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MINERALOGY OF YOREDALE SERIES ROCKS IN UPPER TEESDALE WITH SPECIAL REFERENCE TO CLAY MINERALS

By

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ABSTRACT

A description of the stratigraphy of the area studied is given in which the variations in lithology and sequence are illustrated.

The methods of clay mineral analysis are discussed and it is concluded that X-ray analysis is most useful for the present work.

The argillaceous rocks are classified into types on characteristics recognisable in the field and the results of petrographic and X-ray analyses are tabulated. Apart from kaolinite, which is post-depositional in origin, only three different clay minerals occur: illite, chlorite and mixed-layer.

Different types of argillaceous rock have different clay mineral assemblages.

Associated arenaceous and calcareous rocks are similarly classified on field characters and are analysed petrographically. A system of sandstone classification is proposed.

The relationship between the clay mineralogy of a sedimentary rock and the origin, transport, deposition and post-depositional history of the minerals is discussed.

Recent shallow water marine and deltaic sedimentation are reviewed and the various facies recognised in the Gulf of Mexico and the Mississippi Delta are enumerated.

It is postulated that the Yoredale sediments were deposited in a shallow water marine environment from a series of small shoal-water deltas. The different sediment types and their mineralogies are related to different conditions in the various facies of the environment. Repetition of the sedimentation in the form of cycles is discussed briefly.

An additional chapter deals with the igneous rocks intruding the sediments and the metamorphic effects of the Whin Sill upon the country rocks.

Finally, the distributions of the various sedimentary and metamorphic sulphides in the country rocks are discussed.

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INTRODUCTION

A. AIMS OF THE PRESENT INVESTIGATION

It has been suggested by Millot (1952), Miller (1954) and Potter and Glass (1958) that the petrography and mineralogy of a series of sediments are dependent on the characters of the source areas of the detritus and the environments of deposition of the sediments.

Although shales form a large percentage of the sedimentary column (Pettijohn, 1957, gives the following figures:- limestone 20%, sandstone 22% and shale 58%), they have, until recent years, been much neglected in studies of ancient sediments. This is principally due to the difficulties encountered during use of ordinary petrographic methods.

The aim of the present investigation has been to study the mineralogy of some of the shales from the Carboniferous rocks of the Yoredale facies in Upper Teesdale so as to provide more information on the composition of shales in general. In addition, by combining a study of the shales with a petrographic study of the associated rocks, an attempt has been made to provide a more detailed picture of the depositional environment in which the rocks were formed.

The samples studied were collected from an area of Upper Teesdale, mapped on a 6":1 mile scale, from which detailed evidence of the rock sequences and associations were obtained. The map is reproduced on a 2½":1 mile scale (Figure 0.1).

B. GENERAL GEOGRAPHY AND GEOLOGY OF THE AREA STUDIED

Field observations have been made in an area of over eighty square miles of Upper Teesdale, on the borders of Yorkshire, Durham and Westmorland. The main part of the area studied lies between Mickleton (National Grid Reference: 965237), Mickle Fell (805245), Meldon Hill (772290) and Three Pikes (835343).

The area is dissected by the wide valley of the River Tees and Harwood Beck which trends W.N.W. and the south-western part of the area is cut by the Maize Beck and River Tees drainage system which follows an E. to N.E. direction.

The Tees valley rises from about 700 ft. at Mickleton to over 1500 ft. above Harwood. The fells bordering Teesdale are generally over 1500 ft. with summits over 2000 ft., Mickle Fell, the highest, being 2591 ft.

Most of the area of Teesdale under consideration has a partial cover of Pleistocene and Recent unconsolidated sediments; boulder clay on the valley sides and peat on the fell tops. East of High Force, the Tees valley is floored by alluvium.

There has been extensive working for minerals in the past and although mining is no longer active in Teesdale, shaft and quarry records provide much valuable information.

Previous workers in the area have studied the Lower Palaeozoic Inlier at Cronkley (Gunn and Clough, 1878) and the Whin Sill and its metamorphic effects (Sedgwick, 1827, Clough 1876, Hutchings 1895 and 1898, Smythe 1924, Holmes and Harwood 1928, and Wager 1928).

The Cleveland Dyke has been described by Teall (1884) and again by Holmes and Harwood (1929).

Upper Teesdale was mapped for H.M. Geological Survey by C.T. Clough, maps on 1":1 mile scale being published in 1887.

More recently, the basement beds of the

Carboniferous were examined by Harry in 1950 and some of the beds studied in the present investigation were examined by Jones (Ph.D. thesis, 1956).

Detailed stratigraphical and mining data is to be found in the Northern Pennine Orefield Memoir, vol. I, Dunham (1948), pages 14-25 and 276-318.

In this part of Upper Teesdale, the strata consist of the Carboniferous Limestone Series which rest unconformably on older Palaeozoic rocks and which have a capping of "Millstone Grit". Unconsolidated Pleistocene and Recent beds overly these rocks. The older, Lower Palaeozoic, rocks are cut by minette dykes which are probably of Devonian age. The Carboniferous rocks contain the Carbo-permian quartz dolerite intrusion, the Whin Sill, and are cut by a porphyritic tholeiite, the Cleveland Dyke, which is of Tertiary age. The generalised stratigraphical column can be given as follows -

Unconsolidated Sediments

	Thickness in feet
Recent	
Hill peat	up to 10
Fluviatile alluvium	variable
Pleistocene	
Boulder clay	up to 180

Consolidated sediments

	Thickness in feet
Carboniferous	
Millstone Grit. Coarse sandstones, grits and shales.	up to about 300
Carboniferous Limestone Series. The beds which make up this series of strata are divided into four groups -	
Upper Limestone Group: coarse grits, sandstones, ganisters, thin coals, shales and limestones.	580 - 780
Middle Limestone Group: rhythmic alternation of limestones, shales, sandstones and thin coals.	725 - 850
Lower Limestone Group: Massive limestone, overlain by alternating limestones, shales and sandstones.	300 - 550

Basement Group: Conglomerates overlain by sandstones and shales. about 100

∿∿∿ Unconformity

Lower Palaeozoic.

Slates, possibly Skiddaw Slate Series, and volcanic rocks.

CHAPTER I

The Stratigraphy

a. INTRODUCTION

The beds studied consist of a succession of alternating limestones, shales and sandstones, which form the upper half of the Middle Limestone Group between the base of the Tyne Bottom Limestone and the base of the Great Limestone. The beds are repetitive in character and are of a rhythmic or cyclothemic nature.

b. THE CYCLOTHEM

Previous workers, studying the Carboniferous strata in which deposition has occurred in rhythms, have set out examples of the "typical" cyclothem as it appears in the area which they have studied, viz. Hudson (1924), Brough (1929), Wanless and Weller (1932), Robertson (1948), Dunham (1948,1950), Moore (1959) and Johnson (1959).

Since the concept of a cyclothem has been interpreted in various ways by different authors it is necessary to consider the usage of this term.

According to Wanless and Weller (1932) who first used it, "The word cyclothem is therefore proposed to designate a series of beds deposited during a single sedimentary cycle of the type that prevailed during the Pennsylvanian Period". Much earlier, Miller (1887), writing of the Calcareous Division of the Upper Limestone Series of northwest Northumberland had stated, "The alternations of these beds fall into a certain orderly sequence which marks them out into zones or cycles .... The interval between any one limestone and the next in order above it is thus indicated as forming ....a complete cycle of deposition".

In the northern Pennines, of which Teesdale forms a part, the existence of beds which exhibit a cyclothemic nature has long been recognised. Phillips (1836) obviously recognised the repetitive character of the Yoredale beds although he was not concerned with its significance.

Dunham (1948) illustrated the cyclic character of the beds in the present area of investigation and noted that the cycles were very similar to those described by Wanless and Weller for the Pennsylvanian.

Dunham's "typical" cyclothem consisted of the following succession of beds, starting with the limestone at the base: limestone, calcareous shale, ferruginous shale, sandy shale or shaly sandstone, sandstone, ganister or underclay, coal.

Although the "typical" cyclothem varies from area to area and although different workers may accentuate different aspects of the cyclothem, they are all essentially of the same nature and in their simplest form are very similar to Dunham's cyclothem.

It appears from a consideration of the diagrams and descriptions featured by the afore-mentioned workers, that their "typical" cyclothems are subject to much local lateral variation. This is certainly true of the beds studied in Teesdale. The generalisation of the characteristics of the Yoredale cyclothems and the setting out of "typical" cyclothems, has caused workers to overlook the local variability when it comes to a closer understanding of the conditions under which the sediments were deposited.

The general features of the cyclothems are important in understanding the broader aspects of deposition. The "typical" features are applicable only to the major cyclothems, then not over the whole area, and to some of the minor cyclothems.

The limestones, although generally thin beds, cover very large areas and remain at fairly constant distances apart. The limestones are usually overlain by a shale which is succeeded by a sandstone. Whenever a major sandstone occurs in a cyclothem, its position is generally constant. The coals, wherever they are preserved, usually overlie the main sandstone in the succession and occur at or near the base of a main limestone.

As will be illustrated in the succeeding pages, there are considerable local lateral variations and also many differences in the rock types and their petrography. These can be related, as they are in a later chapter, to differences in the environments of deposition.

Following Dunham (1950), the bases of the cyclothems recognised in the Teesdale area are taken as the bases of the major limestones. The basic structure of the hypothetical, ideal Teesdale cyclothem can be given as:-

7. Coal
6. Ganister or underclay
5. Sandstone
4. Sandy shale, shaly sandstone or shale with thin sandstone bands.
3. Unfossiliferous (? non-marine) ferruginous shale.
2. Marine shale
1. Marine limestone

c. NOMENCLATURE OF THE CYCLOTHEMS

In Teesdale, several cyclothems are recognised in the series of beds studied. These are named after the limestone occurring at the base. Some of the more prominent sandstones in the succession are also named.

Two local terms are frequently encountered in the nomenclature of the beds studied in this area of England. These are the names "hazel" or "hazle" for sandstones and "plate" for shale beds. The term "grey beds" is also often used and indicates a series of thin shales with flaggy sandstone bands.

In some cyclothems there are minor rhythms which are often marked by the presence of thin limestone bands.

The following table indicates the nomenclature of the beds studied and shows that they are divided into five cyclothems:

		Great Limestone.
Four Fathom Cyclothem		Tuft or Water Sill (sandstone)
		Shale (not named)
		Iron Post Limestone
		Quarry Hazle (sandstone)
		Shale (not named)
		Four Fathom Limestone
Three Yard Cyclothem		Nattrass Gill Hazle (sandstone)
		Shale (not named)
		Three Yard Limestone
Five Yard Cyclothem		High Brig Hazle (sandstone)
		Shale (not named)
		Five Yard Limestone
Scar Cyclothem		Low Brig Hazle (sandstone)
		Shale (not named)
		Scar Limestone
Tyne Bottom Cyclothem	Alternating Beds	Copper Hazle (sandstone)
		Shale and sandy shale (not named)
		Cockle Shell Limestone
		Shale and sandy shale (not named)
		Single Post Limestone
		Tyne Bottom Plate (shale)
		Tyne Bottom Limestone

d. VARIATIONS IN THE STRATA

Within the mapped part of Upper Teesdale, exposure is variable. West of High Force (880284), the strata are very poorly exposed and the most valuable sections occur east of the High Force. Even to the east of High Force, the exposures of beds are poor except for those which lie between the Scar and Three Yard Limestones. Dunham (1948) dealt with the stratigraphy of the Yoredale strata in the area studied and Jones (Ph.D. thesis, 1956) gave a fairly comprehensive list of sections from Bow Lee Beck eastwards. For these reasons, the following discussion of the variation of the strata will be based principally on an analysis of the strata between the Scar and Three Yard Limestones and of the Cockle Shell Limestone.

Field observations have shown that there is a considerable variation in any one cyclothem over the whole area and that even closely spaced stream sections may yield very different pictures. Examples of the variation can be seen in the stratigraphical columns given in Figures I,A,1,2 and 3. Whilst the thickness and the character of the limestones vary slightly, the most noticeable effects are in the sandstone and shale content of the cyclothems.

e. VARIATIONS IN THE STRATA BETWEEN THE SCAR AND THREE YARD LIMESTONES

The first fairly complete section of the strata occurring between the Scar and Three Yard Limestones in the western part of the area is seen in Langdon Beck (858314). Here, the succession is at least 86 feet thick and consists of:-

	Three Yard Limestone	
c. 5' 0"	Fine nodular mica shale.	
1' 4"	Coal	
18' 0"	Brown false bedded sandstone-High Brig Hazle	
4' 0"	Calcareous fossiliferous shale	
16' 6"	Fine dark limestone with shale partings - Five Yard Limestone	
	∿∿∿ Erosion surface	
+15' 0"	Massive white sandstone-Low Brig Hazle	
	? Gap	
+25' 0"	Pale crinoidal limestone)	Scar Limestone
1-2' 0"	Arenaceous limestone)	
	on massive false bedded brown sandstone.	

Two and a half miles to the south-east, in Ettersgill (883309), the succession is considerably changed. The Scar Limestone is reduced in thickness but is otherwise little altered. However, the Low Brig Hazle is apparently much thicker and the High

Brig Hazle consists of two sandstone beds separated by a shale sequence which contains a thin coal. Owing to lack of exposure in the intervening ground, it is not possible to say which of the sandstone beds in the High Brig Hazle in this section is equivalent to the High Brig Hazle in Langdon Beck.

The succession between the Three Yard Limestone and the coal below it is notably different in that it is much thicker and contains a sandstone bed. The total thickness of the sequence is about 136 feet and is as follows:-

	Three Yard Limestone	
1' 7"	Brown Sandstone	High Brig Hazle
c.1' 0"	Clayey siltstone	
18' 0"	Massive false bedded sandstone containing clay galls, and with shale fragments in the lower surface.	
9' 0"	Fine mica shale	
0' 9"	Coal	
15' 0"	Massive false bedded sandstone	
c.1' 0"	Shelly calcareous shale	
18' 4"	Fine dark limestone with occasional shale partings - Five Yard Limestone.	
1' 2"	Pale brown sandstone	
1'10"	Micaceous shaly sandstone	
2' 6"	Mica shale with thin sandstone bands.	

18'10"	Massive false bedded and rippled sandstone.	Low Brig Hazle
12' 0"	Mica sandstone with trails and ? algal ? markings, and pit and mound structures.	
3' 0"	Very fissile mica sandstone.	
10' 0"	Fissile mica shale with thin sandstone bands.	
3' 0"	Fossiliferous nodular shale.	
23' 0"	Pale crinoidal limestone - Scar Limestone.	

On shale.

One and a quarter miles to the east of Ettersgill, in Hell Cleugh (910294), the succession has again changed. The shale above the Scar Limestone is much thinner and a minor rhythm occurs above the Low Brig Hazle. The Five Yard Limestone is more shaly. The High Brig Hazle again consists of two sandstone beds. Here, the coal below the Three Yard Limestone is overlain by a nodular shale as in Langdon Beck but this is overlain in turn by a sandstone bed similar to that in Ettersgill.

The total thickness of strata in this section is about 135 feet and consists of:-

Three Yard Limestone.

6-8' 0"	Rotten siltstone with sandstone bands to mica sandstone.	High Brig Hazle.
0-9"	Shaly sandstone.	
1' 6"	Sandstone.	
11' 8"	Black silty mica shale with small pyrite nodules.	
1' 4"	Coal.	
0' 6"	Black carbonaceous shale.	
26' 0"	Soft brown sandstone.	
9"-1'4"	Shale - (obscured).	
17' 0"	Fine dark limestone with shaly partings, lower 1'8" arenaceous.	Five Yard Limestone
2' 8"	Fine sandstone.	
3' 6"	Arenaceous grey shale.	
1' 6"	Pyritous sandstone.	
0' 9"	Arenaceous grey shale.	
0' 9"	Shaly sandstone.	
3' 0"	Obscure.	
3' 0"	Ganister.	
2' 0"	Sandy shale.	
3' 0"	Shaly sandstone.	
30' 0"	Massive sandstone with clay galls in lower 4 ft.	Low Brig Hazle
5' 0"	Mica shale with occasional fossils.	
+15' 0"	Pale crinoidal limestone with top 2 feet shaly - Scar Limestone.	

One half to three quarters of a mile to the south, in Bow Lee Beck, there is only one sandstone forming the High Brig Hazle and the minor rhythm above the Low Brig Hazle is absent, being replaced by a shale succession. The beds immediately below the Three Yard Limestone are not well exposed but probably consist of shale. The total thickness of the cyclothem here is about 150 feet. The succession can be given as follows:-

Three Yard Limestone

15' 0"	Obscured - ?shale?	
25' 0"	Massive false bedded sandstone. High Brig Hazle.	
2' 6"	Obscured - ?shale?	
2' 0"	Calcareous and fossiliferous shale	
18' 9"	Fine dark limestone with numerous shaly partings, base arenaceous.	Five Yard Limestone
6' 0"	Fissile mica shale with fine limestone bands.	
3' 0"	Fine mica shale with fine sandstone lenses.	
1' 0"	Unlaminated calcareous shale and carbonaceous shale.	
28' 0"	Massive sandstone, top false bedded with stigmaria and base calcareous.	Low Brig Hazle
5' 0"	Ferruginous, silty, nodular mica shale.	
5' 0"	Silty nodular mica shale with fossils.	
29' 8"	Massive pale crinoidal limestone, shaly towards the top and with an arenaceous base.	Scar Limestone.

A half mile to the south-east, in Newbiggin Beck, the succession has again changed considerably. Both of the hazles in the succession are thin and their bases are unconformable. There is a large number of minor shales between the sandstones. The sequence is about 126 feet thick and can be given in the following generalised form, the section itself varying considerably along the length of the exposure:-

Three Yard Limestone

1' 6"-4' 0"	Gap- ?shale?	
3' 8"-5' 7"	Massive sandstone with micaceous and shaly base to 12' sandstone with grit base with clay galls and coal fragments.	High Brig Hazle.
	∿∿∿∿∿ ?Unconformity?	
c.6' 0"	Silty mica shale with carbonaceous fragments.	
4' 0"-6' 6"	Massive sandstone.	
c.12'0"	Gap with arenaceous shale fragments.	
8' 0"	Massive sandstone with pyritous base with coal fragments and clay galls.	
	∿∿∿∿∿ ?Unconformity?	
1' 2"	Nodular shale with fossil fragments.	
1' 2"	Calcareous shale.	
17' 0"	Fine dark limestone in several posts with shaly top.	Five Yard Limestone
2"-4"	Dark ferruginous shale.	
8"-10"	Crinoidal, calcareous sandstone.	

8"	Very fine mica shale.	
3' 0"	Rippled silty mica shale with ?algal? structures.	
1' 6"	Sandstone	Low Brig Hazle
1' 0"	Silty mica shale.	
	Small gap.	
1' 0"	Sandstone	
4' 0"	Siltstone	
3' 6"	Sandstone	
? 3' 0"?	Gap	
6' 0"	Massive pale brown sandstone	
3' 0" -6' 0"	Massive white sandstone	
	Unconformity	
2' 0" -4' 2"	Calcareous, fossiliferous nodular shale.	
27' 11"	Pale crinoidal limestone with shaly top.	- Scar Limestone

The strata discussed in the preceding pages serve to illustrate the variability of the beds studied, particularly the sandstones and shales. Some of the strata exhibit even greater variations but few sections of these are exposed. The minor limestones, in particular, are very variable and the Cockle Shell Limestone exhibits numerous variations.

f. VARIATIONS IN THE COCKLE SHELL LIMESTONE AND ASSOCIATED STRATA

The Cockle Shell Limestone in Bow Lee Beck, near its confluence with the Tees, consists of

10 feet of white to pale grey arenaceous and pyritous limestone. Less than a hundred yards away, at Scorberry Bridge, it is argillaceous but not arenaceous and contains the gigantoproductid fauna which is characteristic of this limestone. The succession at the bridge can be given as:-

1' 1"	Fine dark limestone
5' 0"	Fine dark limestone with gigantoproductids and crinoid stems.
2' 0"	Calcareous shale.
3' 6"	Fine dark limestone.

In the Lunedale Quarries, three and a half miles to the south-east, the limestone is represented by the following sequence:-

6"	Hard fossiliferous algal limestone with pyrite.
1' 10"-2' 8"	Fossiliferous calcareous shale.
3' 5"	Shaly limestone
10"-1' 3"	Highly fossiliferous calcareous shale to shaly limestone.
3' 9"	Fine dark limestone.

According to Jones (1956), the above succession is representative of the Single Post Limestone. However, this is believed to be the equivalent of the Cockle Shell Limestone for the following reasons. First, there is a transgression of the Whin Sill on Crossthwaite Common which cuts the Single Post Limestone so that in the south-east this limestone lies below the Whin Sill. There is no evidence of further change in horizon of the sill. Thus the Single Post Limestone must still lie below the Whin Sill in the Lunedale Quarries, where it lies below the limestone succession described above.

Second, in the Lunedale Quarries, there is a thin limestone between the beds described and the Scar Limestone. This is 2'8" in thickness and is fine, dark and shaly, containing gigantoproductids. This limestone also bears cauda galli markings (Wells,1955) on the upper surface and might be thought to be equivalent to the Cockle Shell Limestone. However, a similar limestone post occurs at the base of the Scar Limestone on Crossthwaite Common which becomes separated from the Scar Limestone by a series of thin shales and sandstones thickening to the south-east. The 2'8" limestone bed in the Lunedale Quarries is considered, therefore, to be equivalent to the bottom post of the Scar Limestone to the north-west.

The stratigraphical sections illustrated (Figures I. A, 1, 2, and 3.), and the sections discussed in the previous paragraphs show the extent of the local variability of the strata, particularly in the area studied. This variability serves to emphasise that there is no "typical" cyclothem although there are many variations of the hypothetical ideal cyclothem.

g. VARIATION IN RELATION TO THE THICKNESS OF STRATA

The stratigraphical sections show that the total thickness of sedimentary rock between one limestone and the next in order above it is variable from one section to another, even over the small area studied. There appears (see Figs. I. 1,2, and 3) to be no correlation between the shale or sandstone content and the thickness of strata between the limestones. The type of material deposited and the resultant different relative rates of compaction are, therefore, not wholly, if at all, responsible for changes in thickness of strata from one area to another. It appears that the changes are more probably due to varying rates of sedimentation at different points.

h. VARIATION IN RELATIVE PROPORTION TO ROCK TYPES

Not only are there differences in the thickness of sediment between the limestones but also differences in the proportions of the rock types. The following three sections taken from Teesdale illustrate the variation in proportions of the rock types:-

	Total Thickness	Limestone	Sandstone	Shale
Grasshill shaft, Scar to Three Yard Limestone.	151 ft.	49 ft.	46 ft.	56 ft.
		32.5%	31.5%	36%
Beagill-Hell Cleugh, Scar to Three Yard Limestone.	135 ft.	36 ft.	70 ft.	29 ft.
		29%	50%	21%
Skears Great Rise, Three Yard to Great Limestone.	160 ft.	20 ft.	70 ft.	70 ft.
		12%	44%	44%

Because of the great variability of the Teesdale strata and because much of the area is unexposed, an absolute average for the proportions of the different rock types over the whole area is not assessable. It appears however, from a consideration of many sections, that the abundance of rock types is in the following order: sandstone, shale and limestone.

CHAPTER II

THE ARGILLACEOUS ROCKS: PART I.

The Laboratory Techniques

a. INTRODUCTION

The clay minerals have been studied by means of X-ray powder photography, differential thermal analysis and infra-red absorption spectroscopy.

Petrographic studies, by means of thin sections, have aided other methods of study.

Whilst all the samples of argillaceous rocks have been subject to X-ray powder analysis in their natural state, many separated fine fractions have also been studied.

b. METHODS USED IN THE SEPARATION OF THE FINER CLAY FRACTIONS

Although the underclays examined were generally sufficiently friable to be powdered without crushing, all of the shales were indurated and required crushing to effect disaggregation.

The shales were first crushed in a percussion mortar to a minus 120 mesh size.

Grinding in an agate pestle and mortar was discontinued after the first few separations as it was found that it tended to cause dislocation of the clay mineral lattices, in turn resulting in the production of excessively diffuse X-ray powder photographs. The fine powder, after crushing, was separated by a modified form of the method described by Mackenzie (1956), as follows:

About 50 gms. of the powder obtained from crushing were placed in a quart bottle. The bottle was then half filled with distilled water. Three or four mls. of 0.880 ammonium hydroxide solution were added as a dispersing agent. The bottle was then corked and placed in an end over end shaker. Normally four samples were shaken at the same time for three to four hours. The bottles were then removed, almost filled by adding more distilled water and shaken by hand for one minute. They were then allowed to stand for eight hours. After this time, the top 10 cms of the suspensions were siphoned off into large beakers. Using Stoke's Law:-

$$V = \frac{g}{18} \times \frac{b - p}{n} d^2$$

(where g is the acceleration due to gravity, b is the specific gravity of the solid, p is the specific gravity of the liquid, n is the viscosity of the liquid, d is the diameter of the particles in millimetres and v is the settling velocity in centimetres per second).

it can be calculated that all particles greater than two microns equivalent spherical diameter, 2μ e. s. d. will have settled below the 10 cm level in eight hours.

A further two or three mls of 0.880 ammonium hydroxide solution were added, the bottles were refilled, shaken by hand, and allowed to stand a further eight hours. The procedure was repeated generally four to six times to ensure representative sampling and also to obtain sufficient quantities of the suspensions.

Occasionally longer periods of suspension were used, for example, sixteen hours to collect particle sizes of less than 1.4μ e.s.d.

During the first few separations, the suspensions were neutralised by the addition of 9N acetic acid and were then coagulated by adding 20 mls of 1N magnesium chloride solution. It was found as work proceeded that, in most cases, the addition of acetic acid was not necessary and usually 5 mls or less of magnesium chloride solution were sufficient to coagulate the clay suspensions.

When the supernatant liquid was clear it was decanted and the clay sludge was washed into 25 ml tubes and centrifuged. After pouring off the water, the clay was washed by shaking with distilled water and again centrifuged. This was repeated several times. The clay was then washed and centrifuged twice, using 96% ethyl alcohol, in order to dry it and also to remove any remaining salts. The washings were tested with silver nitrate solution during this last process to detect the presence of chloride ions and the sample was rewashed if the test was positive.

The clay was allowed to dry in air, was crushed gently and placed in sample tubes.

For differential thermal analysis and infra-red absorption analysis, the clays were, at first, placed in a dessicator for four days over a saturated solution of magnesium nitrate (56% relative humidity). In practice, this was found to be unnecessary for most of the routine work of identification. The same procedure as for X-ray samples was therefore used to obtain samples for differential thermal and infra-red absorption spectrographic analysis.

c. X-RAY POWDER ANALYSIS

i. The Cameras

The specimens of separated and natural clay samples were examined by means of X-ray powder photography. Both 9 cm and 11.4 cm powder cameras were used. In the later stages of the investigations an 11.4 cm Phillips powder camera, incorporating the slit collimating system, was obtained and this was used; for the earlier part of the investigation only a 9 cm powder camera was available.

The 9 cm camera was modified for use with clay mineral specimens in the manner described by Brown and Dibley (1956) for unevacuated cameras. With the acquisition of a new X-ray apparatus, the camera was subsequently remodified and the collimating system improved. A slit collimating system was designed to give a beam which was between 0.5 and 1 mm wide at the beam stop. The beam outlet tube was replaced by a narrow, hollow brass trap with a lead backing.

Measurements of spacings up to 27Å with copper radiation and 32Å with cobalt radiation could be made. These limits are more than adequate for the identification of the clay minerals dealt with in this work.

The camera constant of the 9 cm camera was calculated, employing accurate measurements for the camera diameter, width of the knife edges and distance of the knife edges from the camera wall (see Fig. 2.1). Using this calculated constant, the spacings of pure α-quartz (low temperature quartz) were determined from an X-ray powder photograph taken with the camera. The spacings were found to be in close agreement with those given by the National Bureau of Standards, circ. 539, vol. III, the maximum errors being in the order of 0.3%.

ii. Experimental Technique

Specimens of separated and natural clay samples were mounted on pyrex glass hairs, using collodion as the bonding agent. Comparison of photographs of specimens mounted in this manner, with photographs of the same samples mounted in Lindemann glass tubes showed no recognisable differences. The powders can thus be regarded as random aggregates.

Filtered copper (CuKα) and cobalt (CoKα) radiations were used in the investigations although the cobalt radiation was generally only used for the iron-rich samples.

All of the intensities of the lines measured on the films were estimated visually and the linear measurements were accurate to 0.005 cm.

d. CHARACTERISTICS USED IN THE X-RAY IDENTIFICATION OF THE CLAY MINERALS

i. Quartz and other non-clay minerals.

Most of the shales examined contain some quartz. This mineral was distinguished by the presence of strong lines on the films at 4.26, 1.816 and 1.540Å. The strongest line at 3.35Å is near to that of several clay minerals, for example, illite, 3.34Å and kaolinite, 3.36Å.

Calcite was identified primarily in the field (acid test), but its presence is indicated in untreated clay samples by strong lines at 3.03 and 1.917Å.

ii. Illites

Almost all the shales examined contain illite, generally as the major constituent. The term illite was first proposed by Grim, Bray and Bradley (1937) as a general name for the mica-like clay minerals with a 10Å c-spacing and with no expanding lattice characteristics. This definition includes both the dioctahedral muscovite-type lattice and the trioctahedral biotite-type lattice.

Mackenzie (1957 and 1959) has proposed that the dioctahedral and trioctahedral lattice types should be separated and has reserved the name illite only for the iron-poor dioctahedral mica clay minerals. The iron-rich dioctahedral types are the true glauconites, whilst the trioctahedral types he places in a separate group, the ledikites.

Here, the term illite will be used in the sense implied by Mackenzie's definition and the illites are identified primarily by the presence of the basal reflections at 10, 5 and 3.34Å.

Levinson (1955) has shown that the X-ray data can also be used to distinguish the different polymorphs of illite, of which four have been recognised. These can be listed in the following manner:

1. The three-layer trigonal polymorph, 3R (or 3T, see below).
2. The two-layer monoclinic polymorph, 2M.
3. The one-layer ordered monoclinic polymorph, 1M.
4. The one-layer disordered monoclinic polymorph, 1Md.

The use of letters requires careful consideration. Ramsdell (1947), studying the various types of silicon carbide, suggested that the number of layers in the unit cell could be expressed by a number, followed by a capital letter, for example, 'H' for hexagonal, to designate the symmetry of the unit cell. Ramsdell used only 'H' for hexagonal and 'R' for rhombohedral. He chose the letters 'P' and 'T' to express the basic parallelogram and trapezoidal arrangements of the atoms in the 1120 planes.

The present worker believes that the letter 'T' should be used in the sense that Ramsdell originally intended and thus the use of 'T' for trigonal unit cells (Levinson, 1955) or triclinic unit cells

(Mackenzie, 1957 and Frank-Kamenetsky, 1960) should be discouraged.

It is proposed to use the following notation for the various unit cells; previous workers' use of the letters in the same sense is indicated:-

' R' - rhombohedral (Ramsdell, 1947)
' H' - hexagonal (Ramsdell, 1947)
' M' - monoclinic (Levinson, 1955)
'pM' - pseudomonoclinic (Mackenzie, 1957)
'Tc' - triclinic (Mackenzie, 1957)

Levinson (1955) used 'd' after the main notation to indicate disorder in the illites, whilst Mackenzie (1959) used 'γ' before the main notation to indicate disorder in the kandites. Here, it is proposed to follow the earlier worker and use the symbol 'd'.

