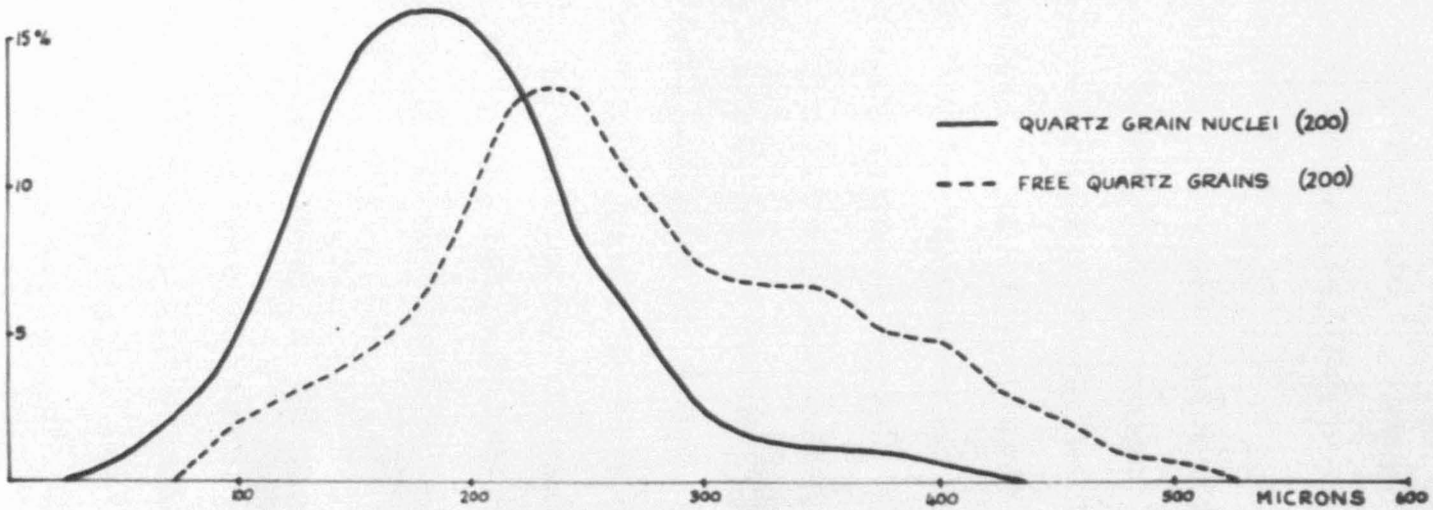


FIG. 2.103 SIZE FREQUENCY CURVES FOR QUARTZ GRAIN NUCLEI AND ASSOCIATED FREE QUARTZ GRAINS

short axis or along zone boundaries.

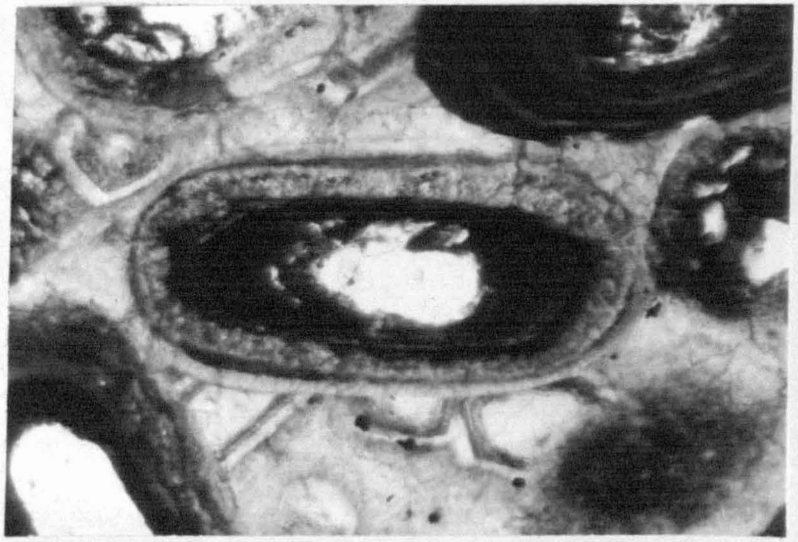
6. Replacement

a) Siderite. Ooliths show partial replacement by siderite in all samples studied, and the various forms in which the siderite occurs are described below.

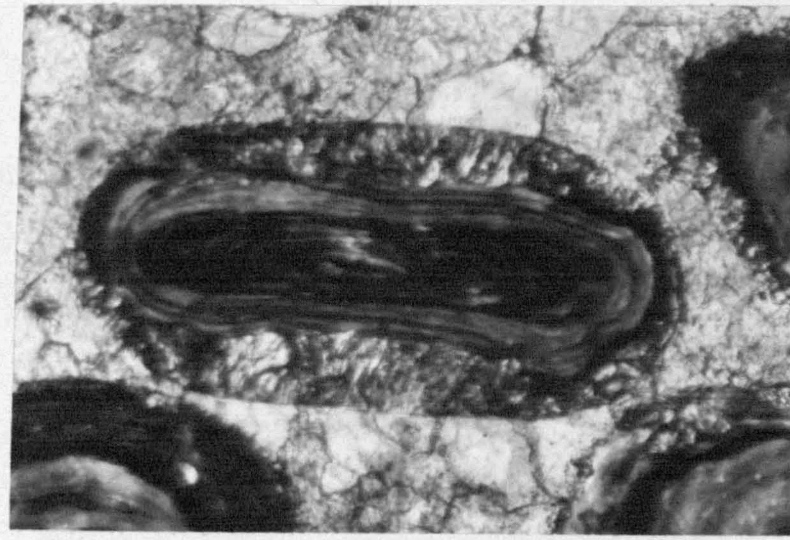
(i) In a few samples the dominant form of replacement consists of marginal rinds of turbid yellow brown siderite, (fig.2.104). At high magnification radially oriented fibres less than five microns across may be distinguished. Under crossed nicols the rind is seen to consist of only slightly elongated roughly radial crystals which have irregular mutual boundaries and which are in optical continuity with adjacent siderite crystals in the matrix, even when these consist of large zoned crystals with spectral extinction. The siderite rinds do not exceed 50 microns in thickness and are generally equally developed around the oolith; sometimes the thickness varies and the rind may be discontinuous, but in neither case does the variation in thickness show any relation to the outline of the oolith. Most rinds follow the oolitic layering but a few transect the laminae.

Bands of siderite identical to the rinds may occur internally and in all such cases the outermost zone of dark turbid chamosite is replaced while a thin layer of clear green chamosite lies outside the siderite and is completely unaffected. Where the outer layers of an oolith have been partially removed by breakage the underlying exposed layers are replaced by the rind which may extend laterally beneath the part of the outer layer that has not been removed.

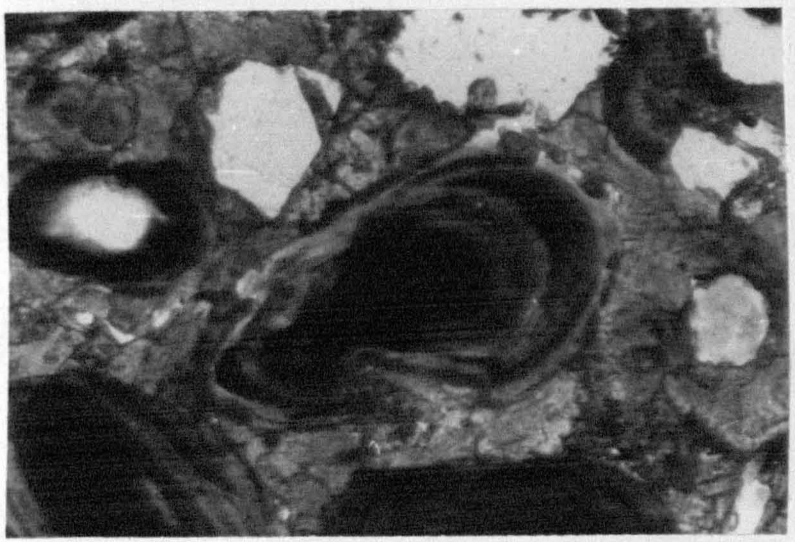
FIG. 2.104	Siderite rind (faint outer zone) with later internal replacement	(x160)
FIG. 2.105	Discontinuous rind showing preferential development on polar surfaces	(x160)
FIG. 2.106	Irregular marginal siderite replacement	(x130)
FIG. 2.107	Partial replacement of recrystallised chamosite oolith by siderite	(x200)



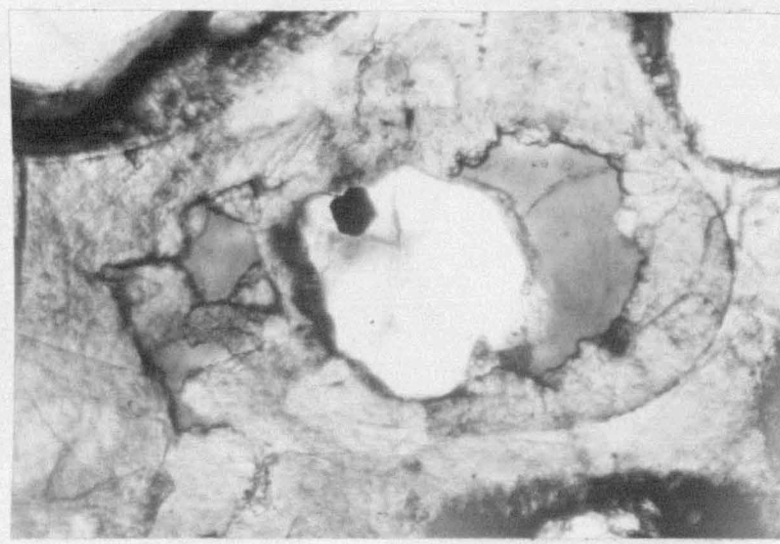
2.104 100μ



2.105 100μ



2.106 100μ



2.107 100μ

(ii) Where the first type of marginal replacement is present only about a half of the ooliths are affected, but a second type is developed which occurs in the remainder of the ooliths and which is occasionally superimposed on the first type. Where both are present the siderite rind is developed as described above, though rather thinly, and the second type occurs within the inner margin of the rind. In this second type the siderite occurs as pale brown to colourless subhedral rhombs about 25 microns across; where the crystals are closely packed they are elongate and parallel-sided and are oriented at a high angle rather than perpendicularly with respect to the oolitic layering. This siderite is patchily distributed and rarely continuous; in each patch the crystals usually have a common extinction, but in larger patches two or more extinction groups are present which are also distinguished by abrupt changes in orientation. The crystal terminations usually produce a saw-tooth inner margin. The siderite may occur anywhere within an oolith but in most cases it is best developed on the polar surface of the ooliths (fig. 2.105). This localisation is not related to the orientation of the oolith with respect to the bedding or to the relations of an oolith with the adjacent ooliths. Occasionally siderite of a similar habit occurs along the interface between a quartz grain nucleus and the envelope.

The occurrence of this type of siderite is accompanied by considerable distortion of the adjacent oolitic laminae, which show considerable thinning out over projections of siderite. At the lateral extremities of the siderite patches the oolitic laminae are clearly truncated but this appears to be largely a displacement effect since all zones and most of the laminae of the outer zone can be traced from one end of a patch to another. The siderite shows a similar relationship to broken ooliths as in the preceding type but again the siderite appears to have displaced rather than replaced the oolitic material whose place it occupies.

(iii) This type of marginal replacement is confined to the ironstone with a truly spherulitic matrix. It is very irregularly distributed and consists of 'dirty' brown siderite crystals which are in optical continuity with adjacent crystals in the matrix, and therefore shows spectral extinction; zoning, although prominent in the matrix, is absent. The most irregular siderite masses are accompanied by extreme distortion of laminae, and as in type (ii) the thickness of the siderite exceeds that of the chamosite replaced (fig.2.106).

Siderite of a similar nature is developed internally around the nucleus in some ooliths. Where the nucleus is chamositic it shows partial or complete replacement and quartz grains show slight replacement. The siderite shows spectral extinction and can sometimes be matched with a large spherulite in the surrounding matrix. The combination of internal and marginal siderite may lead to total replacement of an oolith.

(iv) Another form of marginal siderite affects only recrystallised ooliths (page 207), and may result in total replacement. The siderite occurs as clear crystals, often slightly reddened, and between 50 and 150 microns in length. They pass with little change into crystals of the matrix but the original outline of the oolith is reflected in a line of reddish inclusions (fig. 2.107).

(v) Internal replacement unaccompanied by any form of marginal replacement affects ooliths in the top few inches of the ironstone. Shells and intraclasts are similarly affected. The siderite is in the form of anhedral crystals between 100 and 300 microns in length. Replacement is always from the centre outwards, affecting both the nucleus and the surrounding oolitic material. The siderite has a rather dirty appearance except where it replaces certain nuclei. Where oolitic layers are replaced by large crystals lines of inclusions may remain which correspond to the dark zones of the oolith. The only other siderite replacement to occur in these beds is of type (iv).

(vi) In samples which show total recrystallisation of the matrix to poorly zoned radial siderite aggregates (see type (iv) p.219) total replacement of ooliths may occur. The siderite is in the form of large anhedral crystals which show spectral extinction, being in optical continuity with spherulites in the matrix. Abundant inclusions representing replaced laminae are present and any unreplaced chamosite is very much darkened. Occasionally rinds of type (i) are seen around less replaced ooliths; these may show radial crystal orientation but have commonly recrystallised into optical continuity with the matrix. Type (iv) replacement may also occur with the usual disruption of laminae.

b) Phosphate

(i) Rare ooliths are seen which have been entirely replaced by colourless to dark brown collophane. Almost total loss of oolitic structure has resulted and shrinkage cracks developed.

(ii) At one level in the ironstone ooliths are replaced by pale orange brown slightly birefringent microspherulitic phosphate. The oolitic structure is well preserved and often accompanied by regularly spaced opaque specks in some zones. Replacement may not be complete in which case zones of tangentially oriented chamosite remain. The opaque specks are absent in these unreplaced zones, and seem to be associated with phosphatisation of zones in which the structureless laminae are dominant. The phosphatised ooliths may be apparently isolated or may occur in groups, often associated with shell fragments. The phosphatised ooliths are unaffected by siderite replacement.

(iii) Oolith phosphatisation of a slightly different type is developed on one particular horizon of the ironstone. It differs from type (ii) in being more turbid and in having a deeper colour; it is generally isotropic. The structure of the replaced oolitic material is retained but not with the same perfection as in type (ii). The phosphatisation occurs in a horizon in which a late stage of thick polar accretion of clear chamosite laminae is commonly developed, and this outer zone is unaffected by the replacement.

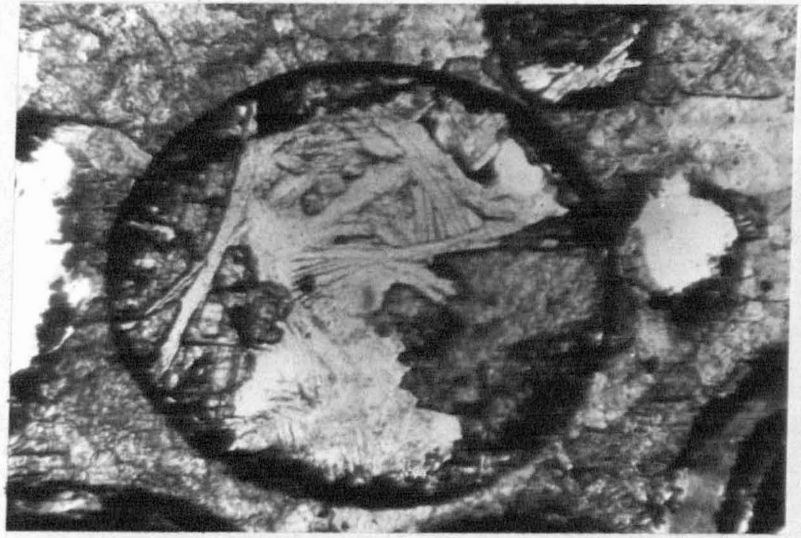
c) Pyrite. The central part of an oolith is occasionally replaced by massive pyrite but some ooliths which have been replaced by collophane are also replaced by disseminated pyrite either in dusty form or as small euhedral crystals.

d) Feldspar Ooliths show replacement by feldspar in the uppermost few inches of the ironstone. The feldspar occasionally occurs as large solitary crystals, but is more commonly in the form of aggregates of disoriented smaller crystals. These smaller crystals range from equant to highly elongate (fig.2.108). Replacement affects both the nucleus and the envelope and in some cases the outer part of the oolith is unaffected when the feldspar crystals can be seen protruding into and cutting the oolitic laminae.

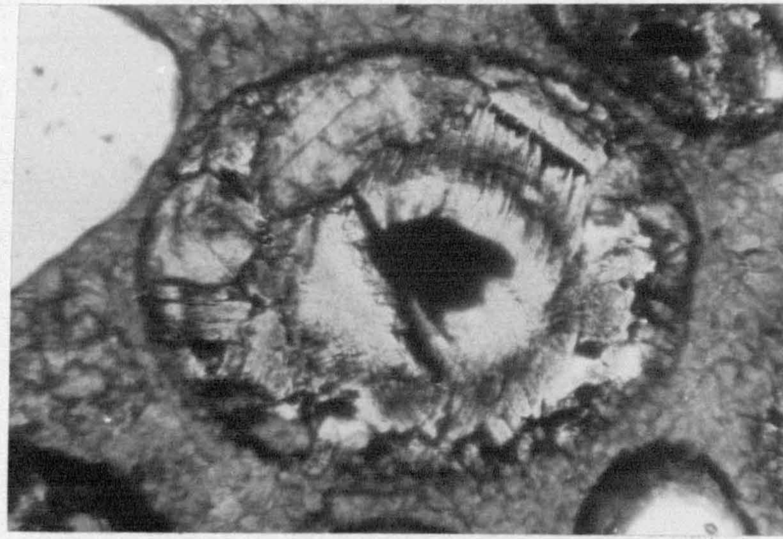
In a few ooliths the nucleus is replaced by a single clear feldspar crystal and this may possess overgrowths of 'dirty' feldspar, crowded with inclusions, which is replacing the oolitic envelope. The overgrowths may have a smooth outer margin or may be highly serrated (fig.2.109) and in a few cases are euhedral (fig.2.110). The inclusions may be arranged in concentric bands, reflecting the original structure. The serrated crystals occur where inclusions are most abundant and since occasionally further non-serrated growth has occurred, the serrated portion represents a very impure chamosite zone. In the feldspar aggregates the inclusions are concentrated interstitially. Except in rare cases where the feldspar crystal boundaries correspond to the original oolitic layering the feldspar tends to be euhedral to chamosite. Feldspar crystals are also euhedral to enclosing replacive siderite crystals. Though partial sideritisation of the feldspar is common feldspar crystals not occurring as overgrowths may show albite twinning and the X-ray diffractometer trace of the insoluble residue of the feldspathic ironstone (Appendix 2 B (vii)) also indicates albite as being the major constituent.

<u>e) Chamosite recrystallisation</u>. A feature of a few ooliths which occurs throughout the ironstone involves complete reorganisation of

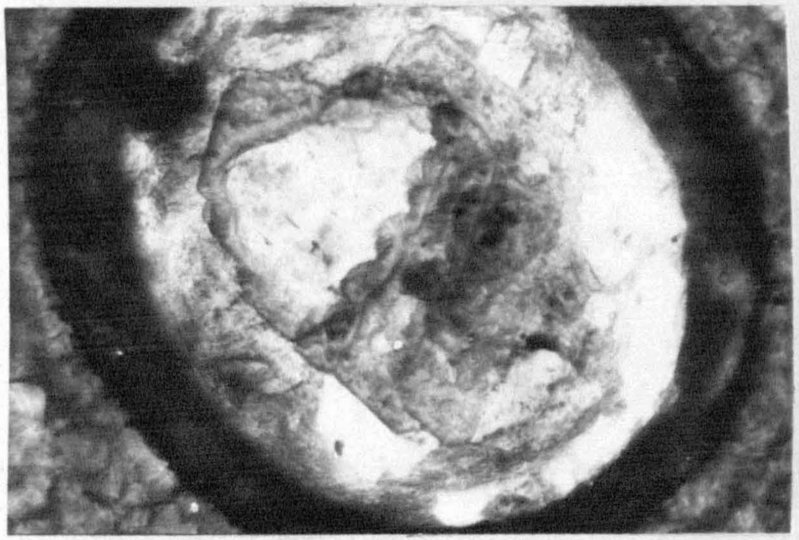
FIG. 2.108	Feldspar laths and siderite replacing oolith	(x120)
FIG. 2.109	Overgrowth of feldspar on oolith nucleus showing serrated margin	(x120)
FIG. 2.110	Euhedral feldspar overgrowth showing replacement by radial aggregates of chlorite	(x350)
FIG. 2.110	Recrystallised chamosite oolith showing almost total loss of internal structure	(x130)



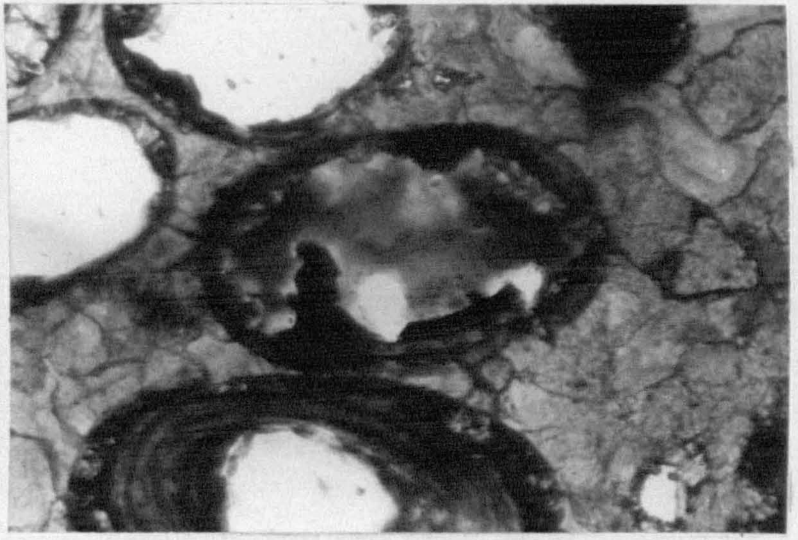
2.108 100μ



2.109 100μ



2.100 100μ



2.111 100μ

the chamosite in the envelope and in the nucleus, if chamositic. This results in the development of green to brown structureless isotropic cryptocrystalline material resembling chamosite mud (fig. 2.111), and the grains could be classed as pseudo-oolites (Carozzi, 1960,p.238). Their oolitic origin is shown by the occasional presence of quartz-grain nuclei, by concentric lines of inclusions sometimes by a faint tangential orientation of the crystals.

Occasionally the structureless chamosite extends beyond the original limit of the oolith and forms a bridge to an adjacent oolith. Bands of inclusions when present show distortion which indicates that the material has been extruded from within the oolith. Sometimes these recrystallised ooliths occur in groups, with mutual flowage of material and these groups act as a single body to siderite replacement.

F. ECCENTRIC OOLITHS

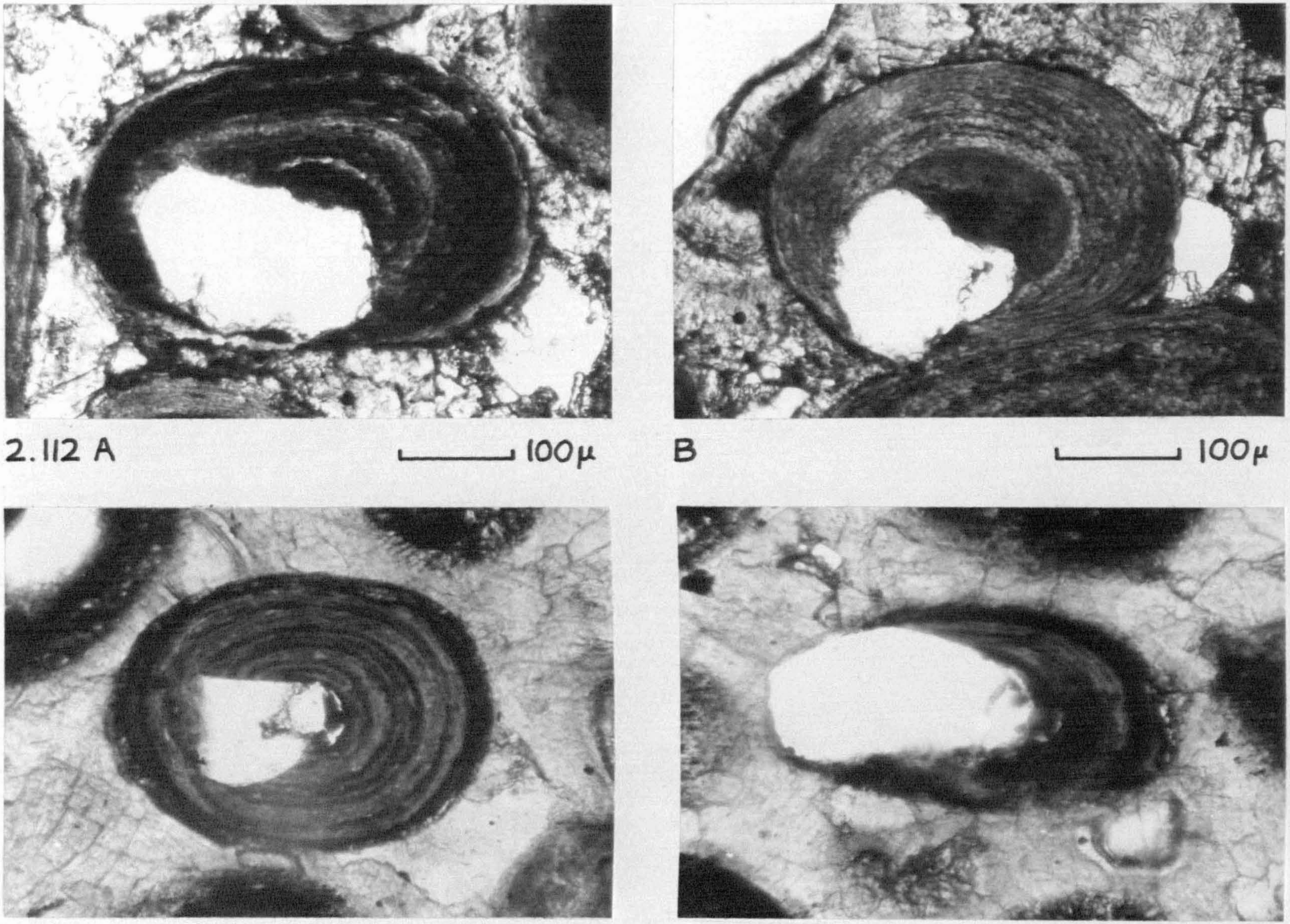
The normal concentric ooliths are accompanied in all samples by ooliths in which the envelope has developed eccentrically to the nucleus (for terminology see figure 2.98). These eccentric ooliths are also characterised by possessing discontinuous zones and laminae which terminate abruptly with strong angular discordance against the surface of the nucleus (fig. 2.112). This feature cannot be attributed to a sectioning effect on concentric ooliths.

The constitution of these ooliths in terms of mineralogy and of structure of the envelope is identical to that of the associated concentric ooliths. The distinguishing feature lies not in the relationship between oolitic laminae themselves but between the

FIG. 2.112 Eccentric ooliths

A.

A. Polar section	(x200)
B. Polar section	(x200)
C. Equatorial section	(x175)
D. Polar section showing one-sided deposition	(x150)



2.112 A 100μ B 100μ

C 100μ D 100μ

laminae and the nucleus. This relationship is described in detail below; the terms 'early' and 'late' are used to aid the description on the assumption that the innermost laminae were laid down first. The limits of the first oolitic zone along the periphery of the nucleus are sharply defined, and concave surfaces lying outside these limits remain uncovered (fig. 2.113a). This contrasts strongly with the initial layer in concentric ooliths (fig.2.113b) which tends to fill in depressions anywhere on the surface of the nucleus and to be thin or absent over all projections. A further contrasting feature in eccentric ooliths is the tendency for reduction in sphericity resulting from a strong outward growth component. The description of subsequent growth stages will be aided by figure 114 in which ooliths observed in thin section are represented and placed in approximate order of maturity.

The second zone is superimposed concentrically on the first, and its outer surface extends along the periphery of the nucleus by an amount depending on the thickness of the zone (figs. 2.114 a - f). Preferential equatorial deposition is usually developed and the equatorial deposition is usually developed and the equatorial plane of the oolith is therefore established at this stage. The oolith increases in size by superimposition of successive zones, with progressive reduction in the area of protrusion of the nucleus (fig.2.114 g - l). In some ooliths the accretion process has stopped before the nucleus has become entirely covered while in others, when a point has been reached where only a small part of the nucleus remains exposed, continuous zones have been laid down of a type indistinguishable

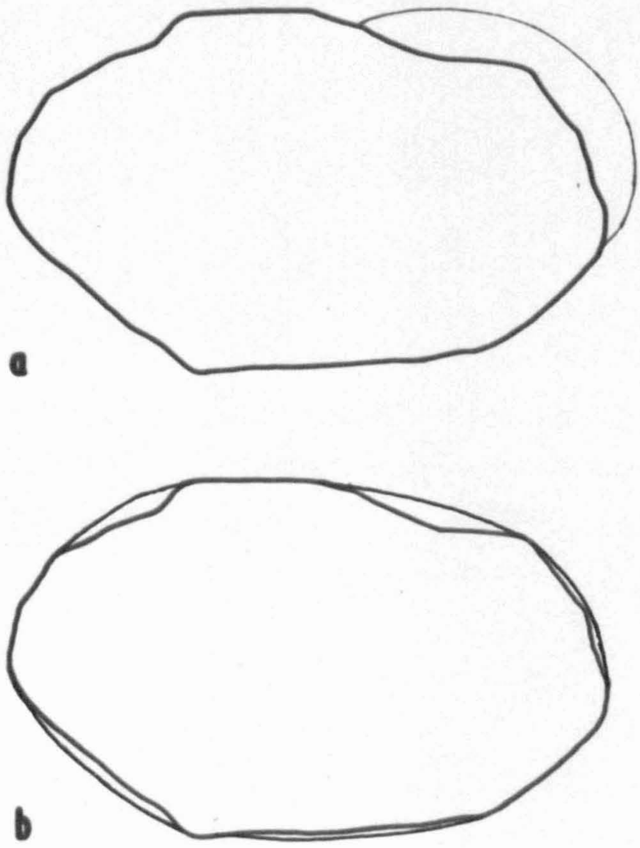


Fig. 2.113 Relationship of first lamina to nucleus in a) eccentric oolith b) concentric oolith.

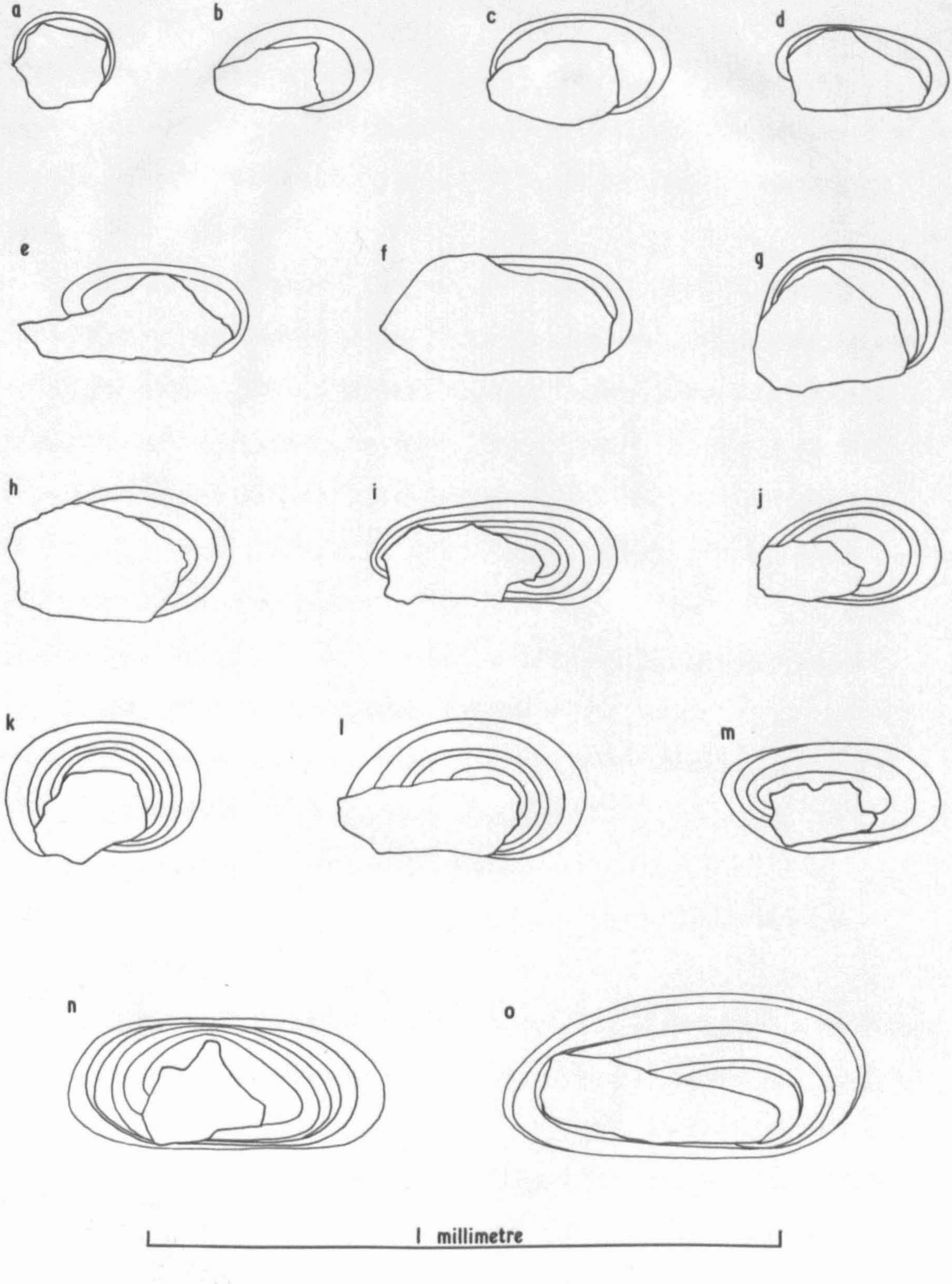


Fig. 2.114 Zone distribution in observed eccentric ooliths in approximate order of maturity.

from those of concentric ooliths (fig.2.114 m - o). The initial continuous zone thickens into depressions and thins over projections on the surface of the eccentric oolith as if the latter were acting as a normal nucleus.

The relationship of zoning to the eccentric nucleus as seen in section varies considerably, the principal forms being represented in figure 2.115; all gradations are found between these end types. Types (i) and (ii) are rare, type (iii) uncommon and only type (iv) abundant. Types (iii) and (iv) are characterised by the displacement of the nucleus from both axes; types (i) and (ii) probably represent equatorial sections of such ooliths. Type (iii) sections differ from type (iv) only in having a more extreme displacement of the nucleus from the minor axis, and the nuclei concerned are usually elongate. It is probable, therefore, that these eccentric ooliths all possess nuclei which are displaced from the centre along both the minor axis and the equatorial plane, with protrusion of the nucleus through the lower polar surface and occasionally through one side of the equatorial surface.

Sections of type (iii) and (iv) show several features in addition to those already described. Early accretion is confined to a relatively small area on the surface of the nucleus, and in relation to the fully developed envelope this area generally corresponds to the junction of the upper polar surface with one side of the equatorial surface. With further deposition the oolith expands upwards and outwards and each successive zone transgresses farther along both

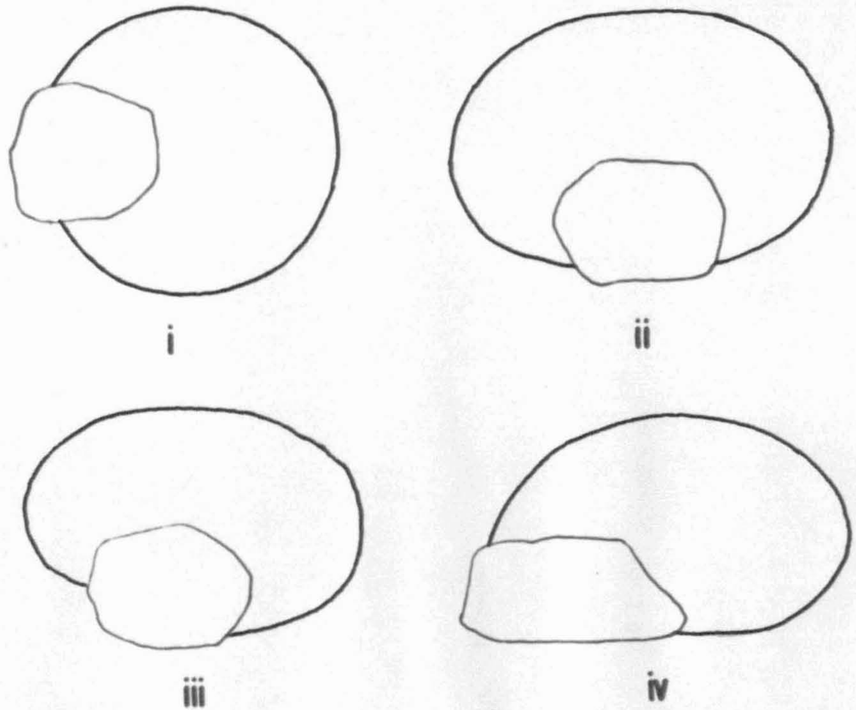


Fig. 2.115 Principal types of apparent envelope to nucleus relationship in eccentric ooliths.

the upper polar surface and equatorial surface of the nucleus. In all of these early stages the zones abut against the nucleus at a high angle. In type (iii) sections deposition is more or less limited to the upper polar surface and one side of the equatorial surface. In type (iv) sections transgression continues along the polar surface and on to the equatorial surface on the side opposite than on which accretion was initiated. On the latter side the zones may pass under the nucleus, where they abut against the surface at a low angle (fig. 2.114 l) a situation also found in some sections of type (iii).

The deposition of discontinuous zones may have ceased at any stage during the transgression of the surface of the nucleus, but there is a fairly definite limit of maximum transpression. This limit is reached when the zones have covered almost the entire equatorial surface (i.e. both sides as seen in section) and part of the lower surface of the nucleus. Zones laid down after this limit has been reached are of concentric type (fig. 2.114 m - o). The asymmetry of the oolitic envelope relative to the nucleus is due mainly to variation in the number of laminae present on either side of the nucleus, but also results from preferential equatorial thickening of the bands on the side of initial deposition. In early stages of accretion the upward growth component may equal the lateral component, but with increases in size there is generally a decrease in the amount of upward growth relative to lateral growth. Occasionally deposition is absent on the upper polar surface in the latest stages. The normal concentric zones which are frequently superimposed on eccentric ooliths generally further increase the elongation of the whole oolith by thick equatorial and thin polar deposition. Polar deposition may even be lacking.

G. ORIGIN OF ECCENTRIC OOLITHS

The eccentricity of the envelope relative to the nucleus suggests a process of formation which differs at least in some respects from that of the associated ooliths with central nuclei. The processes by which initial deposition could be restricted in areal extent may be placed into two main categories;

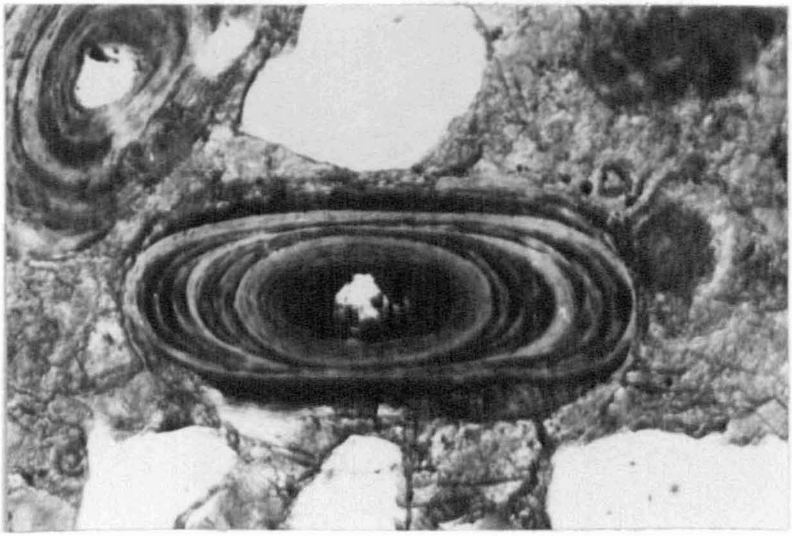
1) The entire surface of the nucleus is available for deposition, but some control is operating which restricts the area over which deposition occurs. Since the nuclei appear to differ in no way from those of normal ooliths, this control is unlikely to be related to surface features of the nucleus. The probability that an external control has had such an influence on accretion implies that the nucleus must have been immobile.

2) Accretion is inhibited over that part of the nucleus which does not receive oolitic deposition.

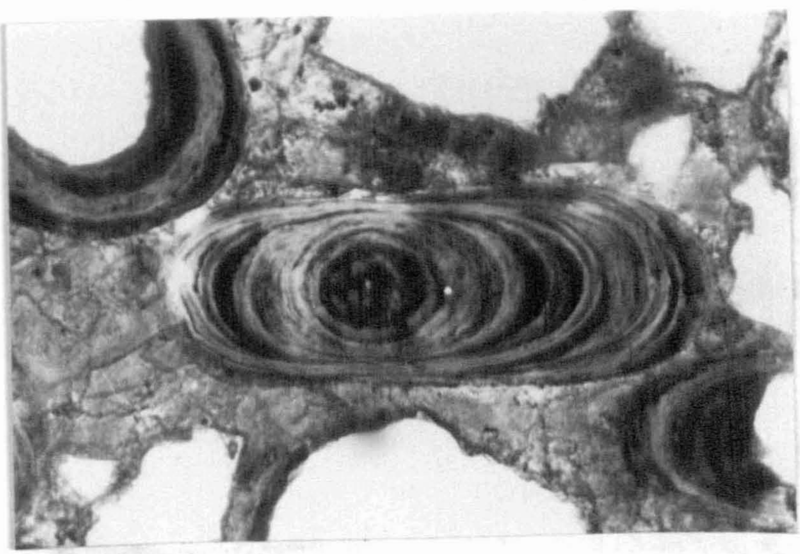
A process of the first type may have been responsible for the development of a few ooliths which show one-sided equatorial deposition (fig.2.117). In these rare cases the strong angular discordance between zones and nucleus is not always observed. The true eccentric ooliths, however, appear to have been produced by a process of the second type. The sharp turning in of the discontinuous zones along their outer margin can only be attributed to the presence

FIG. 2.116 Concentric oolith showing small-scale discontinuities of laminae (x140)

FIG. 2.117 Oolith showing late one-sided deposition on earlier concentric oolith (x140)



2.116 100µ



2.117 100µ

of a well defined barrier limiting the extent of the zones. Furthermore the progressive overlap of successive zones against the nucleus indicates a steady reduction in the area of inhibited deposition. The only possible inhibiting medium which might show these features would be fine-grained sediment in which the nucleus is partially buried. The barrier to oolitic accretion presented by this sediment could be steadily lowered by current winnowing so that each oolitic zone would extend further over the nucleus than the preceding zone (fig.211B). The fine-grained sediment in which the nucleus is buried would presumably consist of chamosite mud. The high angle at which the oolitic zones abut against the nucleus suggests that the outer surface of each zone is turned in sharply at the base where it has been in contact with the sediment. The lower polar surface of the zones would be expected to parallel the sediment surface, and since the equatorial plane parallels the lower surface, the ooliths are thought to have developed with their equators horizontal.