The spacings of the different polymorphs are illustrated in tables 2.1 (from Levinson, 1955) and 2.2 (from Mackenzie, Walker and Hart, 1949). Differentiation of the four polymorphs can be made using the spacings of the lines occurring between 4.4 and 2.6Å. Except for the strong basal reflection (00ℓ) at

3.34Å, this region contains only (02ℓ) and (11ℓ) reflections.

In the one-layer disordered monoclinic polymorph, 1Md, the lines are either very weak or absent in the region between 4.4 and 2.6Å. In particular, lines at 3.07 and 3.62Å are extremely weak or absent.

In the three-layer trigonal polymorph, 3R, the lines in this region, in particular, the line at about 3.85Å, are of medium intensity.

In the two-layer monoclinic polymorph, 2M, the lines in this region are of weak to medium strength. The line at 3.85Å is very weak as is the line outside this region at 1.50Å (060).

In the one-layer ordered monoclinic polymorph, 1M, the lines are very similar in character to those of photographs of the 2M polymorph. Great difficulty may be encountered when attempting to distinguish between these two forms. The presence of a line of weak to medium intensity at about 2.67Å is indicative of the 1M variety.

Walker (1950) and Grim, Bradley and Brown (1951), have stressed that the position of the (060) reflection and the intensity of the basal reflection at about 5Å can be used to distinguish between the dioctahedral and trioctahedral micas.

Nagelschmidt (1937) has shown that for the dioctahedral forms the (060) spacing is near 1.50Å and the basal spacing at 5Å is strong. For the trioctahedral forms the (060) spacing is near 1.53Å and the 5Å basal spacing is weak.

Brown (1951) has emphasised that care must be exercised when using the 5Å basal spacing. If the mica is dioctahedral and aluminous, the 5Å basal reflection is of the same order of intensity as the 10 and 3.3Å basal reflections. If, however, the mica is dioctahedral and is mainly ferriferous in the octahedral positions, the 5Å basal reflection is weak or absent.

For the known trioctahedral micas, the 3.3 and the 2.0Å basal spacings are compared. If these are of similar intensity, the mica is magnesium-rich but if the line at 3.3Å is appreciably stronger, the mica is iron-rich. In the latter comparison the clay must be quartz-free.

iii. Kandites, or Kaolin Minerals

Ross and Kerr (1931) defined the kaolin group of minerals as essentially hydrated aluminous silicates of approximate composition $2H_2O.Al_2O_3.2SiO_2$.. Brindley (1951) has shown that the kaolinite minerals are characterised by strong basal reflections at about 7.1 and 3.57Å. The term kandites was suggested by Brown (1955).

According to Murray (1954), the variations which occur in the X-ray powder photographs from highly crystalline to poorly crystalline kandites form a discontinuous series.

The kandites vary from the poorly crystalline dehydrated halloysite (halloysite which has been heated to 400°C), through the "fireclay" minerals to highly crystalline kaolinite. Murray defined the degree of crystallinity to include disorder within the crystallographic unit layer and also the stacking variations of the unit layers.

Internal disorder is due to the randomness of distribution of the aluminium atoms among the octahedral positions.

Brindley and Robinson (1945, 1946) showed that the kaolinite of a highly ordered nature was triclinic. The kaolinites of lower crystallinity can be indexed as monoclinic or pseudomonoclinic with a unit cell of the same dimensions as the triclinic kaolinite but with α equal to 90°. They have also shown that the structure is highly disordered along the b-axis with unit layers randomly displaced by multiples of b/3. The arrangement along the a-axis is like that of triclinic kaolinite.

Murray (1954) suggested that the X-ray reflections reveal both the internal structural disorder and the random stacking of the unit layers, whereas differential thermal analysis reveals only the orderliness of the internal structure.

In highly crystalline kaolinite, the reflections are sharp but with a decrease in crystallinity by random displacement along the b-axis and rotation of the layers through 120°, there is an increase in the diffuseness and a broadening of the reflections. With a decrease in crystallinity there is also a decrease in the number of reflections. Thus kaolinite has thirty one reflections from the basal 7Å to the (060) line whereas dehydrated halloysite has only seven. There is also an increase in the basal spacing from 7.18Å in the highly crystalline kaolinite to 7.40Å in the dehydrated halloysite. This is due to the presence of water molecules between occasional unit layers of the more poorly crystalline kandites and/or due to either disorder within the unit layers or stacking variations within the unit layers, both of which would be expected in the more poorly crystalline minerals. With decrease in crystallinity, resolution also decreases so that closely spaced lines are either poorly resolved or unresolved. In particular, the $(11\bar{1})$, 4.17Å and the $(1\bar{1}\bar{1})$, 4.12Å lines are unresolved.

Finally, a decrease in crystallinity is marked by the absence of certain reflections due to regular displacement along the b-axis.

Johns and Murray (1959) use an empirical crystallinity index determined from the ratio of the intensities of the lines corresponding to the (021) and the (060) spacings.

Mackenzie (1959) has suggested that the kandites can be classified on a structural basis in a manner similar to that used for the micas. Thus nacrite would become 6M-kaolinite; dickite, 2M-kaolinite; kaolinite, Tc-kaolinite if the mineral had an ordered triclinic structure and ψ M-kaolinite if it had a b/3 disordered monoclinic structure as in the "fireclay" minerals. As indicated above, the use of Md-kaolinite is preferable to ψ M-kaolinite.

The main difficulty encountered in the identification of the clay minerals dealt with in this work was the distinction between kandites and chlorites. Clay minerals, other than illites, only occur in small quantities in most of the shales examined. When a 7Å line of a weak nature was observed in the X-ray photograph, this may have been due to kandite or chlorite minerals of an iron-rich variety in which the 14 and 4.7Å first and third order basal reflections are weak or absent. The (060) line can be used in some cases to distinguish between the two minerals. Acid and heat treatment were generally used in this work where there was any doubt as to the identity of the clay mineral giving rise to a 7Å line. These two techniques are discussed in the next section.

Examples of typical kandite spacings are given in Table 2.3 (taken from Murray, 1954).

iv. Chlorites

Chlorites have been found in small quantities in many of the shales examined but were not identifiable specifically.

Grim (1953) has stressed that the clay size 14Å chlorites occurring in sediments are not well known and it is not definitely established whether they are similar to the well crystallised chlorites of metamorphic and other origins. An examination of the available data shows that they may differ by random stacking of the layers and perhaps occasionally in their state of hydration.

It has been shown by Brindley and Robinson (1951), Grim (1953) and Bradley (1954) that the small size and less regular crystallinity of the sedimentary chlorites causes diffuseness of all of the reflections with the elimination of some of the weaker ones.

Brindley and Robinson (1951) have shown that the chlorites can be identified from the 14 and 7Å lines, the one at 14Å remaining after heating to 550 to 600°C and often increasing in intensity, whereas the 7Å line becomes reduced in intensity. Kandites on heating to this temperature are destroyed due to breakdown of the lattice and thus the lines due to kaolinite completely disappear.

Chlorites, unlike the kandites, are destroyed after heating in dilute hydrochloric acid, the length of time varying with the acid strength and the type of chloritic material being investigated.

Chlorites rich in iron give weak first and third order basal reflections at 14 and 4.7Å, but give strong second and fourth order reflections at 7 and 3.5Å. As shown above, the first and third orders are increased in intensity whilst the second and fourth orders are decreased in intensity on heating.

Some typical examples of chlorite spacings are given in Table 2.4 which is taken from Brindley and Robinson (1951).

v. Mixed-Layer Clay Minerals

The characteristics of the mixed-layer clay minerals, though rather variable, enable them to be distinguished from the pure species. A generalised summary of these characteristics is given in the following paragraphs, based on data from Hendricks and Teller (1942), Brindley (1951), Brown and MacEwan (1951) and Bradley (1953).

Interstratification in mixed-layer clay minerals may be in the form of regular or random stacking of the component clay minerals.

In regular mixed-layer structures, the mineral is composed of different layers which are stacked at regular intervals along the c-axis. From the X-ray powder photographs, the spacings of the (00ℓ) basal reflections are regular and equal to the sum of the basal spacings of the two component layers. This is because, in regular mixed-layer structures, the separate layers contribute to make larger unit cells and, because of the regularity of the layering, all of the larger unit cells formed are identical. The structure of chlorite is an example of a regular mixed-layer structure, being composed of regularly stacked mica-like layers and brucite layers.

Brindley and Robinson (1951) have shown, for example, that a regular mixed-layer mineral composed of chlorite and 2M mica in the ratio 1:2 will give rise to the following series of basal reflections in which only the reflections which are near to those of the 2M mica (in greater quantity) would be expected to be strong:

	Chlorite	2M Mica	Mixed-Layer Mineral
(001)	14Å	20Å	34Å (weak)
(002)	7	10	17 (moderate)
(003)	4.66	6.66	11.3 (strong)
(004)	3.5	5	8.5 (weak)
(005)	2.8	4	6.8 (weak)

In random mixed-layer structures there is no regularity in the stacking of the component layers. A non-integral series of reflections will arise due to reflections from the basal planes. The layers can contribute to the scattering of the X-ray beam only as individuals. Scattering due to mutual orientation of the neighbouring layers as in the muscovites will be eliminated. Basal reflections will be enhanced where they correspond to the reflections due to both component layers but those not common to the two layers will amounts of interstratified material is indicated by a diffuseness and a shading of the 10Å line towards the higher spacings. If, on heating at 500°C for several hours, the 10Å line becomes sharp, the inter-layer material may be montmorillonite or may represent water molecules. If the line remains diffuse, the inter-layer material is probably chlorite.

In the Teesdale rocks, the only mixed-layer mineral found was a random interstratification of illite and a material which the X-ray evidence suggests is chlorite, the chlorite being present in small quantities.

No lines except those due to the clay minerals dealt with above and those due to quartz and calcite were observed. It is not, therefore, proposed to discuss the identification of the other clay minerals such as montmorillonite and sepiolite.

The clay minerals occurring in the Teesdale rocks and their distributions are dealt with in the ensuing chapter. Examples of some typical X-ray powder photographs are given in Fig. 2.2 and tables of spacings are given in Tables 3.10 to 3.15 (Chapter III).

e. DIFFERENTIAL THERMAL ANALYSIS

At the commencement of the investigations, it was hoped that differential thermal analysis might show some differences between the illites and the random mixed-layer minerals in the underclays and the illites in the marine shales. A small number of the Teesdale shales and underclays were, therefore, analysed by this technique. The analyses were kindly carried out in the department by Dr. G.A.L. Johnson.

As most of the samples contain a high percentage of illite, the effects of the other clay minerals tend to be masked and are, therefore, difficult to recognise. Mackenzie (1957) has shown that the temperatures of typical peaks can be given as follows:-

	Endothermic	Endothermic	Endothermic	Exothermic
Kandites	120-130°C	550-600°C	c. 950°C	950-1000°C
Illites	c. 125°C	c. 550°C	c. 900°C	900-1000°C
Chlorites	c. 125°C	c. 600°C	c. 800°C	c. 900°C

As X-ray analysis reveals, kandites are usually absent from the shales studied and the chlorites occurring in them generally have a weak series of basal reflections. Grim (1951) has shown that this type of chloritic material has little effect on the differential thermal curves of illites.

Mackenzie (1957) has shown that if the colour of the fired clay sample is red it may indicate the presence of a chlorite but the presence of carbon and iron-bearing materials may have the same effect.

Thus, although differential thermal analysis has been used for some samples, its usefulness as an aid to the identification of the minor quantities of clay minerals occurring in the shales was very limited.

Some representative curves are given in Fig.2.3. The first curve (T.24) is that of a rather pure illitic shale and shows a typical endothermic peak at about 120°C due to the loss of hygroscopic water. A second endothermic peak occurs at about 600°C due to loss of the lattice hydroxyl groups. The third endothermic peak at about 1000°C due to breakdown of the lattice does not occur.

The second curve (T.75) is of an illitic shale which contains, in addition, quartz and chlorite. Here, the endothermic peak at about 1000°C is present. The effect of chlorite is not seen in this curve, although the small endothermic peak at about 770°C may be due to chlorite.

The third curve (T.99) is of an illitic shale containing some chlorite and quartz. The exothermic peak 'P' at 700°C is due to small amounts of pyrite in the fine fraction.

The fourth curve (T.259) is of an illitic underclay which contains chlorite, quartz and mixed-layer mineral. The curve is rather poor for an illitic shale but otherwise the effects of the mixed-layer mineral and the chlorite are not seen.

Fig.2.4 shows two curves of an underclay (T.364) which contains mixed-layer mineral with minor amounts of chlorite and illite. The curve for the natural clay is typical of the illite curves except for a large exothermic peak at 400-500°C due to the burning of carbonaceous material. The fine fraction is similar but shows that the carbonaceous matter is different from that in the course fraction and its combustion gives rise to a double exothermic peak. The exothermic peak at 1000°C before the final endothermic peak cannot be explained and its origin is not known.

f. <u>Infra-red absorption analysis</u>

It was hoped that where differential thermal analysis had failed to yield more information on the Teesdale clay minerals, infra-red absorption analysis might be of more use. Analyses were made of a number of selected clay samples, using one of the latest models (G.S.2.A.) of the Grubb Parsons grating and prism Infra-red Absorption Spectrograph.

Before analyses of the clay samples from the Teesdale rocks were made, several samples of pure reference minerals (Fithian illite, Supreme Kaolin, chlorite, montmorillonite and quartz) and various weighed mixtures of these clays also were analysed. This was done in order to test the usefulness of the method and also to determine its sensitivity. It was also necessary to obtain a series of reference curves. Reference curves produced by Kerr and Pickett which were published in the A.P.I. Research Project 49 (1951) were not so detailed as those obtained in the present work and it was therefore thought necessary to prepare a new series of standards. More recently Lyon and Tuddenham (1959, 1960) have published curves for the chlorite family.

For the analysis, two milligrams of the clay sample were thoroughly mixed with 0.4 gram of water-free potassium bromide by grinding in an agate mortar. The material was then placed in a press and a pressed disc of 0.5 mm thickness and 15 mm diameter was formed. Absorption curves for the discs were then made.

Dry potassium bromide gives ~~the~~ few absorption peaks in the third and first orders of wavelength (2.5 to 3.7 microns and 5 to 25 microns respectively). The two small peaks at 3.43 and 3.50 microns are distinct and cannot be confused with peaks due to the clay minerals.

i. Absorption peaks of pure clay minerals

The number and positions of the absorption peaks differ from one clay mineral to another and, with any one clay mineral, the heights of the peaks vary with the quantity of the mineral present. Peaks of any diagnostic importance only occur in the third and first orders of wavelength and only these were used. The main differences between the various minerals are seen in the third order absorption curve (Fig.2.5).

Quartz has a strong peak at 2.89 microns and very weak peaks at 3.07 and 3.33 microns.

Fithian illite has a weak peak at 2.67 microns and a medium peak at 2.95 microns, whereas Supreme Kaolin has a strong peak at 2.70 microns with a shoulder at 2.73 and a strong peak at 2.76 microns.

Calcium montmorillinoid, Fuller's Earth, has a medium peak at 2.77 microns and a medium, broad peak at 2.90 microns whilst Penninite has a strong peak at 2.75 microns with a broad shoulder at 2.89 microns.

The differences in the positions of the peaks has been attributed to differences in the bonding of the (OH) groups by Keller and Pickett (1950). Monomeric (OH) gives an absorption peak at about 2.75 microns, dimeric (OH) at about 2.85 microns and polymeric (OH) at about 2.95 microns. Adsorbed water gives absorption peaks at 2.7 to 3.2 microns in the third order and at about 6 microns in the first order.

In the first order, absorption peaks characteristic of the different clay minerals occur, but these have not been related to any particular bondings. However, absorption peaks at 9 microns are believed to be due to the Si-O linkages and peaks at about 10 microns are believed to be due to the octahedral alumina sheets. It has been shown by **Launer** (1952) that the silicon-oxygen groups have different absorption bands in different co-ordinations, viz.

Isolate SiO_4	9.7 to 12.2 microns
Single chain	9.2 to 11.7 microns
Double chain	9.6 to 10.6 microns
Layer	9.3 to 10.4 microns
Framework	9.6 to 10.1 microns
SiO_2	c. 9.2 microns

Thus quartz has a strong absorption peak at 9.27 microns and the clays with layered or sheet structures have strong absorption peaks at 9.85 microns for Fithian illite, 9.7 and 9.9 microns for Supreme Kaolin, 9.62 microns for Fuller's Earth and 9.23, 9.53 and 10.0 microns for penninite.

ii. Absorption peaks due to mixtures of clay minerals

For pure clays, of known characters, the infra-red absorption spectrograph was found to be useful in the determination of relative amounts in mixtures. A series of curves for various mixtures of pure Fithian illite and Supreme kaolin in various proportions by weight were produced. A graph was then constructed, plotting the ratio of the two minerals in the mixtures against the difference in heights (measured in inches for convenience) of a series of absorption peaks due to illite and kaolinite. A series of curves was constructed from which, using the height difference between any two given peaks on the curve, the ratio of illite to kaolinite by weight, could be determined. This graph and some of the characteristic curves are given in Figs. 2.6 and 2.7.

It is realised that for the purer illites and kaolinites the graph is of value, but when using material which is often impure and poorly crystalline, the curve can only be used as an indication of the quantities present. The examination of large numbers of mixtures of kandites and illites of varying crystallinity might possibly show that the method is very useful for quantitative work.

The different absorption curves, although not fully understood, are a useful aid in the identification of the clay minerals as separate species but are of little use in distinguishing the illites of slightly different composition such as occur in the Teesdale rocks. Thus in Fig.2.7, curve T.364 is of an underclay containing mixed-layer mineral, illite, chlorite and quartz, curve T.75 is of a shale containing illite, chlorite and quartz whilst curve T.22 is of a pure illitic shale. Insofar as the clay minerals are concerned, no recognisable difference between these curves occurs. The curves for T.364 and T.75 show small peaks between 8 and 9 microns in the first order wavelength, due to the presence of quartz.

g. FLUORINE ANALYSES OF SOME OF THE ARGILLACEOUS ROCK TYPES

It has been suggested by Keith and Degens (1959) that clay minerals take up fluorine when exposed to sea water for any length of time. Shales which have been deposited in different environments from marine to nonmarine would be expected, therefore, to have different fluorine contents. Fluorine analyses were accordingly made of some representative samples of Teesdale shales. The analyses were kindly carried out in the department by Dr. R.P. Hollingworth.

The method employed was essentially that given by Hollingworth (1957) in Analytical Chemistry, 29, 1130, but with two important modifications -

1. The precipitates of Si, Al, Fe and other elements are, after digestion, by immersion of the beaker and its contents in boiling water, filtered by suction on a fine porosity sintered glass funnel. Separations of the precipitate from the filtrate are made very sharp by fitting the sintered glass plate with a circle of Whatman 42 filter paper on the top of which is layered some filter paper pulp.
2. The titration of the fluoride, as fluosilicic acid, with thorium nitrate in the presence of the indicator SPADNS, is done in a buffered solution, the buffer being prepared by half-acidifying sodium sulpanilate with perchloric acid.

This modified method has given directly comparable results on metamorphic rocks analysed by the original method.

Since the SPADNS method is capable of determining fluorine in 0.25 gram samples of rocks to plus or minus one milligram of fluorine and since 10 ml aliquots out of a 100 ml distillate were taken, the results given in the succeeding chapter should be within plus or minus 0.004 gm of fluorine per 1 gm of sample or 0.4% of fluorine of the most probable results obtained by this method. However, in view of some doubt on the blanks run during the analyses, the poor agreement among the various workers using different methods for fluorine analyses of rocks, and the fact that the thorium nitrate was standardised on 25 ml aliquots, the results for the sake of comparison among themselves and others should be regarded as no better than 1 part in 10. In other words, the fluorine values given may vary within a range of plus or minus 10% of the actual value given.

The results of the fluorine analyses are given in the following chapter where the results of the clay mineral analyses are dealt with at greater length.

h. CONCLUSION

From the various techniques applied to the shales examined, X-ray powder analysis was chosen as the most effective method for clay mineral identification.

Additional aids, such as heat and acid treatment, were also found to be indispensable in some cases.

CHAPTER III

THE ARGILLACEOUS ROCKS: PART II

Results of Field and Laboratory Investigations

a. INTRODUCTION

A total of 140 shale and underclay samples, representative of the main types, were collected from the various stream sections in the field. The division of rocks into types was based on characteristics which are easily recognisable in the field such as mica content, calcite content, (tested with dilute hydrochloric acid), presence or absence of quartz (based on roughness of hand specimen, gauged by feel), nodule content, fossil content and bedding.

In the laboratory, the shale samples were examined for differences in mineralogy and petrography and an attempt has been made to relate the mineralogical differences to variations in field characters and stratigraphical occurrence.

b. FIELD CLASSIFICATION OF THE ARGILLACEOUS ROCKS

The argillaceous rocks were classified into five major and two minor types in the field by means of a few simple tests and observations of easily identifiable characters. These were as follows:

1. Bedding characteristics: the rocks were classed as bedded or unbedded. Bedded rocks were further classified according to their degree of fissility (good or poor) or friability.
2. Calcite content: the rocks were tested with dilute hydrochloric acid and classed as calcareous or noncalcareous according to whether the test was positive or negative.
3. Fossil content: the rocks were classed as highly or sparsely fossiliferous and unfossiliferous according to their fossil content.
4. Mica content: the rocks were classed as nonmicaceous if mica grains were not visible to the unaided eye and fine to coarsely micaceous according to the size of visible micas (less than 0.25 mm, fine; 0.25 to 1 mm, medium; over 1 mm coarse)
5. Presence of quartz: a good guide to the presence or absence of quartz in the field was the feel of the bedding plane in hand specimen. This was

rough if quartz was present in substantial quantities.

6. Pyrite nodules: the rocks were classified into those without nodules or bands, those containing pyrite nodules or bands and those containing ironstone nodules or bands. The latter group was rare.

The relative colour of the beds (light, medium or dark grey) was considered as a useful distinguishing character but apart from the type 6a shales (see below) this depended to a large extent on the moisture content of the argillaceous rocks. As most of the rocks were permanently wet in the field this was not a practical method of distinction.

The types of argillaceous rock and their general characteristics are summarised thus:-

Type 1 Underclays beneath present thin coal seams or in the suspected former positions of coals. Generally non-calcareous and containing a variable amount of mica. Fossils, apart from plant fragments, are absent. Bedding and fissility are absent. No nodules occur. The amount of quartz is variable. These beds may occur below the limestone members of the sequence.

Type 2. Calcareous, non-micaceous and, generally, very fossiliferous shales, having well-developed bedding and poor to good fissility. Quartz, if present, is not detectable and no nodules occur. These shales occur in and immediately above limestones and very occasionally below the limestones.

Type 3. Calcareous to non-calcareous, finely micaceous shales which contain a restricted fauna, mostly brachiopods. Bedding is well developed and fissility is good. Pyrite nodules commonly occur and a small amount of quartz is usually detectable by feel. These shales always occur above the Type 2 argillaceous rocks which overlie the limestones.

Type 4. Non-calcareous, fine to coarsely micaceous shales with no fossils but with some evidence of tracks and trails and pit and mound structures in the more sandy partings. Pyrite nodules often occur at the base and very occasionally ironstone nodules or bands may occur near the top. The bedding is well developed but the fissility is not very marked.

Quartz is very apparent by feel. These shales always occur above the Type 3 argillaceous rocks but below the sandstones.

Type 5. Non-calcareous shales which often contain an abundance of coarse mica. Fossils, part from plant fragments, are absent. Pyrite and ironstone nodules are usually, but not always, absent and the bedding is generally poorly developed. Quartz is always present but is very variable in quantity. The characteristics of these shales are rather more variable than those of the other shales and many of the Type 5 argillaceous rocks show transitions towards other types. These shales generally occur above the main sandstone members of the cyclothems.

Two additional argillaceous rock types were recognised, which were of much rarer occurrence. These are both found within the sandstone members of the cyclothems and are designated Types 6a and 6b.

Type 6a. Thin shale bands occurring in the more massive and well sorted, clean sandstones. They are pale in colour and contain a variable amount of mica which is generally coarse. Bedding is good and the rocks are friable. Fossils and nodules are absent. Quartz is generally absent.

Type 6b. Clay galls occurring in some of the sandstones.

Among the argillaceous rock types, classified on field characters, were included some argillaceous sandstones. Examination of thin sections enabled estimates of the quartz contents of these rocks to be made. For the present purpose, sandstones are regarded as having over 50% quartz by volume and rocks containing less than 50% quartz are regarded as shales. With few exceptions, the field characters were found to be a reliable basis on which to classify the various sandy argillaceous rocks as either arenaceous shales or shaly sandstones.

Whilst much work has been done on the classification of sandstones and limestones, comparatively little has been done on the classification of the argillaceous rocks. No adequate comparisons with similar rocks studied by other workers can, therefore, be made. The classification of the argillaceous rocks studied into Types 1 to 6 developed for the present work is, therefore, peculiar to this study.

The use of definite names such as marine for the Type 2 and 3 argillaceous rocks was considered. However, the Types 4, 5 and 6a cannot be named in a like manner with any certainty and, therefore, the only names retained are those of seat earth or underclay for the Type 1 rocks as these have long been used in geological and mining literature and, in addition, are well known and well defined rock types.

A table of the different argillaceous rock types summarising their field characters is given in Table 3.1.

c. LABORATORY INVESTIGATION OF THE ARGILLACEOUS ROCK TYPES

In the investigation of rocks in the laboratory, the first method which is usually applied is the examination of thin sections. One of the reasons for the comparative rarity of descriptions of argillaceous rocks in geologic literature is that thin section examination presents a number of difficulties. Thin sections of friable shales are difficult to prepare and many friable shales are sufficiently dense to prevent the use of impregnation methods. Even when thin sections of shales can be prepared they are often almost opaque due to fine-grained carbonaceous matter or the clay mineral components are too fine grained to identify them with any degree of certainty.

Similar difficulties were met in the examination of the Teesdale argillaceous rocks. The metamorphosed shales, however, have been partially recrystallised and have lost most of their opacity, probably as a result of the expulsion of carbonaceous or bituminous matter.

Thin sections have been mainly used, therefore, to determine the quartz contents of the rocks although some general features have also been observed. Thin sections have shown that the underclays have no defined orientation of the clay minerals, whereas the shales have a subparallel orientation of the clay minerals. The subparallel texture is best seen under crossed nicols, where it is revealed by partial mass extinction in the parallel position.

The shales studied vary in their degree of orientation of the clay mineral particles and the better bedded shales generally have a more marked subparallel texture. It has not been possible to distinguish any secondary growth of the clay minerals in the shales and, although this may have occurred, it seems likely that the texture is mainly due to the original deposition of the constituent flakes and possibly also some reorientation under pressure. The coarser, detrital micas, in particular, show a marked orientation which is probably original.

The underclays show no orientation of the clay minerals and some of the more arenaceous seat earths, more properly classed as sandstones, show evidence of either recrystallisation or deposition of secondary chlorite. This is shown in Fig. 4.7.

Except for the underclays and the metamorphosed, often markedly recrystallised shales, the identification of the clay minerals by microscopic methods was very difficult and it was therefore necessary to resort to other methods of attack. As already noted the several techniques which were tried included X-ray powder analysis, differential thermal analysis and infra-red absorption spectrographic analysis. As stressed in the previous chapter, X-ray analysis proved to be the most satisfactory and the results of the clay mineral analyses of the argillaceous rocks are based mainly on this technique of investigation. Tables 3.2 to 3.7 summarise the field and laboratory data for the argillaceous rocks.

d. RESULTS OF THE CLAY MINERAL ANALYSES OF THE ARGILLACEOUS ROCKS

Apart from the variations in the quartz, mica and calcite content of the argillaceous rocks recognised in the field and the laboratory, X-ray analyses have shown that there is some diversity in the clay mineral content. Several different types of clay mineral occur in the shales and underclays. Excluding kandite, the clay minerals found in the Teesdale rocks were illite, chlorite and random mixed-layer minerals. Kandites were found in some coarse arenaceous bands in shales above the Three Yard Limestone. Examination of thin sections showed that these bands contained kaolinite in the form of "books" between the quartz grains. In view of the delicate nature of these aggregates which would be broken in transport, it is inferred that the kaolinite here is of secondary origin, probably derived from the weathering of feldspars.

The detrital clay minerals occur in the rocks in the following combinations:-

1. Illite
2. Chlorite
3. Illite plus chlorite
4. Mixed-layer mineral
5. Illite plus mixed-layer mineral
6. Illite plus chlorite plus mixed-layer mineral.

No mixtures of chlorite plus mixed-layer mineral without illite were observed.

The relationships between the clay mineralogy and the different rock types are best expressed in tabular form (Tables 3.8 and 3.9). The results are also expressed graphically in Fig.3.1.

The illites, as shown by X-ray analyses, are of the one layer disordered monoclinic variety (1Md) with possibly some admixed two layer monoclinic (2M) variety. The chlorites are also poorly crystalline and, in general, have weak to very weak first and third order basal reflections and are, therefore, mostly iron-rich.

In the presence of illite, the mixed-layer minerals are rather difficult to identify specifically

The variations between the different samples of the same rock type is large and the variations in the fluorine content which have been observed are of doubtful significance.

f. CONCLUSION

Several different types of argillaceous rock can be recognised both in the field and the laboratory. Differences in clay mineralogy occur which appear to be related to the differences in rock type.

Fluorine analyses have not established a definite relationship between the shale type and the fluorine content.

CHAPTER IV

THE SANDSTONES: PART I

Field and Laboratory Work

a. INTRODUCTION

Several types of sandstone have been recognised in the field and thin sections of these were examined in the laboratory.

Some of the more argillaceous and calcareous rocks were identified in the field as arenaceous shales and arenaceous limestones. These were subsequently placed in their correct classes after laboratory investigation. However, most of the sandstones were recognised as such in the field and could be classified into types. Laboratory investigation has shown that these types have significant mineralogical characteristics.

The pyrite-bearing sandstones were examined by reflected light. Heavy mineral analyses were made of some of the sandstones containing the higher percentages of heavy minerals.

b. FIELD CLASSIFICATION OF THE SANDSTONES

The characteristics used in the field classification of the sandstones were all of a simple nature. The characteristics chosen as being most useful were both structural and mineralogical. Structural characters included bedding, ripple marks and minor internal features such as pit and mound structures. Mineralogical characters used included the amount and size of mica flakes, the amount of clay material, size of the quartz grains and the presence or absence of calcium carbonate (dilute hydrochloric acid test).

In the field, the following types were recognised:-

Type 1. Clean, massive sandstones which show false bedding and occasional ripple marks. These contain very low percentages of mica and clay minerals and are usually calcite-free. They generally form the main sandstones in the succession where they occur but, in some of the cyclothems, are not present in every section. They are seen in some sections to be interleaved lensoid bodies and not continuous sheets.

Type 2. Flaggy sandstones which may contain shale bands at the base and grade up from shales or "grey beds" to the overlying Type I massive sandstones, where these occur. They are often rippled or laminated and contain a high percentage of mica and clay minerals. They are usually calcite-free. Those which occur at the base of the more massive sandstones are usually pyritous. Some of these rocks may contain pit and mound structures. They may also occur at the top of the succession in a cyclothem above the massive sandstones.

Type 3. Argillaceous and non-flaggy sandstones which contain a varying amount of mica. They occur at or near the top of the cyclothems and are generally rotten or poorly consolidated and may contain stigmarian rootlets. They show transition to the underclays and seat earths.

Type 4. Calcareous sandstones which have varying amounts of clay minerals and mica. These often grade laterally or upwards into limestones.

Type 5. Coarse, massive sandstones which are similar in mode of occurrence to the Type I sandstones, but with a noticeably greater argillaceous content.

Unlike the other argillaceous sandstones, the clay material occurs in distinct pockets and is generally pale cream in colour and not grey.

Type 6. Fine to coarse, generally dark grey-green sandstones which often contain clay galls, ferruginous fragments (now limonite, but probably originally clay ironstone) and pyrite. These sandstones have unconformable or abrupt bases. Only the High Brig Hazle, in Newbiggin Beck contains this sandstone; here it is found at the base of each of the two sandstones in the sandstone-shale sequence which makes up the Hazle.

The characteristics of the different sandstone types are summarised in Table 4.1.

This classification is not a rigid one as gradations occur between one type and another. In particular, sandstones occur above the Single Post Limestone which are similar in mode of occurrence to the Type 6 sandstones but mineralogically are similar to the Type 3 and Type 1 sandstones. However, if all the criteria are used, the gradational types can usually be satisfactorily placed.

Characters of each individual sample collected are given in the section on laboratory work, in Tables 4.2 to 4.7.

c. LABORATORY INVESTIGATION OF THE SANDSTONES

As has been stated, the laboratory investigation took the form of the examination of thin sections and polished sections. Some heavy mineral analyses were carried out and X-ray studies were made of certain of the clay minerals and micas.

Two important factors in the variations of the sandstones are the degree of sorting and the maximum and minimum size grades which occur. However, the rocks examined were generally too indurated to effect disaggregation of the constituent grains without fracturing or crushing them. Examination of the rocks in order to determine the grain sizes and the degree of sorting was, therefore, limited to thin section analysis.

Krumbein (1935) has shown that the thin section analysis of sandstones produces fairly reliable data as to the limits of the grain size and the average grain size. After performing a series of experiments on artificial material and unconsolidated sediments, he came to the following conclusions;

The average diameter, as measured on a random section, of an artificially prepared equigranular "rock" was about 0.75 of the true diameter and the grain size varied from 0 to the true diameter. Using natural material with grains which were not all the same size and which were not spherical, he found that the sphericity and degree of sorting had little effect on the estimations. He did, however, use a large number of grains in his determinations.

The estimates used in this thesis were made with a micrometer eyepiece in the microscope and where graded bedding occurred, each layer was estimated individually.

In order to determine the effect on the estimation of grain size of thin section examination, as opposed to examination of separated grains, an arenaceous limestone, T.6a, was placed in dilute acetic acid and the matrix dissolved. The separated grains and a thin section of the rock were then examined. The results of the two examinations can be compared in the following manner:-

	Thin Section	Separated Grains
Range of grain size.	0.025 to 0.45 mm	0.05 to 0.45 mm
General grain size.	0.1 to 0.175 mm	0.125 to 0.25 mm
Description.	Grains subangular to subrounded, some angular to very angular.	Grains subangular to subrounded, few rounded.

N.B. The thin section shows that many of the grains are sutured and are replaced by calcite and that some of the grains show secondary overgrowths of quartz.

It is here suggested that the results of thin section estimations are suitable for comparative studies for the present purposes and can, with caution, be used when comparing the grain size distribution of ancient sediments with modern unconsolidated sediments. It must be borne in mind, however, that the thin sections tend to give an impression of a larger number of smaller grains and a smaller number of larger grains. Thin sections do not affect the estimation of the maximum grain size in the sandstones investigated as even the coarser ones contain sufficient grains to enable some of the grains in the thin section to approximate to the maximum grain size. They do, however, give an impression of a lower minimum grain size because random sections through the grains include those sections which cut the peripheries. The lowering of the minimum grain size can be lessened by not considering the very small grains as seen in thin section which are due to the cutting of the grain peripheries. This would preclude the very poorly sorted sandstones. However, the number of very small grains occurring in most of the thin sections examined was very low and thus most of the rocks are believed to be fairly well sorted.

Thin sections tend, as a consequence of the aforementioned factors, to give an impression of a lower average grain size, although this effect is not large.

Finally, although thin section examination of sandstones does affect the estimation of the degree of sorting, the effect is negligible. This is due to the fact that lowering of the value of the average grain size is small and the fact that the Wentworth grade scale size limits are such that a considerable degree of tolerance occurs and the grain size must vary considerably before the class of the degree of sorting is altered.

d. THE PETROGRAPHY AND MINERALOGY OF THE DIFFERENT SANDSTONE TYPES

As has been stated in the section on the field classification of the sandstones, the different types recognised in the field also show significant variations in texture and mineralogy. These variations are described below.

Owing to the difficulty often incurred in making positive optical identifications of such fine-grained material, the clay minerals can only be tentatively identified on the basis of colour, form, birefringence and other general characters unless they occur in sufficient quantities to permit separation for further optical or X-ray work. For these reasons many specific names are placed in inverted commas.

Type 1. Sandstones

In thin section (Fig.4.1), these sandstones are characteristically well sorted (their constituent grains all lie within small limits in one or two of the size classes on the Wentworth grade scale) although a few have only a fair degree of sorting (their consituents still lie within two classes but the limits may approach a third class). Most of the grains lie between 0.075 and 0.375 mm. Some of the rocks may contain grains as small as 0.05 mm, or as large as 0.75 mm but these are rare. On the Wentworth grade scale, these correspond to very fine to medium grained sandstones.

Most of the rocks contain subangular to subrounded grains although some contain angular to subangular or some subrounded to rounded grains.

Quartz forms 95 to almost 100% of the rocks and the grains are of three different varieties:-

1. Strained grains showing undulose extinction. These form the bulk of the material.
2. Unstrained grains which show complete total extinction. These occur in very minor amounts.
3. Composite grains made up of intergrown quartz in different orientations. These grains are the least abundant of the three types.

Some rocks may contain a few chert grains but these are very rare and always form much less than 1%.

The sandstones are cemented by secondary overgrowths of quartz which are in direct optical continuation with the detrital quartz grains (Fig. 4.2). Occasionally a little calcite cement occurs and may form up to 3% of the rock.

Mica occurs in most of the rocks, the quantity varying from 0 to 2%. The micas are generally well scattered throughout the rocks and are usually coarse, being between 0.25 and 2.5 mm and mostly over 1 mm in length.

Most of the mica is muscovite but rarely leached brown mica or "hydrobiotite" may occur. In a very few of the samples examined biotite occurs but this is generally in the form of altered flakes or shredded aggregates. Most of the biotite is leached to "hydrobiotite" or occasionally has been altered to chlorite.

A typical example of this sort of alteration is seen in a sample of Low Brig Hazle, T.53. Here, the mica is mostly altered biotite. Flakes with residual areas of unaltered biotite show that the original biotite is pale to dark brown or green-brown in colour and shows second order interference colours. The alteration products are of two sorts. Many of the flakes have disintegrated and altered to fine-grained aggregates of a pleochroic, yellow-brown mineral which appears to show first order white birefringence masked by the colour of the material. This is probably fine-grained "hydrobiotite" or "illitic" material. Other flakes have expanded and become paler in colour. The interference colours of this material are reduced to the first order. This material though not identifiable specifically can be grouped under the term "Hydrobiotite". Some of the "hydrobiotite" is partially altered to a pale green, almost colourless, chlorite. Chlorite, which is colourless, often occurs as a secondary interstitial mineral.

Interstitial clay minerals usually form less than 1% of the rocks but may form up to 5%. Optical examination reveals that the clay minerals are mostly fine-grained "sericite" or "illite" but some of the rocks contain small amounts of "chlorite" and, very occasionally, kaolinite. Most of the interstitial clay mineral is secondary and occurs as alteration products of other minerals such as micas and quartz.

The "sericite" occurs in aggregates of colourless to very pale yellow (due principally to "iron" staining) flaky material which invariably shows second order interference colours under crossed nicols.

The "illite" is generally a pale yellow slightly pleochroic colour. It is very fine with poorly developed mutual orientation of the flakes. It generally shows first order interference colours under crossed nicols, probably due to its fine-grained nature.

The "chlorite" is colourless to very pale green-yellow, slightly pleochroic, and is very fine-grained. It shows low first order greys under crossed nicols.

The kaolinite always occurs in typical "book" structures which are too delicate to have been transported. They are thought to have been formed by the alteration of feldspars occurring in the rocks. Kaolinite is very rare in these sandstones.

Amongst the accessory minerals, feldspar is very rare and only one specimen, from the Nattrass Gill Hazle, was found to contain this mineral. The feldspar formed much less than 1% of the rock and was composed of fairly fresh plagioclase.

The detrital heavy minerals usually form less than 1% and never more than 2% of the original minerals in the sandstones. They can be listed in the following order of abundance; zircon, tourmaline, rutile, leucoxene and magnetite.

Zircon occurs in subangular to well rounded grains and is more often well rounded. Two varieties occur, a colourless variety and a less abundant very pale brown variety.

Tourmaline usually occurs as very well rounded grains although a few subrounded prismatic grains occur. Most of the tourmaline is a pale green to olive green or brown pleochroic variety but emerald green and very pale brown varieties also occur.

Rutile occurs only rarely and is a dark brown variety in the form of small acicular crystals. Some of the quartz grains in the sandstones also contain rutile and, rarely, apatite or tourmaline in the form of minute crystals.

Other grains include an opaque mineral with white surface reflection identified as "leucoxene", and also a little magnetite.

A little pyrite occurs which appears to be original but most of this constituent is probably of secondary origin. Small grains and euhedra of pyrite occur in many of the rocks but they are mostly of secondary origin since they replace both the quartz grains and the cement. Some of the sandstones are seen to be spotted in the field, the spots consisting of a mixture of clay mineral and limonite. Polished sections show that some of the spots contain small central grains of pyrite. The spotted rocks are, therefore, considered to have been formed by the weathering of pyrite-bearing sandstones. The origin of the pyrite is obscure but may have been deposited around minute nuclei from connate and groundwater solutions.

Type 2. Sandstones

In thin section (Figs. 4.3 to 4.6), these rocks are characteristically impure and fine-grained, most of the grains lying between 0.025 and 0.25 mm diameter, although the size may vary between diameters less than 0.001 up to 0.375 mm. The degree of sorting is variable but is generally such that most of the particles lie within two classes on the Wentworth grade scale. These are fine to very fine sandstones on the grade scale.

Most of the rocks contain angular to subangular quartz grains although a few may be subrounded.

Quartz grains form 50 to about 95% but generally 75 to 90% of the rocks.

The mineralogy of these rocks is similar to that of the Type 1 sandstones, except for the clay and mica content. The cement of the rock varies considerably and may be quartz, clay or carbonate or a combination of any of these constituents. Secondary silica occurs interstitially and is generally chalcedonic in form.

The micas are more abundant than in the Type 1 sandstones, varying between 1 and 30%, and are concentrated along the bedding planes. The micas vary considerably in size from less than 0.2 to about 2.0 mm, the rocks with the finer quartz grains usually containing the finer micas.

Most of the mica is biotite and "hydrobiotite" (degraded biotite) in the greater proportion of samples. Chlorite is much more common than in the Type 1 sandstones. In these rocks, as in the Type 1, the chlorite is flaky, colourless to very pale green, and shows birefringence in first order greys.

Several interesting relationships are found among the micas which were observed under the microscope. Biotite was seen to be altered to "illite", chlorite, "hydrobriotite", and other less well defined products.

Both colourless and green chlorites occur in the rocks and a third occurrence was noted in some of the rocks, in the form of spheroidal aggregates of colourless to brown chloritic material with a very low birefringence. Most of the spheroidal aggregates are enclosed in thin shells of "illitic" material. This third chlorite was identified on the basis of X-ray analysis (Table 4.8) as an iron-rich chlorite. Other more diffuse patches in the rocks which have a similar appearance are also probably chlorite.

The clay constituents may form as low as 2% but generally form between 5 and 25% of these rocks. The clays are mostly very fine-grained detrital micaceous minerals of the "illite" type. No kandite was observed in any of the sections.

Plagioclase feldspar occurs in some of the rocks and is generally fairly fresh. It always forms much less than 1% of the rocks.

The detrital heavy minerals are the same as occur in the Type 1 sandstones.

One further accessory mineral occurs, usually in the more micaceous rocks, in the form of small rounded grains of a similar size to the quartz grains. It has a low reflectivity and appears black in reflected light with a little white reflection from small isolated points. This material is probably carbonaceous but there was insufficient quantity to establish its specific identity.

Type 3. Sandstones

In thin section (Fig.4.7) these rocks are found to be very fine to fine-grained (Wentworth grade scale) argillaceous sandstones which may contain up to 40% of clay minerals. Argillaceous sandstones are regarded for the purpose of the present work as sandstones containing between 5 and 50% of clay material. The quartz grains vary between 0.01 and 0.25 mm in size and are angular to subangular. A little fine scattered mica may occur but forms much less than 1% of the rocks and the particles are less than 0.5 mm in size. Apart from the clay content, these rocks are similar in mineralogy to the Type 1 sandstones.

Most of the clay is detrital "illitic" material, although spheroidal aggregates of colourless chlorite also occur. These aggregates, as in the Type 2 rocks, are always spheroidal and appear to have formed in situ as the material is too delicate to have been transported along with the rest of the sediment. The origin of this secondary chlorite is obscure.

Type 4. Sandstones

In thin section (Figs. 4.8 and 4.9), these rocks are fine-grained to very fine-grained and are well sorted. The quartz grains are generally between 0.05 and 0.25 mm in diameter and vary from angular to subrounded, the finer rocks containing the more angular grains.

The clay minerals and micas, together, rarely exceed 5% and usually vary from 0 to 5%. The micas are usually less than 0.75 mm in size.

The heavy minerals are the same as occur in the other sandstones, and, as in the other rocks, feldspar may occur in minor amounts. Secondary pyrite may occur and appears to be replacing the calcite.

Apart from the carbonate, which is always calcite (dissolves completely in cold dilute hydrochloric acid) and which varies between 1 and 20%. the rocks are similar to the other sandstone types. Where the amount of calcite is small (it usually forms 3 to 5%), the cement may be quartz or the rock may be bonded by clay mineral.

Two subtypes of this sandstone can be recognised. One represents the more arenaceous end member of the Type 2 limestone, the arenaceous limestone, and is fairly clean, whilst the other more argillaceous subtype represents a more calcareous facies of the Type 2 sandstone.

Type 5 Sandstone

Some of the less argillaceous of these rocks, when examined in the laboratory, were found to be the coarser more argillaceous representatives of the Type 1 sandstones and were subsequently placed in this group.

All of the Type 5 sandstones, recognised as being distinct, both in the field and the laboratory, occur above the Four Fathom Limestone and belong to the Quarry Hazle or Tuft.

In thin section (Fig.4.10) these sandstones are characterised by their poor degree of sorting, the grain size varying from about 0.125 or less, to over 1.0 mm. These correspond to fine to coarse sandstones on the Wentworth grade scale.

The mica is generally small in amount and is generally up to 2 mm in diameter.

"Illite" is found in minor quantities in these rocks but they are unusual in that they are the only rocks examined which were found to contain any substantial amounts of kaolinite. The kaolinite forms pockets up to 1 mm, or occasionally more, in diameter and forms up to 10% of some of the rocks. The kaolinite is in the form of well crystallised "books" and appears to be secondary in origin, possibly derived from the alteration of feldspar.

Type 6. Sandstones

In thin section (Fig. 4.11), these rocks are seen to be very poorly sorted with angular to subrounded quartz grains which are cemented by secondary quartz. The grain size varies between 0.05 and 1.75 mm (from very fine to very coarse on the Wentworth scale). The matrix is argillaceous and forms 10 to 25% of the rocks. The clay minerals forming the matrix are mostly colourless to yellow (stained) or pale green chlorite, but minor amounts of "illite" and kaolinite may also occur.

Mica, apart from a little leached "hydrobiotite", is rare. The heavy minerals are similar to those found in the other sandstone types.

Pebbles of weathered clay ironstone and coal fragments are common. A little calcite and some pyrite euhedra may occur in the interstices of the rocks.

e. CONCLUSION

Several types of sandstone were recognised in the field which were found to differ in mineralogy when examined in the laboratory. There are associated variations in fabric. The diverse types recognised are usually found in different parts of the succession in the cyclothems.

The genetic significance of the various sandstone types is considered in the following chapter.

CHAPTER V

THE SANDSTONES: PART II

Their Classification and their Environmental Significance

a. INTRODUCTION

Previous workers have attempted to classify the sandstones either as a separate group of rocks or on broader outlines which include the limestones and the argillaceous rocks. Various factors have been stressed by different workers, for example:-

1. Environment - Twenhofel, 1950.
2. Tectonism - Jones, 1938, Krynine, 1948.
3. Tectonism and environment - Pettijohn, 1957.
4. Lithologic associations - Krumbein, 1954.
5. Dynamic processes - Griffiths, 1951.

In addition, many different combinations of mineralogical characters have been used in the last few years and a large number of classifications, based on ternary or quaternary diagrams, have been put forward. Some of the more important of these are discussed in chronological order.

b. PREVIOUS CLASSIFICATIONS

(i) Pettijohn, 1948

In 1948, Pettijohn classified the sandstones on the basis of a ternary diagram with three end members: quartz plus chert, feldspar and clay-sericite plus chlorite, (Figure II.1). (The diagrams of the other workers are oriented so that the positions of end members correspond, where possible, to those of Pettijohn.) Pettijohn defined quartzite as a sandstone containing 0 to 10% of feldspar and 0 to 20% of clay matrix with feldspathic quartzite having the same range of clay material but containing 10 to 25% of feldspar. The arkosite also contains 0 to 20% of clay material but has between 25 and 100% of feldspar. The subgraywacke contains 20 to 75% of clay material and 0 to 10% of feldspar, whereas the graywacke has the same range of clay material but contains 10 to 80% of feldspar.

According to Naumann (1858, page 663), who first defined the term, a graywacke consists of angular and rounded grains of quartz, small fragments of siliceous slate, phyllite and other rocks and, in many cases, feldspars, all bound together by a fine-grained matrix.

In the classification given by Pettijohn, the graywacke which he defined does not contain the essential constituents of rock fragments including low grade metamorphic rocks, thus departing from Naumann's definition. Graywacke, as defined by Pettijohn, is in fact an argillaceous feldspathic sandstone.

(ii) Krynine, 1948

Krynine also proposed a classification of sandstones in 1948. In his ternary diagram (Fig.II.2), he included kaolin with feldspar and rock fragments with mica and chlorite. Although he did not give definite percentages on the diagram and only one in the text, it appears that the quartz-chert end member, the orthoquartzite, contains 0 to 15% of clay material plus rock fragments plus feldspar. The arkose contains 10 to 100% of feldspar and kaolin and 0 to 20% of clay plus rock fragments. The impure arkose contains about 40 to 80% of kaolin plus feldspar and 20 to 60% of clay plus rock fragments. The low rank graywacke contains 0 to 20% of kaolin plus feldspar and 10 to 100% of clay micas and rock fragments. The high rank graywacke contains 20 to 40% of feldspar plus kaolin and 20 to 80% of clay micas plus rock fragments.

This classification is very different in its class limits from that proposed by Pettijohn. Krynine did not distinguish between argillaceous sandstone and graywacke proper and he did not specify the type of rock fragments occurring in the sandstones.

(iii) Tallman, 1949

Tallman, 1949, produced further classification which was different, both in class limits and the positions of some of the rock types. Although he gave no diagram it is possible to construct one from the text as in Figure II.3. Here, the constituents are quartz, feldspar and clay plus micas. The orthoquartzite contains 90% or more of detrital quartz and the subgraywacke field includes all rocks which contain between 10 and 25% of clay matrix. The graywacke field includes all rocks in which the clay component makes up more than 25% of the rock. The feldspathic quartzite and arkose both contain 0 to 10% of clay material. The feldspathic quartzite contains 10 to 25% of feldspar, whereas the arkose contains 25 to 100% of feldspar.

A gap occurs between the orthoquartzite, feldspathic quartzite and subgraywacke fields. This classification also contains no true graywackes, as defined by Naumann, but only argillaceous sandstones.

(iv) Dapples, Krumbein and Sloss, 1953

Dapples, Krumbein and Sloss, 1953, produced yet a further classification (Figure II.4.). This diagram is similar to that given by Krynine in that it includes rock fragments with matrix and has, as the other two poles, feldspar and quartz plus chert. However, the orthoquartzite field is much reduced in size and includes rocks with 0 to 5% of matrix and 0 to 5% of feldspar. The feldspathic sandstone contains 5 to 25% of feldspar and up to 5% of matrix whereas the arkose contains 25 to 100% of feldspar and up to 25% of matrix. The subgraywacke field includes the rocks with 5 to 25% of matrix and 0 to 25% of feldspar. The graywacke field includes all rocks with over 25% of matrix. Further subdivision of the rock types was made on the basis of sorting, using a three dimensional diagram in the form of a prism.

(v) Folk, 1954

Folk, 1954, realised the difficulties which occurred by including the detrital clays and produced a diagram, (Figure II.5), in which the chemical cements and detrital clays were disregarded. The end members of his ternary diagram include quartz plus chert, feldspar plus all igneous rock fragments and mica plus metamorphic and metaquartzite rock fragments. The orthoquartzites are defined as rocks which contain 0 to 5% of igneous rock fragments and 0 to 5% of metamorphic rock fragments. The subarkose contains 5 to 25% of igneous rock material and 0 to 10% of metamorphic rock material whereas the true arkose has the same range of metamorphic rock material but has 25 to 100% of feldspar and igneous rock fragments. The impure arkose contains 25 to 90% of igneous rock fragments and feldspar but also contains between 10 and 50% of metamorphic rock material. The subgraywacke and graywacke both contain less than 10% of feldspar and igneous rock fragments. The subgraywacke contains 5 to 25% of metamorphic rock materials and the graywacke, 25 to 100%. The feldspathic graywacke contains 10 to 50% of feldspar plus igneous rock fragments and 25 to 90% of metamorphic rock fragments. Finally there is a feldspathic subgraywacke field which contains 10 to 25% each of metamorphic and igneous rock material.

The difficulty with this classification is that the detrital clays are not included and many argillaceous rocks are, therefore, not placed.

(vi) Gilbert, 1954

Gilbert in Williams, Turner and Gilbert, 1954, classified the sandstones by forming two groups of rocks, the wackes, with 10% or more of detrital clay material, and the arenites, with less than 10% of detrital clay material, (Figures II.6 and 7). The three groups of constituents used for the poles of the diagrams were feldspar, the stable grains of quartz plus chert plus quartzite and the unstable grains of other rock fragments. The various fields are rather loosely defined and many of the rarer rocks are unnamed.

The quartz wackes and arenites contain 0 to 10% each of feldspar and unstable rock fragments. The subfeldspathic lithic wackes and arenites contain 10 to 50% of unstable rock fragments and 0 to 10% of feldspar. The lithic wackes and arenites are placed in a rather vague area which starts at about 10% each of feldspar and rock fragments and approaches 100% of rock fragments and 50% of feldspar. Nearer to these two extremities are located the volcanic arenites and wackes. The feldspathic wacke and arenite contain 10 to 25% of feldspar and up to about 25% of rock fragments. The arkose contains up to about 10% of rock fragments and 25 to about 70% of feldspar. The arkosic wacke and arenite lie in rather vague areas between the arkose and the lithic wacke and lithic arenite.

In this classification there has been some attempt at separating the rock fragments and detrital clays from the other components. However, the true graywackes, as defined by Naumann, are completely neglected.

(vii) Bokman, 1955

The picture was much simplified by Bokman, 1955, who gave as the three components: clay, quartz, and feldspar plus rock fragments, (Fig. II.8). His orthoquartzite includes all rocks which contain over 90% of quartz, whilst the protoquartzite includes all rocks which contain less than 90% of quartz but do not contain more than 20% of clay or more than 25% of feldspar. The area including rocks with more than 20% clay is the graywacke area. The arkose and lithic sandstone occur in the area between 25 and 100% of feldspar plus rock fragments and 0 to 20% of clay. Here, as in many of the other classifications, the defined graywacke is in fact an argillaceous sandstone.

(viii) Pettijohn, 1957

Pettijohn, 1957, gave yet another classification which was very different from the one he produced in 1948, in that it is based on a quaternary diagram. An exploded view of the tetrahedron is seen in Figure II.9. The four components are rock fragments, feldspar, quartz plus chert and detrital matrix. The limits of the types of sandstone are best expressed in tabular form as follows:-

	Quartz	Detrital Matrix	Rock Fragments	Feldspar
	%	%	%	%
Orthoquartzite	80.5-100	0-15	0-5	0-5
Protoquartzite	64-95	0-15	0-25	0-25
Lithic Graywacke	0-64	15-85	4-85	0-42.5
Subgraywacke	0-75	0-15	11-100	0-50
Feldspathic Graywacke	0-64	15-85	0-42.5	4-85
Arkose	0-75	0-15	0-50	11-100
Feldspathic Quartzite	11-85	15-85	0-21	0-21

In using four components, Pettijohn has departed considerably from his earlier classification. Although the rocks are better defined than in the previous classifications, the graywackes are still not the same as the original graywacke of Naumann.

c. GENERAL DISCUSSION OF SANDSTONE CLASSIFICATION

Several terms are used which require definition, the

the most important being sandstone. A sandstone is here considered as a sedimentary rock which contains 50% or more by volume of sand size terrigenous detritus. Sand size material is defined according to the Wentworth grade scale, being composed of particles between 1/16 and 2 mm. Siltstones with grain sizes of 1/256 to 1/16 mm are not considered here but can be classified in a similar manner.

Rocks containing more than 50% detrital clay matter are considered to be argillaceous rocks, while those containing more than 50% of material larger than coarse sand size are considered to be conglomerates.

The term graywacke, as has been shown in the previous paragraphs, has been used by different authors to refer to argillaceous sandstones and graywacke congomerates. Here, it will be used in the original sense which Naumann (1858) intended, that is, a rock consisting of angular and rounded quartz grains, small fragments of siliceous slate, phyllite and other rocks and, in many cases, feldspars, all bound together by a fine-grained matrix. The graywackes are associated with vigorous tectonic activity and the rocks of the graywacke series, that is, rocks with the characters of graywackes, are often found associated with geosynclinal deposits.

An arkose is a rock which is rich in feldspar and often contains the more stable igneous minerals. Rocks of the arkose series are derived from old granite terrains.

Arenite was used as a general term by Pettijohn (1957) to describe the clastic rocks of sand size material without any genetic or mineralogical connotations. Gilbert used the term to delimit sand size sediments containing less than 10% argillaceous matter. "Quartzite" with various prefixes has been used by Krynine and Pettijohn to describe sandstones of epeirogenic origin. However, the term quartzite also refers to rocks of metamorphic origin. The word arenite will be used here in a similar sense as Gilbert's use of the term to include sandstones of epeirogenic origin. Gilbert, however, placed an upper limit of 10% on the clay mineral content whereas the present worker places the upper limit of clay mineral content at 50%.

The arenites usually contain little if any first cycle igneous or metamorphic minerals or rock fragments except occasionally those derived from lavas.

All the classifications which have been put forward by the various workers are different and, as some workers have stated, not only the mineralogy but the degree of sorting, origin of the detritus, environment of deposition and other minor factors are all of importance when attempting to classify the different sandstones.

According to Pettijohn (1948), "No classification independent of rock genesis can be worthy of consideration". Here, three sources of origin of the sandstones are considered. These are the young tectonic areas which give rise to the graywacke series of rocks, old granite terrains which give rise to the arkose series of rocks and epeirogenic areas which give rise to the arenites. The arenites are classified under the group heading of orthoquartzite series. This term, used by Krynine (1941), is retained as the group heading because of its universal usage both in petrographic classifications and in palaeogeologic works.

Tectonically, the different series of rocks are related. Thus sediments from a stable area are of the orthoquartzite series. With increasing tectonic activity, low grade metamorphic rocks are exposed at the source of the material and the sediments enter the graywacke series. The old granite terrains, eroded at the termination of tectonic activity, give rise to rocks of the arkose series.

An examination of the quartz-rock fragment-feldspar triangle, (Figure II.10), indicates one of the inherent difficulties of the classifications which have been put forward. To move from any end of the triangle to another, through a series of rocks, may be the result of change in the tectonic environment. On the other hand, if moving from rocks rich in feldspar or rock fragments to the quartz end, it may equally well be the result of the distance from the source area with weathering effects destroying all but the more stable grains during transport.

An examination of the triangle: sand (quartz plus feldspar) - rock fragments - clay, (Figure II.11), indicates a further difficulty of the classifications. The absence of rock fragments and sand size particles from a sediment may be the result of the distance of transport and the absence of clay or silt size particles from a rock may be due to local sorting in the environment of deposition.

The position of second, third or more cycle sandstones presents a difficulty which, unless the tectonic setting is examined, cannot easily be overcome. Thus an arkose may, under conditions of desert weathering, give rise to a second cycle arkose in a non-tectonic environment.

In view of these difficulties, any classification must of necessity include an indication of the tectonic setting of the rocks. It must also include, amongst the mineralogical characters, detrital clay minerals and rock fragments in separate groups. A description of the type of rock fragments and the type of clay material is also necessary in that both are invariably related to the tectonic setting and the rock types.

If all of these characters are considered, the sandstones can be related to the other detrital rocks of terrigenous origin, the shales and clays and the conglomerates and sedimentary breccias.

It is here proposed to classify the sandstones in the following manner.

d. PROPOSED CLASSIFICATION OF THE SANDSTONES

Three series of sandstone are recognised according to their tectonic environments. The series can be subdivided into rock types on the basis of quaternary diagrams. The three series and the components of the diagrams can be listed as follows:-

i. Non-tectonic or orthoquartzite series: quartz plus chert and clay of variable composition are the important constituents. Some feldspar may occur. The rock fragments are usually of sedimentary origin.

ii. Late tectonic or arkose series: feldspar and kaolinitic clay are the important constituents, whilst quartz and rock fragments, mostly granitic, form the other two constituents.

iii. Early tectonic or graywacke series: low grade metamorphic rock fragments and chloritic clay are important constituents, whilst quartz and feldspar form the other two constituents.

Exploded views of the three quaternay diagrams are given in Figures II.12,13,14. These diagrams also show the trends of the changes in mineralogy due to weathering, transport and sorting. The orthoquartzite series are usually derived from pre-existing sedimentary rocks and thus the effect of weathering on the conglomerates will be negligible in determining the mineralogy of the sandstones derived from them. The environment of deposition of the graywacke series is such that again weathering has little effect because the rocks are usually deposited rapidly.

The heavy mineral suite may often be useful in the determination of the series and in the identification of second cycle rocks. Semi-stable minerals are particularly useful in this respect. The type of quartz may cccasionally be useful in distinguishing grains derived from granite terrains from those derived from metamorphic terrains.