If this is the case the contrast in downward extent of the zones on opposite sides of the equatorial surface must reflect a corresponding variation in the sediment level on opposite sides of the oolith. This variation may be explained by the winnowing hypothesis, since currents are known to cause preferential scouring around obstacles. Where the obstacle is a pebble on a sandy beach the scouring is greatest on the downcurrent side (Twenhofel, 1932; p.668), but the beach slope, which permits gravitational movements of grains, and the partially subaerial conditions make this a specialised environment. Under more normal conditions of unidirectional

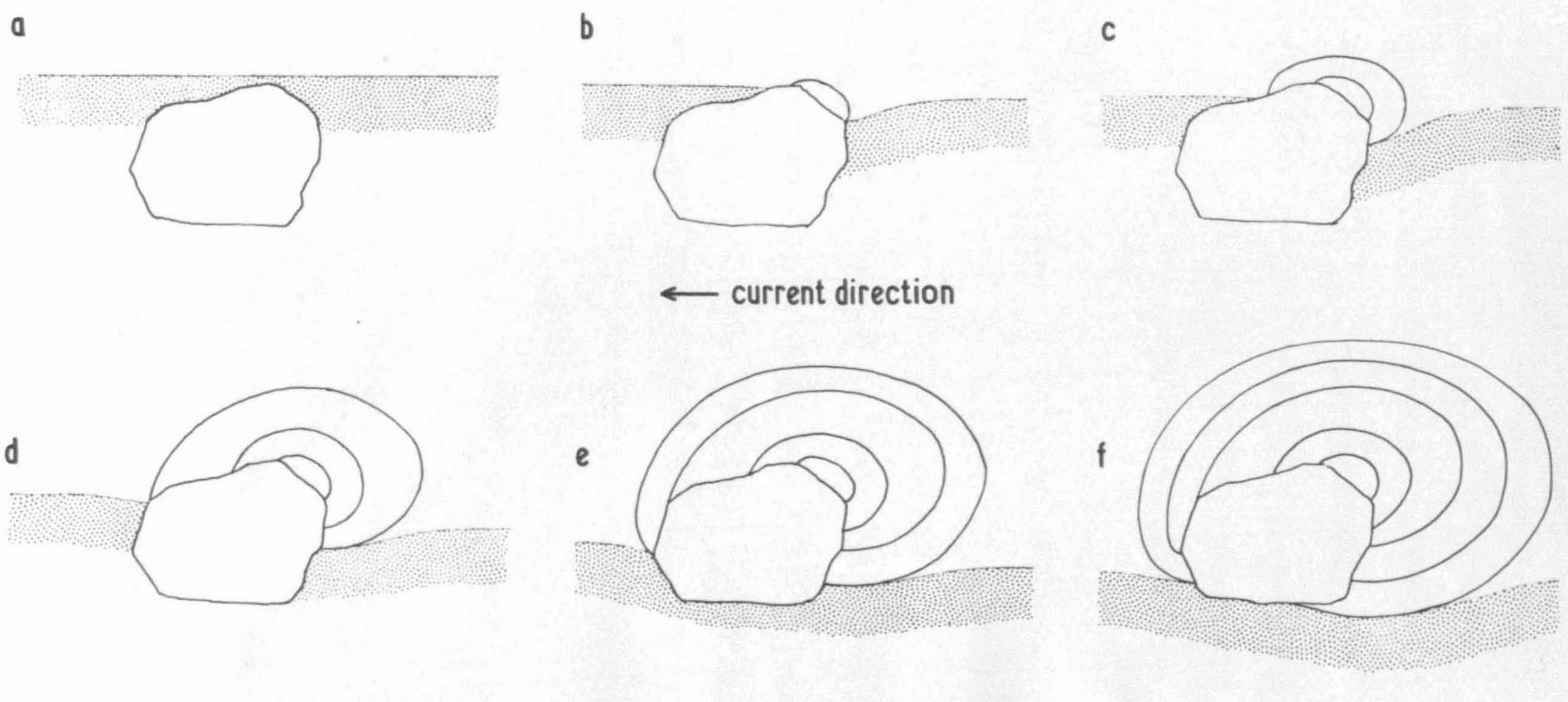


Fig. 2.118 Relationship of eccentric oolith to sediment surface at several stages of oolith formation.

current action along a relatively horizontal surface, excavation of the sediment is greatest on the upcurrent side of an obstacle, producing a crescentic depression or moat (Potter and Pettijohn 1963; p.121). On this basis the side on which the sediment, and therefore the oolitic deposition, is lower represents the upcurrent side. In many cases the zones which comprise the lower polar surface are convex downwards, which is probably a reflection of the cross-section of the upcurrent moat.

The encroachment of the base of the nucleus by late zones must represent the undermining of the nucleus, and the point at which the deposition of discontinuous zones ceases is interpreted as representing the dislodging of the grain from the mud. In many cases subsequent growth has occurred by deposition of normal concentric zones, to which the eccentric oolith has acted as a nucleus, with supression of the irregularities on its lower surface. The abrupt change from eccentric to concentric deposition bears no signs of abrasion, breakage, or change in mineralogy, and is therefore believed to have taken place with little or no break in time, and with no change in chemical conditions.

The current which is responsible for the winnowing of mud around growing ooliths is also apparently a controlling factor in the deposition of the oolitic zones. In most of the observed ooliths the side which has been interpreted here as being upcurrent is also the side on which equatorial thickening of individual zones is most apparent. Occasionally the ooliths reach a stage where the zones do not pass over the upper polar surface, in which case deposition is entirely confined to the upcurrent side. Non-deposition on the polar surfaces is a common feature in the later stages of concentric ooliths.

One-sided equatorial deposition has already been described from a few rare examples, and can presumably also be attributed to current influence. The only difference in these cases is that the nucleus appears never to have been lodged in the sediment, but must have rested on the sediment under the influence of currents insufficiently strong to cause movement of the grain. The length of the polar axis may be the critical factor in producing one-sided deposition. In appears that when this diameter reaches a certain value deposition no longer takes place on either the upper polar surface or the downcurrent side of the equatorial surface, the polar diameter of the concentric oolith which has acted as a nucleus to one sided deposition (fig.2.116) has apparently exceeded this limit.

The possible ways in which current could influence oolitic accretion will now be considered. Deposition of oolitic material may result from a physical, physico-chemical, biochemical or biophysical process, or from a combination of two or more of these.

(i) Physical deposition would be produced by particles carried in suspension adhering to the oolith when brought into contact with it by the current. If such a process were dominant, maximum deposition would be expected to occur immediately above the sediment, where the suspension is most concentrated.

(ii) Physico-chemical deposition would comprise direct precipitation of chamosite from solution. If precipitation were current influenced, maximum accretion would be expected on the upcurrent side of the equatorial surface, and if it were not current influenced an even covering would be expected to form over the exposed surface. In fact, most zones exhibit equatorial thickening on the upcurrent side but at least in the early stages, a strong upward component of growth is also present, although the sediment level is being progressively lowered.

(iii) Biochemical precipitation might result, directly or indirectly, from the activity of organisms such as bacteria or algae. In this case the thickness of each layer should be proportional to the rate of growth of the organisms. The growth rate might be expected to be highest where the current impinges on the oolith since this will be the region of maximum nutritional supply. Other factors such as light, current drag immediately above the sediment, and concentration of suspended sediment might favour an upward component of growth.

(iv) Biophysical trapping of suspended material by organisms would result in maximum accretion where organic growth rate is at a maximum and where there is a sufficient supply of suspended material.

It is probable that no one of these processes is alone responsible for oolith formation. The occurrence of laminae of oriented and disoriented chamosite suggests that either two mechanisms are involved or one mechanism is operating under varying conditions. The upward growth component, common in the early stages of formation, indicates an attempt to reach an optimum level of accretion. The later tendency for polar deposition to become thinner or even absent must indicate the attainment of this level. This feature could be related to the growth of organisms, the level of water saturation, and the concentration of suspended material. Although the limit of upward accretion is variable it is in the order of 150 microns. Organic growth is not likely to be so limited, and such a fine demarcation of chemical saturation is improbable, except possibly in absolutely still water. Conditions of steady gentle winnowing currents, however, might well produce a suspension of very limited height above the sediment surface. If suspended material is an important factor in the development of these ooliths the length of the polar diameter could be limited in this way. Physical accretion alone cannot be responsible for the development of the oolitic envelopes, however, since the level of maximum accretion would be expected to become lower as the sediment is winnowed out. Thus while the upper limit of accretion may depend on the maximum height of suspension, the level of maximum accretion is probably under some organic control.

From all the above observations the following deductions may be made;

(i) The ooliths developed around stationary nuclei;

(ii) The nuclei were partially embedded in mud during the development of the oolith, although occasionally the nuclei appear to have rested on the sediment surface, with the development of one-sided deposition;

(iii) The oolitic layers were deposited under conditions of gentle current action, strong enough to winnow sediment from around the developing oolith. By increasing the exposed surface area of the nucleus available for deposition, the winnowing results in progressive overlap of the oolitic layers. The fact that successive zones always show overlap suggests that winnowing of the sediment may be an essential factor in the formation of these ooliths, especially since accretion appears to result from the trapping of suspended material which could be derived from such winnowing.

(iv) Initial deposition and maximum deposition take place on the upcurrent side of the nucleus.

(v) Accretion is strongly influenced by currents, probably through some organic agency. Non-deposition on the upper polar surface in later stages reflects a limit to upward growth which may be related to the maximum height of suspension above the sediment surface.

(vi) A change from discontinuous to normal continuous laminae marks the dislodging of the eccentric oolith as a result of winnowing.

(vii) The delicate lamination and smoothness of outline exhibited by the envelope of eccentric ooliths is a primary feature of accretion, not modified in any way by rolling or abrasion.

These conclusions as to the mode of formation of eccentric ooliths correspond closely with observations on quiet water calcareous ooliths forming at the present time (Freeman, 1962). Some of these calcareous ooliths show a striking resemblance to the chamosite ooliths described above. They are forming in a muddy environment which is heavily colonised by marine grass, and in which water flow is insufficient to cause rolling of grains.

Freeman attributes the protrusion of nuclei to stationary growth on a muddy floor, and observes that laminae of tangentially oriented and disoriented crystallites are developed, neither of which can be related to rolling or abrasion. Organic material separated from the ooliths suggests that growth is somehow connected with organisms. Some of the ooliths show smooth outlines and a tendency towards flattening, while others have a more irregular shape but are more spherical.

Apart from mineralogy, therefore, the quiet water calcareous ooliths and the eccentric chamosite ooliths differ only in shape. Since current appears to be a controlling factor in the development of shape of the chamosite ooliths, the more symmetrical distribution of 'equatorial' deposition in the calcareous ooliths may be related to a more variable current direction, perhaps under the influence of tides. In general, however, the nature of the physical environmental conditions under which these calcareous ooliths are forming lends support to the proposed conditions of formation of eccentric chamosite ooliths.

H. COMPARISON OF ECCENTRIC OOLITHS WITH CONCENTRIC OOLITHS

The constitution of eccentric ooliths differs in no way from that of concentric ooliths, except in the distribution of the zones relative to the nucleus. The nature of the oolitic zones is similar in both types as is the distribution of zones relative to oneanother. Polar deposition is relatively heavy on the upper surface of the eccentric ooliths while in the concentric ooliths polar deposition is thinner but is developed on both sides. The total length of the minor axis is similar in both types. In early zones of concentric ooliths deposition almost equals the equatorial deposition, but it becomes progressively less and is commonly absent in later zones. An optimum level of accretion is again implied and seems to differ little from that of the eccentric ooliths. Nuclei whose diameter normal to the equatorial plane equals or exceeds the average polar axis length suffer little or no polar deposition. Fluctuations in this level are indicated in some ooliths in which phases of entirely equatorial deposition are followed by renewed polar deposition.

The few ooliths with protruding nuclei which have been separated from the oolite have the same discoidal form as the eccentric ooliths. In thin section they are on average smaller and more spherical than the associated concentric ooliths. This is believed to be a reflection of the mode of formation, in which the eccentric ooliths become dislodged before they are fully grown, and therefore only those which receive subsequent concentric deposition are likely to attain the average oolith size.

Many of what appear superficially to be normal concentric ooliths are seen at high magnification to possess laminae or zones which are discontinuous over one or other of the polar surfaces and which vary in thickness around the equator (fig. 2.116). Normally these variations are on a small scale and are of random distribution;

this results in an overall symmetry. Each polar discontinuity is interpreted as representing a phase of growth in which the oolith was not overturned by the current, though horizontal movement may have occurred. Variation in equatorial thickness indicates truly stationaly phases in which the prevailing current could affect the direction of growth of the envelope; in a few cases such conditions have been sufficiently prolonged to produce large-scale eccentricity in the envelope (fig. 2.117). When, during a stationary phase an oolith comes to rest at an angle to the sediment surface a new eqatorial plane will be established (fig.2.102). This angular shift of the equator is seen only in early stages when presumably the relatively high sphericity of the ooliths permits an oolith to remain stable even when in a tilted position. It is never seen in later stages, which suggests that it results purely from deposition and not by interference of other grains on the sediment surface.

These features all indicate that the concentric ooliths grew in much the same way as the eccentric ooliths, and in similar quiet water conditions. The anchorage afforded by the nuclei of the eccentric ooliths is sufficient to prevent the frequent turning over experienced by the concentric ooliths. Once dislodged from the sediment the eccentric ooliths act as nuclei for concentric deposition with no signs of a change in conditions. A feature which indicates that slightly more agitated conditions may have occurred towards the end of the period of oolith formation is the presence of an outer zone of relatively thick polar deposition in ooliths from the top few inches of the ironstone. The corresponding equatorial deposition is thin and this suggests that current conditions may have become sufficiently agitated to keep the ooliths in continuous motion for much of the time. The same feature is less well developed at lower levels within the ironstone and may represent accretion under relatively agitated conditions as the ooliths neared the outer limit of oolith formation.

The presence at all levels within the ironstone of both concentric and eccentric ooliths in various stages of growth suggests that the two types were developing together. The ooliths appear to have formed during, and possibly as a result of, winnowing of chamosite mud under gentle currents, with all exposed or partially exposed grains acting as nuclei. In both cases the equatorial plane appears to have developed parallel to the sediment surface and since except in a few cases the plane remains constant throughout it is probable that it was a mud surface and not a surface of previously deposited ooliths. In the case of a eccentric ooliths a mud surface was certainly present while for concentric ooliths to develop on top of other ooliths seems unlikely for the following reasons:

(i) Potential nuclei would tend to lodge in interstices between the ooliths and be unable to move around during accretion.

(ii) In more fully developed ooliths tilting, reflected by a shifting equatorial plane, should be common.

(iii) Stationary growth stages should reflect mutual interference by adjacent ooliths.

The ooliths are visualised here as forming under quiet conditions on a muddy floor and sufficiently scattered to prevent mutual interference in growth. Although the structure of the oolitic envelope has not been studied in sufficient detail for the exact method of accretion to be assessed suspended mud and organisms are thought to have been involved. The formation of both types of oolith may therefore depend on the existence of a chamosite mud and its subsequent winnowing. This is supported by the presence of abundant interstitial mud, subsequently replaced by siderite, and by the presence of nuclei which are almost certainly the product of authigenic recrystallisation within a chamosite mud.

I. ENVIRONMENTAL SETTING OF OOLITH FORMATION

If chamosite oolith formation depends on the winnowing of a chamosite mud surface it is clear unless some other factor is involved the surface will soon become covered with ooliths. Since one-sided deposition is displayed in the stationary eccentric ooliths the currents must have been predominantly unidirectional; under such conditions the free concentric ooliths which undergo periodic overturning will migrate away from the site of formation. Eventually the ooliths will accumulate, while more ooliths are forming in an upcurrent direction. In this way, provided that conditions remain stable, ooliths will be continuously produced and deposited, and the nature of the oolith assemblage at the site of deposition will remain fairly constant.

J. ROCKS ASSOCIATED WITH THE WINTERGILL IRONSTONE

In the Winter Gill outcrop a pale gray sideritic sandstone marks the base of the ferruginous beds. The sand grains, though in grain contact, are rather loosely packed but patchily distributed grain clustering has resulted in the local development of sutured contacts. A few chamosite ooliths are present but are highly distorted; they are brown in colour and show little replacement. Shell fragments show sideritisation along laminae and any material not replaced has undergone solution. Siderite occurs as a rim cement on all shell material and intraclasts. The rim consists of fine radially oriented fibrous crystals while tangential colour zoning is present which consists of an inner dark brown zone, a middle colourless zone and an outer brown zone. Post-depositional shell breakage also affects the rim. In the remainder of the rock the siderite occurs as small spherulitic bodies about 150 microns in diameter and showing zoning similar to that of the rims; at the centre of each spherulite is a distinct opaque zone about 75 microns in diameter.

The oolitic ironstone is overlain by a series of thin beds which show a gradation upwards into sandstone. The lowest bed is oolitic with ooliths locally abundant in some lenses; there is however, a sharp break at the top of the ironstone as far as the sand grain to oolith ratio is concerned. Interstitial chamosite mud is present which shows partial replacement by reddish-brown siderite which occurs as rims around all grains. Occasionally large prismatic crystals extend from the rim into the interstices. Where mud is absent clear siderite crystals line the cavities. At higher levels ooliths are rare and show exteme distortion due to penetration by sand grains.

Sideritic oolitic sandstone is found in an outcrop further north and this contains a matrix of siderite crystals up to 200 microns in length which show spectral extinction and fair colour zoning. Ooliths show total replacement.

XII. ORIGIN

A. ORIGINAL NATURE OF SEDIMENT

1. Matrix

The matrix of the true ironstone is entirely sideritic except in rare patches where several ooliths have recrystallised to structureless chamosite, and when the interstitial matrix is also chamositic. At the base of the overlying beds chamosite mud is again present as a matrix though partially replaced by siderite along grain boundaries. It seems probable, therefore, that as in the basal ironstone (see p. 118) almost total sideritisation of an original chamosite mud matrix has taken place. While most of the siderite would appear to represent an original mud matrix in a few cases it is apparently a drusy cement, since a few samples show characters which are distinct from those of the bulk of the samples:

(i) In apperently muddy rocks inclusions are abundant along intercrystalline boundaries and are an important feature in the zonation of radial crystal aggregates; where ooliths form at least 90 percent of the grains the matrix constitutes between 40 and 45 percent of the rock; included ooliths always show marginal replacement with or without internal replacement; phosphate and pyrite occur in small amounts.

(ii) in apparently mud-free rocks siderite is free from inclusions except where traces of earlier crystals are present along grain surfaces; the matrix constitutes between 35 and 40 percent of the rock; included ooliths never show marginal replacement, but internal replacement is always developed and is frequently total; feldspar may occur as a replacive mineral while phosphate and pyrite are absent; drusy cavities may occur.

2. Grains

The proportion of sand to ooliths varies from sample to sample; the sand usually constitutes less than 10 percent of the grains but becomes dominant in the outer parts of the ironstone which represent passage into sandstone. The size of the sand grains relative to that of the associated ooliths also varies considerably (figs.2.92); this is mostly due to variation in sand grain size, but the average oolith size also varies.

The general lack of relationship between the sand grain and oolith sizes suggests separate sources for the two grain types with deposition under conditions which were not sufficiently agitated to produce better sorting. In the sandy laminated samples sorting is more evident within laminae, but the bulk rock remains poorly sorted. The only exceptions are the two samples from the top of the ironstone outcrop (figs.2.92 g,h); in these the two modes show roughly equal spacing which corresponds with that of the modes for a sample of sideritic oolitic sandstone from New Wath Scar. Since the top of the ironstone was apparently mud free and the ooliths were well packed it is probable that current sorting will have taken place, and that the observed sand - oolith size relationship is one of hydrodynamic significance, and this is certainly the case in the sandstone in which both sand grains and ooliths have undergone transport. The general lack of relationship indicates that the present outcrop is close to the site of oolith production.

The average modal size of the sand grains compares closely with that of sand grains from the upper sandstone division in adjacent areas but there is a relatively high proportion of coarser grains, apparently reflecting concentration due to slow accumulation and perhaps selective solution of smaller grains. The average grain size for sand grains acting as nuclei is slightly less than that for the free grains, indicating that only the finer grains were introduced into the area of oolith formation. The modal size of ooliths varies to some extent, the highest values being found in the samples from outcrop with the exception of the top few inches, where the modal size falls off.

In all of the samples studied the sand grains were smaller than the associated ooliths, even where hydrodynamic equivalence has been reached. Since chamosite has a higher specific gravity (3-3.5) than quartz it might be expected that ooliths should be much smaller than the sand grains, especially since they possess low sphericity. This distribution may reflect the scarcity of larger sand grains and smaller ooliths, but it is probable that at the time of transport the density of the ooliths was much reduced by a high content of water or organic material. In contrast Hallimond (1925 p.26) states that ooliths are always smaller than the accompanying sand grains.

B. CURRENT CONDITIONS

The conditions of formation of ooliths have been discussed earlier with the conclusion that they form under gentle, persistent current conditions, and that these currents carry the ooliths to the site of deposition. The muddy nature of the ironstone at the present outcrop suggests deposition under relatively quiet conditions, though intermittent phases of stronger agitation must have produced the cross-bedding seen in a few samples. Since ooliths are found to the north, east, and south of Winter Gill it would appear that the prevailing current was from a westerly direction.

C. ENVIRONMENTAL SETTING

The most significant feature of the Winter Gill ironstone is that it constitutes a body of muddy and almost sand-free sediment which is incorporated within a unit that is otherwise entirely sandy. There are two possible relationships which could account for this situation;

(i) a small area of the sediment surface was in some way kept free of terrigenous sediment while sand was being deposited elsewhere.

(ii) the ironstone has been deposited in a channel as proposed for some of the oolitic ironstone in Rosedale and Bilsdale (Rastall and Hemingway; 1949, p276).

The mechanism proposed here for the formation of ooliths could have proceeded in either case, but there are several features which do not favour accumulation in a late channel. In particular the lateral passage of the ironstone into oolitic sideritic sandstone indicates that the two facies are of the same age. Also the general lack of good grain packing and the lack of any size relationship between sand grains and ooliths distinguishes this ironstone from those which are known to occur in channels.

The nature of the localised muddy environment can only be a matter for conjecture but it is significant that even at the present outcrop, which is not thought to represent the original site of oolith formation, the nine feet of oolitic sideritic sandstone found to the east are represented by 3ft. 9in of ironstone. If one particular site was supplying mud and ooliths throughout this time then this interval should either be represented by a thin horizon of mudstone or not be represented at all. The only way in which this situation could occur would be if the ooliths were forming on substantial bank, and the unusual thickness of the underlying beds suggests that this may have been the case. In comparison with the type area the shale unit and the lower division of the sandstone unit have thickened from between 9 and 10 feet to nearly 19 feet. This is due to the sandstone body forming the upper part of the shale unit and to the thick unlaminated sandstone of the lower sandstone division. The immediate cause of the bank could have been the thickening of the lower sandstone division but this may have developed as a result of the underlying sandstone having acted as a sediment trap during the strong relative compaction effects which must have occurred against the surrounding shale. Alternatively the same condition that produced the first sand body may have been re-initiated in the same locality, but in view of the contrast between the sandstones this seems to be the less likely process.