1. The Non-tectonic or Orthoquartzite Series.

Only the rocks occurring near the quartz pole are common amongst the rocks of sand size range. Amongst these commonly occurring rocks, nine types are recognised. The limits of these types can be listed in tabular form as follows:-

		Quartz + Chert %	Rock Fragments %	Feldspar %	Clay %
1.	Quartz arenite.	95-100	0-5	0-5	0-5
2.	Impure quartz arenite.	90-95	0-10	0-10	0-5
3.	Subargillaceous arenite.	67.5-95	0-9.5	0-9.5	5-25
4.	Subargillaceous feldspathic arenite.	48-85.5	0-9.5	4-24	5-25
5.	Feldspathic arenite.	61.5-90	0-10	5-25	0-5
6.	Argillaceous arenite.	45-75	0-7.5	0-7.5	25-50
7.	Argillaceous feldspathic arenite.	37.5-67.5	0-7.5	2-19	25-50
8.	Pebbly arenite.	61.5-90	5-25	0-10	0-5
9.	Argillaceous pebbly arenite.	37.5-85.5	2-24	0-9.5	5-50

Whilst the limits are not in the form of "round" numbers such as 5, 10, 15, 25%, the limits are accurately fixed by using well defined boundaries on the base of the tetrahedron and joining these lines to the apex.

Any attempt to "round off" the limits of the classes could only be accommodated by distortion of the diagram or overlapping of the different types.

It is important that all rocks with greater than 50% of clay material should be classed as argillaceous rocks as these usually appear to be argillaceous in the field. Similarly all rocks with more than 50% of rock fragments are classed as conglomerates and breccias.

Whatever the composition of the parent material, providing it contains quartz of sand size or larger, the trend in composition of the sediment is towards the quartz-rich rocks,by sorting from the argillaceous rocks, by transport plus sorting from the conglomerates and by weathering plus sorting from the feldspathic rocks.

Weathering without sorting of the feldspathic rocks gives rise to more argillaceous rocks, whilst transport plus weathering without sorting of the conglomerates, which invariably contain some clay matter or shale pebbles, will also tend to result in argillaceous rocks.

2. The Late Tectonic or Arkose Series

Apart from the argillaceous rocks and the granite conglomerates, only three rocks commonly occur. These can be listed in tabular form as follows:-

		Quartz	Rock Fragments	Feldspar	Clay
1.	Arkose	0-75%	0-25%	22-75%	0-10%
2.	Argillaceous arkose.	0-65%	0-22%	12-68%	10-50%
3.	Lithic arkose.	0-50%	22-50%	22-75%	0-10%

Trends can be recognised on the diagram of the arkose series. Movement towards the more argillaceous rocks is due to weathering of the feldspars, whilst movement in the direction of the quartz - feldspar line is due to sorting. Passage from the granite conglomerate to the quartz pole may be accomplished by transport plus sorting and weathering whilst passage towards the arkose may be accomplished by transport without weathering. Movement from arkose to the argillaceous arkose may be accomplished by weathering whilst movement towards the quartz pole is due to weathering plus sorting.

3. The Early Tectonic or Graywacke Series

Apart from the graywacke conglomerates, only four rock types are recognised in this series. These can be listed as follows:-

		Quartz	Rock Fragments	Feldspar	Clay
1.	Graywacke.	0-81%	5-50%	0-8.5%	10-50%
2.	Feldspathic Graywacke.	0-71.5%	5-50%	5-25%	10-50%

(This is shown as (2) on diagram II.14 and lies within the tetrahedron adjacent to the graywacke and above the feldspathic subgraywacke.)

		Quartz	Rock Fragments	Feldspar	Clay
3.	Subgraywacke	30-90%	9-50%	0-10%	0-10%
4.	Feldspathic Subgraywacke.	16.5-81%	9-50%	9-25%	0-10%

Normally the environment of deposition is such that weathering is slight and sorting is not active during the deposition of the graywackes. If, however, the distance of transport allowed sorting and weathering, the sediments would trend in composition towards the quartz pole.

As has been pointed out in the section on the field classification, the Type 5 sandstones, with kaolinite, occur high in the succession studied and are more probably related to changes in the amount of coarser material and feldspar brought in from the source area.

The Type 6 sandstones are, as has been shown on field evidence, the result of local erosion and channelling without accompanying resorting.

The several sandstone types will be more closely related to the environment of deposition in a later chapter, after the limestone members of the sequence have been considered.

CHAPTER VI

THE LIMESTONES

a. INTRODUCTION

Several types of limestone were distinguished in the field on the basis of easily recognisable characteristics. In addition, gradations between limestones and shales or sandstones were recognised. The different types recognised in the field were examined in the laboratory by means of thin sections and their characteristics have been used to fit them into a general classification scheme.

Folk (1959) and Banerjee (1959) have shown that limestones can be classified on their field occurrence and petrography and the resultant classification can be used to elucidate their environmental conditions of deposition. Following these workers, an attempt has been made to relate the Teesdale limestones to conditions of deposition.

b. THE FIELD CLASSIFICATION OF THE LIMESTONES

Several characters were used to classify limestones in the field. These included:-

1. Colour which varied from light to dark grey and pale brown.
2. Texture which varied from fine (calcite mudstone) to coarsely crystalline or sparry.
3. Quartz content: the limestones were examined for quartz grains using a hand lens.
4. Mica content: some of the limestones contained mica which was visible to the unaided eye.
5. Shaliness: many of the limestones were shaly as opposed to thinly bedded.
6. Fossil content: many limestones contained abundant fossils, in particular, crinoids and brachiopods.

On the characters outlined above, six types of limestone were recognised in the field. More detailed description of these types is given below, together with a description of their modes of occurrence.

Type 1. Coarse-grained, pale, non-shaly limestones which contain no apparent quartz but do contain abundant crinoid ossicles. They often contain large corals or large brachiopods.

These form the main body of many of the thicker limestones and the whole of some of the thinner limestones.

Type 2. Coarse-grained, pale limestones which contain noticeable amounts of quartz. They are non-shaly and may contain a few crinoid ossicles or may be unfossiliferous.

These may form the whole of the thinner limestones and often form the base of the thicker limestones. A few thin lenses of this rock type may also occur in fine sandstones or siltstones.

Type 3. Fine-grained, dark grey limestones which contain no apparent quartz. They are non-shaly and contain some crinoid ossicles.

These form the whole of some of the thinner and thicker limestones and often form the tops to the thicker and paler limestones.

Type 4. Coarse-grained, dark limestones which contain no apparent quartz. They are non-shaly and contain some crinoid ossicles.

Only one sample of this rock type has been recognised and occurs as a thin minor limestone.

Type 5. Fine-grained shaly limestones to calcareous shales which contain no apparent quartz. They are dark grey and usually contain an abundant fauna.

These rocks occur as partings in limestones and also at the top of many limestones.

Type 6. Medium grey, shaly and arenaceous limestones which may contain crinoid ossicles and shells.

These rocks occur as thin bands in calcareous shales associated with the Three Yard Limestone.

Table 6.1 summarises the field characters of the various limestones recognised.

c. LABORATORY CLASSIFICATION OF THE LIMESTONES

Folk (1959), showed that limestones are made up essentially of three main constituents. These are the allochems, microcrystalline ooze and sparry calcite. Allochems are transported and otherwise differentiated carbonate bodies in the sediments. Folk recognised only four important allochems. These are intraclasts or reworked fragments of penecontemporaneous carbonate sediment, ooliths, fossil fragments and pellets.

Microcrystalline calcite ooze of 1 to 4 micron size forms the matrix and sparry calcite is coarser and forms the pore-filling cement.

The nomenclature is based on the dominant constituents. Thus a limestone composed of fossils in a microcrystalline ooze is called a biomicrite. This classification is given in Table 6.2.

Further differentiation is accomplished by the use of a grain size scale for both the transported and the authigenic constituents as in Table 6.3.

A difficulty often encountered in the classification of the limestones is the frequent lack of the positive identification of the original constituents of the rocks. Two effects which are encountered are recrystallisation and dolomitisation; both complicate the problem. "Ghost" structures can often be used to determine the original nature of the constituents.

It is here proposed to classify the limestones studied in this work on a basis similar to that used by Folk at the end of each petrographic description.

d. THE PETROGRAPHY OF THE LIMESTONES

A description of the petrography of the six limestone types is given in the following paragraphs and the main features are summarised in Tables 6.4 to 6.9.

Type 1. These rocks all contain sparry calcite and usually contain fossil debris. Some contain microcrystalline ooze.

The percentage of sparry calcite varies considerably not only from one limestone to another or within the same limestone from one locality to another, but also in the different sections taken from the same hand specimen. However, the sparry calcite usually forms more than 80% of the rocks and varies between 0.1 and 1.0 mm in grain size.

Microcrystalline ooze generally forms less than 10% of the rocks and the grain size varies between 1 micron and 5 microns.

Some rocks show evidence of recrystallisation in the occurrence of veinlets of sparry calcite. Most sparry calcite, however, appears to be original as most of the fossils and parts of the fine matrix are not recrystallised.

The Four Fathom Limestone and in some places, the Scar Limestone, is recrystallised with the formation of dolomite, R.I. ordinary ray, 1.6800 (determined also by staining and X-ray analysis). In Hudeshope Beck (grid ref. 948272), a coral band in the Four Fathom Limestone which has been recrystallised with the formation of dolomite, also contains chalcedonic silica in the corals. The dolomite rhombs have, in places, caused dislocation of minor bedding structures and the dolomite is, therefore, thought to have formed before the limestones became fully consolidated.

A little anhedral to euhedral pyrite often occurs, replacing calcite.

Most of the limestones contain crinoid debris and many contain brachiopods or corals.

These limestones vary from crinoidal microsparites to coral or brachiopod biosparites and occasionally micrites. Examples of some of these rocks are illustrated in Figs. 6.1 and 6.2.

Type 2. These rocks contain quartz in a sparry calcite matrix. The quartz is usually angular to subangular but occasionally may be subrounded. The grains vary between 0.025 and 0.5 mm in size but most of the rocks contain well sorted grains which vary between 0.025 and 0.25 mm in size. The quartz content varies between 2 and 50%.

Some rocks contain accessory detrital minerals similar to those found in the sandstones such as tourmaline, zircon and plagioclase. The plagioclase is always fresh and appears, from the few grains measured, to be in the oligoclase-andesine range. Micas such as muscovite, biotite and chlorite also occur in some of these rocks.

The calcite varies from fine to coarse grained sparry calcite. A sample of a calcareous lens of material from the sandstone below the Single Post Limestone in Birk Sike (829334), T.210, was composed of quartz grains and brown micrite aggregates in a calcite matrix. Although the brown carbonate was too fine-grained for optical work, X-ray analysis (Table 6.10) revealed that the rock was composed of quartz, calcite and chalybite and the brown material is therefore believed to be chalybite.

Euhedral to anhedral pyrite, replacing calcite, forms up to 1% of some of the rocks.

Some of the rocks contain crinoid ossicles, foraminifera, bryozoa or, occasionally, corals.

These limestones vary between microsparites and sparites. Examples of these rocks are illustrated in Figs. 6.3 and 6.4.

<u>Type 3</u>. These are fine-grained limestones which contain abundant fossils and little (generally less than 1%), or no quartz.

The calcite varies in grain size between 1 micron and about 15 microns.

A little very fine interstitial opaque matter occurs in the rocks although the percentage (less than 1%) was too small to permit extraction, separation and identification. The optical characters, low reflectivity, black with small points of light reflected from the surface, suggests that the mineral is carbonaceous. Extraction with carbon disulphide was attempted but no results were obtained, suggesting that the limestones are free of bituminous matter. Anhedral to euhedral pyrite grains generally less than 0.25 mm diameter form up to 2% of some of the rocks.

Quartz grains, where they occur, form less than 1% of the rock and are less than 0.15 mm in diameter.

Fossils are common in these limestones and include algae, bryozoa, foraminifera, shell fragments, crinoid ossicles and occasionally corals.

These rocks can be classified as different types of biomicrites and biomicrosparites. An example of this limestone type is illustrated in Fig. 6.5.

Type 4. Only one sample of this rock type was recognised. Although quartz was not apparent in the field, the rock contains about 7% of subangular to subrounded quartz of grain size 0.05 to 0.75 mm diameter.

The rock contains abundant foraminifera, bryozoa, brachiopod fragments, crinoid ossicles and occasional pellets in a matrix of fine sparry calcite and ooze.

The rock can be classified as a foram intramicrite.

This rock is shown in thin section in Fig. 6.6.

Type 5. In thin section these rocks are rather variable but all have the common feature of bearing a medium to large amount of clay mineral. Most of the rocks are very opaque due to the presence of the clay minerals and also possibly due to the presence of carbonaceous material. The matrix is composed of clay mineral and fine carbonate but in some of the rocks intraclasts of coarsely crystalline calcite occur. Fossils include complete shells and shell fragments, crinoid ossicles, bryozoa, foraminifera and,rarely, algae.

Occasionally a very little very fine quartz may occur, the grains being up to 0.0015 mm in diameter.

Pyrite euhedra and anhedral masses occur in variable amounts and may form up to 50% or more in places.

The clay minerals, identified from X-ray analysis, are illite with minor amounts of chlorite.

Most of these limestones can be classed as argillaceous biomicrites. A typical example of this rock type is shown in Fig. 6.7.

Type 6. These are arenaceous argillaceous limestones. Quartz forms up to 30% of the rocks and varies from 0.01 to 0.175 mm in diameter. The quartz varies from angular to rounded but is generally subangular and angular.

The clay minerals are masked by the carbonate but X-ray analyses reveal that the clay minerals are illite with minor amounts of chlorite.

Mica flakes up to 0.25 mm in length invariably occur and usually form up to 1.5% of the rocks. The mica is generally muscovite but may be "hydrobiotite". "Hydrobiotite" occasionally forms up to 4% of some rocks and the flakes may be up to 0.4 mm length. A little pale green or emerald green chlorite may also occur.

Rounded grains of carbonaceous material sometimes occur with the mica but in very small amounts.

Accessory minerals include plagioclase and occasionally tourmaline.

Pyrite occurs in anhedral masses which form up to 3% of the rocks. The pyrite replaces matrix, shells and quartz grains and is therefore believed to be of secondary origin.

Those Type 6 rocks which occur above the Cockle Shell Limestone and the Three Yard Limestone are unfossiliferous whilst those occurring below the Three Yard Limestone are rich in brachiopod fragments.

These rocks can be classed as arenaceous microsparites and biomicrosparites. Typical examples of these rocks are shown in Figs. 6.8 and 6.9.

e. THE ENVIRONMENTAL SIGNIFICANCE OF THE DIFFERENT LIMESTONE TYPES

The interpretation of the depositional environments of the different limestone types is rather more difficult than for other rock types, principally because the rocks are usually monomineralic and also because calcite is often affected by recrystallisation. In accordance with studies made by Folk (1959) and Banerjee (1959), the following comments can be given on the origin of the different types:

Type 1. These contain abundant shell and crinoid fragments and have an abundance of sparry calcite. Argillaceous and other terrigenous detritus is absent or occurs but sparsely. These limestones were probably deposited in clear water in which lived a large number of organisms and which was sufficiently clear to support occasional coral colonies. The unbroken nature of many of the fossils suggests that wave action was insignificant.

Type 2. These contain few fossils and little or no argillaceous material. They occur at the bottom of the thicker limestones and as lateral facies of the thinner limestones. They can often be seen to grade into sandstones.

They were probably deposited in clear marine water areas where current action was sufficiently strong to bring in sand grains from sand banks occurring on the sea floor. At the same time, the water was sufficiently clear and rich in carbon dioxide to allow the chemical precipitation of calcite.

Type 3. These are fine-grained and fossiliferous but contain some argillaceous material. They were probably deposited in areas of quiet water as a foraminiferal ooze into which currents were bringing small quantities of argillaceous material.

Type 4. As has been pointed out, this type of limestone is very similar to the Type 3, except for the presence of quartz and the large number of macro-fossil fragments.

This type was probably deposited in an area of quiet water receiving some argillaceous material but in which gentle current action brought in sand grains from neighbouring banks and also removed most of the calcite mud.

Type 5. These are the shaly limestones to calcareous shales and represent the incoming of land derived detritus in the form of fine argillaceous material. Where this was of a temporary nature the band is succeeded by further purer limestone deposition.

Type 6. These are arenaceous shaly to non-shaly limestones and represent calcareous deposition in the presence of unsorted, land derived detritus.

f. CONCLUSION

Field examination of the limestones in Teesdale reveals that several different types can be recognised. These types are significantly different in their petrography and the differences can be related to variations in the environments of deposition.

CHAPTER VII

THE CLAY MINERALS IN RELATION TO THEIR ENVIRONMENTS OF DEPOSITION

a. INTRODUCTION

When attempting to relate the clay mineral composition of a sediment to its depositional environment, the more important factors to be considered include:-

1. The nature of the original minerals in the source area.
2. The effects of the transporting agent.
3. The physical effects of the depositional environment.
4. The chemical effects of the depositional environment.
5. Diagenesis.
6. Post-diagenetic effects, including weathering.

A difficulty which is encountered in the interpretation of the depositional environments of argillaceous sediments is the use and meaning of the term "diagenesis". This term has been used by different workers in different contexts and thus the descriptions of modern sediments are sometimes difficult to interpret in terms of alteration of the clay minerals.

Von Gümbel (1888) defined diagenesis as "a term denoting the sum of successive changes which take place in exogenetic rocks as a result of continued sedimentation above them, or the percolation of groundwaters through them; e.g. consolidation, development of lamination by increase of vertical pressure due to the superincumbent load, and cementation and recrystallisation due to seasonal and other regular changes in the water content and temperature of the normal exogenetic character". This includes all changes incurred during or after deposition (excluding the effects of metamorphism).

Pettijohn (1954) stated "..... the principal diagenetic processes are cementation, diagenetic reorganisation (authigenesis), diagenetic differentiation and segregation, diagenetic metasomatism (e.g. dolomitisation), intrastratal solution and compaction". Authigenesis and cementation do, in some cases, take place during deposition. Intrastratal solution may also occur during the weathering processes.

In 1953, Strakhov proposed the following three definitions:-

1. Sedimentogenesis - the physical and chemical effects of the environment of deposition during deposition.
2. Diagenesis - the transformation of a sediment into a sedimentary rock.
3. Epigenesis - the stage of alteration of an already constituted rock under tectonically produced changes in the physical and chemical conditions of its existence (excluding metamorphism and weathering).

Epigenesis and diagenesis are difficult to separate in many cases. Not all sediments, for example, are consolidated or transformed into solid rocks. Another difficulty is found in Strakhov's definition of epigenesis. His definition includes tectonically produced changes but does not specifically include epeirogenetically produced changes. He also excludes the weathering processes and to separate epigenetic processes and weathering is difficult in practice. Not only the surface and subsurface weathering may alter the mineralogy of a sedimentary rock. Potter and Glass (1958) have shown that even subvadose weathering may occur. However, Strakhov's definitions are probably the best so far set out and diagenesis as discussed in this chapter will follow his definition as closely as possible.

Changes occurring during and after the burial of the sediments often overlap so that the effects of sedimentogenesis and diagenesis are mutual and continuous as are those of diagenesis and epigenesis and also epigenesis and weathering. Where possible, however, distinctions are made. Thus the syngenetic alteration of sedimentary minerals during deposition (e.g. the formation of glauconite), due to the chemical effect of the environment, is considered as belonging to the process of sedimentogenesis and not diagenesis. In some cases, it is not possible to relate the formation of a particular mineral to any one stage in the history of a sedimentary rock.

b. THE NATURE OF THE ORIGINAL MINERALS IN THE SOURCE AREA

A very important factor determining the final clay mineral assemblage of a sedimentary rock is the nature of the original clay minerals in the source area. This has been emphasised by Weaver (1958) and much of the information here set out is taken from his work.

Clay minerals are formed principally by the chemical weathering of silicates, almost invariably the aluminosilicates, and by the hydrothermal alteration of pre-existing rocks. Of these two modes of origin, only the former is usually important on a regional scale. The clay minerals occurring in soils are, therefore, derived principally from the chemical weathering of silicates and the physical and chemical weathering of the clay minerals occurring in sedimentary and low grade metamorphic rocks.

Grim (1953) (1958) and Keller (1958) have summarised the formation of the various clay minerals under different weathering and hydrothermal conditions.

(i) Hydrothermal alteration

The effects of hydrothermal alteration have been summarised by Grim (1953), (1958). The principle can be given in broad outline thus: the types of clay mineral formed by hydrothermal action are dependent on the pH of the solutions and their alkali-base content.

Thus solutions containing magnesium ions tend to give rise to the development of montmorillonite or, at higher concentrations, chlorite. Caillère and Hénin (1949) have shown experimentally that the interaction of montmorillonite with magnesium-rich solutions produced a mineral tending towards the chlorite structure. Mica minerals, such as muscovite and sericite, form in the presence of potassium-bearing solutions, whilst neutral to acid solutions, devoid of alkalies or alkaline earths, result in the formation of kaolinite.

Grim has shown that the hydrothermal alteration of volcanic rocks, which may take place on a regional scale, produces montmorillonite or halloysite, kaolinite and allophane, depending largely on the magnesium content of the rocks.

(ii) Weathering.

Millot (1952), Rivière (1952), Keller (1958), Milne and Earley (1958), Weaver (1958) and Harrison and Murray (1958) have shown that the types of clay mineral occurring in a soil depends primarily on the nature of the parent rock, the climate, the vegetation cover and the topography.

If the climate and topography favour vigorous leaching action, kaolinite or silica plus iron and aluminium hydrates tend to be formed. When the pH is neutral to acid and the parent rock contains a relatively high concentration of $Al(OH)_3$ and when the hydrogen ions occur in the silicate weathering system, kaolinite tends to be formed. Keller (1958) has shown that various factors may give rise to this condition. These are:-

1. Parent rock rich in Al and possibly Na and K.
2. Silicate parent rock rich in Na and K ions, which stabilise the removal of SiO_2.
3. Parent rock rich in metal ions but subject to intense leaching by groundwaters with removal of the metal ions and alkaline hydrates.

4. Humid climate with efficient drainage.
5. pH neutral to acid.

A relatively low concentration of $Al(OH)_3$ and SiO_4 ions in the weathering system, due to a very humid climate or due to a high abrasion pH (see below), may still result in deposition of bauxite if a slight decrease in pH occurs and the $Al(OH)_3$ reaches its saturation point. Stevens and Carron (1948) define the abrasion pH of a mineral as the characteristic pH developed at the interface between the mineral surface and the surrounding water medium during hydrolysis of the mineral, effected by grinding the mineral under water. Keller uses the term for the naturally developed pH at the surface of colloid size particles in contact with water.

Grim (1958) has pointed out that in the early stages of the leaching process, micas may be degraded and mixed-layer minerals developed by partial removal of the interlayer potassium from the micas and magnesium from the brucite layers of the chlorites. If erosion is sufficiently rapid, the weathering processes may not result in the formation of kaolinite.

Similar results were shown to occur in the weathering of a till by Droste (1956). Here, chlorite was partially leached with the formation of a mixed-layer chlorite-vermiculite and finally vermiculite. Illite became hydrated shortly after in the weathering process and resulted in the formation of a mixed-layer montmorillonite-illite.

If the climate and topography do not favour leaching, alkalies and alkaline earths may be concentrated in the zone of weathering so that illites, montmorillonites, chlorites and sepiolite-attapulgite are formed. The nature of the minerals depends on the amount of potassium and magnesium. Grim (1958) has shown that the presence of large amounts of sodium may result in the formation of zeolite minerals.

The composition of the parent rock may greatly influence the clay mineral composition of the soil derived from it. Milne and Earley (1958), Weaver (1958), Jackson (1958) and Taggart and Kaiser (1960), have made studies on the effect of the source rock and the source area on the derived clay mineral assemblages. To quote an example from Milne and Earley: "Montmorillonite which is the predominant clay mineral of the Mississippi River and Delta sediments, is apparently the stable mineral of soil development and rock weathering in the drainage basin of the Mississippi River. The sediments of the Mississippi Sound-Mobile Bay area, east of the Mississippi Delta and derived ultimately from the Appalachian Province, contain considerably more kaolinite. Transition zones between the two sediments can be differentiated".

Sedimentary and low grade metamorphic rocks may show little alteration on physical disintegration if the effects of chemical weathering are weak or absent. Illites and chlorites generally occur in greater percentage than the other clay minerals in the ancient sedimentary rocks and metamorphic rocks. The derived soils would be expected, therefore, to contain these minerals with additional degraded mixed-layer minerals.

Glacial tills may contain a variable assemblage of physically derived material and the soils formed from the weathering of the tills may, therefore, contain a wide variety of clay minerals.

Volcanic rocks, in particular the basic rocks, tend, if subject to moderate weathering conditions, to give rise to montmorillonite-rich soils.

The high grade metamorphic and crystalline rocks may give rise to a variety of clay minerals, dependent on their compositions and their modes of weathering. Thus a potash-rich granite would give rise to illite on non-acid weathering but would yield kaolinite on vigorous leaching.

In summary, it can be stated that the clay mineral assemblage formed during weathering depends on the following factors:-

1. The nature of the parent rock.
2. The type of climate.
3. The drainage (dependent on topography and vegetation cover).
4. The intensity of weathering (dependent on 1 and 2).
5. The effectiveness of transporting agents in soil removal.

c. THE EFFECT OF TRANSPORTING AGENTS

During transport from their place of origin to their place of final deposition, the clay minerals may be altered in response to the environment provided by the transporting agent. Glacial, aeolian and fluvial transport may be expected to result in some physical abrasion of the minerals but, due to the small particle size, this effect will not be important. From the aspect of chemical change, only the fluvial environment may be expected to have any large effect.

Brown and Ingram (1954), studying the clay minerals of the Neuse River sediments, found that abrasion and "diagenesis" cause a decrease in kaolinite downstream with degradation of the kaolonite near the river mouth. Accompanying this effect, is a downstream increase in the amount of amorphous material and mixed-layer aggregates and the appearance of "chlorite". They postulated that kaolinite and montmorillonite alter to degraded kaolin and amorphous material which alter, in turn, to mixed-layer aggregates. The authors point out, however, that these changes are more noticeable near the estuary and are possibly facilitated by the estuarine environment.

Interrupted transport may, by temporary deposition in a new environment, affect the clay mineralogy considerably. Deposition as alluvium may subject the clay minerals to further weathering, possibly of a different nature to that which took place in the source area. Although interrupted transport by deposition as alluvium may occur quite commonly, the amount of material involved, compared to the amount of material carried without interruption, is probably very small.

Transport by rivers may affect the clay mineralogy of the transported material but the effects are believed to be small. Grim (1958) has pointed out that chemical abrasion may occur with removal of the interlayer potassium from the micas, resulting in degradation and the formation of mixed-layer minerals. Diphormic or two-layer minerals, montmorillonites and degraded micas would probably remain unchanged and any regeneration or formation of triphormic or three-layer minerals is considered improbable.

d. THE PHYSICAL EFFECTS OF THE ENVIRONMENT OF DEPOSITION

The physical and chemical effects of the environment of deposition are generally closely related, particularly in a marine environment. The marine environment only is considered here, as the strata studied are believed to have been deposited in marine, quasimarine and possibly freshwater environments. The factors which can be considered as being purely physical are:-

1. The effect of particle size.
2. The effect of the meeting of two different fluid environments of different salinity and hence density.
3. The effect of current action and turbulence.

Particle size is of great importance since it may influence the distribution of the clay minerals in an environment, particularly when the other two physical factors are prominent. Particle size is generally affected by chemical action in a marine environment as will be discussed later. In any environment of a fluid medium in which small particles are in suspension, providing the current action is negligible, the coarser particles will settle first and the finer particles later and much more slowly. If two different clay minerals are brought into an environment, each with its own particular size range, separation will be effected by their velocities of settling.

A typical example of the effect of two different environments meeting is seen in the Gulf of Mexico. Bates (1953) and Moore and Scruton (1957) have shown that the Mississippi River water moves over the denser saline waters of the Gulf and plumes of sediment are "jetted" up to 65 miles into the Gulf. This prolonged suspension in the marine environment would aid both chemical action and physical separation.

environments it is probably relatively unimportant in the marine environment and, as will be shown below, most of the minerals are reconstituted from degraded materials.

Dietz (1942), discussing the clay minerals occurring in Recent marine sediments, was of the opinion that the clay minerals increased in size from shallow water to the ocean deeps and areas of very slow sedimentation. He also believed that montmorillonite only occurred in nearshore sediments. He therefore postulated that illite may, in part, form from montmorillonite by the absorption of potassium. This would account for the potassium/sodium ratio of seawater as compared with river water and would also account for the increase in clay particle size due to growth in the deeper oceanic areas. Dietz also believed that this resulted in a general absence of montmorillonite from the deep sea sediments and also from most shales.

Murray and Harrison (1956), however, have shown that the most abundant clay mineral in sediments from the Sigsbee Deep is montmorillonite.

Millot (1952) cited the association of ancient shallow water marine lagoons containing illite with "fireclay" deposits containing kaolinite as evidence of the "néoformation" of illite in the lagoons, as both clays had a common origin.

Harrison and Murray (1958) have shown that illitic shales, on weathering, may be leached to yield soils containing clays of the mixed-layer type, the weathering profile being very similar to that of the "fireclays". It does not seem unreasonable to assume that more acid leaching would lead to the destruction of mixed-layer minerals and result in the formation of kaolinite. Thus the "néoformation" of kaolinite from illite is an equally possible theory and the existence of weathering profiles in many "fireclay" deposits suggests that this is the more probable answer.

Grim, Dietz and Bradley (1949) and Grim and Johns (1954) (1958) have shown that the principal effects of the marine environment are regradation of degraded illite and chlorite by absorption of potassium and magnesium respectively. Montmorillonite is apparently unaffected, although Grim, Dietz and Bradley (1949) state that they thought that kaolinite and montmorillonite decreased seawards due to very slow transformation. The regeneration of illite is rapid and thought to be completed before deposition.

Bradley (1953) has shown that the montmorillonite-illite mixed-layer structure, on diagenesis in sea water with absorption of potassium and magnesium forms mixed-layer chlorite-illites. Inaccurate interpretation of the results by many earlier workers may have shown that the montmorillonite was lost in sea water and that the chlorite was entirely "néoformational".

Weaver (1958), has shown that most of the characteristics of the clay minerals occurring in marine sediments are inherited from the source areas of the detritus. He has shown that most of the expandable clay minerals occurring in marine sediments are true montmorillonites and that the mixed-layer minerals rapidly regrade by potassium absorption and fixation. Caillère and Hénin (1949), have shown experimentally that true montmorillonites subject to strong chemical action are altered with great difficulty and do not form true micas or chlorites.

Work by Brown and Ingram (1954), Griffin and Ingram (1955), Powers (1957) (1958) and Keller (1956), lends support to the theory of regeneration.

"Glauconite" is characteristic of the marine environment. Burst (1958), has shown that the mineralogy is very variable and that several different types of "glauconite" occur. These include ordered and disordered mica-type lattices, the former corresponding to true glauconite. Other types exist including interlayered swelling minerals and mixed-layer 14 and 10Å minerals. Burst's survey of the literature shows that all types of "glauconite" are authigenic and can form from biotites, illites and also from faecal pellets. "Glauconite" is not, however, found in the strata studied.

It does appear that, apart from glauconite, most of the clay minerals occurring in marine sediments are inherited and that few are authigenic.

Weaver (1958) has stated that the experimental work by Yoder and Eugster (1955), has shown that the stable low temperature forms of muscovite and illite are the 1M and 1Md varieties, whereas the high temperature form is the 2M variety. In Recent marine sediments, the 1M and 1Md varieties are rare and are practically restricted to the glauconites whereas the illites are mostly of the 2M variety. It appears, therefore, that the 2M illites in sediments are detrital and are not authigenic. However, where ancient sediments contain 1Md illites among the clay minerals, these are mostly land derived from the weathering of non-clay silicates or from pre-existing rocks containing 1Md illites.

Chlorite presents a similar problem. Most of the minerals reported as altering to chlorite are dioctahedral, whereas the chlorite itself is generally trioctahedral. Dioctahedral chlorites are rare or absent from marine sediments and if they are formed in a marine environment, the "diagenetic" process must be relatively ineffectual. Chlorites in sediments are mostly trioctahedral and are, therefore, detrital in origin.

Grim (1958), has shown that diphormic (two-layer, Mackenzie, 1959) clay minerals are not in equilibrium with the marine environment. However, the adsorptive capacity of these minerals is very small and also the structural changes required to form triphormic or three-layer clay minerals are major changes. It might be expected, therefore, that clay minerals such as kaolinite will occur in the marine environment and will alter only very slowly and remain relatively unchanged over long periods of time, the transformation to the triphormic minerals being incomplete. The rate of sedimentation would, therefore, exert a large influence on the amount of material changed. True bentonitic montmorillonite might be expected to remain relatively unaltered in a marine environment and would probably alter only very slowly to regularly stratified mixed-layer minerals, whereas the detrital, generally random, mixed-layer minerals would rapidly regrade to micas.

f. DIAGENESIS

As has already been pointed out in the introduction to this chapter, the term "diagenesis" has been used rather loosely in the past. For the purpose of this work, the only diagenetic factors which are considered in relation to the clay minerals are authigenesis (diagenetic reorganisation), intrastratal solution and compaction.

Weaver (1958), has pointed out that post-depositional alteration of the clay minerals in permeable and porous sands and sandstones is common. Potter and Glass (1958), and Novin (1952), have also illustrated this phenomenon. In shales and clays, however, this effect is small, Ross (1955) and Grim (1958), have pointed out that volcanic material interred with marine sediments may be expected to alter to montmorillonite by reaction with the intrastratal connate water containing potassium and magnesium ions. Bradley (1953) showed that the illite-montmorillonite mixed-layer structures, on diagenesis in sea water, form mixed-layer illite-chlorites by absorption of potassium and magnesium.

According to Murray and Harrison (1956), in the Sigsbee Deep, illite and chlorite develop a better crystallinity a few millimetres below the sediment interface. Milne and Earley (1958), have shown that in buried Pleistocene and Tertiary sediments, a little illite may be developed in the shales, possibly due to absorption of potassium by montmorillonite with pressure and time. The sands of the Miocene sediments show developments of chlorite, possibly from montmorillonite, as montmorillonite occurs in the associated interfingering shales.

Powers (1958), has recognised that magnesium is preferentially adsorbed near the sediment interface and that with increasing depth, more potassium and less magnesium is adsorbed. At a certain depth, both will be adsorbed in equal quantities. This depth is the "equivalence level"which will vary with the physical characteristics of the sediment, the ionic potential of the two ions and their relative concentrations.

Weaver (1958), Burst (1958) and Powers (1958) believe that the depth of burial is an important factor in the destruction of montmorillonite in sediments. Burst showed that montmorillonite becomes less evident in the Gulf Coast Eocene below 3,000 feet. Between 3,000 and 14,000 feet the montmorillonite lattices are interspersed with illite components which increase in frequency with depth. Below 14,000 feet there is a virtual absence of montmorillonite. Chlorite appears to be more predominant and possibly increases in crystallinity with depth.

Weaver (1958), has pointed out that mixed-layer illite-montmorillonite and montmorillonite do increase in amount with depth but this may reflect a difference in the source areas between older and younger sediments. Weaver also pointed out that a review of montmorillonite-bearing rocks suggests that below 10,000 feet, montmorillonite gives way to mixed-layer illite-montmorillonite and that deeper burial causes a reduction in the number of montmorillonite layers.

g. POST-DIAGENETIC EFFECTS

Although they are due to different causes, the epigenetic effects and weathering are often difficult to differentiate in practice. They are, therefore, considered together under the term "post-diagenetic effects". Post-diagenetic effects can be subdivided into:-

1. The effects of groundwater solutions.
2. The effects of weathering.
3. The effects of hydrothermal solutions.

The effects of hydrothermal solutions will not be considered here as the strata studied, with few possible exceptions, have not been affected by such solutions. These exceptions are discussed in a later chapter.

(i) The effects of groundwater solutions and weathering.

The effects of groundwater solutions and weathering on the clay minerals in sediments are inseparable and, in most cases, tend to produce the same effects. The porous and permeable rocks such as the sandstones are particularly susceptible to these effects.

Potter and Glass (1958) have shown that the clay minerals occurring in sandstones may have originated in several different ways. The clay minerals may have been washed into the interstices of the rock either during or after deposition but prior to final burial. In the latter case, the clay minerals may appear to be original but were not deposited contemporaneously with the sand grains. Due to their porous nature, many sandstones may have a constant flow of groundwaters through them. Groundwaters may have a varying pH which may alter the nature of the original clay minerals in addition to causing the formation of new minerals in situ.

Several examples of clay mineral authigenesis can be seen in the Teesdale sandstones including "illites" from biotite and muscovite and fine sericite from quartz. Well crystallised kaolinite occurs in the interstices of some of the more porous sandstones. As such delicate "book" structures are incapable of transport without disaggregation, they are thought to have developed at the expense of other silicates, probably feldspars, and are of secondary origin. Smithson and Brown (1954),(1957), have recorded similar effects in other areas.

Weaver (1958), has shown that the occurrence of secondary kaolinite is dependent on the porosity of the sandstone. This appears to be the case in the strata studied, where the argillaceous sandstones and the shales which are impermeable, or almost so, do not show the development of secondary kaolinite. Potter and Glass (1958), have shown that these secondary effects may be expected to be more pronounced in surface outcrop, where both physical and chemical weathering are intensified. Surface samples of shales and subgraywackes (their nomenclature) may differ in clay mineralogy from subsurface samples but the differences are much less than differences observed in the porous sandstones.

In the present work, most of the shale and argillaceous sandstone samples selected showed no visible signs of weathering and only the freshest samples obtainable were collected. As many of the sandstones and limestones contain interbedded shale bands, these were sampled as representative of the clay minerals deposited during the time of sandstone or limestone deposition.

h. CONCLUSION

The clay mineral assemblage of any sedimentary rock is related to the nature of the original minerals brought into the environment of deposition, the nature of the environment, diagenetic processes and weathering. Where, as in the Teesdale rocks, the clay minerals are believed to have a common source area and where subsequent diagenesis and weathering has been similar in all rock types, apart from the porous sandstones, differences in clay mineralogy strongly suggest variations in the depositional environments of the rock types.

CHAPTER VIII

RECENT SHALLOW WATER MARINE AND DELTAIC SEDIMENTATION

a. INTRODUCTION

The purpose of this investigation has been to interpret the depositional environment of the Middle Yoredale sediments. A review of Recent sedimentation and depositional environments have helped in this interpretation.

Pettijohn (1957), gave the following modified form of Barrel's geographical classification of environments:-

I. Continental (above tidal reach).

  A. Terrestrial.
    1. Glacial.
    2. Aeolian.
    3. Pluvial.
    4. Fluvial.

  B. Paludal.

  C. Lacustrine.
    1. Freshwater
    2. Saline.

II. Mixed Continental and Marine.

- A. Littoral.
  - 1. Tidal lagoon.
  - 2. Beach.
- B. Delta.
- C. Estuarine.

III. Marine.

- A. Offshore (Neritic Facies).
- B. Epicontinental Sea.
- C. Oceanic and Mediterranean (Bathyal and Abyssal).

The following characteristics of the Yoredale rocks narrows the choice of environments:-

1. The occurrence of marine limestones and shales.
2. The occurrence of subaerial deposits (underclays and coals).
3. Shallow water depositional structures.
4. The great variety of rock types.

These characteristics and the lack of characteristics which might indicate the presence of other environments narrows the choice to mixed shallow water marine and continental and marine environments. Several sub-environments may occur in areas of mixed continental and marine environments. For example, an estuary will have as marginal facies, a beach and may also include a tidal lagoon.

The great variety of deposits in the Yoredales, both in the area studied and over the whole of Northern England, and the occurrence of channel deposits in the Upper Yoredales (Dunham, 1948), are more typical of the deltaic environment than any other. The rapid variations in the deposits, in particular the lateral variations, are embraced only by the deltaic environment. It is proposed, therefore, to consider only the shallow water marine and deltaic environments when relating the Yoredale type of sedimentation to Recent deposition.

Perhaps the most intensively studied area of shallow water marine sedimentation is the Gulf of Mexico, which also includes the most intensively studied area of deltaic sedimentation, the Mississippi Delta (Figure 8.1).

b. THE SHALLOW WATER MARINE AREA OF THE GULF OF MEXICO

Gould and Stewart (1955) studied the Continental Terrace sediments in the northeastern area of the

Gulf of Mexico. This is a shallow water area between 0 and 100 fathoms deep. In the east, off the west coast of Florida, the amount of terrigenous detritus brought into the area of sedimentation is negligible. In the northern part of the area, off the coast of Alabama, rivers bring in large amounts of detritus.

During the Pleistocene epoch the sea level was lowered and remained at a lower level for a considerable time. This lower level resulted in the formation of a wave cut platform at about 35 fathoms.

Off the west coast of Florida, limestone deposition has occurred during the Pleistocene and Recent epochs. Near the peninsula, where small rivers empty into the Gulf, terrigenous material forms part of the sediments. The general distribution of sediments in the area is shown in Figure 8.2. The sediments are arranged zonally, parallel to the shore, and consist of the following:-

1. Quartz-shell sand with over 50% quartz.
2. Quartz-shell sand with less than 50% quartz.
3. Shell sand.
4. Algal sand.
5. Oolite sand.
6. Foram sand and silt.

Reef building corals are found in the area and form small mounds, mostly shoreward of the 10 fathom depth, but do not form true reefs. The absence of major reef builders from the area is attributed to the low winter temperature of the water of 18 to 20°C. Consolidated Pleistocene limestone occurs on the shelf, partially covered by unconsolidated sediments.

Off Alabama and northern Florida, sampling of the unconsolidated sediments revealed an absence of the oolite sand (Figure 8.3.). Off Florida the zones recognised were:-

1. Quartz sand.
2. Narrow zone of quartz-shell sand.
3. Zone of algal sand narrowing to the northwest.
4. Foram sand and silt.

The zones of quartz-shell sand and algal sand narrow to the northwest so that off the coast of Alabama only quartz sand and foraminiferal sand and silt occur. The greater quantity of terrigenous material in the northwestern part of the area is due to the greater supply of detritus from the rivers of northern Florida and Alabama. There has also been a more complete removal of older coastal plain deposits from the outer part of the shelf off central Florida during the more recent marine transgression.

c. THE DELTAIC AREA OF SEDIMENTATION, THE MISSISSIPPI DELTA

Fifty miles west of the northern part of the area of marine sedimentation described above is the Mississippi Delta area. This has been intensively studied by Fisk, McFarlan, Kolb and Wilbert (1954), Scruton (1955), Shepard (1956), Moore and Scruton (1957), and Shepard and Lankford (1959).

The modern subaerial birdfoot delta (Figures 8.4 to 8.6) is about 25 miles wide and about 25 miles long. The delta deposits extend beyond the delta front into the Gulf of Mexico to a depth of more than 100 fathoms and into shallower water to the east and north, where the marginal sediments have been studied. Scruton has shown that the total area of deltaic sedimentation is very large, six imbricating deltas being recognised (Figure 8.7).

Two important factors in the study of this area of deltaic sedimentation are the facies of the modern delta and the effects of the reworking of the older deltas.

The area can be described in three sections; the subaerial delta as described by Fisk et al., the marginal facies as described by Shepard, and the broad features of the petrography.

d. THE DELTA FACIES

Fisk and his co-workers recognised several subfacies in the Mississippi Delta, some of which do not form actual subaerial deposits but which are listed here for the sake of completeness. The environments described were:-

1. The prodelta, (this is described with the marginal facies).

2. The delta front, (this is described with the marginal facies).

3. The interdistributary troughs.
   These are found between the main distributary channels and contain deposits which were laid down in very shallow water and which grade into the marsh or bay deposits.

4. The marshes.
   These are found in the silted up areas of the interdistributary troughs, boardering the levees.

5. The channel fill.
   These are found in abandoned minor distributaries.

6. The bar sands.

   Deposits of the bar sands are built out in the front of the distributary channels. More recent work by Scruton (1955), Shepard (1956) and Shepard and Lankford (1959) has led to the belief that the bar sands do not exist as a distinct sedimentary facies.

7. The levees.
   Natural levee deposits are laid down during seasonal floods along the edges of the distributary channels. They pass down into the bar sands and laterally into the marsh deposits.

8. The mud lumps.

These occur near the distributary mouths and marsh and are extruded masses of mud from underlying deposits. These are extruded due to the weight of accumulating bar finger sediments on the underlying clays.

e. THE MARGINAL FACIES TO THE EAST AND NORTHEAST.

The description of these facies is based primarily on work by Shepard (1956). His investigation included, in part, some of the facies of the subaerial delta described by Fisk et al. Shepard, however, did not recognise the existence of the sand bar fingers and stated that, " sand bar finger" is an incorrect term which may apply to the earlier stages of the delta growth but is not confirmed by an examination of the sediments of the advancing distributaries and interdistributary bays. The "sand bar fingers" at the mouths of the distributaries, on the bar slopes, when examined in detail, indicate that generally less than 10% sand occurs in the sediments below a depth of 50 feet, whilst the upper 50 feet of sediment rarely contains more than 25% sand size material.

The following sedimentary environments were distinguished by Shepard:-

1. <u>Delta front platform</u>.
   The delta front platform lies directly adjacent to the forward building passes and is the area containing the mud lumps. It is an area of shallow water. The passes have submarine levees of deposits different from those of the subaerial delta. The levees rise near the distal ends and are connected by an arc-like shoal or lunate bar.

2. <u>Interdistributary bays</u>.
   These include, in part, the trough areas of Fisk et. al. These are shallow water areas which are virtually free from river and tidal currents. They form natural settling basins for sediments washed over the levees in times of flood and from small distributary passes which branch off the main passes.

3. <u>Prodelta slope</u>.
   This lies beyond the platform and is indefinite off the distributary channels.

4. <u>Open shelf with Recent delta influence or "bottomset beds"</u>.
   This lies at the base of the prodelta slope. It is a flat area of Recent sediment transported from the passes of the Mississippi Delta. It is deep to the south and east but shallow to the northeast.

5. Old Shelf deposits.

These deposits, occurring to the east and southeast, are not derived from the Mississippi River and have little in common with the delta deposits.

6. Reworked Mississippi delta area.

This occurs to the north of the "bottomset beds" in shallow water of less than 10 fathoms.

7. Open Inlet.

This occurs to the northeast and is an open area affected by currents passing around Breton Island.

8. Open lagoonal area.

This includes the area of Breton Sound connected to the Gulf of Mexico by large inlets.

f. PETROGRAPHIC DATA USED IN THE DISTINCTION OF THE DIFFERENT FACIES.

Scruton (1955) and Shepard (1956) gave detailed petrographic analyses of the various sediments and it is therefore possible to give detailed data on the sedimentary structures and petrography. Criteria used are the grain size and degree of sorting, the sand, silt,clay ratios, mica distribution, wood (vegetable matter) distribution and aggregate (aggregates of sand grains cemented by various materials) distribution.

Moore and Scruton (1957) have shown that the minor internal structures of the sediments are important and vary from one environment to another. They distinguished five minor structures in the Mississippi Delta area. These are:-

(i) Regular layers which are tabular or thin lenticular bodies, horizontally bedded or crossbedded and varying from a fraction of a millimetre to about 10 centimetres in thickness.

(ii) Irregular layers in which the internal structure is generally heterogeneous and in which each layer is usually ½ to 3 centimetres in thickness.

(iii) Mottled structures which have sinuous contacts and are irregular masses of one grain size in another. Any alignment present is generally vertical. These can be subdivided into mottles with distinct boundaries and those with indistinct boundaries.

(iv) Homogeneous course-grained deposits.

(v) Homogeneous fine-grained deposits.

These are illustrated in Figures 8.8 and 8.9. The distribution of the various minerals and grain sizes in the Mississippi Delta area is given in Figures 8.10 to 8.19, which can be compared with the distribution of the various facies (Figures 8.20 to 8.23).

Regular layers are characteristic of the delta platform and slope deposits, whilst the irregular layers are more characteristic of the margin of the "bottomset beds". The mottled structures occur in sediments around the margin of the area with irregular layers and also in deposits making up part of the reworked Mississippi Delta, the open lagoon inlet and along the margin of the open lagoon. Homogeneous fine-grained deposits occur in the deep "bottomset beds", whilst the homogeneous coarse-grained deposits occur in the centre of the open lagoon and the reworked delta area.

Moore and Scruton considered the possible modes of formation of these structures and these are summarised in Table 8.A. When dealing with consolidated sediments it is possible for the pre-existing structures to be obliterated or somewhat modified during compaction and lithification. It is therefore important to examine all minor structures critically.

The salient features of the petrography of the Mississippi Delta and the marginal sediments are summarised in the succeeding paragraphs.

g. THE PETROGRAPHY OF THE DELTA DEPOSITS

This summary is taken partly from the work of Fisk and his co-workers and partly from Scruton and Shepard.

1. Marsh deposits. These are structureless, organic-rich clayey silts and silty clays, traversed by roots. They contain ferruginous aggregates and commonly contain mica.

2. River Channel deposits. These consist of thin layers of silt or fine sand interbedded with smooth clayey deposits. Occasional lumps of smooth mud in sand, or lumps of sand in clay, occur in the deposits. River channel deposits often contain calcareous aggregates.

3. Channel fill deposits. These consist of massive sands which are very fine-grained at the base,

overlain by bedded sands which pass up into sandy silts and silty clays with a marsh flora. The massive beds have erosive contacts with other deposits.

4. <u>Levee deposits</u>. These consist of laminated or lensoid silty and clayey material near the perimeter of the delta and may be cross-bedded. They contain ferruginous aggregates. They may be covered by growing vegetation and may contain shrinkage cracks. They pass down into bar sands and grade laterally into marsh deposits.

5. <u>Interdistributary trough deposits</u>. These consist of a veneer of layered sandy silts and clayey silts in the marginal sections. The bay bottoms are covered by massive, poorly sorted silty clays. Shallow water crevasse deposits of the marsh front are distinctive, consisting of crossbedded sandy silts and fine sands interbedded with clayey silts. Spits and subaqueous bars are made up of clean, well sorted, fine sands with marginal zones of silty sands. The fauna is brackish water to marine

except for the subaqueous bars and spits where the fauna is similar but has additional marine organisms. The interdistributary bay deposits are similar to the platform deposits but may be unlaminated with lenses or lumps of sand or silt in otherwise clayey deposits.

6. <u>Bar sand deposits</u>. These are made up of fine sands near the surface with very fine sands and silt within deeper water. At the base of the delta there is a bulge in the prodelta zone clays. The sands are massive and locally crossbedded near the top of the section but at depth, thin layers of silty clay and very fine silty sand, with occasional organic-rich laminae, occur.

7. <u>Delta front platform deposits</u>. These contain abundant sand of variable distribution, being about 90% off some of the river bars but scarce in the interdistributary bays. The sediments are generally laminated but a few sands with clayey mottling occur as do clays with sandy mottling. Occasional sand stringers also occur.

Wood is much more common in the subaerial delta than in either the platform or the interdistributary bays. Mica is more abundant in the bays than in the platform deposits and is fairly abundant in the marsh deposits. Molluscs are rare in all of these "topset" environments except the bays.

h. THE PETROGRAPHY OF THE MARGINAL DEPOSITS.

The summary of the main features of the marginal deposits is taken from work by Scruton (1955) and Shepard (1956).

1. The prodelta slope or "foreset" slope deposits. Here, except off the mouths of the main passes, there is an increase in clay and a decrease in sand when passing from the platform deposits. The percentage of wood increases to about 15% of the coarse fraction (compared to the platform with 4%). Aggregates increase in amount but the fraction of mica remains about the same as that of the platform deposits. The number of shells is greatly increased and echinoids occur but are rare.

The lamination of the deposits which is characteristic of the platform sediments is practically absent. Most of the sediments are unstratified and only a few partings of silty layers occur, more commonly off the mouths of the passes.

2. <u>The open shelf with Recent delta influence or "bottomset beds"</u>. In the shallow water area there is an increase in the sand content when compared to the prodelta slope deposits but a decrease in the area of deeper water. The wood content and the aggregate content of the coarse fraction is greatly decreased, whereas the mica content remains about the same. The marine fauna is much increased.

There is a scarcity of stratification and a general homogeneity of the sediments. In the deeper area, the deposits are clayey with occasional lenses of silty material or occasional lumps in the outer areas. The shallower deposits show mottling of silt and sand in a clayey matrix and vice versa.

3. The old shelf deposits. Here, there is a large increase in the sand content and a large decrease in the silt and clay content when compared to the "bottomset beds". Wood, mica and aggregates occur very sparingly, if at all. There is an increase in the shell content when these deposits are compared with the previously described sediments.

4. The reworked deposits. These contain a high percentage of sand, usually over 80%. Silt is very scarce and clay occurs in negligible quantities. Wood is present though it is scarce. Mica and aggregates are very scarce but shells are common.

5. The island deposits. Except for the flats and marshes inside the beaches and dunes, these consist mainly of dune deposits, of more rounded sand containing only a trace of silt or clay.

6. The open inlet deposits. These are intermediate between the reworked deposits and the slope deposits. Wood and mica are abundant but, compared to the platform deposits, the aggregates are scarce. The sediments are mostly of clayey material with lenses, layers and lumps of sand or coarse silt. On core showed the presence of clay balls in the matrix. Shells are of variable abundance.

7. The open lagoon deposits. These deposits vary very little from the inlets in their sand, silt and clay content but aggregates are very scarce. There is a preponderance of sandy or coarse silt sediments with clayey mottling or vice versa.

i. DEPOSITION IN THE MISSISSIPPI DELTA AREA

Scruton (1955) has shown that six deltas have been built up prior to the formation of the modern delta in Post-glacial times, that is, in about 5,000 years. Each delta, as it enlarged, reduced the drainage gradient to a very long low slope. Break through the levees, in time of flood, eventually resulted in the formation of a new delta.

As soon as deposition has ceased and the centre of deposition moves to another area, the shoreline in the area of the abandoned delta retreats due to a combination of compaction and subsidence of the deltaic sediments and marine erosion. This is possibly accompanied by a gradual and continuous regional subsidence. This destructional phase of the sedimentary history of the area is marked by winnowing of the sediments with accompanying resorting which results in the formation of sand sheets which mark the limit of the seaward advance of each subdelta.

As Scruton has pointed out, when referring to the area of sand northeast of the modern delta. "It is important to remember that this mass of sand is not being deposited by the present delta, but was produced by wave and current erosion and winnowing of sediments deposited as the earlier St. Bernard subdelta".

The sediments were built up to sea level in water which deepened to the south and thus each unit consists of a wedge of sediment which thickens to the south.

j. SUMMARY

In the Gulf of Mexico and the Mississippi Delta area many different sedimentary environments and facies can be recognised. Ancient sediments deposited in similar environments might be expected to show similar degrees of sorting, structures and fossil assemblages. This is true of the marine and marginal deltaic sediments. The subaerial deltaic sediments, however, are unlikely to be completely preserved, if at all, due to the erosive activity of the marine environment. It would be exceptional to find complete evidence of a subaerial delta of the type described by Fisk and co-workers.

CHAPTER IX

THE ENVIRONMENT OF DEPOSITION OF THE YOREDALE STRATA

a. INTRODUCTION.

The deposition of the British Lower Carboniferous strata.

The distribution of the Carboniferous rocks has been reviewed by Trotter and Hollingworth (1928) and George (1958). A summary of the significant features of the palaeogeography of Carboniferous sedimentation in the area of Great Britain, taken from these workers, is given in the succeeding paragraphs.

Two large land masses existed at the beginning of the Carboniferous period. To the north, there was a land mass north of Ireland which included, in part, the Highlands of Scotland. To the south, another land mass existed which included the tip of the Southwest Peninsula. Between these two land masses was an area of shallow to moderately deep seas which, in earlier Lower Carboniferous times, contained more rigid areas or blocks which remained above sea level. These were, however, submerged by the time of the deposition of the Middle Limestone Group of the Carboniferous but still exerted an influence as more rigid blocks in the area of subsidence.

Between the two large land masses, several areas of deposition have been identified as separate units. In the north was situated the Northwest Irish-Midland Valley Trough and south of this was the Irish Midland Plain Shelf and the English Central Province. Between these two areas were the Longford-Down and the Southern Uplands Massifs and the Alston and Askrigg Blocks. Between the Southwestern and Central Provinces was a land mass composed of the Leinster Massif, St. George's Land and the Mercian Highlands.

During the deposition of the Middle Limestone Group of the Lower Carboniferous there existed in northern Britain, from the Highlands of Scotland to the Settle area, an area of shallow sea in which the Yoredale facies was deposited. South of this was a deeper-water area of the marine shale facies whilst to the west, in the Lake District and in Ireland, there was an area of marine limestone facies. Trotter (1951) has discussed the relationships between these areas and has shown that the marine limestone facies can be regarded as shallow water shelf deposits of clear seas. The marine shale facies is dominantly argillaceous with thin limestone bands which contain an abundance of thin-shelled organisms such as pelecypods and goniatites. This facies is regarded as the deeper more seaward equivalent of the two shallow shelf facies.

The Yoredale facies, as has been shown, is composed of a series of alternating limestones, shales and sandstones. These beds may include thin coal seams and underclays. Thus, in this shallow shelf area, there was an alteration of marine limestones with deposits which sometimes contained nonmarine, subaerial underclays and coals.

b. DEPOSITION OF THE YOREDALE STRATA

Thus the setting of the depositional area of the Yoredale strata with its marine to nonmarine deposits is similar to the Gulf of Mexico with its Recent marine deposits and the subaerial deposits of the Mississippi Delta which have been described in the preceding chapter. The writer follows Moore (1959) in supposing that the strata of the Middle Limestone Group were deposited in an environment similar to that of the Mississippi

Delta area.

Nevertheless, there are three very significant differences between the modern Mississippi sediments and those of the Yoredale delta. These can be enumerated thus:-

(i) The depth of the marine environment is greater in the Mississippi area than seems to have been the case with the Yoredales. The limestones of the Yoredale succession contain thick-shelled brachiopods and corals which are often broken or "rolled". Similar modern faunas occur in water which is generally less than 150 feet in depth. The limestones of the Yoredales, which form the bases of the cyclothems, were thus probably deposited in less than 150 feet of water. The material deposited on top of the limestones was, therefore, probably deposited in even shallower water. The depth of water off the Mississippi Delta varies from 0 to 240 feet and, in the south reaches over 600 feet.

It is possible, therefore, that the Yoredale delta or deltas were deposited much more rapidly, due to the shallow nature of the water and were subject to rapid changes in the local areas of deposition.

(ii) The size of the channels in the Mississippi Delta are much larger than any known in the Middle Limestone Group. Large channels are known in the Upper Limestone Group and have been discussed by Dunham (1948). It seems probable that in the Middle Limestone Group the terrigenous sediments were deposited from a series of small, imbricating shoal water deltas.

(iii) The quantity of limestone in the Yoredales is much greater than is the case in the area of deltaic sedimentation in the Gulf of Mexico.

The Yoredale strata, in the area studied, are much thinner than in the Northumberland Trough and areas to the east (proved by boreholes) and south, due to the more rigid nature of the fault block and resultant lesser amount of regional subsidence in this area.

c. THE DEPOSITIONAL ENVIRONMENT OF THE MIDDLE LIMESTONE GROUP IN TEESDALE

Although the nature of the environment of deposition cannot be reconstructed in minute detail, a consideration of the different subenvironments of the Gulf of Mexico and the Mississippi Delta, in comparison with the rocks studied, helps to form a broad picture of the Yoredale type of sedimentation.

The comparison here is with the Middle Limestone Group of Upper Teesdale but the deposits are similar to others above and below and cover wider areas.

The following discussion is divided into sections based on the major facies observed in the Mississippi area by Scruton (1955), Shepard (1956) and Moore and Scruton (1957).

(i) The marine facies.

This facies includes all of the marine deposits which are considered to include the limestones, the Type 2 shales and, in part, some of the Type 3 shales.

These rocks were deposited in shallow seas which were very clear during most of the limestone deposition. The depth of the water can be inferred from the presence of numerous thick shelled brachiopods and rolled corals occurring in the limestones. These marine organisms usually live in shallow water.

The purity of most of the thicker limestones suggests that the seas were very clear and that terrigenous detritus was scarce. This is borne out by the presence of the abundant fauna including, in some bands, corals.

Wave and current action was prominent during the deposition of these beds. This is reflected in the presence of broken brachiopods and crinoid ossicles and the comparative rarity of calcareous ooze in most of the limestones.

Not all the limestones are pure and clean and the different limestone types recognised can be equated with the approach of the delta to the area of marine limestone deposition.

The bases of the thicker limestones and, in places, the whole of the thinner limestones may contain abundant quartz grains, as in the Type 2 or arenaceous limestones. In many cases, underlying sandstones can be seen to grade up through calcareous sandstones and arenaceous limestones into the Type 1 limestones which contain no quartz. In cases where the passage from sandstone to limestone is not seen, local derivation of the sand grains by wave and current action from banks of sand on the sea floor is inferred. The well sorted nature of the quartz grains and the absence of clay detritus and micas also suggests a local origin from a previously sorted sand. Primary land derived detritus would probably contain considerable quantities of clay and mica and would not be so well sorted.

The Type 1 limestones, which are very pure, are the result of limestone deposition in the almost complete absence of terrigenous detritus. Accompanying an increase in the amount of clay material brought into the area of deposition was an increase in the amount of clay interred with the carbonate. This resulted in the Types 3 and 4 dark limestones. Still further increase in the amount of clay material resulted in the deposition of the shaly or Type 5 limestones and the calcareous Type 2 shales. An increase, not only in the amount of terrigenous material but also in the carrying power of the currents bringing in the detritus, was responsible for the formation of the Type 6 limestones which are both shaly and arenaceous. Unlike the Type 2 arenaceous limestones, these contain poorly sorted quartz grains and also contain mica.

The occurrence of dolomite in some of the Type 1 limestones, often associated with chalcedonic silica in fossils is similar to occurrences mentioned by Wells (1955) in chert bands found in Carboniferous limestones and these occurrences can be considered as poorly developed chert bands. Both the chalcedony and the dolomite are formed as a result of chemical deposition in shallow marine water, the source of the silica, and possibly some magnesia, perhaps lying in the rivers which brought in the terrigenous detritus. River waters not uncommonly carry high concentrations of silica, (Clarke, 1924). It appears that the silica and magnesia were possibly carried into the sea water and deposited with the limestone due to chemical and physical conditions in the environment, which as yet are not understood.

Recently, Walker (1960) has shown that there may be another source for authigenic silica occurring in rocks, whether it be in the form of overgrowths on detrital grains or in the form of chert. Calcium carbonate may replace quartz or other crystalline silicates such as glauconite. At higher pH values silica is more soluble than calcium carbonate and it is quite possible that the pH of the marine environment may vary quite considerably due to local bacterial action as different bacteria live by different processes. Thus hydrogen sulphide is commonly produced, precipitating sulphides which form pyrite or marcasite but carbon dioxide may also be produced to yield a low pH,or ammonia to yield a high pH. Some of the Teesdale sandstones show replacement of sand grains by pyrite and calcite and it is possible that the quartz dissolved from these rocks has been deposited elsewhere as chert bands. The association of dolomite with chert suggests that the chert was deposited when the pH of the sea water in the area of limestone deposition reached a level which was higher than usual.