At the site of oolith formation the thickening of underlying beds must have been even more pronounced. If the concept of ooliths forming from winnowing of mud is correct then a phase of mud deposition must have occurred between the two sandstone divisions. The very large quantity of mud involved suggests one of three possibilities;

(i) large source area

(ii) thick original deposit of mud

(iii) alternation of phases of mud deposition with phases of winnowing and oolith formation.

The evidence is insufficient to favour any one of these, but it is unlikely that a large source area was involved.

The relatively fine-grained nature of the beds overlying the ironstone at the outcrop suggests that cessation of oolith formation was not due to flooding of the source area by sand. It may have been brought about by change in chemical conditions or by the exhaustion of the supply of the necessary materials. The clean-washed nature of the top of the ironstone together with the partial oxidation and extensive replacement of grains suggests that a pause in deposition may have occurred at this level.

The nature of these deposits indicates that they were not under the influence of wave action and were not exposed at low tide levels. Since they are near the top of the marine bed, however, the depth of water at the later stages cannot have been great.

Ferruginous and oolitic developments are found locally to the south and east of Winter Gill and to a lesser extent to the north.

The prevailing currents which are thought to have produced one-sided deposition in ooliths were therefore probably from the west. Although exposures are poor it would appear that a second area of oolith and chamosite mud formation existed in the region of Danbydale. In this case only slightly oolitic but strongly sideritic sandstone is exposed but a thin parting of practically sand-free oolitic ironstone has been observed at one locality. It is significant that this ferruginous facies also occurs above an area of sandstone development within the shale unit. The lower sandstone division is not seen to be affected by this early sandstone which in Great Fryupdale had ceased before the end of shale deposition. It is possible that elsewhere in this region the sand accumulation continued until the end of the shale unit and that a similar situation to that of Winter Gill might exist.

IV. DIAGENESIS

As previously discussed the original matrix of the ironstone, where present, was believed to have been chamositic. There is little evidence by which the sideritisation could be dated but lack of any compactional features and comparison with the siderite mudstone of the ironstone unit suggests that replacement was early. The sideritised matrix differs from that of the ironstone unit in large fibrous crystal aggregates set in a relatively finely crystalline groundmass. These large crystals could have originated in two ways; either by direct replacement of chamosite representing an initial phase, or by recrystallisation of siderite. The fact that variation of a very local nature occurs in both the zonation and the form of these aggregates implies that this is a feature of early diagenesis. There is, however, a close similarity between the crystal habit of siderite cement and the siderite matrix where both are seen in the same sample. This may indicate that the matrix has undergone partial recrystallisation during crystallisation of the spar, but it is equally possible that the crystal habit of the matrix has influenced that of the later spar.

Diagenetic changes in the grain types mostly involves replacement, but at a very early stage a few ooliths or groups of ooliths have undergone recrystallisation to structureless finely crystalline chamosite and this alteration is clearly earlier than sideritisation since a single siderite rind is present around each recrystallised mass. The development of siderite rinds around ooliths and of marginal replacement of sand grains is confined to those rocks believed to have been originally muddy and is totally absent in the clean washed beds. Marginal replacement is never interrupted by any other form of replacement and appears to be an early diagenetic feature probably closely associated with sideritisation of the matrix. Internal replacement by relatively coarsely crystalline siderite which may begin in the middle of the oolith is probably much later. The fact that internal crystals are sometimes in optical continuity with zoned fibrous aggregates in the matrix may further indicate that these aggregates represent recrystallisation of the siderite matrix during a second phase of sideritisation.

In originally mud free rocks the internal replacement alone is developed, which again may be in optical continuity with the cement. It appears that siderite was introduced at two stages:

(i) early, with replacement of chamosite mud by granular mosaic (and possibly by zoned aggregates, and marginal sideritisation of ooliths and sand grains.

(ii) late, with deposition of spar in interstices, internal replacement of ooliths and sand grains, and probably recrystallisation of the siderite matrix.

An early phase of spar deposition, represented by 'ghosts' of rims and euhedral crystals, occurred in some of the mud free ironstone and this may correspond in age to the sideritisation of chamosite mud.

Replacement by siderite is ubiquitous and it is therefore a convenient standard by which the age of other replacements may be judged. Phosphatisation occurs only in originally muddy rocks and varies to some extent from sample to sample. Two main categories may be clearly distinguished:

(i) grains phosphatised before deposition

(ii) grains phosphatised after deposition

The first category includes a few ooliths and probably shell material. The second, distinguished by its occurrence in patches within the rock and by the delicate preservation of structure, affects ooliths and sand grains. The source of the diagenetic phosphate is probably from solution but in some cases phosphatisation has affected ooliths which surround shell fragments which themselves show some signs of phosphatisation often superceded by siderite replacement. These shells may have acted as foci for phosphate deposition or may themselves have provided the phosphate. Siderite rinds are absent or thin around phosphatised ooliths and phosphatisation is therefore assumed to be of a similar age as marginal sideritisation or slightly earlier. The phosphate may in fact have been mobilised during sideritisation of the matrix.

Pyrite is chiefly associated with sand grains and is often associated with phosphatisation. There is no evidence which might permit the pyritisation to be dated. The only other diagenetic mineral is feldspar which is abundant in the uppermost few inches of the ironstone. This appears to be earlier than the associated replacive siderite where the feldspar laths are included within siderite crystals, while in some cases the two minerals are mutually intergrown. The feldspar which sometimes affects sand grains is assumed to be of a similar age. The feldspar is itself replaced in a few ooliths by fibrous chamosite.

Deformation of ooliths is principally associated with siderite replacement though occasional spastolithisation, which does not affect the siderite rind, is observed. Distortion of laminae occurs around internal replacement crystals and around the more irregular marginal replacements, and is believed to result directly from this replacement. The chemical conditions of deposition and diagenesis are considered in Chapter IX.

V DEPOSITIONAL ENVIRONMENT OF THE SANDSTONE UNIT

Several features of the Winter Gill Ironstone suggest that it accumulated under very gentle current conditions. That similar conditions accompanied deposition of the sandstone is indicated by the horizontal lamination and the presence over a considerable area of a muddy matrix. Ripple-marks are most common in the extreme south and west which indicates slightly more water movement, though probably in the form of oscillation. Cross-bedding is locally developed in the top-most beds, perhaps indicating wave action. The presence of burrowing forms indicates that sedimentation was moderately slow and bottom conditions stable.

Depth of water is not easy to estimate, but a rough maximum figure can be calculated from the maximum sandstone thickness. The sandstone unit represents the transition from shale to a subaerial colonised surface and since compaction in sand after deposition of the whole unit is not likely to be very great the maximum thickness of 18 feet must represent a water depth of less than 25 feet. The thickness variation in the sandstone unit has already been attributed to some extent to differential subsidence. The maximum thickness of sandstone is therefore probably related to maximum subsidence and the maximum depth of water might therefore have been less than 20 feet. Thin sandstone successions (e.g. Hartoft Beck) which cannot be related to early introduction of sand must reflect shallower water and little subsidence. Slight uplift may even have taken place locally. Thickness variation in the Winter Gill Ironstone indicates an upstanding bank of about 12 feet relif at its highest point; possibly two or three feet of sediment may exist above this giving a total thickness of about 15 feet. Since sediment in the bank should have been well compacted before ironstone formation then a fairly accurate maximum depth of about 15 feet is suggested during the deposition of the upper sandstone division.

The restricted fauna probably indicates slightly lowered salinity, and not rapid deposition. The Winter Gill ironstone which represents local slow deposition is totally devoid of fauna in contrast to the highly fossiliferous oolitic ironstones of the Cleveland Ironstone Series for example.

CHAPTER VIII STRATIGRAPHICAL CONCLUSIONS

The variation in thickness of a marine intercalation such as the Eller Beck Bed may be due to several factors:

(i) Irregularity of the underlying surface;

(ii) Slow transgression and for regression resulting in relatively rapid landward thinning;

(iii) Variation in the level of sedimentation in the topmost sediments, either primary or resulting from erosion;

(iv) Rapid facies change with differential compaction;

(v) Variation in tectonic movements.

The upper boundary of the Eller Beck Bed consists of a flat sandstone surface which has been colonised by plants over practically the whole area with the exception of the extreme south-west where a thin carbonaceous bed is, however, present. This, together with the even-bedded nature of the sandstone, suggests that relative to the entire transgression this upper surface represents only a small time interval. Similarly the uniformity of the basal beds and the fineness of grain in both the marine and non-marine strata suggests a small time interval. On this basis transgression and regression appear to have been rapid. This is supported by the fact that although thickness variation does occur, the thickness in the north-east differs little from that in the south-west despite the thickness variation within the individual units.

The nature of the upper and lower contacts also rules out items (i), (iii) and (iv) as causing major thickness variations though locally they may be important. The isopachyte map for the whole of the Eller

Beck Bed (fig. 2.119) clearly indicates that tectonic downwarping produced a well defined north-west to south-east trough which broadens and possibly bifurcates to the south-east.

In individual units thickness variation is due to several causes. In the ironstone unit it is measured only in inches and such a thin deposit is likely to be sensitive to irregularities in the surface of the underlying non-marine beds. The continuity of the ironstone indicates a virtual lack of any such irregularities. The isopachytes for the ironstone unit indicate a general north-west to south-east trend with clear evidence of thinning to the north. The distribution does not conform to that of the whole of the Eller Beck Bed in that the zone of maximum thickness is shifted further north. It does, however, appear that the unit thins out to the southwest and the overall picture is therefore similar. Differences may stem from the fact that the total thickness of sediment is greatly influenced by the clastic sediments of the sandstone unit, and the ironstone unit, being in part of chemical origin, might not be expected to show the same trend. In fact the ironstone unit is dominantly clastic in the region of maximum overall sedimentation and the relative thinness may be due to the fine sediment having been winnowed away. If, as interpreted on page 148, the siderite in the mudstone is the result of early pre-compactional diagenesis then the thickness may be exaggerated with respect to the clastic sediments. The problem still remains however that in Danbydale the ironstone is thin and relatively sand-free although lying along the trough of maximum overall thickness. In this case relative compaction of the underlying beds may be involved since the ironstone rests directly on a thick sandstone

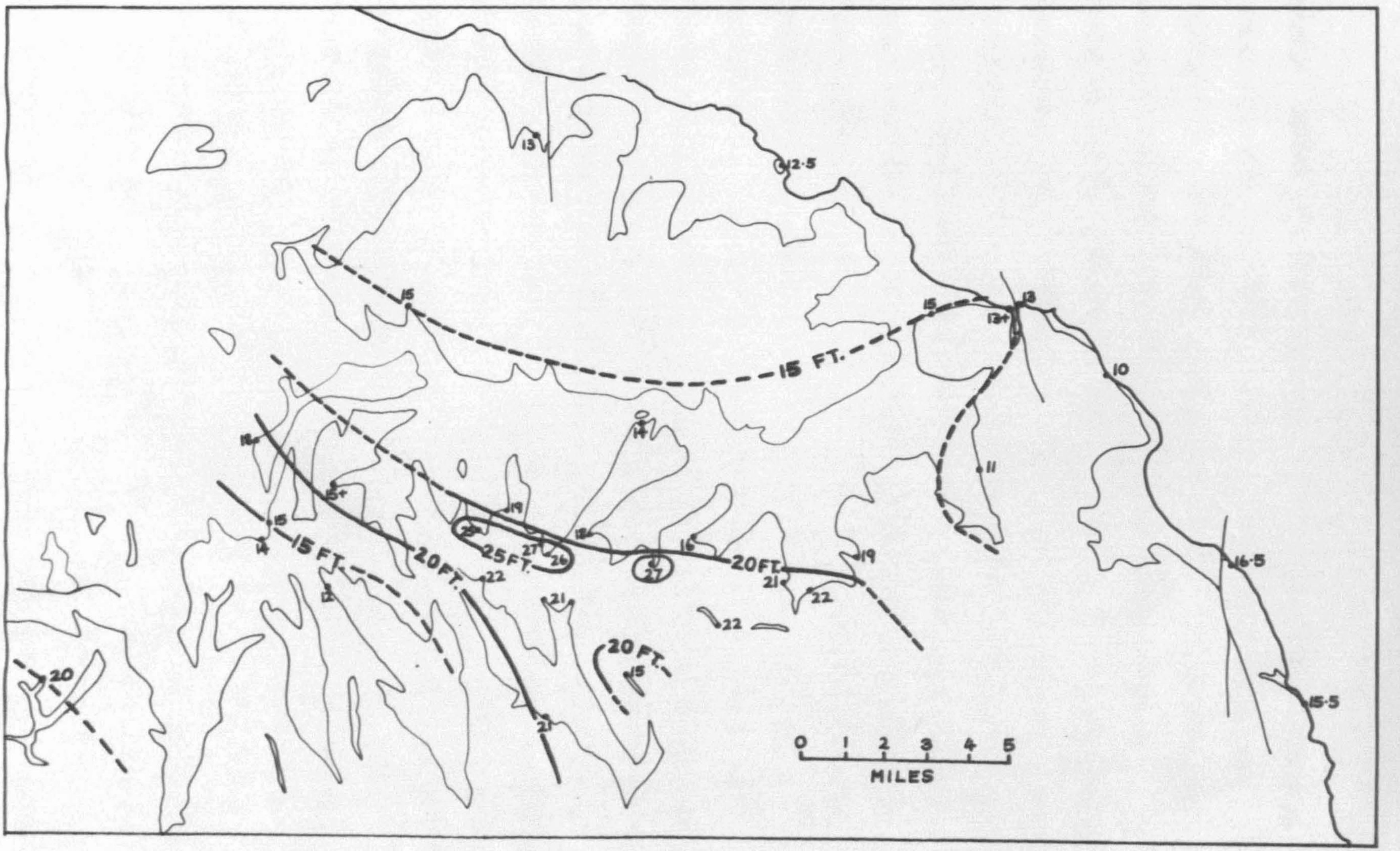


FIG. 2. 119 ISOPACHYTE MAP OF THE ELLER BECK BED

channel. Faunal elements follow the thickness trend quite closely. The burrowing organisms at the base conform to the thickest part of the ironstone unit except where they continue to the south-east. This suggests that the central area was under the influence of brackish or marine water while outlying areas remained unsuitable for colonisation. Marine conditions appear to have been introduced from the south-east probably because this was under the influence of both the general transgression from the south and the downwarping of the trough. The water at this time may have been very shallow or even contained only within the sediment. There is no evidence of marine sediments being laid down before colonisation. The overlying ironstone appears to have been deposited in shallow water and once again the central region corresponding to the axis of the trough, is distinctive in having the most abundant fauna; also the upper part is full of comminuted shell material which suggests that wave action has been active in this region. The lack of shell breakage in surrounding areas is due either to the sediment lying at two great a depth or to the water being too shallow to permit the formation of waves. In view of the previous conclusions the latter seems more probable. The ironstone unit seems therefore to have accumulated under extremely shallow water which with the possible exception of the lowest silty part was fully marine.

The boundary between the shale unit and the ironstone unit is in all cases sharp, sometimes erosional, and it is probable that this is not simply a feature of surface weathering of a gradational contact. Except where erosion of the ironstone has taken place the basal shale is fine-grained and often fossiliferous, the fossils being most abundant in the south and south-east where some forms indicate fully marine conditions. The sudden change from mudstone to shale must represent a deepening of the water and the sharpness of contact may represent an intermediate phase in which the previously deposited sediment was exposed to wave action. The shale would then have started to accumulate as soon as the water depth became sufficient to allow sediment to remain undisturbed.

From this time on the sequence appears to represent a gradual filling of the sea, and this is accompanied by increasing coarseness of sediment. The maximum thickness of the sandstone unit is about 20 feet and since the sand has been deposited relatively slowly with good sorting it is probable that little compaction has taken place. The maximum possible water depth at the end of shale deposition is therefore about 25feet but taking continued subsidence ito account the actual depth was probably less than this. The presence of oscillation ripple marks within the sandstone unit at all levels indicates relatively shallow water, possibly just under the influence of wave action.

The shale facies changes little but the development of sandstone bodies is significantly confined to the area of maximum thickness, suggesting that the sand had been transported along the slight topographic low produced by subsidence. The source of the sand would probably be the north-west. The sandstone is more variable but it is significant that along the line of maximum overall thickness the sandstone shows maximum thickness, greatest abundance of trace fossils and shelly fauna, and development of ferruginous and oolitic beds. This axis of sedimentation is also significant in that it separates off a northern and eastern province of relatively high sandstone - shale ratio from a southern province of low ratio. This is produced by both thickening of the shale and thinning of the sandstone. The axis of maximum subsidence appears to have acted as a clastic trap until a fairly late stage in the deposition of sandstone, since the sudden thinning of the unit between the Fryupdale region and the Farndale - Ingleby region cannot be otherwise accounted for. A steady regression of this type should otherwise produce a sandstone of equal thickness or one which thins regularly away from the source. From the Bluebell Trough section to Blow Gill there is no change in thickness and this area is presumably not under the influence of the trough. The upper part of the shale in this region is thought to be equivalent to the lower part of the sandstone to the north and the lack of gradation between the shale and sandstone is taken to indicate a rapid facies change once the sand managed to be transported beyond the trough.

In the exteme north of the area the sandstone, though probably deposited under the same conditions, underwent occasional phases of emergence and colonisation by plants. The sandstone is less well laminated than to the south, probably due to deposition in extremely shallow water, and it closely resembles the uppermost bed of the central areas which is also colonised by rootlets and represents the final filling in of the Eller Beck Bed sea. Both the shelly and the burrowing fauna are totally absent in the north.

The salinity of the water during deposition of the Eller Beck Bed can be inferred to some extent by the fossils. The ironstone seems to have been deposited under fully marine conditions and the same applies to the lower part of the shale in southern districts. The upper part of the shale is virtually unfossiliferous except for the nodular ironstone at Eller Beck in which fully marine forms are again present. The lower sandstone division contains a locally abundant fauna which is restricted in type while in the upper division the same fossils occur but are rare and are absent at the top. At Commondale, on the other hand, a fauna is found, in position of growth, which suggests fully marine conditions. It would appear that for the most part the shale and sandstone units were laid down under not quite fully marine conditions, which led either to the absence of a fauna or to the development of a restricted fauna. The two exceptions mentioned above are difficult to explain since they are so localised. The most probable factor involved in preventing the attainment of fully marine conditions is an influx of fresh water causing reduction in salinity.

The Eller Beck Bed therefore seems to represent a short lived marine phase in an area of predominatly deltaic sedimentation. The cause of this transgression will be discussed in Part III. The initial transgression clearly extended some distance beyond the northern and eastern boundaries of the present outcrop, although it is in these directions that land appears to have lain. Although the thickness and facies distribution is to some extent valuable in assessing the direction of the shoreline sedimentary structures can also provide information of this kind. In this case these structures are limited to oscillation ripple marks and wash-outs. The former are believed to have been formed by wave movement of the overlying water and might therefore be expected to reflect the distribution of water depth at the time of deposition. This relationship has been observed in recent sediments (Vause 1959 p.561) and in fossil sediments (e.g. Greiner 1962 p.232). In the recent sediments studied by Vause the oscillation ripples occurred at depths of between about 12 and 50 feet. These structures contrast with those of tidal flats (Van Straaten 1953, fig. 7; McKee 1957, fig. 28) which are always asymmentrical ripples which may or may not parallel the strand line. This lends further support to the hypothesis that deposition was not intertidal. Evidence for the trend of the strand line from channel directions is not very reliable; the exact course of these channels is seldom known owing to the lack of internal directional structures. The only certain direction is that of the channel in Reeking Gill, Rosedale, where very consistent small-scale trough cross-bedding indicates a south-south-westerly flow. The channels developed at various levels on the coast trend from between east-west and north-east-south-west and the flow is assumed to have been to the west and south.

The distribution of sedimentary structures is summarised in figure 2.77 and the consistent nature of the trends is believed to indicate that the shore-line lay to the north and east of the present area, though its exact position may have changed considerably during the entire marine interval. The nature of the shore line to the west cannot be determined; the sediments thicken to the south-west while the amount of transgression to the north-west probably depends on the influence of the trough.

The area studied does not extend into the region in which the

Hydraulic Limestone is developed but nodules and a bed of ferruginous argillacious limestone were observed in a section in Blow Gill less than a mile downstream from the measured section. The limestone is assoc with calcarious shale developed at the base of about nine feet of shale with occasional ironstone nodules. The bedded ironstone which reaches 3ft. 6in. thick upstream is not exposed but presumably lies below the calcarious horizon.

The Eller Beck Bed has been stated by Bate (1967, p.134) to belong to the zone of Hyperlioceras discites on the assumption of stratigraphic equivalence to the Hydraulic Limestone, and his proposal is fully accepted here. It is clear that the ironstone unit and possibly to a greater extent the shale and sandstone units extend some distance beyond the present outcrop. On this basis it is doubtful whether a palaeogeographical map such as that produced by Bate (1967; fig.4 p.136) could ever be significantly accurate and it is thought that the trend shown in the isopachyte maps already considered are of more value. The nature of the sediments does, however, suggest that deposition occurred in some form of lagoon rather than along a completely open shore and this would involve enclosure to the west, north, and east as proposed by Bate. The absence of the ironstone and the rapid thinning of the shale unit in the extreme north-east would suggest that the shoreli was closest to the present outcrop in this region.

P A R T T H R E E

G E N E R A L C O N C L U S I O N S

CHAPTER IX GENERAL CONCLUSIONS

I. GRAIN SIZE ANALYSIS OF SANDSTONES

A. RELATIONSHIP OF GRAIN SIZE DISTRIBUTION TO SEDIMENTARY STRUCTURE

In Parts I and II the cumulative curves obtained from the mechanical analysis of both Eller Beck Bed and Lower Deltaic sandstones were represented by the three parameters mean grain size, standard deviation, and skewness. There was little correlation between the parameter values and the everonment of deposition as assessed in the field. It was pointed out that in the cumulative curves studied the values of standard deviation and skewness depended on the position of the inflection point relative to the 84th percentile and, to a lesser extent, to the slope of the fine tail.