A third possible source for the silica is in the ganisters occurring under the coals which show evidence of considerable solution and some redeposition of silica. This has been emphasised by Goldschmidt (1954 Ed.).

Encroachment of the delta onto the area of marine deposition is marked by the incoming of large amounts of clay material which give rise to the Type 2 and Type 3 shales.

The Type 2 shales which lie in and immediately above the limestones may contain very fine mica and silt plus very fine sand size quartz. They are characterised by an absence of mixed-layer clay minerals although it appears, from a consideration of the other shale types, that mixed-layer clay minerals were being brought into the area of deposition. The Type 2 shales were deposited in a marine environment as evidenced by their association with marine limestones and their frequent calcite and fossil content. In this area, the deposition of terrigenous mud was sufficiently slow to enable an abundant marine fauna to exist. The absence of the mixed-layer clay minerals in the Type 2 shales is believed to be due to the slow rate of deposition of the land derived detritus which exposed the clay minerals for long periods to sea water. Under these conditions, the absorption of potassium and magnesium ions would lead to the regeneration of illites and chlorites from the mixed-layer clays.

The Type 2 shales above the limestones are overlain by the Type 3 shales, which represent further approach of the delta or deltas. Deposition appears to have been too rapid for the survival of the marine organisms in most cases and the tops of these shales are unfossiliferous. Pyrite nodules are common, their origin being closely related with the rapid burial of the marine organisms (see Chapter XI). Further indication of the approach of the delta is seen in the increased coarseness of the micas and in the increased amounts of sand in the shales when compared to the Type 2 shales. The marine shales grade upwards into deposits which are equivalent to the marginal facies of the delta.

(ii) The delta marginal facies.

According to Moore (1959), the shales lying above the limestones are equivalent to the prodelta of Fisk et al.(1954), the silty shales are equivalent to the delta front and the flaggy and shaly sandstones are equivalent to part of the interdistributary trough deposits. The writer believes that the shallow nature of the water in which the Yoredale rocks were deposited should be taken into account when studying the beds in relation to the type of sedimentation in the Mississippi Delta. The normal sequence in any area of deltaic sedimentation is one of offshore "bottomset" deposits, followed by nearer shore "foreset" deposits or delta slope deposits, followed finally by the "topset" or delta platform deposits. The "bottomset" deposits are apparently absent from Moore's delta and the "foreset" or delta slope deposits are very poorly developed.

Here, some of the marginal deposits which are of small lateral extent will be considered with the delta deposits. In this section, it is proposed to consider only the reworked deposits, the "bottomset" deposits and the prodelta slope deposits.

(iii) The "bottomset" beds of the marginal facies.

In the area of the Mississippi Delta, these consist of clays and silty clays with sandy lenses and mottles and with a marine fauna.

In the Yoredale beds, a similar type of sedimentary environment gave rise to the Types 2,3 and 4 shales and the Type 2 sandstones. These generally occur in this order in the cyclothems. The sequence can be explained as in the next few paragraphs.

The delta, as it gradually grew out over the area of marine deposition, brought in fine mud which slowly overwhelmed the marine organisms. As the delta grew still further forward, clay was accompanied by silt and the Type 3 shales were thus deposited on the Type 2 shales. Further approach of the delta was marked by the incoming of increasing quantities of coarser detritus and the Type 3 shales were succeeded by the Type 4 shales which are highly arenaceous, and the Type 2 sandstones, the shaly and flaggy sandstones.

During this transgression, with its increasing amounts of detritus, the marine organisms were overwhelmed so that only a few tracks and pit and mound structures are found in the flaggy sandstones. The decay of the buried organisms is believed to be responsible for the presence of sulphide nodules in the shales and pyrite in the sandstones.

The clay mineralogy of the shales also changes. There is a decrease in number of samples containing illite when passing from the Type 2 to the Type 4 shales. This accompanies an increase in the coarseness and quartz content with decrease in fossil content. As shown above in the section on the marine beds, these effects are the result of increased rates of sedimentation. With increased rates of sedimentation, the interaction between ions in the sea water and the clay minerals would be restricted and not all mixed-layer clay minerals present in the clays would be able to react to form illites and chlorites. This would result in the occasional presence of original mixed-layer clay minerals in the Type 4 shales.

On burial, only the entrapped connate water would be able to react with the clay minerals. Assuming that the composition of sea water has changed very little since Carboniferous times, the ratio of potassium to magnesium would be about 1:5. This excess of magnesium would account for the increase in the number of samples containing chlorite when passing from the Type 2 to the Type 4 more rapidly buried shales. Preferential adsorption of potassium by the Type 2 shales during their longer period of exposure to seawater, in addition to the effect of burial, is believed to be the reason for the decrease in the number of samples containing illite when passing from the Type 2 to the Type 4 shales.

(iv) The prodelta slope deposits of the marginal facies.

As has been indicated earlier in this chapter, the writer believes that the Yoredale delta was a shoal water delta which was formed in a very shallow sea. The amount of material deposited in the area of the prodelta slope would be small as the prodelta slope itself would be small in this shallow water setting. In addition, many of the deposits laid down would be susceptible to wave and current action with accompanying resorting.

Some of the Type 2 sandstones, the shaly and flaggy sandstones and the less clean, rippled and stratified Type 1 sandstones were probably deposited in this prodelta slope environment.

According to Moore, these beds were deposited in the delta front and interdistributary trough environments. The delta front deposits are very similar to the prodelta slope deposits and they grade into each other. Some of these deposits may be regarded as delta front deposits and correspond to Moore's delta front deposits.

(v) The reworked deposits of the marginal facies.

These represent areas where deposition has ceased and where wave and current action have been effective and where winnowing and sorting of the sediments has occurred. The resultant deposits are generally well sorted and are represented by very pure sandstones with little or no mica. Most of the main sandstones in the succession belong to this type of depositional environment. These are the Type 1 sandstones which according to Moore were deposited in the interdistributary troughs. It is here considered that these rocks are too pure to have been deposited directly and their general lack of bedding and purity are a consequence of a considerable period of resorting.

These beds are lensoid in character and the lenses are interleaved. It is thought that as deposition of one sand body ceased, it was subject to resorting processes whilst sand was being deposited in adjacent areas. The clay mineralogy of occasional shale bands occurring in these sandstones suggests that the resorting was effected in a marine environment. These shales, the Type 6a, argillaceous rocks, are similar to the Type 2 shales in that they consist of illite or illite plus chlorite but with no mixed-layer clays. They do, however, contain very coarse mica and are probably the result of local settling which occurred during the periods of resorting.

In some places the sandstones are calcareous; these are some of the Type 4 sandstones. These appear to have had interstitial calcite precipitated from the environment of deposition.

(vi) The subaerial delta and the minor marginal deposits.

According to Moore (1959), the delta deposits are represented by the flaggy sandstones, the massive sandstones, the seat earths, coals, minor shales above the coals and the minor rhythms.

As has been pointed out, the writer believes that some of these rock types were deposited in environments different from those envisaged by Moore. Although the area of deposition of the Yoredales was large, the terrestrial detritus was deposited in a series of small imbricating deltas and sorting action was particularly effective in destroying the original features of the beds laid down in shallower water. In consequence, it would rarely be possible to recognise certain of the deposits discovered by Fisk et al. (1954), in the subaerial delta of the Mississippi. However, some of the rock types recognised in the area studied can be equated with the deposits in the subaerial delta and also with the marginal deposits.

The Type 5, poorly sorted arenaceous shales and the Type 3 sandstones which are poorly sorted and 'dirty' or argillaceous appear to have been deposited in fairly quiet water with little resorting but do not have the features of the deeper water sediments such as fossils or evidence of organisms in structures. Pyrite too is rare.

These sediments are thought to have been deposited in natural shallow water settling basins probably during the times of flood. Two environments of deposition within the framework of the delta or the marginal deposits could form settling basins. These are the interdistributary troughs of the imbricating deltas or lagoonal areas between the deltas. Small local distributaries are believed to have brought in sandy material which was deposited as lenses in the shales.

The fact that many of the Type 5 shales contain mixed-layer clays suggests that these deposits were not in contact with sea water for significantly long periods of time. The Type 3 sandstones show lateral and vertical passage into underclays thus indicating the shallow water nature of these deposits.

Rarely, marine inundations of these areas were responsible for the deposition of thin limestone bands and calcareous sandstones or calcareous shales containing a few fossils or pyrite nodules.

There is no evidence in the Teesdale area of the occurrence of large channels in the Middle Limestone Group and a review of the literature has shown that they have not been recognised in other areas at this level; but two types of sandstone infilling small channels have been recognised in the beds studied.

In Ettersgill, a small channel cuts the shales immediately above the Single Post Limestone. This sandstone, a Type 1 to Type 3 sandstone, (it is rather variable in character), is a local channel sandstone which contains nothing extraneous other than a few small clay galls. This is probably the equivalent of a small crevasse channel deposit.

The bases of the two beds of the High Brig Hazle in Newbiggin Beck appear to be parts of true river channel deposits which were probably formed in times of flood so that they cut below the base level of the normal channel into the shales beneath. Another factor which suggests that these were formed in times of flood is the presence of coarse sand and pebble size quartz grains together with coal fragments and shale and clay galls. These are the Type 6 sandstones.

The underclays, ganisters and seat earths, and the coals associated with these minor deposits are the result of colonisation of the subaerial delta and marginal deposits by land plants. The underclays and seat earths have formed in the interdistributary areas whilst the ganisters have formed on the purer resorted sands which probably formed banks near or above sea level. The underclays and coals are poorly developed and thin and were either formed slowly or were in existence for short periods of time only.

In places, shaly coals occur in the succession, for example, the coal in the shale below the Three Yard Limestone in Newbiggin Beck. These coals have no underclays or seat earths and were probably deposited as a vegetable detritus which was washed into the interdistributary troughs.

The underclays are now generally recognised as "fossil" soils. Droste (1956) and Harrison and Murray (1958) have shown that weathering with leaching leads to the formation of mixed-layer clay minerals, whilst Keller (1958) has shown that very vigorous leaching leads to the formation of kandites. Only mixed-layer clays are found with illites in the underclays which were examined but these are thin and poorly developed and the leaching processes were probably only moderate. The lack of chlorites in the underclays appears to be due to the effects of Carboniferous leaching as chlorites are found in the other argillaceous rock types. A similar effect was observed in the weathering of glacial tills by Droste (1956) where chlorite was degraded to mixed-layer clay minerals before illite during the leaching processes.

The absence of primary kandites from the rocks studied and also the absence of montmorillonite appears to be the result of the absence of these minerals from the detritus being brought into the area of deposition.

d. THE POSITION OF THE TYPE 5 SANDSTONES IN RELATION TO THE OTHER ROCK TYPES

The Type 5 sandstones are similar in mode of occurrence to the Type 1 sandstones but are coarser, less well sorted and contain appreciable amounts of secondary kaolinite which is derived from the weathering of feldspars. As deposited, these were coarser, feldspathic "sheet" sandstones. Although they are not so well sorted they were nevertheless clean sandstones and contain very little mica and detrital clay matter.

These sandstones only occur in horizon of the Quarry Hazle and possibly the Tuft sandstone in the western part of the area studied. They appear to have a similar depositional history to the Type 1 sandstones but the detritus is believed to have been derived from a different source area to the west.

e. VARIATIONS IN THE FLUORINE CONTENT OF THE SHALES IN RELATION TO THE ENVIRONMENT

The variations in the fluorine content of the argillaceous rocks can be possibly explained by exchange of hydroxyl ions of the clay minerals for fluorine in the sea water. Thus from Type 2 to Type 4 there is a decrease in the calcite and marine fossil content accompanied by increase in the coarse mica and quartz content. The Type 2 shales are believed to have been subject to long periods of contact with the sea water and this has enabled greater quantities of fluorine to be exchanged. This accords well with the fluorine analyses where the Types 2 and 6a argillaceous rocks contain higher percentages of flucrine than the Type 4 argillaceous rocks. The Type 3 shales which lie between the Type 2 and Type 4 contain intermediate amounts of fluorine.

The Type 5 argillaceous rocks do not appear to follow these trends. However, these rocks are subject to considerable variation and considering their position in the cyclothems and their environments of deposition they may have been subject to varying degrees of exposure to marine waters. Thus of the two analysed, one contained pyrite nodules and the other contained calcite.

The Type 1 argillaceous rocks, the underclays, occurring under the limestones may have been subject to considerable periods of exposure to sea water during the initiation of limestone deposition and this is reflected in their high fluorine contents.

The variations described and their probable causes are not conclusive. There is insufficient difference between the fluorine contents of the different types compared to the variation in different samples of the same type. The analyses cannot, therefore, be regarded as a means to determining whether or not a shale was deposited in a brackish water or marine environment.

f. THE SIGNIFICANCE OF LOCAL LATERAL VARIATIONS IN THE CYCLOTHEMS

The existence of many local lateral variations of the deposits studied has been described by a series of examples in the first chapter of this work.

The major limestones, although variable in the vertical succession, are fairly constant in thickness and character. The minor limestones are much more variable and the Iron Post Limestone is absent in many sections. The occurrence of limestone beds above the Three Yard Limestone and the existence of thin bands of limestone above the Tyne Bottom Limestone in the Alternating Beds in the east of the area indicates that there was less terrigenous detritus being deposited in the southeast.

The local variations in the noncalcareous beds are here considered as important characters of the deltaic facies.

As Scruton (1955) has shown, the various facies of the subaerial delta and the marginal areas of the Mississippi have a large lateral extent. The main sandstones in the area studied are very variable laterally as can be seen from the examples given in Chapter I and from the stratigraphical sections (Figures I.1,2 and 3). These sandstones are seen to be interleaved beds and are not continuous sheet sandstones. The underlying and overlying shales and sandstones associated with them are also very variable. In the higher cyclothems, the main sandstones are absent in some sections.

Whilst the general structure of the cyclothem; limestone, shale, sandstone, seat earth or underclay and coal, indicates that the form of deposition was repeated from cyclothem to cyclothem, that is, marine deposition followed by deltaic deposition, the lateral variations give a more definite picture of the type of deltaic deposition. This has been illustrated in the preceding pages. The numerous and rapid lateral variations are indicative of the existence of numerous shoal water deltas during the deposition of these beds.

In broad outlines, the sedimentary environment of the Middle Limestone Group can be equated with a shallow sea area in which a series of small, imbricating, shoal water deltas periodically grew and filled the area of deposition. At times, parts of the area were built up to sea level or near sea level and were colonised by land plants. At other times, shifting of the shoal water deltas allowed local resorting of the sediments and occasionally local deposition of marine limestones and shales. Very occasionally brackish water lagoons were formed with the deposition of ironstones in the upper part of the succession.

g. THE SOURCE OF THE SEDIMENTS

In the small area investigated, reliable data on the source of the sediments was not forthcoming. Current bedding and ripple marks are only the result of local current and wave movements which may bear no relation to the regional movement of sediment. However, the directions observed in the Teesdale area are in fairly close agreement with those observed in other areas.

False bedding directions in the sandstones indicate that the prominent current directions were roughly south, varying between southeast and southwest.

The coarser feldspathic sandstones are only found in the western part of the area. These are believed to have been derived from the northwest whereas the majority of the sediment is believed to have come from the north and possibly the northeast.

h. CYCLOTHEMIC SEDIMENTATION: THE MECHANISM

It is not proposed, here, to discuss the theories of cyclothemic sedimentation. Many theories have been advanced to account for the Carboniferous cyclic sedimentation viz. Hudson (1924), Weller (1930,1956, 1958), Wanless and Shepard (1936), Robertson (1948, 1952), Weller, Wheeler and Murray (1958), Van der Heide (1950), Moore (1959) and Wells (1960). All of these theories accentuate one aspect or another of cyclothemic sedimentation and to fit the Yoredale sedimentation to any one theory would require much more information than is available in the small area studied.

However, two features of the Yoredale sediments stand out and must fit into any proposed theory of cyclothemic sedimentation applied to these beds. These are the extent of the major limestones and apparent sudden cessation of terrestrial detritus during limestone deposition.

The sudden cessation of the detritus can be caused by the lack of rivers bringing material to the area which could be due to major diversions of the river or rivers (as suggested by Moore (1958)) or due to sudden recession of the coast line. The former cause is unlikely to effect such a complete cessation of detritus and is unlikely to occur with such frequency as would be required to account for the Yoredale cyclothems. The latter cause could be effected by a sudden rise in sea level or a sudden sinking of the area. If either of these two causes is active there should be some reflection of the changes outside the area of Yoredale deposition.

A shallow shelf sea area occurs to the west of the region of Yoredale deposition. Here, in the Workington district, the sediments were deposited beyond the sphere of delta influence and there is little terrigenous clastic material in the succession which is composed chiefly of limestones. Several limestones occur, most of which can be equated with the limestones in the Lower and Middle Limestone Group of the Pennines.

Erosion surfaces are common in the Cumberland succession and it appears that the limestones were deposited in very shallow water which, at intervals, became even shallower or receded completely. The shallow nature of the water which was responsible for the erosional features at the tops of some of the limestones was, at intervals, suddenly succeeded by deeper water in which the succeeding limestone was deposited.

It appears, from an examination of the Yoredale and west Cumberland deposits, that the shelf sea area in which the deposits were laid down was subject to sudden submergence, marked by the deposition of a major limestone.

The minor limestones, such as the Cockle Shell and the Iron Post Limestone do not have such continuous lateral development and even over a small area such as Teesdale may be laterally impersistent. However, these limestones are obviously not of the same origin as the minor limestone bands in the minor rhythms. The Iron

Post Limestone appears to be a poorly developed major limestone and is succeeded by beds which indicate that the limestone may be regarded as the base of a separate cyclothem. Dunham (1948) regarded this limestone as forming the base of the Iron Post cyclothem.

The Cockle Shell and Single Post Limestones form separate limestones in the Teesdale area but in the Wensleydale area they are known to join and in the south form the basal posts of the Middle Limestone (Moore, 1958), which is equivalent to the Scar Limestone. The noncalcareous beds between the Single Post and the Scar Limestones may be regarded as encroachments of shoal water deltas in the area of limestone deposition. The three limestones occur in one cyclothem in the south but in the north the minor limestones form the bases of minor cyclothems below the major Scar cyclothem.

Moore (1958) argued that the uplift which occurred along the Mid-Craven Fault had no effect on the Wensleydale area and that there was no evidence of diastrophic control of the cyclothemic sedimentation.

There appears, however, to be no explanation of the erosion surfaces in the Cumberland succession, other than diastrophic control.

The following sequence of events is suggested for the deposition of the Yoredale cyclothems:

1. Sudden inundation of the area of sedimentation, either by rise in the sea level or by submergence of the shelf, followed by limestone deposition.
2. Growth of the deltas over the area of limestone deposition and gradual overwhelming of the marine organisms.
3. Building up of the area of deltaic sedimentation to sea level or near sea level, and resorting of some of the upper sediments.
4. Weathering of some of the upper surfaces of the sediment, subsequent colonisation by land plants and formation of swamps or marshes over parts of the area.

This sequence of events has been repeated many times, the sediments deposited representing several complete cycles of deposition.

CHAPTER X

IGNEOUS ROCKS AND METAMORPHISM BY THE WHIN SILL

a. INTRODUCTION

Many of the shales studied in Teesdale have suffered some degree of metamorphism during the intrusion of the Whin Sill and, in a very few cases, possibly during the intrusion of the Cleveland Dyke. These two igneous rocks and their effects on the country rocks are considered in this chapter and the next.

b. THE CLEVELAND DYKE

This dyke, of Tertiary age, is exposed in Ettersgill, about half a mile northwest of Dirt Pit Farm, where it is about thirty feet wide and cuts the Whin Sill. Here, the margins of the dyke are amygdaloidal. The dyke is again seen to the east in Smithy Sike, where it is associated with a fault cutting sandstone. In both of these exposures, it trends only a few degrees south of east.

The dyke is also exposed at Mirk Holm, Bow Lee Beck, where it is bifid and occurs along the line of a northwest-southeast trending fault which downthrows to the southwest about 20 feet. The two arms of the dyke are separated by a raft of sediments 5 to 10 feet wide. The northern member of the dyke is about 6 feet thick, whilst the southern member is only about 3 feet thick.

The dyke is also recorded as occurring in the Red Grooves Level and in Coldberry Gutter.

East of Middleton, the dyke crops out in a small sike below the farm of West Stotley and on the north bank of the Tees at and near Intake Sike.

c. PETROGRAPHY OF THE CLEVELAND DYKE

The Cleveland Dyke was described by Teall in 1884 and again by Holmes and Harwood (1929) who classed it as a porphyritic tholeiite.

In Ettersgill the dyke is badly altered and contains large amounts of calcite and chlorite. The rock also contains amygdales of penninite chlorite (R.I. alpha 1.578, gamma 1.580). The larger amygdales of 0.75 mm and over contain centres of quartz.

The dyke is unaltered near Intake Sike where the rock is very fresh and contains plagioclase and pyroxene phenocrysts in a fine-grained groundmass. The plagioclase consists of zoned crystals about 2.0 by 0.8 mm with central labradorite and outer material approaching oligoclase in composition.

The pyroxene phenocrysts are of two types. One is a pale green to pink-brown optically positive clinopyroxene with a 2V of almost 0°, which has rims of colourless pyroxene. A similarly coloured orthopyroxene also occurs in the rock but in lesser amounts.

The groundmass consists of small plagioclase laths and pyroxene with euhedral to anhedral grains of iron oxide minerals. In some of the feldspars cracks occur which have been infilled with zeolite minerals. The matrix of the groundmass is a glassy mesostasis which has partially devitrified.

Rounded, corroded areas of very fine-grained dyke rock occur in the groundmass.

d. THE WHIN SILL

This is a quartz dolerite intrusion believed to be of Upper Carboniferous, pre-Permian age. Only where the sill cuts the strata dealt with in the work on the sedimentary rocks has it been mapped in detail.

East of the Burtreeford Disturbance area, the Whin Sill is intruded into strata between the Tyne Bottom and the Cockle Shell Limestones. Exposures of the Whin Sill are plentiful between Langdon Beck and Middleton, due to the thickness (220 feet was recorded in the Ettersgill Borehole, Dunham, 1948), and its resistance to weathering. Faulting and Pleistocene and Recent erosion have both aided in the production of good exposures.

The sill decreases in thickness from west to east and southeast, being only 14 feet thick in the Lunedale Quarries where its metamorphic effects are negligible.

In places, the sill contains rafts of Yoredale sediments as at High Force, where the following succession is seen:

Whin Sill
1' quartzite to 2' porcellanous shale.
6' Whin Sill.
10' shale.
Tyne Bottom Limestone.

Downstream of Wynch Bridge, a gorge occurs in the Whin Sill along the line of an eroded raft of sediments, of which part is seen at Staple Crag, where at least 8 feet of porcellanous shale with thin sandstone bands occurs. The sediments dip about 30° to the northeast.

In several places at the roof, the Whin Sill gives off leaves of fine to very fine-grained quartz dolerite, which are up to a foot in thickness. Examples of these minor leaves are seen in Ettersgill, near Outberry Bat, and in several stream sections above Holwick.

Above Holwick a transgression of the Whin Sill also occurs. This is marked by the disappearance of the Single Post Limestone to the southeast.

In Rowton Beck, the following succession is seen:

2'2" to 4'	Saccharoidal Single Post Limestone.
c.8'	Porcellanous shale with thin sandstone bands.
1'	Pyritous Whin.
c.3'	Porcellanous shale.
	Whin Sill.

To the southeast in Willy Brig Sike, the thin leaf of Whin is thicker and overlies the Single Post Limestone. Here, the succession is:

Quartzite
7' Fine-grained Whin.
Saccharoidal Limestone
Small gap.
Whin Sill.

South and east of this last exposure, no Single Post Limestone is seen above the Whin Sill and at an exposure about half a mile west of the Middleton Quarries, the Cockle Shell Limestone tends to be saccharoidal, and is underlain by porcellanites indicating the proximity of the Whin Sill.

In the Lunedale Quarries the following succession is seen:

c.8' Dark shaly limestone.
4'4" Pale mica shale.
12-14' Whin Sill
Shales with limestone bands.

Here, the upper 8 feet of limestone is believed, on other evidence (see Chapter I), to be the Cockle Shell Limestone and the transgression is, therefore, complete and the Single Post Limestone here lies below the Whin Sill.

e. PETROGRAPHY OF THE WHIN SILL

The salient features of the petrography have been described by Dunham (1948). Four main types of unaltered quartz dolerite have been recognised. These are the tachylitic marginal facies, the fine-grained dolerite of the thinner parts of the sill and the outer parts of the thicker areas of the sill, medium-grained dolerite of the thicker sill and the pegmatitic varieties, such as are found at High Force Quarry.

The marginal facies and the thin leaves of the Whin Sill contain small phenocrysts of plagioclase and aggregates of pyroxene set in a very fine-grained groundmass of plagioclase, pyroxene and magnetite. There are also large amounts of skeletal ilmenite in the leaves of Whin.

The aggregates of pyroxene have a poikilitic texture containing small crystals of feldspar. These are pale grey-brown clinopyroxenes and are not sensibly pleochroic. The pyroxenes of the groundmass are pale green to brown, slightly pleochroic, clinopyroxenes.

A little brown hornblende ? has developed at the expense of some of the pyroxenes in the groundmass.

The larger crystals of plagioclase, occurring as phenocrysts, are distinctly zoned and show evidence of reaction with inner zones of labradorite and alternating outer zones of more calcic and sodic labradorite. The crystals of plagioclase occurring in the groundmass are of labradorite of a rather more calcic variety than the phenocrysts.

The fine and medium-grained varieties of the Whin Sill are more typical dolerites which consist of three different pyroxenes with plagioclase, opaque minerals, quartz, orthoclase and interstitial micropegmatite. The plagioclase corresponds approximately to labradorite in composition and shows some zoning. The pyroxenes present include grey to pale brown pleochroic clinopyroxene, an almost colourless clinopyroxene and, in some samples, a pale green pleochroic orthopyroxene. The opaque minerals are mostly skeletal and massive ilmenite with a little pyrite in some samples.

Alteration products include calcite, chlorite, green and brown hornblendes, mica minerals (including fine sericite and a little biotite) and some limonite.

The late stage products of the sill, the pegmatites and the zeolite minerals, are not described here as they have been previously described in detail by Smythe (1924) and Dunham (1948).

f. THE METAMORPHIC EFFECTS OF THE WHIN SILL

The metamorphic effects of the Whin Sill in this area have been previously studied by Hutchings (1895 and 1898), Wager (1928) and Dunham (1948). The effects of the metamorphism can be studied in relation to the three main rock types; the limestones, sandstones and shales.

(i) The Limestones.

Fairly pure limestones, such as the Single Post Limestone, are marmorised without the formation of new minerals, the original rock being practically monomineralic. Examples of the saccharoidal Single Post Limestone are seen in Ettersgill and along the banks of the Tees.

The more shaly limestones, such as the Tyne Bottom Limestone at the High Force, show development of calcsilicate minerals such as the garnet, grossularite. Similar effects were recorded by Dunham (1948). These effects are seen in the beds studied as far as the base of the Cockle Shell Limestone on Holwick Fell. Here, a metamorphosed highly fossiliferous band (probably originally a shaly limestone) occurs below the Scar Limestone. Examination in thin section shows a fine-grained mass of material which was optically unidentifiable. X-ray analysis (Table 10.1, and Figure 10.1) reveals this to be composed of chlorite and calcic plagioclase.

(ii) The Sandstones.

These show no apparent effect in the beds studied except for the sandstones very close to the Whin Sill which are indurated and show considerable resistance to weathering and may contain fresh feldspar. Fresh feldspar is rare in the Teesdale sandstones due to weathering and the action of groundwaters.

(iii) The Shales.

The sulphide nodules occurring in the shales are considered separately, together with the sulphides occurring in the unmetamorphosed rocks in the following chapter. Only the effects of metamorphism on the clay and other detrital minerals are considered here.

As has been shown, there are several types of shale recognisable in this area. All of these types show spotting effects (Figures 10.2 and 3) on metamorphism of not too intense a nature. This effect is found in the Bow Lee area to extend to shales occurring above the Scar Limestone, which itself shows no metamorphic effects. The distance of these shales from the Whin contact is about 100 feet. The shales are, therefore, more sensitive to metamorphism than the limestones or sandstones. Similar distances are recorded for the metamorphic effects of the shale sulphides.

As has been pointed out in connection with the limestones, the calcareous shales may develop calcsilicates such as grossularite, chlorite, plagioclase and possibly diopside during metamorphism. Dunham (1948) described a shaly limestone at Falcon Clints which contained garnet, vesuvianite, diopside and chlorite as metamorphic minerals.

Noncalcareous shales near the Whin contact, such as those seen in the Tees exposures, are transformed to porcellanites with recrystallisation of the clay minerals to give sericitic material, discrete areas of pale green to colourless chlorite of extremely low birefringence, small prismatic crystals of rutile and possibly some sphene.

The shales occurring between the outer range of the metamorphism and the porcellanites are spotted. The spots, which vary in size from about 0.08 to 0.4 mm, show considerable concentration of opaque or semi-opaque brown matter, and under crossed nicols are very weakly anisotropic to isotropic. X-ray analyses of the spots from one of the shales (Table 10.2 and Figure 10.4), shows that these are composed of iron-rich chlorite. Dunham (1948) has shown that clear spots, occurring in some of the more highly altered shales, contain andalusite, although none was recognised in the shales examined in this work.

The spotted shales occurring above the Tyne Bottom Limestone at Dine Holm, also show an interesting development of secondary nodules. The nodules have grown in situ during metamorphism distorting the bedding of the shale and have pushed out the spots which have developed at the same time. The nodules are up to 1.5 cm in diameter and are spherical, as distinct from the normal sedimentary nodules which are ellipsoidal and much larger in size.

In thin section (Figure 10.5), several zones can be recognised in the nodules. There is an outer zone of normal shale material with a concentration of opaque matter formed by the coalescing of the opaque spots. The next zone is about 0.5 mm thick and is composed of clay mineral and ?devitrified glass or metakaolin?

The next zone is 3.5 mm wide and consists of irregular aggregates of pyrite up to 0.35 mm in diameter in a groundmass of ?devitrified glass. The innermost zone of the nodule consists of ?devitrified glass with a few very small areas of secondary quartz and colourless chlorite. X-ray analysis (Table 10.3 and Figure 10.6) has revealed that only mica and a 7Å chlorite occur in the ?devitrified glass.

Similar metamorphic effects were recorded as occurring in shales by Hutchings in 1898, who described dark chloritic nodules from a shale at Falcon Clints. Wager (1928) has since shown that these were metamorphosed sedimentary structures. It is believed, however, that the nodules found at Dine Holm show sufficient evidence to be identified as truly metamorphic in origin.

g. SUMMARY

As has been seen, the effect of the Tertiary Cleveland Dyke on the country rocks was negligible.

The Carboniferous intrusion of the Whin Sill had a marked effect on the rocks and in places the effects extended to 120 feet from the junction of the sill.