It would seem therefore that the inflect percentage alone is of equal significance to skewness and standard deviation, though the slope of the fine tail should also be taken into account. A measure of the grain size and sorting of the log-normal section of the curve together with the inflect percentage and measures of the distribution of the fine material would provide an accurate method of recording every feature of the curve. In this study the log-normal curve has been separated and represented by M, ϕ and ,ϕ/and I% it remains to be seen to what extent they reflect the depositional environment. Figures 3.1a,b show the vertical variation of M,ϕ, ,ϕ and I% in the Eller Beck Bed sandstone sections A and B. There is no relationship to be seen within the sections, though as indicated on p. 210 there is a contrast between the sections. The inflect percentages are on average higher in samples B2-12 (88%) than in A4-8 (80%), and this

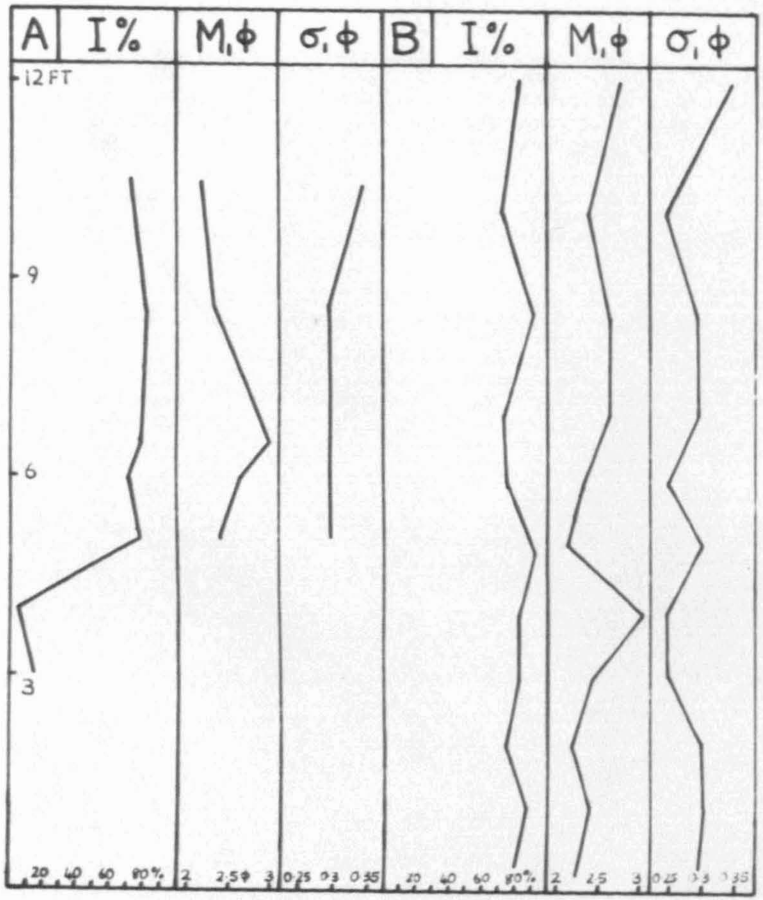


FIG. 3.1 VERTICAL VARIATION IN PARAMETER VALUES, ELLER BECK BED SANDSTONES

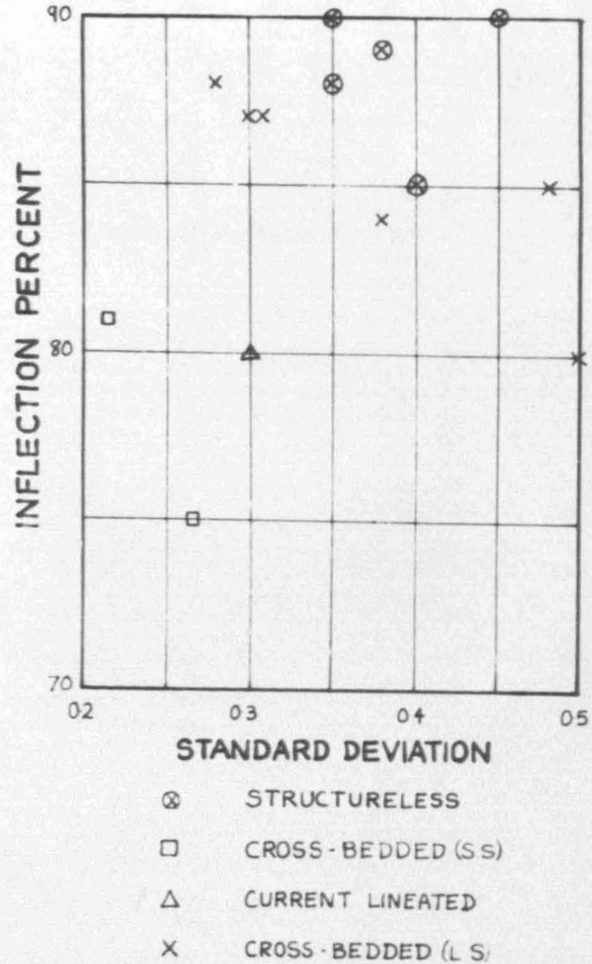


FIG. 3.2 INFLECT PERCENT/STANDARD DEVIATION RATIO RELATED TO STRUCTURE

may reflect the higher energy conditions that field observations would suggest prevailed during deposition of Section B. While the sandstones of Section A are principally horizontally laminated, Section B includes cross-bedded and structureless sandstones. Inflect percentage may therefore be related to structure and hence depositional environment.

The deltaic sandstones are ideally suited for testing this hypothesis since they have been considered throughout as structural groups. The average inflect percentages for each group are as follows (the number of samples is given in brackets)

Small-scale cross-bedded	78%	(2)
Large-scale cross-bedded	85%	(6)
Current-lineated	80%	(1)
Structureless	89%	(5)

The structureless sandstones are seen to have higher inflection percentage than the remaining groups, all of which show some degree of lamination. This suggests a possible relationship between I% and degree of lamination. The two small-scale cross-bedded samples show ripple drift structure but sample 1 shows a lower degree of lamination than sample 2 which corresponds to their I% values - 81 and 75 respectively.

The dependence of I% on degree of lamination can also be seen in a re-examination of the Eller Beck Bed data. As already pointed out Section A comprises horizontally laminated sandstones of lower average inflect percentage than the predominantly cross-bedded or structureless sandstones of Section B. Within Section A itself the same feature may be seen in the form of extremely low inflection percentages of the samples A2 and A3 which mark the downward passage of laminated sandstones into laminated silts.

B. INTERPRETATION OF CUMULATIVE CURVES AS INDICATING MECHANISM OF ACCRETION

A low inflection point reflects the presence of a high percentage of fine grained material in addition to the coarser log-normally distributed sandstone. The distict separation of the fine material from the remainder is unlike the product of simple lack of sorting and, in view of the relationship observed between inflection percentage and degree of lamination, the fine 'tail' is here interpreted as representing the grain size distribution of fine laminae interbedded with the dominant coarse fraction.

Intercalation of fine-grained laminae must result from variations in current strength, and such variations would be expected to reduce the sorting of the sandstone. A graph of I% against $\sigma_1\phi$ has been drawn (fig.3.2) to investigate this possibility. Although the number of samples is low, there appears to be a crude inverse relationship between inflection percentage and standard deviation within any one structural group. In samples showing a high degree of lamination the sorting of the log-normal fraction is relatively poor. This indicates that a current whose competence frequently drops to a level at which fine material may accumulate is of variable strength at higher rates of flow. Inspection of the cumulative curves for these samples shows that in general those with low inflection points have diffuse inflections which probably result from the frequent transition between the low and high energy conditions. This is particularly evident in the current lineated sample (9) in which the laminae show less contrast in grain size than those of cross-stratified samples. Curves with high inflection points (10,12,14) show a relatively sharp break in slope, and tend to show less skewness at the coarse end.

In figure 3.2 the structureless sandstones possible show a trend similar to that of the other groups but, as in the cross-bedded group, one sample shows an anomolously high σ_1:1% value. It is possible that the fine material in the structureless sandstones does not owe its origin to interlamination. Processes such as trapping of fine sediment on the depositional surface or post-depositional filtering could produce a fine tail. Such processes almost certainly operate to some extent in all fluviatile sandy sediments, but whether they could result in the incorporation of 10 to 15 percent of fine material is not certain.

It is clear from inspection of figure 3.2 that the small-scale and large-scale cross-bedded and structureless sandstones fall into reasonably distinct categories. The structureless samples, though having higher inflection percentages that the large-scale cross-bedded sandstones, do not lie on the projection of the trend line of the cross bedded samples. This indicates that they do not represent the limit of reduction of number of interlaminations of fine material in cross-bedded sands, and that they have therefore been deposited under different hydraulic conditions.

C. SUMMARY OF LITERATURE

While a considerable amount of work has been published which deals with the methodology of representing grain size distribution, few authors have attempted to evaluate the significance of the details of the size frequency curves obtained.

Tanner (1964) dealt with the problem in a general manner, considering the possible ways in which sedimentary processes could modify a supposed original log-normal distribution. On this basis the type of curve obtained in this study could be produced by the addition of fine material to a log-normally distributed fraction by some form of mixing.

A detailed insight into the possible mechanisms which might produce such mixing is provided by Moss (1962,3). He distinguishes three populations which are usually present at any one time during the transport of sandy sediments, these being:

A: Grains moving by saltation

B: Grains moving in suspension

C: Grains moving by rolling and creeping.

The separation of A and B in terms of grain size depends on current strength and turbulence. When a current is flowing at a particular speed grains of a corresponding size range are moving by saltation. At moderate and high current speeds the current required to transport grains is similar to that required to bring grains into motion, and rate of accumulation of grains is therefore slow. Grains of population B for the most part remain in suspension but chance landings cause some grains to be incorporated within the interstices between

A population grains. Each sedimentary lamina appears to be characterised by a single population A.

At low velocities, when only small grains are being brought in (or where only small grains are available) the erosional velocity of the grains begins to increase rapidly relative to the depositional velocity. Beyond a certain limit saltation will not be able to occur and deposition will be direct from suspension. The smallest population A grain size measured by Moss was 0.07 mm (equals 3.83ϕ). Moss (1962,p.370) also related his populations A,B and C to common cumulative curves (not, however, plotted on arithmetical probability paper). In some suggested curve forms for single laminae the bulk of the distribution was alotted to population A, the fine tail to population B, and the coarse tail to population C. Population B was represented as varying from less that 5 percent to 20 percent of the total. Moss suggests later (p.371) that "....it appears that sorting measures could be more successfully applied to individual populations than to whole sediments". This has, in effect, been carried out in this study,by assuming log-normal distribution for the population A which is however only an average of a large number of populations for individual laminae.

Moss's classification of transported sediment into three populations was adopted by Visher (1965) and used to interpret grain size frequency curves similar to those obtained in this study. Visher found that the grain size distributions from both recent and ancient fluvial sandstones showed remarkable similarity. This led him to suppose that the slope of the curve segments and the position of the inflection point may be genetically related to transport dynamics. The curves could be represented by two log-normal fractions which overlapped one another in the region of inflection.

The distributions of a hundred samples were treated by factor analysis using the following variables: 1,5,50,95 percentiles; Folk and Ward (1957) values for mean, sorting and skewness; grain size at inflection point; angle between curve segments; and a measure of the mixing of the two populations. Four classes were obtained and these corresponded closely to sedimentary structures:

1) Trough cross-laminated
2) Current laminated
3) Small-scale trough cross-bedded
4) Small-scale trough cross-laminated

Intergradations occurred, but groups 1) and 4), and groups 2) and 3) were always separate. Class 1) is interpreted as resulting from interlamination of populations A and B. Class 2), consisting of alternating coarse and fine laminae shows two distinctly separate log-normal segments. In class 3) the finer population is less distinct, and is interpreted as resulting from the entrapment of fine particles (of Moss's population B) in the interstices of the coarser population. Class 4) is considered as an extreme end member, with the entire distribution composed of the finer, or suspension population.

D. INTERPRETATION OF RESULTS WITH REFERENCE TO LITERATURE

Moss's populations have been adopted, but with slightly different conclusions to those of Visher. The major log-normal segment is interpreted as representing population A. Population C when present produces a slight positive skewness at the coarse end of the curve. Many of the coarser sandstones (other than the coarse basal sandstones) show negative skewness, or truncation, at their coarse end which is interpreted as indicating depletion of population A probably due to lack of available grains of that size.

The grain size at the fine end of the A population may be determined by reading off the point at which the inflection curve meets the roughly log-normal distribution of the fine tail. Values range from 1.55ϕ to 3.85ϕ. The whole range of values for both Lower Deltaic and Eller Beck Bed samples is represented in a histogram (fig.3.3). The fine limit (3.85ϕ) corresponds closely with Moss's value of 0.07 (3.83ϕ) for the finest A population grains measured.

This clearly suggests that the coarser log-normal fraction represents Moss's A population. In samples where the log-normal section approaches this limit the fine tail must therefore represent the B population. In the coarser sandstones the fine tail could represent the B population or both A and B populations. This probably depends on the time involved in the deposition of a fine lamina. If the lamina results from a sudden decrease in current then the grains probably represent the dumping of the suspension population of the previous current regime. If, however, a longer time interval is involved the fine laminae may include A population grains corresponding to the new lower current velocity. It is very difficult to generalise on rates of deposition of laminae (Jopling 1966), but since in thin

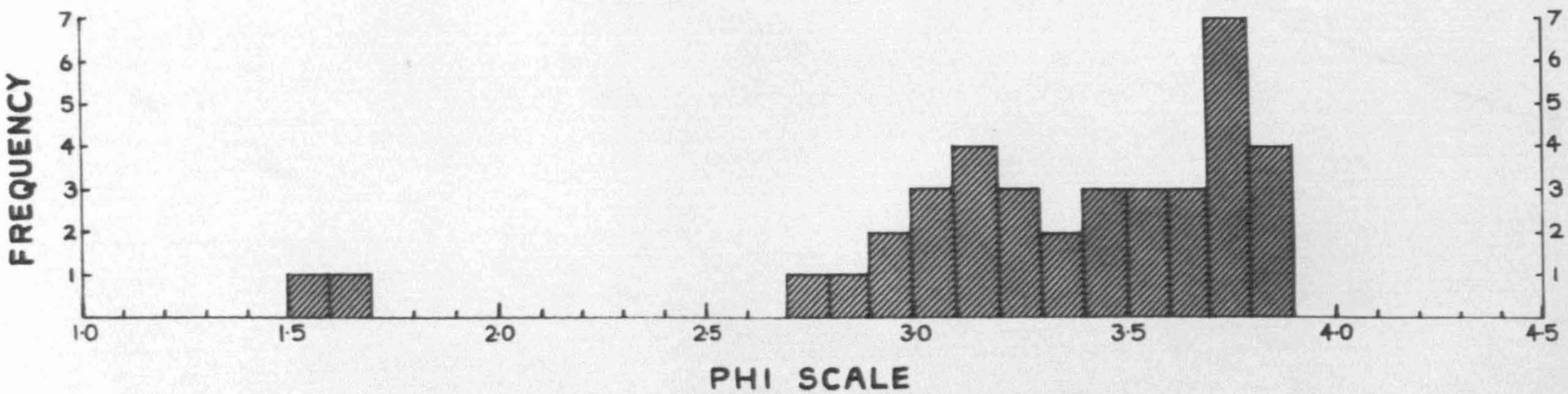


FIG. 3.3 SIZE FREQUENCY HISTOGRAM OF FINE LIMIT OF COARSE LOG-NORMAL DISTRIBUTIONS

section the fine laminae are seen as a gradational decrease in grain size with no apparent change of sorting, they probably represent changing populations A and B.

The large-scale cross-bedded sandstones appear to consist of a coarser population A fraction with a finer fraction consisting of populations A and B which belong to finer grained laminae. The current lineated sandstone, which shows a diffuse inflection point, probably consists of a large number of populations A, varying from lamina to lamina, and some grains of population B which may well be entirely interstitial.

The small-scale cross-bedded sandstones are considered to have a make-up similar to the large-scale cross-bedded ones. This is contrary to Visher's conclusion that they consist entirely of grains deposited from suspension. The limit of the log normal portion of the curve in these samples corresponds with that of the other structural groups, and is here considered as representing the A population. The very nature of cross-bedding implies deposition from populations A or C since particles are continually moving during ripple migration. The fine tail consists of population B, probably in both laminae and interstitial sites.

In the structureless sandstones most of the grains belong to population A. The fine-grained material is presumably not related to lamination, and is therefore interpreted as occupying interstitial sites within the A population. The percentage of fine material varies within the limits implied in Moss's suggested single-lamina grain size distributions.

The samples have only been analysed down to the finest sieve mesh. As pointed out earlier, however, the curves must show an upward inflection towards their fine end, since at the silt grades the curves are almost horizontal. This implies a secondary mode in the clay grade. A problem arises here in that it is difficult to determine whether the clay component is primary or secondary or a combination of the two. Petrographic evidence suggests that the clay in its present form is diagenetic, being either a cement or a product of the recrystallisation of detrital clays.

A comparison of the cumulative curves with Visher's curves for recent fluviatile sandstones suggests that the clay is to some extent secondary. In the two curves representing current laminated sandstones the slope of the fine segments suggest that the finest grains present (at least in appreciable quantities) have diameters between 5.5 and 6 ϕ. Sample 9 of this study, however, shows an initially steep slope which falls off rapidly around 4.25ϕ. By producing the initial slope a fine limit of about 6ϕ is obtained, and at this grain size about 2 percent of clay remains unaccounted for. Considering the probable high flow regime origin of current laminated sandstones it is unlikely that much clay would be incorporated into the sediment, and the 2 percent of clay is therefore interpreted as being authigenic.

In the cross-bedded and cross-laminated samples the fine section of the curve does not show this initial steep slope, but tends to level off at the 90 to 95 percentiles. If the amount of authigenic clay corresponds to that in the current lineated sandstone this implies a primary origin for a clay content forming from 10 to 20 percent of the total. Visher's recent festoon-bedded sample does not, however, show significantly higher clay content than the current laminated samples.

If a second mode is present in the primary clay fraction it may be related to the limiting grain size of population A. Saltation does not occur in appreciable amounts below a certain grain size, probably at about 3.85ϕ, and beyond this limit particles are deposited direct from suspension. Since erosional velocity increases rapidly with respect to depositional velocity beyond this limit it is probable that the finer clay-sized grains are more likely to stay put, once deposited than coarser silt-sized grains.

The slight relative deficiency in silt-sized particles, which corresponds to an apparent universal deficiency (Pettijohn, 1957; p.47), may therefore be related to the limit of substantial saltation. Particles of grain size immediately below this limit probably tend to remain in suspension and would be deposited under relatively quiet water conditions.

This limit to saltation probably also determines the limit to formation of ripple marks, since such structures cannot form without the development of an A population. Although some authors (e.g. Kindle & Bucher 1919 p.633) have pinted out that ripple marks do not form in mud the exact critical grain size has apparently not been determined. Both this problem and that of the clay fraction could be solved by flume experiments on homogeneous and heterogeneous sediments.

II DISTRIBUTION OF IRON

Both the Lower Deltaic Series and the Eller Beck Bed show a moderately high proportion of iron-rich sediments, a feature which also characterised the Dogger and the Cleveland Ironstone Series. In the non-marine beds iron is represented in the shales by siderite nodules and sphaerosiderite and in the sandstones as an intergranular cement. It is, however, not represented in many of the siltstone horizons. The Eller Beck Bed is characterised by the development of one or more bedded ironstones at the base, and of nodules within the shales. The sandstone is generally not iron-rich, but in some localities a part or all of the upper division has a siderite matrix which appears to be related to the development of ironstone near Winter Gill and at other places. Occasionally bands of ironstone are associated with silty beds within the sandstone. The overall distribution of iron may be summarised as follows;

1. In marine beds as continuous (or, occasionally, localised) beds of sideritised chamositic ironstones;
2. As segregations within shales;
3. As replacement of siltstone and shale partings or fragments within sandstone units;
4. As a cement in some non-marine sandstones.

The presence of ironstone in the non-marine beds is to be expected since siderite concretions are characteristic of most fossil delta top deposits. The presence of siderite in the sandstones in large amounts is relatively uncommon. It is clearly of diagenetic origin and often occurs in such quantities that enrichment from some outside source must have taken place. The channel sandstones of this delta differ from those of, for example, the Carboniferous in that they are laterally inextensive; as a result the sandstone is always in close proximity to the surrounding shale and siltstone. The siderite is often accompanied by kalinite, which in some sandstones makes up twenty percent of the rock, and it seems probable that both of these minerals have been precipitated from solutions produced during diagenesis of the shales or siltstones.

Most of the finer grained siltstones and shales are characterised by the development of siderite nodules during diagenesis, whereas most of the siltstones are devoid of concretions. This may suggest that the coarser beds lose their mobilised iron by migration into sandy beds. Porosity could be the controlling factor, with the high clay content of the finest beds permitting only local migration. Alternatively the lack of concretionary iron in the siltstones may reflect an originally low iron content; this would imply a complex process of iron redistribution in the shales.

In the Eller Beck Bed the iron is chiefly associated with bedded ironstones at the base and with localised ironstones associated with the sandstone unit. Siderite nodules are moderately common, but not abundant and are rare or absent in the extreme south. This variation in nodule content must represent either an original variation in iron content of the shale or different degrees of mobilisation of the iron present. As the shale shows no other difference it seems probable that the original iron content was higher in the north. The bedded ironstones seem to have been originally chamositic muds with little other material in some cases. The chamosite appears to have been primary or halmyrolitic, whereas the siderite is thought always to be secondary.

In order to understand the distribution of iron in the sediments it is necessary first to consider in what form the iron was introduced. The principal mechanisms of transport which have been proposed are as follows:

(i) Transport in the solid state;

a) As a ferric hydroxide hydrosol stabilised by organic colloids.

b) As a surface coating on clay minerals.

(ii) In aqueous solution.

Transport in association with organic colloids has been suggested by several workers, and has been supported by experimental results (Moore and Maynard; 1929, p.277). Evidence from natural stream waters, however, appears to be lacking, and this mechanism has not been referred to in recent work as being important in the transportation of iron. Transport in the form of iron oxide coatings on clay minerals is considered to be the principal mechanism by many recent workers and has been considered in detail by Carroll (1958). The carrying of iron in solution was for a long time believed to be of little importance, but Huber and Garels (1953) have shown that under certain naturally occurring conditions, both marine and non-marine, iron may be dissolved in sufficient quantities as to have formed sedimentary ironstone deposits. The nature and distribution of iron in the Lower Deltaic Series and Eller Beck Bed suggests that iron was transported in more than one form.