Near to the Whin Sill, all rock types have been affected. The sandstones have been indurated, the limestones have been marmorised and the shales have been converted to porcellanites. Further away from the contact, towards the outer limit of detectable changes in the country rocks, no effects are seen in the limestones and sandstones, except for the presence of marcasite replacing pyrite. The shales, however, exhibit spotting effects to a distance of 100 to 120 feet from the contact. In addition, as will be seen in the following chapter, a variety of sulphides occurs in the shale nodules including marcasite and pyrrhotite with original pyrite and also rarer chalcopyrite and sphalerite.

Obviously the effects of metamorphism were much greater in the shales than in the limestones or sandstones. However, limestones are usually more susceptible to changes wrought by increased temperature than other rocks, recrystallisation of calcite usually being accomplished with ease.

It is possible that water played a great part in the production of changes in the rocks. The more porous sandstones and limestones probably lost their water early in the history of the intrusion, whereas the more impermeable shales retained theirs and most of the changes were effected using aqueous solutions as diffusional media.

CHAPTER XI

THE SULPHIDES OCCURRING IN THE TEESDALE ROCKS

a. INTRODUCTION

Sulphides occur in several different rock types; they are found disseminated or massive in sandstones, limestones and shales and also in shale nodules. Much of the sulphide is of sedimentary origin.

Hydrothermal sulphides occur in rocks which are faulted and veined, whilst the metamorphio sulphides occur in rocks which have been metamorphosed by the Whin Sill. The nodules which occur in the shales in proximity to the Whin Sill contain a variety of sulphide minerals, including pyrite, pyrrhotite, marcasite, chalcopyrite and others.

Over 50 polished sections have been examined episcopically. The specimens are described in groups arranged on mode of occurrence, i.e. sandstones, limestones and shales. The metamorphic sulphides occurring in shale nodules are considered separately.

The origin of the sulphides is considered and the occurrences of sulphides other than pyrite are related to thermal metamorphism of the rocks.

b. SULPHIDES IN THE SANDSTONES

The occurrence of sulphides in sandstones, except for the hydrothermal sulphides, appears to be restricted to the bases of the sandstones and occasionally the tops, if they are flaggy. The sulphide is pyrite. Pyrite is generally associated with the calcareous facies of the sandstones and often occurs in association with "organic" structures such as pit and mound structures. It invariably occurs as cubes or pyritohedra which replace the quartz and other nonsulphide matter.

A typical example of the mode of occurrence of these sulphides is seen in the base of the Low Brig Hazle in Smithy Sike (892293). This sandstone (T.122), is flaggy and micaceous and contains pit and mound structures. A few pyrite cubes are scattered throughout the rock but are particularly concentrated in the pit and mound structures. The crystals are generally subhedral and are often limonitised. The abundance of the pyrite in the pit and mound structures suggests that the pyrite is, in part, original and was associated with the organic agencies which formed the structures.

A similar occurrence at the top of the Low Brig Hazle on the Tees, near Breckholm Farm (940255) shows the development of pyrite in a flaggy sandstone. The sandstone (T.335) is fine-grained and contains up to 20% pyrite. The pyrite is in the form of cubes, up to 0.25 mm in diameter, replacing both the argillaceous matrix and the quartz grains. Cubes of pyrite may be seen to contain relict quartz grains.

Where the sandstones have been metamorphosed by the Whin Sill, the pyrite may show alteration to marcasite as in the fine-grained sandstone (T.113), which occurs above the Whin Sill in Dufton Sike (871295).

An example of the occurrence of hydrothermal pyrite is seen in a much faulted sandstone, believed to be associated with the Single Post Limestone, in Langdon Beck, immediately below the road bridge,(853311). Here, a pure white quartzose sandstone is cut by numerous small chalybite veins associated with the faults. In the groundmass of the rock there is a development of anhedral, often polycrystalline pyrite, up to 2 mm in diameter, replacing the quartz and forming up to 1 to 2% of the rock (T.172). In many places, masses of pyrite form aggregates up to 4½ inches in diameter which contain only 5% relict quartz, (T.173). The pyrite, here appears to be wholly secondary and is associated with the mineral veins.

c. SULPHIDES IN THE LIMESTONES

Sulphides occurring in the limestones are generally finely disseminated and, where they are original, are associated with, and often replace, fossils. Many examples are seen in thin sections of limestones and also in the few samples of ironstone examined. Two typical occurrences of sulphides in limestones are given as follows.

In the Lunedale Quarries (953238), the Alternating Beds are exposed. A six inch pyritous algal limestone band occurs in the shales below the Cockle Shell Limestone. The polished section (T.361) shows the presence of about 2% pyrite replacing the groundmass of calcite in and near the fossils. The pyrite occurs as very small scattered grains varying in size from 0.001 to

0.5 mm in diameter. The occurrence of the sulphide in small discrete bodies associated with organic remains is indicative of bacterial formation in a manner similar to that described by Baker and Edwards (1951), although some may be of secondary origin as occasional veinlets of pyrite and calcite cut the rock.

In Hudeshope Beck (948271), a thin limestone band occurs below the Three Yard Limestone. It contains large brachiopods which are often pyritised. The polished section (T.372) contains about 3% pyrite in irregular masses which tend towards a euhedral character and which replace the shells. Each mass is composed of several crystals. In the groundmass of the shaly limestone, small irregular masses of pyrite with cube faces and small cubes up to 0.5 mm in diameter occur, replacing and often enclosing pockets of the groundmass.

East of the Burtreeford Disturbance area, the limestones up to and including the Scar Limestone often show replacement of pyrite by marcasite and vice versa. This effect is thought to be due to metamorphism by the Whin Sill.

d. SULPHIDES IN THE SHALES

Pyrite is common in the arenaceous shales where they grade upwards into the bottom of sandstones. Elsewhere, except for the shale nodules and massive pyrite bands which may be regarded as continuous and very large nodules, it is rare. Accumulations of finely disseminated pyrite also occur in material which tends to be nodular but still maintains the fissility of the shale. Occurrences of this material are regarded as poorly developed nodular bands. Two examples are here given of the occurrence of sulphides in shales.

On the Tees, at Broken Way (927266), a pyrite band occurs in shales above the Scar Limestone. The band varies from 0 to 9 inches in thickness and is generally massive, containing only a few small residual pockets of shale material (2 to 5%). The polished section (T.333) shows that the band is composed of large aggregates of pyrite in optical continuity which have replaced the groundmass. The mode of occurrence of this sulphide bed, in the form of a discontinuous "nodular" band, suggests that it is an unusually rich occurrence of sedimentary sulphide. There is no evidence pointing to a hydrothermal or metamorphic origin.

In Ettersgill (883310), a nodular shale occurs above the Three Yard Limestone. The nodules contain small masses of pyrite which, in polished section (T.193), are seen to be composed of anhedral to euhedral grains of pyrite varying from 0.001 to 0.75 mm in diameter.

e. METAMORPHISM OF THE SULPHIDES

The only non-metamorphic and non-hydrothermal sulphide occurring in the Teesdale rocks is the iron sulphide, pyrite. Pyrite occurring in the sandstones and limestones is often partially altered to marcasite as a result of metamorphism and the marcasite may show alteration to pyrite.

Pyrite occurring in the shale nodules yields a variety of minerals on metamorphism. East of the Burtreeford Disturbance Area, shales containing nodules over 100 feet above the Whin Sill may be affected.

Minerals found in the metamorphosed nodules include pyrite, pyrrhotite, marcasite, chalcopyrite, sphalerite and others. Several examples of shale nodules are described in order to illustrate their variation.

A nodular shale occurs below the Cockle Shell Limestone at Dirt Pit, Ettersgill (892299). This horizon is about 40 feet above the Whin Sill. The nodules are usually up to 1 inch thick and 3 to 6 inches in diameter. Polished sections (T.93a) show that near the periphery of the nodules calcite occurs which is interstitial to the clay minerals and which has been partially replaced by quartz. The centres of the nodules are invariably crossed by septarian veinlets containing calcite. The sulphides are arranged zonally.

The central portions contain pyrite grains up to 2 mm diameter, which fill the septarian cracks. In places the pyrite may be replaced by marcasite. There is an outer zone of pyrrhotite to each nodule. Pyrrhotite also occurs in the extremeties of the calcite veinlets. In the distal ends of the veinlets, small amounts of chalcopyrite also occur (Figure 11.1).

Both the pyrrhotite and the chalcopyrite are granular. Occasionally larger masses of pyrrhotite in the groundmass may enclose pyrite and marcasite. Where marcasite and pyrrhotite are in contact, reaction zones may show the presence of remnant pyrrhotite in marcasite. In some parts of the nodules, pyrrhotite is replaced by secondary recrystallised carbonate and quartz. Near to the pyrite veinlets, pyrrhotite is replaced by marcasite.

Similar nodules occur in the shale above the Cockle Shell Limestone at Scorberry Bridge on the Tees (910273). Here, the nodules (T.79) are up to 1 inch thick and 4 inches in diameter. The matrix of the nodules is of shale material replaced, in part, by quartz and carbonate.

The central portions contain pyrite masses which are irregular and up to 0.5 cm in diameter. Embayment of pyrite by marcasite and the presence of few euhedral boundaries to the pyrite suggests the replacement of pyrite by marcasite. The intermediate zones of the nodules contain pyrrhotite which is generally irregular and anhedral and is, in part, replaced by quartz.

The outer zones of the nodules contain vermiform masses of lath-like euhedral pyrrhotite crystals, (Figure 11.2). In places, the pyrrhotite is replaced by more massive pyrite which itself is often altered to marcasite. Many of the pyrrhotite crystals have very fine outer rims of pyrite and some appeared to have rims of ?pentlandite. The ?pentlandite rims were so fine that accurate optical determination was impossible. Staining techniques failed to reveal the presence of nickel and the rims are believed, therefore, to be composed of pyrite which have been obliquely polished with resultant anomalous optical effects. The soft shale matrix which is easily ground down is responsible for the oblique polishing at the margins of the grains.

A similar shale with nodules occurs above the Cockle Shell Limestone in Ettersgill (889292). Here, the nodules are up to 1½ inches by 4 inches (T.142b). The outer edges of the nodules are siliceous with an inner development of spheroidal carbonate. The polished section shows that there are three distinct zones to the nodules.

There is an inner zone of granular to euhedral pyrite which, in some cases, is enclosing grains of marcasite. Individual grains of marcasite are less than 0.17 mm in diameter. The middle zone consists of granular pyrrhotite with occasional grains of euhedral pyrite (Figure 11.3). The pyrrhotite is, in places, replaced by pyrite and marcasite (Figure 11.4). This zone also contains small granules of chalcopyrite. The outer zone contains small lath-like euhedral pyrrhotite crystals, up to 0.1 mm in length, together with a little anhedral pyrrhotite. Aggregates of pyrrhotite replace the groundmass and are intergrown with fine quartz. The quartz has, in places, replaced the pyrrhotite, the outlines of which can be detected in polarised reflected light.

The shale above the Scar Limestone in Smithy Sike (992292), which lies about 80 feet above the Whin Sill, contains similar nodules up to 2 inches by 3 inches (T.99b). These are of spotted shale. Spheroidal carbonate also occurs in the centres of the nodules. A typical polished section shows vermiform aggregates of pyrrhotite with pyrite towards the centres of the nodules. These aggregates are up to 2.5 mm in width and over 1 cm in length and contain euhedral laths (Figure 11.5) of pyrrhotite between 0.02 and 0.5 mm in length. The pyrrhotite occasionally replaces shell fragments. All of the crystals and crystal aggregates show thin outer rims of pyrite. A few aggregates are wholly of pyrite. A few grains of subhedral chalcopyrite occur which are up to 0.1 mm in diameter.

Pyrrhotite is replaced, in some parts, by spheroidal aggregates of carbonate and quartz which show the original outlines of the replaced pyrrhotite. In the centres of the nodules, whole aggregates of pyrrhotite may be replaced by pyrite. A few snall anhedral grains of sphalerite also occur in the nodules. Pyrrhotite, in places, may contain inclusions of ?goethite and under high power objectives may show the presence of "zwischenprodukte" (vide Ramdohr, page 556).

f. THE ORIGIN OF THE SEDIMENTARY SULPHIDES

Tarr (1927, 1928) and Love (1957) have shown that the presence of iron sulphides in sedimentary rocks is often due to local reducing conditions during deposition. The reducing conditions are associated with fairly rapid burial of micro and macro-organisms in areas of shallow water. Decay of the organic matter by putrifying bacteria releases organic acids, hydrogen sulphide and carbon dioxide from the proteins.

Tarr (1928) has shown that the syngenetic pyrite may occur in fossiliferous black shales but not in the overlying and underlying shales. Thus one possible criterion for the identification of an original pyrite-bearing rock is the restriction of the pyrite to one horizon. Many of the pyrite-bearing shales and other rock types in the Yoredales of Teesdale are fossiliferous but some are unfossiliferous. Love (1957) has shown, however, that very small micro-organisms may occur in the sulphide-bearing shales that may be undetected by the usual petrographic methods. Some pyrite samples from the Teesdale shales were dissolved in nitric acid in an attempt to find evidence of micro-organisms but they yielded no recognisable remains. It may be that conditions were not suitable for the preservation of such organisms in the shales examined.

The presence of varying amounts of fine carbonaceous material in all of the shales examined suggests that organic matter was not entirely lacking during the accumulation of the sediments. In addition, the bases of some of the sandstones have concentrations of pyrite associated with organic structures such as pit and mound structures.

Some pyrite may have been derived from groundwaters and deposited around original nuclei, as many sulphide masses have a secondary shell of pyrite. Allen, Crenshaw and Merwin (1914) mention the occurrence of groundwaters carrying sufficient iron and sulphur ions to precipitate pyrite in wells and storage tanks.

The localised occurrence of sulphides to nodules, bases of the sandstones, tops of limestones and other horizons of small vertical but large lateral distributions suggests that most of the pyrite is original. Where mineralisation has occurred with the introduction of pyrite, the occurrence of other minerals in association with the pyrite in faults and fissures helps to distinguish the origin of the pyrite.

Allen, Crenshaw and Merwin (1914) and Buerger(1934) have shown that the acidity of the environment is largely responsible for the type of iron sulphide deposited. Baker and Edwards (1951) have noted occurrences of marcasite and pyrite in relation to environment. Both the experimental evidence and observation of natural occurrences indicate that pyrite is formed under marine alkaline conditions whereas marcasite is formed under freshwater or brackish water acid conditions. Thus it might be expected that limestones contain pyrite whilst shales contain marcasite or pyrite depending on the nature of the depositional environment. However, marcasite may be formed in limestones due to acid solutions arising from decaying organic matter just below the interface of deposition or may be formed later due to circulation of acid groundwaters.

In Teesdale the restricted occurrence of marcasite to the beds which are within 120 feet of the Whin Sill, suggests that the marcasite here was probably of secondary origin and associated with the metamorphism.

The original sulphide was thus probably pyrite, associated with anaerobic bacteria which liberated sulphur and hydrogen sulphide from the organic matter entombed in the accumulating sediments. The reducing conditions, low oxygen content, and high carbon dioxide content of the environment would favour the formation of ferrous carbonate in solution. A reaction between the ferrous carbonate and the hydrogen sulphide would produce ferrous sulphide in the form of pyrite. This reaction can be illustrated by the following equation:

$$FeCO_3 + H_2S + S \longrightarrow FeS_2 + \underset{\displaystyle H_2O + CO_2}{\underset{\downarrow\uparrow}{H_2CO_3}}$$

g. THE ORIGIN OF THE SULPHIDES IN THE METAMORPHOSED ROCKS

Marcasite is found in the sandstones, limestones and shales which have been metamorphosed. Buerger (1934) has shown that pyrite may give rise to marcasite under varying conditions such as the presence of acid solutions, or heat and an excess of iron, or a combination of the two factors. He believed that pyrite contained less iron than marcasite and that the reaction was as follows:

$$\text{Cubic pyrite } FeS_2 \longrightarrow \begin{array}{l}\text{Orthorhombic}\\ \text{marcasite } Fe\end{array} \Big/ \substack{Fe_x\\ \\ S_{2-x}} \Big/ + xS.$$

where x was about 0.004.

However, more recently, Baker and Edwards (1951) have shown that in the sedimentary sulphides which they examined, pyrite contained an excess and marcasite a deficiency of iron when compared to the ideal formula FeS_2.

Marcasite is thus formed by the elimination of some of the iron from the pyrite molecule by heating to moderate temperatures. According to Allen, Crenshaw and Merwin (1914), marcasite forms at "low temperatures" in "weak" acid solutions to higher temperatures in stronger acid solutions, whilst pyrite is formed in alkaline solutions from temperatures as low as 55°C to at least 300°C. The existence of spheroidal carbonate and quartz in some of the nodules suggests that redistribution of the various ions and minerals may have been accomplished in aqueous media which varied in pH from time to time during the period of thermal metamorphism.

Both pyrrhotite and marcasite are found in the metamorphosed shale nodules. The formation of pyrrhotite from pyrite is also promoted by heat thus:

$$\text{Pyrite } FeS_2 \xrightarrow[\text{Heat}]{} \text{Pyrrhotite } Fe_{1-n}S(\text{approx. } Fe_7S_8) + nFe + S.$$

On cooling, the reverse reaction may occur if the products are not removed before this can happen.

According to Tsusue (1961), the minimum temperature of formation of pyrrhotite coexisting with pyrite is about 490°C.

Thus in the shale nodules pyrite is transformed to either marcasite or pyrrhotite and the marcasite may be transformed to pyrrhotite in the following manner.

$$\begin{tikzcd}
\underset{\text{Pyrite}}{Fe_{1+x}S_2} \arrow[rr, "\text{Heat}", "\text{minus } (x+y) \text{ Fe}"'] \arrow[dr, "\text{Heat}", "\text{Minus } (x+n) \text{ Fe and minus S.}"'] & & \underset{\text{Marcasite}}{Fe_{1-y}S_2} + (y+x) \text{ Fe.} \arrow[dl, "\text{Heat}"', "\text{Minus } (y-n) \text{ Fe and minus S.}"] \\
& \underset{\text{Pyrrhotite.}}{Fe_{1-n}S.} &
\end{tikzcd}$$

x and y are very small and n is approximately 0.125. The total result of these reactions is the production of excess iron and sulphur. The minor amount of iron will react with sulphur to form more iron sulphides but providing no complete reversal of the reactions occur there will be an excess of sulphur.

Spectrographic analyses, kindly carried out by P.M. Harris of Oxford and Professor J.E. Hawley of Queens University, Ontario, of two unmetamorphosed shale nodule pyrite samples gave the following results:

Ag	0.003%
Pb	0.038 to 0.073%
Cu	Trace (0.004%)
Ni	0.0010 to 0.0052%
Co	0.007 (0.0035 to 0.0045)%
V	0.0047 to 0.0050%
Cr	0.0015 to 0.0027%
Mn	0.013 to 0.021 (0.04 to 0.055)%
Ti	0.042 to 0.096%
Se	0.0015 to 0.0024%

The figures in brackets are the results of less reliable analyses received. The amounts of copper, cobalt and nickel are very small.

Although copper and zinc are not present in great quantities in the pyrite of the unmetamorphosed shale nodules, sphalerite and chalcopyrite are present in the metamorphosed nodules. The lack of these two minerals in the metamorphosed limestones and sandstones and the practically impermeable nature of the shales suggests that these two sulphides were not deposited from solutions emanating from the Whin Sill. It is here suggested that they were formed by reaction between sulphur and some iron expelled during metamorphism of the pyrite to marcasite and pyrrhotite and traces of copper and zinc occurring in the shales.

There is a distinct difference between the sulphide assemblages in the metamorphosed shales and those in the limestones and sandstones. The lack of zinc and copper-bearing sulphides may be the result of the lack of trace elements in the original constituents; these are usually very low in purer sandstones such as those occurring in Teesdale and in the calcite of limestones. The absence of pyrrhotite is more difficult to explain particularly when both sandstones and limestones near to the Whin Sill may contain marcasite. Pyrrhotite may have been formed but subsequently recombined with sulphur on cooling to yield pyrite and marcasite in the absence of sulphur acceptors such as copper and zinc. These reverse reactions are seen to occur in some of the shale nodules where pyrrhotite is rimmed by pyrite and marcasite. However, there is no evidence of the prior existence of pyrrhotite in the limestones and sandstones, and it is probable that it was never formed during the metamorphism, possibly due to different pressure and solution conditions encountered in these more porous rocks.

REFERENCES

The following list of references have been found to be useful during the period of work. Only the more important references appear in the text. The references are listed under authors, arranged in alphabetical order. Where an author has been responsible for more than one work, the different works are listed in order of printed publication and where more than one author is responsible for the work, the reference is listed under the first name appearing on the title page.

ALLEN, E.T., CRENSHAW, J.L., and MERWIN, H.E., 1914. The effect of Temperature and acidity in the formation of marcasite (FeS_2) and wurtzite (ZnS); a contribution to the genesis of unstable forms. Am. J. Sc. 38, 393-431.

ALLEN, V.T., and JOHNS, W.D., 1960. Clays and Clay minerals of New England and Eastern Canada. Geol. Soc. Am. Bull. 71, 75-86.

ADLER, H., BRAY, E.E., STEVENS, N.P., HUNT, J.M., KELLER, W.D., PICKET, E.E. and KERR, P.F., 1950. Infrared Spectra of Reference Clay Minerals, Rept. 8, Am. Petrol. Inst. Proj. 49.

BAKER G., and EDWARDS, A.B., 1951. Some occurrences of supergene iron sulphides in relation to their environments of deposition. J.S.P. 21, 34-46.

BANERJEE, A., 1959. Petrography and facies of some Upper Visean (Mississippian) limestones in North Wales. J.S.P. 29, 377-390.

*
BARSHAD, I., 1948. Vermiculite and its relation to biotite as revealed by base exchange reactions, X-ray analysis, D.T.A. and water content. Am. Min. 33, 655-678.

BASKIN, Y., 1956. A Study of authigenic feldspars. J. Geol. 64, 132-155.

BATES, C.C., 1953. Rational theory of the delta formation. Am. Assoc. Pet. Geol. Bull., 37, 2119-2162.

BEAL, M.A., and SHEPARD F.P., 1956. A use of roundness to determine depositional environments. J.S. P. 26, 46-60.

BOKMAN, J., 1955. Sandstone classification: Relation to composition and texture. J.S.P. 25, 201-206.

BOSWELL, P.G.H., 1960. The term graywacke. J.S.P. 30, 154-157.

BRADLEY, W. F., 1945. Diagnostic criteria for clay minerals. Am. Min. 30, 704-718.

...................., 1953. Analysis of mixed-layer clay mineral structures. Anal. Chem. 25, 727-730.

...................., 1954, X-ray diffraction criteria for the characterisation of chloritic material in sediments. 2nd Nat. Conf. Clays and Clay Mins., 324-334.

BRINDLEY, G.W., Ed. 1951. X-ray identification and crystal structures of clay minerals. Min. Soc. Monograph. 345p.

...................., and GILLERY, F.H., 1956. X-ray identification of chlorite species. Am. Min. 41, 169-186.

...................., and ROBINSON, K., 1945. Structure of kaolinite. Nature 156, 661-662.

BRINDLEY, G.W., and ROBINSON, K., 1946. The structure of kaolinite. Min. Mag. 27, 242-253.

................................,1946. Randomness in the structures of the kaolinitic clay minerals. Trans. Faraday Soc. 42B, 198-205.

................................,1951. The Chlorite Minerals, in X-ray Identification and Structures of Clay Minerals. Pub. Min. Soc. Ed. G.W. Brindley.

................................, and GOODYEAR, J., 1948. X-ray studies of halloysite and metahalloysite. Min. Mag. 28, 393- 428.

BROUGH, J., 1929. On rhythmic sedimentation in the Yoredale Series. Proc. Univ. Durham Phil. Soc. 8, 116-126.

................, and ROBERTSON, R.H.S., 1958. A chlorite convention for the appraisal of Scottish fireclays. Clay Min. Bull. 3, 221-231.

BROWN, C. Q., and INGRAM, R.L., 1954. The clay minerals of the Neuse River sediments. J.S.P. 24, 196-199.

BROWN, G., 1951. Nomenclature of the Mica clay minerals. See X-ray identification and Structures of Clay Minerals. Pub. Min. Soc. Ed. G.W. Brindley. 155.

BROWN, G., 1955. Report of the Clay Minerals Group Sub-Committee on Nomenclature of Clay Minerals. Clay Min. Bull. 2, 294-302.

............., and DIBLEY, G.C., 1956. Moderately low angle measurements with a 9 cm. powder camera. Clay Min. Bull. 3, 46-47.

............., and MACEWAN, D.M.C., 1951. X-ray Diffraction by structures with Random Interstratification. see X-ray Identification and Structures of Clay Minerals. Pub. Min. Soc. Ed. G.W. Brindley. 266.

............., and NORRISH, K., 1952. Hydrous micas. Min. Mag. 29, 929-932.

BUERGER, M.J., 1934. The pyrite - marcasite relation. Am. Min. 19, 37-61.

BURST, J. F. Jr. 1959. Postdiagenetic clay mineral environmental relationships in the Gulf Coast Eocene. 6th Nat. Conf. Clays and Clay Mins. 327-341.

CAILLERE, S., and HENIN, S., 1949. Transformation of minerals of the montmorillonite family into 10Å micas. Min. Mag. 28, 606-611.

..........................., 1949. Experimental formation of the chlorites from montmorillonite. Min. Mag. 28, 612-620.

CAROZZI, A.V., 1960. Microscopic Sedimentary Petrography. Wiley, New York. 485p.

CARR, K., GRIMSHAW, R.W., and ROBERTS, A.L., 1953. A hydrous mica from Yorkshire fireclay. Min. Mag. 30, 139-144.

CARROLL, D., 1959. Ion exchange in clays and other minerals. Geol. Soc. Am. Bull. 70, 749-780.

CHOUBERT, B., 1952. Recent sedimentation in French Guiana. Cong. Geol. Int. 19th Session. Alger. 65-73.

CLARKE, F.W., 1924. The data of geochemistry. U.S. Geol. Surv. Bull. 770, 814 p., Washington.

CLOUGH, C.T., 1876. The Section at High Force, Teesdale. Quart. Journ. Geol. Soc., 32, 466.

..............., 1880. The Whin Sill as an assimilator of the surrounding beds. Geol. Mag. 7, 433-447.

DAPPLES, E. C., 1947. Sandstone types and their associated depositional environments. J. S. P. 17, 91-100.

..............., KRUMBEIN, W.C., and SLOSS, L.L., 1950. The organisation of sedimentary rocks. J.S.P. 20, 3-20.

.........................., 1953. Petrologic and lithologic attributes of sandstones. J. Geol. 61, 291-317.

DEGENS, E.T., WILLIAMS, E.G., and KEITH, M.L., 1957. Environmental studies of Carboniferous sediments. Part 1. Geochemical criteria for differentiating marine from fresh-water shales. Am. Assoc. Pet. Geol. Bull. 41, 2427-2455.

DIETZ, R.S., 1942. Clay minerals in Recent marine sediments. Am. Min. 27, 219-220.

DROSTE, J. B., 1956. Alteration of clay minerals by weathering in Wisconsin tills. Geol. Soc. Am. Bull. 67, 911-918.

DUNHAM, K.C., 1950. Lower Carboniferous sedimentation in the Northern Pennines (England). Cong. Geol. Int. 18th Session. England. (for 1948) Pt.4, 46-63.

.............., 1948. Geology of the Northern Pennine Orefield. vol. 1. H.M.S.O. 357p.

CROOK, K.A.W., 1960. Classification of arenites. A.J. Sc. 258, 419-428.

EDWARDS, A.B., and CARLOS, G.C. 1954. The selenium content of some Australian sulphide deposits. Austral. Inst. Min. Met. 172, 31-63.

EDWARDS, W., STUBBLEFIELD, C.J. and DUNHAM, K.C., 1947. Marine bands and other faunal marker-horizons in relation to the sedimentary cycles of the Coal Measures of Nottinghamshire and Derby. Q.J.G.S. 103, 209-260.

EMERY, K. O., GORSLINE, D.S., UCHUPI, E., and TERRY, R.D., 1957. Sediments of three bays of Baja, California: Sebastian Viscalno, San Cristobal and Todo Santos. J.S.P. 57, 95-115.

ENNOS, F.R., and SUTCLIFFE, R., On the estimation of chlorite in sedimentary rocks by extraction with dilute hydrochloric acid. Appendix of Spec. Min. Rep. Min. Res. G. B. 29, Iron Ores.

FISK, H.N., McFARLAN, E., KOLB, C.R. and WILBERT, L.J., 1954. Sedimentary framework of the modern Mississippi Delta'. J.S.P. 24, 76-99.

FLEISCHER, M., 1955. Minor elements in some sulphide minerals. Econ. Geol. 50th Anniv. vol. 970-1024.

FOLK, R.L., 1954. The distinction between grain size and mineral composition in sedimentary rock nomenclature. J. Geol. 62, 344-359.

............, 1956. Discussion: The role of texture and composition in sandstone classification. J.S.P. 26, 166-171.

............., 1959. Practical petrographic classification of limestones. Am. Assoc. Pet. Geol. Bull. 43, 1-38.

............., and WARD, W.C., 1957. Brazos River Bar: A study in the significance of grain size parameters. J.S.P. 27, 3-26.

FRANK-KAMENETSKY, V.A., 1960. A Crystallochemical Classification of Simple and Interstratified Clay Minerals. Clay Min. Bull. 4, 161-172.

FRUEH, A.J., 1954. The use of the zone theory in problems of sulphide mineralogy. Geochim. Cosmochim. Acta. 6, 78-89.

GALLIHER, E.W., 1935. Glauconite genesis. Geol. Soc. Am. Bull. 46, 1351-1366.

GARWOOD, E.J., 1912. The Lower Carboniferous succession in the northwest of England. Q. J.G. S. 68, 449-586.

GEORGE, T.N., 1958. Lower Carboniferous palaeogeography of the British Isles. P.Y.G.S. 31, 227-318.

GILBERT, C.M., 1949. Cementation of some California Tertiary reservoir sands. J. Geol. 57, 1-17.

GLASS, H.D., 1958. Clay mineralogy of Pennsylvanian sediments in Southern Illinois. 5th Nat. Conf. Clays Clay Mins. 227-241.

..........., POTTER, P.E. and SIEVER, R.E., 1956. Clay mineralogy of some basal Pennsylvanian sandstones, clays and shales. Am. Assoc. Pet. Geol. Bull. 40, 750-754.

GOLDSCHMIDT, V.M., 1954. Geochemistry. Oxford. Ed. A. Muir. 730 p.

GOULD, H.R., and STEWART, R.H., 1955. Continental terrace sediments in the Northeastern Gulf of Mexico. Soc. Econ. Pal. Min. Spec. Pub. 3, 2-20.

GREENMAN, N. N., and LEBLANC, R. J., 1956. Recent marine sediments and environments of the Northwest Gulf of Mexico. Am. Assoc. Pet. Geol. Bull. 40, 813-847.

GREENSMITH, J. T., 1956. Sedimentary structures in the Upper Carboniferous of north and Central Derbyshire, England. J.S.P. 26, 343-355.

.................., 1957. Lithology, with particular reference to cementation of Upper Carboniferous sandstones in North Derbyshire, England. J.S.P. 27, 405-416.

..................., 1958. Preliminary observations on Chemical data from some British Upper Carboniferous shales. J.S.P. 28, 209-210.

GRIFFIN, G.M., and INGRAM, R.L., 1955. Clay minerals of the Neuse River sediments. J.S.P. 25, 194-200.