The abundance of diagenetic ironstone in a sediment must be related to the amount of iron deposited with the sediment and to the degree of mobilisation of the iron, assuming that there has been no enrichment from outlying sources. The concretions in the non-marine shales are either small, well-defined siderite mudstone nodules or larger, diffuse areas of sphaerosiderite. These seem to be of two ages; the nodules which are never penetrated by rootlets and which show strong compactional features around them appear to be early, whereas the diffuse sphaerosiderite beds are gradational into the shale and appear to have developed after compaction. The nodules appear to represent local redistribution of iron and generally show a relationship to one or more horizons within any bed; the sphaerosiderite developments appear to involve the whole bed and are usually situated around the middle of the bed. The formation of sphaerosiderite involves a high degree of replacement of the detrital constituents of the rock whereas in the nodules a large amount of quartz and clay minerals is retained. It would therefore seem that the nodules were formed essentially by early interstitial cementation by siderite and subsequent alteration of the included grains, whereas the sphaerosiderite has formed mostly by replacement. This later mobilisation could have been accompanied by migration of iron into sandstones.

Although iron is conspicuous within the non-marine sediments it is difficult to compare the amount with that in other fossil deltas such as those of the Carboniferous. This is not the case with the marine beds in which iron is clearly much more abundant to the extent that bedded ironstones seem to replace the limestone of the Carboniferous cycles.

The nodular segregations in the marine beds may be of a similar origin to those of the non-marine beds. They are, however, characteristically regular in form, in contrast to the non-marine nodules, which may reflect somewhat later formation. Sometimes the ironstone is random, at other times, as in Eller Beck shelly horizons have been selected. The development of nodules in the north, but not in the south presents a problem. If the available iron was in solution it implies that either iron was scarcer in the south than in the north, or that conditions were not suitable for precipitation. If the iron was coated on grains then again either it was scarcer in the south or mobilisation or precipitation were unable to take place. In the second case it is probable that original distribution of iron would be fairly uniform and also that solution of iron oxide would take place during diagenesis. The remaining possibilities all imply variation in the chemical properties of the sea water or pore water from north to south. Huber and Garrels (1953, fig.17) considered that 3 parts per million of iron could remain in solution in sea water under intermediate Eh conditions at any natural pH. In fresh water with pH values between 5 and 6 iron may be carried in solution even in well oxygenated waters. Deficiency in calcium carbonate favours the low pH values.

Several explanations may now be put forward for the distribution of nodules:

(i) Sea water approximately neutral in north, but alkaline in south due to presence of dissolved calcium carbonate. If salinity was lowered in the north by influxes of calcium-free fresh water mildly acid conditions may even have prevailed. Dissolved iron trapped in the mud in the northern part would be precipitated as reducing conditions set in, as long as the pore water remained acid. The chemistry of the diagenesis could, of course, be more complex, but in the south no iron would be available for precipitation.

(ii) If the iron were introduced with the detritus mobilisation could only occur under oxidising acid or reducing alkaline conditions. In either case it is difficult to see why there should be any contrast in the chemistry of the diagenesis in the more or less uniform shales unless it was a reflection of an original chemical variation in the sea water as suggested in (i).

The distribution of nodules is therefore attributed to variation in pH of the sea water, probably caused by variation in calcium carbonate concentration. The separation of the calcium deficient water from that to the south corresponds to the southern limit of the subsiding trough, and water flow down this trough may, to some extent have prevented mixing of the waters. The contrast in iron content becomes pronounced above the level of the bedded ironstones at the base of the marine beds. The development of the Hydraulic Limestone to the south indicates that the contrast in calcium content was due both to the incursion of calcium-rich water from the south as well as calcium-poor water from the north. The general absence in the north of calcium carbonate supports this hypothesis. The only occurrences of calcium carbonate are in Winter Gill and Great Fryupdale, and are associated with sandbanks and shell concentrations. The presence of shell material in large amounts in the sandstone unit seems to have set up a local chemical environment in which, as long as free circulation with the sea was maintained, calcium carbonate was precipitated despite its low concentration. The Great Fryupdale sandbank was cemented at first by siderite but later by calcite. This may have resulted from the bank building up to near sea-level and permitting concentration of calcium carbonate by evaporation. The Winter Gill sandbank appears to have undergone early cementation by siderite only, and the later calcite cement appears to be only a reoganisation of calcium carbonate derived from shells. Elsewhere in the Eller Beck Bed shell material is invariably replaced by siderite or kaolinite or represented by voids, indicating that conditions during diagenesis favoured solution of calcite.

The conditions of formation of chamosite are still very much in doubt as lttle is known of the relationships of iron silicates to Eh and pH (Huber and Garrels 1953, p.357). The discovery of chamosite in recent sediments (Porrenga, 1967) may yield some information. In the present study chamosite is believed to have accumulated on the sea floor, and not be of diagenetic origin. Diagenesis clearly

~~forms~~ favours the formation of siderite. The following mechanisms of formation are put forward for discussion:

(i) Physico-chemical precipitation from solution;

(ii) Biochemical halmyrolitic precipitation from solution;

(iii) halmyrolitic alteration of terrigenous mud.

In the preceding paragraphs it was suggested that Eller Beck Bed sediments were influenced by the presence of calcium-rich and calcium-poor waters. In the shale unit those apparently remained separated to some extent, perhaps as a result of the presence of the deeper water trough. This trough may not have been so active during ironstone deposition and a more thorough mixing of waters could have caused precipitation of iron in the form of chamosite. This mechanism is not, however, applicable to the Winter Gill ironstone. Both the bedded ironstones and the Winter Gill ironstone seem to indicate that a low rate of sedimentation is necessary for chamosite formation. Since the sandstone unit is believed to have accumulated under conditions of reduced salinity this does not seem to have been a controlling factor; neither does the presence of a shelly fauna. A low rate of sedimentation does not seem to be the only factor since the fine grained shales are also the product of low sedimentation. The occurrence of chamosite at the base and at the top of the sequence suggest that relatively shallow water is also required.

If a depth control is operating this suggests that organic activity is in some way involved in chamosite formation. The scarcity of fine-grained terrigenous material which is particularly pronounced in the Winter Gill ironstone is either a primary or a secondary feature. The presence only of rather large corroded sand grains suggests that the finer constituents have been removed by solution. Similar features are seen in the Ingleby ironstone seam, but introduction of sand and silt and probable reworking has apparently destroyed this relationship in much of the basal ironstone seam. Provided iron was available it could, therefore, have been concentrated by removal of the terrigenous constituents.

The origin of the iron must now be considered. It was either derived from solution or from the sediment. Carroll (1958) discusses the transport of iron by clay minerals, and considers (p.23) that sufficient iron oxide may be present to form chamosite by combination with alumina and silica of the clay minerals. Thiel is quoted as having shown that removal of alumina and silica is increased by a factor of five in the presence of bacteria. Excess silica and alumina would need to be removed from the clay minerals, and removal of silica is also indicated by the absence of all but the largest, and therefore most resistant, quartz grains. The formation of chamosite by combination of iron and clay minerals was proposed by Hallimond (1925, p.98) and is accepted here. It is also thought probable that organic activity was responsible for setting up a chemical environment in which this reorganisation might take place.

The Eller Beck Bed ironstones invariably show replacement by siderite, which is believed to have taken place during very early diagenesis. The conversion of chamosite to siderite involves a change in iron content from about 30 percent to 48 percent. The sideritisation of a chamosite mud therefore involves a considerable enrichment in iron. The iron may be derived from solution or from clay minerals in surrounding beds, but the Winter Gill ironstone is not associated with clay-rich strata. There is also evidence that sideritisation of the basal ironstone may have preceded shale deposition to a large extent (p. 151), although the sediment was still unconsolidated. The secondary iron must therefore have been introduced from the overlying water, and must have occurred in solution. The concentration of iron in the sediment is visualised as precipitation of siderite either as a pore filling or as a replacement of chamosite. Continued precipitation would set up a gradient whereby ferrous ions would be continuously migrating from the sea-water into the interstitial water and finally to the site of precipitation. The enrichment in iron presumably resulted not only from the instability of the chamosite but possibly liberation of carbon dioxide, together with suitable Eh and pH conditions, favoured siderite formation.

Sideritasation in these beds has been very extensive. In other formations this is not always the case and even in these beds the ooliths have generally resisted total replacement. The degree of sideritisation probably depends on the rate of accumulation; rapid deposition would cut off the free circulation with the sea water and prevent iron enrichment and also perhaps the formation of available carbon dioxide. Some other factor must also operate if the conclusion as to the originof the Winter Gill ironstone are correct. This supposed that oolith formation was preceded by mud formation; sideritisation of the mud is, however, believed to have taken place after its redistribution.

Further enrichment of the ironstone may occur after burial. The distribution of concretionary ironstone in the shale unit on the coast is clearly related to the distribution of the basal ironstone. Concretions are absent where the basal ironstone is present and are present where it is absent. The basal ironstone seems to have attracted mobilised iron from the lower part of the shale, and this has probably resulted in total lithificaation, probably accompanied by replacement of terrigenous grains. These processes appear to have occurred before burial in the Winter Gill ironstone which is believed to have remained exposed for relatively long time before burial beneath subsequent sediments.

The chemistry of ironstone formation is clearly complex, and cannot be determined from one or two examples. The following generalisations are however believed to be applicable to similar ironstones:

(i) iron was introduced both in solution and as grain coatings;

(ii) chamosite mud may be formed under conditions of varying salinity and slow sedimentation, probably by combination of clay minerals and associated free iron. Deficiency in lime and organic activity are probably necessary.

(iii) siderite is always secondary; where the rate of sedimentation exceeds that of iron mobilisation siderite is formed by redistribution of iron within the sediment; at low sedimentation rates chamosite is formed and siderite is formed by ion exchange with the overlying water.

III CHEMISTRY OF DIAGENESIS

In previous individual sections the chemical aspects of diagenesis have not been considered as it was thought that less repetition would be involved if the subject were treated in a section of its own. The relationship of the various iron minerals to one another has been considered in the preceding section, and so they are now considered only in their relationship to other minerals. The exact stability fields of many of the minerals are not fully known and the following discussion is therefore arrived at identifying trends in the chemical diagenesis.

In general terms the basic sequence of diagenetic mineral formation in both marine and non-marine beds is:

(i) Precipitation of iron carbonate - siderite

(ii) Precipitation of clay mineral - kaolinite (rarely dickite).

The first stage may involve two phases represented the non-marine beds by a) nodule formation and b) sandstone cementation and perhaps sphaerosiderite formation, and in the marine beds by a) precipitation of siderite as bedded or concretionary mudstone b) rim cementation.

The full sequence of diagenesis shown by the non-calcareous marine ironstone will be considered in detail. This sequence appears to be:

(i) Pyritisation of aragonite shells in part

(ii) Solution of aragonite shells in part

(iii) Precipitation of siderite as finely crystalline matrix

(iv) Solution of aragonite and calcite shells

(v) Precipitation of drusy siderite

(vi) Precipitation of clay mineral in remaining rocks, and replacement of chamosite

(vii) Local replacement by pyrite.

This sequence may be related to the Eh and pH changes which are characteristic of the sequence of deposition and burial of modern sediments (Fairbridge in Larsen and Chilingar 1967, p.27) except for the early phase of pyritisation. The first agents in promoting diagenetic changes in fine-grained sediments are the aerobic bacteria which attack organic material and use up all of the available oxygen in the process. Their action results in neutral and slightly reducing conditions.

Under such conditions both aragonite and calcite might be expected to undergo solution but in most cases only partial solution of aragonite shells occurs before phase (iii) siderite precipitation. Since further solution of shell material occurs after sideritisation of the matrix it is possible that siderite is precipitated very rapidly relative to a slow continuous shell solution process. The critical point is probably dependent on the Eh, pH and the production of large amounts of CO_2 by bacterial action. The large amount of iron required must be extracted from solution; this iron may result from solution of oxide films on clay particles to form nodules , but in the bedded ironstones must be derived from chamosite and from the sea-water itself. Siderite precipitation may be expected to take place, so long as the sea water is saturated in iron and the interstitial water is more alkaline than the sea water, provided that no impervious clay cover is deposited such as would prevent the diffusion of iron ions into the sediment. Deposition of siderite must have continued until a fairly dense crystalline aggregate was produced and dehydration, indicated by shrinkage cracks preceded the main phase of shell solution. This phase of solution involved both calcite and aragonite, and the interstitial water must still have been neutral to mildly alkaline at this stage. The shell voids and shrinkage cracks usually possess a rim of siderite which is only of limited thickness, probably reflecting a depletion in the supply of iron or carbon dioxide.

All remaining voids have been subsequently filled by authigenic kaolinite or, rarely, dickite. These clay minerals were probably formed under continuing alkaline conditions. Cavity fill is accompanied by replacement of chamosite by kaolinite. The final phase appears to be pyritisation, representing the maximum in reducing conditions produced by bacterial. Sphalerite may also form at this stage.

In the non-marine beds kaolinite similarly always follows siderite in order of precipitation. This suggests either some difference in conditions of precipitation for the two minerals or similar conditions with varying rates of attainment of saturation of the minerals. As discussed earlier, a probable source for the siderite and kaolinite in the deltaic sandstones lies in the surrounding shales. The iron, if introduced as iron oxide on clay particles, would be readily liberated into solution under the reducing and acid conditions that usually prevail in vegetated back-swamp areas. With compaction these iron-bearing solutions would rend to pass into the porous sandstone bodies whose interstitial water would probably be of lower acidity, causing precipitation of iron. The kaolinite-bearing solutions, on the other hand, may be generated at a later stage, probably through hydrolysis of fine-grained feldspar under acid conditions. A variation in the ease, and therefore, rate of formation of iron and clay-bearing solutions may well account for the ubiquitous precedence of kaolinite by siderite.

The marine beds show a few local variations from the usual mineral assemblage. In some places phosphate appears to have been an early-formed mineral, possibly replacing chamosite mud both before and during early stages of sideritisation. Phosphate is particularly associated with burrows. The formation of phosphate is believed to take place under acid and slightly reducing conditions (Fairbridge, op.cit. p.74). This indicates early sea-floor reaction before bacterial action has begun to raise the pH and lower the Eh.

The Eller Beck Bed also shows local calcareous developments, a particularly in Great Fryupdale and Winter Gill. In the former case the development of a sandstone lens in the shale unit seems to have caused the local calcareous environment since although the associated ironstone and sandstone units show all of the normal characteristics except that shell material has not been leached out, and a calcite cement is present in the sandstone. The sandstone lens has a predominantly calcite cement, though a thin rim of early siderite cement, is present.

This would seem to indicate precipitation of iron in a mildly alkaline reducing environment. The change to calcite must therefore reflect a change to more alkaline conditions when perhaps the available carbon dioxide was preferentially taken up by calcite and the iron incorporated into the calcite lattice (the diagenetic calcite is invariably ferroan). The sand lens appears to have been sufficiently large for it to have maintained its calcareous conditions and to affect the overlying sandstone unit. It may be important in this respect that the lens is locally overlain directly by thin-bedded sandstones, so that it was never sealed off by an impervious shale layer.

Features associated with the calcareous facies include replacement of aragonite by ferroan calcite which is simply a transition from a metastable to a stable form. Early replacement of calcite by quartz seems to indicate an initial phase of relatively low pH. Dolomite is locally formed and appears to be controlled by the occurrence of chlorite or chamosite which presumably supplies the magnesium required, or at least acts as focus for crystallisation.

The Winter Gill sandstone lens is superficially similar to that of Great Fryupdale. Siderite is, however, more abundant and in the basal part at least its precipitation is separated from that of ferroan calcite by a phase of shell fracture. It is quantitatively possible that the calcite cement in the lower part of the sandstone is derived by leaching of the shells in the upper part under simultaneous conditions of high and low pH respectively.

The calcite cement in these cases seems to take the place of kaolinite in other beds. Perhaps the Eh and pH conditions are similar in both cases but the sand banks have been able to maintain open connection with the sea water to allow concentration of calcite in the manner proposed for siderite precipitation.

One other interesting phenomenon remains to be considered, namely the occurrence of authigenic albite at the top of the Winter Gill Ironstone. It occurs as a replacement of chamosite and quartz, and its restriction to the top few inches of the ironstone indicates sea-floor alteration at a time of non deposition. The formation of feldspar presumably requires an alkaline environment and it may be that the sodium has been introduced by prolonged contact with sea water. This process has been suggested in other cases of authigenic feldspar (Crowley 1939).

In summary it would appear that the rate at which an impervious clay layer is developed over a body of sediment undergoing diagenesis is extremely significant as regards the nature of the final product. In this particular study isolation of a porous body in shale results in the introduction of clay bearing solutions, whereas if an open connection is maintained with overlying water precipitation of carbonates takes place.

Looking at the sediments as a whole it would appear that a considerable amount of siderite has been precipitated. In the case of the marine beds much of this siderite appears to have been derived from solution in sea water. Siderite always precedes calcite as a cementing material, and the calcite is always ferroan. In sediments of similar depositional character but of other formations siderite is often rare or absent. Conversely calcite is more abundant, both as a diagenetic cement and as a primary precipitate. These contrasting features suggest that the Eller Beck Bed sea contained an unusually low proportion of calcium carbonate which would in itself favour a greater concentration of iron in solution.

IV SUMMARY OF HISTORY OF SEDIMENTATION

Deposition of the Lower Deltaic beds appears to have been preceded by local reworking of Dogger sediments in the east. In the north and west there is no evidence of this and in one exposure ease of Roseberry Topping the contact appears to be gradational. This is particularly interesting since at Roseberry Topping itself and on the adjacent escarpment the *Thinnfeldia* plant bed is developed and the Dogger absent. This seems to indicate that a Dogger shore-line could be represented here. The field relationship of the plant bed to the Dogger might well be discovered by excavating the base of the Deltaic Beds along this escarpment.

The Dogger surface must have been at or near sea level when the non-marine sediments were first introduced into the area since shales or siltstones occur immediately above the Dogger. Sedimentation of the deltaic beds was principally of interdistributary type, resulting in the formation of successive beds of shale, siltstone, and thin sandstone. Except at the base and top these beds are mostly of small lateral extent, and erosion due to drainage seems to have been prevalent. At three main levels sandstone-filled channels are developed; these are mostly strongly erosional into the underlying beds, and little overbank sedimentation appears to be associated with them. They probably had a somewhat meandering form but lateral incision is not extensively developed and the streams appear to have been short-lived. One or two localities display sandstones which appear to represent almost continuous deposition, and these are thought to mark the site of relatively large distributaries. Even these are not developed at the extreme base and at the top of the series and the sediments as a whole are thought to have been laid down between large permanent distributaries to the east and west.

The close of on-delta sedimentation is marked by widespread and uniform siltstone which locally passes up into shale. The top of these beds is extensively burrowed by marine organisms and appears to have become inundated by the sea with little or no erosion. Subsequently the depositon of mud, probably chamositic, became widespread and a lamellibranch fauna became established. The deposition of these basal muds was probably slow but influxes of sand and silt appear to have occurred throughout. The upper contact with shale appears to be sharp and this may represent increased influx of silt and clay, with more rapid deposition. Shale deposition continued for a long time in some areas but in the extreme north and northeast sand began to spread in. This incursion of sand must have been quite rapid but did not get into extreme south and southwestern districts until late; the subsiding trough appears to have acted as a trap for the sand. The sea was probably very shallow throughout with the maximum depths occurring along the axis of subsidence. The sandstone deposition appears to have been relatively rapid and resulted in the filling up of the marine basin. At the same time salinity appears to have been reduced, as indicated by the sparse fauna in the sandstone and in the equivalent shale in the extreme south-west. The fauna is chiefly developed along the subsiding trough, along which a flow of fully marine water may have been maintained, possibly beneath a cover of less saline water. Shelly sandbanks occur in a similar position in the shale unit, and the presence of crinoid remains indicates more fully marine conditions than are indicated by the fauna elsewhere.

The sandstone appears to have built up to water level and over most of the area was colonised by plants. Some pause in deposition appears to have occurred before the first non-marine beds were laid-down; these consist of shales and siltstones as at the base of the Lower Deltaic beds.

Although the Eller Beck Bed sandstone represents a regression of the sea, and the readvance of the delta, transport appears to have been along the depositional strike although the bulk advance was southward. The sand is probably derived from an advancing major distributary whose sediment is being transported along the shore after having been introduced into the marine environment. It is therefore essentially a 'marine' deposit and once sea-level has been reached the fine-grained interdistributary sediments advance over the sand surface. No subsidiary channels are developed at this stage, except on a very small scale.

The base of the Lower Deltaic beds cannot be directly compared with that of the Middle Deltaic beds. The former represents the initiation of deltaic conditions while the latter represents a re-establishment of these conditions. Although the Dogger surface has acted in a similar way to the Eller Beck Bed sandstone surface the sediments themselves represent a completely distinct phase of sedimentation. The facies distribution bears no relationship to that of the Eller Beck bed and the time span is far greater, ranging from the Ludwigia murchisonae zone to the Lioceras opalinum zone.

V INFLUENCE OF TECTONICS ON SEDIMENTATION

The probable role of earth movements in affecting sedimentation of the Lower Deltaic beds has already been discussed (p. 74) Slight uplift and very gentle folding may have occurred before the spreading in of on-delta conditions. In the Scar section near Whitby localised downwarping appears to have provided an early stream course in which the sandy sediments appear to have slid during subsequent movements. Several rather poorly-defined cycles may be detected within the deltaic beds, consisting of four phases of interdistributary sedimentation and three of erosional channel development. The fine-grained sediments are believed to have been deposited between widely spaced distributaries during fairly stable tectonic phases; the channel horizons are believed to represent uplift of the hinterland and to some extent of the depositional area itself.

The on-delta sedimentation finished with a phase of uninterrupted subsidence during which uniformly fine-grained beds were laid down, and the upper part developed laminated shales in places. The topmost beds are overlain with apparent conformity by marine mudstones and siltstones. The marine transgression could have resulted from an increase in the rate of regional subsidence. Alternatively it may have resulted from lack of hinterland uplift producing a reduction in river activity; if subsidence continued at its normal rate recession of the delta would take place. These two mechanisms might be expected to have a similar effect on the distribution of non-marine and marine beds. One feature, however, indicates that subsidence did not continue entirely as before; the non-marine beds are thinnest in the centre of the region while the overlying marine beds are show their thickest development. This may imply a change in the relative subsidence pattern but may also reflect a topography related to the large-scale interdistributary area believed to cover the whole region.

As may be seen from the isopachyte maps in figure 3.4 the Eller Beck Bed thicknesses reflect subsidence along a north-west south-east trough. The ironstone unit shows slight thickening into the trough and the isopachytes are sufficiently regular in character for the thickening to be related to subsidence rather than to irregularity of the drowned delta surface. Shale unit thicknesses vary to a small extent over much of the area. The thinning in the extreme north and east seems to reflect relatively little subsidence on the margins of the basin but at the same time sedimentation of sand may have begun earlier in these areas. The axis of the trough seems to have been the site of occasional accumulations of coarse shelly detritus which built up as banks though possibly originally of a channelling nature. Relative subsidence is particularly clearly represented by the thickness variation of the sandstone unit, although the base of the sandstone may to a limited extent by diachronous. The areas of maximum subsidence seem to contain the most abundant fauna. The subsiding trough appears to have had a profound effect on sedimentation to the south-east. Not only does the sandstone thin in this direction, but the trough appears to

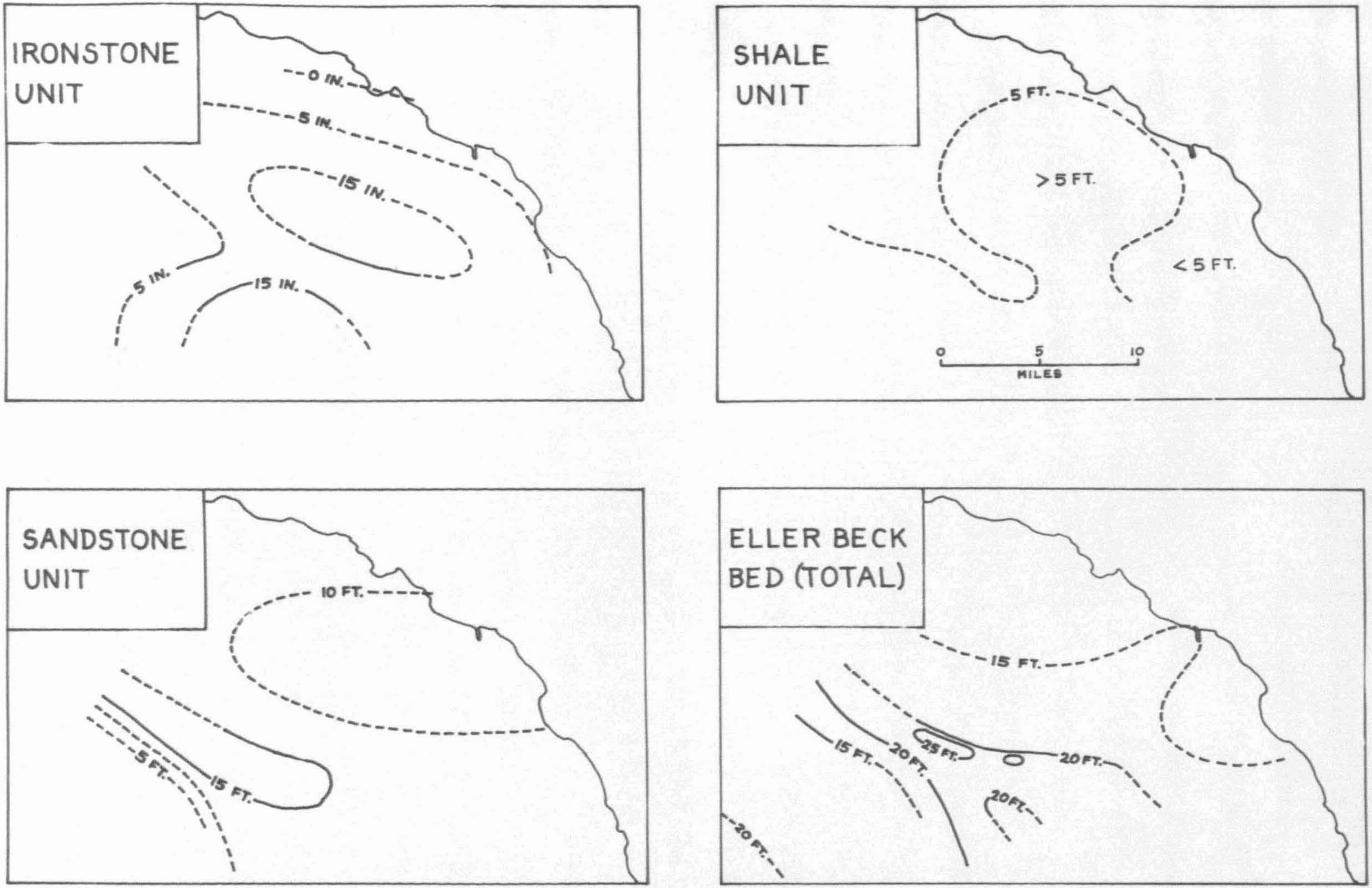


FIG. 3.4 ISOPACHYTE MAPS FOR THE ELLER BECK BED

have acted as a trap for detritus transported as bed-load so that the sandstone unit is to a large extent represented by shale. Sand appears to have spread south-eastwards only when filling of the trough was virtually complete. The limited field evidence seems to indicate relatively uniform thicknesses in the extreme south-east, and this presumably reflects uniform subsidence.

The readvance of the delta indicates a lessening of the subsidence rate and presumably a return to a tectonic environment similar to that under which the earlier non-marine beds were deposited.

It has been suggested that movement along the Peak Fault was contempporaneous with sedimentation (Wilson 1948, p.84) in order to explain the striking difference in thickness of the beds of either side of the fault. This was later ascribed to transcurrent movement (Hemingway in Howarth 1962) which brought together successions from areas of differing degrees of subsidence. There is no evidence from either the non-marine or the marine beds that throws any light on the matter. There is no evidence of contemporaneous erosion or of post-depositional distrubances associated with the fault.

A similar thickness variation may be seen in the Eller Beck Bed on either side of the Whitby fault (figs.2.2, 2.56). The shale unit is four feet thick on the south side and only two on the north side. On either side the thicknesses are maintained for some several miles. The lower sandstone division is thicker to the north, while the ironstone unit shows variation in facies, with only rare ooliths and shells in the north. There is again no sign of contemporaneous erosion of a subaqueous fault scarp and unless transcurrent movement had occured there seems to be little reason for a change of facies. Although in the non-marine beds rapid lateral variation is common, it is striking that the East Cliff section is particularly free of sandstone while it is such a short distance from the large distributary represented in the West Cliff.

The scanty evidence available suggests that transcurrent movement may have taken place on the Whitby fault. The ironstone unit in the East Cliff reaches a maximum of eight inches, and since this exceeds any other thickness on the coast it probably represents sedimentation in the axis of the trough. This would imply sinistral movement, such as would have occurred if the Peak Fault were transcurrent. Certain joint patterns and horizontal slickensiding observed in some inland sandstone outcrops could also point to the presence of transcurrent faults. It is possible that the striking difference between the Eller Beck Bed sections at Battersby Crags and Ingleby may be partly due to transcurrent faulting.

APPENDIX I

ELLER BECK BED SECTIONS

The following sections are accompanied by their respective grid re ferences, but their approximate locations may be determined from figure 2.1. Total section thicknesses are given to the nearest three inches.

1. IRON SCAR (015968)

	ft.	in.
Flat-bedded sandstone with ripple-marked partings	5	6
Thin-bedded sandstone with intercalations of silty shale and beds of impersistent silty ironstone; occasional shelly bands	6	0
Thin-bedded laminated sandstone interbedded with silty shale and ironstone	1	0
Silty shale with laminae of fine sand in upper part; shell casts in lower part	4	0
Ironstone, slightly oolitic siderite mudstone with scattered fossils. Abundant ironstone-filled burrows at base, penetrating the underlying beds		0-3
Silty shales, buff coloured, laminated (non-marine)		
TOTAL	16	9

2. RAVENSCAR (990014)

Flat-bedded laminated sandstone	6	0
Thin-bedded laminated sandstone, with intercalations of silty shale and silty ironstone near the base	7	0
Shale, silty in upper part with laminae of fine sand; shell casts in lower part	3	6
Ironstone, oolitic siderite mudstone with scattered fossils. Ironstone-filled burrows at base		
. TOTAL	16	9

3. HAWSKER (SOUTH) (939085)

	ft.	in.
Sandstone, penetrated by rootlets	1	0
Grey sandy carbonaceous silt	2	0
Flat-bedded sandstone, loaded at base locally	5	0
Grey silty shale		0-9
Sandstone with channelling base; loaded at base internally	2-4	
Grey shale well laminated and fissile in lower part, tough and silty in upper part with abundant carbonaceous material and coal scares	4-7	
Sandstone, tough and fine-grained, with partings of silty shale		3-10
Grey silty shale (non-marine) TOTAL	c.16	0

4. HAWSKER (NORTH) (938086)

Silty sandstone, with rootlets	1	0
Silty shale	2	3
Sandstone, thin bedded; thicker unit at top	5	0
Shale, with silty laminae in upper part	3	0
Ironstone, unfossiliferous fine-grained siderite mudstone		4-6
Shale	1	6
Ferruginous sandy silts (non-marine) penetrated by burrows filled with slightly oolitic ironstone		
TOTAL	c. 12	6

5. WHITBY - EAST CLIFF (902114)

	ft.	in.
Sandstone	3	0
Laminated sandstone, thin-bedded with occasional thicker units with channelling bases	5	0
Silty shale with sandy laminae	2	8
Ironstone nodule horizon	max.	3
Shale	1	6
Ironstone, impersistent; fossiliferous and oolitic with ironstone-filled burrows at the base penetrating underlying non-marine beds	max.	8
TOTAL c	14	0

6. WHITBY - WEST CLIFF (898113)

Laminated sandstone, generally flat-bedded but with some shallow trough cross-bedding	min.	5	3
Soft sandstone with laminae of silty shale		2	0
Silty shale		2	0
Ironstone, imperistent; siderite mudstone with sand grains and rare ooliths		max.	3
Soft dark silty sand with occasional small nodules of siderite mudstone		max.	2
Silty shale (non-marine)			

7. NEWHOLME BECK (863106)

	ft.	in.
Sandstone, rather thin-bedded with small-scale trough cross-bedding	7	0
Thin-bedded fine-grained laminated sandstone with abundant partings of silty shale and occasional ironstone nodules	5	3
Impersistent ironstone, unfossiliferous pale grey siderite mudstone	max.	3
Shale, silty in upper part	2	9
Ironstone, shelly pale grey siderite mudstone; irregularly bedded	max.	4
Poorly bedded silts, locally thick with developments of shelly siderite mudstone; silt-filled burrows at base		3-7

8. HINDERWELL (808168)

Laminated sandstone, blocky weathering	4	0
Thin-bedded laminated sandstone with silty partings; pale grey siderite mudstone bed with small fossils at base	6	9
Ferruginous sandstone, impersistent	max.	2
Silty shale		3
Ironstone, unfossiliferous fine-grained siderite mudstone	max.	2
Shale, silty at base with sand grains; occasional shell casts in lower part	1	5
Silty shale (non-marine) TOTAL	12	6

9. KILTON BECK (706177)

Sandstone, erosional base	3	0
Sandstone flat-bedded with occasional silty partings; shale flakes at base	7	6
Shale, with sharp contact with sandstone above; pale grey siderite mudstone nodules	2	3
Ironstone, impersistent, pale grey siderite mudstone with small shells and scattered ooliths, thin silty sand at base	max.	2½
Soft sandstone with irregular burrows, filled with silty shale, to depth of 1 foot; Occasionally 'nodules' of weathered siderite mudstone near top with rather haphazard distribution.		
TOTAL	c.13	0

10. LITTLE BECK (887049)

	ft.	in.
Sandstone, very thin-bedded at base	4	0
Shale, with sandy laminae in lower part	2	6
Poorly bedded silty sandstone with scattered shells and occasional shelly lenses; shale flakes and carbonaceous material present. Masses of ironstone (3in. max.) in horizon few inches from base, shelly siderite mudstone.	2	6
Sandy siltstone passing down into pale silty shale; shell casts present throughout	2	3
Soft dark silt with shell casts		3
Buff silty shales (non-marine)		
TOTAL	c.11	6

11. ELLER BECK (832022)

		ft.	in.
Soft, thin-bedded sandstone (non-marine)			
Flat-bedded sandstone, thin bedded at base; occasional ripple marks; U tubes and rare shell casts		10	0
Thin-bedded laminated sandstone with partings of silty shale which became thicker and more abundant towards the base		3	4
Ironstone, impersistent, siderite mudstone with abundant fossils; silty in lower part		max.	10
Silty shale with ribs and laminae of fine sand, grading downwards into shale; occasional nodules at base		3	1
Unbedded soft grey clay			1
Shale		1	4
Ironstone, siderite mudstone, becoming increasingly silty downwards; abundant comminuted shell material at top; larger complete shells in lower part, some in growth position.			9
Dark fossiliferous sandy silt with small siderite nodules; silt-filled burrows at the base penetrate underlying beds			6
Soft grey shale (non-marine)	TOTAL	19	9

12. NEW WATH SCAR (818004)

Flat-bedded sandstone locally sideritic	9-12	
Thin-bedded laminated sandstone with partings of silty shale	4	0
Soft unbedded grey clay		½
Silty shale with unfossiliferous nodules of siderite mudstone		3
Soft umbedded grey clay		½
Silty shale with fine sand laminae	3	9
Ironstone, nodular fossiliferous siderite mudstone	max.	9
Shale		

CONT......

		ft.	in.
Shale			6
Soft unbedded grey clay			1
Shale		1	4
Ironstone, shelly siderite mudstone			8
Dark sandy silt, with silt-filled burrows penetrating beds below			4
Buff grey silty shales (non-marine)	TOTAL	23	0

13. <u>WHEELDALE GILL (788990)</u>

		ft.	in.
Sandstone, poorly bedded, thin bedded at base		10	0
Shale		1	5
Ironstone, siderite mudstone with occasional shells			3
Ironstone, shelly siderite mudstone, with much comminuted shell material; silty at base			10
Dark sandy silt with silt-filled burrows penetrating underlying beds			4
Pale grey silty shales (non-marine)	TOTAL	12	6

14. <u>GRAIN BECK (809007)</u>

		ft.	in.
Ferruginous sandstone with scattered shells		7	0
Ironstone, siderite mudstone with plant material			8
Thin-bedded sandstone, with occasional shelly layers, fine grained near base		5	6
Laminated silty shales with ribs of fine grained sand in upper part		1	0
Shale			5
Ironstone nodules, weathered		max.	$1\frac{1}{2}$
Shale, tough and silty near top		5	4
Ironstone, shelly siderite mudstone, silty at base			9
Dark sandy silts with small U-tubes at base			4
Buff-grey shales (non-marine)	TOTAL	20	3

15. EGTON GRANGE (774025)

		ft.	in.
Flat-bedded sandstone, fine grained at the base, medium grained above; top not seen	min.	6	6
Sandstone, medium grained occasionally cross-bedded; shelly and pebbly at base and scattered irregular beds of siderite mudstone		4	0
Shale		1	10
Ironstone, shelly siderite mudstone			6
Fossiliferous silty clay passing down into sandy siltstone with small U-tubes at the base			7
Buff-grey siltstone (non-marine)			

16. WINTER GILL (759020)

		ft.	in.
Massive sandstone, penetrated by rootlets		2	6
Dark laminated sandstone			7
Dark soft sandy silt			2
Pale-grey laminated sideritic sandstone			9
Black oolitic ironstone		3	9
Thin-bedded pale grey sandstone, shelly		1	10
Massive white sandstone, thin bedded at base; scattered rare fossils		6	0
Laminated micaceous silts, with shelly bands		3	3
Sandstone interbedded with siderite mudstone passing down into calcareous sandstone		4	0
Shelly and sandy sideritic limestone with scattered ooliths		1	0
Very shelly limestone			11
Sandy shelly silts with wood fragments			4
Shale, with unfossiliferous siderite mudstone nodules at top		1	4
Ironstone, shelly siderite mudstone		6	6
Dark sandy silt, with U-tubes at base			6
Pale grey silty shale (non-marine)	TOTAL	27	6

17. GLAISDALE HEAD (741015)

		ft.	in.
Thick-bedded sandstone			
Thin-bedded sandstone with silty partings		1	0
Shales with occasional sideritic nodules		4	0
Ironstone, shelly siderite mudstone			7
Dark sandy silt			5
Soft grey shale, weathered (non-marine)			

18. BIRK WATH BECK (734028)

		ft.	in.
Flat-bedded sandstone, rather massive in upper part		8	0
Thin-bedded sandstone		4	0
Shale, silty and fossiliferous in basal 3 in.		5	0
Ironstone, weathered			4
Greenish silts, fossiliferous			8
Buff silty shales (non-marine)	TOTAL	18	0

19. STONEGATE GILL (77082)

		ft.	in.
Dark laminated shale		c. 7	0
Pale grey silty shale with rootlets (non-marine)		1	1
Sandstone, thin and flat bedded with a few ripple-marked surfaces; penetrated by rootlets at top. Inclined U-tubes 1ft. 6in. from base		3	9
Shale, with occasional ironstone nodules; silty laminae and sandy ribs at top, passing rapidly into overlying sandstone		7	0
Ironstone, tough and dark with scattered shells and ooliths			10
Silty shale			2
Ironstone, soft and silty with occasional small shells and no ooliths			4

CONT........

	ft.	in.
Shaly silt with ironstone nodules at top, passing down into silty shale with occasional fossil casts		6
Pale grey shale with plant remains (non-marine) TOTAL	12	6

20. BUSCO BECK (759065)

	ft.	in.
Sandstone	7	0
Pale grey clay		4
Dark laminated carbonaceous silt with wood fragments		4
Pale grey laminated silt (non-marine)		6
Grey silty sandstone (c.8in) penetrated by rootlets and passing down into soft silts with irregular burrows	1	1
Soft sideritic silts, penetrated by rootlets, with burrows in lower part; occasional small oolitic lenses		9
Silty sandstone with wood fragments, penetrated by rootlets; passes down into laminated shale	2	4
Ironstone, nodular, with carbonaceous fragments	max.	4
Tough white fine-grained sandstone with burrows at the top		6
Shale with ironstone nodules and occasional lenses of tough fine-grained sandstone	6	9
Ironstone, silty siderite mudstone; small shells	1	5
Dark sandy silt, with U-tubes at the base		2
Pale grey clay (non-marine) TOTAL	13	0

21. GREAT FRYUPDALE– YEW GRAIN (718018)

		ft.	in.
Sandstone massive, ferruginous in part, unfossiliferous		12	0
Flat-bedded sandstone with occasional vertical burrows especially in the lower part; occasional shells		4	3
Thin-bedded sandstone with shelly layers; irregular silty burrows abundant		1	7
Thin-bedded sandstone with silty micaceous partings; some bedding planes crowded with shells	min.	2	3
Gap – probably very small			
Greenish sandstone, unfossiliferous	min.	2	6
Thin-bedded sandstone with greenish clay matrix, pebbly at top; shelly		3	0
Soft pale grey clay (non-marine)			
Blue-green silty sandstone with rolling base			
White to grey silty clay with plant material	TOTAL	25	6

22. GREAT FRYUPDALE – THE SCAR (713023)

		ft.	in.
Massive sandstone, flat-bedded		12	0
Sandstone with abundant burrows and shells		4	9
Thin-bedded sandstone with thin shell-beds		1	7
Silty shales with laminae of fine sand		2	6
Tough homogeneous grey sandstone		2	6
Sandstone with greenish clay matrix and occasional shells; pebbles at base			10
Pale grey silty shale		1	0
Shelly ironstone with small U-tubes at base			11
Poorly-bedded pale grey silty clay (non-marine)	TOTAL	27	3

23. DANBYDALE - OLD HANNAH'S NICK (701036)

		ft.	in.
Flaggy sandstone	min	7	0
Soft sandstone		2	9
Silty shale with laminae of fine white sand, and occasional lenses of close-grained greenish sandstone		1	6
Sandstone, ferruginous at top, with occasional shells		7	0
Gap -	max.	2	0
Grey carbonaceous silts (non marine)			

24. DANBYDALE - THE CRAGS (688027)

		ft.	in.
Soft greenish sandstone, thin-bedded	min.	2	0
Shelly oolitic ironstone, fine-grained siderite mudstone with green ooliths; burrowed			8
Green shelly sandstone with occasional U-tubes			5
Silts and sandy silts with occasional shells		1	9
Ferruginous silty sandstone			3
Green shelly sandstone, silty in lower part		1	2
Greenish-brown sandstone, fossiliferous			3-6
Poorly bedded shelly silts, pale purplish tint in parts		1	3
Shelly ironstone, fine-grained siderite mudstone			5
Soft sandstone with reddish stained laminae (non-marine)			9
Thin-bedded sandstone with sun-crack and footprint horizon		2	0

25. DANBYDALE - SANDSTONE QUARRY (684030)

	ft.	in.
Sandstone, thick-bedded	4	0
Flat-bedded ferruginous sandstone with scattered green ooliths	1	6
Massive ferruginous sandstone with occasional rubbly bands comprising wood fragments and shell casts	3	6
Thin-bedded ferruginous red weathering sandstone		
with lenses composed of shells and wood fragments; base mildly erosional	2	8
Flat-bedded ferruginous sandstone, red weathering in part, with scattered ooliths; partings (2in. maximum) developed which consist almost entirely of flattened ooliths	1	6
Flat-bedded ferruginous sandstone with shells	2	0
Thin-bedded shelly sandstone min.	1	0

26. WESTERDALE QUARRY (674026)

Sandstone	1	6
Flaggy sandstone, reddish-brown, with green ooliths	2	6
Thin-bedded sandstone	2	0
Dark red flaggy sandstone with ooliths and shells, especially near the top; more massive in lower part	2	8
Thin-bedded sandstone with partings of silty shale min.	1	6

27. COMMONDALE (662807)

	ft.	in.
Thin and rather irregularly bedded sandstone with abundant partings of dark silty shale; nodular sandy ironstones in lower part; at top sandstone is penetrated by rootlets; unfossiliferous	8	0
Ironstone, silty siderite mudstone with large shells locally abundant		9
Dark thin-bedded sandstone, medium grained with silty partings; fossiliferous	1	9
Soft dark silty sandstone passing downwards into unbedded clay	1	0
Ironstone, pale grey siderite mudstone; nodular with well defined laminae of sand; unfossiliferous. max.		2
Sandstone, dark and medium grained, alternating with finely laminated shale with rare shell casts. Sandstone ribs poorly bedded, and show signs of burrowing activity within the sandstone and also from above (shown as small shale filled burrows).	2	6
Tough dark sandy silts, poorly bedded passing down into dark laminated silty shale; shell casts common	1	1
Tough finely laminated sandstone penetrated by silt-filled U-tubes; also partings of dark silty shale not apparently related to burrows; pass downwards into pale silty shale and thin fine-grained sandstone ribs, (non-marine with perhaps marine intercalations at top).		
TOTAL	15	0