GRIFFITHS, J.C., 1951. Size versus sorting in some Caribbean sediments. J. Geol. 59, 211-243.

GRIM, R.E., 1934. The petrographic study of clay minerals - a laboratory note. J.S.P. 4, 45-46.

................., 1951. The depositional environment of red and green shales. J.S.P. 21, 226-232.

................., 1953. Clay Mineralogy. McGraw-Hill, New York. 384 p.

................., 1958. Concept of diagenesis in argillaceous sediments. Am. Assoc. Pet. Geol. Bull. 42, 246-253.

................., and ALLEN, V.T., 1938. Petrology of the Pennsylvanian underclays of Illinois, Geol. Soc. Am. Bull. 49, 1485-1514.

GRIM, R.E., BRADLEY, W.F. and BROWN, G., 1951. The Mica Clay Minerals in X-ray Identification and Structures of Clay Minerals. Pub. Min. Soc. Ed. G.W. Brindley.

............, BRAY, R.H. and BRADLEY, W.F., 1937. The mica in argillaceous sediments. Am. Min. 22, 813-829.

............, DIETZ, R.S., and BRADLEY, W.F., 1949. Clay mineral composition of some sediments from the Pacific Ocean off the California coast and the Gulf of California. Geol. Soc. Am. Bull. 60, 1758-1808.

............, and JOHNS, W.D., 1954. Clay mineral investigation of sediments in the Northern Gulf of Mexico. 2nd Nat. Conf. Clays Clay Mins. 81-103.

.............................., 1958. Clay Mineral composition of Recent sediments From the Mississippi River Delta. J.S.P. 28, 186-199.

GRIM, R.E., KERR, P.F., and BRAY, R.H., 1935. Application of clay mineral technique to Illinois clay and shale. Geol. Soc. Am. Bull. 46, 1909-1926.

..........., LAMAR, J.E., and BRADLEY, W.F., 1937. Clay minerals in Illinois limestones and dolomites. J. Geol. 45, 829-843.

GRUNER, J.W., 1934. The structures of vermiculites and their collapse by dehydration. Am. Min. 19, 557-575.

GUNN, W. and CLOUGH, C.T., 1878. Discovery of Silurian beds in Teesdale. Q.J.G.S. 34, 27-34.

HALL, A.J., 1941. The relation between colour and chemical composition in biotites. Am. Min. 26, 29-33.

..........., 1941. The relation between chemical composition and refractive index in the biotites. Am. Min. 26, 34-41.

HALLIMOND, A.F., 1925. Staining techniques for carbonates in thin section. Geol. Surv. Min. Res.

HARRISON, J.L., and MURRAY, H.H., 1959. Clay mineral stability and formation during weathering. 6th Nat. Conf. Clays Clay Mins. 144-153.

HARRY, W.T., 1950. Basement Carboniferous in Upper Teesdale, N. Yorks, Geol. Mag., 87, 297-299.

HEALD, M.T., 1956. Cementation of Simpson and St. Peter Sandstones in parts of Oklahoma, Arkansas and Missouri. J. Geol. 64, 16-30.

HEINRICH, E. W., 1946. Studies in the mica group: the biotite-phlogopite series. Am. J. Sc. 244, 836-848.

HENDRICKS, S. B. and ROSS, C.S., 1941. Chemical composition and genesis of glauconite and celadonite. Am. Min. 26, 683-708.

................, and TELLER, E., 1942. Random layer lattices. J. Chem. Phys. 10, 147-167.

HEY, R.W., 1956. Chert and limestones from the Crow Series near Richmond, Yorkshire. P.Y.G.S. 30, 289-299.

HEYSTEK, H., 1954. An occurrence of a regular mixed-layer clay mineral. Min. Mag. 30, 400-408.

............, 1955. Some hydrous micas in South African clays and shales. 3rd Nat. Conf. Clays Clay Mins. 337-355.

HIND, W., 1902. On the characters of the Carboniferous rocks of the Pennine system. P.Y.G.S. 14, 422-464.

HOLLINGWORTH, R.P., 1957. Spectrophotometric determination of fluorine in rocks. Anal. Chem. 29, 1130-1133.

HOLMES A., and HARWOOD, H.F., 1928. The Age and Composition of the Whin Sill and related dykes of the North of England. Min. Mag. 21, 493-542.

............................, 1929. The tholeiite dykes of the North of England. Min. Mag., 22, 1-52.

HOPKINS, M.E., 1958. Geology and petrology of the Anvil Rock Sandstone of Southern Illinois. Ill. State Geol. Surv. Circ. 256.

HUDSON, R.G.S., 1924. On rhythmic succession of the Yoredale Series in Wensleydale. P.Y.G.S. 20, 125-135.

HUTCHINGS, W.M., 1895. An interesting contact-rock, with notes on contact-metamorphism. Geol. Mag. 2, 4th decade, 122-131, 163-169, 314-317.

................, 1898. The contact-rocks of the Great Whin Sill. Geol. Mag. 5, 4th decade, 69-82, 123-131.

JACKSON, M.L., 1959. Frequency distribution of clay minerals in major great soil groups as related to the Factors of soil Formation. 6th Conf. Clays Clay Mins. 133-143.

JOHNS, W.D., and GRIM, R.E., 1958. Clay mineral composition of the Recent sediments from the Mississippi River Delta. J.S. P. 28, 186-199.

............, GRIM, R.E., and BRADLEY, W.F., 1954. Quantitative estimations of clay minerals by diffraction methods. J.S.P. 24, 242-251.

JOHNS, W.D., and MURRAY, H.H., 1959. Empirical crystallinity index for Kaolinite. - 40th Ann. Meeting of Min. Soc. of America.

JOHNSON, G.A.L., 1959. The Lower Carboniferous stratigraphy of the Roman Wall district in western Northumberland. P.Y.G.S. 32, 83-130.

................, 1960. Palaeogeography of the Northern Pennines and Part of North Eastern England during the Deposition of Carboniferous Cyclothemic Deposits. Int. Geol. Cong. Rept. of 21st. Session. Sect. 12., 118-128.

JOHNSTON, W.A., 1922. The character of the stratification of the sediments in the Recent delta of the Fraser River, British Columbia, Canada. J.Geol. 30, 115-129

JONES, H.LI,L., 1956. The Stratigraphy and Structure of the Area between Middleton in Teesdale and Woodlands. Ph.D. Thesis, Durham.

JONES, O.T., 1938. On the Evolution of a Geosyncline. Proc. Geol. Soc. Lond. 94, 62-66.

JULIAN, Y., 1952. Observations sedimentologiques sur Transgression marine.Cong. Geol. Int. 19th Session - Alger. Pt. 8, 235-240.

KEITH, M.L., and DEGENS, E.T., 1959. Geochemical indicators of marine and fresh-water sediments. Researches in Geochemistry. Ed. P.H. Abelson. Wiley, New York. 43-44.

KELLER, W.D., 1946. Evidence of texture on the origin of the Cheltenham Fireclay of Missouri and associated shales. J.S.P. 16, 63-71.

.............., 1953. Illite and montmorillonite in green sedimentary rocks, J.S.P. 23, 3-9.

.............., 1953. Clay Minerals in the type section of the Morrison Formation. J.S.P. 23, 93-105.

KELLER, W.D., 1956. Clay minerals as influenced by environments of their formation. Am. Assoc. Pet. Geol. Bull. 40, 2189-2210.

.............., 1958. Argillation and direct bauxitisation in terms of concentration of hydrogen and metal cations at surface of hydrolyzing aluminium silicates. Am. Assoc. Pet. Geol. Bull. 42, 223-245.

.............., and CHUEN PUTING, 1950. The petrology of a specimen of the Perry Farm Shale. J.S.P. 20, 123-132.

.............., and PICKETT, E.E., 1950. The absorption of infrared radiation by clay minerals. Am. J. Sc. 248, 264-273.

KERR, P.F., 1951. Ed., Reference clay minerals. A.P.I. Research Project. 49.

KOKUBU, N., 1956. Fluorine in rocks. Mem. Fac. Sci., Kyushu Univ., Ser. C, 2, 95.

KRUMBEIN, W. C. 1935. Thin section mechanical analysis of indurated sediments. J. Geol. 43, 482-496.

..............., 1947. Shales and their environmental significance. J.S.P. 17, 101-106.

..............., 1954. The tetrahedron as a facies mapping device. J.S.P. 14, 3-19.

..............., and SLOSS, L.L., 1959. Stratigraphy and Sedimentation. Freeman, San Francisco. 497 p.

KRYNINE, P.D., 1948. The megascopic study and field classification of sedimentary rocks. J. Geol. 56, 130-165.

KULBICKI, G., and GRIM, R.E., 1959. A new method for thermal dehydration studies of clay minerals. Min. Mag. 32, 53-62.

KULLERUD, G. and YODER, H.S., 1959. Pyrite stability relations in Fe-S system. Econ. Geol. 54, 533-572.

LALOU, C., 1957. Studies on bacterial precipitation of carbonates in sea water. J.S.P. 27, 190-195.

LAUNER, P.J., 1952. Regularities in the infrared absorption spectra of silicate minerals. Am. Min. 37, 764-784.

LEVINSON, A.A., 1955. Studies in the mica group: polymorphism among the illites and hydrous micas. Am. Min. 40, 41-49.

LLOYD, R.M., 1954. A technique for separating clay minerals from limestones. J.S.P. 24, 218.

LOVE, L.G., 1957. Micro-organisms and the presence of sygenetic pyrite. Q.J.G.S. 113, 429-440.

LYON, R.J.P., and TUDDENHAM, W.M. 1960. Infrared determination of the kaolin group minerals. Nature. 185, 835-837.

MACEWAN, D.M.C., 1946. The nomenclature of the halloysite minerals. Min. Mag. 28, 36-44.

MACKENZIE, R.C., 1956. Methods of separation of soil clays in use at the Macaulay Institute for Soil Research. Clay Min. Bull. 3, 4-6.

MACKENZIE, R.C., Ed., 1957. The Differential Thermal Investigation of Clays. Min. Soc., London. 456 p.

................, 1957. The illite in some O.R.S. soils and sediments. Min. Mag. 31, 681-689.

................, 1959. The Classification and Nomenclature of Clay Minerals. Clay Min. Bull. 4, 52-66.

................, and FARQUHARSON, K.R., 1956. A method for concentration of dilute clay suspensions without coagulation. Clay Min. Bull. 3, 7-13.

................, WALKER, G.F. and HART, R., 1949. Illite occurring in decomposed granite at Ballater, Aberdeenshire. Min. Mag. 28, 704-713.

MATHER, K., 1955. Discussion: Terminology of limestone and related rocks: an interim report. J.S.P. 25, 304-305.

MATHIAS, H.E., 1928. Syngenetic origin of pyrite concretions in the Pennsylvanian shales of North-central Missouri. J. Geol. 36, 440-450.

MILLER, D.N., 1957. Authigenic biotite in spheroidal reduction spots, Pierce Canyon Red Beds, Texas and New Mexico. J.S.P. 27, 177-180.

MILLER, H., 1887. The Geology of the country around Otterburn and Elsdon. Mem. Geol. Surv. Gt. Brit., London.

MILLER, R.L., 1954. A model for the analysis of environments of sedimentation. J. Geol. 62, 108-113.

MILLOT, G., 1952. Heritage et neoformation dans la sedimentation argileuse. Cong. Geol. Int. 19th Session - Alger. Fas. 18. 163-177.

MILNE, I.H., and EARLEY, J.W., 1958. Effect of source and environment on clay minerals. Am. Assoc. Pet. Geol. Bull. 42, 328-338.

MILNE, I.H., and SHOTT, W.L., 1958. Clay mineralogy of Recent sediments from the Mississippi Sound area. 5th Nat. Conf. Clays Clay Mins. 253-265.

MILNER, H.B., 1940. Sedimentary Petrography. 3rd. Ed. Murby, London. 666p.

MITCHELL, B.D., and MACKENZIE, R.C., 1954. The removal of free iron oxide from clays. Clay Min. Bull. 2, 115.

MOORE, D., 1958. The Yoredale Series of Upper Wensleydale and adjacent parts of north-west Yorkshire. P.Y.G.S. 31, 91-148.

..............., 1959. Role of deltas in the formation of some British Lower Carboniferous cyclothems. J. Geol. 67. 522-539.

..............., 1960. Sedimentation units in sandstones of the Yoredale Series (Lower Carboniferous) of Yorkshire, England. J.S.P. 30, 218-227.

MOORE, D.G., and SCRUTON, P.C., 1957. Minor internal structures of some Recent unconsolidated sediments. Am. Assoc. Pet. Geol. Bull. 41, 3723-2751.

MOORHOUSE, W.W., 1959. The Study of Rocks in Thin Section. Harper, New York, 514 p.

MORETTI, F.J., 1957. Observations on Limestones. J.S.P. 27, 282-292.

MORRIS, W.J., 1957. Effects of sphericity, roundness and velocity on traction transport of sand grains. J.S.P. 27, 27-31.

MUDGE, M.R., 1956. Sandstones and channels in Upper Pennsylvanian and Lower Permian of Kansas. Am. Assoc. Pet. Geol. Bull. 40, 654-678.

MUELLER, J.C., and WANLESS, H.R., 1957. Differential compaction of Pennsylvanian sediments in relation to sand-shale ratios. J.S.P. 27, 64-88.

MURRAY, H. H., 1954. Genesis of clay minerals in some Pennsylvanian shales of Indiana and Illinois. 2nd Nat. Conf. Clays Clay Mins. 47-67.

..............., 1954. Structural variations of some kaolinites in relation to dehydrated halloysite. Am. Min. 39, 97-108.

..............., and HARRISON, J.L., 1956. Clay mineral composition of Recent sediments from the Sigsbee Deep. J.S.P. 26, 363-368.

..............., and SAYYAB, A.S., 1955. Clay mineral studies of some Recent marine sediments off the North Carolina Coast. 3rd Nat. Conf. Clays Clay Mins. 430-441.

McKEE, E. D., 1939. Some types of bedding in the Colorado River Delta. J.Geol. 47, 64-81.

..............., 1957. Primary structures in some Recent sediments. Am. Assoc. Pet. Geol. 41, 1704-1747.

NAGELSCHMIDT, G.,1937. X-ray investigations on clays, Part 3. Zeit. f. Krist., 97, 514-521.

NAGELSCHMIDT, G., and HICKS, D., 1942. The mica of certain Coal Measure shales in South Wales. Min. Mag. 26, 297-303.

NAUMANN, C.F., 1858. Lehrbuch der Geognosie. Engelmann, Leipzig. 960p.

NELSON, B.W., and ROY, R., 1954. New data on the composition and identification of the chlorites. 2nd Nat. Conf. Clays Clay Mins. 335-348.

NEWHOUSE, W.H., 1927. Some forms of iron sulfide occurring in coal and other sedimentary rocks. J.Geol. 35, 73-83.

NOVIN, E.S., 1952. The sediments of the Central Tyrrhenian Sea. Cong. Geol. Int. 19th Session - Alger. Sect. 4. 87-97.

OINUMA, K., KOBAYASHI, and SUDO, T., 1959. Clay mineral composition of some Recent marine sediments. J.S.P. 29, 56-63.

PASSEGA, R., 1957. Texture as characteristic of clastic deposition. Am. Assoc. Pet. Geol. Bull. 41, 1952-1984.

PETTIJOHN, F. J., 1948. Studies for students: A preface to the classification of sedimentary rocks. J. Geol. 56, 112-117.

................., 1954. Classification of sandstones. J. Geol. 62, 360-365.

................., 1957. Sedimentary Rocks. 2nd Edition. Harper. N.Y. 708p.

.................., 1960. Discussion: The term Graywacke. J.S.P. 30, 627.

PHILLIPS, J., 1836. Illustrations of the geology of Yorkshire. Part 2. The Mountain Limestone District. London.

POTTER, P.E., and GLASS, H.D., 1958. Petrology and sedimentation of the Pennsylvanian in Southern Illinois: A vertical profile. Illinois State Surv. Rep.204.

............., and OLSEN, J.S., 1954. Variance components of cross-bedding directions in some basal Pennsylvanian sandstones of the Eastern Interior Basin: Geological application. J.Geol. 62, 50-73.

POTTER, P.E., and SIEVER, R., 1956. Sources of Basal Pennsylvanian sediments in the Eastern Interior Basin. J. Geol. 64, 225-455.

............., NOSOW, E., SMITH, N.M., SWANN, D.H., and WALKER, F.H., 1958. Chester cross-bedding and sandstone trends in the Illinois Basin. Am. Assoc. Pet. Bull. 42, 1013-1046.

POWERS, M.C., 1953. A new roundness scale for sedimentary particles. J.S.P. 23, 117-119.

..............., 1954. Clay mineral diagenesis in the Chesapeake Bay area. 2nd Nat. Conf. Clays Clay Mins. 68-80.

..............., 1957. Adjustment of land derived clays to the marine environment. J.S.P. 27, 355-372.

................, 1959. Adjustment of clays to chemical change and the concept of the equivalence level. 6th Nat. Conf. Clays Clay Mins. 309-326.

PRYOR, W.A., and GLASS, H.D., 1961. Cretaceous - Tertiary clay mineralogy of the Upper Mississippi embayment. J.S.P. 31, 38-51.

QUAIDE, W., 1957. Clay minerals from the Ventura Basin, California. J.S.P. 27, 336-341.

RAMDOHR, P., 1960. Die Erzmineralien und ihre verwachsungen. Akademie Verlag Berlin. 1089 p.

RAMSDELL, L.S., 1947. Studies in Silicon Carbide. Am. Min. 32, 64-82.

RAYNOR, D.H., 1953. The Lower Carboniferous rocks in the North of England: A review. P.Y.G.S. 28, 231-315.

READING, H.G., 1957. The stratigraphy and structure of the Cotherstone Syncline. Q.J.G.S. 113, 27-56.

RICHARDSON, W.A., 1919. On the origin of septarian structure. Min. Mag. 18, 327-338.

RICHARDSON, W.A., 1921. The relative age of concretions. Geol. Mag. 58, 114-124.

RIVIERE, A., 1952. Sur l'origine des argiles sedimentaires. Cong. Geol. Int. 19th Session - Alger, Fasc. 18. 177-181.

............., and VERNHET, S., 1952. Influence des matieres organiques sur la sedimentation des argiles. Cong. Geol. Int. 19th Session - Alger. Fasc. 4. 105-108.

ROBBINS, C., and KELLER, W.D., 1952. Clays and other non-carbonate minerals in some limestones. J.S.P. 22, 146-152.

ROBERTSON, R.H.S., 1956. Preparatory techniques. Clay Min. Bull. 3, 2-3.

...................., BRINDLEY, G.W., and MACKENZIE, R.C., 1954. Mineralogy of kaolin clays from Pugu, Tanganyika. Am. Min. 39, 118-138.

ROBERTSON, T., 1948. Rhythm in sedimentation and its interpretation: with particular reference to the Carboniferous sequence. Trans. Geol. Soc. Edin., 14, 141-175.

.............., 1952. Plant control in rhythmic sedimentation. Third Cong. Avanc. Etudes Strat. Geol. Carbon., 515-521.

RODGERS, J., 1950. The nomenclature and classification of sedimentary rocks. Am. J. Sc. 248, 297-311.

..........., 1954. Terminology of limestone and related rocks: An interim report. J.S.P. 24, 225-234.

ROMO, L.A. and ROY, R., 1955. F-OH Exchange in layer silicates. Abstract. 3rd Nat. Conf. Clays Clay Mins. 402.

ROSS, C.S. and KERR, P.F., 1931. Kaolin minerals. U.S. Geol. Survey. Prof. Paper 165E, 151-176.

RUSNAK, G.A., 1957. A fabric and petrologic study of the Pleasantview Sandstone. J.S.P. 27, 41-55.

RUSSEL, R. J., 1958. Geological geomorphology. Geol. Soc. Am. Bull. 69, 1-22.

RUTHRUFF, R.F., 1941. Vermiculite and hydrobiotite. Am. Min. 26, 478-484.

SCHULTZ, L.G., 1958. Petrology of underclays. Geol. Soc. Am. Bull. 69, 363-402.

SCRUTON, P.C., 1950. The petrology and environment of deposition of the Warner, Little Cabin and Hartshorne Sandstones in Northeast Oklahoma. Am. J. Sc. 248, 408-426.

.............., 1955. Sediments of the Eastern Mississippi Delta. Soc. Econ. Pal. Min. Spec. Pub. 3. 21-51.

SEDGWICK, A., 1827. On the Association of Trap Rocks with the Mountain Limestone Formation of High Teesdale. Camb. Phil. Trans. 2, 139.

SHEPARD, F.P., 1954. Nomenclature based on sand - silt - clay ratios. J.S.P. 24, 151-158.

SHEPARD, F. P., 1956. Marginal sediments of the Mississippi Delta. Am. Assoc. Pet. Geol. Bull. 40, 2537-2623.

..............., and LANKFORD, R.R., 1959. Sedimentary facies from shallow borings in the Lower Mississippi Delta. Am. Assoc. Pet. Geol. Bull. 43, 2051-2067.

..............., and MOORE, D.G., 1955. Central Texas Coast sedimentation: Characteristics of sedimentary environment, Recent history and diagenesis. Am. Assoc. Pet. Geol. 39, 1463-1593.

SHEPHERD, E.S., 1940. Note on the fluorine content of rocks and ocean-bottom samples. Am. J. Sc. 238, 117-128.

SHROCK, R.R., 1948. A classification of sedimentary rocks. J. Geol. 56, 118-129.

.............., 1948. Sequence in Layered Rocks. Harper, New York. 507p.

SIDWELL, R., and WARN, G.F., 1951. Pennsylvanian sedimentation in Northeastern Socorro County, New Mexico. J.S.P. 21, 3-11.

SIEVER, R., 1949. Trivoli Sandstone of Williamson County, Illinois. J.Geol. 57, 614-618.

.........., 1953. Petrology and sedimentation of the Upper Chester Sandstones. J.S.P. 23, 207-219.

..........., 1957. Pennsylvanian sandstones of the Eastern Interior Coal Basin. J.S.P. 27, 227-250.

............., and GLASS, H.D., 1957. Mineralogy of some Pennsylvanian carbonate rocks of Illinois. J.S.P. 27, 56-63.

SLOSS, L.L., 1947. Environments of limestone deposition. J.S.P. 17, 109-113.

SMITH, J.V., and YODER, H.S., 1956. Experimental and theoretical studies of the mica polymorphs. Min. Mag. 31, 209-235.

SMITH, S., 1912. Carboniferous Limestone of the North of England, Newcastle.

SMITHSON, F., and BROWN, G., 1954. The petrography of dickitic sandstones in North Wales and Northern England. Geol. Mag. 91, 177-188.

SMITHSON, F. and BROWN, G., 1957. Dickite from sandstones in northern England and North Wales. Min. Mag. 31, 381-391.

SMYTHE, J.A., 1914. On some inclusions in the Great Whin Sill in Northumberland. Geol. Mag. 4, 86-116.

.............., 1924. Minerals of the North Country: Silicates. The Vasculum, 10, 100.

STEVENS, R.E., and CARRON, M.K., 1948. Simple field tests for distinguishing minerals by abrasion pH. Am. Min. 33, 31-49.

STRAKHOV, N.M., 1953. Diagenesis in sediments and its significance in sedimentary ore formation. Izvest. Akad. Nauk. SSSR Ser. Geol., №5, 12-49.

................, 1957. Methodes d'etudes des roches sedimentaires. Moscou (trad. BRGGM), t. 1, 542p.

STUBICAN, V., 1959. Residual hydroxyl groups in the metakaolin range. Min. Mag. 32, 38-52.

TAGGART, M.S., Jr., and KAISER, A.D., Jr. 1960. Clay mineralogy of Mississippi River deltaic sediments. Geol. Soc. Am. Bull. 71, 521-530.

TALLMAN, S. L., 1949. Sandstone types: Their abundance and cementing agents. J. Geol. 57, 582-591.

TARR, W.A., 1927. Alternating deposition of pyrite, marcasite, and possibly melnikovite. Am. Min. 12, 417-421.

..........., 1928. Syngenetic pyritisation in local reducing areas of Pennsylvanian Shales in Missouri. J. Geol. 36, 434-439.

TAYLOR, J.H., 1952. Clay minerals and the evolution of sedimentary rocks. Clay Min. Bull. 1, 238-243.

TEALL, J.J.H., 1884. Petrological Notes on some North of England Dykes. Quart. Journ. Geol. Soc. 40, 209.

.............., 1884. On the Chemical and Microscopical Characters of the Whin Sill. Quart. Journ. Geol. Soc. 40, 640.

TORRENCE, M.R., 1955. Reference chlorite characterisation for chlorite identification in soil clays. 3rd Nat. Conf. Clays Clay Mins. 117-145.

TREFETHEN, J.M., 1950. Classification of sediments. Am. J. Sc. 248, 55-62.

TROTTER, F.M., 1951. Sedimentation facies in the Namurian of Northwest England and adjoining areas. Liv. and Manch. Geol. Jour. 1, 77-112.

.............., and HOLLINGWORTH, S.E., 1928. The Alston Block. Geol. Mag. 65, 433-448.

TUDDENHAM, W.M., and LYON, R.J.P., 1959. Relation of infrared spectra and chemical analyses for some chlorites and related minerals. Anal. Chem. 31, 377-380.

TWEN HOFEL, W.H., 1950. Principles of Sedimentation. 2nd Edition. McGraw-Hill, New York. 674p.

VAN DER HEIDE, S., 1950. Compaction as a possible factor in upper Carboniferous rhythmic sedimentation. Rept. 18th Int. Geol. Cong. part 4, 38-45.

VAN STRAATEN, L.M.J.U., 1960. Some recent advances in the study of deltaic sedimentation. Liv. and Manch. Geol. Jour. 2, 411-442.

......................., and KUENEN, P.H., 1958. Tidal action as a cause of clay accum~~m~~ulation. J.S.P. 28, 406-413.

VIVALDI, J.L.M. and MACEWAN, D.M.C., 1957. Triassic chlorites from the Jura and the Catalan Coastal range. Clay Min. Bull. 3, 177-183.

WAGER, L.R., 1928. A metamorphosed shale previously described as a "spotted" metamorphic rock. Geol. Mag. 65, 88-91.

WALKER, G.F., 1949. The decomposition of biotite in the soil. Min. Mag. 28, 693-703.

............., 1950. Trioctahedral minerals in the soil-clays of north-east Scotland. Min. Mag., 29, 72-84.

............., 1957. On the differentiation of vermiculites and smectites in clays. Clay Min. Bull. 3, 154-163.

WALKER, T.R., 1960. Carbonate replacement of detrital crystalline silicate minerals as a source of authigenic silica in sedimentary rocks. Geol. Soc. Am. Bull. 71, 145-152.

WALTON, E. K., 1955. Silurian greywackes in Peebleshire. Proc. Roy. Soc. Edin. 65, 327-357.

..............., 1956. Limitations of graded bedding and alternative criteria of upward sequence of rocks in the Southern Uplands. Trans. Geol. Soc. Edin., 16, 262-271.

..............., 1957. Two Ordovician conglomerates in Southern Ayrshire. Trans. Geol. Soc. Glasgow. 22, 133-156.

..............., and MUIR, R. O., 1957. The East Kirton Limestone. Trans. Geol. Soc. Glasgow. 22, 157-168.

WANLESS, H.R., 1955. Pennsylvanian rocks of the Eastern Interior Basin. Am. Assoc. Pet. Geol. Bull. 39, 1753-1820.

WANLESS, H.R., and SHEPARD, F.P., 1936. Sea level and climatic changes related to late Palaeozoic cycles. Geol. Soc. Am. Bull. 47, 1177-1206.

.............., and WELLER, J.M., 1932. Correlation and extent of Pennsylvanian cyclothems. Geol. Soc. Am. Bull. 43, 1004-1015.

WEAVER, C.E., 1955. Mineralogy and petrology of the rocks near the Quadrant-Phosphoria Boundary, in Southwest Montana. J.S.P. 24, 163-192.

.............., 1958. A discussion on the origin of clay minerals in sedimentary rocks. 5th Nat. Conf. Clays Clay Mins. 159-173.

.............., 1958. The effects and geologic significance of potassium "fixation" by expandable clay minerals, derived from muscovite, biotite, chlorite and volcanic material. Am. Min. 43, 839-861.

WEAVER, C.E., 1958. Geological interpretation of argillaceous sediments. Am. Assoc. Pet. Geol. Bull. 42, 254-309.

............., 1959. The clay petrology of sediments. 6th Nat. Conf. Clays Clay Mins. 154-187.

WEEKS, L.G., 1953. Environment and mode of origin and facies relationships of carbonate concretions in shales. J.S.P. 23, 162-173.

WEINTRAUB, J., and SAND, L.B., 1959. Alteration of micaceous minerals by sulfide solutions. 6th Nat. Conf. Clays Clay Mins. 369-377.

WELLER, J.M., 1930. Cyclical sedimentation of the Pennsylvanian Period and its significance. J. Geol. 38, 97-135.

............., 1956. Argument for diastrophic control of late Palaeozoic cyclothems. Am. Assoc. Pet. Geol. Bull. 40, 17-50.

WELLER, J.M., 1958. Cyclothems and larger sedimentary cycles. of the Pennsylvanian. J. Geol. 66, 195-207.

............., WHEELER, H.E., and MURRAY, H.H., 1958. Cyclothems: A discussion. Am. Assoc. Pet. Geol. Bull. 42, 442-452.

WELLS, A.J., 1955. The development of chert between the Main and Crow Limestones in North Yorkshire. P.Y.G.S. 30, 177-196.

.............., 1957. The stratigraphy and structure of the Middleton-Tyas-Sleightholme Anticline, North Yorkshire. Geol. Assoc. Proc. 68, 237-254.

..............., 1960. Cyclic Sedimentation: A Review. Geol. Mag. 97, 389-403.

WENTWORTH, C.K.,1922. A scale of grade and class terms for clastic Sediments. J.Geol. 30, 377-392.

WESTOLL, T. S., ROBSON, D.A. and GREEN, R., 1955. A guide to the geology of the district around Alnwick, Northumberland. P.Y.G.S. 30, 61-100.

WHEELER, H.E., and MURRAY, H.H., 1957. Base-level control patterns in cyclothemic sedimentation. Am. Assoc. Pet. Geol. Bull. 41, 1985-2011.

WHITEHOUSE, U.G., and McCARTER, R.S., 1958. Diagenetic modification of clay mineral types in artificial seawater. 5th Nat. Conf. Clays Clay Mins. 81-119.

WILLIAMS, H., TURNER, F.J. and GILBERT, C.M., 1954. Petrography. Freeman, San Francisco. 406p.

WOLF, K.H., 1960. Simplified limestone classification. Am. Assoc. Pet. Geol. Bull. 44, 1414-1416.

WOOD BASS, N., 1939. Verden Sandstone of Oklahoma, an exposed shoestring sand of Permian Age. Am. Assoc. Pet. Geol. Bull. 23, 559-581.

YODER, H.S., 1959. Experimental studies on micas: A synthesis. 6th Nat. Conf. Clays Clay Mins. 42-60.

YODER, H.S., and EUGSTER, H.P., 1954. Phlogopite synthesis and stability range. Geochim. Cosmochim. Acta. 6, 157-185.

.................................., 1955. Synthetic and natural muscovites. Geochim. Cosmochim. Acta. 8, 225-280.

*BANNISTER, F.A., 1942. Brammalite - A new sodium illite mineral from Llandebie, South Wales. Min. Mag. 26, 304-307.