28. BATTERSBY CRAGS (605065)

		ft.	in.
Flat-bedded sandstone, thin bedded in lower part; few sideritic horizons with scattered shells	max.	18	0.
Soft pale grey siderite mudstone with sandy laminae	max.	1	0
Sandstone, impersistent, with irregular silt-filled burrows	max.	2	0
Pale grey silty shales (non-marine)			
	TOTAL	21	0

29. INGLEBY ESCARPMENT - BURTON HOWE (607025)

Ironstone, shelly oolitic siderite mudstone, highly oolitic at top	1	8
Silty shale		11
Ironstone, micaceous silty siderite mudstone, fossiliferous		11
Soft shelly silts with purplish tint, and tough dark grey shelly silty shale at base	1	10
Ironstone, fine-grained shelly siderite mudstone		4-5

30. INGLEBY ESCARPMENT - BLUEBELL TROUGH (608020)

	ft.	in.
Coaly shale (non-marine)		
Tough fine-grained whitish sandstone, thin-bedded, unfossiliferous	3	6
Blocky grey silts passing rapidly downwards into dark laminated silty shales with internally cross-laminated lenses of tough fine-grained sandstone	2	6
Shale, finely laminated, unfossiliferous apart from occasional _Zamia_ leaves, interlamination at base with pale purplish silty shale with shells	4	6
Ironstone, shelly oolitic fine-grained siderite mudstone	1	2
Pale purplish unbedded silts, shelly; non-oolitic siderite mudstone nodules with shells, near top.	2	8
Ironstone, shelly fine-grained siderite mudstone		3
Grey blocky micaceous silts with _Baciria_; highly carbonaceous lenses near top (non-marine) TOTAL	14	6

31. FARNDALE (626002)

	ft.	in.
Tough fine-grained thin-bedded sandstone, unfossiliferous min.	1	6
Fissile dark grey shales, silty at top, unfossiliferous	5	9
Ironstone, shelly oolitic fine-grained siderite mudstone		8
Soft unbedded pale purplish shelly silts	1	6
Ironstone, shelly fine-grained siderite mudstone		5
Soft grey silts with _Baciria_ and rootlets (non-marine)		

32. ROSEDALE - CASTLE CRAG (686008)

	ft.	in.
Thin-bedded ripple-marked sandstone with inclined U-tubes and occasional shells; flakes of silty shale on one horizon min.	2	0
Dark grey Shale, with laminae and ribs of fine-grained sandstone in upper part. At least six nodule horizons, some with rare shells (up to 4ft. 9in. from base	7	8
Dark silty shale with shells (2ft. 8in) passing down into poorly-bedded grey-green silts	3	6
Soft shelly silts with ironstone, slightly oolitic siderite mudstone (I) at top; small U-tubes at base	2	6

33. ROSEDALE - BERKING GILL (692007)

	ft.	in.
Sandstone with small-scale trough cross-bedding from NE; unfossiliferous, penetrated by rootlets at top	13	0
Dark grey silty shale with sand laminae passing down into shale	1	3
Ironstone, pale grey unfossiliferous siderite mudstone		2
Shelly sandstone		2
Grey silty shale	1	2
Sandstone, silty in lower part		9
Shale, silty laminae in upper part	1	6
Unbedded silty shale with unfossiliferous siderite mudstone 3 in from top		9
Ironstone, pale grey shelly and slightly oolitic siderite mudstone (I)		4
Grey silty micaceous shale (7in). passing down into soft pale fossiliferous silts		10
Ironstone, fossiliferous siderite mudstone (B)		4
Soft grey unbedded silt with occasional nodules of pale grey siderite mudstone; shelly		9
Tough micaceous carbonaceous silt with wood fragments and occasional shells		6

Cont.........

	ft.	in.
Soft grey shale with plant material (non-marine) TOTAL	21	6

34. ROSEDALE CRAGS (701967)

	ft.	in.
Sandstone, fine-grained and thin-bedded at the base; penetrated by rootlets	11	0
Dark grey shale with laminae and small lenses of sand near the top; rare ironstone nodules	7	0
Ironstone, pale grey slightly oolitic siderite mudstone with intraclasts of soft siderite mudstone (I)		5
Dark, poorly-bedded fossiliferous silty shale	1	0
Grey ferruginous silty sandstone, soft and poorly-bedded, fossiliferous in upper part with impersistent siderite mudstone (B)	1	8
TOTAL	21	0

35. ROSEDALE MINES (S.E.) (728945)

	ft.	in.
Sandstone, massive weathering, penetrated by rootlets	10	0
Dark grey silty shales with sandy laminae, passing downwards into shale		
Ironstone, shelly and slightly oolitic		
Poorly-bedded sandy shelly silts		

36. ROSEDALE - NORTHDALE (generalised)(719995) (723992) (725994)

	ft.	in.
Massive sandstone, penetrated by rootlets	4	0
Flat-bedded ferruginous sandstone with occasional shells sideritic and oolitic developments common at any level	8	0
Thin-bedded sandstone with fossiliferous bands; bedding may be destroyed by extensive burrowing	5	9
Dark grey shale, with sand laminae in upper part and ironstone nodules throughout; fossiliferous in lower part	4	6
Ironstone, impersistent fossiliferous slightly oolitic siderite mudstone max.		3
Silty shale, fossiliferous, with poorly-bedded silt at base		6-18

Cont.........

	ft.	in.
Shelly unbedded silts with impersistent fossiliferous ironstone at top; small U-tubes at base	1	0
TOTAL	24	3

37. HARTOFT BECK (757960)

	ft.	in.
Sandstone, thin-bedded with abundant partings of silty shale and occasional ironstone bands. 1 - 4 ft. non-shaly sandstone at top	6-7	
Shale, silty in upper part; scattered ironstone nodules	c 5	0
Horizon of laterally extensive siderite mudstone nodules		2½
Shale, laminated	1	3
Soft dark poorly-bedded silty shale with fossils; abundant small burrows		6
Slightly oolitic siderite mudstone with occasional shells (I)		2-3
Soft silty shale		1-3
Soft silty siderite mudstone with shells, chiefly gastropods		2
Shale		8
Ironstone, silty siderite mudstone with abundant comminuted shell material and occasional larger shells, including <u>Pholadomya</u> in growth position. Impersistent nodular nodular ironstone with scattered ooliths at top		8
Soft grey silty shale with plant remains		
TOTAL	15	9

38. BLOW GILL (530940)

	ft.	in.
Sandstone, fine-grained and flaggy	3	6
Shale	c.12	0
Ironstone, soft and clayey in upper part dense and sideritic at base; abundant <u>Pholadomya</u> and <u>Pleuromya</u> in vertical position 2ft. 9in to	3	6
Dark silty shale with shell casts; pebbles at base	1	4
Pale grey shale (non-marine)		
TOTAL	20	0

APPENDIX 2 LABORATORY METHODS

A. MECHANICAL ANALYSIS

(i) Sample is broken up by light hammering and is then crushed gently in a pestle and mortar, the disaggregated material being sieved off from time to time. Crushing is continued until all grains are disaggregated.

(ii) The disaggregated material is then transferred to a crucible and heated with 50% HCl, to remove iron carbonate and oxide, and calcite.

(iii) The material is then filtered and washed thoroughly and sieved on a 230 mesh sieve by means of a jet of water.

(iv) The sand retained on the sieve is dried and transferred to a stack of sieves ranging from 230mesh to the largest mesh required, and is sieved on a sieve shaker for ½ hour.

(v) Material which passed through the 230 mesh sieve in stage 3 is washed on successively finer sieves and the several fractions dried and weighed. The suspension which passed through the 400 mesh sieve was dried and weighed. Size analysis was not carried out beyond the limit of sieving.

B. X-RAY DIFFRACTION

The material to be analysed was ground to a fine powder with an agate pestle and mortar, and the powder transferred into a beaker of water and stirred into suspension. The suspension was then allowed to settle for about an hour so that the larger and heavier particles settle out. A glass slide was then placed horizontally within the remaining suspension and left until a thin coating of clay had accumulated. The liquid was then pipetted off as rapidly as possible and the slide removed and dried.

The X-ray traces of some of the clay samples examined are given in figure A.1. Figure A.2 shows the comparison of the diagnostic peaks of dickite and kaolinite. Figure A.3 gives the trace from which feldspar was confirmed from the Winter Gill ironstone. The key/to the trace numbers is as follows:

(i) Interstitial clay from channel sandstone (kaolinite)

(ii) Clay from non-marine shale (kaolinite)

(iii) Insoluble residue from non-marine ironstone concretion (kaolinite).

(iv) Clay from marine shale. (kaolinite).

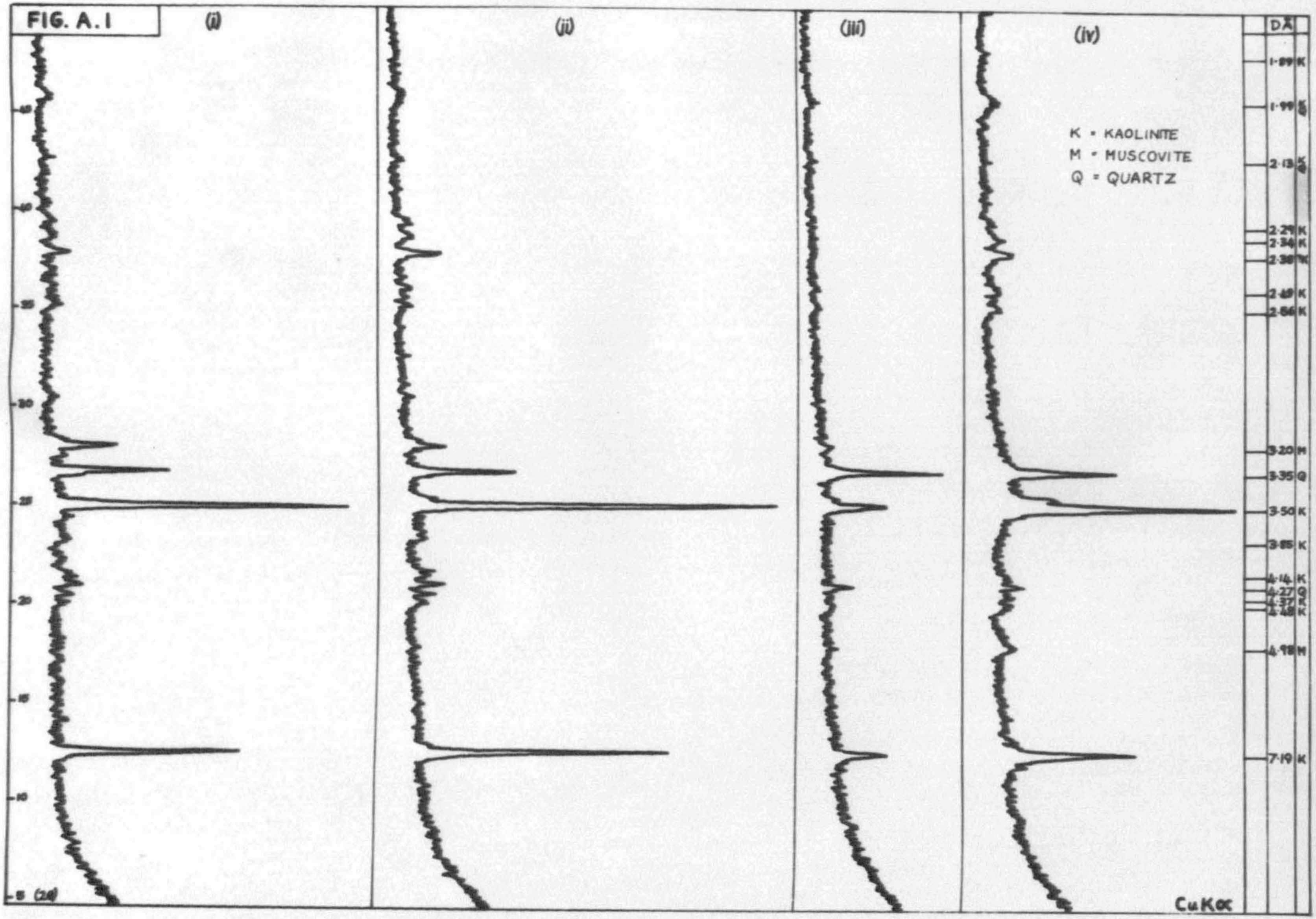
(v)a Kaolinite, basal ironstone, Iron Scar.
b Kaolinite basal ironstone, Egton Grange.

(vi)a Dickite nodular ironstone, Eller Beck
b Dickite basal ironstone, Grain Beck
c Dickite sandstone unit, Grain Beck

(vii) Dickite with kaolinite, sandstone unit, New Wath Scar.

(viii) Albite with quartz (insoluble residue), top of Winter Gill Ironstone.

FIG. A.1



(i)
(ii)
(iii)
(iv)
K = KAOLINITE
M = MUSCOVITE
Q = QUARTZ
DÅ
1·89 K
1·99 K Q
2·13 K Q
2·29 K
2·34 K
2·38 K
2·49 K
2·56 K
3·20 M
3·35 Q
3·50 K
3·85 K
4·14 K
4·27 Q
4·37 K
4·48 K
4·98 M
7·19 K
CuKα
5 (2θ)
10
15
20
25
30
35
40
45

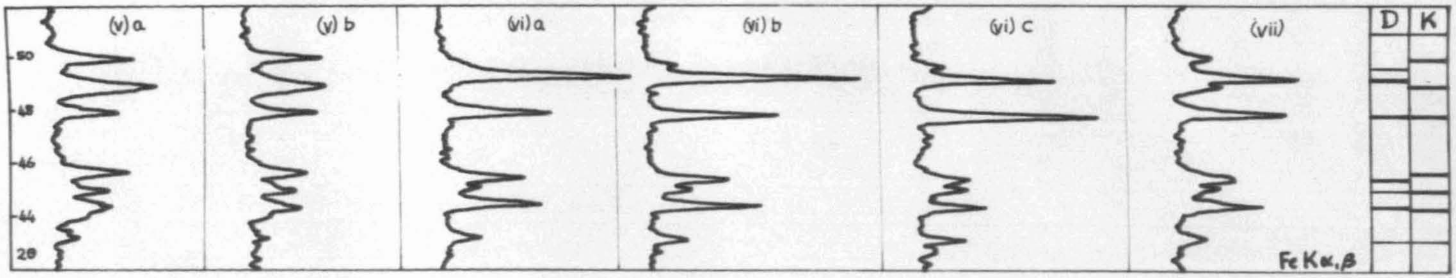

(v) a
(v) b
(vi) a
(vi) b
(vi) c
(vii)
D
K
50
48
46
44
2θ
Fe Kα, β

FIG. A.2

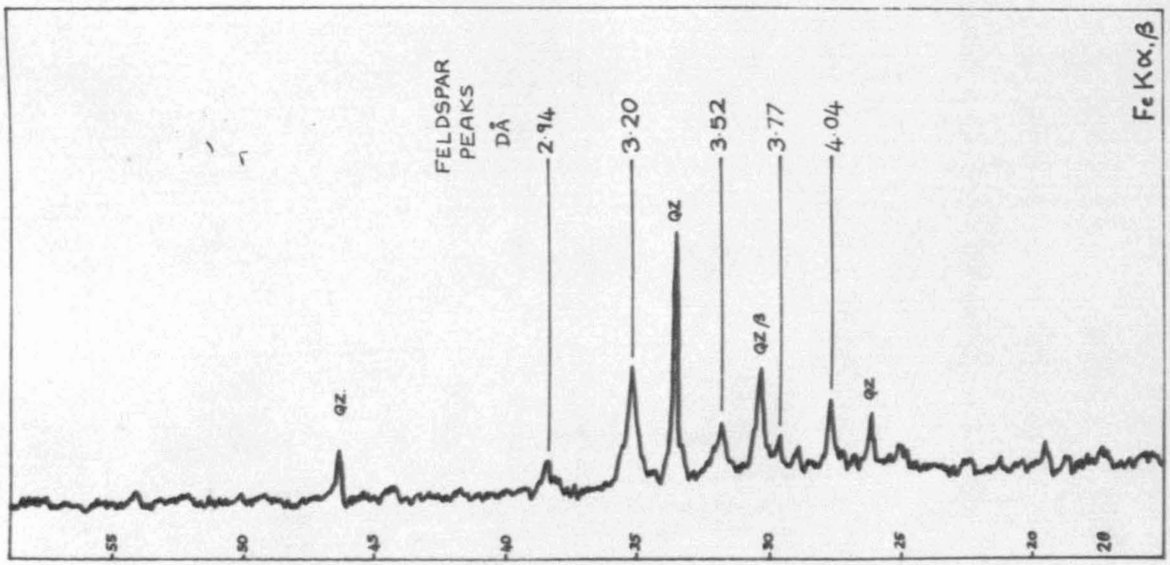

FELDSPAR
PEAKS
DÅ
2·94
3·20
3·52
3·77
4·04
QZ
QZ β
Fe Kα, β
55
50
45
40
35
30
25
20
2θ

FIG. A.3

C. X-RAY PHOTOGRAPHY

X-ray photography was carried out on two single ooliths from the Winter Gill ironstone. These were obtained by crushing the rock and examining it under a binocular microscope in order to select ooliths which showed the least amount of marginal sideritisation. The dA values obtained and the line intensities are given below and compared with values given by Brindley (1951) for orthogonal 7Å chamosite.

BRINDLEY (1951)			OBSERVED	
I (CALC)	hkl	d(A)	I	d(A)
			M	14.05
70	001	7.026	VS	7.04
21	020	4.457	M	4.65
25	021	3.880		
38	002	3.513	S	3.53
7	022	2.804	VW	2.82.
8	200	2.688	VW	2.69
88	201	2.511	S	2.51
2	041	2.210		
48	202	2.135	M	2.13
3	042	1.941		
30	203	1.766	M	1.77
2	240	1.760		
3	241	1.707		
2	043	1.651		
21	060	1.552	S	1.55
11	061	1.516	M	1.51
23	204	1.471	W	1.474
6	062	1.420	W	1.421
3	005	1.405		
2	044	1.403		
2	400	1.344	VW	1.342
14	401	1.320	W	1.321

No impurites could be detected on the photographs and it would seem that the 14A line is a feature of the chamosite itself. The 003 reflection of chlorite is not developed and it would therefore appear that, in part, the chamosite has a 14A structure as described by Von Englehardt (1942). The diffuse nature of the 201 reflections suggests a certain amount of disorder as described by Youell (1955).

D. WET CHEMICAL METHOD FOR PHOSPHATE DETERMINATION

About 1 gm. rock powder is weighed out into a crucible and 15 cc. conc. HCl and 5 cc. conc. HNO_3 added; this mixture is then evaporated to dryness. Evaporation is repeated twice with small quantities of conc. HNO_3 to decompose fluorides. A few drops of ammonium nitrate are added before the final evaporation to aid in driving off HF.

When completely dry the crucible is heated until the contents become brown, and then cooled. The crust is moistened with 10cc. 1:4 HNO_3 and the liquid gently boiled for a few minutes. The liquid is then filtered (No.40) into a 250cc. beaker. The crucible is raised and the filter washed six times with some warm dilute acid, making sure that the bulk of the liquid does not exceed 50cc.

About 25cc ammonium nitrate solution (340 gm./l.) is added and the liquid heated to boiling. A hot mixture of 20 cc. each of ammonium molybdate solution (30 gm./500cc warm water) and 1:3 nitric acid is added with constant stirring. The liquid is filtered and the filtrate tested with more ammonium molybdate and any precipitate filtered off. This process is repeated until no further precipitate comes down. The precipitate in the filter is then washed with a 1:1:1 solution of ammonium nitrate HNO_3 and water, washing down the sides of the beaker until the addition of ammonia in excess produces no permanent precipitation in a few drops of filtrate placed in a watch-glass. About 50cc of washing mixture should suffice.

The filter paper is filled with 1:1 NH_4OH and put the dissolved precipitate into the original beaker. Ammonium hydroxide is add d until solution is complete. 10 cc magnesia mixture is added, through the same filter, to the solution in the beaker and stirred vigorously to produce a crystalline precipitate. The beaker is allowed to stand for 12 hours and the contents then filtered through a small paper, the material adhering to the sides of the beaker being rubbed off. The precipitate is washed well in 3% ammonia solution.

The filter paper and its contents are placed in a weighed crucible, carbonised and then ignited at bright red heat. The weight of the yellow precipitate multiplied by a factor of 0.638 gives the weight of P_2O_5 present in the original sample.

E. STAINING METHOD FOR CARBONATES IN THIN SECTION (modified from Dickson, J.A.D. (1965), Nature, 205, p.587.)

Reagents

A: solution 0.2g alizarin red S per 100c.c. 1.5% HCl.

B: solution 2.0g potassium ferricyanide per 100c.c. 1.5% HCl.

C: solution 1.0g alizarin red S per 100c.c. water.

Solutions A and B are mixed in the ratio A:B = 3:2. The thin section is immersed in this mixture for 20 seconds, and is then removed and washed gently. Staining is carried out for a further 5 to 10 seconds in solution C, and followed by a second washing.

Etching of the slide before staining was not carried out as this causes a very pronounced relief between minerals of contrasting solubility, which leads to focussing difficulties in photography of the thin sections, especially at high magnification. Incipient cracking of the blue stain seems to be common; the second stage of the staining is therefore carried out in aqueous, rather than acid, solution, in order to prevent accentuation of this cracking by effervescence.

9

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ADDENDUM - 'OPAL' IN CHAMOSITE AND KAOLINITE OOLITHS.

Sectioning of oolitic ironstones subsequent to the preparation of much of this thesis threw some light on the feature previously ascribed to the presence of finely divided opal (pages 126, 137,etc.). In order to avoid oxidation of the siderite matrix some sections were cover-slipped quickly and at low temperature, using Canada Balsam. Ooliths, which in previous sections of the same rock had appeared clear, were to a large extent opaque, and white in reflected light. On heating and the old Canada Balsam being scraped off and replaced with fresh material, the opacity disappeared. After several experiments it was decided that the opacity was caused by minute particles of air being trapped within the porous chamosite envelopes because at the time of cover-slipping the balsam had not become sufficiently fluid to displace the air. The effect of trapping of air is similar to that seen when the section is examined dry, prior to cover-slipping. Sections should therefore be left on the hot plate with the balsam until all opacity is seen to have disappeared. Alternatively mounting may be carried out at low temperatures with more fluid media, such as Caedax or Depex. There is, however some indication that the Depex reacts with the Lakeside mounting medium as some ooliths cracked and tended to float up out of the plane of section.